

Upper mantle beneath foothills of the western Himalaya: subducted lithospheric slab or a keel of the Indian shield?

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SUMMARY

The fate of the mantle lithosphere of the Indian Plate in the India–Eurasia collision zone is not well understood. Tomographic studies reveal high P velocity in the uppermost mantle to the south of the western Himalaya, and these high velocities are sometimes interpreted as an image of subducting Indian lithosphere. We suggest that these high velocities are unrelated to the ongoing subduction but correspond to a near-horizontal mantle keel of the Indian shield. In the south of the Indian shield upper-mantle velocities are anomalously low, and relatively high velocities may signify a recovery of the normal shield structure in the north. Our analysis is based on the recordings of seismograph station NIL in the foothills of the western Himalaya. The T component of the P receiver functions is weak relative to the Q component, which is indicative of a subhorizontally layered structure. Joint inversion of the P and S receiver functions favours high uppermost mantle velocities, typical of the lithosphere of Archean cratons. The arrival of the P_s converted phase from 410 km discontinuity at NIL is 2.2 s earlier than in IASP91 global model. This can be an effect of remnants of Tethys subduction in the mantle transition zone and of high velocities in the keel of the Indian shield. Joint inversion of SKS particle motions and P receiver functions reveals a change in the fast direction of seismic azimuthal anisotropy from 60° at 80–160 km depths to 150° at 160–220 km. The fast direction in the lower layer is parallel to the trend of the Himalaya. The change of deformation regimes at a depth of 160 km suggests that this is the base of the lithosphere of the Indian shield. A similar boundary was found with similar techniques in central Europe and the Tien Shan region, but the base of the lithosphere in these regions is relatively shallow, in agreement with the higher upper-mantle temperatures. The ongoing continental collision is expressed in crustal structure: the crust beneath NIL is very thick (58 ± 2 km), and the S velocity in the intermediate and lower crust is around 4.0 km s^{-1} . This anomalously large velocity and thickness can be explained by scraping off the lower crust, when the Indian lithosphere underthrusts the Himalaya.

Key words: asthenosphere, body waves, collision belt, crustal structure, lithosphere, seismic anisotropy.

1 INTRODUCTION

Various data suggest that the Himalayan belt was created and is being created mainly by underthrusting of the crust of the Indian Plate, accompanied by some form of absorption of the Indian mantle lithosphere. Since the onset of continental collision about 50 Ma, crustal shortening of the Himalayan belt has absorbed several hundreds of kilometres of Indian lithosphere (e.g. Coward *et al.* 1987; Searle *et al.* 1988; Shelling & Arita 1991; DeCelles *et al.* 2002).

The fate of the absorbed mantle lithosphere of the Indian Plate is unclear. Seismic body wave tomography is usually efficient in tracing subducting mantle lithosphere, but the results for the Indian lithosphere are rather unclear. In particular, Zhou & Murphy (2005) interpreted their tomographic model as evidence of flat subduction,

resembling that proposed by Argand (1924), whereas Replumaz *et al.* (2004) explained their image in terms of steeply plunging Indian lithosphere, similar to that, proposed by Mattauer (1986). The problems might arise from a low accuracy of the input traveltime data or/and the complexity of the actual deep structure. Non-unique interpretation of the available seismic data is reflected in the physical models, which demonstrate either horizontal (Chemenda *et al.* 2000) or dipping (Touissaint *et al.* 2004) continental lithosphere in the India–Asia collision zone.

A common feature in the latest tomographic models (Replumaz *et al.* 2004; Zhou & Murphy 2005; Li *et al.* 2006) is a high P velocity in the uppermost (50–250 km depths) mantle in a zone a few hundred kilometres wide to the south of the western Himalaya. These high velocities were interpreted as an image of plunging mantle

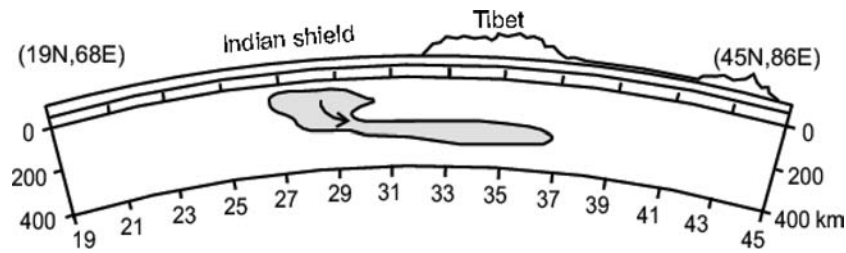


Figure 1. Contour of the high-velocity body in the upper mantle; modified from Zhou & Murphy (2005). Topographic profile is shown with a 20:1 exaggeration.

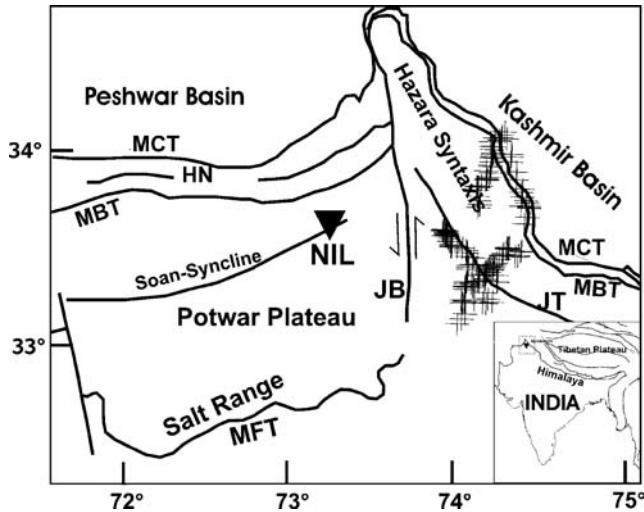


Figure 2. Map of the study region. Inset shows its position within a larger region. Tectonic features (DiPietro & Pogue 2004): Main Frontal thrust (MFT), Main Boundary Thrust (MBT), Main Central Thrust (MCT). Hissartang-Cherat-Nathia Gali Thrust (HN), Jhelum-Balakot Fault (JB), Jammu Thrust (JT). Projections of Ps piercing points at a depth of 410 km for the receiver functions in Fig. 3 are shown by crosses.

lithosphere (Fig. 1). Here we present a detailed analysis of recordings of the IRIS station NIL which is located close to this zone and provides independent data on its deep structure. The importance of this station is determined not only by its location, but also by the fact that it is the only long term digital broad-band seismograph station in a large region. Long term operation is a pre-requisite for our analysis of anisotropy, because it requires observations of distant seismic events in a broad range of azimuths, whereas short term observations in central Asia produce too few recordings for some directions.

Station NIL (Fig. 2) is located to the south of the Main Boundary Thrust (MBT) of the Himalaya. This fault is steep near the surface, but flattens at depth (Molnar 1990; DiPietro & Pogue 2004). Its gently dipping part is still active, but the shallow deformations occur now at the Salt Range Thrust, alternatively mentioned as the Main Frontal Thrust. The overall rate of convergence between the northern and southern boundaries of the Potwar plateau, where NIL is located, is thought to be around 10 mm yr^{-1} , and the sedimentary cover is detached from the underlying basement of the Indian shield (Molnar 1990). The Himalayan belt, which has a general NW–SE trend, in the region of NIL takes a sharp bend, which is known as Hazara Syntaxis.

The main results of our work are presented in Sections 2, 3 and discussed in Section 4. In Section 2, we present *P* and *S* receiver functions and a joint inversion of them for the isotropic velocity model. *P* receiver functions are widely used in seismology for imag-

ing structures in the crust and the upper mantle. This technique relies mainly on the transmitted Ps phases that are converted from *P* to *S* at seismic discontinuities in the vicinity of the seismograph station. Interference between multiple reflections from shallow boundaries and Ps phases from deeper boundaries is a major shortcoming of this technique. In particular, multiple reflections from the base of the sediments beneath NIL may severely distort the record of the Ps phase from the Moho. *S* receiver function technique (Farra & Vinnik 2000) which relies on the Sp (*S*-to-*P*) converted phases is free from this shortcoming, because Sp phases from deep discontinuities arrive much earlier than the multiples from shallow boundaries. *P* and *S* receiver functions are complementary: non-uniqueness of independent inversion of either *P* or *S* receiver function for the velocity model is minimized by a joint inversion of them (Vinnik *et al.* 2004) and can be further reduced by constraining the solution with teleseismic *P* and *S* traveltime residuals (Vinnik *et al.* 2006). The isotropic model thus obtained is further used in Section 3.

Section 3 contains our analysis of seismic azimuthal anisotropy in the crust and the upper mantle beneath and in the vicinity of NIL. Observations of seismic anisotropy provide a unique opportunity to study rock fabric and the related rock flow at large depths (e.g. Nicolas & Christensen 1987; Savage 1999). Analysis of shear wave splitting in SKS and similar seismic phases, a standard tool for mapping azimuthal anisotropy, was previously applied to the recordings of seismograph stations in and in the vicinity of the western Himalaya (Sandvol *et al.* 1994; Herquel and Tapponnier 2005) and elsewhere. Implicit in these techniques is the assumption of a single anisotropic layer, which is generally hard to justify. In principle, SKS particle motion can be inverted for a two-layer anisotropic structure (e.g. Menke & Levin 2003), but these methods require high-quality recordings of SKS in many azimuths, whereas in India and neighbouring regions the available recordings are mostly from one very narrow azimuthal sector. Even in the case of one-layer anisotropy, SKS techniques provide no constraints on the depth of the anisotropic layer, which undermines scientific significance of these data.

Here we apply an upgraded SKS technique that can resolve anisotropic parameters of a few layers (Vinnik *et al.* 2007). The idea of the method is to combine observations of shear wave splitting in SKS and *P* receiver functions. This approach was developed in several studies (Kosarev *et al.* 1984; Vinnik & Montagner 1996; Girardin & Farra 1998; Farra & Vinnik 2000; Vinnik *et al.* 2002). Other examples of application of *P* receiver functions to the problems of seismic anisotropy can be found elsewhere (e.g. Bostock 1997; Levin & Park 1997).

2 JOINT INVERSION OF *P* AND *S* RECEIVER FUNCTIONS

A construction of accurate *P* and *S* receiver function requires several tens of teleseismic recordings in the same azimuthal sector.

Therefore, our analysis is conducted in the azimuthal sector centred near 100° , where the number of recordings is sufficiently large.

A gradational layer generates the transmitted Ps phase only if the wavelength is not smaller than the thickness of the layer. For the reflected phases a similar relation holds for the quarter wavelength. Optimum results for the crust can be obtained at the periods of 4–8 s (S -wave wavelength around 25 km). To provide high sensitivity of the P receiver function to gradational layers several kilometres thick and suppress effects of wave scattering at small scale random heterogeneities, the raw records were low-pass filtered with the corner at around 5 s.

P receiver functions are obtained following Vinnik (1977). The record is presented by particle motion projections on the axes L , Q and T . Axis L is parallel to the principal P -wave particle motion direction in the wave propagation plane. The angle i_P between the L axis and the vertical direction is determined by analysis of the related covariance matrix. Axis Q is normal to L in the same plane and is nearly optimum for observing Ps phases in an isotropic Earth. The axes L , Q and T (transverse) form a right-hand coordinate system. To equalize recordings of different events, all components of each recording are deconvolved by the L component. The deconvolution is performed in time domain with a proper regularization. It transforms the L component of each recording into standard bell-shaped pulse. Similar signals in the other components are transformed by the deconvolution into standard form as well. To detect the weak Ps phases from deep discontinuities, the individual Q components are stacked with moveout time corrections. Like in other similar studies the reference slowness (with zero move-out corrections for any conversion depth) is fixed at 6.4 s deg^{-1} . The result is presented as a set of traces, corresponding to different trial conversion depths. The depth of the discontinuity can be found accurately from the arrival time of the Ps phase, or it can be evaluated crudely as the trial depth, at which the signal amplitude attains its maximum. If both estimates are mutually consistent, interpretation of a seismic arrival as the Ps phase is correct.

The stack of the P receiver functions (Fig. 3a) is obtained by summing 162 individual receiver functions in a distance range between 42° and 86° in the backazimuth sector between 70° and 127° . Within the first 20 s it contains a wave train with the largest amplitudes in the traces corresponding to the smallest conversion depths. This wave train is a signature of crustal structure. For the sign convention adopted here positive polarity of the Ps phase corresponds to positive discontinuity (with a higher S velocity on the lower side). A prominent phase with positive polarity and the largest amplitudes at a trial depth of 300–500 km arrives at a time of 41.8 s. This is the Ps phase from the global 410 km discontinuity, and it arrives 2.2 s earlier than predicted by IASP91 global model (Kennett & Engdahl 1991). The piercing points of this phase at a depth of 410 km are beneath the Himalaya (Fig. 2). A converted phase from another prominent mantle discontinuity at a depth around 660 km is too weak for a robust detection.

Calculation of the S receiver function (Farra & Vinnik 2000) involves low-pass record filtering and record decomposition into Q , L , T and M components. Q corresponds to the principal S particle motion component in the wave propagation plane. The angle i_{SV} between the principal direction of the S -wave motion in the wave propagation plane and the radial direction is determined by analysis of the related covariance matrix. L is perpendicular to Q in the same plane and is nearly optimum for detecting the Sp converted phases. T is perpendicular to the wave propagation plane, and M is the principal S particle motion component in the plane containing the Q and T components. The angle θ between the axes Q and M is con-

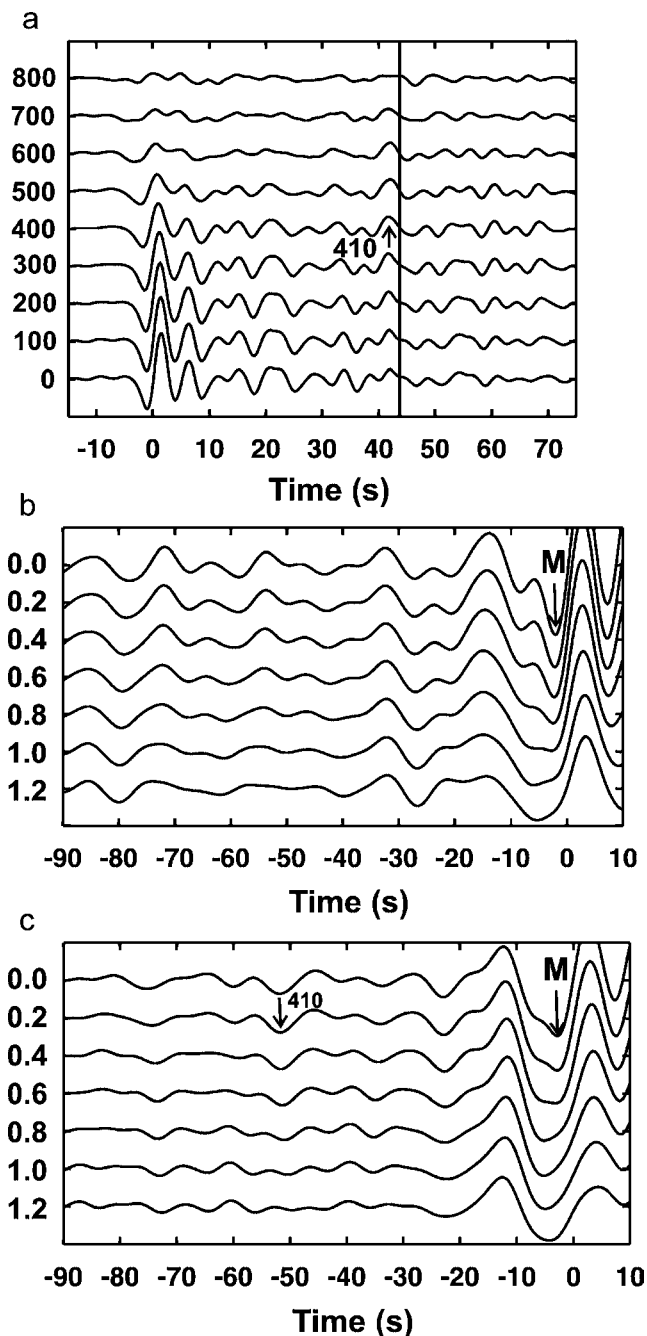


Figure 3. Stacked P (a) and S (b and c) receiver functions. The stacks of S receiver functions are obtained for epicentral distances from 65° to 75° (b) and 75° to 90° (c). Each trace in (a) corresponds to a trial conversion depth in km shown on the left. Vertical line at 44 s indicates the P410s arrival time for IASP91 (Kennett & Engdahl 1991). Each trace in (b) and (c) corresponds to the differential slowness in s deg^{-1} shown on the left. The arrows labeled 410 are for P410s (a) and S410p phases (c). The arrow labeled M in (b) and (c) corresponds to the Sp phase produced by a few crustal discontinuities.

trolled by the focal mechanism of the earthquake. All components are deconvolved by the M component.

The Sp converted phases can be generated by the SV and SH (in the case of anisotropy) components of the incoming S wave. To isolate the Sp phases generated by SV, the individual deconvolved L components of many events are stacked with weights depending on θ and level of noise. The weights are obtained as the solution

of a linear system of equations; the number of equations is determined by the number of recordings (Farra & Vinnik 2000). The resulting L component is normalized to the Q component. The procedure involves evaluation of σ , the rms value of noise in the stack. To account for the difference in slowness between the S and Sp phases from deep discontinuities, the individual receiver functions are stacked with time corrections (slant stacking). The correction is calculated as the product of the trial differential slowness and the differential epicentral distance (difference between the epicentral distance of the seismic event and the average epicentral distance of the group of events). The stack is calculated for a number of values of the differential slowness.

S receiver function technique is most efficient in the distance range between 65° and 90° . In the distance range between 75° and 90° SKS seismic phase arrives within 30 s of S , and at around 83° they even arrive at the same time. How the potential interference between S and SKS affects performance of the S receiver function technique? The Sp and SKS converted phases from discontinuities at depths less than 150 km arrive with a negligible difference in time (small relative to the dominant period of 10 s), but at larger depth the difference becomes significant, around 10 s for a depth of 400 km. 410 km discontinuity is global, and the converted phases from this discontinuity are of great diagnostic value. Numerical simulations reveal three possible kinds of interaction between S and SKS in S receiver functions: (1) contribution of SKS to the wave train used is small relative to S ; (2) contribution of S is small relative to SKS; (3) contributions of S and SKS are comparable. In the first case the detected signal is S410p, in the second case this is SKS410p. In the third case no signal is detected, because the waveform used for deconvolution is strongly different from both S and SKS.

To judge on the possible S /SKS interference problem we divided the set of the available individual S receiver functions in the distance range between 65° and 90° into two subsets: from 65° to 75° and from 75° to 90° . The number of receiver functions in each subset is nearly 60, the average backazimuth is close to 100° , the average distances are 68° and 83° , the rms values of noise are 0.008 and 0.007. For comparison, peak amplitudes of the L component in the interval from 0 s to -20 s in Figs 3(b and c) are around -0.1 and 0.05 . For the adopted sign convention negative polarity of the Sp converted phase implies positive discontinuity. The negative polarity phase with the largest amplitude in the 0 s deg^{-1} trace at a time near -3 s, marked M , is a combined effect of the Moho and discontinuities within the crust. The positive polarity phase at a time near -14 s is mainly a side lobe of the phase marked M .

In the distance range between 65° and 75° the slowness of S (around 12 s deg^{-1}) is too large for generating S410p, and this explains absence of S410p in Fig. 3(b). In the epicentral distance between 75° and 90° the slowness of S (around 10 s deg^{-1}) is sufficient for generating S410p. S410p in this stack at a time near -51.7 s is very clear (Fig. 3c). Its amplitude (0.021) is comparable with the theoretical amplitude of S410p for IASP91 standard model, with implication that the effect of interference between S and SKS in our data is small. The largest amplitude S410p is observed at the differential slowness of 0.2 s deg^{-1} , close to the theoretical slowness of 0.6 s deg^{-1} for horizontal layering. The difference can be explained by a tilt of 410 km discontinuity of around 2° or/and by the S velocity in the mantle rising from the east to the west.

To invert the P and S receiver functions for velocity models, we assume that the Earth is laterally homogeneous and isotropic. We use the 0 km trace of P receiver functions and 0 s deg^{-1} trace of S receiver functions. Other traces could be more appropriate for depths around 200 km, but at the relatively long periods of our receiver functions

this effect can be neglected. A trial model is described by the P and S velocities V_P and V_S , density and thickness of each plane layer. Density is derived from P velocity by using Birch law. For the P and S receiver functions we calculate synthetic Q and L components, respectively, as:

$$Q_{P,\text{syn}}(t, \mathbf{m}, c_P) = \frac{1}{2\pi} \int_{-\infty}^{+\infty} \frac{H_{P,Q}(\omega, \mathbf{m}, c_P)}{H_{P,L}(\omega, \mathbf{m}, c_P)} L_{P,\text{obs}}(\omega) \times \exp(i\omega t) d\omega \quad (1)$$

$$L_{SV,\text{syn}}(t, \mathbf{m}, c_S) = \frac{1}{2\pi} \int_{-\infty}^{+\infty} \frac{H_{SV,L}(\omega, \mathbf{m}, c_S)}{H_{SV,Q}(\omega, \mathbf{m}, c_S)} Q_{SV,\text{obs}}(\omega) \times \exp(i\omega t) d\omega. \quad (2)$$

Here t is time, ω is angular frequency, $\mathbf{m}$ is the vector of unknown model parameters, c_P and c_S are the adopted apparent velocities for the P and S receiver functions, respectively, indices 'obs' and 'syn' correspond to the actual receiver functions and their synthetic analogues, respectively, and H are theoretical transfer functions for the stack of plane layers. Angles i_P and i_{SV} are known for each individual receiver function but not for the stacked ones. In the inversion these angles are included in $\mathbf{m}$ and allowed to vary within a few degrees of their respective average values. The synthetics are calculated with the aid of Thomson–Haskell matrix algorithm (Haskell 1962). Earth flattening transformation is applied as in Biswas (1972). Note that the deconvolution means filtering of various components of the same recording by the same filter. With the L component of the P receiver function ($L_{P,\text{obs}}$) in the input, the resulting synthetic Q component (Q_{syn}) is automatically deconvolved, and it can be compared directly with the actual Q component of the P receiver function. A similar relation between the observed and synthetic components holds for the S receiver functions.

Let $E_P(\mathbf{m})$ and $E_S(\mathbf{m})$ be misfits for the stacked Q components of the P receiver functions and L components of the S receiver functions, respectively. We define the misfit (cost) as the rms difference between the observed and synthetic functions. The inversion is then to manipulate $\mathbf{m}$ in order to minimize $E_P(\mathbf{m})$ and $E_S(\mathbf{m})$ simultaneously. We conduct a search for the optimum models by using an interactive algorithm, similar to Simulated Annealing (Mosegaard & Vestergaard 1991). In this technique, the cost functions are minimized by applying a set of moves, that is, a set of model perturbations, and accepting or rejecting the moves according to the Metropolis rule (Metropolis *et al.* 1953). The Metropolis rule is formulated by using a parameter termed temperature. We use the Metropolis rule in cascade (Mosegaard & Tarantola 1995) for the two cost functions.

Our search technique is described in detail by Vinnik *et al.* (2007). We use a step-wise temperature function. The value of temperature at each step is chosen by the condition that the search, when initialized at an arbitrary point $\mathbf{m}_0$ within chosen bounds, converges to the same point for any $\mathbf{m}_0$. In a general case, the inversion procedure may require several temperature iterations. For the next iteration we set narrowed bounds resulting from the previous iteration, and repeat manual adjustment of temperature. The procedure is finished when the cost function in the convergence point has reached a sufficiently small predefined level.

We tested 4×10^5 models and selected those 5 per cent of them (Fig. 4), which provide a sufficiently small misfit between the observed and synthetic components of the receiver functions (less than 0.017 for S receiver functions). We divide the parameter space into cells, and present the models by the number of hits in each cell. These numbers are shown by colour code, and we demonstrate

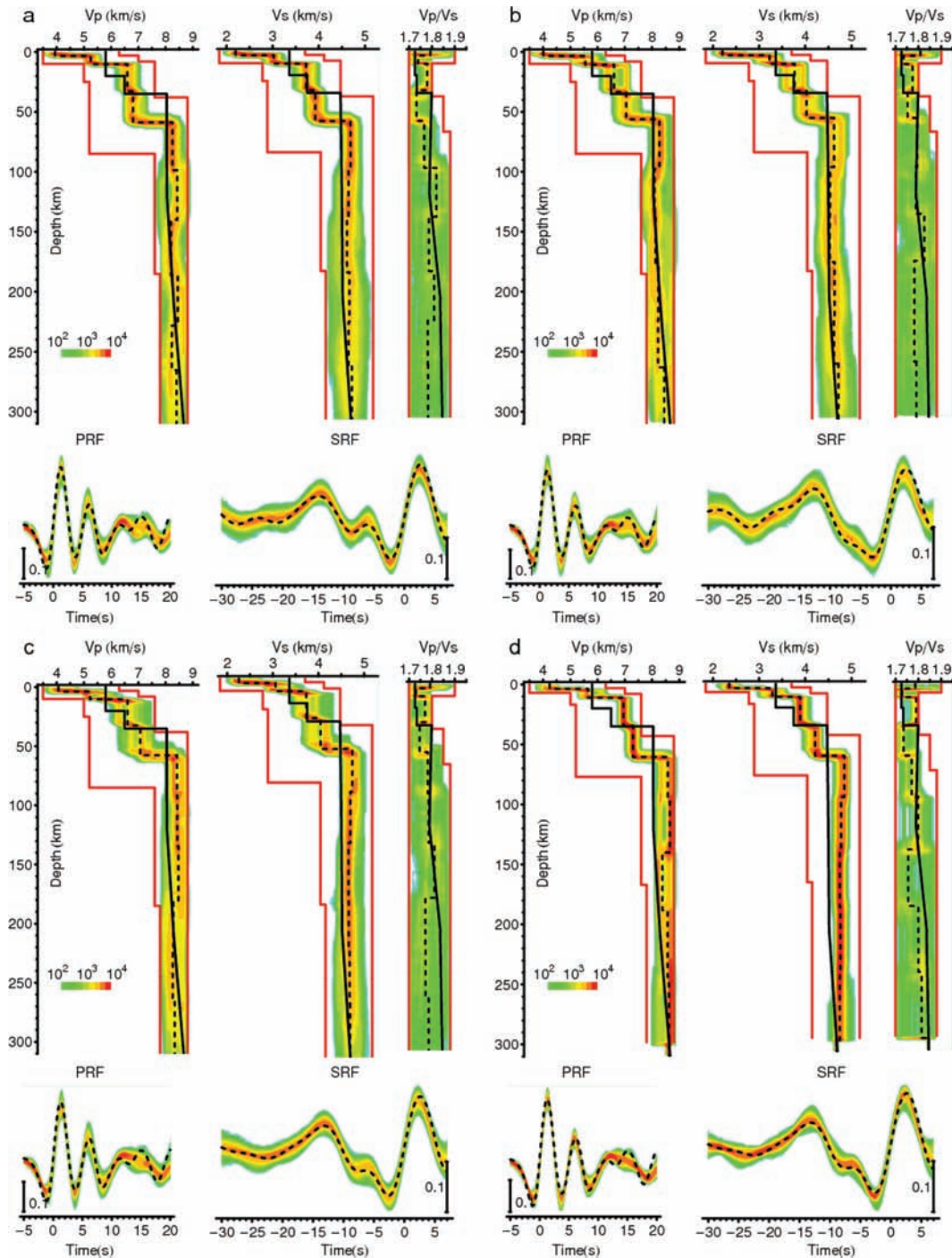


Figure 4. Probability density functions (colour code) of the isotropic velocity models obtained by joint inversion of P and S receiver functions. Search limits are shown by red lines. IASP91 velocities and their ratios are shown by solid lines; dash lines are for the medians. The same colour code shows probabilities for the synthetic P and S receiver functions (PRF and SRF, respectively); dash lines are the actual receiver functions. (a) S receiver functions are for the epicentral distance interval from 65° to 75° ; (b) is the same as (a) but for the interval from 75° to 90° ; (c) is the same as (a) but for the interval from 65° to 90° ; (d) is the same as (c), but the velocity models are additionally constrained by hypothetical P and S traveltime residuals of -0.8 and -1.9 s, respectively.

similar statistics for the data space. The trial models consist of 4 layers in the crust and five layers in the mantle. The values of c_P and c_{SV} are fixed at the average values of apparent velocities of the P wave and S phase, respectively. The search was conducted in a broad range of P and S velocities. The layer thickness could be up to 50–70 km in the mantle and up to 10–30 km in the crust. The obtained probability density functions for V_S contain a narrow band

of S velocities with the highest probability. The crust beneath NIL inferred from both subsets is 58 ± 2 km thick. The same crustal structure is obtained from the whole set of 119 individual S receiver functions (Fig. 4c). A 10 km thick low velocity zone in the upper crust is underlain by a layer with the S velocity of 3.7 km s^{-1} . The rest of the crust is represented by anomalously high S and P velocities (ca. 4 and 7 km s^{-1} between 30 and 58 km). S and P

velocities in the uppermost mantle are higher than in IASP91 by a few percent.

The idea of constraining the solution by the absolute (relative to IASP91) teleseismic P and S traveltime residuals (Vinnik *et al.* 2006) is based on the assumption that the residuals are accumulated in the layer sampled by the receiver functions. The joint inversion of the receiver functions and of the residuals can be performed with the technique, very similar to that, described for the receiver functions. If topography on 410 km discontinuity and lateral velocity variations at depths exceeding 300 km can be neglected, P and S residuals can be inferred from the residual of the P410s phase. This discontinuity is related to the phase transition from olivine to wadsleyite with a positive Clapeyron slope (Katsura & Ito 1989), and its depth is affected by temperature. Accordingly, the residual of P410s can be large and positive (in the hot mantle plumes) or negative (in the cold subduction zones). The latter is possible in the Himalaya and the region to the south of the Himalaya (Van der Voo *et al.* 1999).

Assuming a standard depth of 410 km discontinuity, the teleseismic P and S residuals can be inferred from the residual of P410s by adopting the standard ratio of 2.7 between the S and P residuals (Vinnik *et al.* 1999). To account for the difference in lengths of the wave paths of the P and S waves in the P receiver functions (10 per cent), the standard ratio should be reduced to 2.4. The time of the P410s phase is measured relative to the time of the P wave. Therefore, the residual of P410s (−2.2 s) is the difference between the teleseismic S and P residuals, −3.8 and −1.6 s, respectively. We could not obtain satisfactory results for these residuals, because the models thus obtained contained a strong artificial discontinuity at a limiting depth of 250–300 km. This result demonstrates that the residual of −2.2 s is, at least partly, an effect of uplift of the 410 km discontinuity or/and high velocities at depths exceeding 300 km. The receiver functions can be reconciled only with a fraction of this residual. This is demonstrated by example (Fig. 4d), where the adopted S and P residuals are reduced by one half, to −1.9 s and −0.8 s. Then for the whole S receiver function set we got the models, qualitatively similar to those obtained without the residuals.

3 DEPTH LOCALIZED AZIMUTHAL ANISOTROPY

Our data analysis follows the method of Vinnik *et al.* 2007. We assume that anisotropy is hexagonal with either horizontal or tilted symmetry axes, and it varies between a few layers. The receiver functions are obtained for discrete values ϕ_i of the backazimuth with a small sampling interval. For the $Q_i(t)$ and $T_i(t)$ components of the receiver functions, where t is time, we define weighting coefficients W_i^Q and W_i^T :

$$\begin{aligned} W_i^Q(\psi) &= -\cos 2(\psi - \phi_i) / \sum_i \cos^2 2(\psi - \phi_i), \\ W_i^T(\psi) &= \sin 2(\psi - \phi_i) / \sum_i \sin^2 2(\psi - \phi_i). \end{aligned} \quad (3)$$

The Q and T sums are obtained as

$$\begin{aligned} QF(t, \psi) &= \sum_i W_i^Q(\psi) Q_i(t), \\ TF(t, \psi) &= \sum_i W_i^T(\psi) T_i(t). \end{aligned} \quad (4)$$

Variations of QF and TF as functions of ψ can be caused by anisotropy and/or lateral heterogeneity of the isotropic medium. In the case of azimuthal anisotropy, $QF(t, \psi)$ and $TF(t, \psi)$ are similar, and this similarity presents a criterion of anisotropy. In the inversion of the receiver functions for the anisotropic models, the similarity

of $QF(t, \psi)$ and $TF(t, \psi)$ permits us to use, instead of $QF(t, \psi)$ and $TF(t, \psi)$, their average $SF(t, \psi) = [QF(t, \psi) + TF(t, \psi)]/2$.

We invert the receiver functions and SKS waveforms by fitting the theoretical SF components of the receiver functions and T components of SKS to the observed ones. Like in Section 2, the Earth medium is modelled by a stack of plane layers. For each trial model we calculate the synthetic Q and T components of the receiver function by using its L component as an input:

$$Q_{\text{syn}}(t, \mathbf{m}, \mathbf{c}) = \frac{1}{2\pi} \int_{-\infty}^{\infty} \frac{H_Q(\omega, \mathbf{m}, \mathbf{c})}{H_L(\omega, \mathbf{m}, \mathbf{c})} L_{\text{obs}}(\omega) \exp(i\omega t) d\omega \quad (5)$$

$$T_{\text{syn}}(t, \mathbf{m}, \mathbf{c}) = \frac{1}{2\pi} \int_{-\infty}^{\infty} \frac{H_T(\omega, \mathbf{m}, \mathbf{c})}{H_L(\omega, \mathbf{m}, \mathbf{c})} L_{\text{obs}}(\omega) \exp(i\omega t) d\omega. \quad (6)$$

Here ω is angular frequency, $\mathbf{m}$ is vector of unknown parameters of the model, $\mathbf{c}$ is vector of fixed parameters, subscripts obs and syn correspond to the observed receiver functions and their synthetic analogues, H_Q , H_T and H_L are theoretical transfer functions. Q_{syn} and T_{syn} are transformed into SF_{syn} by (4). Earth flattening transformation is applied as in Biswas (1972). The synthetics are calculated by using anisotropic version of the Thomson–Haskell algorithm (Keith & Crampin 1977; Kosarev *et al.* 1979). By using an expression similar to (5) and (6) we calculate the theoretical T component of each SKS phase, $T^{SKS}(t, \text{baz})$, where baz is backazimuth, by using its observed R component and the corresponding theoretical transfer functions. We define the misfit (cost) as the rms difference between the observed and synthetic functions, and conduct a search for the optimum models with the method outlined in Section 2.

To obtain the receiver functions, the raw recordings in a broad range of backazimuths (Fig. 5) were processed as described in Section 2. Azimuthal distribution in Fig. 5 is very inhomogeneous. To fill the sector between 340° and 360°, we used PP waves at the epicentral distances more than 100°. In order to obtain a homogeneous distribution in the azimuth domain, the individual receiver functions were stacked without moveout corrections in bins 20° wide.

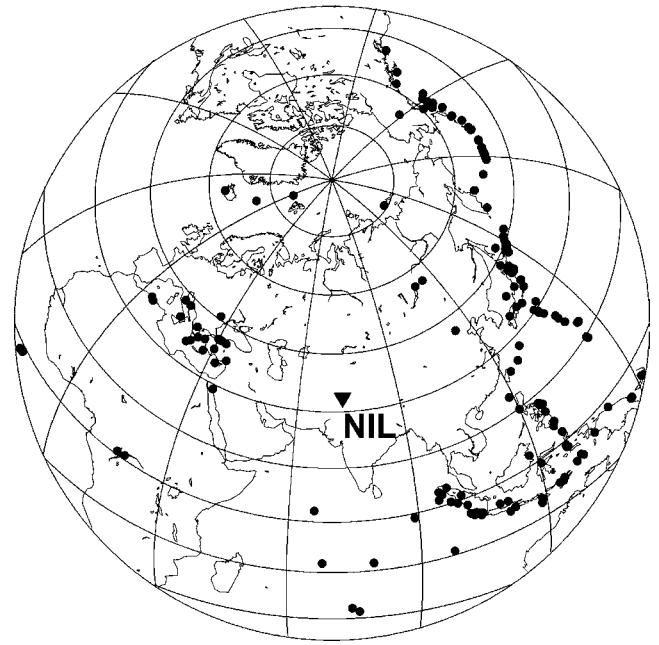


Figure 5. Locations of epicentres of seismic events used for obtaining P receiver functions for the analysis of anisotropy; those at distances more than 90° are not shown.

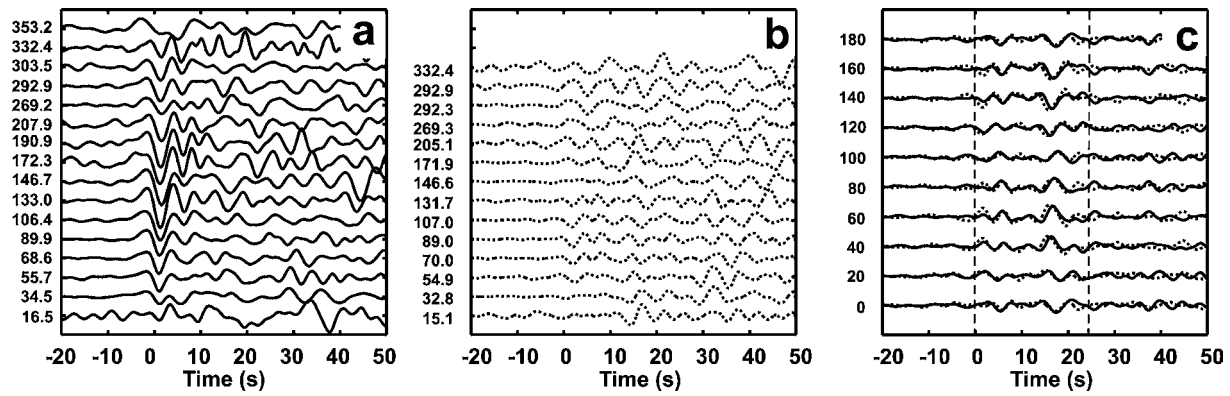


Figure 6. Q (a), T (b), QF and TF (c) components of the P receiver functions. Numbers on the left are backazimuths (a and b) and ψ (c) in degrees. Note coherence between QF and TF in the interval from 0 to 25 s (marked by vertical lines).

The average number of stacked receiver functions is around 10 per bin. Maximum amplitudes of the Q component in the time interval from 0 to 10 s are a few times larger than those of the T component (Fig. 6). Strong lateral heterogeneity of the lithosphere usually results in amplitudes of the T component in this interval comparable to those in the L components (see e.g. Vinnik *et al.* 2007). Small T component amplitudes in our data imply that lateral heterogeneity in the vicinity of the station is relatively weak. The resulting $QF(t, \psi)$ and $TF(t, \psi)$ (4) are coherent within the first 25 s, which is the criterion of azimuthal anisotropy.

Almost all recordings of the SKS phase come from a narrow azimuthal sector, about 10° wide. The T component amplitude in this group is small in comparison with noise. To improve signal-to-noise ratio we deconvolved both components of SKS (R and T) by the R component and stacked the deconvolved components (Fig. 7). The signal-to-noise ratio in the stack is sufficiently large. Inversion of these data with the SKS technique yielded a fast direction of 40° clockwise from north and a delay of the slow split wave of 0.4 s.

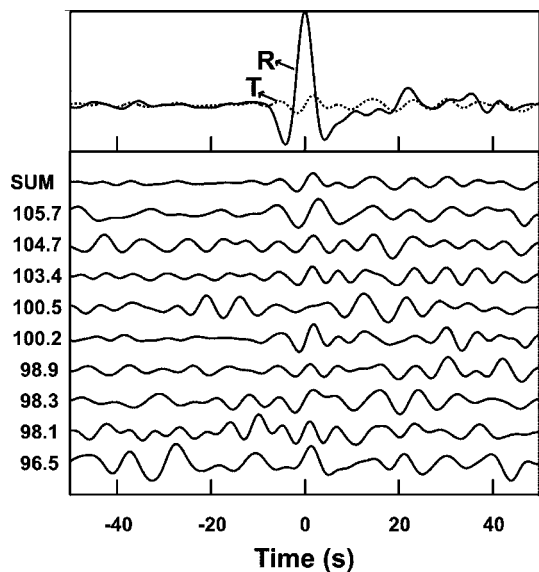


Figure 7. T components of the SKS waveforms deconvolved by the R components; numbers on the left are backazimuths of the seismic events. The deconvolved T components are stacked (trace SUM) to reduce noise. Top panel: deconvolved and stacked R and T components of SKS, used in constructing the models of anisotropy.

To invert the data in Figs 6 and 7, isotropic velocities V_P and V_S , density and the depth of the Moho were fixed at the values corresponding to the model in Fig. 4(c). Apparent velocities and angles i_p were fixed at their average values in each bin. Anisotropic parameters to be found for each layer are $\delta V_S/V_S$, where δV_S is the difference between the fastest and the slowest S velocities, and azimuth of the symmetry axis (fast azimuth). A description of anisotropy requires two other parameters: $\delta V_P/V_P$, which is defined like $\delta V_S/V_S$, and η , which controls velocity variations between the fast and slow directions. The values of $\delta V_P/V_P$ and η were fixed at 1.5 and 1.0. The inversion is relatively insensitive to them; for justification of these numbers see Vinnik *et al.* (2007). The crust of our model consists of 2 anisotropic layers of fixed thickness. The mantle consists of three anisotropic layers of unknown thickness, which may vary in a broad range. The adopted upper limit for S wave anisotropy ($\delta V_S/V_S$) is 10 per cent for both, the crust and the mantle. The range of fast azimuths is between 0° and 180° .

We tested around 10^5 models and selected those 700, which provide sufficiently small misfit between the observed and synthetic SF (less than 0.02) and between the observed and synthetic T components of SKS. Like in Section 2 (Fig. 4) we divide the parameter space into cells, and present the results by the numbers of hits in each cell, which are shown by colour code. A similar statistics is shown for the data space. In the selected models (Fig. 8), mantle S -wave anisotropy is large (5–7 per cent) in the depth interval between 80 and 220 km. No anisotropy is found at a larger depth. A pronounced change in the fast directions of anisotropy occurs at a depth of 160 km. In the upper and the lower layer the fast azimuths are around 60° and 150° . The Ps phase from this boundary is seen in QF and TF at a time of 16 s.

In the crust and a thin subcrustal layer anisotropy is about twice weaker than in the two other layers and the estimates of $\delta V_S/V_S$ are rather uncertain. The fast azimuth in the lower crustal layer is around 160° . We tested models with the isotropic crust. For most of them a misfit for SF in the interval from 0 to 8 s was relatively large but the optimum models of the mantle remained nearly the same. In the model with the smallest misfit thus obtained (Fig. 8) the 160 km discontinuity is at a depth of 150 km; magnitudes of S -wave anisotropy in the third and second mantle layers are *ca.* 5 and 6 per cent.

4 DISCUSSION AND CONCLUSIONS

The T component of the P receiver functions is weak relative to the Q component in all azimuths (Fig. 6), which is indicative of a

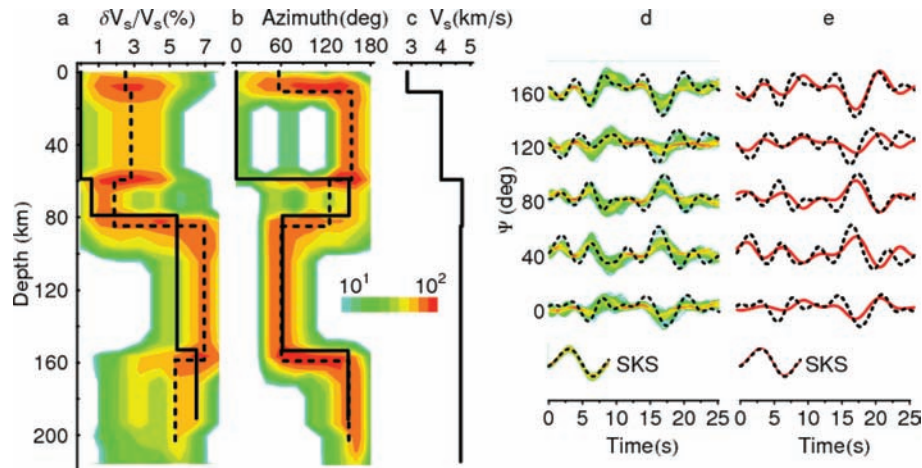


Figure 8. Joint inversion of the *SF* components of the *P* receiver function and the deconvolved and stacked *T* component of SKS. (a) Probability density function (colour code) of dV_s/V_s (%) for selected models and the median (dash line); the example of the model with isotropic crust is shown by solid line. (b) The same as a, but for the azimuth of fast axis. (c) Isotropic *S* velocity model, used in the inversion. (d) Observed *SF* (dash lines) and probability density functions (colour code) of theoretical *SF* for the models in (a) and (b); the lower trace is the same but for the *T* component of SKS. (e) Observed (dash) and synthetic (red) *SF* for the model with the isotropic crust in (a) and (b).

subhorizontally layered structure in the vicinity of station NIL. Joint inversion of the *P* and *S* receiver functions favours high uppermost mantle *P* and *S* velocities relative to IASP91 standard model (Figs 4a–c). The velocities higher than standard are characteristic of the mantle lithosphere of the Archean cratons (Jordan 1988). However, the velocities in the lithosphere of the Indian shield present an exception: whereas in the typical shields P410s phase arrives 1–2 s earlier than predicted by IASP91, at station HYB (Hyderabad) it arrives at a standard time (Chevrot *et al.* 1999), and similar estimates are known for other southern stations (Kumar & Mohan 2005, Kiselev *et al.* 2007). Joint inversion of the *P* and *S* receiver functions of the southern stations (Kiselev *et al.* 2007) reveals standard velocities in the uppermost mantle, a few per cent lower than in Fig. 4. A recovery of normal cratonic mantle structure in the north of the Indian shield is consistent our data and it might explain the relatively high upper-mantle velocities at depths less than 250–300 km in the tomographic models by Replumaz *et al.* (2004), Zhou & Murphy (2005) and Li *et al.* (2006).

Whereas a fraction of the –2.2 s residual of P410s at NIL can be attributed to the shield structure in the vicinity of the station, the rest is indicative of anomaly within and near the mantle transition zone (MTZ). Similar early arrivals of P410s were found in the Himalaya a few hundred kilometres to the SE of NIL by Oreshin *et al.* (2007), and it is likely that the observations in both regions sample the same anomalous structure. Therefore, interpretation of these observations by Oreshin *et al.* (2007) is applicable to the data of NIL. In particular, early arrivals of P410s seem to be associated with anomalously thick transition zone (Oreshin *et al.* 2007; Prakasam *et al.* 2007). The discontinuities bounding the MTZ are caused mainly by phase transitions from olivine to wadsleyite at 410 km and from ringwoodite to perovskite and magesiowüstite at 660 km, with positive and negative Clapeyron slopes, respectively (Katsura & Ito 1989). Beneath the Himalaya to the SE of NIL, the depth interval between the two discontinuities is larger than the standard interval of 250 km by 10 km, with implication that this is an effect of low temperature. For the Clapeyron slopes of about $\pm 3 \text{ MPa K}^{-1}$ the anomaly of 10 km corresponds to a temperature anomaly of about 100 °C. A low temperature leads to uplift of the 410 km discontinuity and early arrivals of the P410s seismic phase. In fact, the anomalous interval between

the two discontinuities can be a direct consequence of the uplift of the 410 km discontinuity. High velocities associated with the low temperature can be another reason for the early P410s arrivals.

Underthrusting of the Indian Plate at MCT started roughly 20 Ma (Molnar 1988) and, assuming a steep subduction at a rate of around 2 cm yr^{-1} , the plate could already reach 410 km discontinuity. However, the Indian Plate would be cooler than the ambient mantle at this depth by several hundred degrees C. Moreover, the phase transition at this temperature can be kinetically hindered (Kirbi *et al.* 1996), and a wedge-shaped zone of metastable peridotite may persist to large depths. Not only thickening of the MTZ would be unobservable, but even a detection of P410s would be problematic, as demonstrated by observations in a typical subduction zone (Vinnik *et al.* 1996). We conclude that the observed phenomenon, perhaps, is not an effect of the ongoing intracontinental subduction. More likely, this is an effect of Tethys subduction, active a few hundred kilometres to the north of the MCT and extinct since about 20 Ma. As the Indian Plate is moving to the northeast with a velocity of about 4 cm yr^{-1} , it now overrides remnants of subduction left at a depth of several hundred kilometres (Van der Voo *et al.* 1999).

Our analysis of anisotropy reveals a pronounced change in the fast direction of anisotropy in the mantle at a depth around 160 km. Previous analysis of SKS in the neighbouring region in Pakistan demonstrated no *S*-wave splitting, which might be explained either by isotropy of the Earth's medium or by a two-layer anisotropic structure, in which the effects of the two layers cancel out (Sandvol *et al.* 1994). Our analysis favours the second option.

Fast direction of anisotropy in the lower layer is parallel to the trend of the Himalaya (150°). A similar alignment is known in some other compressional mountain belts, but the depth of the effect is usually a matter of speculation and its origin is a subject of dispute (Nicolas 1993; Tommasi *et al.* 1999). However, application of the same method at the stations in and near the Tien Shan mountains has shown that the alignment of the fast direction of anisotropy with the trend of the belt, like in our observations, is mainly the asthenospheric effect (Vinnik *et al.* 2007). According to the rock physics the fast direction in the mantle tends to be aligned with the flow or/and at right angle to the uniaxial shortening direction.

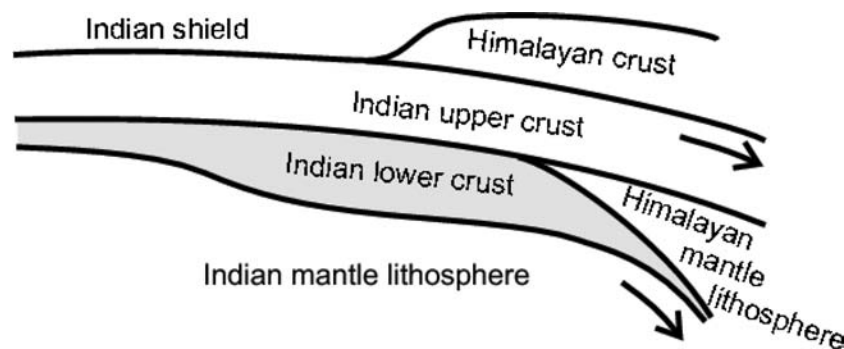


Figure 9. Schematic model of the continental collision that may explain origin of the thick high-velocity crustal layer (grey).

A similar two-layer structure was previously found with a similar approach in south Germany (Farra & Vinnik 2000), where the fast direction in the asthenosphere is aligned with the trend of the Alps. The boundary between the two anisotropic layers in south Germany and in the Tien Shan is found at a depth of 80–90 km and around 100 km, respectively. This depth corresponds to temperature of around 1000 °C, in agreement with the definition of anisotropic lithosphere and asthenosphere (Estey & Douglas 1986). The larger depth of this boundary beneath NIL is very likely an effect of a cooler shield geotherm.

Anisotropy in the subcrustal lithosphere can be either a result of current deformations, or it can be inherited from the geological past and ‘frozen’ for long time. The current deformations should be, most likely, in the form of subhorizontal simple shear in the azimuth of plate motion (Tommasi *et al.* 1999). This deformation would lead to subhorizontal foliation planes with a small splitting for the S waves propagating in the subvertical directions. Large $\delta V_S/V_S$ in the subcrustal lithosphere of our model suggests a subvertical foliation and favours either ‘frozen’ anisotropy or current deformations, but very different from subhorizontal shearing.

Our analysis in Section 2 assumes that the Earth’s medium is isotropic, which is a simplification. The results of this analysis could be seriously affected by anisotropy if the backazimuth of our seismic events was coincident with either fast or slow direction in one of the anisotropic layers. Our backazimuths are intermediate, and our isotropic models don’t contain any significant discontinuity at a depth of 160 km, where anisotropy changes its direction.

To summarize, our data are consistent with the high velocities to the south of the western Himalaya being an effect of a shield like structure in the uppermost mantle, and this interpretation may present a viable alternative to the models, where this structure is interpreted as an image of a steeply plunging Indian lithosphere. Velocities in the uppermost mantle of this region can be higher than in the rest of the Indian shield, because in the south they are anomalously low. High velocities at larger depths are probably related to the remnants of Tethys subduction.

The ongoing continental collision in the Himalaya is expressed in crustal structure: the crust beneath NIL is very thick (around 58 km), and at depths exceeding 30 km is represented by very high P and S velocities (around 7 and 4 km s⁻¹, respectively). Previously, similar crustal structures (large thickness and anomalously high velocity at moderate depths) were found beneath the northern and southern margins of Tarim (Vinnik *et al.* 2006; Li *et al.* 2002, respectively). The model in the north was obtained with the method, similar to the method of the present study; the data for the south is a result of wide-angle seismic sounding. Tectonic setting of these regions is very similar to that in our study: in the north and south, Tarim

underthrusts the Tien Shan and the Kunlun, respectively. Indications of a similar structure are found beneath the lesser Himalaya, a few hundred kilometres to the SE of NIL (Oreshin *et al.* 2007). The lower continental crust consists of high-velocity ductile rocks, and it is possible that this ductile layer is partly scraped off by underthrusting and accreted in front of the continental subduction zone in about the same way as sediments of the subducting oceanic lithosphere (Fig. 9). Apparently, this phenomenon is reproduced in some numerical experiments of Toussaint *et al.* (2004).

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