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The structure and dynamics of the lithosphere beneath Tibet from seismic surface-wave analysis

by
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A dissertation submitted to the University of Dublin, Trinity College
for the degree of Doctor of Philosophy

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and

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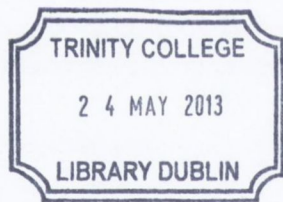
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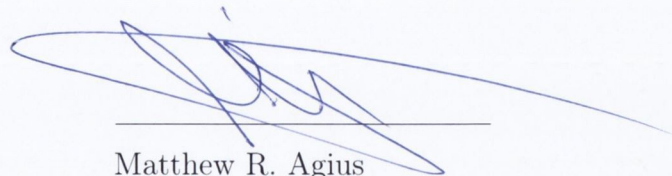
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Matthew R. Agius

January 14, 2013.

Summary

Despite the numerous studies undertaken to investigate the underlying structures beneath the Tibetan Plateau, fundamental questions about the mechanism of lithospheric convergence between India and Asia (establishing if convergence is from: underthrusting, north/south subduction, lithospheric thickening, or convective removal), and, the internal lithospheric dynamics within Tibet (rigid block or continuous deformation), persist until now. This study aims to determine which of the different mechanisms are at work beneath Tibet, keeping in mind that elements of different mechanisms may be present beneath different parts of the plateau.

The data set is broad-band Rayleigh- and Love-wave phase-velocities measured from pairs of station across the plateau. A series of rigorous analyses is performed on the data: (1) direct comparison of inter-station dispersion, (2) inversion for 1-D radially anisotropic S -velocity models, (3) inversion for 1-D azimuthally anisotropic S -velocity models (4) modelling of regional averages, and (5) tomographic models. All of these techniques yielded consistent results regarding shear-velocity structure and anisotropic properties of the crust and upper mantle beneath Tibet.

The Tibetan crust has expected low shear velocities, however, with strong north-south variations across the plateau. Mid-crustal LVZ (and inferences of melt and low viscosity) and high conductivity are found across large areas of northern Tibet. The consistent lateral distribution of the data and the coherent pattern of anisotropy within the northern regions, strongly suggest that crust is homogeneous and that deformation is diffused across large areas. Strong radial anisotropy is observed beneath western Tibet and Yunnan, both regions experiencing extension (flattening), whereas north-eastern Tibet has very weak radial anisotropy, strike-slip fault mechanisms, and no extension. The ongoing crustal thinning in the west probably causes the anisotropic mica crystals to become near-horizontally oriented (*Shapiro et al.*, 2004), whereas in the east, horizontal pervasive flow may align micas in the vertical plane, resulting in weak or absent radial anisotropy. The flow direction, derived from crustal azimuthal anisotropy, show W–E and NW–SE fast directions in central and eastern Tibet, respectively. Special focus is given to north-eastern Tibet, where the inferred fast directions are aligned southeast rather than

northeast, as would be expected from an elevation-gradient induced flow (*Clark and Royden, 2000*). The fast azimuths are parallel to the extensional component of the current strain rate across Tibet, strongly suggesting similar deformation through the entire crust. Despite the mid-crust's greater susceptibility to deformation and flow, the correlation of azimuthal anisotropy with surface strain indicates that the mid-crust still holds some degree of coupling with the adjacent layers. The close agreement of anisotropy and extension component of strain with the traces of sutures implies that the dominant deformation mechanism within the plateau has not changed since initiation of continental collision and is still governed by the northward push of India.

The upper 75 km of the mantle beneath Tibet is made up of an Indian lithosphere in the west and southwest and a Tibetan lithosphere and asthenosphere elsewhere. Strong, cold, cratonic Indian lithosphere underthrusts southwestern Tibet (up to the BNS at 85°E), and warm Tibetan lithosphere and asthenosphere lay further north, up to the Kunlun Fault. The Tibetan lithosphere and asthenosphere have low-average S -velocities, indicative of warmer temperatures. Although the finer structure of the Tibetan lithospheric mantle remains hard to resolve using surface waves alone, a thick layer of low-average V_S in the uppermost mantle is difficult to explain from high temperatures generated by crustal radioactivity and reconcile with the presence or formation of a thick cratonic lithosphere at depths down to 200 km (*McKenzie and Priestley, 2008*). Surface-wave data can fit a series of other seismic observations such as S_n , P_n , and a shallow LAB discontinuity, which together add support for a thin Tibetan lithosphere underlain by an asthenosphere. The dynamics of the asthenosphere is revealed by azimuthal anisotropy beneath central Tibet, characterised by fast SSW–NNE direction. The amplitude of the anisotropy increases from south to north and is parallel to the direction of India's plate motion, suggesting that asthenospheric flow is pushed outward by India's northward subduction.

Cold Indian lithosphere subducts beneath the Tibetan asthenosphere under the central and eastern plateau. The lithospheric convergence mechanisms varies from west to east; steep-angle subduction of India beneath west-central Tibet and shallow-angle subduction of India in eastern Tibet, with the subducting Indian lithosphere reaching as far north as northern Qiangtang–Songpan–Ganzi Terrane.

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*Dedicated to my fiancée Louise
and to my parents Angelo and Elizabeth*

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List of Abbreviations

1-D	1 Dimensional
2-D	2 Dimensional
3-D	3 Dimensional
AMI	A utomated M ultimode waveform I nversion
ATF	A ltyn T agh F ault
BNS	B angong- N ujiang S uture
GPS	G lobal P ositioning S ystems
HI-CLIMB	H imalayan-Tibetan C ontinental L ithosphere during M ountain B uilding
INDEPTH	I Nternational D Eep P rofilng of T ibet and the H imalaya
JRS	J insha R iver S uture
KF	K unlun F ault
LAB	L ithosphere A sthenosphere B oundary
LVZ	L ow V elocity Z one(s)
Moho	M ohorovičić discontinuity
MT	M agneto T elluric
My	M illion y ears
P	P rimary (compressional) seismic body wave
s	s econds
S	S econdary (transverse) seismic body wave
SB	S ichuan B asin

V_P	P -wave velocity
V_S	S -wave velocity
V_{SH}	Horizontally polarised S -wave velocity
V_{SV}	Vertically polarised S -wave velocity
YZS	Y arlung- Z angbo S uture

1

Introduction

Tibet, the largest and highest plateau on Earth (*Fielding et al.*, 1994), is sandwiched between the northward moving India (*Wang et al.*, 2001) and relatively stable Eurasia. This spectacular structure (Figure 1.1) was formed by successive collisions of different micro-continents with the southern margin of Eurasia and subduction of oceans between them since the early Paleozoic (e.g., *Allègre et al.*, 1984; *Searle et al.*, 1987; *Dewey et al.*, 1988; *Murphy et al.*, 1997; *Yin and Harrison*, 2000; *Zhang et al.*, 2007). The last in the series of accretion events is the collision of the India continent with Eurasia that started at ~ 55 –34 million years ago (e.g., *Besse et al.*, 1984; *Patriat and Achache*, 1984; *Searle et al.*, 1987; *Garzanti and Van Haver*, 1988; *Rowley*, 1996, 1998; *Aitchison et al.*, 2007; *Green et al.*, 2008, or perhaps even more recently at 25–20 My ago, *van Hinsbergen et al.* (2012)). India's convergence rate with Asia has decreased as it underthrusts southern Tibet (e.g., *Powell and Conaghan*, 1973; *Molnar and Tapponnier*, 1975; *Armijo et al.*, 1986; *Ni and Barazangi*, 1984). It is estimated that India and Asia converged by up to 3,600 km since 52 My (*van Hinsbergen et al.*, 2011), with at least 1,400 km of

north-south shortening absorbed by the Tibetan orogen (*Yin and Harrison, 2000; Guillot et al., 2003*).

The relatively smooth topography of the plateau blankets complex underlying structure. Over the last century, various models of the Tibetan continental collision mechanism have been proposed, which invoke one or more of the following basic modes: (1) underthrusting of India beneath entire Tibet (e.g., *Argand, 1924*), or (2) beneath southern Tibet only (e.g., *Ni and Barazangi, 1984; Owens and Zandt, 1997; Nábelek et al., 2009*), (3) variants of lithospheric subduction, such as steep-angle northward Indian subduction (e.g., *Tilman et al., 2003; Li et al., 2008*), shallow-angle northward subduction (e.g., *Friederich, 2003; Priestley et al., 2006; Zhou and Murphy, 2005*), or southward subduction of Asia’s lithosphere (e.g., *Willett and Beaumont, 1994; Tapponnier et al., 2001; Kumar et al., 2006; Zhao et al., 2010*), (4) viscous thickening of the lithosphere (*McKenzie and Priestley, 2008*), and (5) convective removal of a thickened lithosphere (e.g., *Houseman et al., 1981; Molnar et al., 1993; Hatzfeld and Molnar, 2010*).

Strong debate also revolves around the internal deformation within the Tibetan Plateau. Two main end-member models include: “rigid blocks”, where discrete tectonic blocks slide past each other with relatively little internal deformation (e.g., *Tapponnier et al., 1982; Avouac and Tapponnier, 1993*), or “continuous deformation”, where the deformation is diffuse and distributed broadly (e.g., *Dewey and Burke, 1973; England and Houseman, 1986; England and Molnar, 1997a*). Processes within the deep crust and upper mantle play a key role in the plateau dynamics. These processes can now be revealed in increasing detail by seismic observations, thanks to the recent deployments of numerous seismic stations across the plateau (e.g., *Yang et al., 2010, 2012*).

This study aims to determine which of the aforementioned mechanisms are at work beneath Tibet, keeping in mind that elements of each mechanism may be present beneath different parts of the plateau. The main data set is formed by inter-station Rayleigh and Love surface-wave dispersion measurements, determined from a selection of stations across Tibet and then inverted for one-dimensional (1-D) shear-velocity profiles. Shear velocities are sensitive to temperature and composition and can offer important information on the state of the rocks at depth. The fundamental advantage of inter-station measurements compared to the source-

station measurements, normally used in large-scale tomography, is that they can be performed in broader frequency bands. Furthermore, the use of local measurements brings the advantage of their simple, clear relationship to the local Earth structure. The inferred 1-D profiles are free from lateral smoothing that characterises 3-D models, and hence are used to resolve trade-offs between model parameters in the crust and in the upper mantle and to determine seismic structure within both with higher accuracy.

In this thesis, Chapter 2 provides general background on surface-wave analysis. Chapters 3, 4 and 5 that follow are manuscripts prepared for submission in peer-reviewed journals, with each manuscript targeting one aspect of the problem. In Chapter 3 the Tibetan upper mantle structure is under focus; broad-band data is used to constrain the 1-D models from the Moho down to 300 km depth. A series of rigorous tests are designed to establish the velocity ranges of low and high-velocity anomalies, detected on top of each other, and to establish estimates of uppermost mantle radial anisotropy. The configuration of the anomalies and estimates of temperatures within them lead us to conclusions on the origin of the anomalies and on the complex morphology of the subducted Indian lithosphere in the upper mantle beneath Tibet.

In Chapter 4, phase-velocity data and 1-D shear-velocity profiles are used to constrain the structure of the Tibetan crust. Robust ranges of shear velocity and radial anisotropy are determined using the short-period band of the data. Melt fraction estimates based on the mid-crustal shear-velocities indicate regions with low viscosity that are prone to flow. The internal deformation is analysed from the comparison of fault systems and earthquake focal mechanisms, characterising surface deformation, with the lateral distribution of radial anisotropy, characterising the deformation in deep crust.

Finally, in Chapter 5, a thorough investigation is carried out on the patterns of azimuthal anisotropy, inferred from analysis of various seismic data types: shear-wave birefringence, receiver functions, and surface waves. The lack of agreement between published models prompted a detailed study on the azimuthal dependence of the dispersion curves. Various methodologies are adopted to establish the amplitude and fast-propagation direction at different periods, such as two-dimensional tomography, regional averages, direct comparison of azimuth-dependent curves,

and inversion for azimuthally anisotropic shear-velocity models — all of which show consistent patterns. Multi-layered azimuthal anisotropy is established in central and northeastern Tibet, indicative of distinctly different patterns of flow in the lithosphere and asthenosphere.

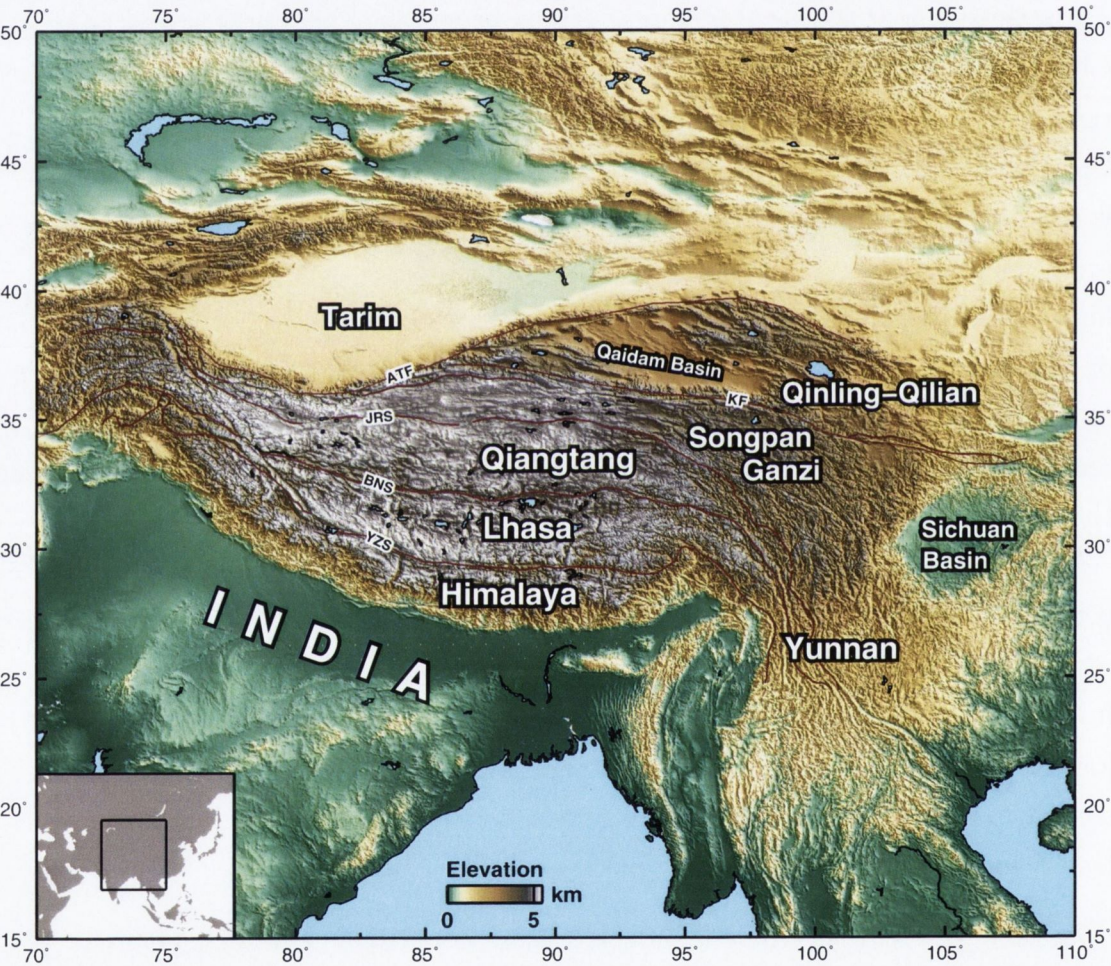


Figure 1.1: Map of Eurasia showing relative elevation variations. Topographic map highlighting the dramatic elevation changes; Green shade indicates low lying regions; Grey shade indicates very high elevation. Black lines are coastlines of oceans, lakes, and rivers. Dark red lines are sutures and major faults across Tibet and surrounding regions: YZS (Yarlung-Zangbo Suture), BNS (Bangong-Nujiang Suture), JRS (Jinsha River Suture), KF (Kunlun Fault) and ATF (Altyn Tagh Fault), which divide Tibet in major terranes. Inset shows location on a regional map of Asia.

2

Seismic surface-wave analysis

2.1 Seismic surface waves

When an earthquake occurs, seismic waves radiate away from the hypocentre in all directions. There are two main types of waves: body and surface waves. Whilst body-waves travel deep inside the Earth, surface waves are confined to outer layers. Body waves consist of compressional waves, popularly known as primary (P) waves, and shear waves, also known as secondary (S) waves. The two types of surface-waves are Love wave, with particle motion perpendicular to the great-circle plane, and Rayleigh wave, with particle motion within the great circle plane. Love waves depend on the horizontally polarised shear waves propagation, whereas Rayleigh waves depend on the vertical polarised shear waves and P waves propagation. The different nature of propagation and speed of these waves result in distinct seismic phases arriving at different times at a seismic station; the P body wave is recorded first followed by S , Love, and Rayleigh surface waves (Figure 2.1).

Surface-wave propagation is dispersive; the waves travel at different velocities

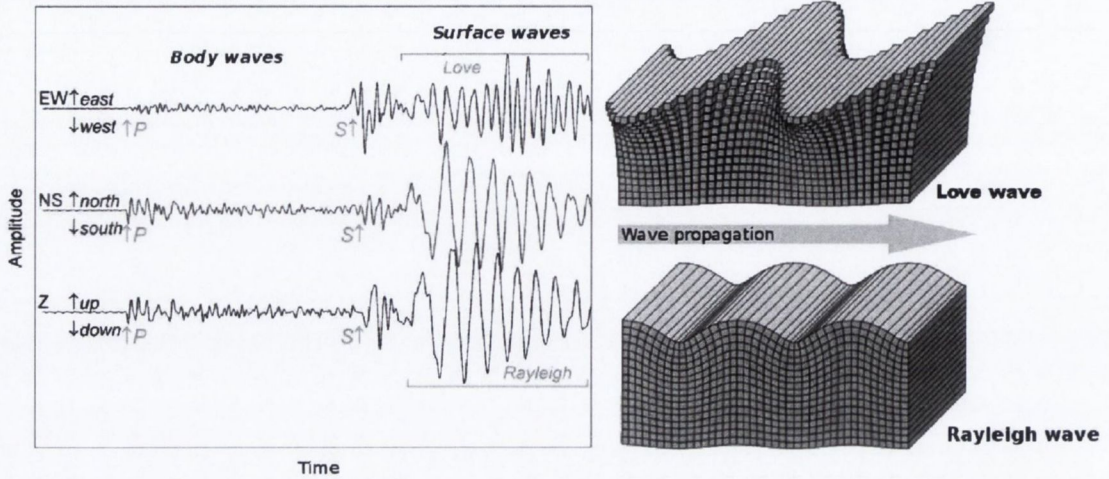


Figure 2.1: Seismogram of a teleseismic earthquake and illustration of surface-wave propagation. Left: A seismogram showing the recorded oscillations of a teleseismic earthquake in three components (two horizontals: east-west and north-south, and vertical: up-down). The seismogram shows two types of waves: body (P and S) and surface wave (Love and Rayleigh). Image adapted from <http://web.ics.purdue.edu/~braile/> accessed on the 22nd October 2012. Right: Illustration of surface-wave propagation. Love-waves oscillate in the horizontal domain; Rayleigh-waves oscillate in the vertical domain. Image adapted from *Introduction to Seismology* (figure 8.4), by Peter M. Shearer, 1999.

for different frequencies. This frequency-depth dependence of the waves makes it possible to infer the shear velocity as a function of depth, and hence the structure below. In this study, the seismograms of teleseismic earthquakes are used to determine surface-wave dispersion across Tibet. Efforts are made to acquire accurate broad-band data in order to have the ability to determine shear-velocity structures across a broad depth range, from the shallow crust down through to the base of the upper mantle.

2.2 Data retrieval

In the last two decades there has been a rapid increase in the deployment of temporary and permanent seismic stations across the Tibetan Plateau and the surrounding regions. The data from most of these instruments is in the public domain and available from the Incorporated Research Institutions for Seismology (IRIS) Data Management Center (DMC) (<http://www.iris.edu/dms/dmc/>). For

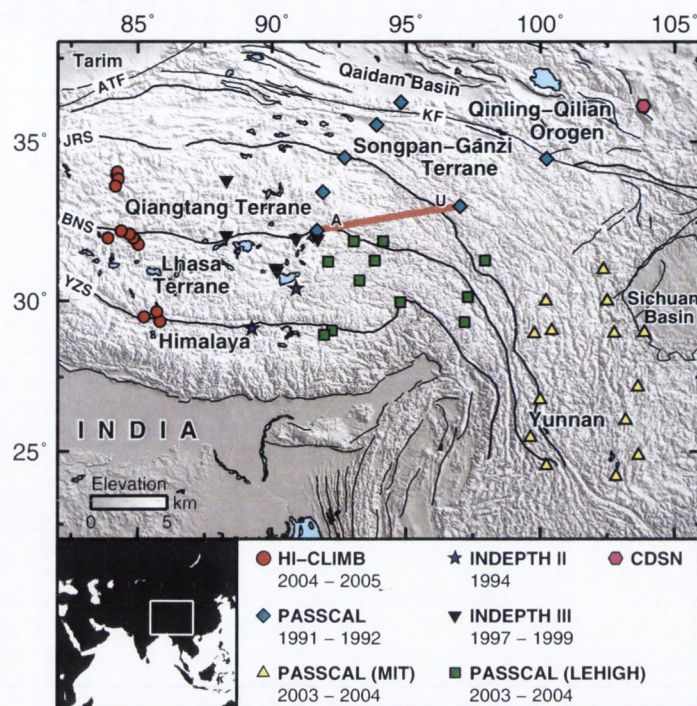


Figure 2.2: All the seismic stations used for the inter-station measurements. Different symbols correspond to different seismic networks operated at different times. Red path: Paired stations AMDO (A) and USHU (U). Black lines: Sutures and faults: YZS (Yarlung-Zangbo Suture), BNS (Bangong-Nujiang Suture), JRS (Jinsha River Suture), KF (Kunlun Fault) and ATF (Altyn Tagh Fault). Grey shade highlights the elevation. Bottom left map: The location of the map region in Asia.

a requested region and time window, the DMC provides a list of networks, station names, station coordinates, dates of operation, sampling streams (channels), channel components, location code, station description, and the data availability (public or restricted).

Pairs of stations with paths crossing the same tectonic blocks or similar terranes, and with both stations having a suitable number of same earthquake recordings that occurred along the pair's great circle path were selected. In total 56 stations, which coincidentally make up 56 station pairs, were used. The stations are broadband instruments belonging to various temporary networks deployed over the years: PASSCAL 1991–1992 (*Owens et al.*, 1993), INDEPTH II 1994 (*Nelson et al.*, 1996), INDEPTH III 1997–1999 (*Huang et al.*, 2000), HI-CLIMB 2004–2005 (*Nábělek et al.*, 2005), PASSCAL (Lehigh) 2003–2004 (*Sol et al.*, 2007), PASSCAL

(MIT) 2003–2004 (*Lev et al.*, 2006), and from the permanent China Digital Seismic Network (Figure 2.2). A total of 959 global earthquakes that occurred between 1991–2005 were used. These have magnitudes between 5 and 9, and epicentral distances to the stations between 4.6° and 150° .

2.3 Data processing

Seismograms at 1 sample per second are sufficient to make the required measurements; the north, east, and vertical components of stream L were the first preference for the data retrieval. When this stream was unavailable higher sample rates such as 20 or 100 (channels B and H) were selected and down sampled to 1 sample per second.

Several automated quality checks and pre-processing steps are performed on the data. Seismograms are first checked for their integrity such as incomplete or erroneous data, data gaps, incorrect timing, and clipping. The waveforms are then corrected for instrument response to produce displacement seismograms. Finally, the horizontals are rotated to represent the radial and transverse components.

2.4 Measurement of surface-wave phase velocity

Rayleigh- and Love-wave fundamental-mode dispersion are measured using the well established two-station method (*Sato*, 1955; *Knopoff*, 1972; *Meier et al.*, 2004; *Endrun et al.*, 2008; *Deschamps et al.*, 2008; *Yao et al.*, 2006, 2010; *Yang et al.*, 2010). In this study, broad-period phase velocities between two stations are obtained using two independent approaches: cross correlation of two seismograms (station-station) and derived inter-station phase velocities from source-station waveform fitting. In order to reduce effects from unmodelled surface-wave diffraction, the selected earthquakes have epicenters closely aligned to the great-circle path between the two stations, less than 10° difference to the back azimuth (azimuth towards the earthquake epicentre).

2.4.1 Cross correlation of seismograms

Inter-station phase velocities are determined by analysing the phase difference of teleseismic surface waves. Assuming a Jeffreys-Wentzel-Kramers-Brillouin (JWKB) approximation for a spherically symmetric isotropic Earth model, the phase velocity $c(\omega)$ is obtained from the phase of the cross-correlation function $\Phi(\omega)$ of the vertical component recordings of the same earthquake at two stations with the same back azimuth (*Meier et al.*, 2004):

$$c(\omega) = \frac{\omega(\Delta_1 - \Delta_2)}{\arctan\{\Im[\Phi(\omega)]/\Re[\Phi(\omega)]\} + 2n\pi}$$

where Δ_1 and Δ_2 are the epicentral distances to the two stations.

Figure 2.3 is an example of an inter-station phase-velocity measurement using cross-correlation between the stations AMDO and USHU (their location is indicated in Figure 2.2). The vertical component waveform for each is shown along with their corresponding spectrum showing amplitudes as a function of frequency and time obtained from multiple Gaussian filters. Assuming that the fundamental mode has the largest amplitude on the seismogram, the group travel time of the Rayleigh-wave fundamental mode for the different frequencies is identified by the highest amplitudes in the spectrum (solid white lines). The cross-correlation function is also plotted with its corresponding amplitudes as a function of frequency and time. Strong, high-frequency scattering and noise are easily noticed as correlated patches at different times in the cross-correlation spectrum. Effects of noise and interference are reduced by filtering the cross-correlation function with a frequency-dependent Gaussian bandpass filter. The signal-to-noise ratio is enhanced by applying frequency-dependent time windows to the filtered cross-correlation function, consequently decreasing the effects of side lobes likely to be caused by correlation of the fundamental mode with higher modes or other scattered waves. The cross-correlation function is then transferred into the frequency domain and the phase velocity is computed from the phase of the cross correlation $\Phi(\omega)$.

The array of phase-velocity curves from the $2n\pi$, ($n \in N$), is shown in Figure 2.3 (top right, blue curves). The synthetic Rayleigh-wave fundamental-mode dispersion curve of a global model is plotted as a visual reference (thin dashed grey

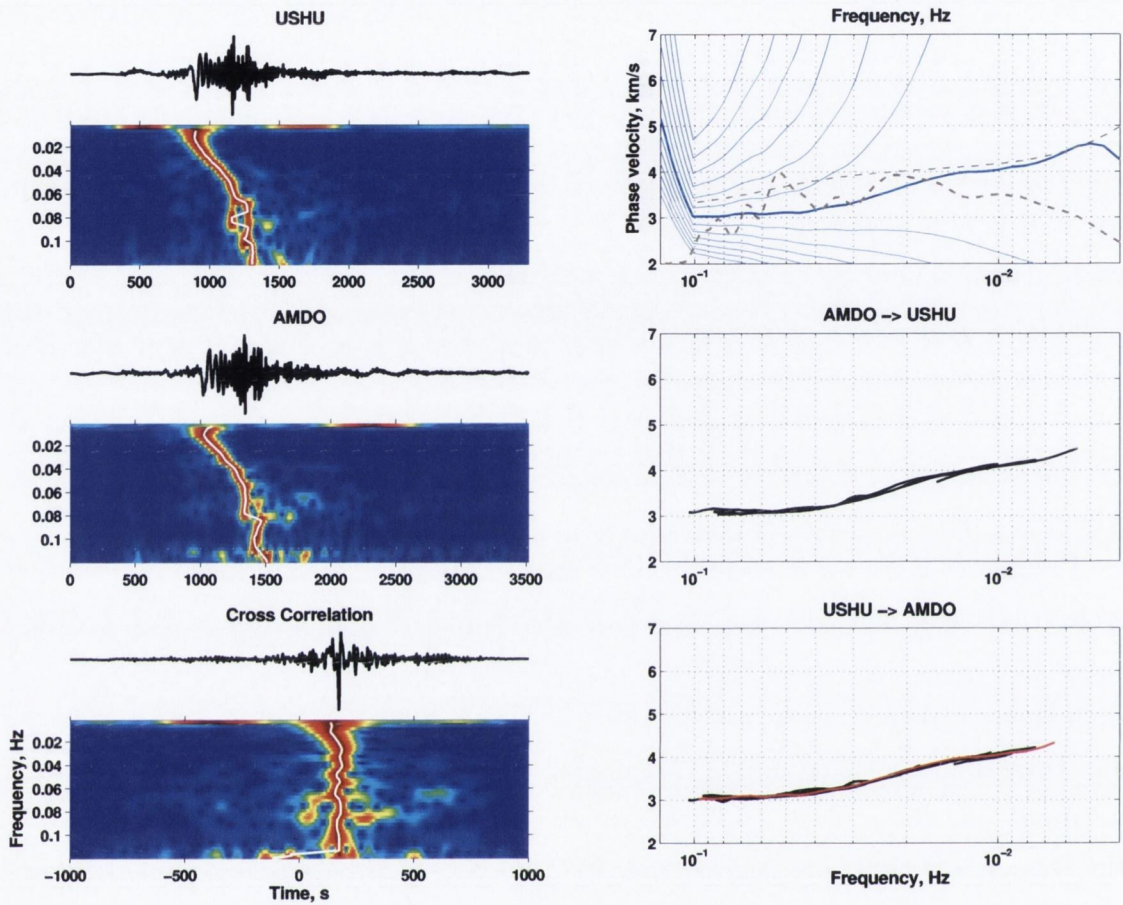


Figure 2.3: Phase-velocity measurements from the two-station cross-correlation method of paired stations AMDO and USHU (Figure 2.2). Detailed description in *Meier et al. (2004)*. Top left: Vertical seismograms of an earthquake recorded at both stations (black waveforms) and the corresponding amplitudes as a function of time and frequency calculated using multiple filters. Solid white lines in the time-frequency frame is the automatically determined group travel times of the Rayleigh-wave fundamental mode. Bottom left: The cross-correlation function of the two waveforms with applied frequency-dependent filters and time windows. Top right: Solid blue lines: The series of phase-velocity curves calculated from the phase of the cross-correlation function. Thin dashed grey curve: The synthetic fundamental-mode Rayleigh-wave dispersion computed from a global model, plotted to act as a guide in the manual selection of the accepted phase velocities. Thick dashed grey curve: The normalized amplitude spectrum of the cross correlation function. In an interactive process the user selects the phase-velocity curve within a frequency range. Lower right frames: Selected dispersion curves (black) sorted according to the wave propagation direction. Red curve: the current selected phase-velocity.

curve) to aid in the manual selection of the accepted phase velocity. Due to the very thick crust and different temperature and composition beneath Tibet, phase-velocities sampling through the plateau are expected to be low. This is evident from the curve that has the velocities close to the reference for the broadest period range (bold blue curve). Only smooth portions of the curve are selected (frame below, red curve) as these are the least likely to be contaminated with diffraction or with interference of different modes that usually result in narrow-band roughness. In order to make the selection of the curve less subjective, the user is assisted with the plot of the normalized amplitude spectrum of the cross-correlation function (thick dashed grey curve) indicating the frequency range that is most constrained by the data. Measurements made from earthquakes occurring on opposite sides of the inter-station paths have similar dispersion (shown separately in Figure 2.3 bottom right frames).

The advantage of this technique is that measurements are successful even when waveforms appear complex. On the other hand, long-period surface waves can interfere with body waves and distort the measurements made from the cross-correlation function. The dispersion of such waves are better measured using full waveform fitting. Love-wave dispersion is determined using an analogous procedure, however with the transverse components.

2.4.2 Automated multimode waveform inversion

The Automated Multimode Inversion (AMI) by *Lebedev et al.* (2005) is a waveform fitting technique utilizing synthetic seismograms to simultaneously constrain Earth structure from information contained in S , multiple S , and surface waves. In order to account for the Earth's heterogeneity along the source-station propagation path, AMI utilises a 3-D reference model, the construction of a series of complex time-frequency windows, and a strict misfit criteria to generate accurate waveform fits. Adopting the JWKB approximation, the synthetic seismogram is computed as:

$$s(\omega) = \sum_m A_m(\omega) \exp[i\omega\Delta / (C_m^0(\omega) + \overline{\delta C_m(\omega)})]$$

where the sum is over modes m , and Δ is the epicentral distance. $A_m(\omega)$ is the frequency-dependent, complex amplitude of the modes which depends on the source

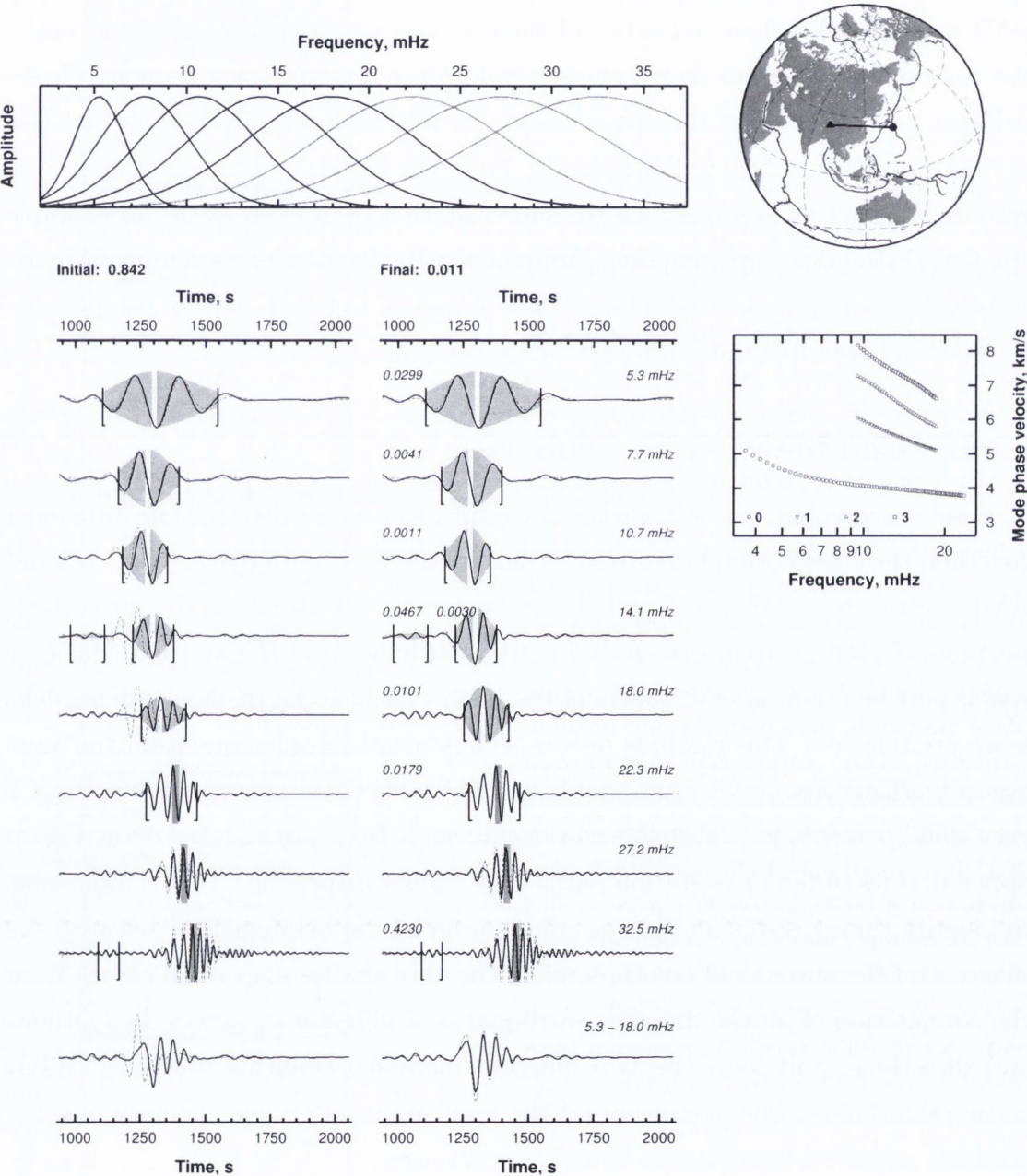
mechanism, radiation pattern, and assumed Earth model. Reference phase velocities $C_m^0(\omega)$ and their Fréchet derivatives $\delta C_m^0(\omega)/\delta\beta(r)$ within an approximate sensitivity area are extracted from the 3-D reference model, and averaged together to represent the structure along the source-station path. Figures 3.6, 4.4 and 5.4 show examples of Fréchet derivatives for surface-wave phase velocity, with respect to shear speed and depth. The perturbations needed for the synthetic waveforms to fit the data are controlled through the relationship of the Fréchet derivatives with the perturbation in S velocity, $\delta\beta(r)$:

$$\overline{\delta C_m(\omega)} = \int_0^a \frac{\delta C_m^0(\omega)}{\delta\beta(r)} \overline{\delta\beta(r)} dr$$

where a is the radius of the Earth. AMI undergoes a rigorous, automated procedure for each seismogram, with a case-by-case selection of weighted time-frequency windows using different Gaussian filters. The final waveform fit is generated through simultaneous non-linear inversion of all time-frequency windows, while enforcing a strict misfit criteria of 5% in each.

An example of a waveform fit is shown in Figure 2.4 for an earthquake recorded at station AMDO. The vertical seismogram shown at the bottom of the waveform columns (the same for the initial and final fit) is filtered using eight Gaussian band-pass filters, with peaks between 5.3 to 32.5 mHz (shown in the top frame). The

Figure 2.4 (facing page): An example of automated multimode waveform inversion for an earthquake recorded at station AMDO. Detailed description in *Lebedev et al.* (2005). Top right: Event and station location on a map and their corresponding great-circle path. Top left: Eight Gaussian filters used to filter the seismogram. Left waveform column: Initial fit of filtered waveforms (solid lines) with synthetics (dashed lines) computed from a 1-D background model. The time-frequency windows are indicated by half-brackets and by shading of the signal envelope. Fundamental-mode wave packets are identified by a white vertical line at the maxima of the envelope. At the bottom are the band-pass filtered (using those frequencies successfully fit) seismograms. Right waveform column: Same as left but for the final fit, minimized through path-averaged perturbations in $\overline{\delta\beta(r)}$. The values of the misfit are given next to each window; initial and final global misfits are given above the waveform columns. Right graph: Phase velocities for the fundamental and higher modes extracted from the solution of the final waveform fit.



initial misfit of the synthetic waveform to the data is clearly seen as the discrepancy between the dashed and solid waveforms, at the bottom band-pass filtered seismograms, or at each filter frequency. In this example, the seismogram was fit using 5 band pass filters (5.3–18.0 mHz) and 4 modes. After waveform fitting is complete, AMI measures the phase velocities of those modes which contribute significantly. An example of these are shown at the right. Only the fundamental-mode phase-velocity curve is used in this dissertation. In this example, the dispersion has a period range between 40 to 280 seconds. From the successful fit of the same event recorded at another station along the source-station great circle path (for example station USHU), the corresponding inter-station Rayleigh-wave fundamental-mode dispersion is computed. The same method is applied to the transverse components in order to determine the Love-wave phase velocities.

2.4.3 Combined measurements

The measurements from both methods complement each other and are combined together; the cross-correlation produces most of the short-period measurements and AMI contributes to the long-period measurements. In an interactive procedure, portions of each curve are reselected; curves outlying from the average, kinks, or rough portions (not smooth) parts of the curves likely to be frequency-dependent noise are rejected. This rigorous process is important in order to obtain the most accurate dispersion curves that enable us to observe patterns and differences of a very small percentage ($<1\%$). Measurements made from earthquakes occurring on opposite sides of the inter-station paths have similar dispersion. This is important to identify timing errors and biases arising from lateral heterogeneity between the source and the stations of one direction. The very similar dispersion curves from the compilation of all the different earthquakes of different distances, magnitudes and directions, and from the two different methods, confirms that the JWKB assumption holds. The average of all the measurements produce smooth, robust, accurate phase velocities across broad period ranges.

Figure 2.5 shows an example of combined, superimposed inter-station measurements from AMI and cross correlation. Each individual dispersion curve for Love and Rayleigh waves is colour coded to represent the technique and the source direction. Most of the earthquakes are from western Pacific. The short period mea-

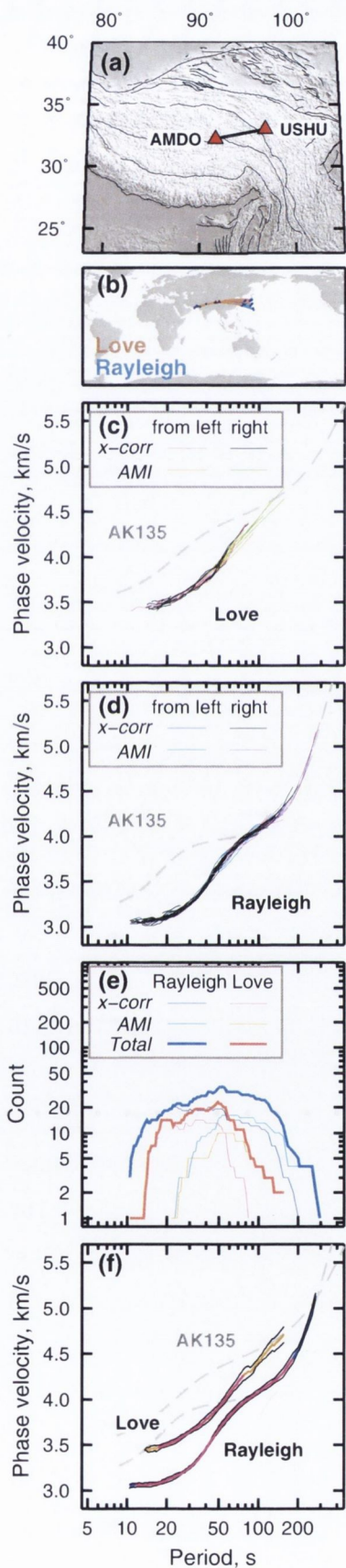


Figure 2.5: Combined inter-station pair phase-velocity measurements. Top map (a) shows the location of the station pairs (red triangles), and the corresponding inter-station paths (black path). The global map (b) shows earthquakes (dark blue dots) and their corresponding great circle paths aligned along the station paths which produced the Love- (brown) and Rayleigh-wave (cyan) measurements. Love- and Rayleigh-wave dispersion curves are shown in separate plots (c, d) each showing superimposed measurements. Different colours indicate the method and direction each measurement belongs to. AMI: Measurements from the Automated Multimode Inversion technique (*Lebedev et al., 2005*). x-corr: Measurements from cross-correlation (*Meier et al., 2004*). Dashed grey curves show the phase-velocities computed for the AK135 global reference model (*Kennett et al., 1995*). Count graph (e) shows the total number of measurements at each period for Love and Rayleigh wave (red and blue respectively); different colours show the count from different methods. The bottom panel (f) shows the average dispersion curves for Love and Rayleigh waves and their standard deviations; robust phase-velocities selected for further processing are highlighted in pink.

surements (<50 s) are mostly produced by cross correlation, whereas the longer periods are from AMI. In general, Rayleigh-wave measurements outnumber Love-wave measurements as can be seen in the count panel. In some cases, measurements exceed well over 200 seconds (Figures 2.5, 3.4, and 4.2). The average and the standard deviation of all the measurements at each period are calculated (bottom frame). Only the smooth portions of the average Love and Rayleigh-wave dispersion curves are selected for further processing (pink curve, bottom frame). Note the very smooth, broad-band (10–200 s) phase-velocity curves.

The dispersion curves are much slower than the synthetic dispersion curves computed for the AK135 global-average reference model (*Kennett et al.*, 1995). This is typical for Tibet where short-period phase velocities (<70 s) are low due, primarily, to a thicker crust. At longer periods, though the variation in the dispersion curves is less dramatic, small differences away from the reference are indicative of deep structural anomalies.

2.4.4 Phase-velocity distribution

The phase-velocity measurements at different periods and along different paths reveal first-order indications of anomalies across the Tibetan Plateau. Figure 2.6 shows phase-velocity variations at 20 s period for Rayleigh and Love waves. Rayleigh-wave phase velocities of paths within the plateau have relatively low velocities compared to paths sampling across lower elevation regions, such as Yunnan, Sichuan Basin and Qinling-Qilian. Minor differences between paths within the high plateau could either be due to shallow crustal heterogeneity or anisotropic properties. For example, there seems to be a systematic pattern of west-east aligned paths that tend to be faster than north-south aligned paths. Love-wave phase velocities at the same period have more variation, and the relationship of the velocities with structure, particularly within the high plateau, is less clear. Interestingly, the Love-wave phase velocities across Songpan-Ganzi, Qinglin-Qilian, and Sichuan Basin have relative anomalies similar to the Rayleigh-wave dispersion, but, in western and southern Tibet (including Yunnan), Love-wave phase velocities are relatively faster to those of Rayleigh wave.

The phase velocities at different period ranges will be investigated in the following chapters. In order to determine the crustal and mantle structure beneath

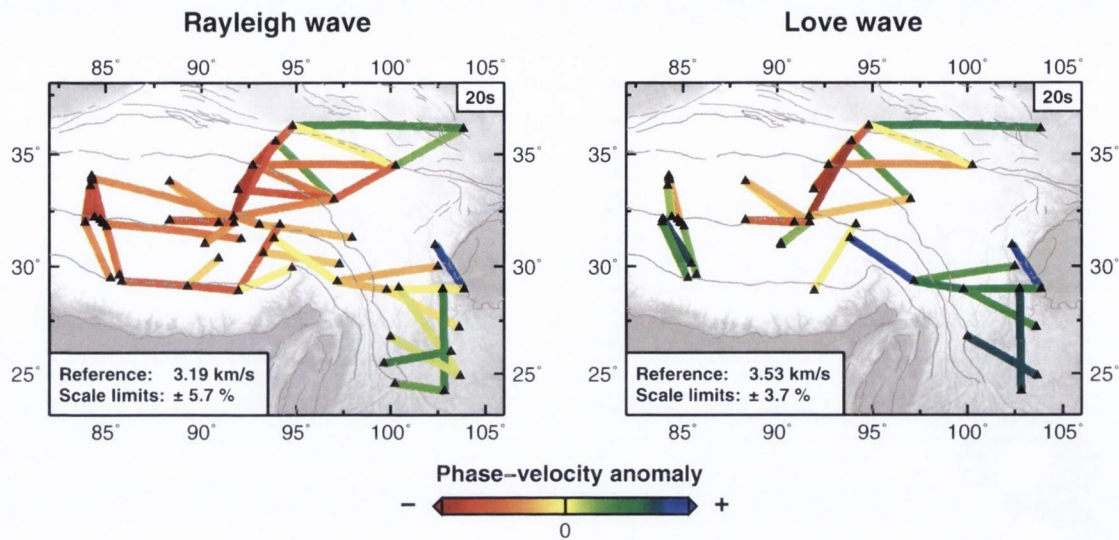


Figure 2.6: Rayleigh- and Love-wave fundamental-mode phase-velocity variations across the Tibetan Plateau. Coloured lines show the measured paths between stations (black triangles). Different colours correspond to the phase-velocity difference from the reference velocity within the scale limits at 20 seconds period.

the paths, an inversion of phase velocity for shear velocity with depth is needed. Radial and azimuthal anisotropy are taken into account, as these are important to deduce constraints on the dynamics of the plateau.

3

Tibetan and Indian lithospheres in the upper mantle beneath Tibet

3.1 Introduction

The uplift and evolution of the Tibetan Plateau (Figure 3.1) has long been recognised to be a result of India's collision with Asia (*Argand*, 1924; *Molnar and Tapponnier*, 1975). The plateau has been assembled of a number of tectonic blocks, now separated by narrow sutures (*Allègre et al.*, 1984). Deformation of these blocks associated with India-Asia convergence started already during the Creta-

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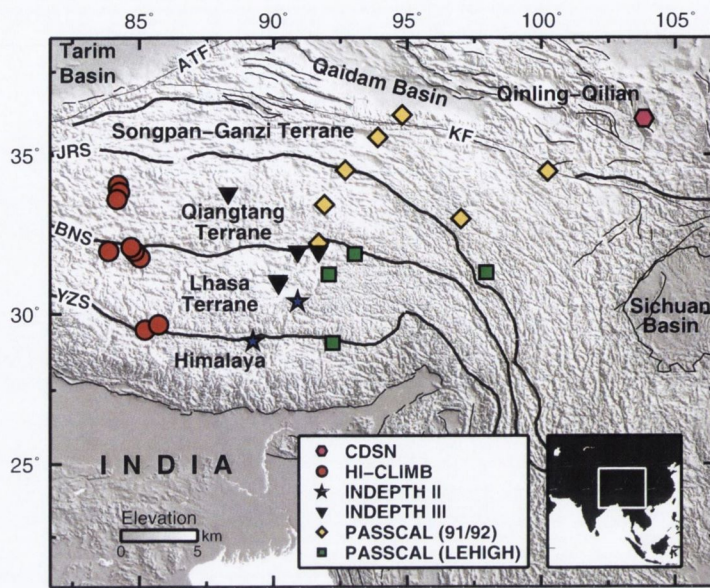


Figure 3.1: Map of Tibet and surrounding regions. Different symbols identify stations of various networks used in this study. Black lines indicate sutures and faults: YZS (Yarlung-Zangbo Suture), BNS (Bangong-Nujiang Suture), JRS (Jinsha River Suture), KF (Kunlun Fault) and ATF (Altyn Tagh Fault), which divide Tibet into major terranes. Inset shows the location of the map region in Asia.

ceous northward subduction of the Indo-Australian oceanic lithosphere beneath them (*England and Searle, 1986*). Following the closure of the Tethys ocean, the thrusting of Indian continental lithosphere beneath the Himalayas and the Lhasa Terrane is thought to have been taking place for the last ~ 50 My (determined using radiometric dating of minerals e.g., *Allègre et al., 1984; Yin and Harrison, 2000*), although it has also been proposed that the collision of thick, contiguous Indian continental lithosphere with Asia occurred only around 34 Ma (*Aitchison et al., 2007*, determined from faunal assemblages) or even 25–20 Ma (*van Hinsbergen et al., 2012*, with added constrain from climate change).

The oceanic Tethyan slabs have now sunken into the deep mantle (*Replumaz et al., 2004, 2010a; Van der Voo et al., 1999*). It is estimated that India and Asia converged by up to 3,600 km since 52 My (*van Hinsbergen et al., 2011*), with at least 1400 km of north-south shortening absorbed by the Tibetan orogen (*Yin and Harrison, 2000; Guillot et al., 2003*). Different mechanisms for the convergence have been proposed. The three main end-member models include underthrusting,

subduction and viscous thickening of the lithosphere. We now discuss six different scenarios which fall under one of the three types of models (Figure 3.2), while keeping in mind that elements of all three end-member models may be present beneath different parts of the plateau.

Underthrusting of India far north beneath the Tibetan Plateau has been proposed by *Argand* (1924). In this paper we define underthrusting as the scenario in which the Indian lithosphere slides directly beneath the Tibetan lithosphere, with no asthenospheric window between them (e.g., *Ni and Barazangi*, 1984; *Owens and Zandt*, 1997). The underthrusting of India beneath the entire length of the Himalayas, for example, is beyond doubt (e.g., *Ni and Barazangi*, 1984; *Nábelek et al.*, 2009). Further north, it may occur only in certain parts of the plateau. Beneath western Tibet, high seismic velocities have been detected in the shallow mantle using both body waves (*Barazangi and Ni*, 1982; *McNamara et al.*, 1997; *Huang and Zhao*, 2006; *Li et al.*, 2008) and surface waves (*Shapiro and Ritzwoller*, 2002; *Friederich*, 2003; *Priestley et al.*, 2006; *Lebedev and van der Hilst*, 2008; *Kustowski et al.*, 2008), suggesting underthrusting of the Indian lithosphere beneath western Tibet as far northward as the Tarim Basin (*Barazangi and Ni*, 1982; *Li et al.*, 2008). Consistent with this scenario, “flat-lying” *S*-to-*P* phase conversions from *S* receiver functions in western Tibet have been interpreted as the Indian and Asian lithosphere-asthenosphere-boundaries (LAB), meeting near the northern boundary of the plateau (*Zhao et al.*, 2010; *Kind and Yuan*, 2010).

The second end-member model is lithospheric subduction, which we define as the lithosphere sliding into the asthenosphere (the Tibetan lithosphere above and the Indian lithosphere below are separated by a layer of asthenosphere). The possible modes of subduction beneath Tibet include shallow-angle, steep-angle, northward and southward subduction (Figure 3.2). Steep-angle, northward subduction of India has been reported in parts of southern Tibet based on *P*-wave, travel-time tomography (*Tilmann et al.*, 2003; *Huang and Zhao*, 2006; *Li et al.*, 2008). The tomographic models showed high-velocity anomalies extending from north-Indian uppermost mantle down to 300 km depth beneath southern Tibet, north of the Yarlung-Zangbo Suture (YZS), although only in one particular segment within Tibet (87°E to 90°E). In northern Tibet, north of the Bangong-Nujiang Suture (BNS), these models show low wave speeds in the entire upper mantle.

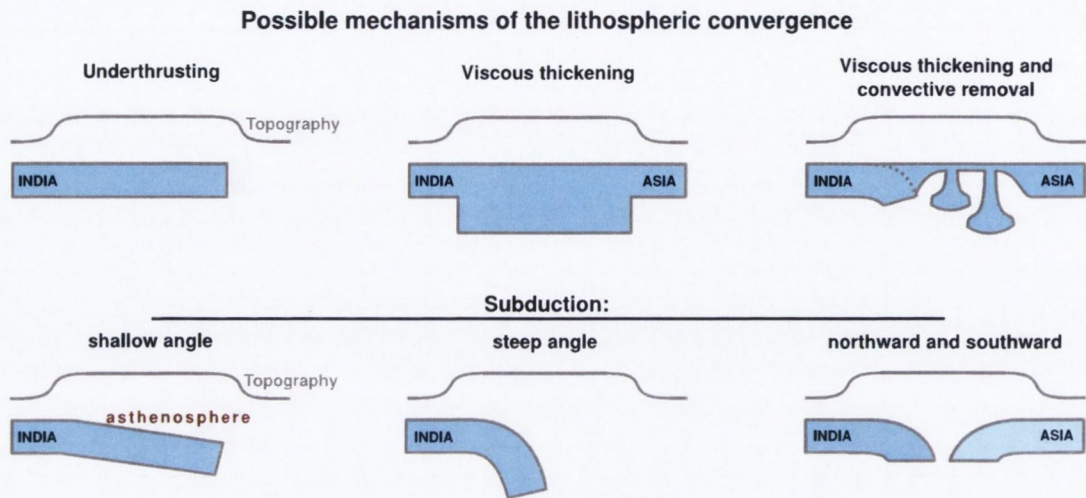


Figure 3.2: Conceptual models (not to scale) for the India-Asia lithospheric convergence. The three end-member models include underthrusting of India’s lithosphere beneath Tibet, viscous thickening of the lithosphere, and subduction.

Southward subduction beneath north-eastern Tibet has been proposed based on evidence from P receiver functions (Kosarev *et al.*, 1999; Kind *et al.*, 2002). Subduction reaching as far south as southern Qiangtang Terrane has recently been inferred from S receiver functions as well (Kumar *et al.*, 2006; Zhao *et al.*, 2010, 2011).

Surface-wave tomographic models, both global (Shapiro and Ritzwoller, 2002; Lebedev and van der Hilst, 2008, Figure 3.3) and regional (Debayle *et al.*, 2001; Friederich, 2003; Priestley *et al.*, 2006; Kustowski *et al.*, 2008; Panning *et al.*, 2012), show low shear speeds in the uppermost mantle beneath northern and eastern Tibet, underlain by high S velocities (V_S) at ~ 200 km depth. While the high-velocity mantle lithosphere beneath the Himalayas and southernmost Tibet is likely to be Indian (Friederich, 2003; Priestley *et al.*, 2006), the lithospheric configuration beneath the central, northern and eastern Tibet is debated.

The third end-member model of lithospheric convergence beneath Tibet is viscous lithospheric thickening. In one of the proposed scenarios, the mantle lithosphere thickens and remains intact, while in another scenario the thickened lithosphere becomes destabilised, with large parts of it removed convectively and sinking into deep mantle (Figure 3.2). McKenzie and Priestley (2008) compared the

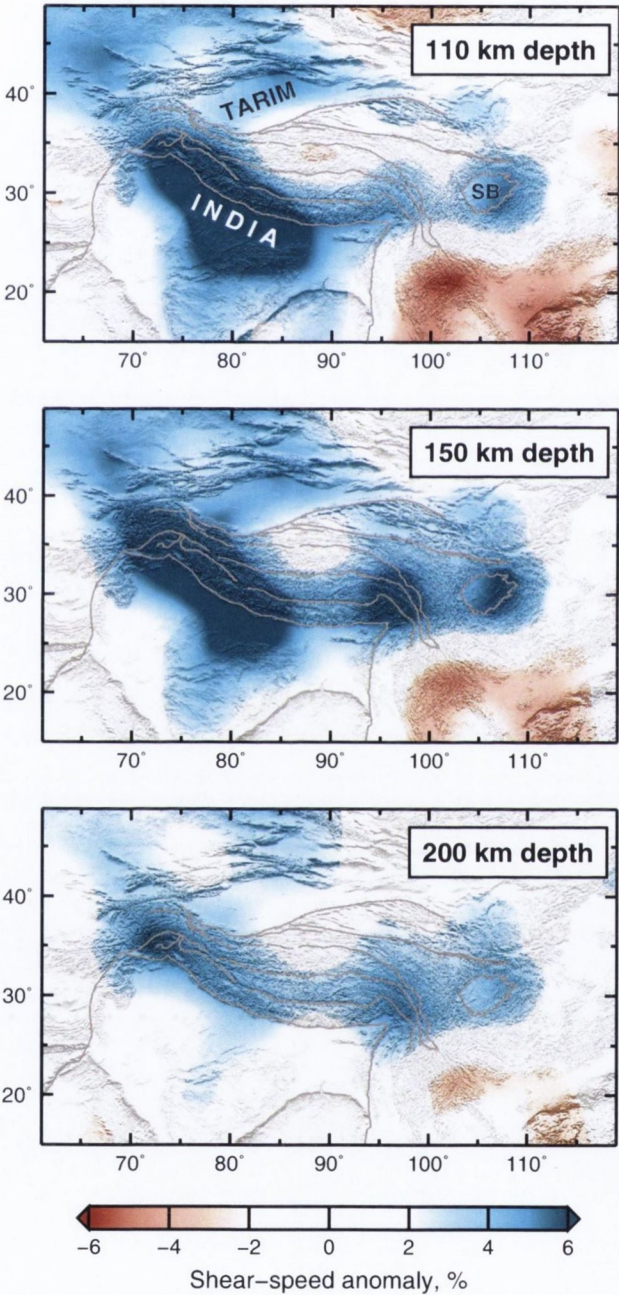


Figure 3.3: Surface-wave tomography across Tibet and surroundings from the global model of *Lebedev and van der Hilst* (2008). The three depth slices show fast and slow shear-speed anomalies beneath Tibet (blue and red, respectively). SB: Sichuan Basin.

lithospheric structure of Tibet to that of cratons and suggested that a craton was forming beneath Tibet today. They proposed that shortening of the lithospheric mantle would thicken the lithosphere and transport depleted material downwards, to form a cratonic root. They argued, further, that the crust was insulated from the convecting mantle and that crustal radioactive heating was responsible for the lower V_S in the uppermost mantle, just beneath the Mohorovičić discontinuity (Moho). The alternative scenario (e.g., *Houseman et al.*, 1981; *Molnar et al.*, 1993; *Hatzfeld and Molnar*, 2010) invokes convective removal of Tibet's thickened mantle lithosphere, with the cold lower lithosphere dripping into sinking plumes and the warm asthenosphere flowing in to replace it.

In spite of decades of intensive study, there is surprisingly little agreement, as of today, regarding even the most basic features of the lithospheric configuration beneath Tibet. What is the fate of the Indian lithosphere after it descends beneath the Himalayas? Is it still within the upper mantle beneath Tibet, and how does it influence the dynamics of the plateau? What is the thermal structure and thickness of the lithosphere of Tibet itself? What are the basic mechanisms of the lithospheric convergence? Surface-wave tomographic models show, in mutual agreement with each other, high-velocity anomalies in the upper mantle beneath much of Tibet. Most of the recent body-wave models, in contrast, do not display such anomalies (e.g., *Li et al.*, 2008).

In this study we aim to resolve the controversy by measuring accurate surface-wave phase velocities using stations within and around Tibet and deriving robust shear-velocity profiles for different parts of the plateau. The one-dimensional, radially anisotropic (1-D) V_S models are obtained in small, simple inverse problems, with features of the models showing clear relationships to anomalies in the data. The data require substantial high-velocity anomalies in the upper mantle beneath much of the plateau. Using petro-physical models, we then estimate the large temperature anomalies necessary to account for the observations and show that they are consistent with the presence of subducted Indian lithosphere in the upper mantle beneath most of the plateau. The character of subduction, however, shows significant changes across Tibet.

3.2 Data and methods

The fundamental advantage of inter-station measurements of surface-wave phase velocities, compared to the source-station measurements normally used in large-scale tomography, is that they can be performed in broader frequency bands (*Meier et al.*, 2004; *Lebedev et al.*, 2006). This allows us to resolve trade-offs between model parameters in the crust and in the upper mantle and to determine seismic structure within both with higher accuracy. Here we aimed to obtain tight constraints on V_S profiles beneath a selection of locations across Tibet, in order to derive inferences on the structure and dynamics of the upper mantle beneath it.

We chose 29 pairs of broadband seismic stations, including both permanent stations and the stations from recently deployed temporary networks (PASSCAL (91/92) (*Owens et al.*, 1993), INDEPTH II and III (*Nelson et al.*, 1996; *Huang et al.*, 2000), HI-CLIMB (*Nábělek et al.*, 2005), PASSCAL (Lehigh) (*Sol et al.*, 2007), Figure 3.1). Each pair was such that numerous suitable earthquake recordings were available for the measurements of inter-station surface-wave dispersion, resulting in robust average phase-velocity curves. The thousands of three-component seismograms used in the analysis were retrieved from the Incorporated Research Institutions for Seismology (IRIS) database.

3.2.1 Phase-velocity measurements

Robust inter-station phase-velocity measurements were performed in broad period ranges using a combination (*Lebedev et al.*, 2006) of cross-correlation (*Meier et al.*, 2004) and the Automated-Multimode-Inversion (AMI) of surface and S waveforms (*Lebedev et al.*, 2005). The cross correlation of the vertical components from two stations aligned along the same great circle path with an earthquake measures the phase delay of Rayleigh waves. The Rayleigh-wave fundamental-mode inter-station phase velocity is then calculated using the difference of the distances from the source to the two stations. Similarly, Love-wave phase velocity is measured using the transverse components. The main advantage of the cross-correlation technique is that it can measure accurate phase velocities for broad period ranges (5–300 s), even at short periods where seismograms are often complex due to the diffraction of the surface-wave fundamental mode.

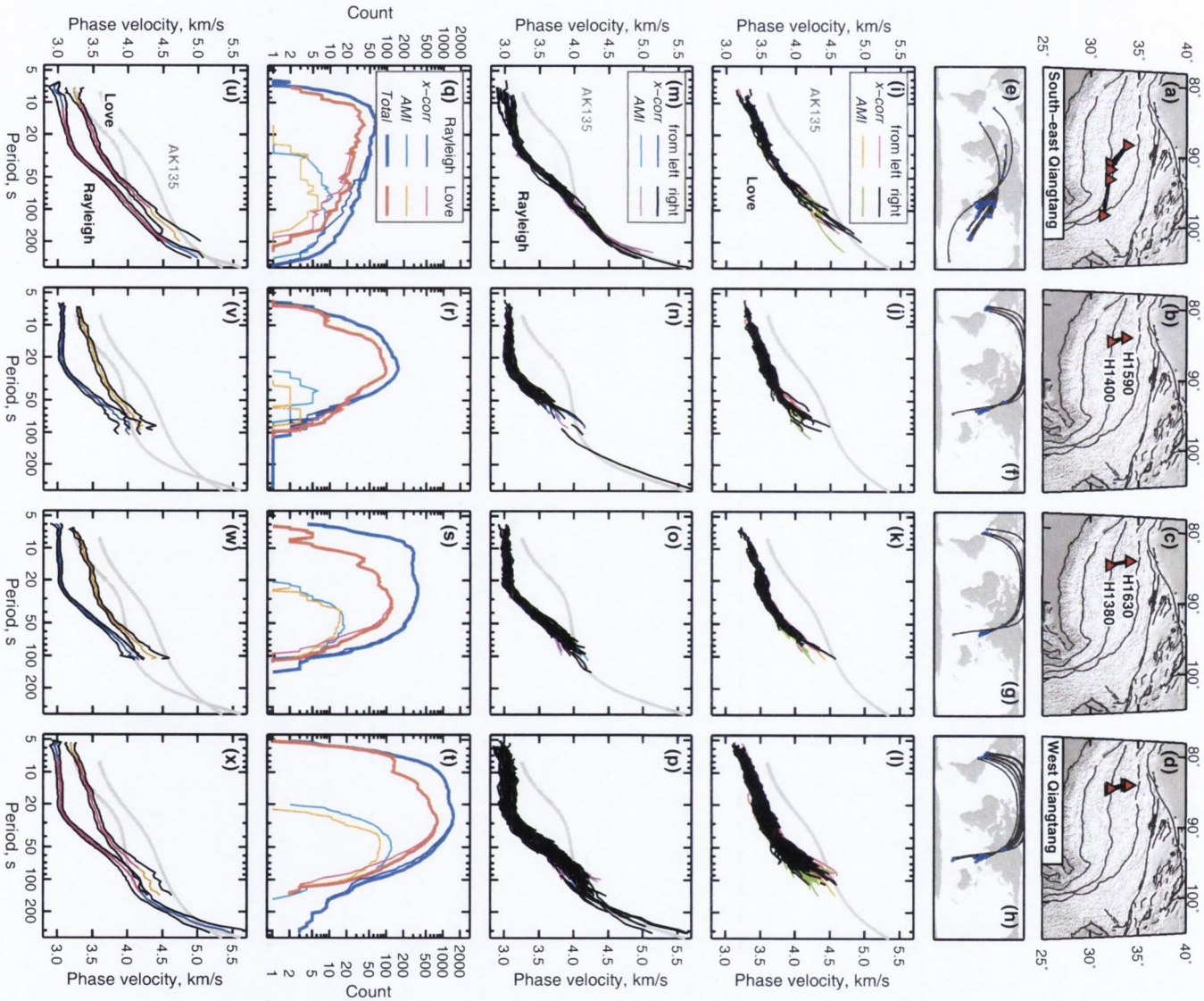
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Phase velocities can be difficult to extract using cross correlation at long periods, at which surface waves often interfere with energetic S and multiple S waves, especially for Love waves. These are better calculated using source-station phase-velocity measurements derived from full waveform fits computed by AMI. S , multiple S , and surface waves are simultaneously fit within a set of time-frequency windows, using synthetic seismograms generated through normal-mode summation. The fundamental-mode station-station dispersion curve can therefore be determined from pairs of the source-station measurements (same event, two station on the same great-circle path).

Only events with the difference between the back azimuths and station-station azimuths of less than 10° were included. This great circle path condition is used not because we assume that the surface waves propagate exactly along the station-station azimuth but because errors due to unmodeled surface-wave diffraction increase with an increasing difference between the event-station and station-station azimuths.

The earthquakes contributing to our measurements are from both directions along the inter-station paths, although predominantly from the east (from western Pacific and southeast Asia, Figure 3.4, e-h). The similarity of phase-velocity mea-

Figure 3.4 (facing page): Phase-velocity measurements for inter-station pairs and grouped regional averages. Top (a-d): the location of the stations (red triangles), and inter-station paths (black lines). The global maps (e-h) show earthquakes (blue dots) and earthquake-station great circle paths (black lines). Love- and Rayleigh-wave dispersion curves are displayed in separate plots (i-l, m-p), each showing superimposed one-event measurements from different methods: cross-correlation (x-corr) and Automated-Multimode-Inversion (AMI). Different colours indicate the measurement method and wave-propagation direction. Grey curves show the dispersion curves computed for the AK135 reference model (*Kennett et al.*, 1995). The total number of measurements at each period for Love and Rayleigh waves are shown in red and blue, respectively, in the count graph (q-t). Different colours show the measurement count from different methods. The bottom frame shows the resulting average dispersion curves for Love and Rayleigh waves and their standard deviations (black curves) (u-x); pink curves show the most robust measurements selected to be used for the inversions. The two columns in the middle illustrate inter-station measurements for individual station pairs. The columns on the left and on the right show examples of region-average phase-velocity measurements, using all the data from all inter-station pairs within the regions.



measurements calculated from earthquakes located on opposite sides of the inter-station paths is important, as systematic differences indicate timing errors or measurement biases arising from lateral heterogeneity between the source and the stations. For all the station pairs in our dataset, the measurements from the opposite directions are in mutual agreement (Figure 3.4, i-p).

The fundamental-mode dispersion curves measured by cross-correlation and AMI were combined together. The measurements from both methods are consistent and complement each other — cross-correlation measurements contribute to the short periods and AMI measurements provide more longer-period dispersion curves. Only smooth portions of each of the one-event curves were selected. Rough (not smooth) curves, rough portions of curves, and obvious outliers (measurements away from the general trend) were excluded. The selection of smooth curves is important because errors due to diffraction of the fundamental mode (*Pedersen, 2006*) and errors due to interference with higher modes are frequency-dependent and often result in roughness of measured dispersion curves.

The selection of smooth curves and the averaging over hundreds of measurements from earthquakes from different directions and distances enhance the accuracy of the phase-velocity measurements. The average phase velocity and the standard deviation are computed for each period. In the final selection step, only the smoothest parts of the average dispersion curves, with smallest standard deviations, were selected for further processing (Figure 3.4, u-x).

3.2.2 Phase-velocity curves

The shape of the dispersion curves reflects the distribution of elastic parameters within the Earth. Short-period waves travel closer to the surface, whereas long-period waves travel at greater depths and sample broader depth ranges. In the 7–50 second period range, both our Love- and Rayleigh-wave phase-velocity curves are well below the curves computed for global-average AK135 reference model (*Kennett et al., 1995*), due to the very thick crust beneath Tibet (*Romanowicz, 1982*). Our focus here is on longer-period surface waves, more sensitive to the deeper, mantle structure. We observe systematic differences between dispersion curves from different parts of Tibet at periods longer than 40 seconds, at which Rayleigh waves sample the upper mantle.

Comparing our measurements, we find that neighbouring station pairs show similar inter-station dispersion curves. Station pairs H1400-H1590 and H1380-H1630 (Figure 3.4, b and c), for example, are close to each other within western Qiangtang Terrane; as expected, they have very similar dispersion curves (Figure 3.4, j-k and n-o). In addition to these two pairs, 6 other station pairs along the northern HI-CLIMB network also show very similar dispersion curves. These measurements can thus be grouped together, to characterise the structure of this region within Tibet, West Qiangtang (Figure 3.4d). With hundreds of “one-event” Love- and Rayleigh-wave dispersion curves measured within this region (Figure 3.4, l, p and t), the average phase-velocity curves for it are robust in a broad period range, for both Rayleigh and Love waves (Figure 3.4x). Similarly, three inter-station pairs across South-east Qiangtang are also grouped together (Figure 3.4a). Although the stations are relatively far apart, they yield very similar Rayleigh- and Love-wave dispersion curves (Figure 3.4, i and m). Unlike in West Qiangtang, the inter-station paths in South-east Qiangtang have different azimuths, so that the earthquakes contributing signal for the measurements are distributed in more diverse source regions around the globe (Figure 3.4e).

3.2.3 Regions within Tibet

Guided by the degree of similarity of phase-velocity curves, we identify seven distinct regions within Tibet that our measurements sample: West Lhasa, Central Lhasa, West Qiangtang, South-east Qiangtang, East-central Qiangtang, Songpan-Ganzi and Qinling-Qilian. Figure 3.5a shows the grouped inter-station pairs, indicating the location of each region. The dispersion curves from West Lhasa and Qinling-Qilian have the most distinctive curves when compared to the other regions across Tibet (Figure 3.5b). Between 10 and 50 seconds, phase velocities from the two paths sampling Qinling-Qilian orogen (region 7) are higher than from other paths (Figure 3.5, b and h); this is due to the thinner crust in Qinling-Qilian, just outside of the high plateau. Within the high plateau (regions 1–6), Rayleigh and Love waves have very similar phase velocities in all regions between 10 and 40 s, with variations of less than 0.2 km s^{-1} (Figure 3.5b). Beyond 40 s, pronounced contrasts in the rates of phase-velocity increases with period in different regions indicate strong heterogeneity in the upper mantle.

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Within each region, phase-velocity differences between individual inter-station pairs are mostly at short periods. The thin, pale-coloured curves in Figure 3.5, c–h, show the within-region velocity variations, most visible at 7–25 seconds. At very short periods these variations may be an effect of sediments; at intermediate periods (10–25 seconds) the anomalies are likely due to deeper crustal heterogeneity. Beyond 30 seconds, individual Rayleigh-wave measurements are remarkably consistent with each other within each region. Individual Love-wave inter-station dispersion measurements tend to be noisier and plot within somewhat broader bands. Bold, dark coloured phase-velocity curves in Figure 3.5 are the averages over all the station pairs within a region. These averages yield robust phase-velocity curves in period ranges broader than those of the curves for individual inter-station pairs (pale curves). This is because the region averages are computed from the larger number of all the “one-event” measurements from the region.

For the purposes of this study, we focus primarily on measurements at periods longer than 40 s. Depth sensitivity kernels (Fréchet derivatives of surface-wave phase velocity with respect to the shear speed at different depths, Figure 3.6) show how surface waves at longer periods are sensitive to greater depths and broader depth ranges. At 50 s, Rayleigh-wave phase velocity has a peak sensitivity at 65 km depth, just above the Moho, and at 130 s the maximum sensitivity is at about 190 km. Love-wave peak sensitivity is shallower than the sensitivity of Rayleigh waves for the same periods.

A comparison of region-average curves at intermediate and long periods indicates strong lateral heterogeneity in the upper mantle beneath Tibet. The neighbouring West Lhasa and West Qiangtang regions show the largest contrast in the Rayleigh-wave dispersion curves, with a sharp phase-velocity increase at periods above 30 s in West Lhasa indicating uppermost-mantle seismic velocities in this region that are much higher than those beneath West Qiangtang (Figure 3.5c). Between 40 and 100 seconds, Rayleigh-wave phase velocities in Central Lhasa are lower than in West Lhasa (Figure 3.5d) but still slightly faster than in South-east Qiangtang (Figure 3.5e). The Rayleigh-wave phase velocities in East-central Qiangtang are slightly higher than in South-east Qiangtang between 50 and 110 seconds, but both exceed the global reference values between 110 and 180 s. The measurements for the Songpan-Ganzi region show that beyond 70 seconds the

Rayleigh-wave phase-velocities are lower than in the neighbouring eastern Qiangtang (Figure 3.5, b and g). At 40–60 s, Rayleigh-wave phase velocities in West Qiangtang are lower than in eastern Qiangtang and the dispersion curve then keeps to the global reference for periods beyond 100 seconds (Figure 3.5b). Dispersion measurements from paths sampling the Qinling-Qilian region, north of the Kunlun Fault and with a crust thinner than beneath the other regions, show substantially higher phase velocities of both Love and Rayleigh waves, at periods up to 100 s (Figure 3.5h).

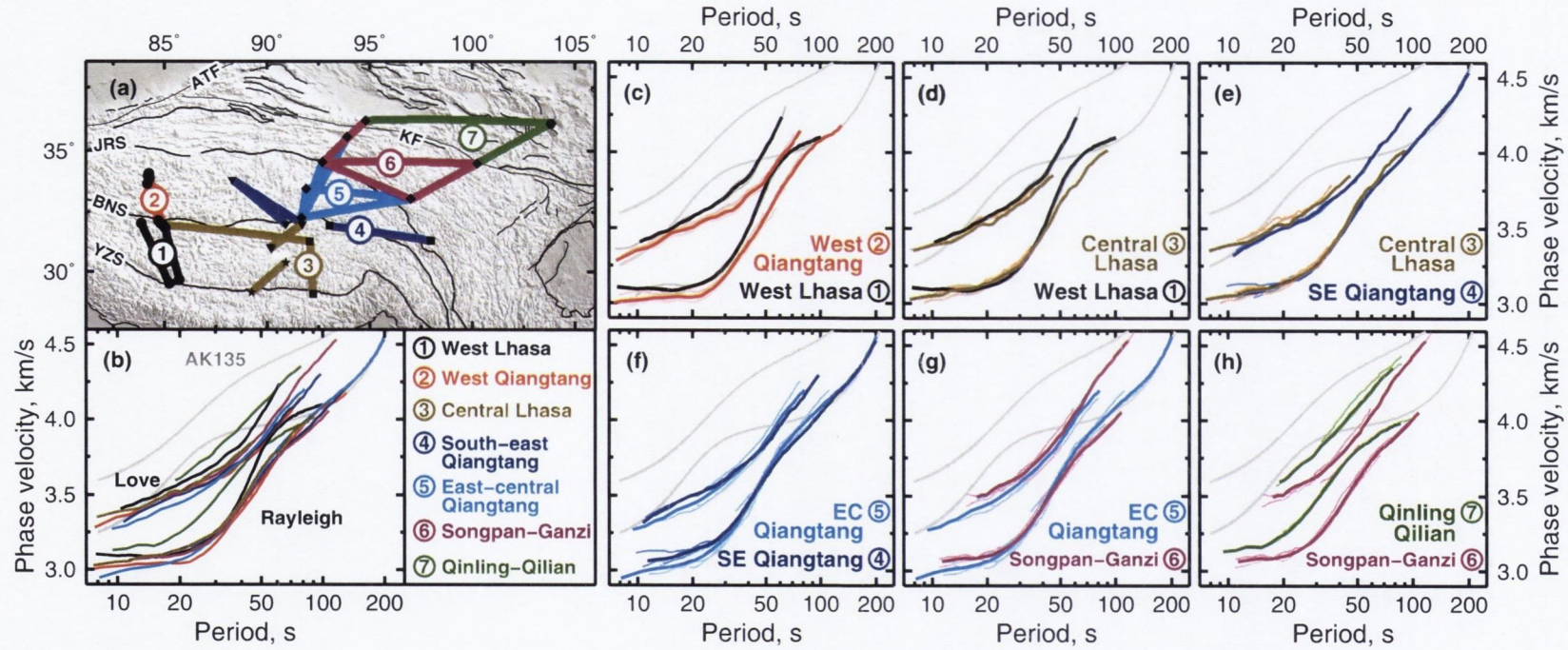


Figure 3.5: Broad-band phase-velocity curves measured across Tibet. (a) Inter-station paths within 7 colour-coded regions. Different symbols indicate stations of different networks as in Figure 3.1. (b) The 7 Love-wave and 7 Rayleigh-wave, region-average phase-velocity curves. Grey lines: Dispersion curves computed for the AK135 reference model (*Kennett et al., 1995*). (c-h) Comparisons of region-average dispersion curves. The colour code is as in (a). Pale, thin curves are the phase-velocity measurements from individual inter-station paths within the region, bold curves are region averages.

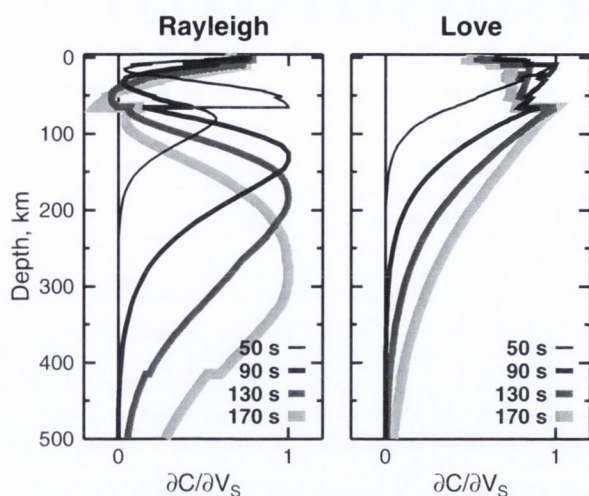


Figure 3.6: Depth sensitivity kernels for the fundamental mode Rayleigh- and Love-wave phase velocities. Each curve is scaled independently. These Fréchet derivatives of phase velocities are computed for a 1-D V_S profile from South-east Qiangtang, with a four layer crust and a 65-km deep Moho.

3.2.4 Inversion of phase-velocity curves for shear-velocity profiles

We invert Rayleigh- and Love-wave dispersion curves simultaneously for a 1-D profile of the isotropic average shear speed $V_S = V_{S(\text{avg})} = (V_{SV} + V_{SH})/2$ and radial anisotropy $(V_{SV} - V_{SH})/2$, where V_{SV} and V_{SH} are the vertically and horizontally polarised shear speeds, respectively. The inversion is performed using a fully non-linear, Levenberg-Marquardt gradient search. The search direction is computed at each iteration with the synthetic phase velocities computed directly from the V_{SH} and V_{SV} models, using a suitably fast version of the MINEOS modes code (Masters, <http://geodynamics.org/cig/software/mineos>). Compressional velocity V_P has a small but non-negligible influence on the Rayleigh-wave phase velocity. V_P is assumed isotropic and the ratio between the isotropic-average shear speed $V_{S(\text{iso})}$ and V_P is kept fixed during the inversion. In Figures 3.7, 3.8, and 3.9 we plot the Voigt isotropic average $V_{S(\text{iso})} = (2V_{SV} + V_{SH})/3$ instead of the arithmetic average $V_{S(\text{avg})}$.

Perturbations in the model are controlled by basis functions, boxcar-shaped in the crust and triangular in the mantle (*Bartusch et al., 2011*). These define the sensitivity depth range of two independent inversion parameters: one for the isotropic average V_S and one for the amount of radial anisotropy. The depth of the Moho and three intra-crustal discontinuities are additional inversion parameters. Slight norm damping is applied in order to avoid physically unreasonable models.

The inversions are fundamentally non-unique. Rather than selecting one pre-

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ferred model, we obtain a suite of models that fit the data. The bundles of V_S profiles are generated in grid searches, complemented by targeted test inversions designed to answer specific questions, as described below.

Although our primary aim is to investigate the structures of the upper mantle, the influence of the crust is significant and is taken into account. All the V_S profiles start from the surface, at 0 km depth. Topography is taken into account. The parameterisation for the crustal layers allows for radial anisotropy and isotropic-average shear-speed perturbations. The synthetic dispersion curves from all the S -velocity models fit the data within the standard deviation at all periods. A detailed investigation of the crust is beyond the scope of this study and will be presented in a separate study (Chapter 4).

We plot the models for each region colour-coded according to the data-synthetic fit they provide. In Figure 3.7, both the non-uniqueness of the models and their robust features are readily apparent. The colour of each profile and synthetic dispersion curve is determined from the phase-velocity, data-synthetic misfit computed over the frequency band most sensitive to the upper mantle ($T > 40$ s). The coloured depth ranges of the V_S profiles indicate the depth sensitivity of the data. The data-synthetic misfits for the Love and Rayleigh waves are shown in Figure 3.7c and d, respectively. The remaining misfits for the best fitting models (darkest lines) are due to noise in the data (given the broad depth ranges of phase-velocity sensitivity (Figure 3.6), sharp kinks in phase velocity curves cannot be explained by any realistic structure). Note that for most of the better-fitting models, the misfits are within the range of only $\pm 0.5\%$.

Overviewing the models, we note that the profiles for West Lhasa show mantle lithosphere that is 100 km thick and characterised by very high shear speeds, down to 175 km depth (Figure 3.7, region 1). In West Qiangtang (region 2), in contrast, the uppermost mantle layer (from the Moho down to 100–125 km depth) shows low average V_S . This layer, however, is underlain by a ~ 100 km thick, high velocity anomaly with V_S exceeding 4.7 km s^{-1} . A similar high-velocity anomaly is present beneath central and southeastern Qiangtang (regions 4 and 5), where it is deeper and broader (150–300 km depth). Although less prominent, the high-velocity anomaly is also observed beneath Songpan-Ganzi (region 6), below 150 km depth, but not beneath Qinling-Qilian (region 7).

Uppermost-mantle S velocities in central and northeastern Tibet are relatively low, but not as low as those in West Qiangtang. In contrast to West Lhasa, Central Lhasa displays no craton-like, high V_S in the uppermost mantle, with the best fitting profiles there close to the global average values.

3.2.5 Targeted test inversions and parameter-range estimation

Shear-velocity models yielded by inversions of phase-velocity curves are non-unique. For example, the LAB is not always well defined by the V_S models as the data can be fit equally well with or without a sharp discontinuity (e.g., *Bartusch et al.*, 2011). Similarly, a deeper Moho trades-off with faster lower crust and upper mantle, and *vice versa* for a shallower Moho.

In order to determine if a specific lithospheric configurations fits the surface-wave data, information on the Moho depth and S_n velocity were included in the background model and were not allowed to change during the inversion (e.g., Figure 3.9). These joint inversions are important to find a holistic solution to the different data sets and lead to a better constraint on the surface-wave data.

In another approach, a series of targeted test inversions to delimit ranges of shear speed and radial anisotropy were made by strong damping of the V_S or anisotropic parameters within a tested depth range. Thus, the inversion is encouraged to make use of the other less damped parameters to find a model that fits the data in the set conditions. These test inversions are performed in a systematic way, analogous to a “grid search” approach, with the test parameter set at a new step for the next inversion. Tests for V_S ranges were performed in steps of 0.025 km s^{-1} . The better or worse fit from each test is visible in the data-synthetic misfit as the affected periods reflect the depth range sensitivity (e.g., Figures 3.7 and 3.8).

For each inverted model we calculate the data-synthetic root mean square (RMS) of the combined Love and Rayleigh waves for periods longer than 40 s. The advantage of using RMS is that it accumulates misfit over a number of samples rather than being biased by one noisy sample. As a criteria to discriminate between acceptable and rejected models we allow for RMS values of up to 3.5 m s^{-1} above the best fitting regional model.

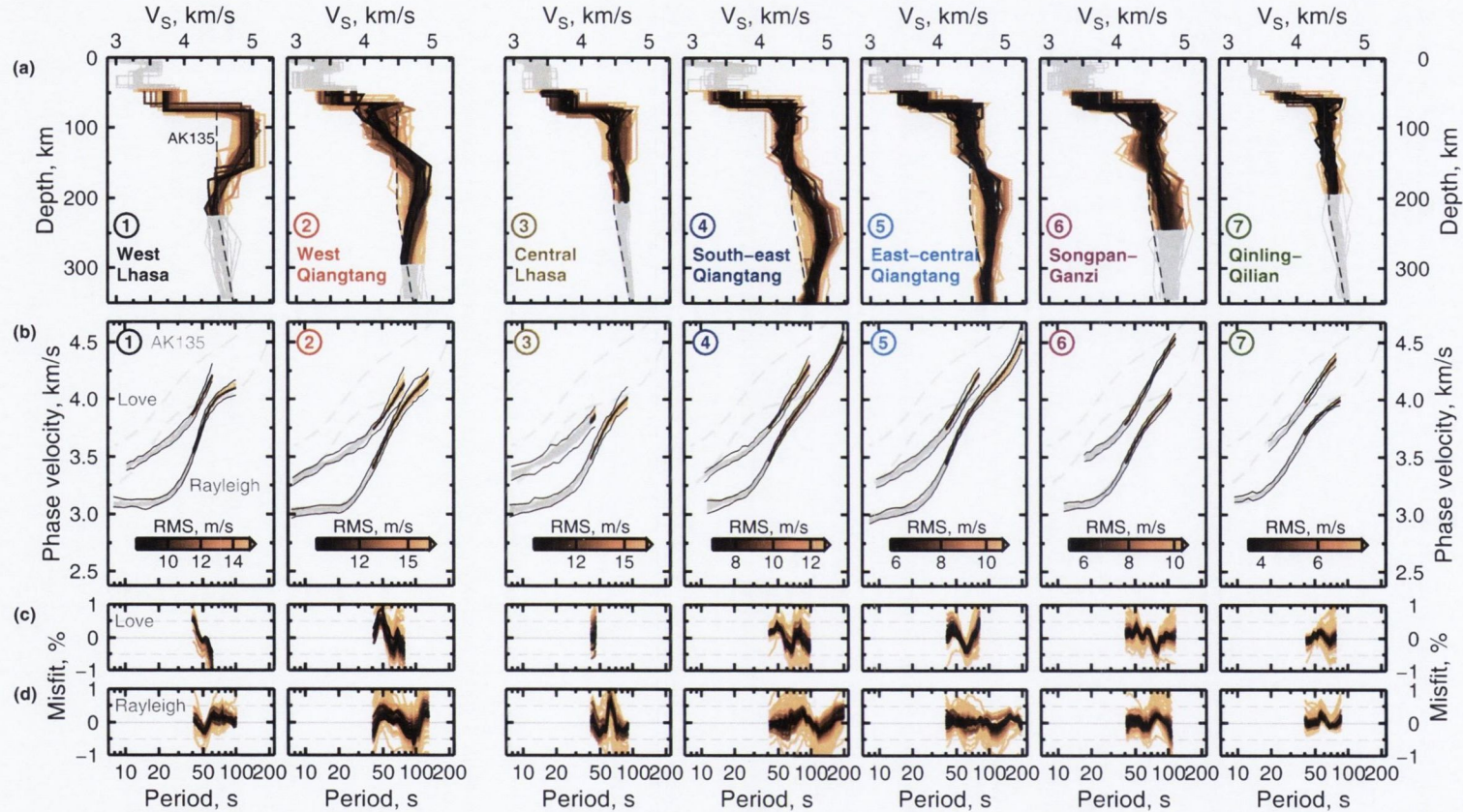


Figure 3.7: Shear-speed profiles for the seven regions within Tibet. (a) Each 1-D V_S profile is obtained in an independent, non-linear, gradient-search inversion with a generally different parameterisation. (b) Synthetic phase velocities corresponding to the 1-D V_S profiles. (c, d) The data-synthetic misfit for Love- and Rayleigh-wave phase velocity, respectively. The colour scale indicates the data-synthetic RMS misfit for the combined Rayleigh- and Love-wave dispersion curves for periods >40 s. The coloured depth range of the V_S profiles shows the depth sensitivity for periods from 40 s up to the long-period cut-off for these data. All synthetic dispersion curves fit within the measured standard deviation (solid black curves). Dashed lines are the AK135 reference model (a) and synthetic phase-velocity curves computed for it (b) (Kennett *et al.*, 1995).

3.3 Upper mantle structure

The upper mantle beneath the Tibetan Plateau is strongly heterogeneous with the most striking feature being across western profiles (Figure 3.5c and Figure 3.7, regions 1 and 2). Beneath West Lhasa all the 1-D V_S profiles require high upper-mantle S velocities close to 5 km s^{-1} whereas in West Qiangtang the uppermost 75 km mantle appears complex having broad low-velocity variability between 4.0 – 4.5 km s^{-1} . This contrast across the BNS is in agreement with the tomographic model in Figure 3.3 (110 km depth) which shows a dramatic shear-velocity decrease at $85^\circ\text{E } 31^\circ\text{N}$. At 150 and 200 km depths, high S velocities are now beneath central and eastern Tibet also in agreement with profiles 4, 5 and 6 (Figure 3.7).

In general body- and surface-wave tomographic models agree on the shallow upper-mantle architecture of West Lhasa with both techniques showing fast seismic velocities. This is not the case for the northern and eastern Tibetan regions such as West Qiangtang as body-wave tomographic models show low velocities in the upper mantle (*Huang and Zhao, 2006; Li et al., 2008*) in contrast to high velocities observed by surface-wave models (*Debayle et al., 2001; Friederich, 2003; Priestley et al., 2006; Kustowski et al., 2008; Panning et al., 2012*). In practice, surface-wave models have good resolution to vertical heterogeneity within lithospheric depths whereas body-wave models, using near-vertical rays and sampling through a very thick crust, find it hard to resolve vertical anomalies. On the other hand thin, shallow, uppermost-mantle discontinuities modelled by receiver functions are difficult to resolve by surface waves. Thus models of low S velocity just beneath the Moho increasing with depth below (*Friederich, 2003; Priestley et al., 2006*), challenge models that have a thin lid on top of a warm asthenosphere (*Houseman et al., 1981; Molnar et al., 1993; Hatzfeld and Molnar, 2010; Zhao et al., 2010; Zhao et al., 2011*). In this study we investigate the different configurations and consider only the models that fit the surface-wave data.

3.3.1 Underthrusting of India beneath West Lhasa

To determine the average shear velocity and hence the strength of the uppermost mantle beneath the Tibetan Plateau we run a series of specific test inversions with set V_S values in the upper ~ 70 km of the mantle. Parameterisation above and

below the test depths allows the inversion to perturb the model to fit the data. Figure 3.8a shows test models for East-central Qiangtang, their corresponding synthetic dispersion, and data-synthetic misfits. The S velocity at the lower crust and below the test depths appears to trade off with the upper and lower V_S ranges. The change in the data-synthetic dispersion fit can be clearly seen between 50–80 seconds for both Love and Rayleigh waves in the misfit frame.

In West Lhasa, between the Moho and 155 km depth, the average shear velocities of the better fitting models are between $4.87\text{--}5.01\text{ km s}^{-1}$ (Figure 3.7, Table 3.1, region 1). These velocities are approximately 5% faster than the average V_S of cratons (*Lebedev et al.* (2009), also illustrated in Figure 3.11a) and highlights the strong rigid nature of north western India. Across Central Lhasa, as expected from the surface-wave tomography (Figure 3.3), the S velocities are close to continental average (Figure 3.7, Table 3.1, region 3). It is unclear whether the V_S is the true velocity throughout the entire area beneath the inter-station paths or whether it is a result of averaging across a fragmented upper mantle as suggested by *Liang et al.* (2012).

The upper mantle high S velocities in West Lhasa contrasts with the very low V_S of nearby West Qiangtang in the north ($4.18\text{--}4.33\text{ km s}^{-1}$, Table 3.1, region 2). This dramatic change in shear speed is also mapped in the surface-wave tomography (Figure 3.3, 110 km depth). Our S velocities for the northern Tibetan regions, Qiangtang and Songpan-Ganzi, are also very low but increases in average V_S towards the north east (Table 3.1, Figure 3.12a). The sharp contrasts of fast and slow V_S indicate that the regions have different lithospheric mantles: one we identify as Indian lithospheric mantle (West Lhasa) and the other as Tibetan lithospheric mantle (north of the BNS) (Figure 3.12a). The immediate, large V_S increase at the Moho beneath West Lhasa fits our definition of underthrusting, hence Indian lithospheric mantle underthrusts the Tibetan Plateau only as far as the BNS along 85°E (Figure 3.13, central profile).

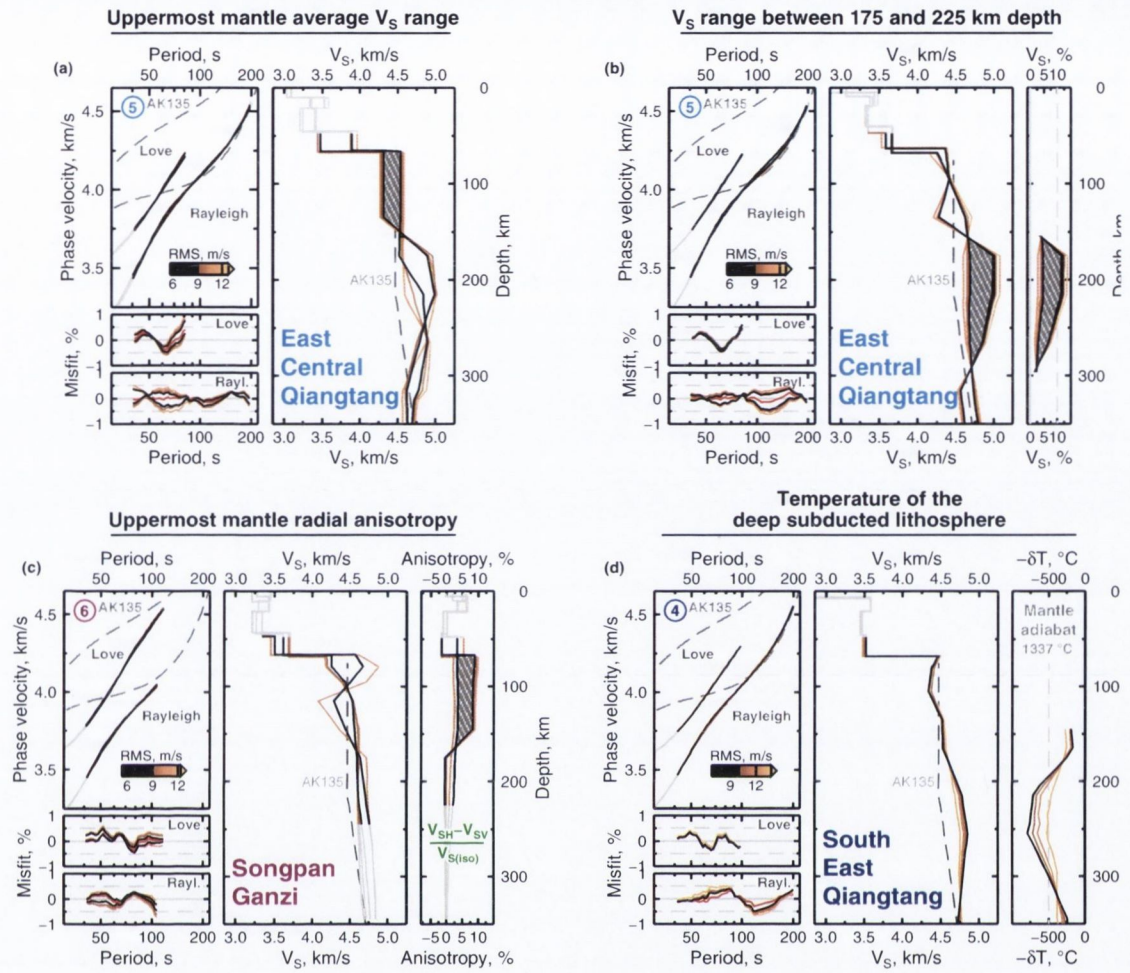


Figure 3.8: Four examples of the grid-search test inversions, performed to delimit V_S ranges, amounts of radial anisotropy, and temperature anomalies. The colour scale indicates the data-synthetic RMS misfit of the combined Rayleigh- and Love-wave dispersion curves for periods >40 s. Black V_S profiles and the corresponding synthetic phase velocities are the upper and lower bound determined for the average V_S or radial anisotropy within the depth range tested. The RMS misfit given by models within the ranges are up to 3.5 m s^{-1} larger than the global minimum misfit. Thin red curves in the misfit frames are the data-synthetic misfits given by the best-fitting profiles. (a) The test for the uppermost-mantle-average V_S range beneath East-central Qiangtang. (b) The test for the V_S range between 175 and 225 km depths beneath East-central Qiangtang; anomaly in percent relative to AK135 references is also shown. (c) The test for uppermost mantle radial anisotropy in the Songpan-Ganzi region. The isotropic average $V_{S(\text{iso})} = (2V_{SV} + V_{SH})/3$. (d) The temperature anomaly beneath South-east Qiangtang. Temperature anomalies are relative to the mantle adiabat with T_P 1337°C (see Figure 3.10). Dashed curves: synthetic dispersion curves computed for AK135 (Kennett *et al.*, 1995).

Region	Region No. (Fig. 3.5)	Uppermost Mantle		Subducted Lithosphere			
		Depth range (km)	Average V_S range (km s ⁻¹)	Depth range (km)	V_S range (km s ⁻¹)	V_S anomaly (%)	Temp. anomaly from 1337 °C mantle adiabat
West Qiangtang	2	Moho – 115	4.18 – 4.33	150 – 200	4.70 – 4.84	5.1 – 8.3	-580
South-east Qiangtang	4	Moho – 135	4.26 – 4.45	215 – 265	4.82 – 5.10	6.6 – 12.9	-740
East-central Qiangtang	5	Moho – 135	4.32 – 4.54	175 – 225	4.68 – 5.00	4.7 – 11.8	-490
Songpan-Ganzi	6	Moho – 135	4.32 – 4.53	185 – 235	4.63 – 4.88	3.2 – 8.9	-300
West Lhasa	1	Moho – 155	4.87 – 5.01				
Central Lhasa	3	Moho – 145	4.41 – 4.65				

Table 3.1: Average shear velocity and temperature anomalies for the lithospheric mantle beneath Tibet.

3.3.2 Warm, low S_n -velocity Tibetan lithospheric and asthenospheric mantle

The investigation for the finer structure of the Tibetan lithospheric mantle requires that we perform a joint analysis of surface-wave data with data of other types such as S_n velocities (uppermost mantle velocities just below the Moho) and Moho depths. Regional travel-time tomography for S_n propagation within northern Tibet show slow-to-average S_n velocities with minimum V_S ranging from 4.33 km s^{-1} (Ritzwoller *et al.*, 2002), 4.45 km s^{-1} (Sun *et al.*, 2008a,b) and 4.47 km s^{-1} (Pei *et al.*, 2007). Low frequency S_n waves (0.2–1.0 Hz) have been observed to propagate efficiently across the northern part of the plateau with a velocity of $\sim 4.7 \text{ km s}^{-1}$ (Barron and Priestley, 2009). At higher frequencies (>1 Hz), S_n coda propagates inefficiently through vast areas of northern Tibet and neighbouring regions (Barazangi and Ni, 1982; McNamara *et al.*, 1995; Barron and Priestley, 2009).

P_n velocities across north of BNS are close to continental average velocities with V_P at a typical 8.0 km s^{-1} (McNamara *et al.*, 1997; Hearn *et al.*, 2004; Liang *et al.*, 2004; Meissner *et al.*, 2004; Phillips *et al.*, 2005; Liang and Song, 2006; Sun and Toksöz, 2006; Pei *et al.*, 2007). Hearn *et al.* (2004) differentiates P_n velocities propagating between an upper and a lower lid and concludes that uppermost mantle P -wave velocity increases slightly with depth.

In Figure 3.9 we select three different 1-D V_S models that represent possible uppermost mantle structures for the northern Tibetan regions. These velocity models match average Moho depths from receiver functions (Kind *et al.*, 2002; Kumar *et al.*, 2006; Tseng *et al.*, 2009; Nábelek *et al.*, 2009; Yue *et al.*, 2012) and deep seismic sounding (Zhao *et al.*, 2001; Zhang and Klemperer, 2005; Liu *et al.*, 2006; Li *et al.*, 2006), and, accommodate for average S_n velocity (Ritzwoller *et al.*, 2002; Sun and Toksöz, 2006; Sun *et al.*, 2008a,b). We show that surface-wave data can fit models with S_n values just below the Moho as well as with models that have the shear velocity matching S_n at 20 and 35 km deeper; this to take into consideration the frequency-depth dependence of propagating S_n waves. Hence, the latter models comply with observations of inefficient high-frequency propagation of S_n in northern Tibet and with observations of faster P_n at sub-Moho depths. The data can be fit equally well with models that have a sub-Moho monotonic

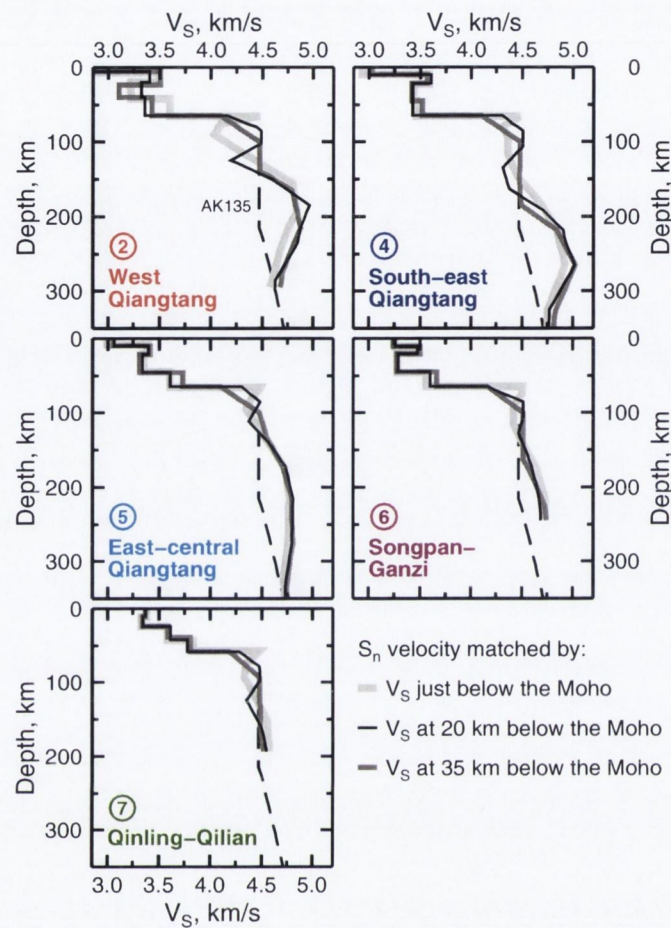


Figure 3.9: Example V_S profiles for northern Tibet that fit the surface-wave data and the published Moho depths and S_n velocities. S_n velocity is matched at Moho depth (light grey), at 20 km below the Moho (black), and at 35 km below the Moho (dark grey). The depth extent of the profiles varies according to the sensitivity (frequency band) of phase-velocity curves for the regions. Dashed line is the AK135 reference model (Kennett *et al.*, 1995).

S -velocity and with models with an uppermost mantle low-velocity zone; such a LVZ coincides with the negative phase discontinuities observed by S -to- P receiver functions across western and central Tibet (Zhao *et al.*, 2010, 2011).

The bottom of the low S -velocity varies from 125 km depth in West Qiangtang to 160 km depth in east Qiangtang (Figure 3.7). These depths are similar to a lithospheric interface modelled by S -to- P conversions by Wittlinger *et al.* (2004) and Yue *et al.* (2012).

Irrespective of the fine or average structure in the upper 75 km of the mantle, all our models for northern Tibet require low S -wave velocity (Figure 3.7 and 3.9, Table 3.1, regions 2, 4, 5 and 6). The sensitivity of S velocities to temperature suggests that these regions have high temperatures, hence a warm Tibetan lithospheric and asthenospheric mantle (Figure 3.12a).

3.3.3 Shallow-angle subduction of India

Fast upper mantle shear-velocity anomalies beneath the Tibetan Plateau have been subject to various discussions particularly on their origin and role in the continental collision between India and Asia. Our south-north regional V_S profiles (Figure 3.7, regions 1–2 and 3–6) show a 75–100 km thick high S -velocity “deepening” from West Lhasa to West Qiangtang and deeper down to 300 km in east Qiangtang — deeper than the fast anomalies mapped by surface-wave tomographic models (*Shapiro and Ritzwoller, 2002; Friederich, 2003; Priestley et al., 2006; Panning et al., 2012*). The deepening fast anomaly in the south-north profiles automatically rule out models of complete Indian underthrust beneath the plateau.

The high shear speeds at ~ 200 km depth mapped by regional surface-wave tomographic horizontal cross sections (*Debaille et al., 2001; Friederich, 2003; Priestley et al., 2006; Kustowski et al., 2008; Panning et al., 2012*), indicate that the anomaly is a large-scale feature beneath vast areas of the plateau. Hypothesised models describing the upper mantle’s low and high V_S anomalies from convective removal of mantle lithosphere or from craton formation of a thickened lithosphere are speculative and with little or no evidence. In the case for convective removal (*Molnar et al., 1993; Hatzfeld and Molnar, 2010*), irrespective of whether the dripping process was from an individual or several downwelling plumes, or whether the dripping process have taken place at different times, the large volume downwelling over a few million years, would have led to catastrophic lithospheric processes. At the surface, no colossal volcanism is observed (*Chung et al., 2005*). Beneath the high S -velocity anomaly our profiles show a decrease in V_S values down to typical global upper mantle averages, hence we see no obvious reason for retarding or halting of sinking lithospheric mantle. In the case of craton formation from a thickened lithosphere, the deep, thick lithospheric mantle acts as a thermal lid from the hot asthenosphere below, and therefore, the only source for internal lithospheric heat

is crustal radioactivity (*McKenzie and Priestley, 2008*). The decrease of V_S due to the conduction of heat as illustrated by *McKenzie and Priestley (2008)* does not resemble our better fitting profiles for northern Tibet; our profiles have very low velocities up to $\sim 4.5 \text{ km s}^{-1}$ from the Moho down to $\sim 150 \text{ km}$ depth whereas *McKenzie and Priestley (2008)* numerical models have a minimum of 4.5 km s^{-1} starting from the Moho increasing with depth. It is unlikely that heat solely produced by a radiogenic crust is sufficient to warm a 75 km thick layer below the Tibetan Moho.

Vertical cross sections from surface-wave tomography show high shear-speeds crossing from south to north Tibet at a shallow angle and with the highest velocity contours localised beneath north India (*Shapiro and Ritzwoller, 2002; Friederich, 2003; Priestley et al., 2006; Panning et al., 2012*). High S -wave velocities are indicative of low temperatures, typical for subduction, and hence to resolve for the origin of the high S -velocity material, we infer temperature. Here we determine quantitative temperature estimates for the deep high V_S and compare it to the expected temperature of subducted Indian lithosphere.

Our S -velocity estimates for the deep lithosphere show that across Qiangtang V_S anomalies between $150\text{--}265 \text{ km}$ depth are $4.7\text{--}12.9\%$ faster than global values (Table 3.1). Figure 3.8b illustrates the V_S tests for East-central Qiangtang region, each inverted independently with a specific fixed V_S between $175\text{--}225 \text{ km}$ depth. The misfit frame clearly shows the response of each of the V_S models with the synthetic dispersion fit worsening between 100 and 200 s. Note that the Love-wave synthetic phase velocity is unaffected by the deep V_S variations.

We select the least anomalous 1-D V_S model from the best fitting regional bundle and compare it to a series of V_S profiles generated from assorted mantle adiabats. We use the software package LitMod (*Afonso et al., 2008; Fullea et al., 2009*) to generate six geotherms defined by potential temperatures T_P ranging from 537 to 1337°C . LitMod combines petrological and geophysical modeling of the lithosphere and sub-lithospheric upper mantle. For each adiabat, LitMod calculates the shear velocity, density and attenuation as a function of temperature, pressure and bulk composition. Stable mineral assemblages and their physical properties are computed by solving a Gibbs energy minimisation scheme (i.e., *Perple_X, Connolly, 2005*). Bulk density and seismic velocities are calculated according to different

averaging of the mineral physical properties as described in *Afonso et al.* (2008) and *Fullea et al.* (2009). Figure 3.10 shows the output relationship of temperature with density, attenuation and shear velocity. With the decrease in temperature, the depth of the phase transformations from orthopyroxene to high-pressure Mg-rich clinopyroxene between 200–300 km (*Stixrude and Lithgow-Bertelloni*, 2007), the dissolution of the pyroxene into garnet on top of the ‘410’ discontinuity, and, the depth at which olivine goes to wadsleyite, are seen rising in the V_S and density models (Figure 3.10). The mantle adiabats are parallel and consistent with those of *Cobden et al.* (2008). We use our 1-D V_S model that has the lower-bound deep shear-velocity anomaly to estimate the temperature difference from the 1337°C adiabat (Figure 3.8d and 3.10). Temperature anomalies for the deep lithosphere beneath Qiangtang are within 490–740°C cooler than the 1337°C mantle adiabat (Table 3.1). Comparing the AK135 V_S model to the temperature estimates gives an uncertainty of $\pm 100^\circ\text{C}$.

A conceptual temperature estimate for the subducted Indian lithosphere beneath Tibet can be illustrated by “pushing down” the geotherm of the stable Indian Shield by *Priestley et al.* (2008). The S -velocity band of the better fitting models for West Lhasa (Figure 3.7a, region 1) are $\sim 5\%$ faster than the average V_S of cratons (Figure 3.11a, *Lebedev et al.* (2009)), suggesting that the lithospheric mantle is stronger and probably colder than the typical craton. Assuming that the subducting Indian slab had a similar geotherm to continental India prior the collision with Eurasia, and, that it will take tens of millions of years for the slab to internally warm-up to the surrounding temperatures (e.g., *Deal and Nolet*, 1999), we can then estimate the expected temperature anomaly between the deepened and the present-day Indian Shield geotherm. The Moho beneath West Lhasa is at about 75 km depth, 40 km deeper than continental Indian Moho (~ 35 km depth, e.g., *Gupta et al.* (2003); *Rai et al.* (2003)). Thus if the shield geotherm is deepened by 40 km, the Indian lithosphere below West Lhasa would have a temperature anomaly of $\sim 250^\circ\text{C}$ cooler (Figure 3.11a).

The deeper high shear speeds found beneath West Qiangtang have very similar velocities to those found in West Lhasa but at ~ 55 km deeper (Figure 3.11a and b). Hence, if the Indian geotherm is “pushed down” further to mimic India subducting underneath West Qiangtang, the temperature anomaly is about 700°C (Figure

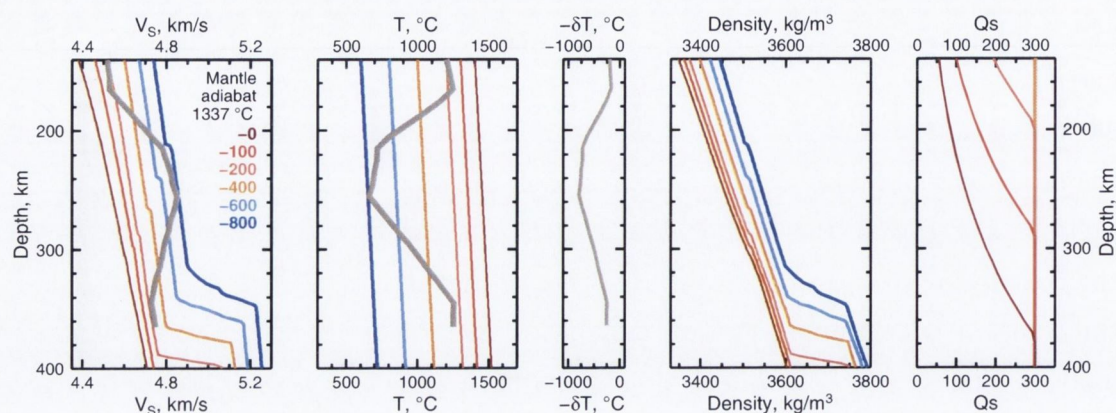


Figure 3.10: Estimation of temperature anomalies in the upper mantle. First, a set of reference 1-D models, each comprising the profiles of temperature, density, Q_s and V_s , is computed numerically using LitMod (Afonso *et al.*, 2008; Fullea *et al.*, 2009). For each assumed adiabat, LitMod calculates shear speed, density and attenuation as a function of temperature, pressure and bulk composition; an average mantle composition is assumed here (e.g., Fullea *et al.*, 2012). Note how the decrease in temperature causes decreases in the depths of the pyroxene phase transformation at 200–300 km, of the dissolution of the pyroxene into garnet above the ‘410’ discontinuity, and of the olivine-wadsleyite phase transformation nominally at 410-km depth. The profiles are calculated for a reference mantle adiabat with potential temperature T_p 1337°C (red) and for a series of temperature profiles that are cooler than the reference adiabat by the same amount at all depths, with the temperature decrease of up to 800°C (blue). The set of petro-physical models provides a mapping of shear velocities into temperature. Dark grey profiles: V_s for South-east Qiangtang (best-fitting model in Figure 3.8d) and the estimated temperature profile and temperature anomaly.

3.11b). This conceptual estimate for the subducted Indian lithosphere temperature anomaly is remarkably similar to the independent, more accurate petrophysical estimate (Figure 3.10). Taking into account the close proximity of both regions (West Lhasa and West Qiangtang), we consider that the similarities of the fast S velocities and layer thickness, and the agreement in the temperature estimates with that of a subducted Indian lithosphere as a signature that links the high V_s lithosphere below Qiangtang with the lithospheric mantle of West Lhasa and continental India. In view that the anomaly is north of the BNS, models of Indian steep-angle subduction are eliminated.

Measurements across Songpan-Ganzi are similar to central Qiangtang, low shear speed in the upper 100 km of the mantle underlain by high V_s below. Since

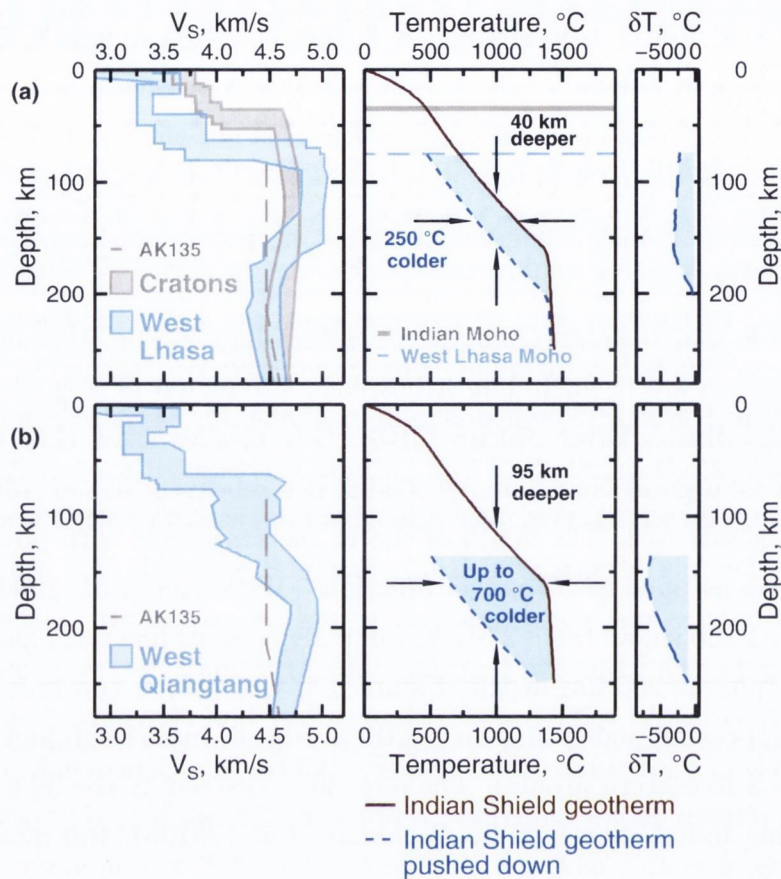


Figure 3.11: The thermal origin of high-velocity anomalies beneath the West Lhasa and West Qiangtang regions. (a, left): The S -velocity profile range for West Lhasa, with an average Moho at 75 km depth (light blue), and the S -velocity profile range for cratons, with an average Moho depth at 35 km (light grey) (Lebedev *et al.*, 2009). (a, centre): the Indian Shield geotherm (bold solid line) (Priestley *et al.*, 2008), and the same geotherm pushed down by 40 km (dashed line). 40 km is an estimate of how much the Indian lower crust and mantle lithosphere need to go down to underthrust the Himalayas and western Lhasa, with the upper crust removed. (a, right) The estimated temperature anomaly at depth. (b): Same as (a), but for West Qiangtang, with the Indian Shield geotherm pushed 95 km down.

the phase-velocity curves for this region are less broad, to reduce the misfit influence from the shorter periods, the data-synthetic RMS for the high V_S anomaly inversion tests are calculated using dispersion >70 s. Between 185–235 km depth the V_S is high with values between 4.63–4.88 km s^{-1} (Table 3.1). Note that the anomaly depth range is at the weaker end of the Rayleigh-wave dispersion sensitivity, and al-

though the S -velocity tends to be high similar to the profiles across Qiangtang, the V_S , temperature and bottom of the anomaly are weakly constrained. Thus southward subduction of Asia beneath Songpan-Ganzi is not ruled out. Further north, across the Kunlun Fault, 1-D S -velocity models for Qinling-Qilian region have continental-average mantle velocities and show no pronounced anomalies (Figure 3.7, region 7).

We conclude that beneath Qiangtang the deep high S -velocities and cold temperature estimates are of an Indian lithosphere subducted at a shallow angle, at least up to the Jinsha River Suture (JRS). Our profiles show that the high V_S beneath East Qiangtang and Songpan Ganzi lies relatively flat at 150 km depth (Figure 3.7, regions 4-6). The depth of the V_S increase agree with positive lithospheric interfaces mapped by S receiver functions (Wittlinger *et al.*, 2004; Yue *et al.*, 2012). Figure 3.12b shows the +2.5% V_S anomaly contour line from global surface-wave tomography at 200 km depth (Figure 3.3) which seems to border with the Kunlun Fault, possibly indicating the northern edge of India's subduction beneath Tibet (Figure 3.13 eastern profile). The large area covered by the high V_S can well be the "missing India" identified by Replumaz *et al.* (2010b), still intact, beneath the Tibetan Plateau.

3.3.4 Subducted Indian lithosphere-asthenosphere boundary

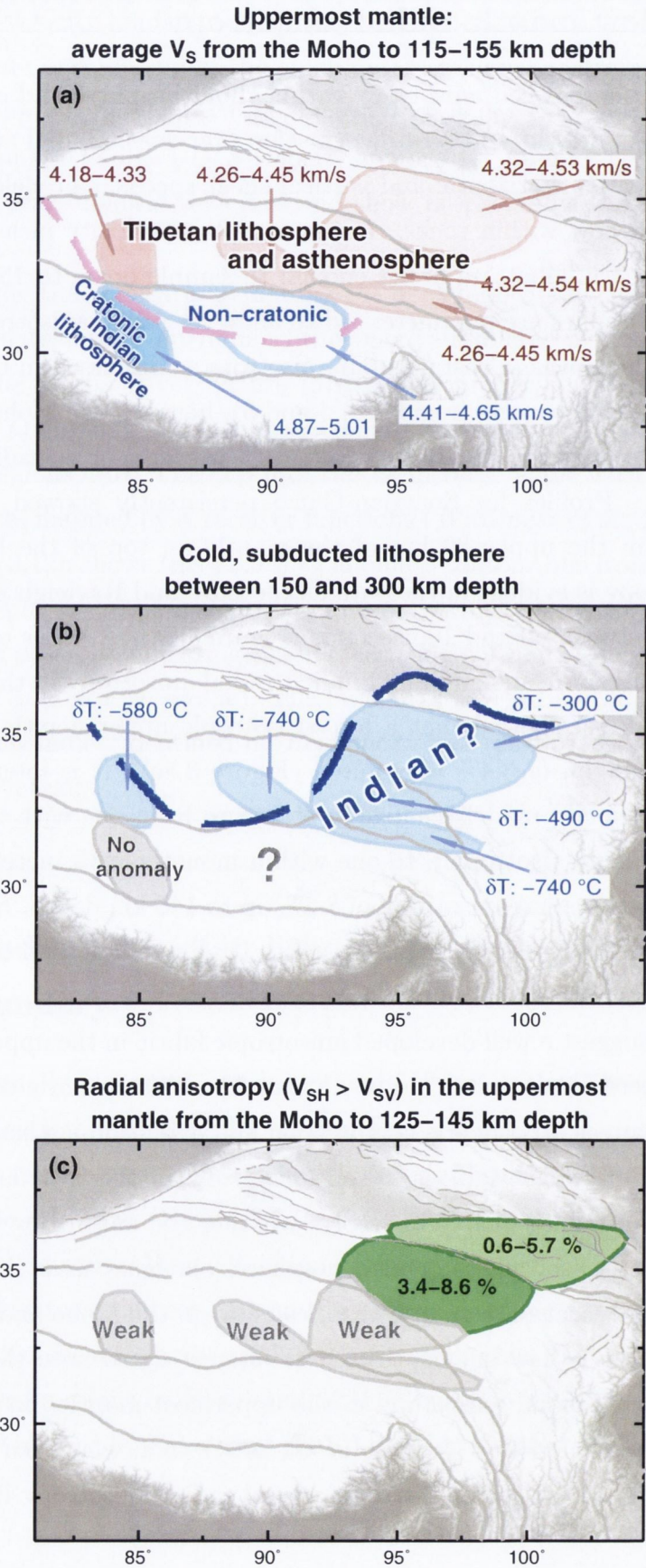
The LAB, normally identified as a reduction of V_S below the lithosphere, is noticeable in many regional profiles. In western Lhasa the LAB is clearly seen at the sharp velocity step between 150–175 km depth establishing the lithospheric mantle thickness to be ~ 100 km (Figure 3.7, region 1). The bottom of a similar 100 km fast layer deepens across West Qiangtang (region 2) down to 250 km and further down to 300 km towards east and north-eastern Tibet (regions 4-6). The depth of these negative shear-velocity gradients are deeper than the interfaces modelled from S receiver functions by Zhao *et al.* (2011), but, can be compared to the deepening interface imaged by Zhao *et al.* (2010) for central Tibet marked as the Indian LAB.

3.3.5 Shallow mantle radial anisotropy

Measurement of shear-speed anisotropy within the lithosphere and asthenosphere help us investigate present and past deformation processes. Radial anisotropy, the difference between the horizontal and vertical shear speeds, can yield information about strain and flow within rocks. Love-wave phase-velocity measurements for central and northern Tibet are broad enough to sample down to 150 km depths. Each of our 1-D profiles are parameterised to allow for radial anisotropy at crustal and upper mantle depths. Crustal radial anisotropy is discussed in Chapter 4.

Within Qiangtang terrane (regions 2, 4 and 5), between the Moho and 150 km depth, the best fitting profiles require weak ($<2\%$ average) or no radial anisotropy to fit the data. Profiles for Songpan-Ganzi persistently showed strong radial anisotropy within the upper 80 km of the mantle on top of the high S velocity. The anisotropy is evident in the data as the Love and Rayleigh phase-velocity curves diverge between 60 and 100 seconds (Figure 3.5g). A series of specific test inversions for Songpan-Ganzi, adding implicit radial anisotropy in the background V_S model ($V_{SH} > V_{SV}$), showed that a forced isotropic upper mantle fits the data badly and a minimum of 3.4% is required (Figure 3.8c). It is interesting to see how the test S -velocity models change in structure from one with a low-velocity zone at 125 km depth (isotropic), to one with a monotonic V_S increase below the Moho having strong radial anisotropy of 8.6% up to 145 km depth. Shallow upper mantle radial anisotropy is also required across Qinling-Qilian and is in the range of 0.6–5.7% between the Moho and 125 km depth.

Our models suggest a well developed anisotropic fabric in the upper mantle layers of north-eastern Tibet that could be a result from different or a combination of mechanical scenarios: lithospheric compression and extension, asthenospheric flow at the edge of the Indian slab, or frozen anisotropy in the Tibetan lithospheric mantle. Although we are unable to resolve whether the anisotropy is confined to the Tibetan mantle lithosphere or asthenosphere, our tests show that the strong radial anisotropy beneath Songpan-Ganzi lies between the Moho and above a high S -velocity anomaly (~ 145 km depth). It is interesting to note that where our models show strong radial anisotropy in the uppermost mantle, no crustal radial anisotropy has been modelled (*Shapiro et al.*, 2004), and, where our models show regions of weak or no anisotropy, strong crustal radial anisotropy has been indi-



cated — predominantly in western Tibet. *Shapiro et al.* (2004) suggested that the strength of crustal radial anisotropy is consistent with the thinning of the crust and flow within it. Outward flow of material from western Tibet might not only be restricted to the mid-lithosphere, but perhaps extend deeper across the surrounding northeastern regions. East-west fast propagation azimuthal anisotropy, considered as a record of the flow direction, have been modelled by different seismic studies: shear-wave splitting show an increasing delay time towards northern Tibet (*McNamara et al.*, 1994; *Sandvol et al.*, 1997; *Huang et al.*, 2000; *Herquel and Tapponnier*, 2005; *Lev et al.*, 2006; *Sol et al.*, 2007; *Chen et al.*, 2010; *Leon Soto et al.*, 2012), and, surface-waves studies suggest possible large deformation in the asthenosphere beneath the northeastern Tibetan margin (*Zhang et al.*, 2011). From our 1-D S -velocity profiles for Songpan-Ganzi we conclude that probable lateral flow is confined within the upper 80 km of the mantle, on top of a high V_S anomaly.

Figure 3.12 (facing page): Shear-speed distributions, temperatures, and radial anisotropy in Tibetan upper mantle. Shaded areas indicate the regions within Tibet for which the estimates were derived. (a) Uppermost-mantle-average V_S ranges constrained in this study. The depth ranges for which the averages are determined sample the Indian cratonic lithosphere in the West Lhasa and the Tibetan lithosphere and asthenosphere in central, northern, and eastern Tibet. Specifically, the depth ranges extend from the Moho down to 115 km (West Qiangtang), 135 km (East-central Qiangtang, South-east Qiangtang, and Songpan-Ganzi), 145 km (Central Lhasa), and 155 km (West Lhasa) (Table 3.1). Dashed, thick pink line: the approximate northern boundary of the high-velocity Indian lithosphere beneath western and southern Tibet at 110 km depth according to global surface-wave tomography (Figure 3.3, *Lebedev and van der Hilst* (2008)). (b) Temperatures within the anomalously cold upper mantle zones at 150-300 km depth, interpreted to contain the subducted Indian lithosphere. Thick, dashed blue line is a +2.5% anomaly contour at 200 km depth beneath Tibet from the global surface-wave tomography (Figure 3.3). (c) Uppermost mantle radial anisotropy in northeastern Tibet, in the depth ranges from the Moho to 125 km depth (Qinling-Qilian) and from the Moho to 145 km depth (Songpan-Ganzi).

3.4 Conclusions

In order to determine the mechanisms of lithospheric convergence between India and Eurasia, we measured accurate, broadband phase-velocity curves across central and northeastern Tibet, inverted them for 1-D shear-velocity profiles, and derived temperature estimates based on petro-physical modelling. According to the degree of similarity of measurements from neighbouring station pairs, we identified 7 regions within Tibet, each with relatively homogeneous structure within it.

For each region we performed an extensive series of non-linear, gradient-search inversions resulting in bundles of shear-velocity profiles characterising the mantle structure. We constrained the range of possible shear speed and radial anisotropy values within the upper mantle.

Our models suggest that the upper 75 km of the mantle beneath Tibet is made up of the Indian cratonic lithosphere in the west of the plateau and the Tibetan lithosphere and asthenosphere elsewhere. A cold, 75 km thick Indian lithosphere, comprising mantle lithosphere and, possibly, lower crust, underthrusts West Lhasa at 85°E to as far north as the BNS (Figure 3.13 central profile). This is evidenced by the pronounced high-velocity anomaly beneath the West Lhasa region. Under Central Lhasa, in contrast, V_S average over the uppermost-mantle depth range is close to continental-average values (Figure 3.12a).

The Tibetan mantle lithosphere, present beneath the Moho in the central, northern and eastern plateau, is warm and thin. The S -velocity average over the depth span including the mantle lithosphere and shallow asthenosphere are in the 4.18–4.54 km s⁻¹ range (Figure 3.12a), indicating relatively high temperatures. These low S -velocities across northern Tibet contrast with the high velocities in the southwest and, hence, render unlikely the models with the underthrusting of India beneath all or most of Tibet.

Surface-wave measurements are easily reconciled with published Moho depths and S_n wave velocities. Models with the S_n values equal to V_S just below the Moho, as well as models that have shear velocity matching S_n at depths 20 or 35 km below the Moho, are consistent with surface-wave data. The latter models also comply with the observations of inefficient propagation of high-frequency S_n and with observations of relatively fast P_n at sub-Moho depths. Surface-wave data

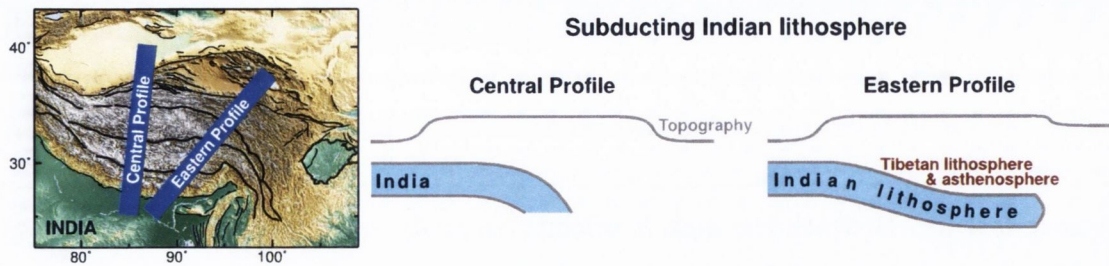


Figure 3.13: The complex subduction of Indian lithosphere beneath Tibet. Left: Elevation map of Tibet and the locations of the central and eastern profiles. Right: the steep-angle subduction of India beneath central Tibet and the shallow-angle subduction of India beneath eastern Tibet.

can also be fit with or without a low velocity zone below a thin Tibetan mantle lithosphere. For models with a low-velocity zone in the upper 100 km of the mantle, the top of the lower velocity zone (assumed LAB) shows general consistency with interfaces inferred from S receiver functions in western and central Tibet (Zhao *et al.*, 2010).

Shallow upper mantle radial anisotropy, indicative of the deformation pattern, is weak beneath the Qiangtang Terrane but strong beneath eastern Songpan-Ganzi ($>3.4\%$). Radial anisotropy is also required to fit the data to the north of the Kunlun Fault, beneath Qinling-Qilian (Figure 3.12c). Our models suggest a well developed anisotropic fabric in the upper layers of the mantle in northeastern Tibet, a region which, unlike western Tibet, has weak or no radial anisotropy in the crust and is experiencing crustal thickening (Shapiro *et al.*, 2004). Our tests confirm that the strong radial anisotropy beneath Songpan-Ganzi lies within the upper 80 km of the mantle, the depth range that contains the Tibetan lithosphere and asthenosphere and is above the high S -velocity anomaly located below 145 km depth.

At 150–300 km depth beneath central and northeastern Tibet, phase-velocity data require a major high-velocity anomaly, with a thickness and a shear-speed increase within it similar to those found within the Indian lithosphere underthrusting West Lhasa. The depth to the top of the fast anomaly shows agreement with the depths of interfaces mapped by S receiver functions (Wittlinger *et al.*, 2004; Yue *et al.*, 2012). The location of the high-velocity anomaly is also consistent with

that given by large-scale tomography (*Lebedev and van der Hilst*, 2008), although the depth range and amplitude of the anomaly are resolved much more accurately using the broad-band dispersion inversions in this study. The presence of high S velocities beneath both Qiangtang and Songpan-Ganzi confirms that this is an extensive feature. Beneath the high S -velocity anomaly, our profiles show a decrease in V_S values, down to global upper-mantle averages.

Petro-physical temperature estimates for the high S -velocity zone suggest that the deep lithosphere below Qiangtang is about 490–740°C cooler than normal asthenosphere at the same depth (Figure 3.12b). These temperature anomalies are consistent with those that would occur within subducted Indian lithosphere.

We thus conclude that the lithospheric convergence mechanisms beneath Tibet are as follows: underthrusting of India beneath Tibet in the west (as already evidenced by published tomographic models); underthrusting followed by steep-angle subduction beneath west-central Tibet (around 85°E; the West Lhasa and West Qiangtang regions in this study); and shallow-angle subduction of India in eastern Tibet, with the subducting Indian lithosphere reaching as far north as northern Qiangtang–Songpan-Ganzi.

4

Shear-velocity structure and radial anisotropy of the Tibetan crust

4.1 Introduction

Tibet is the world's largest and highest plateau (*Fielding et al.*, 1994), sandwiched between the northward-moving India to the south and the relatively stable Eurasia to the north (*Molnar and Tapponnier*, 1975). The tectonic blocks that make up the plateau (*Allègre et al.*, 1984), Figure 4.1, have been subjected to considerable deformation since ~ 50 My ago, following the continental collision (e.g., *Besse et al.*,

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1984; *Patriat and Achache*, 1984). How this convergence is accommodated today is still debated.

One end-member model invokes deformation primarily at narrow boundaries between “rigid blocks”, the boundaries being active faults (e.g., *Avouac and Tapponnier*, 1993). The other end-member model invokes “continuous deformation”, with viscous behaviour of the lithosphere (e.g., *Dewey and Burke*, 1973; *England and Houseman*, 1986; *England and Molnar*, 1997a). The different scenarios proposed for the continuous deformation of the lithosphere include “vertically coherent deformation”, in which the lithosphere deforms in approximately the same way within its entire thickness (e.g., *Flesch et al.*, 2005; *Wang et al.*, 2008), and mid-crustal “channel flow”, in which mechanically weak mid-lower crust undergoes flow that is distinctly different from the motions of the (stronger) layers above and below (e.g., *Royden*, 1997; *Beaumont et al.*, 2001). Detailed discussion of the different deformation mechanisms can be found in a number of reviews (e.g., *Klemperer*, 2006; *Searle et al.*, 2011). In this study we investigate the crustal shear-velocity structure beneath the plateau, and, interpreting our findings together with other geophysical and geological observations, aim to determine which mechanisms are most consistent with the data.

Shear velocities are commonly used to infer the composition and physical state of the rocks at depth, including their temperature; estimates of the melt fraction and viscosity can also be derived. An intra-crustal low velocity zone (LVZ) has been inferred for Tibet from intermediate- and short-period surface-wave observations, using one-dimensional (1-D) crustal models (e.g., *Chun and Yoshii*, 1977; *Romanowicz*, 1982; *Kind et al.*, 1996; *Cotte et al.*, 1999; *Rapine et al.*, 2003) and regional tomographic maps (e.g., *Guo et al.*, 2009; *Yang et al.*, 2012). The LVZ and the low resistivity zone at the same depth detected in magnetotelluric studies (MT) (e.g., *Chen et al.*, 1996; *Wei et al.*, 2001; *Arora et al.*, 2007) have been interpreted as evidence for the existence of a weak, partially molten layer within Tibetan mid-lower crust that is prone to viscous flow (*Nelson et al.*, 1996). Whether such a low-viscosity layer is present beneath the entire plateau is debated. Uncertainty remains over how representative the seismic and MT models are for the whole of Tibet, and to what extent they may be affected by anisotropy.

Tibet’s crust is characterised by strong seismic anisotropy, both radial (*Shapiro*

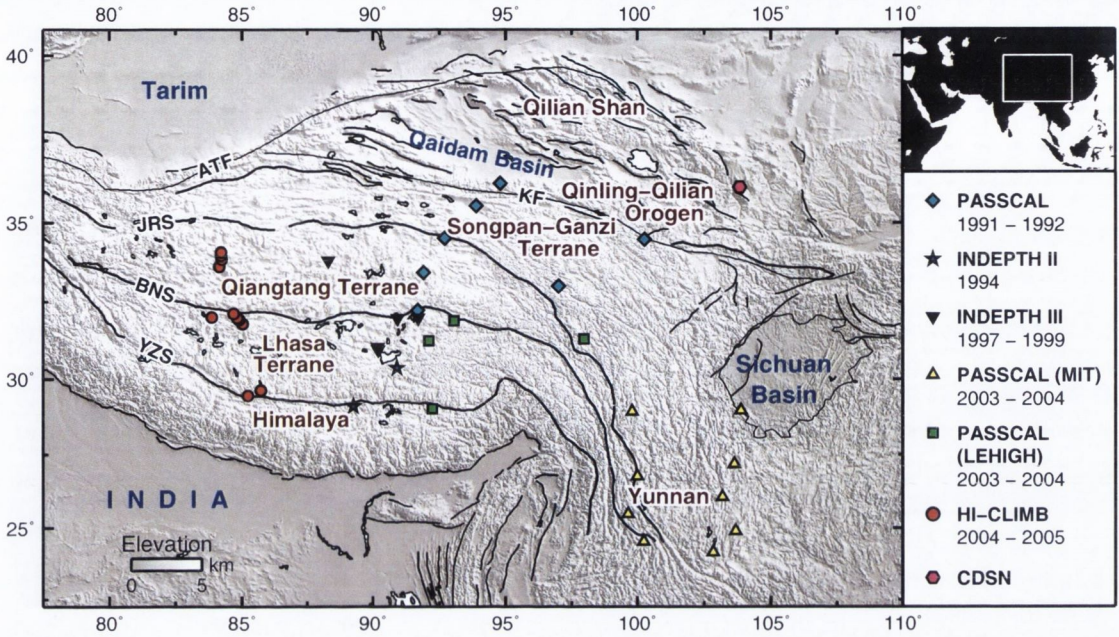


Figure 4.1: Seismic stations used in this study. Different symbols correspond to different networks deployed across the Tibetan Plateau. Black lines indicate sutures and faults: YZS (Yarlung-Zangbo Suture), BNS (Bangong-Nujiang Suture), JRS (Jinsha River Suture), KF (Kunlun Fault) and ATF (Altyn Tagh Fault), which divide Tibet in major terranes. White shade highlights high elevation. Top right map: Location of the map region in Asia.

et al., 2004) and azimuthal (*Huang et al.*, 2004; *Su et al.*, 2008). Most published shear-speed models, however, have been computed from Rayleigh-wave dispersion and, thus, represent vertically polarised shear waves only (V_{SV}). Hence, it is still uncertain if the reported decrease in shear velocity within the Tibetan mid-lower crust reflects the isotropic-average seismic velocity (which can be used as a proxy for partial melt and viscosity), or whether the low V_{SV} velocities are a result of strong radial anisotropy, instead. Low S velocities arising from higher temperatures and melts would decrease both V_{SV} and V_{SH} , whereas radial anisotropy, due to deformation, would result in divergence of the V_{SV} from V_{SH} values.

The 1-D model for western Tibet presented by *Shapiro et al.* (2004), inferred from both Love- and Rayleigh-wave data, did not require a LVZ for both V_{SV} and V_{SH} . 1-D models by *Duret et al.* (2010), averaging over long paths across Tibet, show only a small decrease in V_{SV} and an increase in V_{SH} from the upper-

middle to the lower-middle crust. In contrast, the recent tomography of *Yang et al.* (2012), performed using ambient-noise, cross-correlation measurements for numerous inter-station paths across Tibet, shows a very pronounced LVZ in V_{SV} , laterally continuous across large parts of the plateau and with no obvious correlation to surface tectonics.

The distribution and, in particular, magnitude of the radial anisotropy itself are also still uncertain. *Duret et al.* (2010) confirmed the occurrence of strong radial anisotropy in Tibet, as previously reported by *Shapiro et al.* (2004), by computing average anisotropic profiles along long paths across the plateau. Tighter constraints on the shear-velocity and radial-anisotropy ranges within Tibetan crust and their lateral variations are important in order to advance our understanding of the current state and the dynamics of the orogen.

The main purpose of this study is to determine the isotropic structure and radial anisotropy and their lateral variations within the Tibetan crust. To this end, we measure accurate, broadband Love and Rayleigh surface-wave phase velocities in different regions across the plateau. Inverting the measurements for 1-D, radially anisotropic, shear-velocity models, we determine the ranges of S -velocity and radial anisotropy values consistent with the data. We detect a pronounced low-velocity layer within the crust across the high plateau and derive estimates of partial melt within it. While the probable very low viscosity within this layer shows potential for pervasive flow within it, radial anisotropy characterises the laterally varying pattern of the deformation that has actually occurred.

4.2 Data and measurements

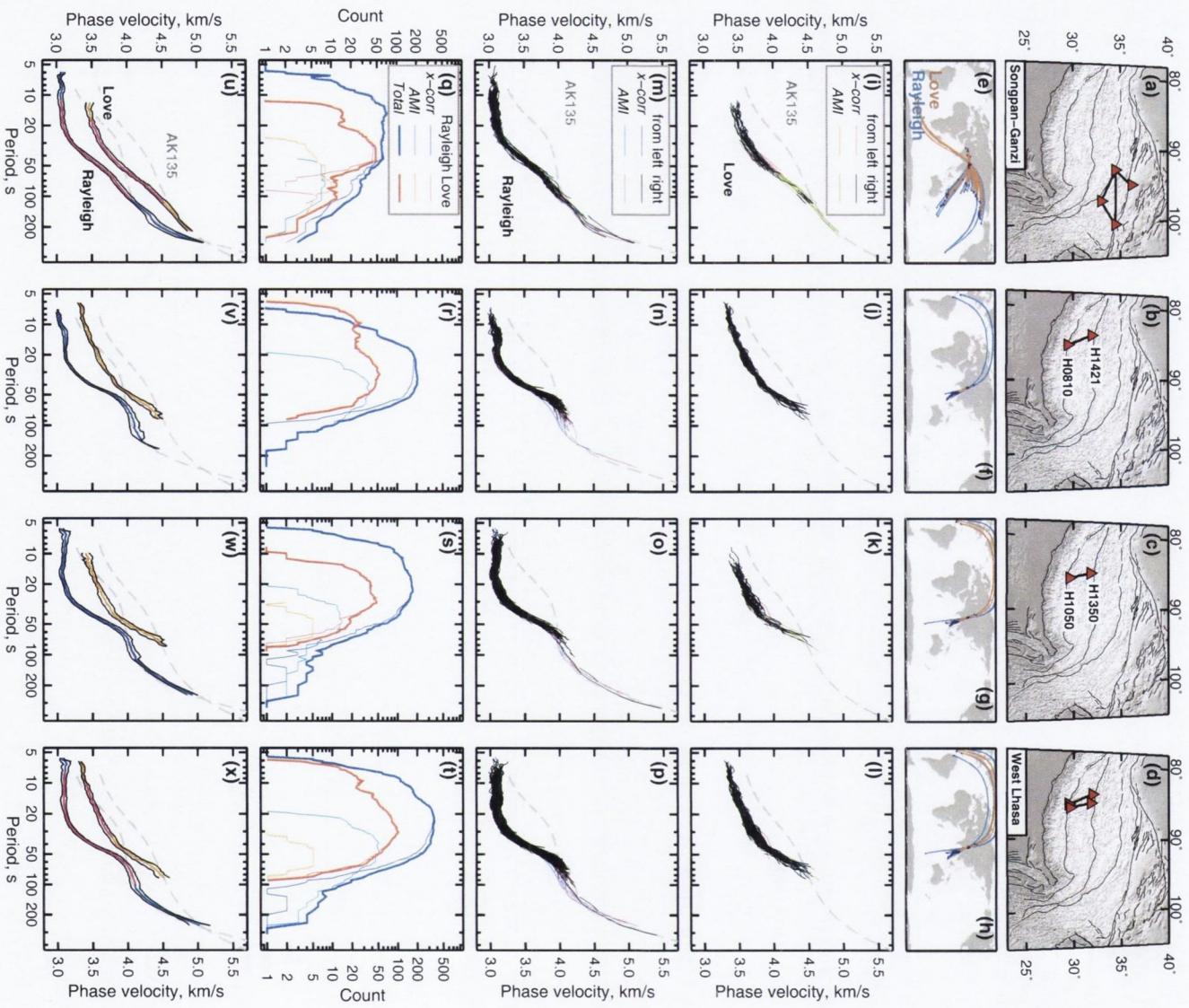
The recording of seismic surface waves propagating between two stations can be used to derive important information about the structure, geophysical properties and rheology of rocks between the stations. We measured broad-band, inter-station, phase-velocity curves of Love and Rayleigh surface waves for central and eastern Tibet and surrounding regions. The pairs of stations were selected such that a large number of earthquake recordings suitable for the measurements were available. In total, 34 station pairs were selected. The stations are broadband instruments belonging to various temporary networks (PASSCAL (91/92) (*Owens*

et al., 1993), INDEPTH II and III (*Nelson et al.*, 1996; *Huang et al.*, 2000), HICLIMB (*Nábělek et al.*, 2005), PASSCAL (Lehigh) (*Sol et al.*, 2007), PASSCAL (MIT) (*Lev et al.*, 2006)) and from the permanent China Digital Seismic Network (Figure 4.1). The seismograms were retrieved from the Incorporated Research Institutions for Seismology (IRIS) Data Management Center (DMC). Automatic quality control on the data is performed to check for clipping and incomplete seismograms. The seismograms are then response-corrected to displacement, and the horizontal components are rotated to obtain the transverse component.

Surface-wave dispersion measurements are made using a combination of two different, independent techniques: direct station-station cross-correlation of seismograms (*Meier et al.*, 2004) and the derivation of inter-station phase velocities from source-station measurements, performed using automated, multi-mode waveform fitting (*Lebedev et al.*, 2005). In the first method, two vertical-component (for Rayleigh waves) or transverse-component (for Love waves) seismograms with recording of the same earthquake at both stations of a pair are cross correlated. The cross-correlation function is filtered with a frequency-dependent bandpass filter. In order to enhance the signal-to-noise ratio and to down-weight side lobes resulting from correlations of the fundamental mode with scattered waves or higher modes, a frequency-dependent time window is applied. The fundamental-mode, Rayleigh-wave phase velocity is then calculated from the phase of the cross-correlation function and the difference in source-station distances (*Meier et al.*, 2004). The strength of this technique is that it can measure accurate phase velocities at both short (~ 5 s) and long (> 200 s) periods.

In the second method, we use the Automated Multimode Inversion (AMI) of surface and S -wave forms (*Lebedev et al.*, 2005) to simultaneously fit S , multiple S and surface waves, using synthetic seismograms generated by mode summation. AMI is run on each of the two stations of a pair for the same earthquake. From the waveform fits of both seismograms, we extract the fundamental-mode phase velocities and then calculate the inter-station phase-velocity curves. AMI's key strength, for our purposes here, is in measuring phase-velocities of long-period surface waves even when they interfere with energetic body waves.

To decrease the effects of surface-wave diffraction we restrict the selection of earthquakes to those approximately on the same great circle path with the two



stations (station-earthquake back azimuth within less than 10° from the station-station back azimuth). We select earthquakes listed in the Centroid Moment Tensor catalogue (e.g., *Dziewonski et al.*, 1994), with moment magnitudes larger than about 4.8, and epicentral distances to the stations from 860 to 16,400 km. In an interactive selection, the smoothest portions of the fundamental-mode Rayleigh- and Love-wave dispersion curves are selected. Kinks or roughness in the curves are usually caused by diffraction of the fundamental mode or its interference with higher modes; rough portion of the curves are not accepted. Obvious outlier curves, away from the average, are also not accepted. Careful attention was paid to visually identify systematic differences in phase-velocity measurements that are propagation-direction dependent; these would be indicative of incorrect station timing or a measurement bias due to wave propagation, from at least one direction. For the station pairs selected, no direction-dependent biases were detected. Averaging over a large number of one-event measurements, made using recordings of earthquakes in different source regions, in different directions from the station pair, results in robust measurements in broad period ranges. The rigorous selection of only the smooth parts of the dispersion curves increases the accuracy of the phase-velocity

Figure 4.2 (facing page): Phase-velocity measurements for station pairs and for region averages. Top maps (a–d) show the location of the station pairs (red triangles), and the inter-station paths (black lines). The global maps (e–h) show earthquakes (dark blue dots) and source-station great circle paths corresponding to Love-wave (brown) and Rayleigh-wave (cyan) measurements. The measured Love- and Rayleigh-wave dispersion curves are shown in (i–l) and (m–p), respectively. Different colours indicate the method and the direction each measurement corresponds to. AMI: Measurements from the Automated Multimode Inversion technique (*Lebedev et al.*, 2005). x-corr: Measurements from the cross-correlation technique (*Meier et al.*, 2004). Dashed grey curves show the phase-velocities computed for the AK135 reference model (*Kennett et al.*, 1995). Count graphs (q–t) show the total number of measurements at each period for Love and Rayleigh waves (red and blue respectively); different colours show the count from different methods. The bottom frame shows the resulting average dispersion curves for Love and Rayleigh waves and their standard deviations (black curves) (u–x); pink curves show the most robust measurements selected to be used for the inversions. The two columns in the middle illustrate inter-station measurements for individual station pairs. The columns on the left and on the right show examples of region-average phase-velocity measurements, using all the data from all inter-station pairs within the regions.

curves.

Figure 4.2 shows examples of Rayleigh- and Love-wave dispersion measurements. HI-CLIMB station pairs H0810-H1421 and H1050-H1350 are on nearly the same great-circle paths (Figures 4.2b and c) and thus have almost the same set of earthquakes used in the measurements (Figures 4.2f and g). Their close proximity also means that the surface waves sample nearly the same structure, resulting in very similar dispersion curves (Figures 4.2 j-k and n-o). Note the numerous, superimposed, consistent measurements from the different methods and opposite source directions (Figures 4.2r and s). Many of the phase-velocity measurements start from very short periods (<10 s), sampling primarily the shallow, upper-crustal structure. Between 10 and 40 s periods, phase velocities are very low compared to global averages, due to the very thick crust and low shear velocities within it. Between 40 and 60 seconds, phase velocities increase rapidly with period, approaching the reference values computed for the 1-D global reference model AK135 (*Kennett et al.*, 1995). For each path, the average phase velocity and the standard deviation is calculated at each period (Figures 4.2v and w). Dispersion differences between the two paths are mostly at periods <20 s, likely due to shallow, upper crustal heterogeneity.

Although the average dispersion curves derived from our measurements are very smooth, and mostly with small standard deviations, there may be undetected biases remaining in the measurements. In this paper, we base our inferences and conclusions only on robust, persistent patterns that are present in the data measured along multiple inter-station paths.

4.2.1 Region-average measurements

Nearby station pairs are often contained within the same tectonic units and exhibit similar phase-velocity curves (Figures 4.2b and c). Thus, station pairs with similar dispersion curves between 15 and 50 seconds period were grouped together, with the measurements combined so as to characterise the entire region. Figures 4.2l and p, show the combined Love- and Rayleigh-wave measurements from all the individual station pairs across West Lhasa (Figures 4.2b and c). Combining dispersion curves together yields more robust and broadband measurements, averaged from many hundreds of Love- and Rayleigh-wave curves. The region-average

phase-velocity curve and the corresponding standard deviation are calculated at each period (Figure 4.2x). Only the smoothest part of the curve is then selected, to be used in inversions (pink curve in the bottom frame).

In another example, Figure 4.2a shows the measurements for Songpan-Ganzi, combined from a number of differently aligned station pairs. Unlike West Lhasa, the earthquakes used for the measurements are distributed more widely across the globe since the inter-station paths have very different azimuths (Figures 4.2a and e). The dispersion curves from each individual station pair are very similar, even though the pairs are at significant distances from each other. The similarity of the dispersion curves indicates relatively homogeneous structure across the entire region.

We identify nine sub regions, with weak heterogeneity within each: West Lhasa, Central Lhasa, West Qiangtang, East-central Qiangtang, South-east Qiangtang, Songpan-Ganzi, North Yunnan, South Yunnan and Qinling-Qilian. Figure 4.3 shows the inter-station pairs within each region and also a comparison of the dispersion curves from different regions. All the 9 regions have very low Love- and Rayleigh-wave phase velocities for periods less than 50 seconds, when compared to phase velocities computed for the global reference model AK135 (*Kennett et al.*, 1995). Regions at highest elevations within Tibet display the slowest phase velocities (Figure 4.3, regions 1–6). Between 10 and 20 s, Rayleigh- and Love-wave phase-velocity variations of the 9 regions are less than 0.2 km s^{-1} and are likely to be due to sedimentary-layer and other upper-crustal heterogeneity.

The Rayleigh-wave dispersion curve from West Qiangtang shows a dip at ~ 20 seconds. Between 30 and 40 second period, all six regions in the central, high-elevation part of the plateau are characterised by remarkably similar Rayleigh-wave phase velocities, suggesting a largely homogeneous structure within the deep crust. At periods longer than 40 s, the phase velocity in West Lhasa increase rapidly with period, to exceed the global reference at 60-second period. Phase velocities in other high-elevation parts of Tibet increase with period more gently.

The dispersion curves from Qinling-Qilian and South Yunnan are distinguished by their high velocities between 20 and 50 second periods and are very similar, indicating very similar crustal structure at the northeastern and southeastern margins of the plateau. Dispersion measurements from North Yunnan display intermediate

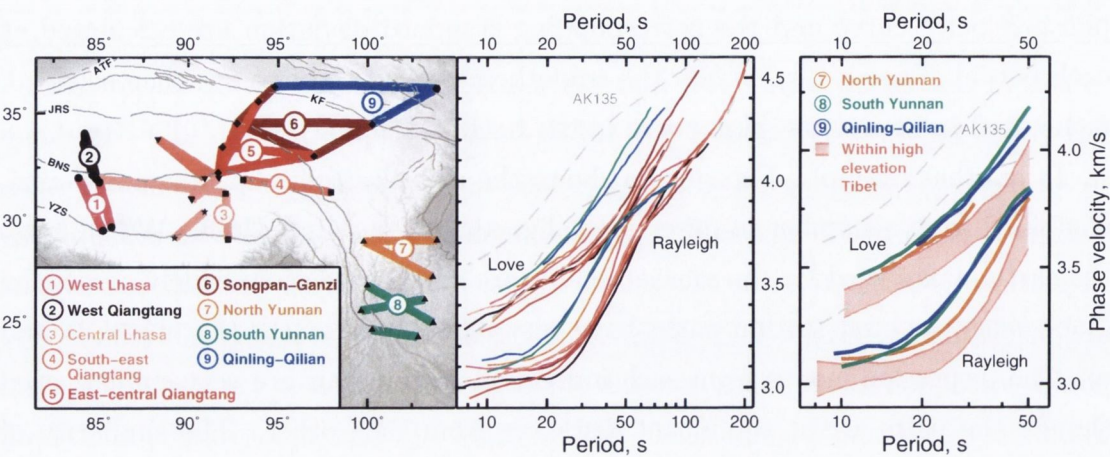


Figure 4.3: Broad-band phase-velocity curves measured in regions across the Tibetan Plateau. Left: Map showing inter-station paths within 9 colour coded regions. Different symbols indicate stations of different networks, as in Figure 4.1. Centre: the 9 Love-wave and 9 Rayleigh-wave, region-average, phase-velocity curves. Right: Comparison of region-average dispersion curves within the high-elevation part of the plateau (pink shade) and those from outside the high plateau. Dashed grey curves: Dispersion curves computed for the AK135 reference model (*Kennett et al., 1995*).

phase velocities, higher than in the central but lower than in the outermost parts of the plateau, indicating gradual changes in the crustal thickness and shear-velocity structure from the central towards the southeastern part of Tibet (Figure 4.3).

4.2.2 Surface-wave depth sensitivity

Rayleigh and Love waves sample the Earth differently and are sensitive to different depth ranges at the same period. Figure 4.4 shows Fréchet derivatives of the fundamental-mode Rayleigh and Love phase velocities with respect to shear velocity at depth, for 4 different periods at which the waves sample the thick Tibetan crust. The normalised sensitivity kernels are computed for a 1-D shear-velocity model of West Lhasa. At a 10 s period, the peak sensitivity of Rayleigh waves is at about 10 km depth; the maximum sensitivity depth and the depth span of sensitivity increases with period. Love waves are sensitive to shallower depths at the same periods. Joint inversion of broad-band, Love and Rayleigh dispersion curves can put tight constraints on isotropic-average shear speeds and radial anisotropy within the entire crust.

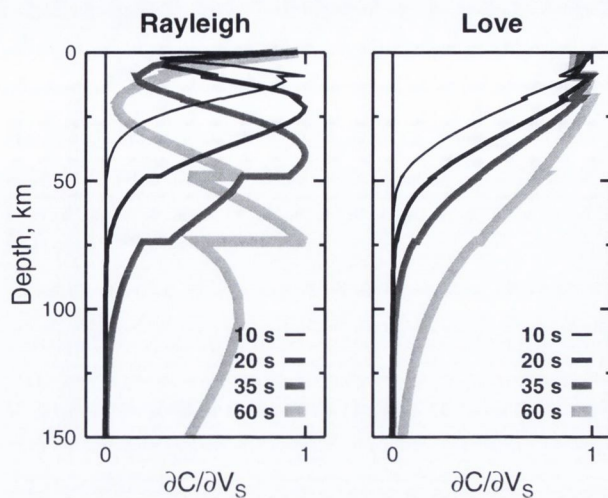


Figure 4.4: Depth sensitivity kernels for phase velocities of the fundamental-mode Rayleigh and Love waves. The curves for different periods are scaled independently. These kernels are computed for a 1-D shear-velocity profile of West Lhasa, with a four layer crust and a Moho discontinuity at 74 km depth.

4.3 Inversion of phase-velocity curves for shear-velocity profiles

Shear velocities are sensitive to temperature and composition and can offer important information on the state of the rocks at depth. The advantage of inverting an inter-station phase-velocity curve for a 1-D profile is that the model is free from lateral smoothing that characterises 3-D models. Furthermore, the use of local measurements brings the advantage of their simple, clear relationship to the local Earth structure. Rayleigh and Love-wave dispersion curves are inverted simultaneously for an isotropic average shear speed profile ($V_{S(\text{avg})} = (V_{SV} + V_{SH})/2$) and a radial anisotropy profile ($(V_{SV} - V_{SH})/2$), where V_{SV} and V_{SH} are the vertically and horizontally polarised shear speeds, respectively. The phase-velocity curves used for the inversion are re-sampled at a logarithmic sample spacing to balance the structural sensitivity along the broad-band dispersion curves.

The inversion is a non-linear, least square, Levenberg-Marquardt gradient search, run from MATLAB®. For each perturbation in the shear-speed profile, the Love- and Rayleigh-wave synthetic phase velocities are computed from the V_{SH} and V_{SV} models, respectively, using a suitably fast version of the MINEOS modes code (Masters, <http://geodynamics.org/cig/software/mineos>). The non-linear inversion algorithm minimises the data-synthetic misfit and converges to the true best-fitting solution. The small but non-negligible influence of the compressional-wave velocity ($V_{P(\text{iso})}$) on the Rayleigh-wave dispersion is taken into account by

keeping a fixed ratio to the isotropic-average shear speed ($V_{S(\text{avg})}$), same as in the starting (background) model.

The inversion is parameterised using boxcar and triangular basis functions, each used for two independent parameters, one for the isotropic average V_S perturbation and the other for radial anisotropy. In most inversions, the depth of the Moho and 3 intra-crustal discontinuities are also inversion parameters. With 15–20 depth basis functions spanning the crust and upper mantle and mild norm damping, the inversions can fit the data closely with relatively smooth, 1-D shear-velocity models.

4.3.1 Key parameters for the crust

Changes in the shape of the dispersion curve at short and intermediate periods indicate heterogeneity in the crust. How many parameters are necessary to capture the crustal shear-velocity structure? In order to determine what general structure best describes the data, we first perform an over-parameterised test inversion with no intra-crustal discontinuities. Ten triangular basis functions within crustal depth range provide ample freedom for the V_{SH} and V_{SV} models to perturb and fit the short period data (Figure 4.5a). Another 10 triangular basis functions are defined to parameterise the upper mantle down to the transition zone, in order to fit any anomalies required by the longer periods. The depth of the crust-mantle discontinuity is also included in the parameterisation.

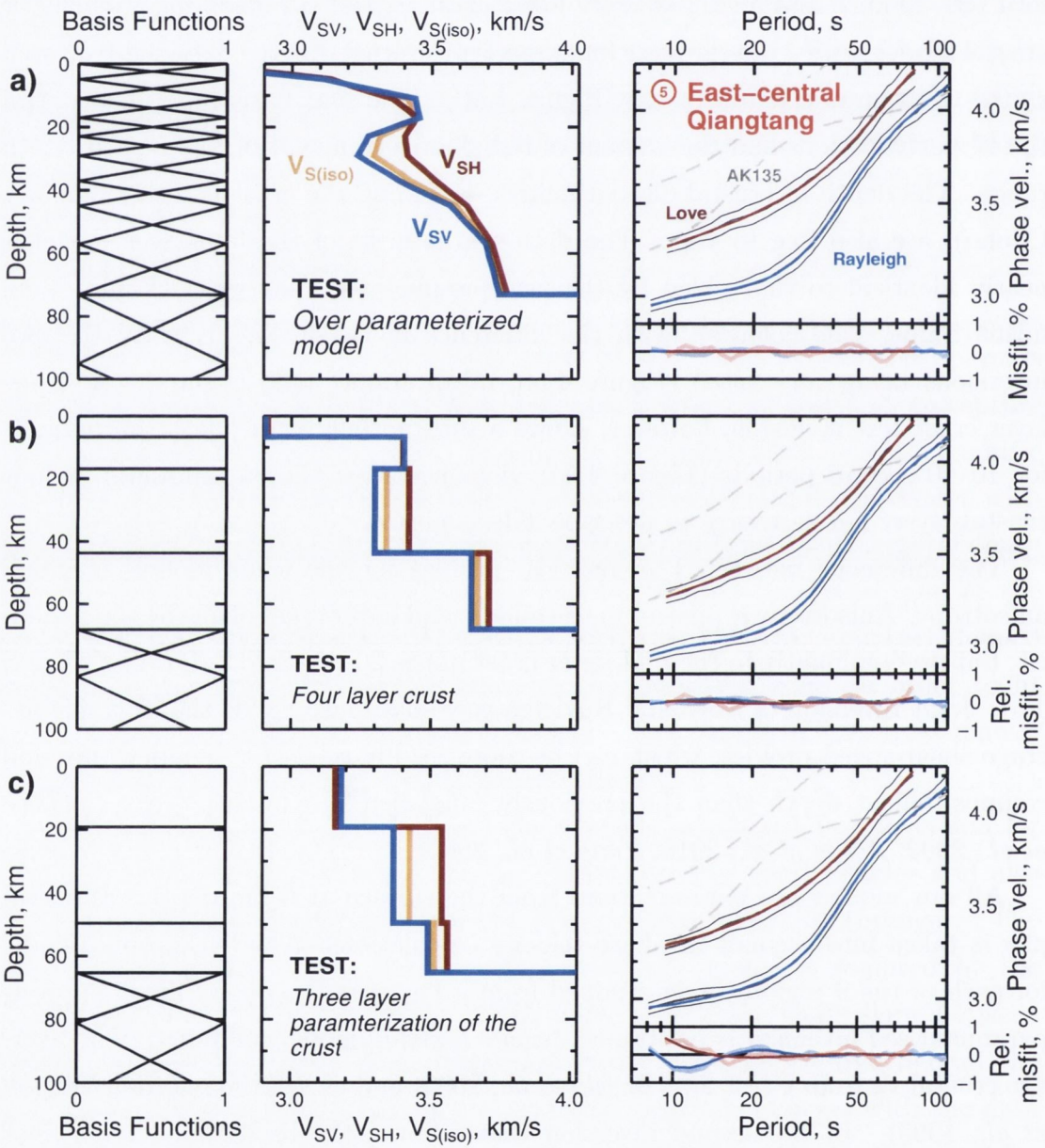
Figure 4.5a shows the model from the over-parameterised test inversion for East-central Qiangtang, showing the V_{SV} (blue) and V_{SH} (red) profiles and also the Voigt isotropic average shear speed ($V_{S(\text{iso})}=(2V_{SV}+V_{SH})/3$) (orange). The conspicuous reduction in V_S between 20 and 40 km depths is required by the Rayleigh-wave phase-velocity curve, which shows a decrease in its slope between 12 and 20 seconds (Figure 4.5a, dispersion frame, dark grey curve behind the synthetic dispersion). The misfit of the synthetic dispersion curves with the data is very low for both Rayleigh and Love wave, well below $\pm 0.5\%$ (Figure 4.5a, misfit frame, Love (pale red) and Rayleigh wave (pale blue)). The remaining data-synthetic misfit, larger for Love waves, is due to noise in the measurements. The $V_{S(\text{iso})}$ (and V_{SV}) model in Figure 4.5a has structure similar to that of the published V_{SV} models for northwest Himalaya by *Caldwell et al.* (2009), although somewhat

faster.

The results of the over-parameterised inversion show four basic layers within the crust: a 7–10 km thick upper crust with relatively low velocities in it, higher-velocity upper-middle crust (10–20 km), lower-middle crust with lower velocities in it (20–40 km), and higher-velocity lower crust. A test inversion for a four-layer crustal model, with 4 boxcar basis functions for the crust, fits the data nearly as well as the over-parameterised model (Figure 4.5b). The four basis functions control the V_S perturbations and the amount of radial anisotropy within the boxcar depth range. The depth of crustal discontinuities, including the crust-mantle boundary (Moho), are also free to vary. The data-synthetic fit of the four-layer model is nearly identical to that given by the over-parameterised inversion (Figure 4.5b, misfit frame, pale colours), with the difference in the misfit given by the two inversions being very small (Figure 4.5b, misfit frame, bold colours). A three-layer crust test inversion, however, shows a substantially worse data-synthetic fit for 10–20 second periods (Figure 4.5c), demonstrating that a minimum of four crustal layers are necessary to describe Tibetan crust.

The difference between V_{SH} and V_{SV} profiles for the same models is radial anisotropy. Anisotropy is present in the middle and lower crust of all the three models, but predominantly in the mid-lower crust in the 20–45 km depth range (Figure 4.5). Joint inversion of Love and Rayleigh-wave data also yields the isotropic average shear speed profiles, which can be more readily related to temperature and composition at depth than the previously published V_{SV} models (e.g., *Caldwell et al.*, 2009; *Acton et al.*, 2010; *Yang et al.*, 2012).

All our models are parameterised from the surface at 0 km depth. Topography is taken into account in phase-velocity calculations. The background model for each of the 9 regions is determined from a two-step inversion. First, an over-parameterised inversion is performed using a starting model composed of an average crustal V_S from Crust 2.0 (*Bassin et al.*, 2000) and an AK135 mantle (*Kennett et al.*, 1995). In the second inversion that follows, the background model has crustal discontinuities, Moho depth, and the average crustal V_S determined in the first inversion, and with the mantle set, again, to the AK135 reference model. All the models shown fit the data within the standard deviation.



4.3.2 Investigation of parameter ranges

The shear-velocity models yielded by the inversions are non unique; many different models can fit the data almost equally well. Small variations in the thickness and S velocity of a layer can be compensated by the velocities and thicknesses of the layers above and below. For example, the Moho depth from the inversions of East-central Qiangtang trades off with the lower crust shear velocity, shallower for a low V_S and deeper for a high velocity, with little effect on the misfit for periods >40 s (Figures 4.5b and c).

Rather than inverting dispersion curves for a single best-fitting S -velocity profile, we then perform a series of test inversions, comprising multiple grid searches within specially selected parameter sub-spaces. These inversions are targeted to determine ranges of V_S and radial anisotropy within specific crustal depth ranges and are designed to answer questions such as: how low or high is the shear velocity

Figure 4.5 (facing page): Three test inversions using different crustal parameterisation for East-central Qiangtang. (a) An over-parameterised test inversion. Left: Ten triangular basis functions parameterise the crust, to allow the inversion to fit the data. One discontinuity parameter is used to allow Moho depth changes. Center: The best-fitting radially anisotropic S -velocity model, comprising the profiles of V_{SV} (blue), V_{SH} (red), and the Voigt isotropic average $V_{S(iso)}$ $((2V_{SV}+V_{SH})/3)$, orange). The model does not include any intra-crustal discontinuities. Right, upper frame: Synthetic Rayleigh- and Love-wave dispersion curves. Dark grey curve: measured dispersion curves (hidden behind the synthetic curves). Thin black lines show the standard deviations of the data. Dashed grey curves: Dispersion curves computed for the AK135 reference model (*Kennett et al.*, 1995). Right, lower frame: The data-synthetic misfit for Love waves (thick pale red) and Rayleigh waves (thick pale blue). (b) Test inversion with a four layer crust. Left: Four boxcar basis functions parameterise the crust to allow for V_S perturbations; 4 discontinuity parameters allow for depth changes of three intra-crustal discontinuities and the Moho. Center: Best-fitting S -velocity model, depicted as in (a). Right: Corresponding synthetic phase velocities and data-synthetic misfit (thick pale curves) and the misfit relative to the synthetic dispersion curves from (a) (thin bold curves). (c) Test inversion with a three-layer crust. Left: Three boxcar basis functions parameterise the crust to allow for V_S perturbations; 3 discontinuity parameters allow for depth changes of two intra-crustal discontinuities and the Moho. Center: Inverted S -velocity model, shown as in (a) and (b). Right: dispersion curves and misfits, shown as in (b). All the inversions have upper mantle parameterisation sufficient to fit the long period data.

within the middle crust, or, what is the strength of the radial anisotropy in a depth range? The inversions are parameterised in a way that the property under investigation is fixed within the depth range in question, while all the other parameters remain free to perturb. Hence, the inversion is set up to find the best model that fits the data under the restricted condition. For example, an inversion is set with a fixed V_S between 20–45 km depth while the other parameters, such as the V_S of the other crustal layers, Moho and the upper crustal discontinuity, as well as the anisotropy within all the crustal layers are free to change. For each new inversion the V_S in the background model is incremented by a small step of 0.025 km s^{-1} . The data-synthetic fits achieved in each inversion provide the quantitative criterion for which models are the most consistent with the data.

This approach is different from the grid search applied by *Caldwell et al.* (2009), who performed a series of inversions by varying the V_S gradient from the surface down to 100 km depth of different starting models with no Moho discontinuity. Our approach also differs from Monte-Carlo methods (e.g., *Shapiro et al.*, 2004; *Yao et al.*, 2008, 2010; *Huang et al.*, 2010; *Duret et al.*, 2010; *Yang et al.*, 2012), although, like them, it produces a suit of best-fitting models. The main advantage of our approach is that it explores the complete ranges of parameters identified as most relevant, with no *a priori* assumptions on how broad these ranges may be.

Figure 4.6 shows three test inversion series set up to delimit the ranges of shear velocity between 20 and 45 km depth and radial anisotropy averages over the 20–45 km depth range and over the entire crustal thickness. For each test we plot the Voigt isotropic average shear speed ($V_{S(\text{iso})}$), the radial anisotropy ($V_{SH} - V_{SV} / V_{S(\text{iso})}$), the corresponding Love- and Rayleigh-wave synthetic dispersion curves, and the data-synthetic misfits. Plots of V_S , dispersion curves, and misfits zoom on the crustal depths and periods. The tests for the mid-crustal shear-velocity range show expected trade-offs between shear velocities in neighbouring layers and, also, radial anisotropy. When the S velocity between 20 and 45 km depth is set high, both the upper and lower crustal velocities decrease whilst the radial anisotropy becomes stronger in order to push the synthetic Rayleigh-wave phase velocity lower and fit the data (Figure 4.6, Songpan-Ganzi). Similarly, a very low V_S within the mid-crust results in a higher V_S in the adjacent upper and lower crustal layers and a decreased radial anisotropy. The best-fitting S -velocity profile,

shown in red, shows an intermediate V_S in the mid-crustal depth range and the adjacent layers. The divergence of the synthetic dispersion curves corresponding to the different test models is apparent at 20–40 s periods, most sensitive to the LVZ depth range, especially for Rayleigh waves. The same approach is used to determine the ranges for mid-crustal radial anisotropy and crustal-average radial anisotropy.

The data-synthetic misfit given by the best fitting model (Figure 4.6, red curve in the misfit frame) is small, much less than 0.5% at all periods. The remaining misfit is due to noise in the data. In order to establish a quantitative criterion for which models are most consistent with the measurements, we, instead of setting a data-synthetic misfit threshold for all the samples (which could be biased by a single noisy sample), use the root-mean-square (RMS) misfit computed over the length of both the Love- and Rayleigh-wave curves, up to the 60 second period. Parameter ranges determined in the tests are such that all the models with the parameter in question (e.g., lower-middle-crustal S velocity) within the range yield a misfit smaller than the global-minimum misfit plus a constant (2.5 m s^{-1}). The colour scale of the profiles and the corresponding dispersion curves correspond to the RMS misfits, with darker colours indicating better fits (Figure 4.6).

The shear-velocity tests for the mid-crustal LVZ of Songpan-Ganzi, 20–45 km depths, yield the lower and upper S velocity limits of 3.08 and 3.43 km s^{-1} , respectively, with a best fitting V_S of 3.18 km s^{-1} (Figure 4.6 and Table 4.1, Songpan-Ganzi). In this and following figures, V_S and radial anisotropy profiles are plotted in pale grey where the structure cannot be reliably resolved by the Rayleigh and Love data in the period range of the measurements.

In contrast to Songpan-Ganzi, the best-fitting V_S profile for West Lhasa shows no LVZ in isotropic-average S velocities but features, instead, strong mid-crustal radial anisotropy of $\sim 12\%$ (Figure 4.6, West Lhasa, red best fit). Estimates of the radial anisotropy within 20–45 km depth were derived from test inversions with anisotropy fixed in the background model ($V_{SH} > V_{SV}$) and with the adjacent layers allowed to have between 0 and $+3\%$ anisotropy. Mid-crustal radial anisotropy in West Lhasa has a lower and upper limit between 8.5 and 13.8%. Crustal-average radial anisotropy estimates (depth span between 10 km depth and the Moho) give a range of 5.5 to 8.6% (Figure 4.6, West Lhasa).

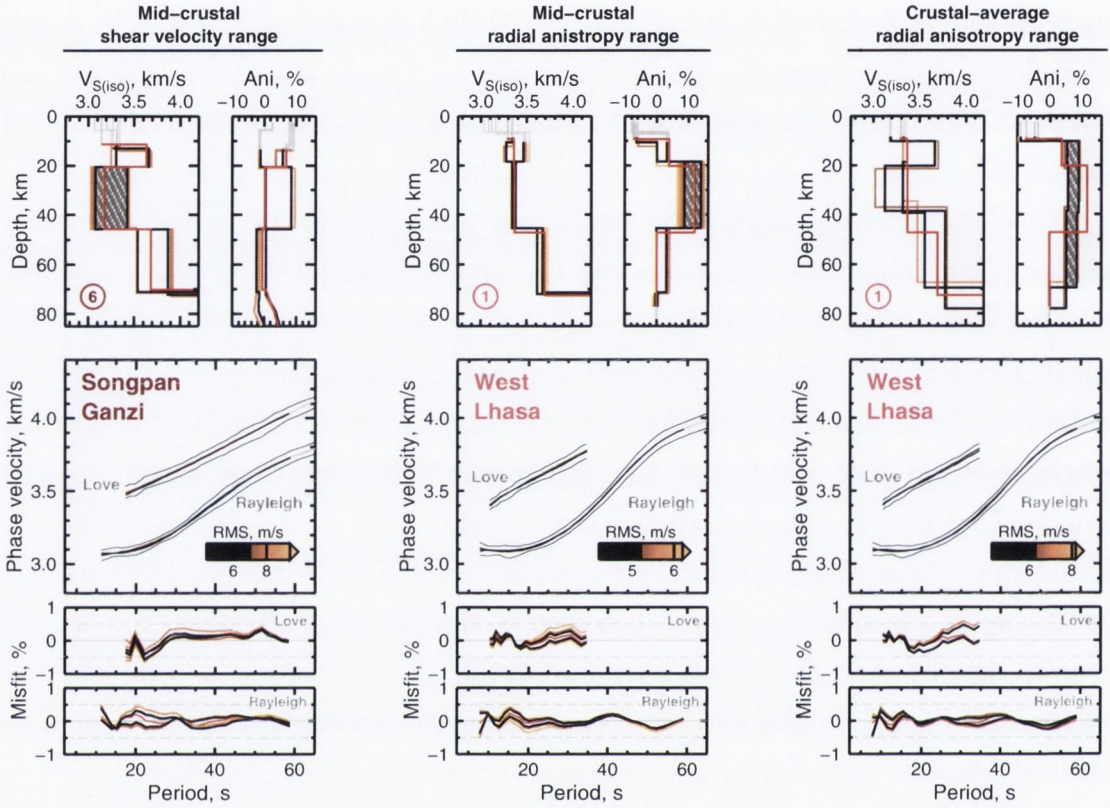


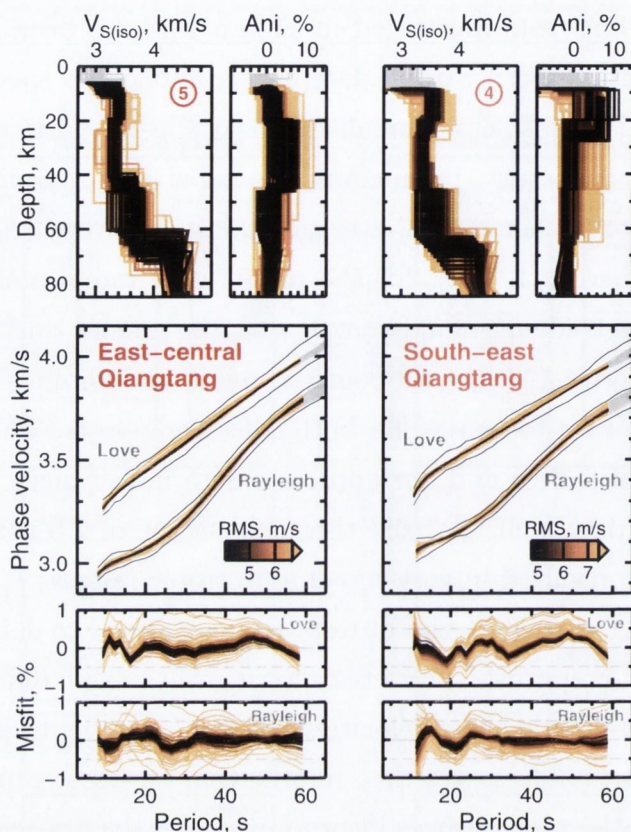
Figure 4.6: Grid-search test inversions performed to delimit the ranges of V_S and radial anisotropy in 20–45 km depth range, and the range for crustal-average radial anisotropy, between 10 km depth and the Moho. Next to each of the isotropic average $V_{S(iso)}$ models ($(2V_{SV} + V_{SH})/3$) is the radial anisotropy profiles $((V_{SH} - V_{SV})/V_{S(iso)})$. Grey shade in the models show the range of the tested values that turned out to be consistent with the data. Synthetic dispersion curves for each of the V_S models is plotted in the frame below. The colour scale indicates the data-synthetic RMS misfit over the combined Rayleigh- and Love-wave dispersion curves, for periods up to 60 s. Black curves represent the accepted data-synthetic fits which have a RMS value between the best fitting regional profile (red) and a set RMS threshold. Thin black curves show the standard deviation of the data. The profiles are coloured in depth ranges where they are reliably constrained by the data. The misfit frames show the data-synthetic misfits at different periods for each V_S model. Left: The V_S range at 20–45 km depths beneath Songpan-Ganzi. Centre: The range for radial anisotropy at 20–45 km depths beneath West Lhasa. Right: Crustal-average radial anisotropy, between 10 km depth and the Moho, West Lhasa.

The tests illustrated in Figure 4.6 have been performed for all of our regions within Tibet. Strong damping is applied to specific crustal layers and discontinuities that show unrealistic shear speeds, particularly when running tests for an isotropic crust. In an effort to narrow down parameter ranges, the RMS threshold factor for the crustal-average radial anisotropy tests is set to 2 m s^{-1} and reduced further to 1 m s^{-1} for the mid-crustal radial anisotropy ranges. Because of the large trade-off of anisotropy with the crustal thickness, regions which have a thinner crust (North and South Yunnan, and Qinling-Qilian) have the RMS threshold factor set to 1 m s^{-1} for both anisotropic tests. The selection of the RMS threshold is subjective and our approach is to have ranges from reasonable models that fit the data well. A fixed threshold factor of 2.5 m s^{-1} on all the three tests would have resulted in very broad anisotropic ranges.

A separate series of tests was performed to determine shear-velocity ranges for the lower crust. These tests were unsuccessful, due to the strong trade-offs between deep-crustal shear velocities and the Moho depth and uppermost mantle velocities. To try and resolve this, information on the regional Moho depth and uppermost mantle velocities was incorporated into the inversions. The lower crustal velocities were kept below 4.0 km s^{-1} and the Moho depths were not allowed to be shallower than published depths: 70 km for Lhasa (e.g., *Priestley et al.*, 2008; *Schulte-Pelkum et al.*, 2005; *Nábelek et al.*, 2009), 60 km for Qiangtang and Songpan-Ganzi (e.g., *Zhao et al.*, 2001; *Kind et al.*, 2002; *Kumar et al.*, 2006; *Nábelek et al.*, 2009; *Tseng et al.*, 2009; *Yue et al.*, 2012), 50 km in Qinling-Qilian (e.g., *Liu et al.*, 2006; *Shi et al.*, 2009; *Karplus et al.*, 2011), and 50 km and 40 km in North and South Yunnan respectively (e.g., *Kan et al.*, 1986; *Li et al.*, 2006; *Xu et al.*, 2007). Nonetheless, equally well fitting models could be obtained for lower-crustal V_S values in ranges too broad to be useful.

Plotted together, the results of all the inversions for each region display both the uncertainty of the shear-speed and radial-anisotropy profiles, and the robust features within them. Figure 4.7 shows examples of the bundles of isotropic V_S and anisotropy profiles, as well as the corresponding dispersion curves and frequency-dependent misfits, all colour-coded according to misfit. The well fitting (dark coloured) $V_{S(\text{iso})}$ models for East-central Qiangtang have mid-crustal shear speeds between 3.2 and 3.5 km s^{-1} , with 0–8% radial anisotropy. The synthetic dispersion

Figure 4.7: Bundles of radially anisotropic S -velocity profiles compiled from all the test inversions for East-central Qiangtang and South-east Qiangtang. For each profile, the corresponding synthetic Rayleigh- and Love-wave dispersion curves and the data-synthetic misfits are plotted in the frames below, as described in Figure 4.6. The colour of each profile reflects the misfit. The profiles are coloured in depth ranges where they are reliably constrained by the data. All the profiles fit the data within the standard deviation (thin black lines) at all periods.



curves and data-synthetic misfits show how narrow the range of best-fitting phase-velocity curves is. South-east Qiangtang displays mid-crustal velocities similar to East-central Qiangtang, but with shallower radial anisotropy. The broad ranges of lower-crustal shear speeds and the Moho depths are due to the Moho- V_S trade-off.

4.4 Results

We present suites of 1-D radial anisotropic shear-velocity models for nine regions across the Tibetan Plateau, displayed in Figure 4.8. For each region we show the isotropic shear velocity profiles as a function of depth from all the test inversions, and the corresponding radial anisotropy. The profiles are colour-coded according to their fit, dark being the better fitting models. The white loops on the map enclose the different regions represented by the profiles, approximated by the inter-station pairs (Figure 4.3).

The plateau is characterised by low mid-crustal shear velocities ($\leq 3.5 \text{ km s}^{-1}$),

across vast areas from west to south east Tibet, even though in Yunnan the crust thins significantly. Internally within the plateau, strong velocity contrast exists between different regions; mid-crustal shear velocities decrease towards the north of the plateau along West Lhasa to West Qiangtang, from 3.37 to 3.14 km s^{-1} ; along Central Lhasa to Songpan-Ganzi, from 3.44 to 3.18 km s^{-1} ; and from South to North Yunnan with velocities decreasing from 3.50 to 3.34 km s^{-1} (Figure 4.8 and Table 4.1). Although the suite of better fitting models for the regions north of the BNS have a mid-crustal LVZ, only West Qiangtang, Songpan-Ganzi and South-east Qiangtang show robust patterns of a decrease in shear velocity between 10 and 30 km depth (Figure 4.8, regions 2, 4 and 6). The contrast between the monotonic velocity of Central Lhasa and the LVZ across Songpan-Ganzi (regions 3 and 6) is also highlighted by *Yang et al.* (2012). In the case of West Lhasa, which also has no isotropic LVZ in the best fitting profile (Figures 4.6 and 4.8, region 1), the strong radial anisotropy in the same depth range results in a large amplitude V_{SV} LVZ — as observed in previous studies (e.g., *Rapine et al.*, 2003). The crustal structure and shear velocities of North Yunnan are similar to the regional profiles of central Tibet, including a small amplitude mid-crustal LVZ, despite this region being a considerable distance away from the central regions and having a shallower Moho.

At the northern and southeastern margins of the plateau the crustal structure is different to the profiles found within Tibet. In the north, the Kunlun Fault appears to be a sharp boundary between the low velocities of Songpan-Ganzi and the high velocities of Qinling-Qilian, as V_S increases to about 3.6 km s^{-1} , eliminating the LVZ (region 6 and 9). Similarly across Yunnan, as anticipated from the very similar dispersion curves to Qinling-Qilian (Figure 4.3, regions 8 and 9), the mid-crustal S velocity increase diminishes the LVZ amplitude.

Radial anisotropy differs between regions. In western Tibet the better fitting models for West Lhasa and West Qiangtang have the strongest mid-crustal radial anisotropy with minimums of 9 and 5%, respectively (Figure 4.8, Table 4.1, regions 1 and 2). Radial anisotropy at similar depths but with a lesser amplitude is observed in Central Lhasa, East-central Qiangtang, North and South Yunnan, and Qinling-Qilian. The dispersion curves for Songpan-Ganzi and South-east Qiangtang require weak-to-no radial anisotropy in the middle and lower crust, and

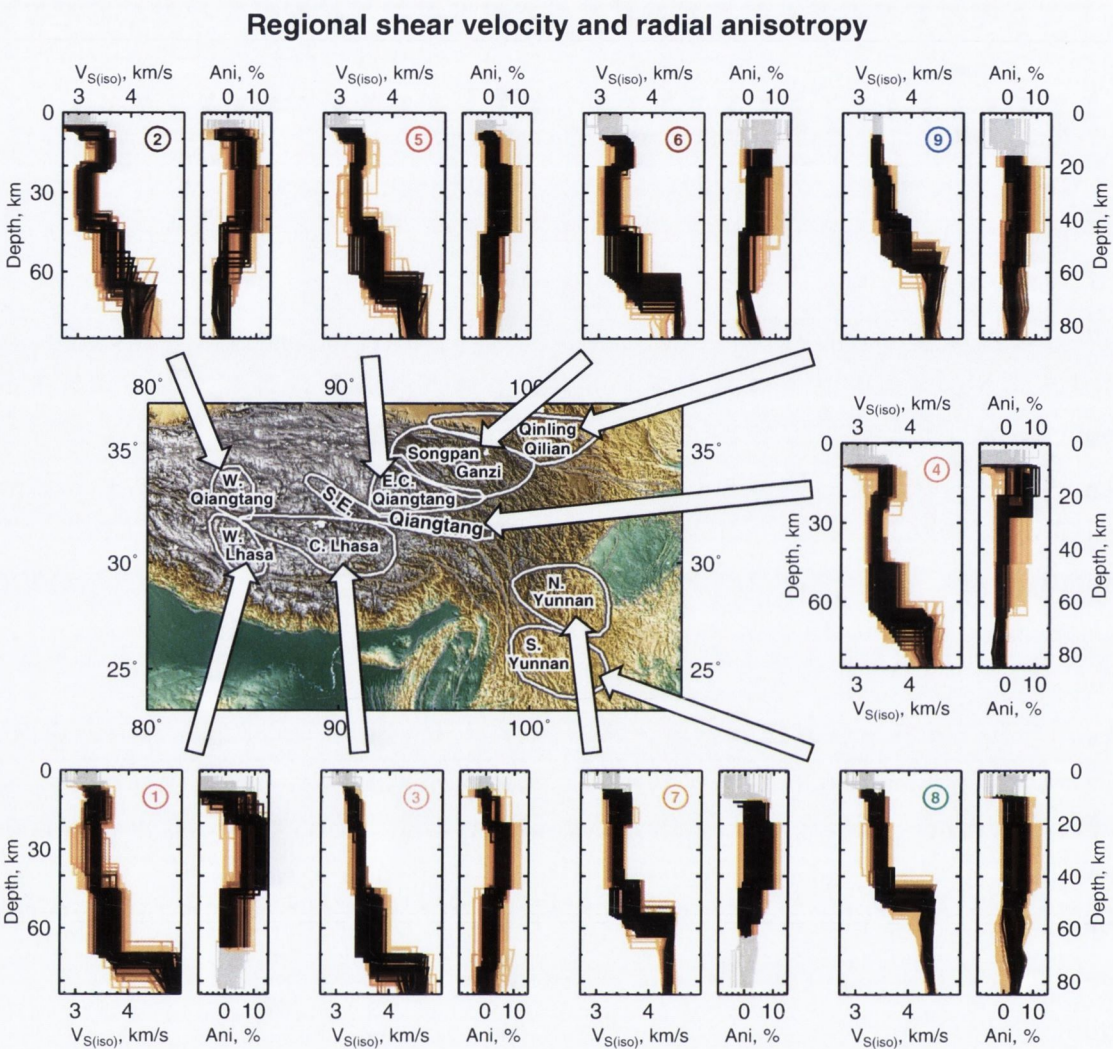


Figure 4.8: Bundles of radially anisotropic, *S*-velocity profiles compiled from all the test inversions. The models represent average structure over regions shown on the map. Each individual profile is colour coded to show data-synthetic misfit; darker colours indicate better fit. The profiles are coloured in depth ranges where they are reliably constrained by the data. All the profiles fit the data within the standard deviation at all periods.

anisotropy seems to occur shallower (Figures 4.6, 4.7 and 4.8, Table 4.1, regions 4 and 6). The contrast in radial anisotropy between West Lhasa and Songpan-Ganzi can be observed directly from the phase velocities (Figure 4.2 u and x). The Rayleigh- and Love-wave dispersion curves for West Lhasa, when compared to Songpan-Ganzi, diverge between 20 and 30 s period, which is sensitive to the

middle crust. Similarly, radial anisotropy across South Yunnan is evident due to similar divergence of Love-wave dispersion from Rayleigh-wave, when compared with the dispersion of Qinling-Qilian (Figure 4.3, regions 8 and 9). The divergence takes place for periods longer than 30 seconds, which suggest that radial anisotropy occurs at greater depth.

Although no specific tests were performed to directly resolve the crustal thickness, and keeping in mind the strong trade-off of lower crustal shear velocities with Moho depth, variations in the regional crustal thickness from the better fitting profiles is observed. In southern Tibet the crustal thickness is between 70–80 km (regions 1 and 3), between 60–70 km in northern Tibet (regions 2, 4–6), between 50–60 km thick in Qinling-Qilian (region 9), and thins gradually towards the south-east to about 40–50 km depth (regions 7 and 8). The Moho depths of all the nine regions in general agree with the average depths inferred from other seismic studies: a deep Moho beneath the southern Tibet (e.g., *Priestley et al.*, 2008; *Schulte-Pelkum et al.*, 2005; *Nábelek et al.*, 2009) and a shallower Moho across northern Tibet (e.g., *Zhao et al.*, 2001; *Kind et al.*, 2002; *Kumar et al.*, 2006; *Nábelek et al.*, 2009; *Tseng et al.*, 2009; *Yue et al.*, 2012), a sudden jump across the Kunlun Fault (*Liu et al.*, 2006; *Shi et al.*, 2009; *Karplus et al.*, 2011), and a gradually rising Moho towards southeast Tibet (*Kan et al.*, 1986; *Li et al.*, 2006; *Xu et al.*, 2007).

4.5 Discussion

4.5.1 The shear velocities within Tibet's crust

Many shear-velocity models inferred from surface-wave measurements are based on Rayleigh-wave studies only. Modelled shear speeds from Rayleigh waves in regions that have significant radial anisotropy are only representative of the vertically polarised shear waves V_{SV} . Our 1-D radial anisotropic models fit both wave types, and hence the isotropic models are expected to represent a more accurate shear-velocity structure.

In order to benchmark our models, we compare the best-fitting mid-crustal velocities estimates (V_{SV}) with those of other 1-D models also inferred from surface waves. For example in Songpan-Ganzi, which has weak radial anisotropy, the mid-

	Region	Middle crust V_S (km s^{-1}) 20–45	Middle crust radial anisotropy (%) 20–45	Crustal average radial anisotropy (%) 10–M
Depth (km)				
West Lhasa	1	3.37 3.21–3.41	9–14	6–9
West Qiangtang	2	3.14 3.02–3.36	5–9	5–8
Central Lhasa	3	3.44 3.31–3.49	4–8	3–5
S.E. Qiangtang	4	3.43 3.27–3.49	0–1	0–2
E.C. Qiangtang	5	3.33 3.25–3.44	0–8	2–4
Songpan Ganzi	6	3.18 3.08–3.43	0–5	0–5
North Yunnan	7	3.34 3.23–3.52	0–10	3–6
South Yunnan	8	3.50 3.32–3.61	0–9	4–7
Qinling Qilian	9	3.59 3.38–3.63	1–8	2–5

Table 4.1: Shear velocity and radial anisotropy estimates for different regions within Tibet. Mid-crustal S velocities and radial anisotropy are from the best fitting profiles. Velocity and radial anisotropy ranges are from test inversions as illustrated in Figure 4.6.

crustal V_{SV} is 3.18 km s^{-1} — a close match with the 1-D model of *Yang et al.* (2012). Similarly, Central Lhasa has a mid-crustal V_{SV} of 3.39 km s^{-1} , which is also in good agreement with the average value of *Yang et al.* (2012), however in our case the mid-crust has a radial anisotropy of 4–8%. Another good agreement in S velocity is with that of *Rapine et al.* (2003) for West Lhasa, a region with even stronger radial anisotropy: our mid-crust V_{SV} is 3.23 km s^{-1} . Across Qiangtang, *Rapine et al.* (2003) report an increase in V_{SV} whereas our profiles show a drop in S velocity. The increase in V_{SV} could be due to parts of the source-station paths sampling the strong Tarim basin and may also be due to azimuthal anisotropy biases favouring fast propagation in the west-east orientation across northern Tibet (Chapter 5). 1-D radial anisotropic models from paths crossing the entire plateau along the northern and southern regions have mid-crustal V_{SV} in the range of $3.15\text{--}3.3 \text{ km s}^{-1}$ (*Duret et al.*, 2010), within our range of values for the different regions.

Likewise, consistency exists between our 1-D mid-crustal V_{SV} with the velocities of recent tomographic models using hundreds of paths crossing the entire plateau (*Acton et al.*, 2010; *Yang et al.*, 2012), and across Yunnan (*Yao et al.*,

2008, 2010). Isotropic V_S estimates for Yunnan are within average velocities of the radial anisotropic tomographic model by *Huang et al.* (2010). Other tomographic maps either show weak lateral changes in the mid crust (*Sun et al.*, 2010) or show very low velocities located in the eastern region of the plateau (*Li et al.*, 2012), resulting in differences to our and other published models.

The very good agreement of our vertically polarised shear velocities with V_{SV} of other models demonstrate that our inversions are robust, and can infer shear speeds requiring varying strengths of radial anisotropy. This gives us confidence that the V_S estimates for the different regions across the plateau represent the true isotropic shear velocity. Differences between V_{SV} and $V_{S(iso)}$ in regions that have strong radial anisotropy can be significant. For example in West Lhasa the mid-crustal isotropic velocity is faster than V_{SV} by about 4% — this is significant enough to distort inferences on the geophysical state and composition of rocks if V_{SV} is assumed to be the true shear speed.

Figure 4.9 (cross-sections AA', BB', and CC') compares mid-crustal shear speeds of various tomographic models (*Yao et al.*, 2008, 2010; *Huang et al.*, 2010; *Acton et al.*, 2010; *Yang et al.*, 2012) with our estimates along three north-south profiles across Tibet. Although the different shear speeds have similar trends, the isotropic V_S is notably faster beneath regions with strong mid-crustal radial anisotropy such as West Qiangtang and Lhasa. Interestingly, West Qiangtang and Songpan-Ganzi have similar very low velocities, but, the former has strong radial anisotropy ($V_{SH} > V_{SV}$, 5% minimum) and the latter does not. Profiles AA' and BB' clearly show strong velocity variations between the different terranes. Along profile BB' radial anisotropy decreases from south to north as reflected in the converging shear velocities towards Songpan-Ganzi. The regional inter-station pairs for Central Lhasa miss a low V_S anomaly in southern Tibet (*Acton et al.*, 2010; *Yang et al.*, 2012) probably originating from north of the Himalayas and across the Yarlung-Zangbo Suture (*Unsworth et al.*, 2005; *Rippe and Unsworth*, 2010). In the north all three velocities come together beneath Songpan-Ganzi, and further north, across the Kunlun Fault, our V_S highlights a dramatic velocity increase. Shear velocities inferred from controlled-source experiments across south and north of the Kunlun Fault are significantly higher and are not plotted (*Mechie et al.*, 2012). Along profile CC' tomography suggests that there is strong local heterogeneity

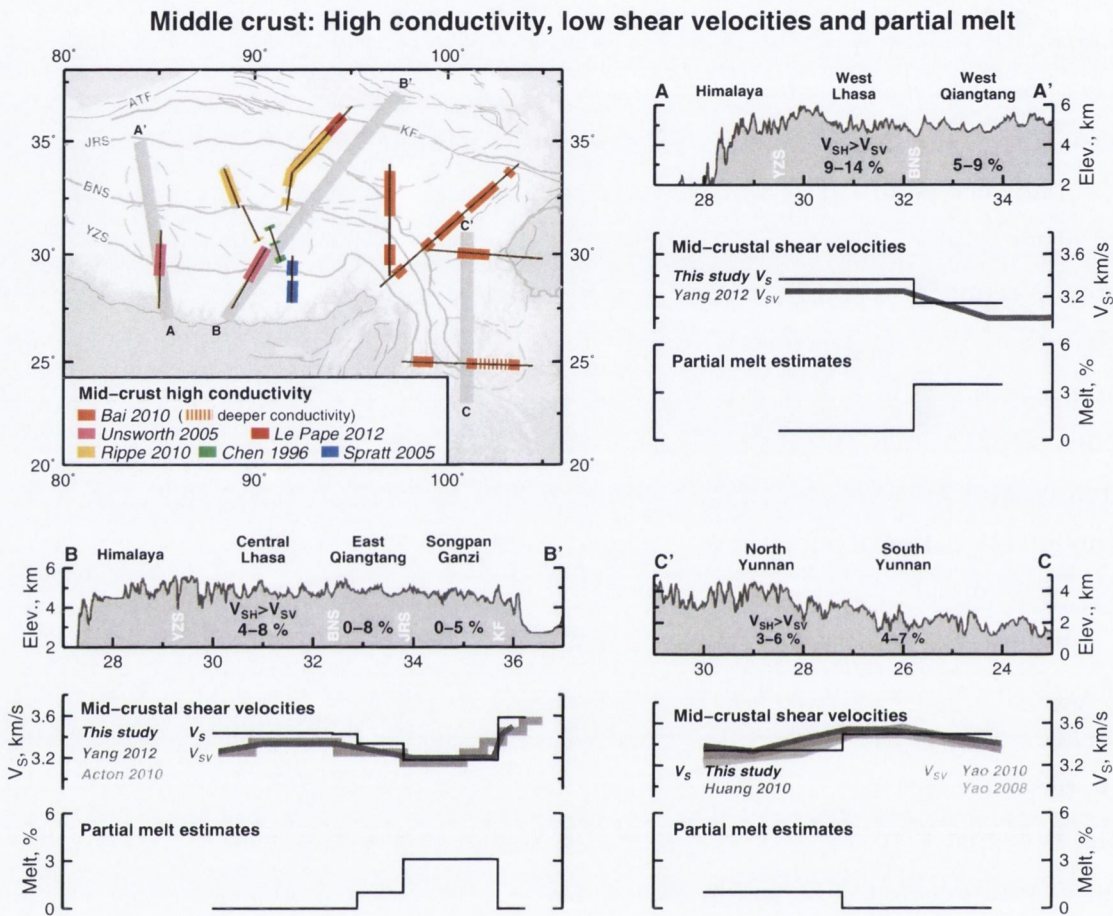


Figure 4.9: Mid-crustal high conductivity, shear velocities and partial melt across the Tibetan Plateau. Map shows the location of different MT experiments (dark brown thin lines) and the location of three vertical cross sections (thick grey lines). Highlights along the MT lines indicate regions of mid-crustal high conductivity from various studies (*Chen et al.*, 1996; *Spratt et al.*, 2005; *Unsworth et al.*, 2005; *Rippe and Unsworth*, 2010; *Bai et al.*, 2010; *Le Pape et al.*, 2012). For each cross section the elevation is shown along with mid-crustal average S velocities and partial melt estimates (20–45 km depth range). Our best-fitting shear velocities (V_S , bold thin black lines) are shown along with other estimates of shear velocities (thick grey lines, V_{SV} : *Yao et al.*, 2008, 2010; *Acton et al.*, 2010; *Yang et al.*, 2012, and V_S : *Huang et al.*, 2010). The differences between velocities can be explained, in part, by mid-crustal radial anisotropy, indicated, as measured in this study, in the elevation frame. Mid-crustal partial melt estimates are plotted in the bottom frames.

(Yao *et al.*, 2008, 2010; Huang *et al.*, 2010). Our profiles average out local shear-velocity anomalies and other V_S changes arising from an alternating positive and negative radial anisotropy (Huang *et al.*, 2010).

The shear-velocity models represent the robust structures beneath the regions, and provide a more accurate estimate of the crustal S velocities appropriate to assess the heterogeneity beneath Tibet. Mid-crustal velocities vary across the plateau decreasing from $\sim 3.4 \text{ km s}^{-1}$ in the south to 3.14 and 3.18 km s^{-1} in the north beneath West Qiangtang and Songpan-Ganzi, respectively. Similarly, in southeastern Tibet, S velocities decrease from South to North Yunnan from 3.50 to 3.34 km s^{-1} , respectively (Figure 4.11a).

4.5.2 Regional distribution of low-velocity zones

Recent geophysical models have kept alive the debate regarding Tibet's internal structure and deformation, between the two end-member modes: "rigid blocks" and "continuous deformation". Topography-induced models for Tibet by Clark and Royden (2000) indicate that the morphology of the eastern plateau reflects large-scale fluid flow within the underlying crust that oozes around strong rigid boundaries, creating broad, gentle margins. On the other hand, MT studies cutting across the various terranes of eastern Tibet proposed two localised channels of high conductivity that can act as shear zones and/or accommodate flow (Bai *et al.*, 2010).

Inferences from seismic shear velocities have been few. A mid-crustal LVZ, indicative of a vertically confined weak layer, has been determined in the south (e.g., Kind *et al.*, 1996; Cotte *et al.*, 1999; Caldwell *et al.*, 2009), and across the plateau (e.g., Romanowicz, 1982; Rapine *et al.*, 2003), too sparse to discriminate between any mechanisms. Only recently, with the rapid increase in the number of stations, it is becoming possible to map the entire plateau. South-north vertical cross sections from tomographic models of Tibet have two mid-crustal low-velocity anomalies, one on each side of the BNS, with the northern anomaly having the lowest V_{SV} (Acton *et al.*, 2010; Yang *et al.*, 2012). However, the lack of Love-wave information in these models makes it unclear if the decrease in V_{SV} reflects the true isotropic velocity or whether the low velocities are a result of strong radial anisotropy. 1-D models that fit both Love- and Rayleigh-wave dispersion and have

strong radial anisotropy between 20 and 45 km depth, do not have an explicit mid-crustal LVZ (*Shapiro et al.*, 2004; *Duret et al.*, 2010). Albeit all the available models, it is still not clear if the LVZ is plateau-wide or if the LVZ is isolated beneath specific regions.

The trend of better fitting models within the compilation of regional profiles highlights robust V_S structures, including a LVZ (Figures 4.7 and 4.8). The good agreement of the best fitting models with the velocities of other models, discussed in the previous section, adds trust to subsequent inferences. Estimates for the LVZ amplitude from the V_S decrease between 15 and 30 km depth, $((V_{S(15)} - V_{S(30)})/V_{S(15)})$, indicate that LVZ amplitudes vary substantially across the plateau: strongly pronounced in West Qiangtang and Songpan-Ganzi ($>9\%$ V_S reduction), weaker in central and eastern Qiangtang ($<3\%$), and unnecessary across Lhasa. The western profiles have a similar V_{SV} reduction to the 1-D profiles of north-west Himalaya (*Caldwell et al.*, 2009). The crustal structure and shear velocities of North Yunnan are similar to the regional profiles of central Tibet, including the mid-crustal LVZ. The increase in V_S from North to South Yunnan diminishes the amplitude of the LVZ in the south (Figure 4.8, regions 7 and 8) resulting in a similar structure to Qinling-Qilian, as anticipated from the alike dispersion curves of both regions (Figure 4.3). The profiles for North Yunnan agree well with 1-D models from receiver functions (*Xu et al.*, 2007) but contrast with the tomographic models that either lack a distinctive LVZ or have very low shear speeds just above the Moho (*Yao et al.*, 2008, 2010; *Huang et al.*, 2010). Unlike southeast Tibet where the velocity seems to increase gradually, the firm V_S increase beneath Qinling-Qilian from Songpan-Ganzi suggest that the Kunlun Fault is a sharp boundary for the LVZ in north Tibet.

The regional distribution of our LVZ agrees qualitatively with the tomographic map of a LVZ by *Yang et al.* (2012). They show large LVZ amplitudes ($>9\%$, V_{SV}) around the periphery of Tibet, particularly in western Qiangtang and Songpan-Ganzi, regions which we also identify with strong LVZ amplitudes (Figure 4.8, regions 2 and 6; Figure 4.11b, red shaded regions). In south and central Tibet, the model of *Yang et al.* (2012) has an interconnected LVZ in-between large areas without an LVZ, also similar to regions we identify with none or a weak LVZ (Lhasa and central Qiangtang). We note that any existence of lateral channels or local

discontinuous LVZ (*Hetényi et al.*, 2011) are smoothed in our profiles. Nonetheless, similar dispersion of the individual inter-station pairs within the same region, such as Songpan-Ganzi (Figure 4.2a, i and m), indicate that the crust is uniform within that region. Hence, the low shear velocities and other inferred properties such as melts, viscosity and flow (discussed below) are the same beneath the inter-station path coverage of the region. A homogeneous crust beneath northern Tibet has also been modelled by tomography (*Yang et al.*, 2012). We conclude that on a regional scale a mid-crustal low shear-velocity zone varies across the plateau in a more systematic way from south to north, with the decrease in velocities towards the north, up to the Kunlun Fault. Northeastern Tibet, Songpan-Ganzi, is characterised by a laterally broad LVZ that is prone to widespread deformation.

4.5.3 Low shear velocities due to high temperatures

With little knowledge about the temperature, composition and water content beneath Tibet we have to rely on shear-velocities, seismic observations, and laboratory experiments to infer other geophysical properties. Precise temperature-depth estimates for Tibet are determined from the α to β quartz transition in quartz bearing rocks. The transition takes place at the boundary between the upper and middle crust, generating a measurable seismic signature. Its detection yields precise temperature estimates and marks specific temperature-depth contours. Beneath north-central Lhasa the transition occurs at 32 km depth (0.85 GPa) and has a corresponding temperature of 800°C (*Mechie et al.*, 2004). In order to estimate the expected shear velocity at this depth and temperature we make use of laboratory measurements and relationships — an approach also adopted by *Caldwell et al.* (2009) and *Yang et al.* (2012). Likewise we use the laboratory measurement of *Christensen* (1996) performed on a variety of dry rocks at room temperature (r.t.) and at 1000 MPa (representative of pressures at 30 km depth). We select metamorphic rocks with the lowest shear velocities: metagraywacke, phyllite, granite gneiss, biotite (tonalite) gneiss, mica quartz schist, and paraganulite — all in the range of 3.51–3.66 km s⁻¹, with an average of 3.62 km s⁻¹. Velocity reductions due to high temperatures are estimated by applying the V_S -temperature relationship of 2×10^{-4} km s⁻¹ per degree Celsius determined from temperature-pressure experiments on a variety of gneisses (*Kern et al.*, 2001). Thus at 800°C the expected

V_S is 3.46 km s^{-1} , close to the V_S in Central Lhasa (3.44 km s^{-1}) and South-east Qiangtang (3.43 km s^{-1}). Plutonic rocks such as granite-granodiorite have a high V_S which at this depth and temperature would not match the V_S for Central Lhasa.

Across Qiangtang (along the INDEPTH III profile) *Mechie et al.* (2004) determines that the $\alpha - \beta$ transition occurs shallower at 18 km depth (0.48 GPa) with a corresponding temperature of $\sim 700^\circ\text{C}$. Such a transition is also thought to occur north of Songpan-Ganzi, at a similar depth, along the INDEPTH IV profile (*Mechie et al.*, 2012). Other temperature estimates from nearby xenoliths indicate that temperatures reach 1000°C at a depth of 50 km (1.3 GPa) (*Hacker et al.*, 2000). At these temperatures the calculated reduced shear velocity for hot solid rocks, including pelitic rocks (*Hacker et al.*, 2000), is still above 3.4 km s^{-1} , much higher than V_S across west and central Qiangtang.

Laboratory experiments show that following the $\alpha - \beta$ transition, further increase in temperature continues to decrease S velocities at a higher rate, followed by rapid decrease at the onset of fluid induced partial melting (e.g., *Mueller and Massonne*, 2001). The dehydration-melting solidus for muscovite bearing schist is in the range of $820\text{--}900^\circ\text{C}$ between 1.0–1.6 GPa (e.g., *Clemens and Vielzeuf*, 1987; *Patiño Douce and Harris*, 1998; *Patiño Douce and McCarthy*, 1998; *Holyoke and Rushmer*, 2002; *Litvinovsky et al.*, 2000). Assuming such a composition and high temperatures in the range of $900\text{--}1000^\circ\text{C}$, partial melting starts to occur and the shear velocities are reduced substantially. Our results indicate that mid-crustal temperatures of about $\sim 800^\circ\text{C}$ can explain the low shear velocities in Central Lhasa, South-east Qiangtang and South Yunnan whereas in regions with lower mid-crustal velocities cannot be explained by higher temperatures alone and require some other velocity-reducing process such as partial melting.

4.5.4 Lower shear velocities due to partial melt

A partially molten middle crust beneath southern Tibet has been hypothesised from various geophysical studies (*Nelson et al.*, 1996). The presence of fluids have been inferred from strong seismic reflections (bright spots) observed across the Himalaya at 15–20 km depth using the INDEPTH I and II data. The strong interfaces are thought to be produced by magmas or partial melting (*Brown et al.*, 1996) and also water (*Makovsky et al.*, 1996). *Makovsky and Klemperer* (1999)

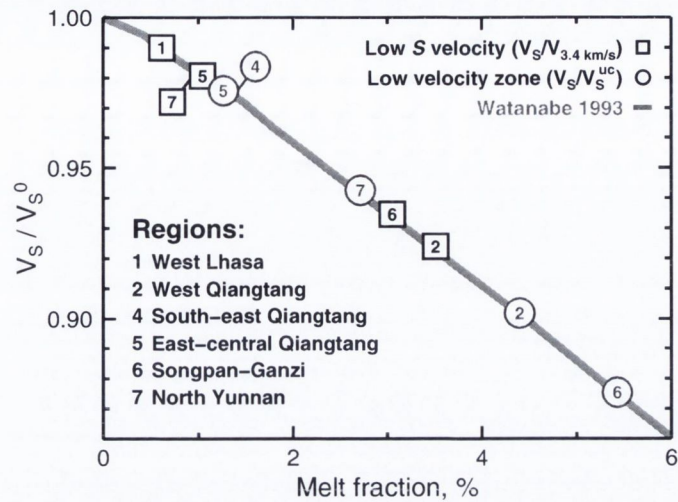


Figure 4.10: Mid-crustal partial melt estimates for different regions across the plateau. Estimates are based on the analytical shear-velocity relationship of V_S/V_S^0 and rock fluid fractions by *Watanabe* (1993) (light grey curve). V_S is the velocity within the mid-crustal (30 km depth), according to the best-fitting, region-average profile. Black squares: Partial melt estimates using a reference S velocity of crustal rocks on the onset of melting at 30 km depth, $V_S^0=3.4 \text{ km s}^{-1}$ (see text for details). Dark grey circles: Partial melt estimates based on the decrease in velocity from the layer above the LVZ.

suggest that 10% by volume of free aqueous fluids is required to produce the observed bright reflections. In contrast to southern Tibet, no strong shallow seismic reflections have been observed in north-central Lhasa and south Qiangtang (along the INDEPTH III profile). The lack of bright spots suggests the absence of fluid concentrations in these regions (*Haines et al.*, 2003). Estimates of melt are sparse and vary in numbers. It is still unclear whether melts are plateau-wide (*Nelson et al.*, 1996) or whether melts vary between south and north (*Owens and Zandt*, 1997) or are only localised beneath rifts like the Yadong-Gulu. Furthermore it is also uncertain how much percentage of melt is involved.

As the shear-velocities in western and northern Tibet are lower than that of hot solid rocks and temperatures are higher than the solidus, we consider the effect fluid-induced partial melting has on the S velocities. We derive estimates for the fluid volume in the rocks assuming that the lower shear velocities are due solely to the presence of either aqueous fluids or melts, or both, and not due to the (poorly known) composition. We determine two estimates for melt fraction, one

based on the assumption that the V_S^0 of dry crustal rocks equals the velocity just on top of the LVZ (similar to the approach adopted by *Caldwell et al.*, 2009), and the other estimate is based on an unconstrained reference V_S^0 of 3.4 km s^{-1} — a velocity well below the expected V_S of hot dry solid rocks at 30 km depth (a conservative approach also applied by *Yang et al.*, 2012). Mid-crust V_S/V_S^0 ratios are compared to an analytical relationship small fluid fractions of silicate melts and water have on seismic velocities (*Watanabe*, 1993, Figure 4.10). These relationships are similar to other V_S -melt studies involving organic earth-analogue experiments (*Takei*, 2000) and theoretical calculations for horizontally-aligned melts (*Taylor and Singh*, 2002), as also illustrated by *Caldwell et al.* (2009). Although we are unable to distinguish between the effects of melt or water, the estimates of partial melt are critical to infer important information on the rheology of rocks.

Our estimates for mid-crustal partial melt fractions are completely dependent on shear velocities and hence have a similar distribution to the low S velocities (Figure 4.9). Percentages of melt derived from best fitting profiles show that West Qiangtang and Songpan-Ganzi have melt fractions between 3 and 6%, whereas North Yunnan, eastern Qiangtang and West Lhasa have melts less than 3, 2 and 1%, respectively (Figure 4.10). Regions Central Lhasa, South Yunnan and Qinling-Qilian are unlikely to have any significant volume of melt. Although the two definitions we adapt for melt estimates are different, they have similar regional distribution and melt fraction patterns, and show a correlation between percentage melt and the LVZ amplitude (Figure 4.11b). The partial melt estimates for West Qiangtang and Songpan-Ganzi are within the range of inferred melt for northwest Himalaya from 1-D V_{SV} seismic models (2.5–7%, *Caldwell et al.*, 2009). Melt estimates for North Yunnan are within suggested amounts of less than 4% from the V_S comparison with laboratory experiments (*Xu et al.*, 2007).

4.5.4.1 Electrical resistivity

Magnetotelluric studies for Tibet's crust show a drop in electrical resistivity at about 20 km depth. The resistivity (or its inverse, the electrical conductivity) of the rocks can be influenced by high temperatures, melts, hydrous minerals or a combination of all. Mid-crustal conductivity is not the same across the entire plateau and variations takes place within the same terranes (Figure 4.9). Between

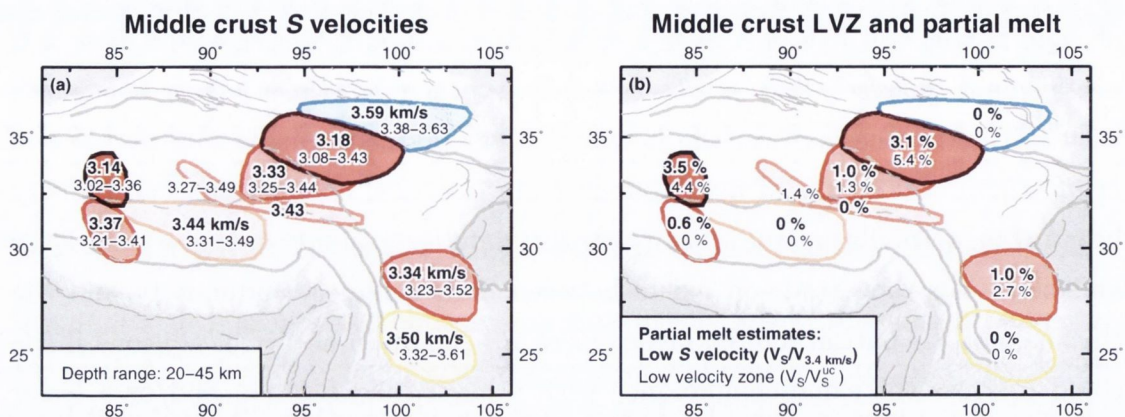


Figure 4.11: Maps of mid-crustal shear velocities, low-velocity zones and estimated partial melt. (a) Best fitting, isotropic-average S velocities (bold) and V_S ranges for the middle crust in the 20–45 km depth range (Table 4.1). (b) Estimates of mid-crustal partial melt based on the analytical shear-velocity relationship of V_S/V_S^0 and rock fluid fractions by Watanabe (1993) (Figure 4.10). Values in bold font: Partial-melt estimates from best fitting regional shear velocity and a reference S velocity of 3.4 km s^{-1} for hot solid rocks. Values in light font: Partial-melt estimates from the decrease in V_S from the layer above the LVZ. Red shaded regions indicate areas that have a mid-crustal LVZ.

20–45 km depth very high conductivity is observed beneath north of the Himalayas and southern Lhasa (Chen *et al.*, 1996; Unsworth *et al.*, 2005; Spratt *et al.*, 2005), sparsely across Central Lhasa (Chen *et al.*, 1996; Rippe and Unsworth, 2010), highly conductive beneath Qiangtang (Rippe and Unsworth, 2010), beneath Songpan-Ganzi up to the Kunlun Fault (Le Pape *et al.*, 2012; Bai *et al.*, 2010), and across southeast Tibet (Bai *et al.*, 2010). The distribution of conductivity agrees with the mid-crustal shear velocity not only in terms of depth, but also with the north-south pattern of decreasing V_S and increasing conductivity, from Central Lhasa towards northern Tibet up to the Kunlun Fault. East-central Qiangtang, Songpan-Ganzi and North Yunnan are characterised by strong conductivity, lower V_S , a LVZ, and significant partial melt. Likewise, regions with relatively high V_S , no LVZ, and no melt have intermittent or no conductivity such as West and Central Lhasa, South-east Qiangtang, Qinling-Qilian, and South Yunnan. The latter has deeper conductivity, below 40 km depth (Bai *et al.*, 2010).

Partial melting has long been thought to be responsible for the high conductivity in southern Tibet — across the Himalayas (Pham *et al.*, 1986). The coincidence

of conductivity with bright seismic spots added support for regionally interconnected fluid phases (*Chen et al.*, 1996; *Li et al.*, 2003; *Unsworth et al.*, 2005), with further synthesis of ductile flow of a partially molten middle crust (*Nelson et al.*, 1996). Estimates of fluid content from electrical conductivity can be very broad. Magnetotelluric studies can only estimate the conductance of a fluid layer and hence a strong trade-off exists between melt volume, conductivity and the thickness of the partial molten layer (*Li et al.*, 2003). In addition, aqueous fluids could produce the same conductance with a lower fluid fraction and/or a lesser layer thickness. *Li et al.* (2003) proposes that a layer of 10–15% aqueous fluid a few hundred metres thick overlying a thick layer of partial melt gives the most consistent explanation to both the MT and seismic data for the Himalayas and southern Lhasa. Melt estimates based on MT observations are between 2 to 4% melt for the Himalaya and 5–14% for southern Lhasa (*Unsworth et al.*, 2005), 2–6% for northern Tibet (*Unsworth et al.*, 2004), and 5–20% across southeast Tibet (*Bai et al.*, 2010). Melt estimates based on shear velocities are always at the lower end of the ranges based on MT. A joint analysis of shear velocities and conductivity for melt estimates remains desirable.

4.5.4.2 Distribution of melt across Tibet

Different geophysical observations consistently show north-south contrasts in the crustal properties across the plateau particularly distinguished by dry, wet or no melts. Wet melts across the Himalayas are inferred from bright seismic reflection (*Brown et al.*, 1996; *Makovsky et al.*, 1996) and high conductivity (e.g., *Pham et al.*, 1986; *Unsworth et al.*, 2005). Across Central Lhasa, melts are unlikely as deduced from lack of bright seismic spots (*Haines et al.*, 2003), high resistivity (INDEPTH 500 line, *Rippe and Unsworth*, 2010), and moderately low shear velocities (this study). Dry hot melts in northern Tibet are attributed to lack of bright seismic spots (*Haines et al.*, 2003), high temperature and absence of hydrous minerals (*Hacker et al.*, 2000), high conductivity (e.g., *Unsworth et al.*, 2004; *Le Pape et al.*, 2012), and low *S* velocities (this study, *Yang et al.*, 2012). Inferences of potentially less melt in Lhasa than northern Tibet is also noted in tomography (figure 12a in *Yang et al.*, 2012). In southernmost Tibet, wet melts can be a result of dehydration associated with eclogitization reaction in the lower crust (*Hetényi et al.*, 2007). In

the north, crustal radiogenic sources (*McKenzie and Priestley, 2008*) and a warm lithospheric mantle (e.g., *Barron and Priestley, 2009*, Section 3.3.2) are likely to be the cause for an elevated temperature, low V_S and melts.

4.5.5 Implications for large-scale, mid-crustal channel flow

Despite the increasingly corroborated conclusions of a weak crustal layer (see previous section), evidence of large-scale viscous flow will remain hard to attain. Southward flow beneath southern Tibet has been derived from thermal-mechanical numerical models (*Beaumont et al., 2001*) based on geophysical observation of melt-weakened rocks (*Nelson et al., 1996*). The model has been able to explain high-grade metamorphic rocks exhumed at the Greater Himalayan Sequence (*Jamieson et al., 2004*). Northward flow is likely to encounter resistance due to the structurally stronger Tarim and Qaidam Basin, but still, injection of Tibetan crust beneath Qaidam has been imaged from seismic (*Karplus et al., 2011*) and MT observations (*Le Pape et al., 2012*). The penetration of Tibetan crust into the Asian lithosphere may be responsible for ongoing thickening in the north (Figures 4.12 a and b). On a larger scale, plateau-wide deep crustal flow has been deduced to explain the regional topography and crustal thickness (*Zhao and Morgan, 1987*; *Royden, 1997*; *Clark and Royden, 2000*; *DeCelles et al., 2002*; *Royden et al., 2008*).

Laboratory experiments by *Rosenberg and Handy (2005)* show that as little as 5% melt can drop viscosity by an order of magnitude — this is enough to trigger channel flow beneath the Tibetan Plateau as inferred from numerical models (e.g., *Beaumont et al., 2001*). Evidence for a viscosity drop has been gathered from partial melt estimates based on MT studies (*Unsworth et al., 2005*) and seismic velocities (*Caldwell et al., 2009*). Since our partial-melt estimates are based on regional-average S velocities, we prefer to look at the spatial patterns that would be susceptible to a low or high viscosity. West Qiangtang and Songpan-Ganzi, which are likely candidates of high volume fraction of melt, are the most prone to have a low viscosity whereas Lhasa probably has a higher viscosity. Viscosity is expected to increase in south-east Tibet, across North to South Yunnan to correspond with our V_S increase. Our inferences for regional low and high viscosity are comparable to the compiled estimates of *Rippe and Unsworth (2010)* that show high viscosity for northern Lhasa and low viscosities for the southernmost, northern, and eastern

regions of Tibet.

One of the best ways to detect active or frozen deformation, such as flow, is from coherent patterns of seismic anisotropy, radial and azimuthal. Deep crustal strain induced by flattening and extension (flow) can cause anisotropic mineral crystals such as micas (*Weiss et al.*, 1999; *Nishizawa and Yoshino*, 2001; *Meltzer and Christensen*, 2001) and amphiboles (*Barruol and Kern*, 1996; *Tatham et al.*, 2008) to realign. Azimuthal anisotropy from fast shear-wave propagation in a preferred direction has been detected seismically across Tibet; short-period Rayleigh-wave group-velocity show a west-east fast propagation orientation in central Tibet (*Huang et al.*, 2004; *Su et al.*, 2008). The pattern of azimuthal anisotropy, indicative of the flow direction, rotates clockwise north of the eastern Himalayan syntaxis and align in the direction of southeast Tibet. On the other hand, radial anisotropy from shear-speed differences between horizontal and vertical shear-wave propagation, gives an indication of vertical strain. Radial anisotropy by *Shapiro et al.* (2004) show strong mid-crustal radial anisotropy in the middle-to-lower crust across northwestern Tibet ($V_{SH} > V_{SV}$ by 10%) and no radial anisotropy across east Songpan Ganzi. They compare the anisotropy distribution with the moment tensor of earthquakes and argue in favour of anisotropy caused by crustal thinning.

We establish ranges for the crustal radial anisotropy from the 1-D shear velocity profiles. Mid-crustal radial anisotropy is strongest in West Lhasa, West Qiangtang and moderately in Central Lhasa. Specific inversions testing for crustal-average radial anisotropy, between 10 km depth and the Moho, give tighter values (Figure 4.12c, Table 4.1). The spatial pattern of the regional radial anisotropy is similar to that inferred from the tomography of *Shapiro et al.* (2004), in particular the lack of anisotropy across Songpan-Ganzi. The 1-D models from *Duret et al.* (2010) are likely to be an average of very strong radial anisotropy in western Tibet and weaker anisotropy in eastern Tibet.

We reiterate that the very similar dispersion curves of the individual interstation pairs within a region (e.g., Songpan-Ganzi, see earlier discussion) indicates a homogeneous crust. Northeastern Tibet is characterised by a strong LVZ, high mid-crustal conductivity, low viscosity (*Rippe and Unsworth*, 2010), and weak radial anisotropy. Anisotropic crystals can become near-horizontally oriented as a result of ongoing crustal thinning, such as in western Tibet; likewise, horizontal pervasive

flow may align micas in the vertical plane resulting in a weaker radial anisotropy similar to what is observed in Songpan-Ganzi. Hence our results favour models of broad internal deformation which is also corroborated with the low topographic gradients and smooth topography across eastern Tibet (*Clark and Royden, 2000*).

4.5.5.1 Crust-mantle decoupling

A weak intra-crustal layer can result in decoupling upper crustal deformation from the underlying mantle (*Jin et al., 1994; Royden, 1997*). Numerical models that have depth-dependent rheology for the Tibetan lithosphere (e.g., *Beaumont et al., 2001; Clark and Royden, 2000*) are opposed by other models with vertical coherent deformation (e.g., *Flesch et al., 2005; Wang et al., 2008*). *Copley et al. (2011)* models the different tectonic regimes of strike-slip and normal faulting and concludes that a mechanical coupling is required between the upper crust of southern Tibet and the underthrusting Indian crust. Strong crust-mantle mechanical coupling has been proposed for central and eastern Tibet by comparing surface observations with mantle deformation inferred from *SKS* shear wave splitting (*Flesch et al., 2005; Wang et al., 2008*). *Holt (2000)* also correlates crustal and mantle strain fields for the same regions, but remarks that the vertical coherence may be influenced more by the velocity boundary conditions imposed on both crust and upper mantle, and thus, strong coupling might not be necessary between crust and upper mantle.

If the lithosphere is undergoing uniform deformation it is expected to have similar anisotropic properties. In a related study using a subset of this data set, the long-period dispersion is used for a detailed investigation of the lithospheric-mantle shear-velocity structure (Section 3.3.5). In contrast to the crustal radial anisotropy across Qiangtang, the upper 100 km of the mantle has weak radial anisotropy. Conversely Songpan-Ganzi has weak crustal radial anisotropy but strong uppermost mantle radial anisotropy (3–8%). Hence, we find no correlation between crustal and mantle radial anisotropy that would give support for a coupled mechanism. Alternatively mineral alignment in crust and mantle may produce different anisotropy when under similar deformation.

4.5.6 Deformation mechanisms

The channel flow models listed in the previous section share different views on the initiation of flow. These divide between models that have flow initiated by stronger Indian crust injected into a weaker lower crust, analogous to a “hydraulic piston” (e.g., *Zhao and Morgan*, 1987; *DeCelles et al.*, 2002), and topography-induced lateral extrusion of the lower crust (e.g., *Molnar and Tapponnier*, 1978; *Bird*, 1991; *Beaumont et al.*, 2001; *Clark and Royden*, 2000; *Klemperer*, 2006). Both of these scenarios base their arguments on the smooth topography characterising the plateau, with high elevations in the west and a smooth descent towards the east. However, despite the smooth morphology, internal structures inferred from the shear velocity models here and in other studies (e.g., *Acton et al.*, 2010; *Yang et al.*, 2012) indicate a more complex scenario. Models of deep crustal flow must take into account the heterogeneous north-south internal structure. Furthermore, at the surface, the plateau and surrounding regions have distinct fault systems indicating different deformation regimes (e.g., *Tapponnier and Molnar*, 1977; *Taylor and Yin*, 2009). The mechanism of normal faults in west and central Tibet is attributed to two opposing perspectives that relate to the different flow initiation type: gravitational collapse (*Molnar and Tapponnier*, 1978), or intense compression from India on the Himalayan arc (*Kapp and Guynn*, 2004).

Molnar and Tapponnier (1978) matched the surface fault systems with earthquake fault plane solutions. They established that compression earthquakes are located on the margins of the high plateau, and that earthquakes within central Tibet include large components of normal faulting, with the extensional component oriented east-west. They explain that normal faulting in the highest parts of the plateau is associated with the adjacent thrust faulting; the high plateau has reached its maximum height and gravitational potential, and therefore, the continuous elevation from thrust faulting yield the high plateau to grow outward instead of increasing elevation. In Figure 4.12a we show the different fault systems colour coded by their type (*Taylor and Yin*, 2009; *Styron et al.*, 2010) superimposed on an orange shade indicating high elevation >5 km. The fault systems match the regional earthquakes also colour-coded according to their mechanism (Figure 4.12b). A strong correlation seems to exist between the surface fault types, earthquake focal mechanisms and high topography in western Tibet, as suggested by *Molnar and*

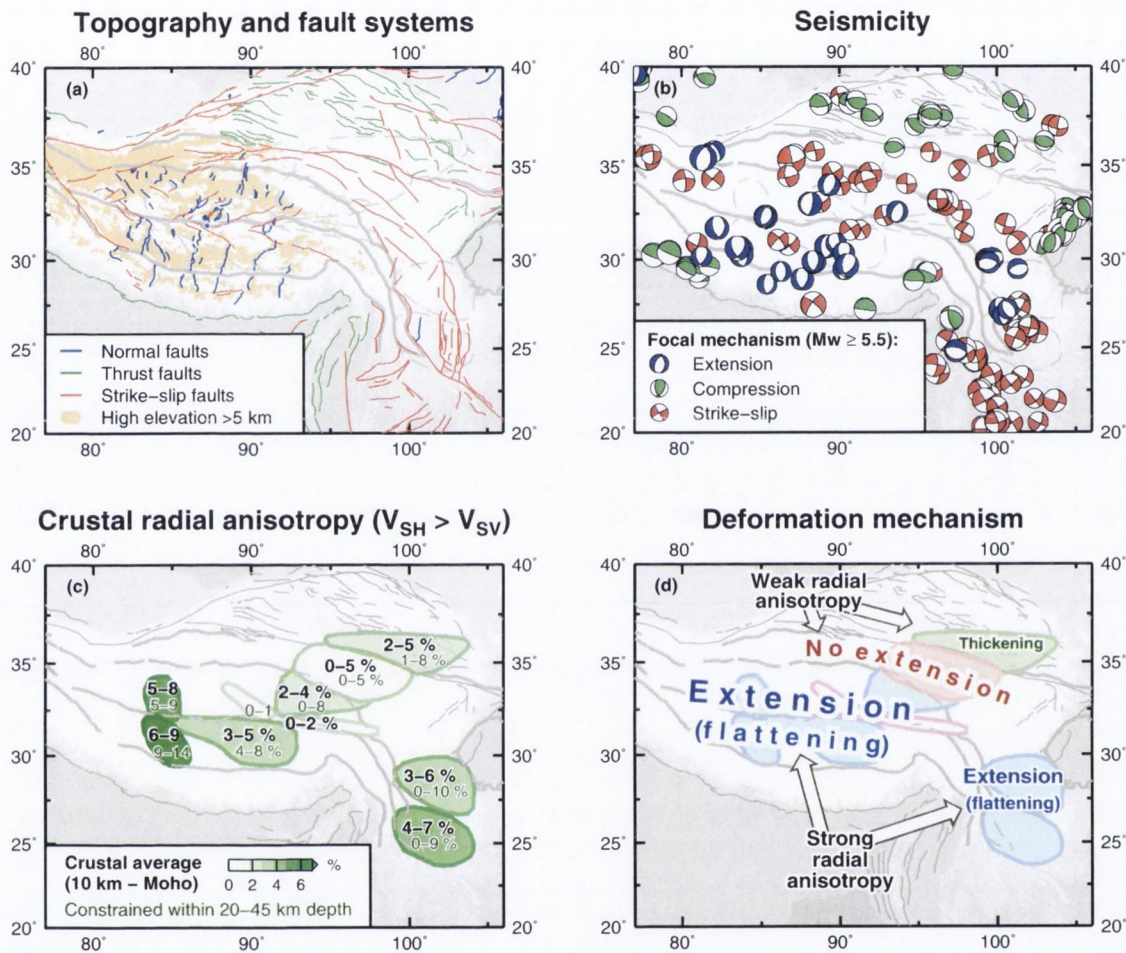


Figure 4.12: Deformation and radial anisotropy in the Tibetan crust. (a) Different fault systems colour-coded according to their type (normal, thrust, and strike-slip) from *Taylor and Yin (2009)*. Orange shade highlights regions of elevation higher than 5 km (using ETOPO1 elevation data set, re-sampled at 5×5 km grid by GMT, *Wessel and Smith, 1998*). Thick grey lines indicate sutures, dashed grey loops indicate regions. (b) Focal mechanism of shallow earthquakes (< 50 km depth) with moment magnitude ≥ 5.5 . The focal mechanisms are well-constrained double couple CMT solutions (<http://www.globalcmt.org>), colour-coded according to their type (extension, compression, and strike-slip). (c) Ranges of crustal-average radial anisotropy ($(V_{SH} - V_{SV})/V_{S(iso)}$), between 10 km depth and the Moho (bold, black numbers), and ranges of radial anisotropy constrained in the 20–45 km depth range (dark green numbers). (d) The correlation of the deformation regime and radial anisotropy within Tibetan crust.

Tapponnier (1978), but not in southeast Tibet. Such a correlation and proposed mechanism argues for a homogeneous crust beneath the high elevations, which is not observed in tomography (*Yang et al.*, 2012) or in our 1-D models (Figure 4.11).

Kapp and Guynn (2004) model the strike of normal faults in Tibet from continued insertion of Indian crust into Tibet. They argue that the onset of extension from the normal faults is not a proxy for when the plateau achieves a maximum elevation and that the plateau crustal thickness and elevation may be increasing concurrent with extension if the flow of Indian crust into Tibet exceeds that being displaced away from the plateau.

We do not find any correlation between the internal high elevation of the plateau and low shear velocity (and its proxies, partial melt and viscosity), and also no correlation between the high elevation and radial anisotropy. We find that radial anisotropy reflects the deformation — the focal mechanism of earthquakes and surface fault types across the plateau. Strong radial anisotropy is observed in regions experiencing crustal extension (flattening) in west-central Tibet and southeast Tibet, whereas regions experiencing only horizontal lateral deformation (strike-slip) have no radial anisotropy (Figure 4.12c and d). Anisotropy is also detected in the northern margins of the plateau beneath regions experiencing crustal thickening. Our results favour the hypothesis of *Kapp and Guynn* (2004) in which the regional extension is a consequence of continued insertion of Indian crust, not only because it accounts for the strong radial anisotropy beneath the normal faults, but also because it explains the relative higher V_S (and lower temperatures) across Lhasa.

4.5.7 Lower crust shear velocities of southern Tibet

The non-unique models from the lower crust test inversions limit us in making robust quantitative conclusions. Nonetheless, these tests can still contribute to the discussion about shear velocities at these depths. In order to get an estimate for the expected lower-crust V_S beneath Lhasa, we infer velocities from the adjacent Indian Shield, assuming that the underthrusting crust beneath southern Tibet is of Indian origin, and that prior to subduction the lower crust had had a very similar lower-crustal shear velocity to those found beneath undeformed India today.

Indian lower crust is considered to be basaltic (*Kaila et al.*, 1968; *Dube et al.*, 1973) and has shear velocities between $3.8\text{--}4.0\text{ km s}^{-1}$ (*Narain*, 1973; *Rai et al.*,

2003; *Acton et al.*, 2010, 2011). These velocities are within the range of velocities of lower crustal rocks measured in laboratories simulating similar depth and temperature conditions. Rocks such as mafic granulite, gabbro-norite-troctolite, and greenschist facies basalt have V_S of 3.85 km s^{-1} , 3.93 km s^{-1} , and 3.96 km s^{-1} respectively, at 1000 MPa and at r.t. (*Christensen*, 1996). The decrease of shear velocity due to the increase in temperature for lower crustal rocks is in the range of $1\text{--}2 \times 10^{-4} \text{ km s}^{-1}$ per degree Celsius (*Nishimoto et al.*, 2008). At temperatures of 400°C , the V_S is between $3.85\text{--}3.95 \text{ km s}^{-1}$ similar to those observed in India's lower crust.

As the lower crust is deepened by approximately $\sim 30 \text{ km}$ (e.g., *Nábelek et al.*, 2009), in order to determine an estimate for the expected S velocity at the new depth, we extrapolate the measurements of *Christensen* (1996) to a pressure of 2000 MPa. The anticipated temperature beneath Lhasa is about 1000°C (*Jiménez-Munt et al.*, 2008; *Chan et al.*, 2009) and the shear velocity reduction due to the high temperature is 0.5 km s^{-1} ($\sim 5 \times 10^{-4} \text{ km s}^{-1}$ per degree Celsius, *Nishimoto et al.*, 2008). The calculated shear velocity for the deepened Indian lower crust is $\sim 3.5 \text{ km s}^{-1}$, well within our better fitting profiles for Lhasa (Figure 4.8, regions 1 and 3).

Fast shear-velocity rocks within the deep crust of southern Tibet have been hypothesized from images of large impedance contrasts (*Hirn et al.*, 1984; *Kind et al.*, 2002; *Schulte-Pelkum et al.*, 2005; *Nábelek et al.*, 2009; *Wittlinger et al.*, 2009; *Nowack et al.*, 2010). *Schulte-Pelkum et al.* (2005) and *Nábelek et al.* (2009) suggested that the deep crustal interfaces imaged by receiver functions are from an eclogitized lower crust. Recently, outcrops of gabbro, diabase and basalt associated with eclogite and ultramafic rocks have been discovered in southern Tibet (*Yang et al.*, 2009). The shear velocity of mafic eclogite at 1000 MPa and r.t. is $\sim 4.6 \text{ km s}^{-1}$ (*Christensen*, 1996). Assuming similar relationships to the inferred deepened lower crust, at deep crustal pressures and temperatures the shear velocity for eclogite decreases to $\sim 4.2 \text{ km s}^{-1}$, far higher than our observed V_S . In this regard we do not observe the very high shear-velocities inferred by *Wittlinger et al.* (2009). *Schulte-Pelkum et al.* (2005) suggests that 30% of the lower Indian crust undergoes the eclogite phase transition. Hence the expected average velocity of a 30 km thick lower crust would be $\sim 3.71 \text{ km s}^{-1}$ — also within our better fitting

models for West and Central Lhasa (Figure 4.8, regions 1 and 3).

Although the better fitting regional profiles for northern Tibet have similar lower-crustal shear velocities to those in Lhasa, we are unable to distinguish differences between terranes (e.g., *Chan et al.*, 2009), and does not necessary mean that Indian lower crust underthrust so far north (e.g., *DeCelles et al.*, 2002).

4.6 Conclusions

We have measured accurate, broadband, Rayleigh- and Love-wave phase-velocity curves using pairs of stations located in various parts of the Tibetan Plateau. At short periods, dispersion curves from within the high-elevation plateau show very low phase velocities, whereas the curves from lower-elevation margins of the plateau show higher phase velocities. In a joint inversion of Rayleigh and Love waves we compute 1-D, radially anisotropic shear-velocity models and run a series of rigorous tests to determine the ranges of crustal S velocity and radial anisotropy most consistent with the data.

Shear velocities in the middle crust are anomalously low everywhere within the high-elevation, central part of Tibet, while also showing strong lateral variations across the plateau. Mid-crustal S velocities decrease towards the north, from $\sim 3.4 \text{ km s}^{-1}$ in Lhasa to as low as $\sim 3.1\text{--}3.2 \text{ km s}^{-1}$ in West Qiangtang and eastern Songpan-Ganzi. This decrease in V_S results in a pronounced mid-crustal low-velocity zone (LVZ) in the northern regions. Velocities increase gradually towards southeast Tibet ($\sim 3.5 \text{ km s}^{-1}$) and increase abruptly across north of Kunlun Fault (Figure 4.11a). Comparing the mid-crustal shear velocities with the expected velocities of metamorphic rocks at corresponding depths and temperatures, we find that high temperatures alone ($\sim 800^\circ\text{C}$, *Mechie et al.*, 2004) can explain the low mid-crustal velocities across Central Lhasa, South-east Qiangtang and South Yunnan. Regions with very low velocities are in locations that have higher temperatures ($800\text{--}1000^\circ\text{C}$, *Hacker et al.*, 2000) in the middle crust, exceeding typical melting temperatures of rocks at these depths (e.g., *Patiño Douce and McCarthy*, 1998). Our conservative estimates yield between 3 and 6%, melt fractions for West Qiangtang and Songpan-Ganzi and melt fractions less than 3, 2 and 1% for North Yunnan, eastern Qiangtang and West Lhasa, respectively (Figure 4.11b). In South

Yunnan, significant melting is unlikely. The higher shear velocities within the thinner crust of Tibet's margins could be explained by the smaller amounts of crustal radiogenic heating and, thus, lower temperatures there.

The depth and lateral distribution of the LVZ show agreement with the depth and lateral distribution of a high conductivity layer, mapped in MT studies. Regions with no LVZ, such as Lhasa, South Yunnan and Qinling-Qilian, have at most intermittent high-conductivity zones, with no apparent melt, whereas north Qiangtang, Songpan-Ganzi and North Yunnan are characterised by a LVZ, high conductivity, and melt. Melt estimates based on V_S are at the lower end of melt estimates typically inferred in published magnetotelluric studies (e.g., *Unsworth et al.*, 2005; *Bai et al.*, 2010).

Dispersion curves measured at different station pairs within broad regions within Tibet are remarkably similar. This indicates, in particular, that the mechanically weak middle crust is present and relatively uniform beneath entire regions, not at narrow zones of deformation. Thus, deformation and flow in the middle crust are likely to be diffuse and distributed over broad areas, rather than taking place in localised channels, along faults (*Bai et al.*, 2010) or elsewhere. The extent of such diffuse deformation is corroborated by the low topographic gradients and smooth topography across Tibet (*Clark and Royden*, 2000). Crustal flow may be driven, in large part, by extensional deformation (crustal flattening) in the western and southern Tibet, possibly resulting in large-scale channel flow towards eastern Tibet (*Clark and Royden*, 2000; *Royden et al.*, 2008).

Our region-average 1-D models confirm the pattern of lateral distribution of radial anisotropy across Tibet determined previously by surface-wave tomography (*Shapiro et al.*, 2004). The strongest radial anisotropy is observed across western Tibet, with moderately strong anisotropy in Central Lhasa, East-central Qiangtang and Yunnan. Interestingly, the strength of radial anisotropy within the crust and mantle lithosphere in central and northern Tibet do not correlate. This suggests that deformation in the crust and mantle lithosphere may be significantly different.

Radial anisotropy in the crust does not correlate with low shear velocities (and, thus, with partial melts and low viscosities) or with elevation. Instead, radial anisotropy reflects the deformation pattern and is the strongest in regions experiencing extension (crustal flattening). Crustal thinning causes the radial anisotropy

probably by aligning mica crystals in horizontal planes (*Shapiro et al.*, 2004). Similarly, horizontal flow may align micas in a vertical plane, resulting in a weak radial anisotropy, as observed in Songpan-Ganzi, a region experiencing strike-slip earthquakes and no overall extension.

5

Multi-layered azimuthal anisotropy and crustal and mantle deformation beneath Tibet

5.1 Introduction

Following India's collision with Asia, the Tibetan Plateau has undergone significant, north-south shortening (e.g., *Yin and Harrison, 2000*). Global Positioning Systems (GPS) measurements show that India is still moving north today. The velocity vectors align north-northeast at stations in central Tibet, with active N–S shortening, and show a smooth, clockwise rotation round the eastern syntaxis

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(e.g., Wang *et al.*, 2001; Zhang *et al.*, 2004). How the deformation is accommodated beneath the surface, within the deep crust and upper mantle, is still poorly understood, however.

A record of deformation at depth can be obtained from the seismic investigation of strain-induced anisotropy in crustal and mantle rocks. Studies of shear-wave splitting (birefringence of shear waves) have revealed strong anisotropy beneath Tibet. Figure 5.1 shows a comprehensive compilation of the shear-wave splitting measurements in and around Tibet to date (McNamara *et al.*, 1994; Hirn *et al.*, 1995; Sandvol *et al.*, 1997; Huang *et al.*, 2000; Herquel and Tapponnier, 2005; Lev *et al.*, 2006; Singh *et al.*, 2006; Singh *et al.*, 2007; Sol *et al.*, 2007; Fu *et al.*, 2008; Kumar and Singh, 2008; Wang *et al.*, 2008; Chen *et al.*, 2010; Zhao *et al.*, 2010; Huang *et al.*, 2011; Leon Soto *et al.*, 2012). The distribution of the fast-propagation directions shows a broad regional trend, with the directions rotating around the east Himalayan syntaxis. A detailed examination, however, reveals that, in many locations, measurements from stations close to one another show different fast directions and split times, the latter ranging from null to >2 seconds. The origins of these apparent inconsistencies are still unclear.

Most shear-wave splitting analyses to date have been performed under the assumption that the splitting originates within a single anisotropic layer. The shear-wave birefringence is predominantly due to anisotropic fabric of the rocks at depth, induced by finite strain. The strain beneath Tibet can be associated with deformation of the continental lithosphere, for example, its N–S horizontal shortening in central Tibet (e.g., McNamara *et al.*, 1994; Hirn *et al.*, 1995; Davis *et al.*, 1997; Huang *et al.*, 2000; Leon Soto *et al.*, 2012), or flow in the asthenospheric mantle (e.g., Leon Soto *et al.*, 2012).

Variations in shear-wave splitting parameters as a function of back azimuth (azimuth towards the earthquake epicentre) are, normally, evidence for multiple anisotropic layers below the station, with different orientation of anisotropic fabric within each (e.g., Silver and Savage, 1994; Savage, 1999; Levin *et al.*, 1999). Gao and Liu (2009) reported two layers with significant anisotropy beneath the station Lhasa, previously marked with null measurements when the data were processed assuming a single anisotropic layer. They concluded that the upper layer had a fast direction consistent with the GPS-constrained surface movement, and the

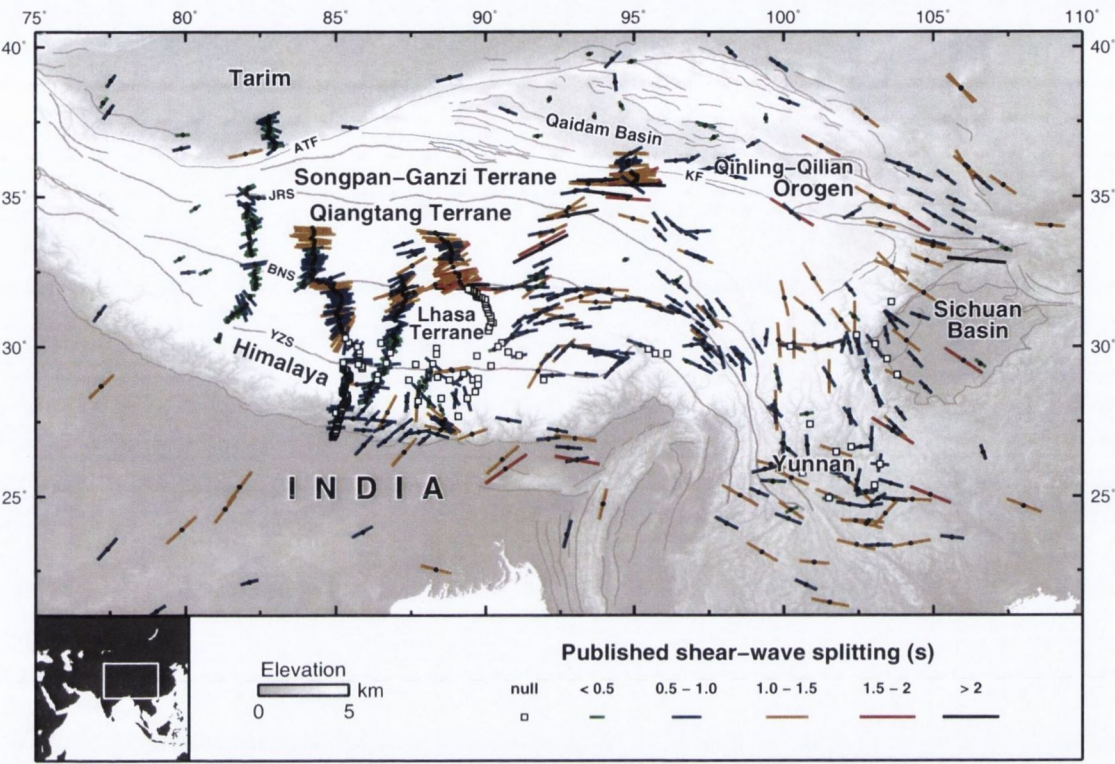


Figure 5.1: Dynamics beneath the Tibetan Plateau and surrounding regions as inferred from shear-wave birefringence. The topographic map shows the contrasting elevation superimposed by a compilation of various shear-wave splitting time and fast polarisation direction from teleseismic waves such as *SKS* and *SKKS* (McNamara *et al.*, 1994; Hirn *et al.*, 1995; Sandvol *et al.*, 1997; Huang *et al.*, 2000; Herquel and Tapponnier, 2005; Lev *et al.*, 2006; Singh *et al.*, 2006; Singh *et al.*, 2007; Sol *et al.*, 2007; Fu *et al.*, 2008; Kumar and Singh, 2008; Wang *et al.*, 2008; Chen *et al.*, 2010; Zhao *et al.*, 2010; Huang *et al.*, 2011; Leon Soto *et al.*, 2012). Colour-coded length-varying bars indicate the split time. Black closed circles are the location of seismic stations. Black open squares are stations with null measurements. Dark grey lines show sutures and faults: YZS (Yarlung-Zangbo Suture), BNS (Bangong-Nujiang Suture), JRS (Jinsha River Suture), KF (Kunlun Fault) and ATF (Altyn Tagh Fault). Bottom left map: Location on a regional map.

lower layer had a fast direction that reflected either N-S shortening due to the India-Eurasia collision or flow in the asthenosphere related to the absolute motion of Eurasia towards the west (Gripp and Gordon, 2002). Similar two-layer modelling has been carried out at a number of sites within the plateau (Levin *et al.*, 2008; Huang *et al.*, 2011; Li *et al.*, 2011; Levin *et al.*, 2012), although with little agreement, so far, on the fast-propagation directions in the layers and on the

interpretation of the anisotropy.

Surface waves provide depth resolution of seismic-velocity structure and anisotropy in the entire lithosphere-asthenosphere depth range. In a number of Rayleigh-wave studies, depth-dependent azimuthal anisotropy beneath Tibet has been reported (*Huang et al.*, 2004; *Su et al.*, 2008; *Yao et al.*, 2010; *Yi et al.*, 2010). Understanding the depth dependence of anisotropy is essential in order to determine the mechanisms of Tibet's deformation, because the deformation itself is likely to be strongly depth-dependent. For example, models with large-scale channel flow within a low-viscosity, middle-crustal layer have been proposed to explain the lateral growth of the plateau (e.g., *Bird*, 1991; *Clark and Royden*, 2000). The weak, mid-crustal layer beneath Tibet (e.g., *Nelson et al.*, 1996) could mechanically decouple the upper crust from the mantle, while, also, accommodating pervasive flow within it quite different from the motions in the layers above and below (e.g., *Royden*, 1997).

In this study we investigate depth-dependent azimuthal anisotropy beneath Tibet using inter-station measurements of phase velocities of Rayleigh surface waves. Using a few tens of highly accurate, broadband dispersion curves from across central and eastern Tibet, we apply a number of alternative surface-wave analysis approaches and determine robust patterns of phase-velocity anisotropy as a function of period. We then use these results to constrain the direction and amplitude of shear-velocity anisotropy as a function of depth and infer the dominant deformation mechanisms that operate in the crust and upper mantle beneath Tibet.

5.2 Measurements

Seismic surface-waves propagating away from an earthquake, because of their dispersive nature, travel at different velocities for different frequencies. Their frequency dependency enable us to infer shear velocity as a function of depth and determine important elastic properties of the rocks. We use accurate, broadband, Rayleigh-wave phase-velocity measurements to detect azimuthal variations within the lithosphere and uppermost mantle. The data is the same as that used in Chapter 3 and 4, and, in addition we include new dispersion measurements for north-eastern Tibet (Songpan-Ganzi), south-east Tibet (across the Yunnan province),

round the eastern syntaxis, and a few west-east paths across central Tibet. Measurements from all data sets are generated using the same methods. Dispersion curves from the new station pairs have similar phase-velocities to adjacent paths from the combined data set.

5.2.1 Two-station method

We use the well-established two-station method to measure the Rayleigh-wave fundamental-mode dispersion (e.g., *Sato*, 1955; *Knopoff*, 1972; *Meier et al.*, 2004; *Endrun et al.*, 2008; *Deschamps et al.*, 2008; *Yao et al.*, 2006, 2010; *Yang et al.*, 2010). We apply two independent techniques to extract broader-period phase-velocities; cross correlation of two seismograms (station-station) (*Meier et al.*, 2004), and, derived phase velocities from source-station waveform fitting (*Lebedev et al.*, 2005).

Inter-station phase velocities are determined by analysing the phase difference of teleseismic surface waves, assuming a Jeffreys-Wentzel-Kramers-Brillouin (JWKB) approximation for a spherically symmetric, isotropic Earth model. In the cross-correlation method, the vertical component seismograms of an earthquake recorded at two stations with the same back azimuth are cross correlated. A series of frequency-dependent band-pass Gaussian filters and frequency-dependent time windows are applied to the cross-correlation function to enhance the signal-to-noise ratio and reduce side lobe energy resulting from correlations of the fundamental mode with higher modes or with other scattered waves such as body waves. The Rayleigh-wave phase velocities are then computed from the phase of the cross correlation function and the stations separation. In an interactive interface, the correct dispersion curve of the fundamental mode is selected manually (*Meier et al.*, 2004).

Long-period surface waves can interfere with body waves and possibly distort the measurement from the cross correlation function. The dispersion of these waves are better measured using full waveform fitting. We use the Automated Multimode Inversion (AMI) by *Lebedev et al.* (2005) to simultaneously fit source-station phase coda including S, multiple S, and surface waves. From the successful fit of an event recorded at both stations we extract the source-station dispersion curve, and then derive the corresponding inter-station Rayleigh-wave fundamental-mode dispersion.

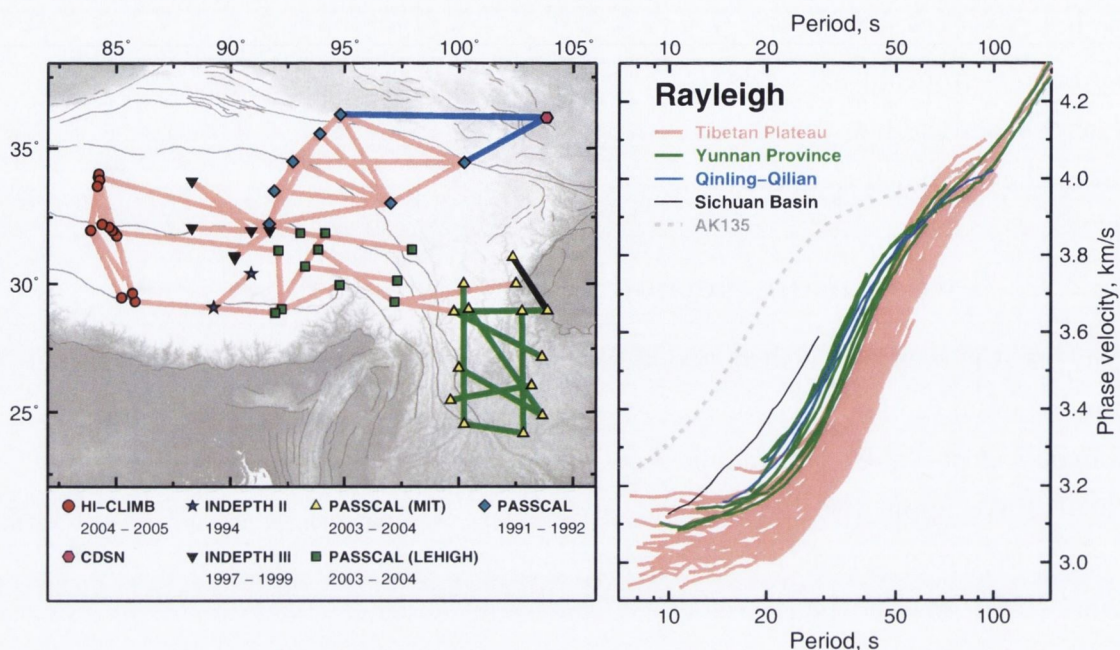


Figure 5.2: Broad-band Rayleigh-wave fundamental-mode phase-velocity measurements from inter-station paths across the Tibetan Plateau. Left: All the seismic stations used for the measurements. Different symbols correspond to different seismic networks operated at specified times. Right: Dispersion curves of all the paths on the map, colour coded to represents different regions of the plateau.

In order to reduce effects from unmodelled surface-wave diffraction, the selected earthquakes have epicenters closely-aligned to the great-circle path of the two stations, less than 10° station-source back azimuth, respectively. These earthquakes have magnitudes between 5 and 9, and epicentral distances to the stations between 4.6° and 150° . Measurements made from earthquakes occurring on opposite sides of the inter-station paths have similar dispersion. The measurements from both methods complement each other and are combined together; the cross-correlation produces most of the measurements, even when waveforms appear complex, and, AMI contributes to the longer period measurements. All the dispersion curves are reselected manually; outlying curves, or narrow-band kinks likely to be frequency-dependent noise or interference from different modes are rejected. This rigorous process is important in order to obtain the most accurate dispersion curves that enable us to observe weak patterns and differences of a small percent ($\sim 1\%$).

The similar dispersion curves from the compilation of all the different earth-

quakes of varying distances, magnitude and directions, and from the two different methods confirms that the JWKB approximations holds. Thus the dispersion curves are representative of the path between the stations. From the average of all the measurements, we produce smooth, robust, accurate phase velocities for broad period ranges. Figures 2.5, 3.4, and 4.2 show examples of combined measurements of AMI and cross correlation, resulting in broadband, smooth dispersion curves from hundreds of consistent measurements.

In total, the combined data set encompasses 56 station pairs distributed across the Tibetan Plateau. The stations are broadband instruments belonging to various temporary networks that operated at different times (PASSCAL (1991–1992) (*Owens et al.*, 1993), INDEPTH II (1994) (*Nelson et al.*, 1996), INDEPTH III (1997–1999) (*Huang et al.*, 2000), PASSCAL (Lehigh, 2003–2004) (*Sol et al.*, 2007), PASSCAL (MIT, 2003–2004) (*Lev et al.*, 2006), and HI-CLIMB (2004–2005) (*Nábělek et al.*, 2005)) and from the permanent China Digital Seismic Network. All the seismograms were retrieved from the Incorporated Research Institutions for Seismology Data Management Centre. Figure 5.2 shows the stations location sorted by their respective networks. Inter-station lines indicate the paired stations used to obtain the dispersion measurements plotted on the adjacent graph. At short periods, between 20 and 40 seconds, paths sampling outside high elevation (across Qinling-Qilian and Sichuan basin) have higher phase velocities than of those paths sampling the high Tibetan Plateau. Yunnan, which has a smooth changing elevation, has intermediate phase velocities of central Tibet and lower elevation (green curves). At very short periods, <20 s, dispersion curves vary strongly from one another indicating significant heterogeneity in the shallow crust, particularly in southern Tibet (close to the Himalayas) and at the eastern syntaxis.

5.2.2 Rayleigh-wave dispersion across Tibet

Phase-velocity maps at different periods show relative velocity changes that highlight contrasting regions within the plateau (Figure 5.3). At 25 s the dispersion measurements of different station pairs in central Tibet have very similar phase velocities, suggesting a homogeneous structure. Velocities increase gradually towards South Yunnan in contrast to an abrupt change across the Kunlun Fault and across Sichuan Basin. The increase in velocity indicates that the crust either thins

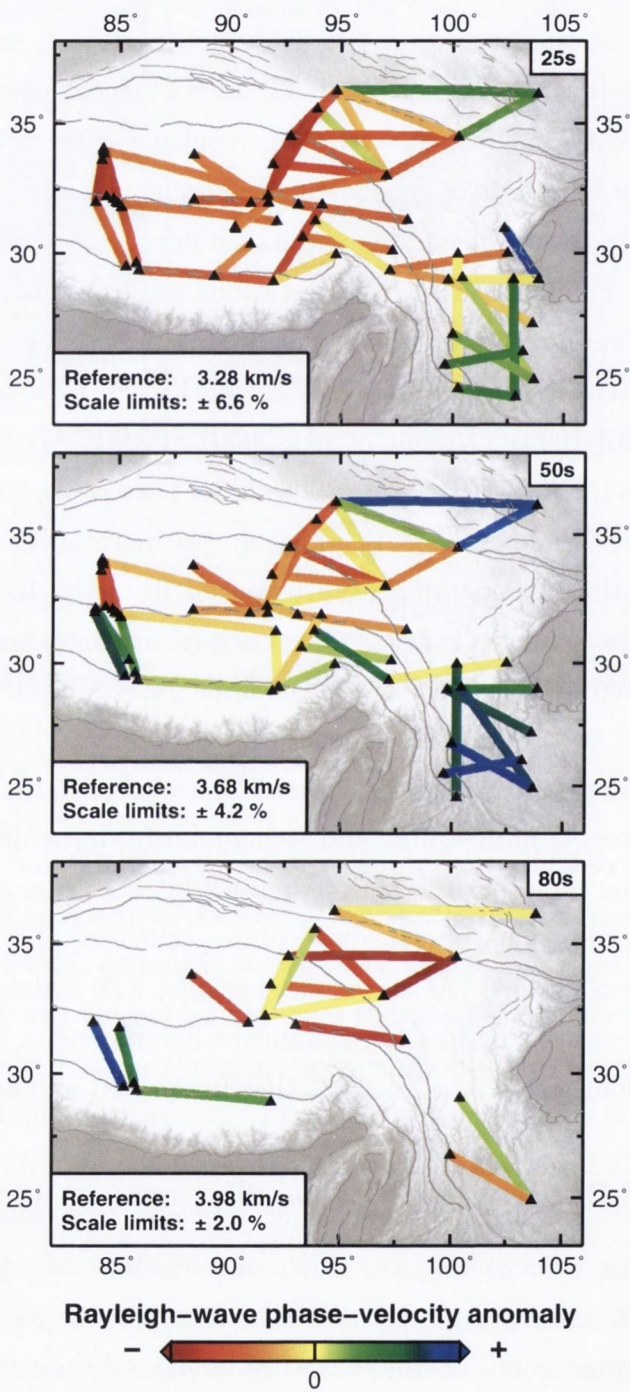


Figure 5.3: Rayleigh-wave fundamental-mode phase-velocity variations across the Tibetan Plateau. Coloured lines show the inter-station paths between stations (black triangles). Different colours correspond to the phase-velocity difference from the reference velocity within the scale limits, at 25, 50 and 80 seconds period.

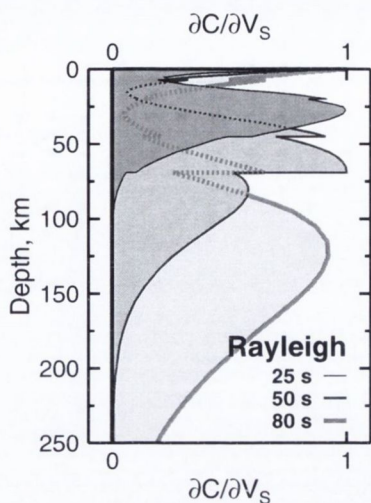


Figure 5.4: Depth sensitivity kernels for the Rayleigh-wave fundamental-mode phase velocities. Different curves represent the Fréchet derivatives of phase velocities with respect to shear velocities and depth for periods 25, 50 and 80 seconds, all scaled independently. These kernels are computed from a 1-D isotropic shear-velocity profile that has a four layer crust and a Moho discontinuity at 69 km depth, typical for East-Central Qiangtang.

or changes in composition, or a combination of both. Upon closer inspection of central Tibet, west-east aligned station pairs have faster velocities than the north-south aligned pairs — possibly an effect of azimuthal anisotropy assuming that the increase in velocity is only due to the alignment of the station pairs. A high-velocity anomaly is also noted for northwest-southeast aligned paths in Songpan-Ganzi.

At longer periods the influence of the lower crust and upper mantle on the surface-wave dispersion becomes evident (Figure 5.3). In central and northern Tibet, phase-velocities are still relatively slower to the faster velocities in the south and outside the high-elevated plateau. In West Lhasa the fast velocities are probably influenced by the strong, cold underthrusting India whereas in northern Tibet the low phase velocities are probably due to a warm shallow upper mantle (Barazangi and Ni, 1982; Priestley *et al.*, 2006; Barron and Priestley, 2009, Section 3.3.2). At 50 s period, regions with lower elevation than central Tibet (north of Kunlun Fault and southeast Tibet) have phase velocities close to global-average models. At 80 s period, phase velocities of adjacent paths within east-central Tibet differ significantly. The paths are in close proximity to each other; at long periods they are likely to sample the same structures.

5.2.3 Depth sensitivity of Rayleigh waves

The frequency-depth sensitivity of the Rayleigh-wave fundamental-mode phase velocities is illustrated in Figure 5.4. The different curves represent the Fréchet derivatives of phase velocities with respect to shear velocities and depth. These

kernels are computed from a typical 1-D four-layer crust shear-velocity model for East-central Qiangtang. Each curve is normalised to highlight the broad sensitivity. A short period 25 s Rayleigh wave has a peak sensitivity between 20–30 km depth. At 50 s, the sensitivity broadens and deepens with a maximum depth sensitivity between 40–70 km. At longer periods the Rayleigh waves are now sensitive primarily to mantle structures. The shades within the curves highlight the phase-velocity depth-sensitive integral of each respective period.

5.3 Estimating azimuthal anisotropy

Ongoing shear strain on crustal rocks can cause anisotropic mineral crystals such as micas (*Weiss et al.*, 1999; *Nishizawa and Yoshino*, 2001; *Meltzer and Christensen*, 2001) and amphiboles (*Barruol and Kern*, 1996; *Tatham et al.*, 2008) to realign. Similarly, upper mantle crystals such as olivine, when under strain, deform to have a lattice preferred orientation parallel to the flow direction (*Zhang and Karato*, 1995). The coherent anisotropic alignment is detected seismically from the speed differences of shear-wave propagation. Hence anisotropic properties of rocks can yield important information on the past or present deformation processes. We estimate the azimuth, amplitude, and depth range of two anisotropic layers beneath Tibet using various methodologically independent approaches.

5.3.1 2-D phase-velocity tomographic maps

The effect of elastic anisotropy on the propagation of surface waves gives rise to an azimuthal dependence on the phase velocities (*Smith and Dahlen*, 1973). Therefore, phase-velocity measurements of individual inter-station pairs are prone to propagation biases arising from their aligned orientation if they coincide with fast anisotropic directions. The regional azimuthal anisotropic pattern is best analysed by a tomographic inversion which determines a solution that describes the dispersive data from structural and azimuthal variations simultaneously (e.g., *Deschamps et al.*, 2008; *Darbyshire and Lebedev*, 2009).

For a specific frequency ω we solve for the anisotropic and isotropic phase-

velocity anomalies, $\delta C(\omega)$ and $\delta C_{iso}(\omega)$ respectively:

$$\begin{aligned}\delta C(\omega) = & \delta C_{iso}(\omega) + A_1(\omega) \cos 2\psi + A_2(\omega) \sin 2\psi \\ & + A_3(\omega) \cos 4\psi + A_4(\omega) \sin 4\psi\end{aligned}$$

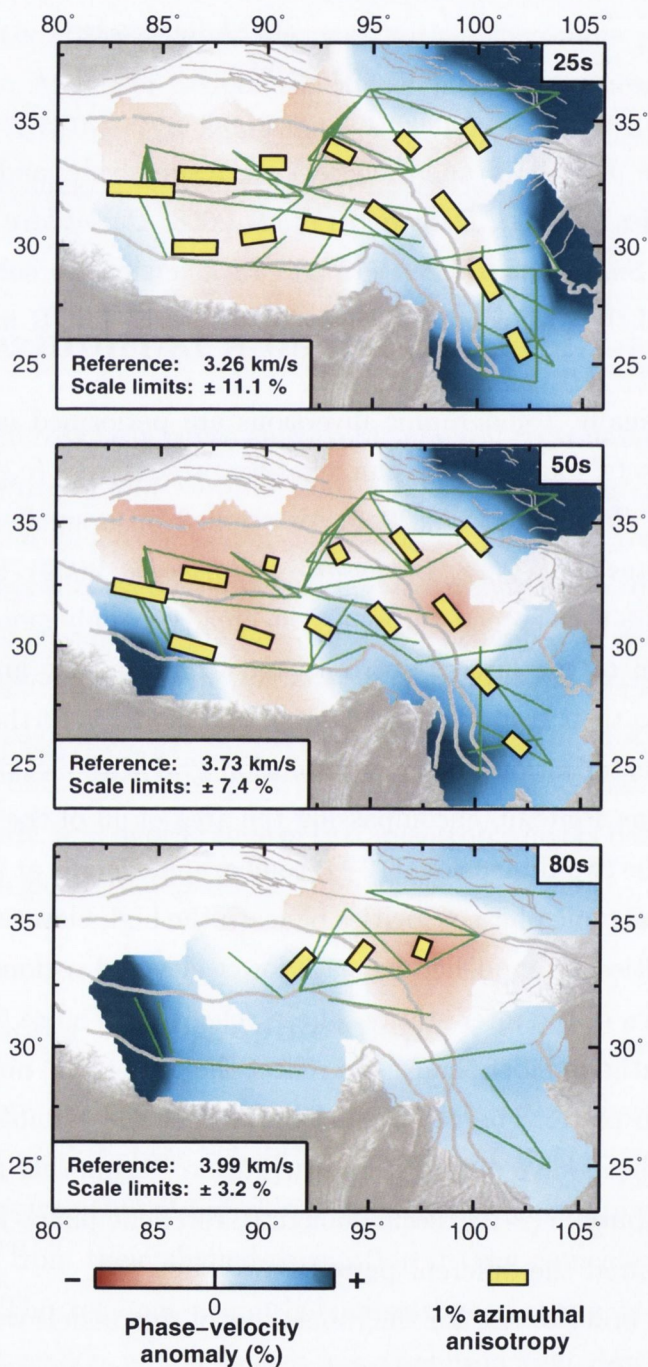
Azimuthal wave-propagation variations are represented by the 2ψ and 4ψ terms to fit the periodic π and $\frac{\pi}{2}$ variations, respectively, and by their corresponding amplitudes A_n (e.g., *Deschamps et al.*, 2008; *Darbyshire and Lebedev*, 2009), adapted from *Smith and Dahlen* (1973). In inversion we solve linearly for a model that fits all the data simultaneously by using the LSQR method (*Paige and Saunders*, 1982).

Usually, tomographic inversions are performed using large regional or global data sets, much larger than what we present here (e.g., *Lebedev and van der Hilst*, 2008; *Priestley et al.*, 2006), or applied on a smaller geographic scale (e.g., *Deschamps et al.*, 2008; *Darbyshire and Lebedev*, 2009). Since we have a small number of pairs that spread across a vast area, we apply smooth and damped parameterisation to the lateral isotropic and azimuthal 2ψ and 4ψ anisotropic variations, and to their respective gradients. The model is further smoothed by applying the parameterisation on a low-resolution grid with ~ 300 km knot spacing across the Tibetan Plateau, encompassing the area of all of the 56 inter-station paths.

The two-dimensional (2-D) tomographic model at 25 s period (Figure 5.5) shows low isotropic phase velocities beneath the high-elevated plateau (pink shade). High velocities are modelled beneath lower-elevated regions (blue shade), similar to the pattern of the inter-station phase velocity in Figure 5.3. At 50 and 80 s the shade in southern Tibet changes to reflect the dispersion, now sensitive to the subducting Indian plate. The yellow sticks represent the azimuthal anisotropy, the preferred fast shear-wave propagation orientation. Azimuthal anisotropy of a small percent in amplitude ($\sim 1\%$) seem enough to reconcile phase-velocity differences within the plateau at the different periods.

In order to test for the robustness of the models we run a series of test inversion using different levels of damping on all the parameters. The models were also tested by running inversions with less data (the removal of inter-station paths that have outlier measurements). Solutions from the test inversions consistently showed the

Figure 5.5: Tomographic maps for Tibet using 25, 50 and 80 seconds period Rayleigh-wave fundamental-mode phase velocities. Thick grey lines are major boundaries across the plateau. Green lines indicate the inter-station paths used. Red and blue shade show relative lower and higher isotropic phase-velocity variation to the reference, respectively. Length-varying yellow sticks show the fast 2ψ anisotropic direction and amplitude.



expected isotropic contrasts, and robust 2ψ azimuthal anisotropic patterns. At 25 and 50 s period frames the preferred fast-propagation orientation is E–W for west and central Tibet, and, NW–SE in eastern Tibet. At 80 s, east-central Tibet has a SSW–NNE orientation for faster propagation. The pattern of azimuthal anisotropy at each period can be visually noticed directly from the fast and slow phase velocities of the inter-station paths (Figure 5.3).

5.3.2 Regional average azimuthal anisotropy

A change in azimuthal anisotropy with period is identified across east-central Tibet; the fast direction changes orientation between 50 and 80 s (Figure 5.5). In order to determine a more accurate azimuth and amplitude of anisotropy at different periods, we run a series of specific tests on the regional measurements only. We compile a data set made up of selected inter-station paths that are adjacent to each other and that together have an azimuthal coverage of about 360° to represent fast propagation in all directions (Figure 5.6). We assume that all the measurements

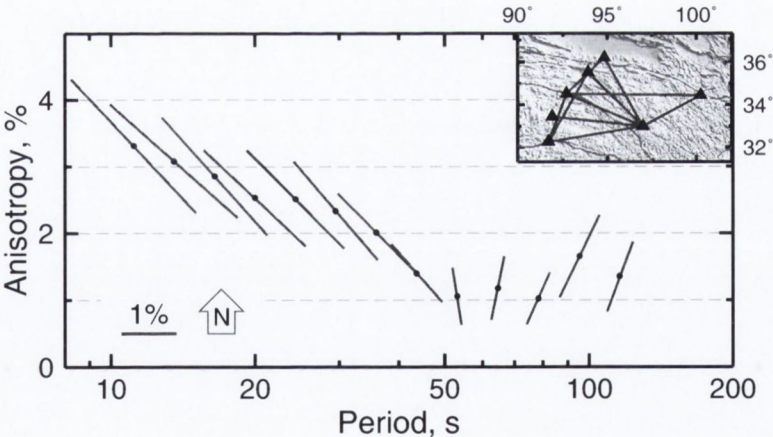


Figure 5.6: Variation of Rayleigh-wave fundamental-mode phase-velocity azimuthal anisotropy across north-eastern Tibet. The regional-average dispersion data is from selected inter-station paths that make up a 360° azimuthal coverage (inset). The measurements of the individual inter-station pairs are weighted at each period, and are assumed to sample a uniform anisotropic structure beneath the paths (*Adam and Lebedev, 2012*). The vertical axis and the length-varying bars represent the strength of the azimuthal anisotropy. The orientation of the bars show the fast propagation direction as viewed on a map (N indicates the North direction).

sample a uniform structure and that variation in velocity is due to an azimuthal dependence. The data set is sorted by azimuth, weighted, and averaged using a 15° sliding window (*Adam and Lebedev, 2012*). At each period a least-squares linear inversion is run to solve for the azimuthal anisotropy that best describes the average velocity. The outcome from all the independently processed periods are two distinct anisotropic patterns (Figure 5.6). At short periods up to 50 s, the fast orientation is NW–SE with a diminishing strength from ~ 3.5 to 1%. The azimuth changes orientation to SSW–NNE within a narrow period range between 50 and 60 seconds. The fast direction remains coherent for periods well over 100 seconds. In a similar analysis for a region farther east to this regional average (north of Sichuan Basin), *Zhang et al. (2011)* finds fast directions in the NWW–SEE direction at all periods with a rapid increase in anisotropy amplitude from 1 to 3% at long periods > 60 seconds.

5.3.3 Direct measurement of azimuthal anisotropy

Applied smoothing during inversions results in averaged estimates and diminished values, particularly of weak-signaled properties like azimuthal anisotropy. The anisotropic strength in the tomographic models is less than 2% whereas in the regional-average anisotropy model the strength reaches a maximum of 3.5%. In order to establish the true anisotropic strength we make direct measurements from the dispersion curves of two adjacent inter-station paths. Assuming that nearby paths sample similar tectonic settings, we calculate the percentage difference from the two Rayleigh-wave phase-velocity curves. After comparing different paths with different azimuths to each other we found that the paths with fast directions match those inferred from the inversions.

Figure 5.7 shows examples of phase-velocity differences at short periods of paired Rayleigh-wave dispersion curves, colour coded to correspond to the mapped inter-station paths. The dispersion differences from inter-station pairs within central Qiangtang have a consistent 3% peak between 20 and 35 s period, the period range most sensitive to the Tibetan mid-crustal depths (Figure 5.4). This azimuthal anisotropy is considered as the minimum strength and could increase if the selected fast and slow paths are not aligned exactly to the real fast and slow directions. Different inter-station pair combinations sampling across different sur-

face tectonics in central Tibet show a similar pattern, which adds confidence to our estimates. The change in azimuthal anisotropy across eastern Qiangtang and Songpan-Ganzi inferred in tomography (Figure 5.5) is evident in the direct measurements. Selected pairs reflecting the azimuthal change show a different pattern in the dispersion differences.

Collectively, the dispersion of near-W-E and near-N-S aligned pairs split naturally between ones with high and low phase velocities for periods between 20 and 35 s (Figure 5.11, dispersion curves) — supporting the assumption that the dispersion sample a relatively uniform crust and that the differences in phase velocity at mid-crustal periods are of an anisotropic nature. We disregard any significance of peaks or rough changes in the dispersion difference which could potentially be due to noise in the phase-velocity curves; therefore, we determine the average anisotropic estimates between 20 and 35 s (grey shade). In Figure 5.11 we plot length-varying arrows between the paired inter-station paths representing their average phase-velocity percentage difference (green arrows). Each arrow is oriented mid-angle between the parallel and perpendicular direction of the fastest and slowest path, respectively, resulting in a similar plot to the tomographic inversion. Beneath the arrows are the paired inter-station paths that show the faster E-W trending paths (pale blue) and the slower N-S trending paths (pale red).

A similar approach is adapted for the longer periods (>50 s) to investigate azimuthal anisotropy in the upper mantle. A region of interest is East-central Qiangtang which has paths with significant phase-velocity differences (Figure 5.3, 80 s). Whilst the uppermost mantle beneath Tibet is expected to be heterogeneous, particularly because of India's northward subduction, we deem that long-period dispersion of adjacent paths located in very close proximity to one another sample similar (same) deep structures. Figure 5.8 shows comparisons of paired broadband Rayleigh-wave dispersion curves zoomed in the longer periods. North-south trending station paths in East-central Qiangtang (AMDO-ERDO, AMDO-BUDO, and AMDO-USHU) have very similar fast dispersion curves whereas in-between west-east trending station paths (USHU-WNDO, ERDO-USHU, and BUDO-USHU) have slow phase velocities. The paths of fast dispersion are sub-parallel to the fast SSW–NNE direction inferred from the inverted models (Figures 5.5 and 5.6). In Figure 5.12 we plot the fast and slow dispersion curves (for periods greater than

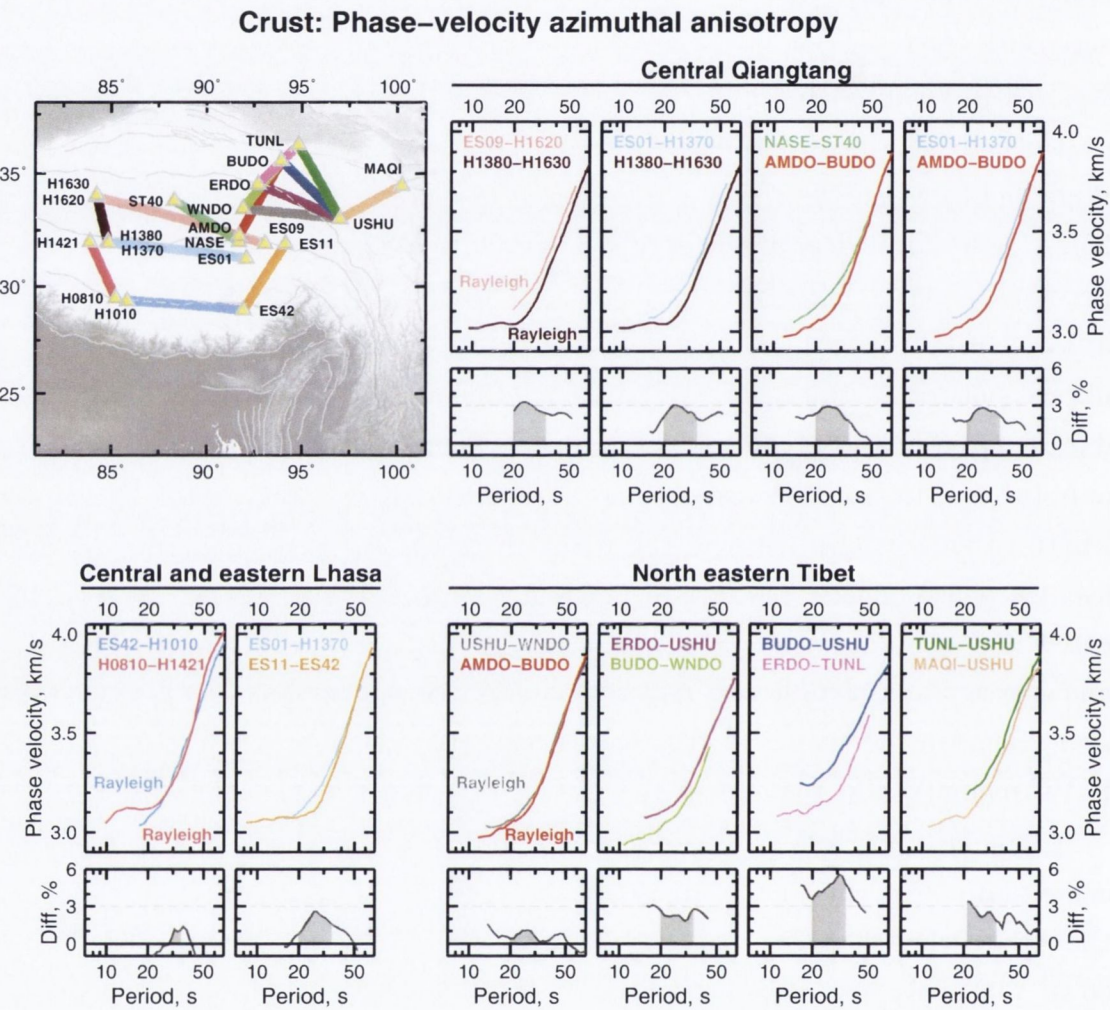


Figure 5.7: Direct measurements of crustal azimuthal anisotropy across central and eastern Tibet from selected Rayleigh-wave dispersion curves. These estimates are based on the assumption that velocity differences at middle-band periods are due to the path's orientation. Top: Rayleigh-wave phase-velocity curves of fast propagating (west-east) and slow propagating (north-south) aligned station pairs across Central Qiangtang. The colour of each dispersion curve corresponds to the inter-station path on the map. Below the dispersion curves is the corresponding phase-velocity percentage differences between the two paths. Grey shade highlights the phase-velocity percentage difference between 20 and 35 seconds period range, sensitive to the middle crust (Figure 5.4). Dashed grey line marks 3% difference. Bottom: Same as top, but for Lhasa and north eastern Tibet.

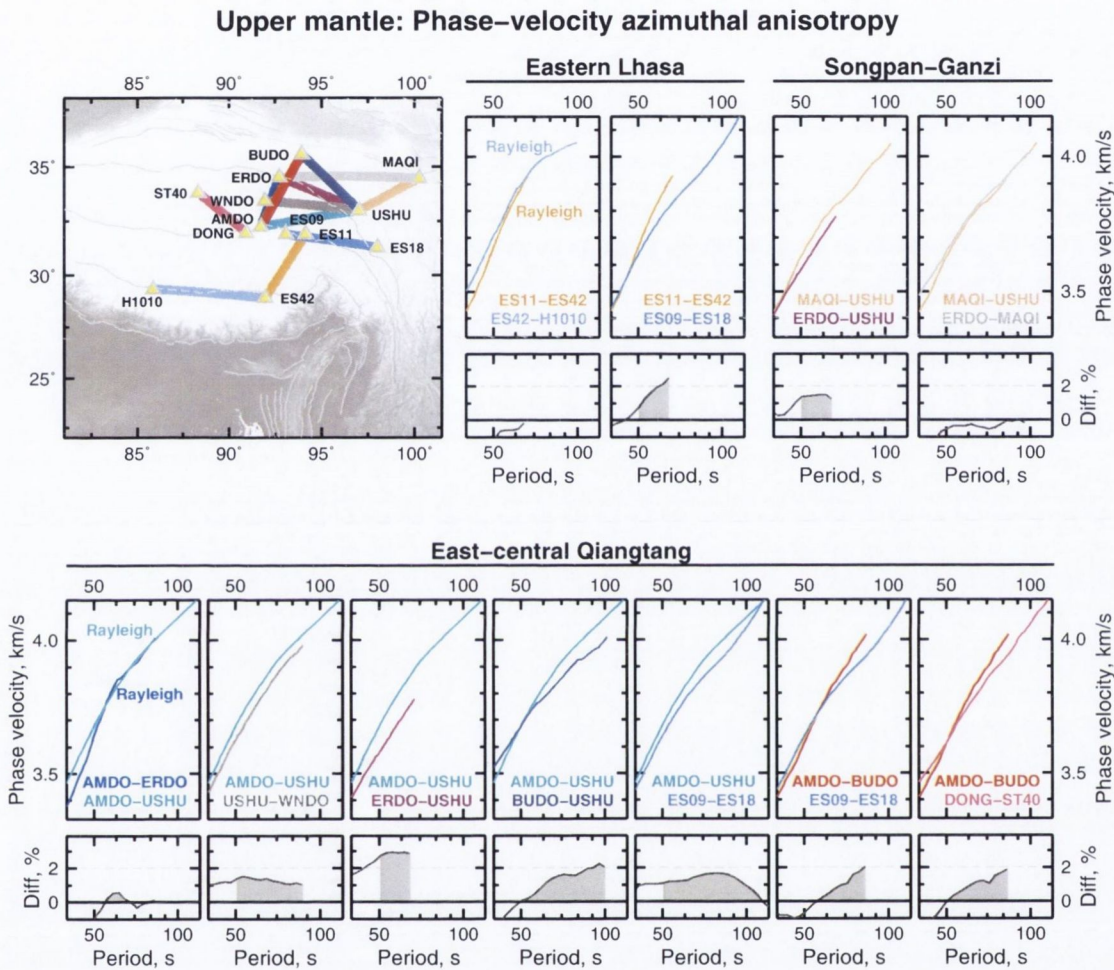


Figure 5.8: Direct measurements of upper mantle azimuthal anisotropy across eastern Tibet from selected Rayleigh-wave dispersion curves. These estimates are based on the assumption that velocity differences at long periods are due to the path’s orientation. Top: Rayleigh-wave phase-velocities curves of fast propagating (north-south) and slow propagating (west-east) aligned station pairs across Eastern Lhasa and Songpan-Ganzi. The colour of each dispersion curve corresponds to the inter-station path on the map. Below the dispersion curves is the corresponding phase-velocity percentage differences between the two paths. Grey shade highlights the phase-velocity percentage difference for periods ≥ 50 seconds, sensitive to the upper mantle. Dashed grey line marks 2% difference. Bottom: same as top, but for East-central Qiangtang.

50 s), the corresponding fast-slow station paths, and the average phase-velocity percentage difference of the paired paths (purple arrows).

5.3.3.1 1-D shear-velocity azimuthal anisotropy model

To constrain the depth range sensitive to the difference between two dispersion curves, we invert both Rayleigh-wave phase-velocities for a 1-D shear velocity with depth profile. Both dispersion curves are fit simultaneously during inversion by solving for a 1-D profile of an average shear speed $(V_{SV1}+V_{SV2})/2$ and anisotropy $(V_{SV1}-V_{SV2})/2$. Both velocity profiles are representative of the vertically polarised shear velocity V_{SV} corresponding to each of the inter-station paths, respectively. The difference in V_{SV} is assumed to be due to azimuthal anisotropy.

The inversion is a least squares, gradient search algorithm based on the Levenberg-Marquardt implementation. This method is ideal for curve fitting problems such as these dispersion curves. The model space is explored with small perturbations in the background shear-velocity model. At each perturbation, synthetic phase velocities are generated (Masters, <http://geodynamics.org/cig/software/mineos>) and compared to the data, making the approach fully non linear. At the end of each iteration, a solution with a lower data-synthetic residual is determined.

The flexibility of the inversion to perturb the background model is defined by basis functions. These define the model space and also control the V_S isotropic and anisotropic changes within the model. Although the increase of basis functions result in a model that better fits the data, attention needs to be made not to over fit parts of the dispersion curves contaminated with frequency-dependent noise (usually manifested as roughness in the curves). A balance needs to be found in what is realistically resolved by the data. Deep, thin, anomalous layers are unlikely to be robust features and can easily be fit by a less anomalous alternative model. A detailed investigation of phase- to shear-velocity inversion trade-offs and non-unique 1-D S -velocity models for Tibet has been demonstrated in Chapter 4 where it was determined that a four-layer crust has an equally good fit to a model with significantly more parameterisation. In this study we take a similar approach; four layers parameterise the crust to fit the short-period data, and 10 basis functions are defined throughout the upper mantle to accommodate the deeper sensitivity of

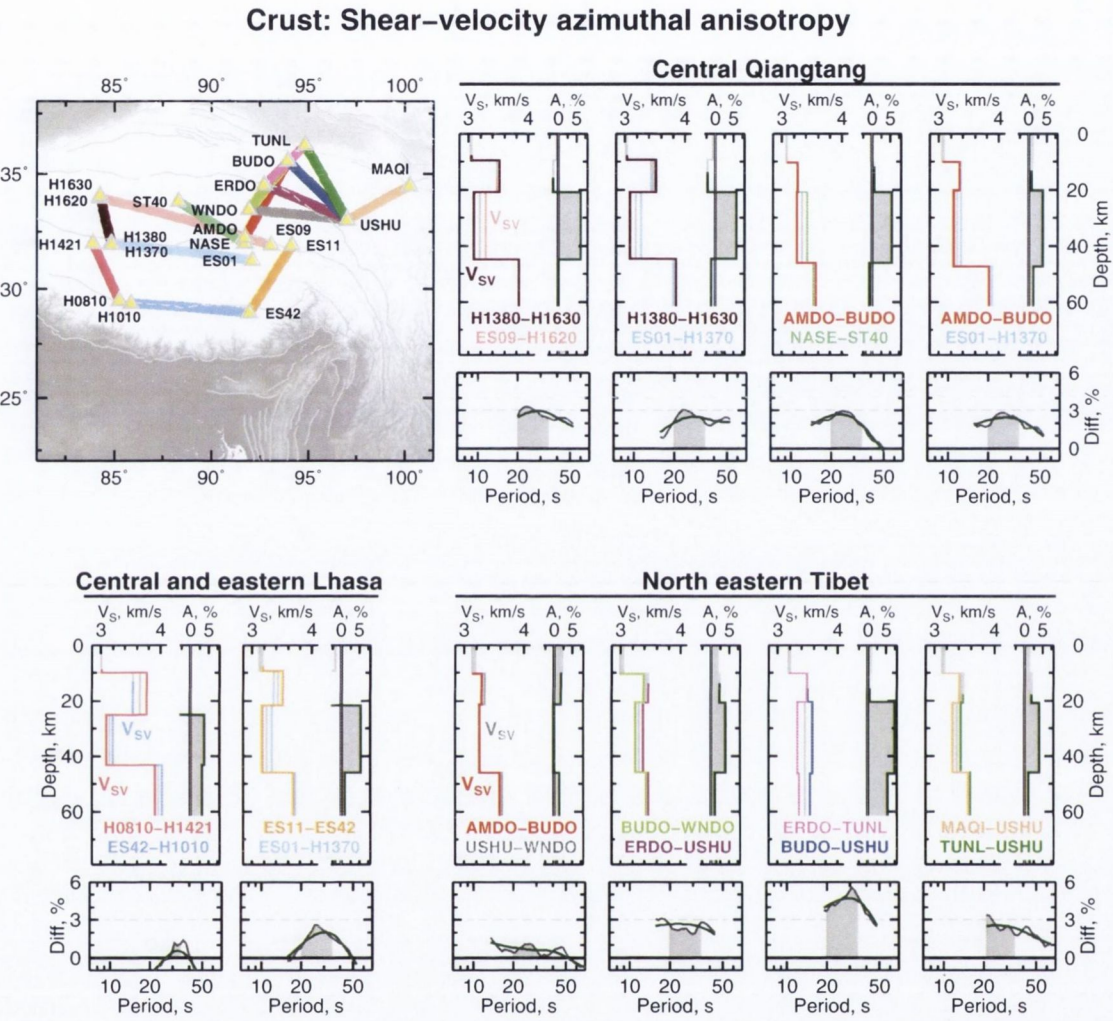


Figure 5.9: Crustal 1-D shear-velocity azimuthal anisotropy across central and eastern Tibet from the joint inversion of two Rayleigh-wave dispersion curves to S velocity. Top: 1-D S velocities of fast propagating (west-east) and slow propagating (north-south) aligned station pairs across Central Qiangtang and the corresponding 1-D anisotropy and dispersion percentage difference. The selected paths are the same as in Figure 5.7. Each colour-coded model corresponds to an inter-station path on the map. Coloured 1-D profile depth ranges reflect the depths most sensitive to the Rayleigh waves. Grey shade in the 1-D anisotropy highlights faster propagation along west-east oriented paths. Shaded phase-velocity percentage difference indicate periods most sensitive to the middle crust (Figure 5.4). Dark green curve is the synthetic dispersion percentage difference. Dashed grey line marks 3% dispersion difference. Bottom: Same as top, but for Lhasa and north eastern Tibet.

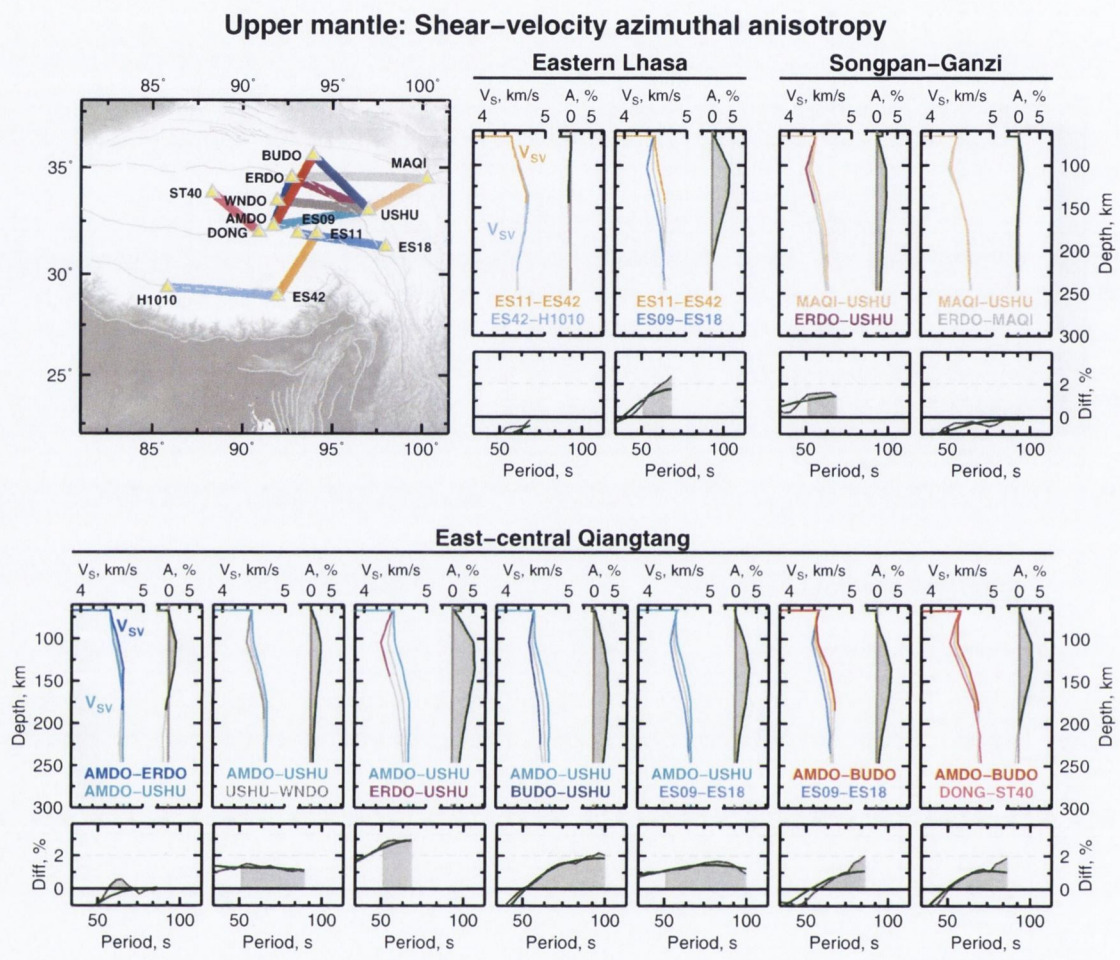


Figure 5.10: Upper mantle 1-D shear-velocity azimuthal anisotropy across central and eastern Tibet from the joint inversion of two Rayleigh-wave dispersion curves to S velocity. Top: 1-D S velocities of fast propagating (north-south) and slow propagating (west-east) aligned station pairs across Eastern Lhasa and Songpan-Ganzi. Each model has a corresponding 1-D anisotropy profile and dispersion percentage difference. The selected paths are the same as in Figure 5.8. Each colour-coded model corresponds to an inter-station path on the map. Coloured 1-D profile depth ranges reflect the depths most sensitive to the Rayleigh waves. Grey shade in the 1-D anisotropy highlights faster propagation along north-south oriented paths. Shaded phase-velocity percentage difference indicate periods most sensitive to the upper mantle (Figure 5.4). Dark green curve is the synthetic dispersion percentage difference. Dashed grey line marks 2% dispersion difference. Bottom: same as top, but for East-central Qiangtang.

longer periods. Four crustal depth discontinuities are parameterised and are free to perturb during the inversion. In order to attain a very good fit to the data and yet achieve smooth models, slight damping was employed during the inversions.

Figures 5.9 and 5.10 show examples of 1-D shear-velocity models for the crust and mantle from the inversion of paired Rayleigh-wave phase velocities (using the same dispersion curves shown in Figures 5.7 and 5.8). To highlight the parts of the models that constrain the data, the plots of the 1-D models and dispersion difference are zoomed into the sensitive depths and periods. The percentage difference in the shear speeds is shown in the corresponding 1-D anisotropy profiles (grey shade). The synthetic dispersion of each of the V_{SV} models fits well with the respective data at all periods (within the standard deviation of the data and with $<1\%$ data-synthetic misfit). The dispersion difference between the two synthetic Rayleigh-wave phase velocities are very similar to the difference of the measurements (green curve in the period-percentage difference frame). Narrow-band, oscillatory variations between the measured and synthetic dispersion differences are likely to be from frequency-dependent noise in the data. The coloured depth ranges of the V_S and anisotropy profiles indicate the depth sensitivity of the data. Figures 5.11 and 5.12 show the spatial distribution of the 1-D S -velocity azimuthal anisotropy, the same as that of the average dispersion differences.

5.4 Results

In this study we present various estimates for azimuthal anisotropy across the Tibetan Plateau. We use accurate phase-velocity measurements to determine the depth-dependent azimuthal anisotropy across Tibet and to establish the amplitude of the anisotropy.

The tomographic model (Figure 5.5) shows low, isotropic phase velocities beneath the high-elevated plateau in contrast to the neighbouring lower-elevation high velocity regions: the Sichuan Basin, Qinling-Qilian and South Yunnan. The smooth, weak isotropic variations at 25 s period within central Tibet suggest a very homogeneous middle crust. At 50 s fast velocities are observed in West Lhasa probably due to the strong Indian lithosphere underthrusting beneath the plateau (e.g., *Nábělek et al.*, 2005, Section 3.3.1). Coherent 2ψ azimuthal anisotropy of a

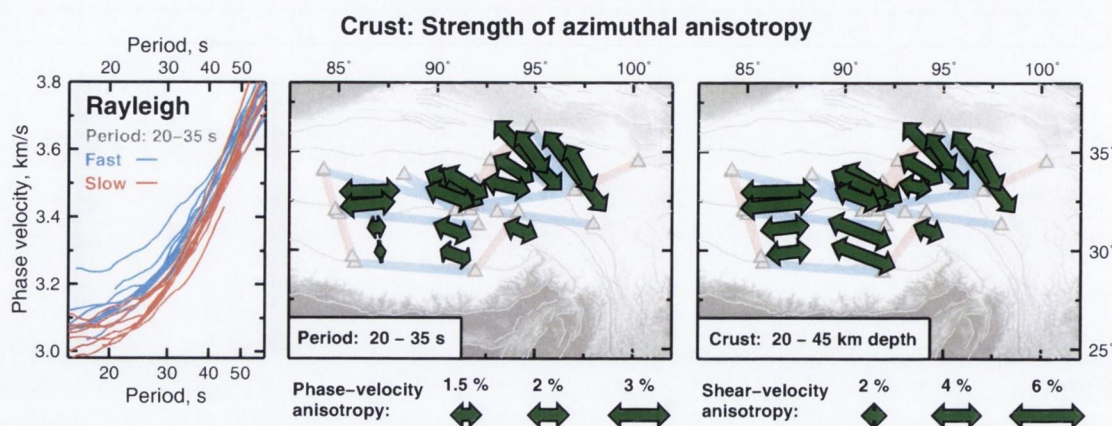


Figure 5.11: Mid-crustal azimuthal anisotropy across Tibet. Left: Selected Rayleigh-wave phase-velocity curves of paths that have fast propagating dispersion (blue curves) paired with paths that have slower propagation (red curves) for periods between 20 and 35 seconds (similar paths as in Figures 5.7 and 5.9). Maps: Paired inter-station paths colour coded for fast (pale blue) and slow (pale red) propagation corresponding to the Rayleigh-wave dispersion curves. Left map: Green arrows represent the Rayleigh-wave average phase-velocity percentage difference between 20–35 s period of specific oriented paths (grey shade in Figure 5.7). The arrows are plotted between the two paired inter-station paths and oriented mid-angle between the parallel and perpendicular direction of the fastest and slowest paths, respectively. Right map: Same as left map but for 1-D shear-wave azimuthal anisotropy estimates for the middle crust (~ 20 –45 km depth) from the joint inversion of two Rayleigh-wave dispersion curves for shear velocity (Figure 5.9).

few percent is required to reconcile phase-velocity differences within the plateau. In west and central Tibet azimuthal anisotropy indicates an E–W orientation for fast surface-wave propagation. Towards the east, the azimuthal pattern rotates clockwise and align to a NW–SE direction, north of the eastern Himalayan syntaxis and in south-east Tibet. At 80 s the azimuth beneath East-central Qiangtang is SSW–NNE. The sudden change in direction takes place within a narrow frequency band between 50 and 60 s (Figure 5.6).

The amplitude of the anisotropy is determined from the direct comparison of dispersion measurements. Paths with fast directions match closely the fast directions inferred in tomography and regional average inversions. The direct measurements also narrow the period range that is mostly affected by the anisotropic velocities. At crustal periods W–E aligned paths in central Tibet have 3% faster phase velocity between 20 and 35 seconds. The anisotropic characteristic is persis-

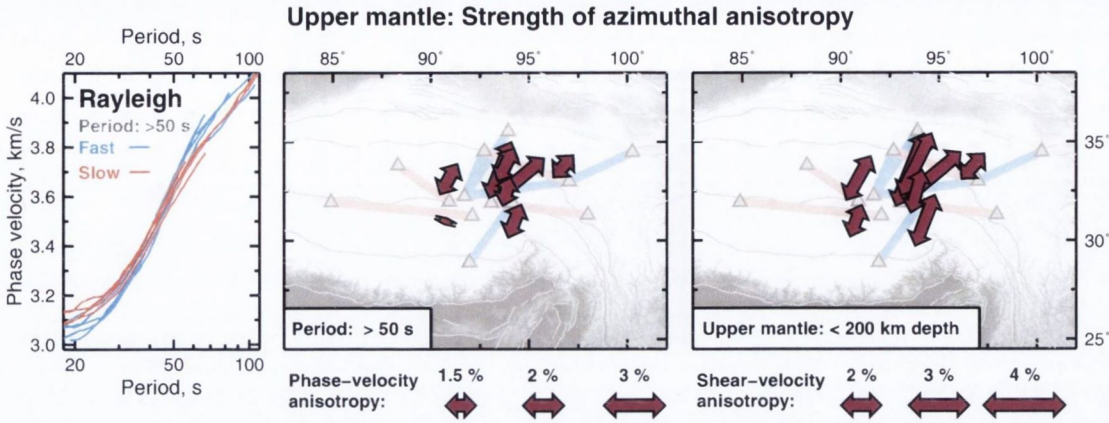


Figure 5.12: Upper mantle azimuthal anisotropy across east-central Tibet. Left: Selected Rayleigh-wave phase-velocity curves of paths that have fast propagating dispersion (blue curves) paired with paths that have slower propagation (red curves) for periods ≥ 50 seconds (similar paths as in Figures 5.8 and 5.10). Maps: Paired inter-station paths colour coded for fast (pale blue) and slow (pale red) propagation corresponding to the Rayleigh-wave dispersion curves. Left map: Purple arrows represent the Rayleigh-wave average phase-velocity percentage difference for periods ≥ 50 seconds of specific oriented paths (grey shade in Figure 5.8). The arrows are plotted between the two paired inter-station paths and oriented mid-angle between the parallel and perpendicular direction of the fastest and slowest paths, respectively. Right map: Same as left map but for average 1-D shear-wave azimuthal anisotropy estimates for the upper mantle (< 200 km depth) from the joint inversion of two Rayleigh-wave dispersion curves for shear velocity (Figure 5.10).

tent even when paths are across different terranes (Lhasa and Qiangtang). Towards north-eastern Tibet the anisotropic strength varies with the orientation change of the paths, from 1 to 5% peak when in the NW–SE direction (Figure 5.7). In inversion of phase velocity for 1-D shear velocity, the models for central Tibet have a mid-crustal anisotropic layer between 20 and 45 km depth, coinciding with a LVZ (Figure 5.9). The faster propagating paths have V_{SV} faster by about 5%. In eastern Tibet, as anticipated from the dispersion, the anisotropy varies with the change in alignment of the station pairs, from 2 to $\sim 7.5\%$ (Figure 5.11).

Azimuthal anisotropy is also detected in the upper 150 km of the mantle beneath east-central Tibet. At long periods (> 50 s) the fast orientation beneath east-central Qiangtang is SSW–NNE (Figure 5.5) and has an amplitude of about 2% (Figures 5.6 and 5.8). 1-D shear-velocity models indicate that the azimuthal

anisotropy is within a LVZ presumed to be the Tibetan asthenosphere (Figure 5.10). A similar anisotropic layer is also inferred at other paired combinations close to east-central Tibet, but not in southern Lhasa (Figure 5.12). The average amplitude of the S -velocity anisotropy in east-central Tibet is in the range of 3–4%.

We stress that these estimates are under the assumption of a uniform structure sampled beneath the matched inter-station pairs, and that the difference in phase velocities and inferred shear velocities are of an anisotropic nature. We cautiously note that the amplitude of the isotropic heterogeneity from the collective pairs is comparable to the amplitude of the anisotropy (overlapping fast and slow dispersion curves in Figures 5.11 and 5.12). Hence, under the set assumptions, these estimates represent the minimum azimuthal anisotropy strength and could increase if the selected fast and slow paths are not aligned exactly to the real fast and slow directions. Isotropic heterogeneity, likely within the upper 20 km of the crust and at the peripheral of the plateau would distort these estimates. Nonetheless, the different station-pair combinations sampling across different surface tectonics display robust patterns.

The various estimates of azimuthal anisotropy provided here show coherent results of a multi-layered anisotropic structure. Figure 5.13 shows two examples of 1-D shear-velocity models for East-central Tibet. Although the anisotropy in the mantle has a broader depth range than that within the crust, the crustal anisotropy is significant. More importantly is the change in the fast direction between the two layers by about 90°.

5.5 Discussion

Seismic azimuthal-dependent anisotropy, faster seismic wave propagation in a specific direction, is often interpreted as a record of present-day deformation (e.g., *Becker et al.*, 2012; *Endrun et al.*, 2011). Seismic experiments have successfully detected azimuthal anisotropy beneath Tibet using various techniques such as regional surface-wave dispersion (e.g., *Griot et al.*, 1998; *Huang et al.*, 2004), shear-wave birefringence (e.g., *McNamara et al.*, 1994; *Leon Soto et al.*, 2012) and receiver functions (e.g., *Vergne et al.*, 2003; *Levin et al.*, 2008). Albeit the numerous studies, the pattern and magnitude of the azimuthal anisotropy inferred from the

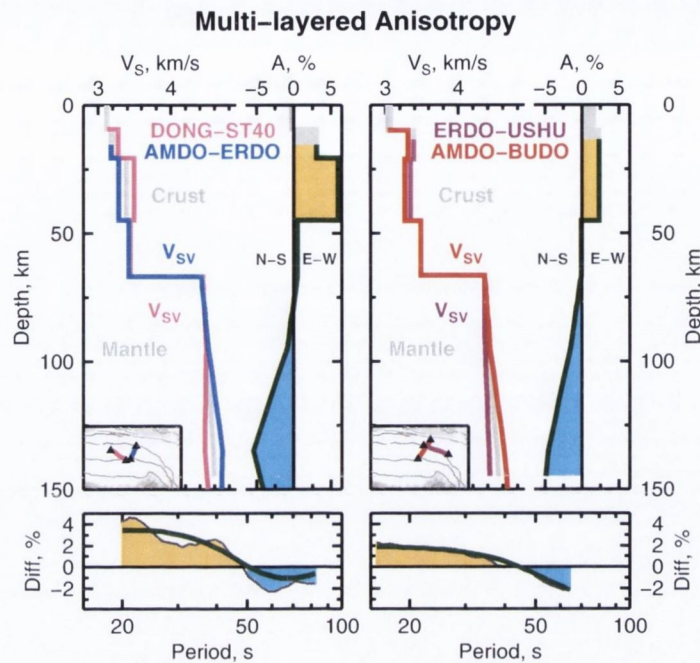


Figure 5.13: Shear-velocity and azimuthal anisotropy with respect to depth. 1-D shear-velocity models from the inversion of paired Rayleigh-wave dispersion curves of near-perpendicular selected inter-station paths located in east-central Tibet. Inset map: The location of the paired inter-station paths, colour coded to match the 1-D models. Positive and negative anisotropy reflects different directions of fast propagation, east-west and north-south trending orientations, respectively. Orange and blue shade indicate the period range most sensitive to the crustal and mantle depths, respectively. Dark green curve is the synthetic dispersion percentage difference from the corresponding models.

different techniques differ, and furthermore, it is still unclear what is the depth range and cause of the anisotropy. Our results show two anisotropic layers confined within the lithosphere and asthenosphere, respectively, both governed by the India-Eurasia convergence.

5.5.1 Amplitude and azimuth of anisotropy constrained by Rayleigh-wave dispersion

Patterns of azimuthal anisotropy inferred from regional tomographic Rayleigh-wave group velocity (*Huang et al.*, 2004; *Su et al.*, 2008) and phase velocity (*Yao et al.*, 2010; *Yi et al.*, 2010) differ not only in amplitude but also with the inferred

fast directions at different periods. At ~ 10 s within the plateau *Su et al.* (2008) has a smooth changing azimuthal anisotropy rotating clockwise from SSW–NNE in far west Tibet to NW–SE in eastern Tibet with an amplitude varying from 1–5%, whereas *Huang et al.* (2004) has sudden variations in amplitude and azimuth in central Qiangtang. At ~ 60 s, *Huang et al.* (2004) has an azimuthal clockwise rotation around east-central Lhasa ($90\text{--}95^\circ\text{E}$) to align from SW–NE in the west to NW–SE in the east with very strong amplitudes of 10%, whereas *Su et al.* (2008) has a wider rotation round $85\text{--}95^\circ\text{E}$ and weaker anisotropy of less than 5%. At ~ 150 seconds, both *Su et al.* (2008) and *Huang et al.* (2004) have weak WSW–ENE anisotropy in central Tibet, but the latter model has strong anisotropy in Tarim and Burma, whereas the former maintains a weak anisotropy throughout. Phase velocities disagree too. For example, at ~ 30 s beneath North Yunnan, *Yao et al.* (2010) has a $\sim 2\%$ fast N–S direction whereas *Yi et al.* (2010) has very weak anisotropy. In a qualitative comparison with group velocity, at ~ 30 s *Su et al.* (2008) has a 5% fast NW–SE direction for the same region. (At short periods, group-velocity dispersion are sensitive to a slightly shallower structure than phase-velocity dispersion (*Smith et al.*, 2004; *Lebedev et al.*, 2012)). At 100 s, *Yao et al.* (2010) and *Yi et al.* (2010) disagree significantly, particularly in eastern Qiangtang with both models having different, strong variations.

Our azimuthal anisotropy inferred from the tomographic maps (Figure 5.5) shows different anisotropic patterns than those previously published. In west-central Tibet, at 85°E , our models for 25 and 50 s (the periods sensitive to the mid-lower crust) show relatively strong W–E anisotropy whereas others show fast N–S directions at similar depth sensitivity (*Huang et al.*, 2004; *Su et al.*, 2008; *Yi et al.*, 2010). We note that these models have few stations located in west Tibet, particularly they lack the stations from the HI-CLIMB profile. This profile certainly adds more constraint to the models. Similarly, at longer periods > 80 s, our tomography and regional-average phase-velocity models for north-eastern Tibet have very good local station coverage (Figures 5.5 and 5.6); both our models indicate a SSW–NNE fast directions which differ from the WSW–ENE and N–S fast directions inferred from the long period group and phase velocity, respectively.

The different methods applied in this study show how the amplitude of the anisotropy strongly depends on the technique and parameterisation adapted. Though

the addition of more paths in the tomographic model increases the azimuthal resolution, consequently the model results in strongly averaged isotropic and anisotropic estimates. At 25 s, the anisotropy has an amplitude of $\sim 1.5\%$ (Figure 5.5). The anisotropy determined from the more specific regional-average phase velocity has less paths to fit and assumes a uniform structure; the recovered amplitude at 25 s is now 2.5% (Figure 5.6). Finally, from the direct comparison of just two inter-station paths and without any averaging, the ‘true’ azimuthal anisotropy at 25 s in Central Qiangtang is about 3% and up to 5% in north-eastern Tibet (Figures 5.7 and 5.11). Likewise, at longer periods the phase-velocity anisotropy has a maximum amplitude of 3% (Figures 5.8 and 5.12).

5.5.1.1 Depth-dependent azimuthal anisotropy

The depth dependence of the dispersion data gives an indication of the structural azimuthal variation and in turn helps in understanding depth-varying deformation. *Huang et al.* (2004) divide the anisotropic pattern at different periods into three-layers: upper crust, lower crust and sub-Moho, and mantle lithosphere. They suggest that the anisotropy of each layer reflects different deformation regimes such as surface faults, deep crustal flow and mantle compression. *Su et al.* (2008) concludes that strong azimuthal anisotropies reside within the upper 100 km of the crust and mantle. *Yao et al.* (2010) make a tighter constraint on the depth of the anisotropic structures by inverting the dispersion data for shear velocity. They determine azimuthal anisotropy in the uppermost crust, mid-crust, and in the uppermost mantle (80 and 110 km depth). They suggests that at 10 km depth the fast polarisation axes are near-parallel to the major strike slip fault systems and also resembles the pattern of surface motion from GPS. The anisotropy in the mid-crust has complex patterns and with an amplitude less than 3% , lower than our 1-D V_S estimates of $3\text{--}6\%$ (Figure 5.11).

The upper 250 km of the mantle, modelled from large-scale V_S tomography, has azimuthal anisotropy on the order of $\sim 2.5\%$ (*Griot et al.*, 1998; *Priestley et al.*, 2006). Our 1-D V_S models have upper mantle (<200 km depth) anisotropy in the range of $2\text{--}4\%$ with a SSW–NNE azimuth (Figure 5.12). The fast directions compare well with those inferred by tomography of *Priestley et al.* (2006) ($1\text{--}2\%$ at 200 km depth) but not with those of *Griot et al.* (1998). At shallower depth

(100 km) the tomographic models are characterised by strong complex patterns of anisotropy. These are likely an artifact of the strong shallower anisotropy, which, in our models, is resolved to be within the crust, distinct from the mantle anisotropic layer.

5.5.2 Shear-wave birefringence

Numerous studies have successfully detected the splitting of shear waves that traverse beneath Tibet (Figure 5.1). Despite the fast polarisation orientation of the measurements appearing to show a coherent clockwise rotation around the eastern Himalayan syntaxis, measurements of neighbouring stations show large splitting-time differences that are hard to understand. A series of stations in the south have null split time immediately next to stations that have splitting time >1 s. Similarly in northern Tibet, stations that exhibit delay times of over 2 s are juxtaposed to stations with a much lesser splitting time (e.g., *McNamara et al.*, 1994; *Huang et al.*, 2000; *Leon Soto et al.*, 2012). These measurements were made assuming a single anisotropic layer.

The delay time of a near-vertical propagating phase, in its simplest configuration of a single anisotropic layer, is dependent on the anisotropic strength, shear velocity, and thickness of the medium travelled in. Observations turn complex when phases traverse through more than one anisotropic layer. The splitting time as well as the polarisation of the fast wave becomes strongly back-azimuth dependent as illustrated in the review by *Savage* (1999). Hence, in order to determine if the observed birefringence reflects the splitting time of a single anisotropic layer or multiple layers, one requires numerous measurements from different back azimuths to detect periodic patterns of splitting parameters. This may be difficult to achieve because either earthquakes may not occur at specific back azimuths and/or short-term instrument deployments may not have enough time to record adequate seismicity.

Many of the shear-wave splitting measurements across Tibet are from temporary arrays and thus risk that most of the teleseismic sources are limited from the east, from the active subduction zones in the West Pacific. Studies on the permanent station in Lhasa (LSA, orange triangle in Figure 5.14) have previously reported to have not observed any splitting time and hence null fast directions

(*McNamara et al.*, 1994). Recently, because of its long operation, a detailed investigation using more phase arrivals provided from a wider back-azimuth spread revealed that LSA indeed has significant anisotropy (*Gao and Liu*, 2009). Results suggest that the phases travel through two anisotropic layers and that splitting times can be up to 1.5 seconds.

To overcome the lack of vertical constraints provided by shear-wave birefringence, inferences based on the near-vertical path traversed by the phases can help establish a depth range for the anisotropy. For example, because of the variation in splitting parameters across the plateau, the lower mantle is ruled out as a dominant anisotropic source (*McNamara et al.*, 1994). The crustal contribution to the anisotropy can be determined from the analysis of Moho converted *P*-to-*S* phases. Delay times in the range of 0.2–0.3 s have been reported for Tibet’s crust (*McNamara et al.*, 1994; *Herquel et al.*, 1995; *Chen et al.*, 2012). Although the crustal splitting time is considered small when compared to the longer lag time of deeper phases, it still has a significant contribution.

5.5.2.1 Shear-wave splitting with two anisotropic layers

Here we estimate splitting times for the two anisotropic layers determined in our 1-D models, for the crust and mantle. For each layer we use an approximate relationship between the anisotropic layer thickness L and shear velocity β to determine the shear phase split time δt :

$$\delta t = \frac{L\delta\beta}{\beta_o}$$

where β_o is the isotropic shear velocity of the layer, and $\delta\beta$ is the difference in *S* velocity (*McNamara et al.*, 1994). Figure 5.14 shows selected estimated splitting time for the crust and mantle. The time is calculated from the anisotropy and *S* velocity of the four crustal layers and from the Moho down to 200 km depth, for the crust and mantle respectively. The orientation of the arrows are inferred from the direct comparison of the dispersion curves and are only meant for a qualitative comparison (Figures 5.11 and 5.12). The predicted crustal splitting time varies between 0.2 and 0.5 s for Lhasa and Qiangtang, close to measured *Ps* delay times from the same regions (*McNamara et al.*, 1994; *Herquel et al.*, 1995). Predicted

values for Songpan-Ganzi tend to be higher. The predicted split time within the mantle is in the range of 0.5–1.2 seconds. The two time estimates cannot be added together to match the observed delay time at the surface as shear-wave splitting from more than one layer are subject to back azimuth periodic patterns (*Silver and Savage, 1994; Savage, 1999; Levin et al., 1999; Gao and Liu, 2009*). Likewise we do not expect our inferred fast directions to match directly those of shear-wave splitting.

Efforts to resolve for two or more anisotropic layers beneath Tibet from the shear-wave splitting parameters are few and these have been challenged by the limited back-azimuth coverage (e.g., *Lev et al., 2006; Wang et al., 2008*). Interestingly those with confident models show similar patterns for eastern Tibet — a fast SW–NE polarisation upper layer and a fast NW–SE polarisation in a lower layer (*Gao and Liu, 2009; Huang et al., 2011; Li et al., 2011*), a stark contrast to our models. The layers seem to switch in western Tibet with fast NW–SE direction in the upper layer and fast SW–NE direction in the lower layer (*Levin et al., 2008, 2012*).

In order to determine a qualitative approximate range of the expected splitting parameters from our azimuthal anisotropy, we compare our results to those of *Gao and Liu (2009)*. The azimuthal difference between our two anisotropic layers in East-central Tibet are close to the azimuthal configuration of *Gao and Liu (2009)*, but, the synthetic splitting parameters have to be rotated (shifted) by about $+70^\circ$ to match our upper and lower fast directions. The shift in the synthetic pattern is justified as the fast axes of each layer with respect to each other are not changed and hence the $\pi/2$ pattern is maintained. Earthquakes typically used for shear-wave splitting measurements in Tibet occur in the Tonga subduction zone at about 125°N back azimuth from the stations (e.g., *Chen et al., 2010*). The *Gao and Liu (2009)* synthetics suggest that at about 195° back azimuth ($70^\circ + 125^\circ$), the fast polarisation is at about 20° with a corresponding split time of 1 s. A change in the back azimuth of $\pm 15^\circ$ will result in fast directions that can alternate between 90° and 0° or in the range between 0° to 50° , and with a delay time between 0.5 to 2.0 seconds. These ranges for azimuth and split time resemble the various different shear birefringence for East-central Tibet (Figures 5.14). In another example, synthetic patterns of splitting parameters for a two-layer azimuthal anisotropy by

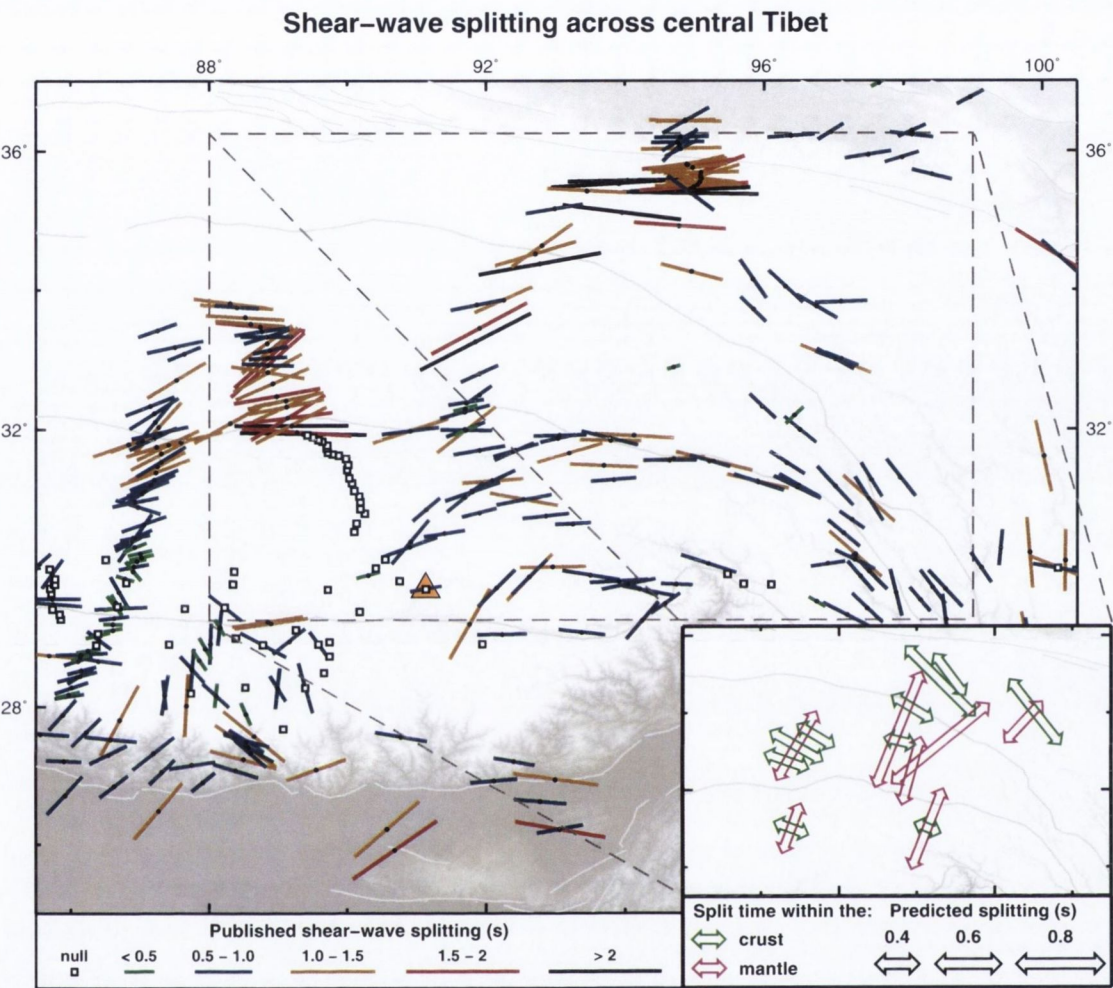


Figure 5.14: Published shear-wave splitting from teleseismic waves and predicted splitting from 1-D shear velocity models. Large map: A compilation of various *S*-wave splitting estimates across central Tibet as in Figure 5.1. Colour-coded length-varying bars indicate the split time and fast polarisation direction. Black closed circles are the location of seismic stations. Black open squares are stations with null measurements. Orange triangle is station Lhasa (LSA). Grey dashed rectangle is the location of the inset. Inset: Shear-wave splitting time predicted from selected 1-D *S*-velocity models for east-central Tibet. Green arrows represent splitting time calculated from crustal *S* velocity and anisotropy (Figure 5.9). Purple arrows represent splitting time calculated from upper mantle *S* velocities and anisotropy (<200 km depth, Figure 5.10). The arrows are plotted between the belonging paired inter-station paths and oriented mid-angle between the parallel and perpendicular direction of the respective fastest and slowest path.

Levin et al. (1999) also result in a delay time jump of ~ 1 s and large changes in the fast polarisation direction when shifted to represent our configuration.

5.5.3 Anisotropic interfaces

Seismic interfaces imaged by receiver functions indicate strong anisotropic surfaces within the crust. The azimuth-dependent amplitude of strong P -to- S conversions obtained for Songpan-Ganzi were modelled using one or two anisotropic layers (*Vergne et al.*, 2003). Highly deformed flyschs and schists were suggested to explain the large amplitude anisotropy of about 15% in the upper 25 km of the crust. An advantage of this technique is the parameterisation for the downward plunge in the direction of the fast direction. Hence the anisotropy can alternatively be explained, to some extent, by a dipping interface. Nonetheless, *Vergne et al.* (2003) conclude that the phases are best described by anisotropy. Similar strong anisotropy ($>10\%$) with a more detailed crustal structure has been modelled across the southern terranes (*Sherrington et al.*, 2004), at the BNS (*Ozacar and Zandt*, 2004), across the Himalaya (*Schulte-Pelkum et al.*, 2005), and in western Tibet using a joint application of teleseismic shear wave splitting with receiver functions (*Levin et al.*, 2008). Predicted birefringence in teleseismic shear waves from these anisotropic crustal layers contribute to about ~ 0.5 s in Songpan-Ganzi (*Vergne et al.*, 2003) and up to ~ 0.3 s in western Tibet (*Levin et al.*, 2008).

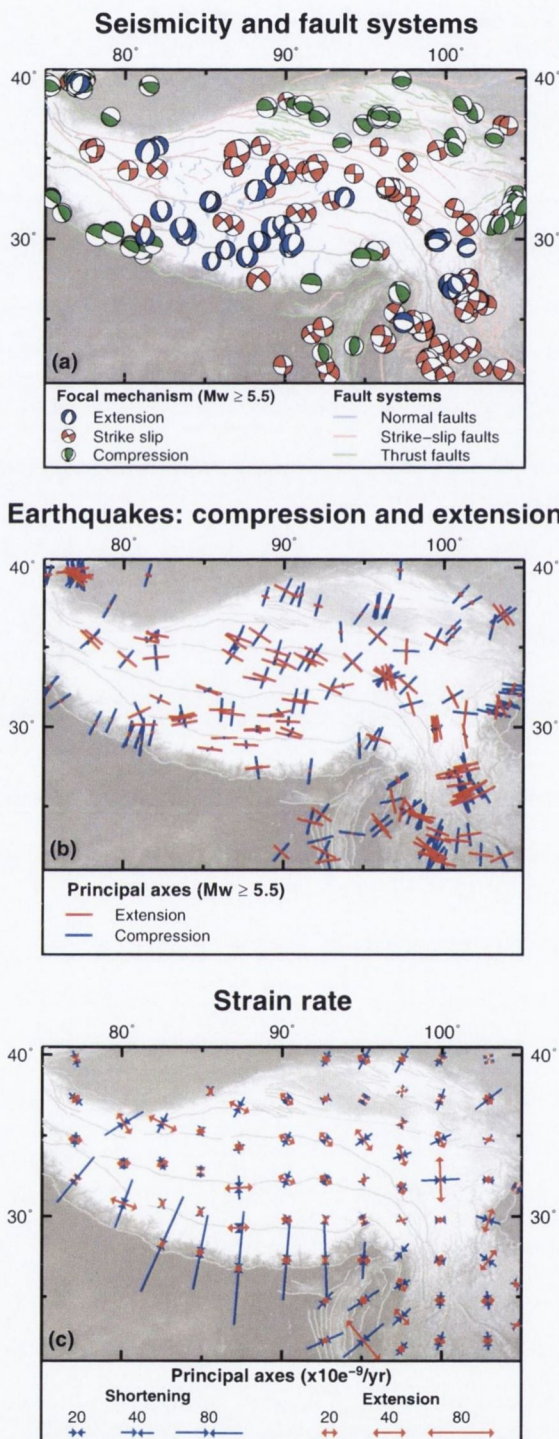
Many of the shear-velocity models derived from the receiver functions are very different from our 1-D models and also between each other. They either have high crustal S velocities or have very narrow LVZs, or both. Interfaces from within the shallow crust are highly sensitive to localised features (just below the station) and may not reflect the regional structures. Our anisotropic models have a smooth transition from west to east Tibet both in amplitude and direction and do not observe the detailed changes suggested by, for example, *Sherrington et al.* (2004). Hence we conclude that the very strong anisotropic pattern is either very shallow in the uppermost crust such that it is not detected by our short period data, or that the very large amplitude anisotropy are local features and are averaged-out in our inter-station measurements.

5.5.4 Tibetan lithospheric deformation

The present-day upper crustal deformation of the plateau is best observed by earthquake source mechanisms. Disregarding deep events related to India's subduction, earthquakes across Tibet and the surrounding regions occur only within the upper brittle crust (e.g., *Sloan et al.*, 2011). The different earthquake mechanisms reveal an interesting pattern of deformation; compressional earthquakes at the periphery of the plateau, and extensional and strike-slip earthquakes within the high-elevated interior of the plateau. The fault plane solutions of earthquakes match the thrust, normal and strike slip fault systems (e.g., *Molnar and Tapponnier*, 1978; *Taylor and Yin*, 2009) (Figure 5.15a). The contribution of extension and compression from earthquakes is shown in Figure 5.15b, where we plot the principal axes of the horizontal components from the earthquake moment tensors. The edges of the plateau are dominated by N–S compression, whereas extension is prevalent within the plateau.

In Figure 5.15c we plot the strain rate model from the Global Strain Rate Map (GSRM) by *Kreemer et al.* (2003). Because this model integrates geodetic and geologic data with seismic information, a similar but more robust pattern of deformation to the one from focal mechanisms is anticipated. The model shows strong variations in strain rates across the plateau; arc-normal contractions in the Himalayas (~ 100 nanostrain per year, long blue arrows) and W–E trending dilation characterises the central part of the plateau (~ 20 – 40 nanostrain per year, red arrows). Strong deformation is not only taking place in southern Tibet but also in eastern Tibet. Crustal stretching within the high plateau is largest in the east (~ 50 nanostrain per year) where the directions of the axes change from NW–SE in northern Tibet to N–S west of Sichuan (e.g., *England and Molnar*, 1997a,b; *Holt et al.*, 2000; *England and Molnar*, 2005; *Allmendinger et al.*, 2007). We note that the orientation of the extensional component of the strain is parallel to the traces of sutures from west to east of the plateau all the way to North Yunnan. The similarity of the present day strain with ancient marks of deformation suggest that the mechanism of deformation within the plateau has not changed since continental collision begun and is still governed by the northward push of India. In central Tibet the different terranes are being squashed against the more rigid north and experiencing E–W extension, whereas in the east the terranes bend round the

Figure 5.15: Tectonic regime of the Tibetan Plateau from fault mechanics, seismicity, and integrated strain rate models. (a) Different fault systems colour-coded according to their type (normal, thrust, and strike-slip) from *Taylor and Yin (2009)*. Focal mechanism are of shallow earthquakes (<50 km depth) and with moment magnitude ≥ 5.5 . The focal mechanisms are well-constrained double couple CMT solutions (<http://www.globalcmt.org>) colour-coded according to their mechanism (normal, compression, and strike-slip). Thick grey lines indicate sutures. (b) Principal axes of the horizontal components of the earthquake moment tensor, compression (blue sticks) and extension (red sticks), each normalized to the length of the largest axis. (c) Strain rate principal axes from the Global Strain Rate Map (GSRM) model (*Kreemer et al., 2003*). Blue, inward arrows denote compression (shortening). Red, outward arrows denote extension.



eastern syntaxis.

Since the strain rate models are inferred using upper crustal information, our models of azimuthal anisotropy add new constraints about the depth extent of the deformation. The orientation of the extensional component of the current strain field has a striking resemblance to the pattern inferred from the tomographic model at 25 seconds (Figure 5.5), indicating coherent deformation within the deeper crust. The inverted 1-D models show that the strength of anisotropy varies with depth and in different parts of the plateau (Figures 5.9 and 5.13). The strength of the anisotropy seems to be weaker in the deeper crust (anisotropy decreases towards 50 s period, Figure 5.6), however this might not necessarily indicate less deformation, but may be the result of absence of anisotropic minerals. The increase in pressure and temperature can make compositional changes to the hydrous minerals mica (e.g., *René et al.*, 2008) or amphibolites (e.g., *Spear*, 1981), both of which are prone to dehydration. Another reason for the lack of strong azimuthal anisotropy in the lower crust may be an artifact from the inversion; the shear velocity has large trade off with the Moho discontinuity and anisotropic structure. This makes it hard to resolve whether the anisotropy is confined only within the crust or extends further into the lithospheric mantle. Taking into consideration that the short-period dispersion has a coherent fast direction up to 50 seconds that is sensitive to depths >70 km (Figures 5.4, 5.5 and 5.6), we deem that the deformation pattern in the deep crust is similar to that observed in the mid crust. We concluded that anisotropy reflects W–E extension in central Tibet and NW–SE extension in eastern Tibet (Figure 5.17), and that the extension is likely to be of a lithospheric scale with depth-dependent strengths.

5.5.5 Implications for mid-crustal channel flow

Several studies have shown that the Tibetan crust is underlain by a weak mid-crustal layer characterised by low shear velocities (e.g., *Kind et al.*, 1996; *Yang et al.*, 2012), high conductivity (e.g., *Chen et al.*, 1996; *Wei et al.*, 2001) and high temperatures (e.g., *Hacker et al.*, 2000; *Mechie et al.*, 2004). The combination of these geophysical properties strongly suggests that the mid-crust is partially molten and is prone to flow (*Nelson et al.*, 1996). Estimates for the viscosity of the mid-crust beneath Tibet indicate that it can be of an order of magnitude lower than

adjacent rocks (e.g., *Beaumont et al.*, 2001; *Unsworth et al.*, 2005; *Caldwell et al.*, 2009), enough to trigger flow if under high gravitational potential (e.g., *Clark and Royden*, 2000; *Rippe and Unsworth*, 2010; *Royden*, 1997). The wide extent of the LVZ has been recently revealed by a high resolution tomographic model inferred by Rayleigh-wave dispersion (*Yang et al.*, 2012). As the synthesis of a weak viscous layer is becoming increasingly accepted, the role such a layer has on the dynamics of the plateau has left researchers divided between two end-member views: depth-dependent, decoupled lithospheric deformation, or vertically coherent lithospheric deformation. In the latter, surface deformation from GPS and Quaternary fault slip rate data are matched with the mantle strain inferred from *SKS* shear wave splitting (typically assuming a single mantle anisotropic layer) and suggest that the lithosphere deforms coherently (e.g., *Flesch et al.*, 2005; *Wang et al.*, 2008). In the former, the lithospheric vertical strength is interrupted by a weakened mid-lower crust resulting in decoupling the upper crust deformation from the mantle (e.g., *Royden*, 1997). Our models show varying anisotropic strength with depth that are likely to represent elements of the two end-member deformation models.

Large-scale lower crustal flow induced by lateral pressure gradients has long been inferred for the expansion of the plateau (*Bird*, 1991). *Clark and Royden* (2000) propose that a viscous lower crust oozes around the strong rigid boundaries of Tarim and Sichuan creating gentle margins at the eastern regions of the plateau. Such a model has been used to explain the uplift in southeastern Tibet (*Royden et al.*, 2008). Likewise, one would expect similar deformation in regions with alike morphology such as that in Yunnan — specifically in northern Tibet where, in an elevation-induced flow model, the lower crustal flow is in the NE direction (*Clark and Royden*, 2000; *Klemperer*, 2006; *DeCelles et al.*, 2002). We do not observe any anisotropic pattern pointing towards the NE at lithospheric depths (Figure 5.5, 5.6, and 5.11), and we do not find any correlation between topography and anisotropic azimuth or strength. The rapid southeast flow adopted by *Royden et al.* (2008) is facilitated by trench rollback in the western Pacific and Indonesia, a tectonic feature that is non-existent in north eastern Tibet. The pattern of the azimuthal anisotropy reflects the surface extension suggesting that the vertical distribution of the strain is under the same influence (Figure 5.17). Indeed, many of our 1-D models have a mid-crustal low shear-velocity zone that is characterised by

strong anisotropy. Thus we conclude that the mid-crust is ductile and experiences a more rapid deformation (extension) than the brittle upper crust, but, because of the coherent vertical deformation (extension in the same direction), it is likely that the mid-crust still holds some degree of coupling with the adjacent layers. We envisage that large-scale outward flow does not take place north of Kunlun Fault, but is retained within the high plateau. Furthermore the spatial distribution of the azimuthal pattern and strain across the vast terranes suggest that the deformation is over broad areas beneath the plateau and not localised at the edges of deformation zones (e.g., *Avouac and Tapponnier, 1993*).

5.5.6 Tibetan asthenospheric deformation

Shallow mantle azimuthal anisotropy is detected in our measurements and has been inferred to reside within a LVZ between 100–200 km depth (Figure 5.10). The LVZ is the Tibetan asthenosphere underlain by a high shear velocity (e.g., *Zhao et al., 2011*, Section 3.3.2). The SSW–NNE fast directions derived beneath central Tibet are parallel to the inferred direction of India’s plate motion with respect to a stable Eurasian plate from GPS observations (N20°E–N21°E, *Sella et al., 2002*; *Wang et al., 2001*). The anisotropy is weak beneath south central Lhasa and increases towards the north (Figure 5.12). We are inclined to explain the asthenospheric anisotropy as induced by the subducting Indian slab, sliding beneath the Tibetan asthenosphere, and the ongoing N–S shortening of the plateau. In contrast to the conclusions of *Griot et al. (1998)* that mantle anisotropy is strongly coupled with surface motions, we favour arguments by *Priestley et al. (2006)* that the azimuthal anisotropy from strained olivine crystals undergoing simple shear at the base of a plate follows the principal extension direction and aligns with the flow direction (*Zhang and Karato, 1995*).

In Figure 5.17 we show 3-D cross sections for central and northeastern Tibet. The cartoons (not to scale) are derived from all the results presented here and highlight two anisotropic layers. The upper layer resides in the crust and correlates with the extensional component direction of the strain and also with ancient sutures. The lower anisotropic layer is within the Tibetan asthenosphere and is parallel to the direction of Indian plate motion with respect to a stable Eurasia. Overlain is a panel showing *SKS* measurements, processed assuming a single anisotropic layer.

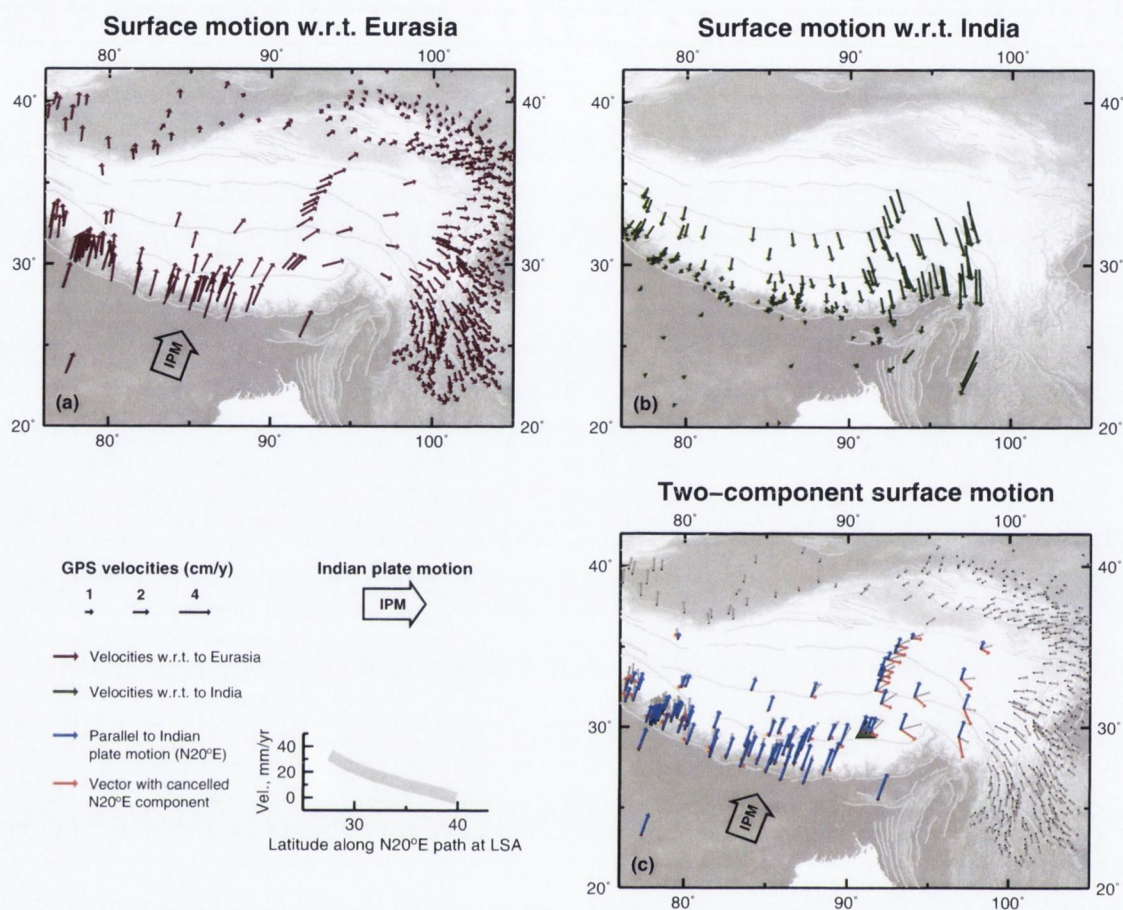


Figure 5.16: Different perspectives of surface deformation across Tibet and surrounding regions from Global Positioning System (GPS) measurements. (a) GPS velocity vectors with respect to stable Eurasia from *Zhang et al.* (2004). IPM indicates India's plate motion ($\sim$ N20°E). (b) GPS velocities vectors with respect to stable Indian plate from *Banerjee et al.* (2008). (c) Decomposition of GPS velocity vectors into two components with respect to stable Eurasia. Blue arrows: Predicted velocity parallel to $\sim$ N20°E, the assumed direction of the Indian-Eurasian collision (*Zhang et al.*, 2004). The velocity at each respective station is determined from the velocity profile along the N20°E direction over passing station Lhasa (LSA, green triangle), using a smooth fit to the data in *Zhang et al.* (2004) (grey curve on the left). Red arrows: The resultant vector following the subtracted N20°E velocity component from the original GPS vectors at each station.

5.5.7 Perspectives from Global Positioning System measurements

The GPS measurements from across Tibet are strongly dominated by relative motions of adjacent tectonics, particularly India. Whilst India's motion towards Eurasia is thought to be a simple northward motion of the entire Indian plate, at least up to the Himalayas, the smooth changing velocity vectors both in amplitude and direction across Asia makes it unclear what is the contribution of India beyond the Himalayas and what is the deformation from strictly within the plateau (*Zhang et al.*, 2004) (Figure 5.16a). Towards northeastern Tibet, the north component of the vectors decreases, vectors change direction to point towards the east, and eventually point to the south in south-eastern Tibet. GPS vectors in west Tibet, remain predominantly pointing northward with a slight inclination towards the west. At the northern edge of the plateau the vectors have a very small magnitude when compared to those in the south suggesting strong absorption being taken by the plateau. Figure 5.16b shows the GPS velocity with respect to a stable India reference frame using the data set from *Banerjee et al.* (2008). The arrows in west and central Tibet point towards India and their amplitudes decrease farther south towards the Himalaya. In contrast, the vectors at the eastern syntaxis point towards the SSE and increase in amplitude towards the south indicating the presence of a strong southern component. GPS vectors in a Tibetan Plateau fixed reference frame show west and south Tibet rotating anti-clockwise, with strong velocities towards south-east in eastern Tibet (*Gan et al.*, 2007). Here we take a different view. In order to investigate the plateau's internal deformation without the influence from India, we decompose the GPS velocity vectors relative to Eurasia by subtracting the predicted northward Indian motion ($\sim N20^\circ E$, *Zhang et al.*, 2004). Hence the resultant vector is assumed to represent the plateau's internal contribution (Figure 5.16c, red arrows).

The GPS model we present here assumes that the vectors across the high plateau are composed of two components, Indian and Tibetan (Figure 5.16c). The Tibetan vectors divert between westerly pointing or easterly pointing at around $\sim 85^\circ E$ in the Himalaya, similar to the patterns inferred by *Styron et al.* (2011). Of interest are the Tibetan vectors in eastern Tibet which have directions remark-

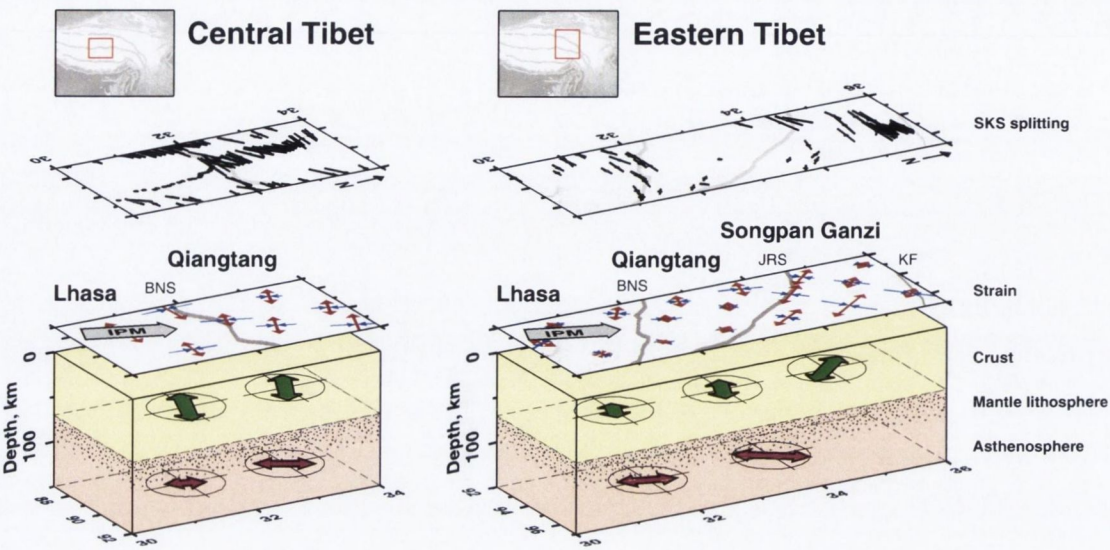


Figure 5.17: Three-dimensional view of the deformation beneath central and eastern Tibet. Not to scale. Red rectangles in the small maps indicate the location of the 3-D cross sections. Thick arrows show azimuthal anisotropy in the Tibetan crust (green) and asthenosphere (purple), averaged from the shear-velocity anisotropy in Figures 5.11 and 5.12, respectively, plotted in the shown projection. The panel on top of the crust shows the surface terranes separated by sutures and faults (grey lines). Superimposed on the topography is the strain rate principal axes from the Global Strain Rate Map (GSRM) model (*Kreemer et al.*, 2003). Blue, inward arrows denote compression (shortening). Red, outward arrows denote extension. IPM: Indian plate motion. Top frame: Published shear-wave splitting from various works, same as in Figure 5.1.

ably aligned parallel to the sutures and faults. Contrary to the impression of an outward, east directed flow in northeastern Tibet and a southward directed flow in south-eastern Tibet as in Figure 5.16a (arrows start pointing south at about $\sim 30^\circ\text{N}$ west of Sichuan), the Tibetan vectors shown in Figure 5.16c suggest that the southerly pointing vectors are located further north up to the Kunlun Fault. The smooth clockwise rotation of the Tibetan vectors and their parallel direction to active faults in the region suggests that the terranes are ‘bending’ around the eastern syntaxis as India keeps on advancing northward.

5.6 Conclusions

We measure robust, broad-band Rayleigh-wave phase-velocities between pairs of stations located across the Tibetan Plateau. The data set is explored rigorously using different techniques, including 2-D tomography, regional-average phase velocities, and also direct dispersion comparison. In central Tibet we identify two period bands that have distinct azimuthal anisotropy patterns. Inferred 1-D shear-velocity models show strongly anisotropic layers in the Tibetan mid-lower crust and asthenosphere.

The crustal azimuthal anisotropy is W–E oriented in central Tibet and NW–SE oriented in eastern Tibet. The azimuths are parallel to the extensional component of the current strain rate field across Tibet, strongly suggesting similar deformation from the surface down to the deep crust. The 1-D shear-velocity models show varying azimuthal anisotropy with depth. In central Tibet anisotropy is particularly strong within a mid-crustal LVZ (about 5% in the ~20–45 km depth range) that coincides with the depth of high conductivity and partial melt (*Nelson et al.*, 1996). Despite the mid-crust’s greater susceptibility to deformation and flow, the correlation of the azimuthal anisotropy orientation with surface strain indicates that the mid-crust still holds some degree of coupling with the adjacent layers (Figure 5.17). The close agreement of anisotropy and extension with the traces of sutures implies that the dominant deformation mechanism within the plateau has not changed since initiation of continental collision and is still governed by the northward push of India.

The crustal flow direction inferred from the azimuthal anisotropy does not indicate any SW–NE flow as in the channel-flow model for northern Tibet (*Clark and Royden*, 2000) (Figure 5.17). Furthermore, we do not observe any correlation between topography and azimuthal anisotropy. We also note that the azimuthal anisotropy pattern is coherent across different terranes, its amplitude strongest in northeastern Tibet (where shear velocities in the crust are particularly low, suggesting partial melting and low viscosities, Chapter 4). These results show that the deep crustal deformation is diffuse (e.g., *England and Houseman*, 1986; *Zhao and Morgan*, 1987), and that flow, rather than taking place at localised channels (*Bai et al.*, 2010), is likely to be distributed across broad areas.

Azimuthal anisotropy with a SSW–NNE fast-propagation direction has been detected within the asthenosphere beneath central and eastern Tibet (100–200 km depth). The match of the anisotropic azimuth with the direction of India’s plate motion leads us to conclude that asthenospheric flow is driven by India’s northward subduction.

The two-layer anisotropic structure beneath Tibet (Figure 5.17) can also explain qualitatively the complexity and variability of published shear-wave splitting measurements.

6

Conclusions

This dissertation consists of three major seismic investigations about Tibet, aimed to (1) establish the mechanism of lithospheric convergence between India and Eurasia, (2) determine the seismic-velocity structure and physical and rheological properties of the crust, and (3) investigate the dynamics of the Tibetan lithosphere and asthenosphere. The following is a summary of the work and conclusions.

The data from selected station pairs across the plateau were used to determine the fundamental-mode dispersion of Love and Rayleigh waves in the period range of 5–200 seconds. All data were manually selected in order to achieve the highest quality phase velocities. The data has undergone a series of rigorous analyses: (1) direct comparison of inter-station dispersion, (2) inversion for 1-D radially anisotropic *S*-velocity models that simultaneously fit the Love- and Rayleigh-wave dispersion of an inter-station pair, (3) inversion for 1-D azimuthally anisotropic *S*-velocity models that simultaneously fit both Rayleigh-wave dispersion of two inter-station paths, (4) modeling of regional averages, and (5) tomographic models. All of these techniques yielded consistent results regarding shear velocity structure

and anisotropic properties of the crust and upper mantle beneath Tibet.

The Tibetan crust has, as expected, low shear velocities, however, there are strong variations across the plateau. Mid-crustal LVZ are only found in northern Tibet, north of the BNS, as the S velocities decrease northward, towards the Kunlun Fault (Figure 4.11). High melt fractions ($<6\%$) and low viscosity are thus likely in the middle-lower crust beneath West Qiantang and Songpan-Ganzi. The crust of the Lhasa Terrane is unlikely to have any significant melt, as the low shear velocities within it can be explained by relatively high temperatures alone. Furthermore, the distribution of the LVZ coincides with the top of a high-conductivity layer, supporting the argument for the occurrence of a partially molten layer within the middle-lower crust. The similarity of inter-station dispersion curves within whole regions, such as in eastern Songpan-Ganzi, implies nearly homogeneous structure within them; therefore, the inference of a weak, low viscosity layer, prone to deform, applies to the entire regions. This is further substantiated by the coherent radial and azimuthal anisotropy across the respective areas. Thus, on a regional scale, the geophysical properties are consistent with models of diffuse internal deformation (e.g., *England and Houseman*, 1986), particularly in northern Tibet.

Deep crustal deformation inferred from radial anisotropy does not correlate with low seismic velocities (and, hence, partial melts and low viscosities) or elevation, but instead reflects the deformation pattern at the surface, evidenced by focal mechanisms of earthquakes (Figure 4.12). Strong radial anisotropy is observed beneath western Tibet and Yunnan, both regions experiencing extension (flattening), whereas north-eastern Tibet has very weak radial anisotropy, strike-slip fault mechanisms, and no extension. The ongoing crustal thinning in the west probably causes the anisotropic mica crystals to become near-horizontally oriented (*Shapiro et al.*, 2004), whereas in the east, NW-SE shear may align micas in the vertical plane, resulting in weak or absent radial anisotropy. The flow direction is derived from crustal azimuthal anisotropy, with W-E oriented fast directions in central Tibet, and NW-SE fast directions in eastern Tibet. Special focus is given to northeastern Tibet, where the inferred fast directions are aligned southeast rather than northeast, as would be expected from an elevation-gradient induced flow (*Clark and Royden*, 2000). The fast azimuths are parallel to the extensional component of the current strain rate across Tibet, strongly suggesting

similar deformation through the entire crust. Despite the mid-crust's greater susceptibility to deformation and flow, the correlation of azimuthal anisotropy with surface strain indicates that the mid-crust still holds some degree of coupling with the adjacent layers (Figure 5.17). The close agreement of anisotropy and extension with the traces of sutures also implies that the dominant deformation mechanism within the plateau has not changed since initiation of continental collision and is still governed by the northward push of India. The intense compression on the Himalayan arc from the continued insertion of cold Indian crust into Tibet produces the normal faults across southern and central Tibet (*Kapp and Guynn, 2004*), and may also be responsible for the higher shear velocities and colder temperatures in southern Tibet than those found across northern Tibet.

The upper 75 km of the mantle beneath Tibet is made up of an Indian lithosphere in the west and southwest and a Tibetan lithosphere and asthenosphere elsewhere. Strong, cold, cratonic Indian lithosphere underthrusts southwestern Tibet (up to the BNS at 85°E), and warm Tibetan lithosphere and asthenosphere lay further north, up to the Kunlun Fault (Figure 3.12). The Tibetan lithosphere and asthenosphere have low-average S velocities, indicative of warmer temperatures. Although the finer structure of the Tibetan lithospheric mantle remains hard to resolve using surface waves alone, a thick layer of low-average V_S in the uppermost mantle is difficult to explain from high temperatures generated by crustal radioactivity and reconcile with the presence or formation of a thick cratonic lithosphere at depths down to 200 km (*McKenzie and Priestley, 2008*). Surface-wave data can fit a series of other seismic observations such as S_n , P_n , and a shallow LAB discontinuity, which together add support for a thin Tibetan lithosphere underlain by an asthenosphere. The dynamics of the asthenosphere beneath central Tibet is revealed by azimuthal anisotropy, characterised by fast SSW–NNE direction (Figure 5.17). The amplitude of the anisotropy increases from south to north and is parallel to the direction of India's plate motion, suggesting that asthenospheric flow is pushed outward by India's northward subduction.

The Indian lithosphere subducts beneath the Tibetan asthenosphere under the central and eastern plateau (Figures 3.12). The thickness and shear speed of the deep high-velocity anomaly beneath central Tibet is similar to those found within the Indian lithosphere underthrusting West Lhasa. Petro-physical temperature es-

timates for the deep, high S velocity are consistent with those that would occur within subducted Indian lithosphere. Hence, from the 1-D models, the lithospheric convergence mechanisms varies from west to east; steep-angle subduction of India beneath west-central Tibet and shallow-angle subduction of India in eastern Tibet, with the subducting Indian lithosphere reaching as far north as northern Qiangtang–Songpan–Ganzi Terrane (Figure 3.13).

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*Oh Dublin, cold Dublin,
I shall miss you.*