

# Depth localized azimuthal anisotropy from *SKS* and *P* receiver functions: The Tien Shan

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## SUMMARY

Shear wave splitting in the seismic *SKS* phase provides a unique possibility to judge on deformations at depths inaccessible for direct observations. Fast *S* wave polarization direction in collisional belts is often parallel to the trend of the belt, although deformations of the mantle lithosphere in low-angle thrusts would lead to the fast polarization direction normal to the trend of the belt. These considerations suggested that the upper mantle in collisional belts is decoupled from the crust. However, *SKS* technique is notable by a poor depth resolution, and usually it assumes that the fast polarization direction is the same at any depth, which is hard to justify. Here, to investigate depth dependent azimuthal anisotropy in the mantle, we invert jointly *P* receiver functions and *SKS* particle motions at a number of seismograph stations. The technique involves azimuthal filtering of the receiver functions and provides a criterion to discriminate between the effects of azimuthal anisotropy and lateral heterogeneity of isotropic medium. A search for the optimum models is conducted with a technique similar to simulated annealing. Testing with synthetics demonstrates that this approach is robust. The results for 10 seismograph stations in the Tien Shan, the world's most active intracontinental collisional belt in Central Asia, reveal a pronounced change in the patterns of azimuthal anisotropy at a depth around 100 km. In the mantle lithosphere (at depths less than 100 km), anisotropy is relatively weak and fast wave polarization direction varies laterally in a broad range. This layer is not necessarily decoupled from the crust: its anisotropy can be a combined effect of present day thrusting and of deformations of the geologic past. In the lower layer (asthenosphere) the average azimuth of fast wave polarization is close to the trend of the belt, whereas magnitude of *S* wave anisotropy is stable and large (between 5 and 6 per cent). This anisotropy is a likely result of recent uniaxial shortening at right angle to the trend of the belt. At some stations the data require anisotropy in the crust. There is no evidence for anisotropy at depths exceeding 150–250 km.

**Key words:** asthenosphere, lateral heterogeneity, lithospheric deformation, plate convergence, seismic anisotropy, shear-wave splitting.

## 1 INTRODUCTION

Seismic anisotropy presents a unique possibility to study rock fabric and the related deformations at depths inaccessible for direct observations (e.g. Nicolas & Christensen 1987; Savage 1999). The available seismological tools differ by the wave types, depth ranges, accuracy and resolution. *SKS* and similar seismic phases (*SKKS*, *PKS*) are shear waves propagating in a near-vertical direction and, in the absence of azimuthal anisotropy, polarized in the wave propagation plane (*SV*). If *SKS* propagates through anisotropic zone, the *SV* wave splits into two quasi-shear waves with differing velocities and polarizations nearly perpendicular to the ray-path and

to each other. At long periods, the resulting transverse (*T*) motion component looks like the derivative of the radial (*R*) component: the signal in the *T* component is shifted by quarter period relative to the *R* component, and the *T/R* amplitude ratio is proportional to frequency. This relationship between the components of *SKS* provides an effective diagnostics for azimuthal anisotropy (Vinnik *et al.* 1984). *SKS* particle motion can be interpreted in terms of two parameters: fast polarization direction and delay of the slow split wave. In comparison with other seismological tools, lateral resolution of the *SKS* techniques in the upper mantle is very high: it may resolve lateral variations of the parameters of anisotropy on a scale of about 100 km (Rümpker & Ryberg 2000) versus several hundred kilometres for surface waves.

Implicit in the classic *SKS* technique is the assumption of a single anisotropic layer. In principle, *SKS* particle motion can also be inverted for a two-layer anisotropic structure (e.g. Menke & Levin

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2003), but the absolute depths remain unknown. Moreover, the analysis requires high-quality recordings of *SKS* in many azimuths. Usually this is not the case: in Central Asia, for example, most of the available *SKS* recordings are from the same narrow azimuthal sector corresponding to events in the Fiji-Tonga region. Even in the case of one-layer anisotropy, *SKS* techniques provide only very rough constraints on the depth of the anisotropic layer, which undermines scientific significance of the data thus obtained.

Here we demonstrate an upgraded *SKS* technique that can detect both lateral and vertical variations of seismic azimuthal anisotropy. The idea of the method is to combine observations of shear wave splitting in *SKS* and *P* receiver functions. The first application of *P* receiver functions to the analysis of mantle anisotropy (Kosarev *et al.* 1984) was later perfected in several studies (Vinnik & Montagner 1996; Girardin & Farra 1998; Farra & Vinnik 2000; Vinnik *et al.* 2002). For other applications of *P* receiver functions to the problems of anisotropy see, e.g. Bostock (1997, 2003), Levin & Park (1997). Vinnik *et al.* (2002) conducted a search for anisotropic models of the crust and mantle beneath the Tien Shan in Central Asia with the aid of a genetic algorithm. The present work can be viewed as a development of that study. Whereas the main ideas and the database remain practically the same, a search for the optimum models is conducted with a better technique and a significant attention is paid to the problems of uniqueness and accuracy. Whereas previously used isotropic models of the crust and mantle of the Tien Shan were very rough, those used in the present study are obtained by a joint inversion of the *P* and *S* receiver functions and of the teleseismic *P* and *S* residuals of each station (Vinnik *et al.* 2006). The aim of the paper is two-fold: we describe methodology that can be used worldwide and investigate deep structure of a unique region.

The region of our study (Fig. 1) belongs to the Tien Shan, the world's largest and most active intracontinental mountain belt. It was initially created by convergence of a few continental blocks in the Paleozoic (Burtman 1975) and transformed into a peneplain in the Mesozoic. A small volume of basaltic magmas that erupted in the late Cretaceous and early Paleogene (Dobretsov & Zagruzina 1977) is an effect of a plume (Grachev 1999). Tectonic activity resumed at about 20 Myr (Sobel & Dumitru 1997), apparently in re-

sponse to the India-Eurasia collision (Molnar & Tapponnier 1975). At present, the Tien Shan is shortening at a rate of about 20 mm/yr (Abdrakhmatov *et al.* 1996; Reigber *et al.* 2001). At this rate, with a total shortening of about 60 km, which is accommodated at several thrust or reverse faults (Abdrakhmatov *et al.* 2001), the Tien Shan could be constructed in a few Myr. Crustal thickness beneath the Tien Shan, evaluated by a joint inversion of *P* and *S* receiver functions, varies between about 40 and 60 km. The upper mantle is laterally heterogeneous: at a depth exceeding 100 km, the dominant feature is a broad low velocity zone oriented at a right angle to the strike of the belt (Vinnik *et al.* 2006). Seismic anisotropy in the upper mantle of the Tien Shan as found with *SKS* techniques generally follows the pattern found in several other collisional belts: fast split *S* wave at most stations is parallel to the trend of the belt (Makeyeva *et al.* 1992; Wolfe & Vernon 1998; Roecker 2001).

Our analysis of seismic data permits us to address the problem of parallelism between polarization of the fast split wave and the trend of the belt, a subject of dispute for many years (e.g. Nicolas 1993). In thrust belts like the Tien Shan, deformations in the lithosphere can be described as near-horizontal simple shear. This means that fast *S* waves in *SKS* should be polarized parallel to the shear direction, i.e. normal to the belt. The discrepancy between the actual and predicted directions suggested that orogenic domains deforming by thrusting and crustal thickening are decoupled from the upper mantle (Tommasi *et al.* 1999). We will check if this suggestion is correct.

## 2 METHOD OF DATA PROCESSING: DIAGNOSTICS OF AZIMUTHAL ANISOTROPY

Methodology of our work is close to (Farra & Vinnik 2000; Vinnik *et al.* 2002).

The method of data processing involves construction of *P* receiver functions. *P* receiver function is the response of the earth medium in the vicinity of the seismograph station to excitation by the *P* waves of distant earthquakes. This response consists of the *Ps* converted phases, reflected and scattered waves. In the method (Vinnik 1977), which we use, this response is presented by projections of particle motion on the axes *P* (or *L*), *SV* (or *Q*) and *T*. Axis *L* is parallel to the principal *P* wave particle motion direction in the wave propagation plane. Axis *Q* is normal to *L* in the same plane and is nearly optimum for observing the *Ps* converted phases in the isotropic earth. Horizontal projections of the axes *L* and *Q* are positive in the direction from the source to the receiver. The axes *L*, *Q* and *T* form the right-hand coordinate system. A typical angle between the vertical axis and *L* is around 20°. Axis *T* is perpendicular to *L* and *Q*. All components are deconvolved by the *L* component. In an isotropic laterally homogeneous earth, the *P* wave is coupled only with *SV*. Consequently, the converted and reflected waves are polarized strictly in the plane of wave propagation. In the presence of azimuthal anisotropy or/and lateral heterogeneity, the converted and reflected waves are recorded in all three components, and their amplitudes depend on the azimuth.

The wave phenomena that make *P* receiver functions very useful in studies of depth dependent azimuthal anisotropy are illustrated by theoretical seismograms for the plane *P* wave propagating through the stack of plane layers over the isotropic half-space. For simplicity the model (Table 1) includes only one anisotropic layer. The wave field at the earth's surface (Fig. 2) is calculated with the Thomson-Haskell-Crampin algorithm (Keith & Crampin 1977; Kosarev *et al.* 1979) for the apparent velocity of 18.5 km s<sup>-1</sup>. Anisotropy is

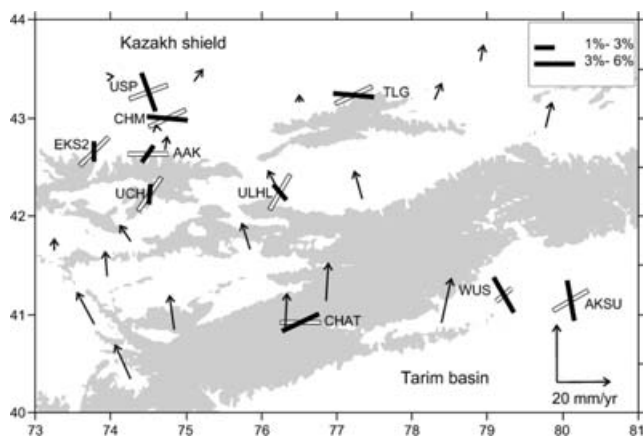
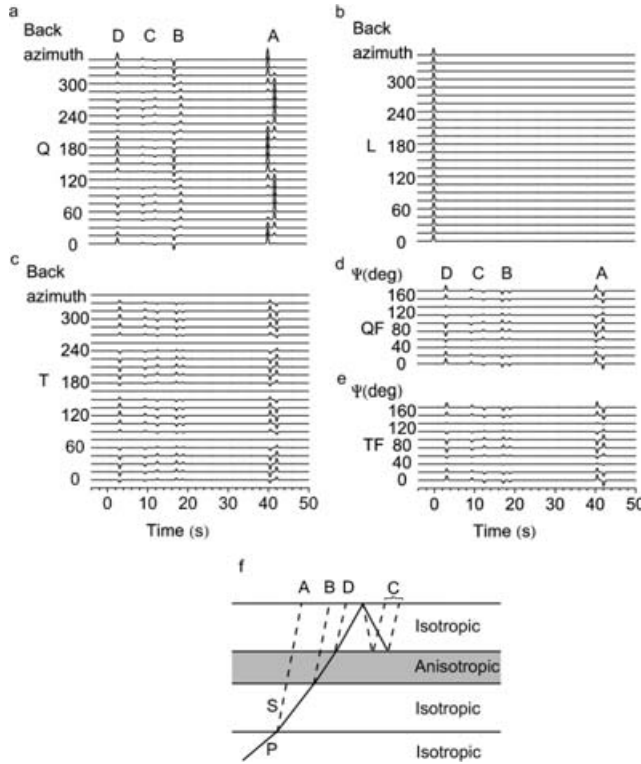


**Figure 1.** Map of the Central Tien Shan. Altitudes exceeding 3000 m are shown grey. The parameters of anisotropy are obtained in the present study. Azimuth of polarization of the fast split wave is shown for the mantle lithosphere (filled bars) and the asthenosphere (open bars); the length of a bar depends on magnitude of *S* wave anisotropy. Location of the seismograph station corresponds to the point of intersection of the bars. GPS velocities (Reigber *et al.* 2001) are shown by arrows; the scale is in the lower right corner.

**Table 1.** Parameters of model 1.

| $V_p$ | $V_s$ | $\rho$ | Depth | Azimuth<br>deg | $dV_s/V_s$<br>percent |
|-------|-------|--------|-------|----------------|-----------------------|
| 8.5   | 4.7   | 3.45   | 0     |                | 0                     |
| 8.5   | 4.7   | 3.45   | 30    |                | 0                     |
| 8.5   | 4.7   | 3.45   | 30    | 0              | 5                     |
| 8.5   | 4.7   | 3.45   | 180   | 0              | 5                     |
| 8.5   | 4.7   | 3.45   | 180   |                | 0                     |
| 8.5   | 4.7   | 3.45   | 410   |                | 0                     |
| 9.15  | 5.0   | 3.80   | 410   |                | 0                     |

**Figure 2.** Synthetic seismograms of  $Q$  (a),  $L$  (b),  $T$  (c) components and transformed  $QF$  (d) and  $TF$  (e) components for model 1 (Table 1). Ray diagrams of seismic phases A, B, C and D are shown in (f). Amplitude of the  $L$  component is shown decremented 10 times relative to the other components. Note similarity between  $QF(t, \psi)$  and  $TF(t, \psi)$ .

hexagonal with a horizontal symmetry axis. Whereas other symmetries can be present in the Earth, hexagonal symmetry is the simplest, and at the same time able to capture the major observable effects (Becker *et al.* 2006). The isotropic component of the model is described by velocities  $V_p$ ,  $V_s$  and density  $\rho$ . The anisotropic component is defined by the values of  $\delta V_p/V_p$ ,  $\delta V_s/V_s$ ,  $\eta$  and azimuth of the fast direction  $\alpha$ . Here  $\delta V_p$  and  $\delta V_s$  are the differences between the  $V_p$  and  $V_s$  velocities parallel (fast) and perpendicular to the symmetry axis (slow),  $\eta$  is the parameter which controls velocities for the directions intermediate between the fast and slow. These parameters are related to the five elastic parameters  $C$ ,  $A$ ,  $L$ ,  $N$ ,  $F$  for hexagonal anisotropy with a vertical symmetry axis:

$$\begin{aligned}\sqrt{C/\rho} &= V_p(1 + \delta V_p/2V_p), \quad \sqrt{A/\rho} = V_p(1 - \delta V_p/2V_p), \\ \sqrt{L/\rho} &= V_s(1 + \delta V_s/V_s) \\ \sqrt{N/\rho} &= V_s(1 - \delta V_s/V_s), \quad F = \eta(A - 2L).\end{aligned}\quad (1)$$

The resulting  $L$  component (Fig. 2) contains practically only the pulse of the incoming  $P$  wave, whereas in the  $Q$  and  $T$  components there are several arrivals. The group labelled A consists of two pulses. They are generated by conversion from the  $P$  to  $S$  at the isotropic lower boundary of the model and subsequent splitting of the  $S$  wave in the anisotropic layer. This wave pair is analogous to  $SKS$ : they differ only in the depth of conversion (upper mantle for A and the core-mantle boundary for  $SKS$ ). The wave pair B is converted from  $P$  to  $S$  at the lower boundary of the anisotropic layer and further modified by propagation of the resulting  $S$  in the anisotropic layer. The wave pair C consists of the pulses propagating in the layer between the upper boundary of the anisotropic layer and the free surface. The first wave is reflected as  $P$  from the free surface and converted to  $S$  by reflection at the layer bottom. The second wave is reflected as  $S$  from the free surface and the layer bottom. Similar waves reflected from the lower boundary of the anisotropic layer arrive outside the shown time interval. Finally, the wave D is the  $Ps$  converted phase from the upper boundary of the anisotropic layer. This wave propagates in the isotropic medium, but its polarization differs from  $SV$  owing to the conversion from the anisotropic boundary. Note that the waves B, C and D are generated at the boundaries of the anisotropic layer and thus are sensitive to the distribution of anisotropy with depth.

In the real earth, the secondary phases like  $Ps$  are affected by both anisotropy and lateral heterogeneity. We assume that the effects of both are small and, for this reason, additive. Then, if they vary in the azimuth domain with different periodicity, the signal can be separated from noise by azimuthal filtering. In the case of hexagonal anisotropy with a horizontal axis of symmetry, the anisotropic components of  $Q$  and  $T$  vary in the azimuth domain with a dominant period of  $180^\circ$  (Fig. 2). There is no clear reason for the effects of lateral heterogeneity to vary with the same periodicity: if heterogeneity is in the form of a dipping interface, the related effects vary in the azimuth domain with a period of  $360^\circ$ . Then filtering out the 2<sup>nd</sup> harmonic of the Fourier series in both the  $Q$  and  $T$  components of receiver functions may suppress the effects of lateral heterogeneity and isolate the effect of azimuthal anisotropy.

We assume that the receiver functions are obtained for discrete values  $\phi_i$  of the back azimuth with a small sampling interval, and extract the 2<sup>nd</sup> harmonics by summation of the components of individual receiver functions with weights depending on the azimuth. For the  $Q_i(t)$  and  $T_i(t)$  components, where  $t$  is time, we define weighting coefficients  $W_i^Q$  and  $W_i^T$ :

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

The  $Q$  and  $T$  sums are obtained as

$$\begin{aligned}QF(t, \psi) &= \sum_i W_i^Q(\psi) Q_i(t), \\ TF(t, \psi) &= \sum_i W_i^T(\psi) T_i(t).\end{aligned}\quad (3)$$

Application of these transformations to  $Q_i(t)$  and  $T_i(t)$  demonstrates that  $QF(t, \psi)$  and  $TF(t, \psi)$  are similar (Fig. 2), 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  $SF(t, \psi) = (QF(t, \psi) + TF(t, \psi))/2$ . Effect of seismic noise is demonstrated by example, where a strong noise is added to the theoretical  $Q_i(t)$  and  $T_i(t)$  (Fig. 3a) corresponding to the model 2, which contains two anisotropic layers (Table 2). Spectral densities of

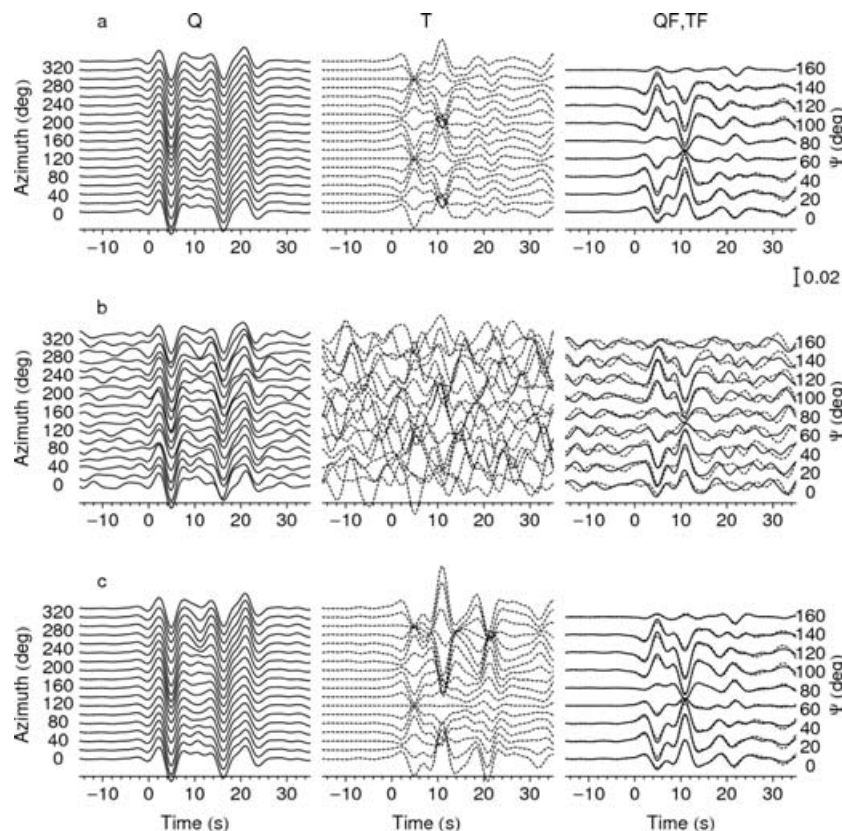


**Figure 3.** Synthetic  $Q$ ,  $T$ ,  $QF$  and  $TF$  components for model 2 (Table 2). Amplitudes of the  $T$ ,  $TF$  and  $QF$  component are shown incremented 5 times relative to the  $Q$  component. The scale bar corresponds to  $T$ ,  $TF$  and  $QF$ . Upper row: synthetics for the unperturbed model. Middle row: the result of summation of the  $T$  and  $Q$  components in the upper row and noise. Bottom row: the same as in (a), but for the symmetry axis in the lower layer tilted by  $45^\circ$ .

**Table 2.** Parameters of model 2

| $V_p$<br>km s $^{-1}$ | $V_s$<br>km s $^{-1}$ | $\rho$<br>g cm $^{-3}$ | Depth<br>km | Azimuth<br>deg | $dV_s/V_s$<br>per cent |
|-----------------------|-----------------------|------------------------|-------------|----------------|------------------------|
| 6.4                   | 3.7                   | 2.9                    | 0           |                | 0                      |
| 6.4                   | 3.7                   | 2.9                    | 40          |                | 0                      |
| 8.1                   | 4.5                   | 3.3                    | 40          | 30             | 4                      |
| 8.1                   | 4.5                   | 3.3                    | 100         | 30             | 4                      |
| 8.1                   | 4.5                   | 3.3                    | 100         | 100            | 2                      |
| 8.1                   | 4.5                   | 3.3                    | 200         | 100            | 2                      |
| 8.5                   | 4.72                  | 3.4                    | 200         |                | 0                      |

the signal and noise are similar, and there is no correlation between the signal and noise. The RMS amplitude of noise is around 2 per cent of the  $L$  component. The results of transformation of the noisy  $Q_i(t)$  and  $T_i(t)$  (Fig. 3b) into  $QF$  and  $TF$  demonstrate that whereas the signals are beyond recognition in the  $T_i(t)$ , they are very clear and almost similar in  $QF(t, \psi)$  and  $TF(t, \psi)$ . The use of  $SF(t, \psi)$  instead of  $QF$  and  $TF$  would facilitate a further reduction of the noise.

In the actual earth the symmetry axes can be tilted, and this might change the 2<sup>nd</sup> azimuthal harmonic in the receiver functions. However, the example, where the axis in the lower layer of the model 2 is tilted by  $45^\circ$ , demonstrates, that the criterion of similarity between  $QF(t, \psi)$  and  $TF(t, \psi)$  is robust (Fig. 3c). Apparently, the effect of the tilt is mainly in the first azimuthal harmonic, which is filtered out by transformation (3).

### 3 METHOD OF INVERSION OF RECEIVER FUNCTIONS AND SKS PARTICLE MOTIONS

We invert the receiver functions and  $SKS$  waveforms for azimuthal anisotropy by fitting the theoretical  $SF$  components of the receiver functions and  $T$  components of  $SKS$  to the observed ones. The earth medium is modelled by a stack of plane layers. A trial model is described by isotropic and anisotropic parameters of hexagonal anisotropy for each layer (see Section 2). For each trial model we calculate the synthetic  $Q$  and  $T$  components of the receiver function by using its  $L$  component as an input:

$$Q_{syn}(t, \mathbf{m}, c) = \frac{1}{2\pi} \int_{-\infty}^{\infty} \frac{H_Q(\omega, \mathbf{m}, c)}{H_L(\omega, \mathbf{m}, c)} L_{obs}(\omega) \exp(i\omega t) d\omega \quad (4)$$

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

Here  $\omega$  is angular frequency,  $\mathbf{m}$  is vector of parameters of the model,  $c$  is apparent velocity, 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  $SF_{syn}$  as described in the preceding Section. The theoretical transfer functions are calculated with the Thomson-Haskell-Crampin algorithm (Keith & Crampin 1977; Kosarev *et al.* 1979). Earth flattening transformation is applied as in Biswas (1972).

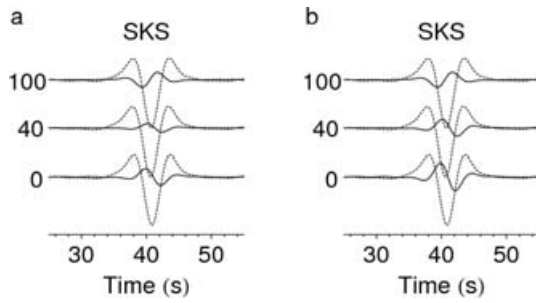


**Figure 4.** Synthetic seismograms of the SKS phase for the unperturbed model in Table 2 (a) and for the symmetry axis in the lower layer tilted by 45° (b). The radial and transverse components are shown by dash and solid lines, respectively.

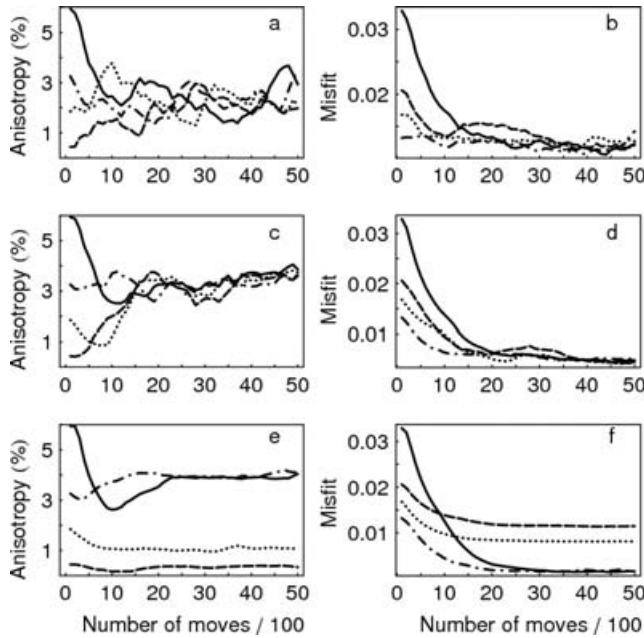


**Figure 5.** Effect of the temperature on the search process. Anisotropy ( $\delta V_s/V_s$ ) in the upper anisotropic layer of Table 2 (left) and cost function (right) versus the number of moves for the temperatures: 10 (a, b), 1 (c, d) and 0.1 (e, f). The misfit is between the synthetic  $SF(t, \psi)$  for model 2 (Table 2) and for the trial models. Each curve corresponds to a randomly generated starting point. Note that at the intermediate temperature all estimates of anisotropy converge to the same value (3 per cent), and cost function becomes less than 0.01 for all starting points; at the low temperature (0.1), anisotropy estimates do not converge and the cost remains large for some starting points; at the high temperature (10) cost function remains high at any number of moves, and the estimates of anisotropy fluctuate with a large variance.

By using an expression similar to (4, 5) we calculate the theoretical  $T$  component of each SKS wave by using its observed  $R$  component and the corresponding theoretical transfer functions. Various samples of the SKS waveform differ by the back azimuth (baz) of the respective seismic events. In our notations the theoretical  $T$  component of SKS for a certain back azimuth is  $T_{syn}^{SKS}(t, baz)$ .

Let  $E_P(\mathbf{m})$  and  $E_{SKS}(\mathbf{m})$  be misfits for  $SF(t, \psi)$  and  $T_{syn}^{SKS}(t, baz)$ . 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  $E_{SKS}(\mathbf{m})$  simultaneously. We conduct a search for the optimum models by using an interactive algorithm, similar to Simulated Annealing (SA) (Mosegaard &

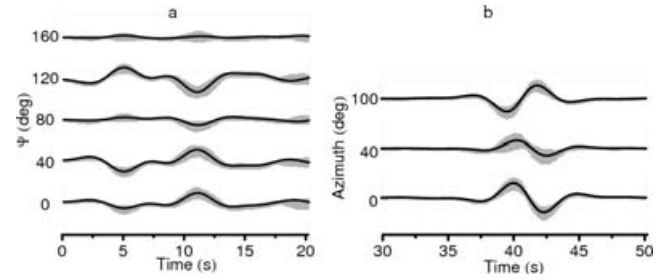


**Figure 6.** Inversion of the data (Fig. 3, upper row, and Fig. 4a) for model 2 (Table 2): the input  $SF$  (a) and  $T$  components of SKS (b) are shown by black lines; the synthetics for selected trial models are grey.

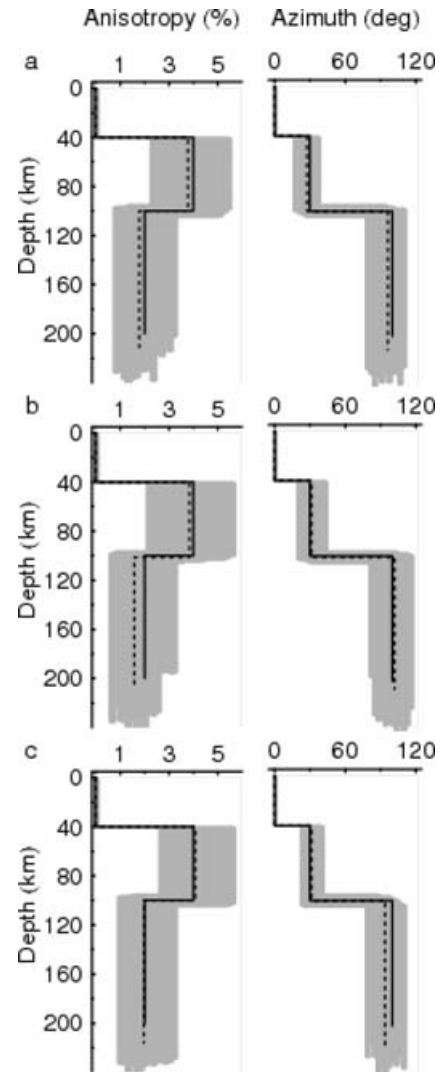


**Figure 7.** Results of inversion of the data for model 2 (Table 2): a – models for  $QF$  and  $TF$  (Fig. 3, upper row) and  $T$  components of SKS (Fig. 4a); b – models for the same data as in (a), but with noisy  $QF$  and  $TF$  components (Fig. 3, middle row); c – models for the same data as in (a), but for the tilted symmetry axis in the lower layer (bottom row in Fig. 3 and Fig. 4b). The actual parameters and the medians of the parameters found by inversion are shown by solid and dash lines, respectively. The models found by inversion are shown grey.

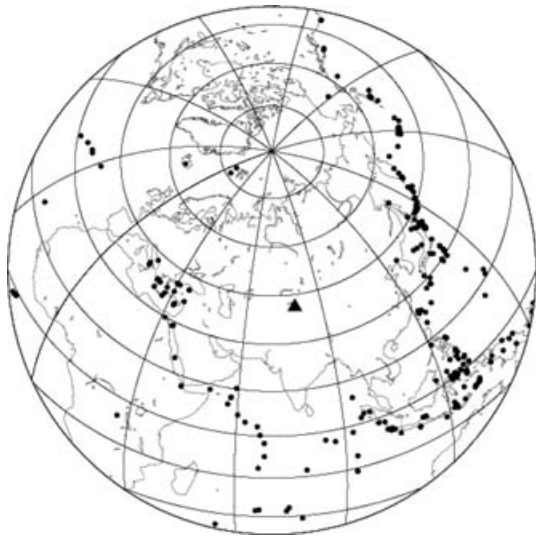


**Figure 8.** Epicentres of seismic events (filled circles) used in the calculations of the receiver functions of station WUS (triangle). Some events to the north of WUS at distances exceeding  $100^\circ$  and not shown.

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). If  $\mathbf{m}_c$  is the ‘current model’ then the ‘attempted model’  $\mathbf{m}_a$  is accepted when the condition  $E_i(\mathbf{m}_a) \leq E_i(\mathbf{m}_c)$  is satisfied, where  $i$  stands for ‘ $P$ ’ or ‘ $SKS$ ’. Otherwise the move is accepted with probability  $\exp(E_i(\mathbf{m}_c)/T_i - E_i(\mathbf{m}_a)/T_i)$ , where  $T_i$  is temperature.

We use the Metropolis rule in cascade (Mosegaard & Tarantola 1995) for the two cost functions. Each move in the model space consists of perturbing a randomly selected single component of vector  $\mathbf{m}$ . The perturbation is proportional to a random number, chosen uniformly at random between  $-1$  and  $1$ , multiplied by the length between prior bounds. The value of the proportionality coefficient (equal to  $0.1$  as a rule) should be small enough to ensure correlation between the successive values of the cost functions (Tarantola 2005). Temperature schedule is an essential problem of a practical application of the SA techniques. We use 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, converge to the same point for any  $\mathbf{m}_0$ .

We demonstrate the search process by using synthetics for the model 2 (Table 2). In addition to the synthetic  $SF(t, \psi)$ , we synthesized the  $R$  and  $T$  components of the  $SKS$  phase for three back azimuths:  $0^\circ$ ,  $40^\circ$  and  $100^\circ$  (Fig. 4). These synthetics were calculated for the initial model (a) and the model with a tilted symmetry axis (b). The adopted apparent velocity of  $SKS$  is equal to  $18.5 \text{ km s}^{-1}$ . The unknowns in this experiment are the parameters of two anisotropic layers:  $\delta V_s/V_s$ , azimuth of symmetry axis (fast axis)  $\alpha$  and depth of the layer bottom. We have found by numerical experiments that the cost function is insensitive to  $(\delta V_p/V_p)/(\delta V_s/V_s)$  and  $\eta$ , if they vary within reasonable bounds ( $\pm 0.2$  and  $\pm 0.1$ , respectively). Therefore we fixed  $\eta$  at the usual value of  $1.0$ , as in PREM (Dziewonski & Anderson 1981) and the ratio between  $(\delta V_p/V_p)/(\delta V_s/V_s)$  at  $1.5$ . The latter value is based on the analysis of published data for the upper mantle (Oreshin *et al.* 2002b). The number of trial models in each experiment is of the order of several thousands.

We show only the search process for  $\delta V_s/V_s$  in the upper layer and only for  $SF(t, \psi)$ . In the case of low temperature (bottom row in Fig. 5) the solution depends on starting point  $\mathbf{m}_0$ . At high temperature (upper row) the search for different  $\mathbf{m}_0$  does not converge to the same solution. The optimum temperature (middle row) provides convergence to, roughly, the same value of the unknown. Similar regularities are observed for the other unknowns and for the  $T$  component of  $SKS$ .

In a general case, the inversion procedure may require several temperature iterations. At each 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 sufficiently small level. Since we minimize two cost functions simultaneously, the two temperatures are adjusted in parallel. In our case, the predefined levels ( $0.01$  for the receiver functions and  $0.1$  for  $SKS$ ) are reached at the first temperature iteration. We verify that the resulting synthetics are indeed close to the input  $SF(t, \psi)$  and  $T^{SKS}(t, \text{baz})$  (Fig. 6). Medians of the corresponding parameter distributions (Fig. 7a) are close to the input model. Similar results (Fig. 7b) are obtained for the noisy receiver functions (Fig. 3b) and for the receiver functions in Fig. 3(c) and  $SKS$  waveforms in Fig. 4(b) for the tilted symmetry axis (Fig. 7c). The results look similar, because the errors of inversion are large relative to the expected effects. Note that the inversion is performed by assuming a horizontal position of the symmetry axis. These experiments suggest that the results of the inversion are insensitive to the noise in the receiver function and tilt of the symmetry axes. In other words, the method is robust.

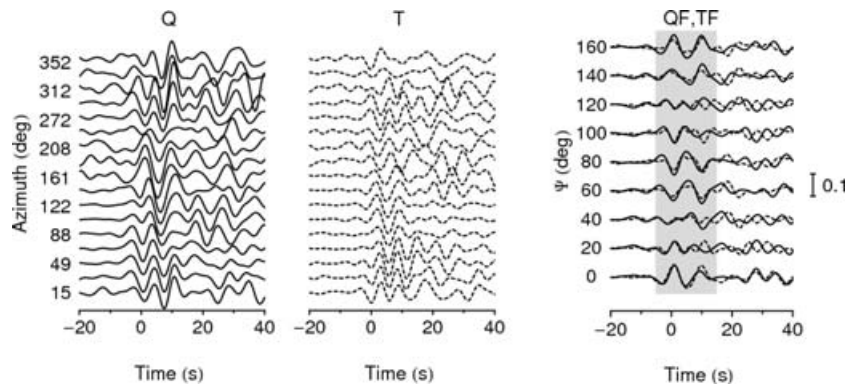


**Figure 9.**  $Q$ ,  $T$ ,  $QF$  and  $TF$  components of the receiver functions of station WUS.  $Q$  and  $T$  components are stacked in azimuthal bins. Mean values of the back azimuth for each bin (in degrees) are shown on the left.

#### 4 DATA AND RESULTS

The method described above was applied to the recording of almost 40 broadband stations in the Tien Shan. To reduce the contribution of randomly scattered waves in the coda of the  $P$  wave, the raw records were low-pass filtered with the corner period around 5 s.

Anisotropic models are obtained for 10 stations (Fig. 1). The GHENGIS network (Roecker 2001), consisting of 29 stations, operated for about 18 months. This time span is too short for obtaining a good azimuthal distribution of the receiver functions, because large earthquakes are rare to the north and west. From this network, satisfactory data are provided only by stations AKSU and CHAT. The other 8 stations belong to the long-term networks: IRIS/GSN

(stations AAK and TLG), Geoscope (station WUS), KNET (stations CHM, USP, EKS2, ULHL and UCH). At 4 other stations of KNET, coherence between  $QF$  and  $TF$  was too low for the diagnostics of azimuthal anisotropy.

We present the data for station WUS in detail. The receiver functions of WUS, like those of the other selected stations, are from a broad range of back azimuths (Fig. 8), but the distribution of the recorded events is very inhomogeneous. In the azimuths from  $30^\circ$  to  $150^\circ$  the number of well-recorded events in the optimum distance range from  $30^\circ$  to  $90^\circ$  is very large, but in some other sectors the events in this distance range are few or totally missing. To fill the sector between  $320^\circ$  and  $360^\circ$ , we used  $PP$  waves at the distances more than  $100^\circ$ . In order to obtain a homogeneous distribution of

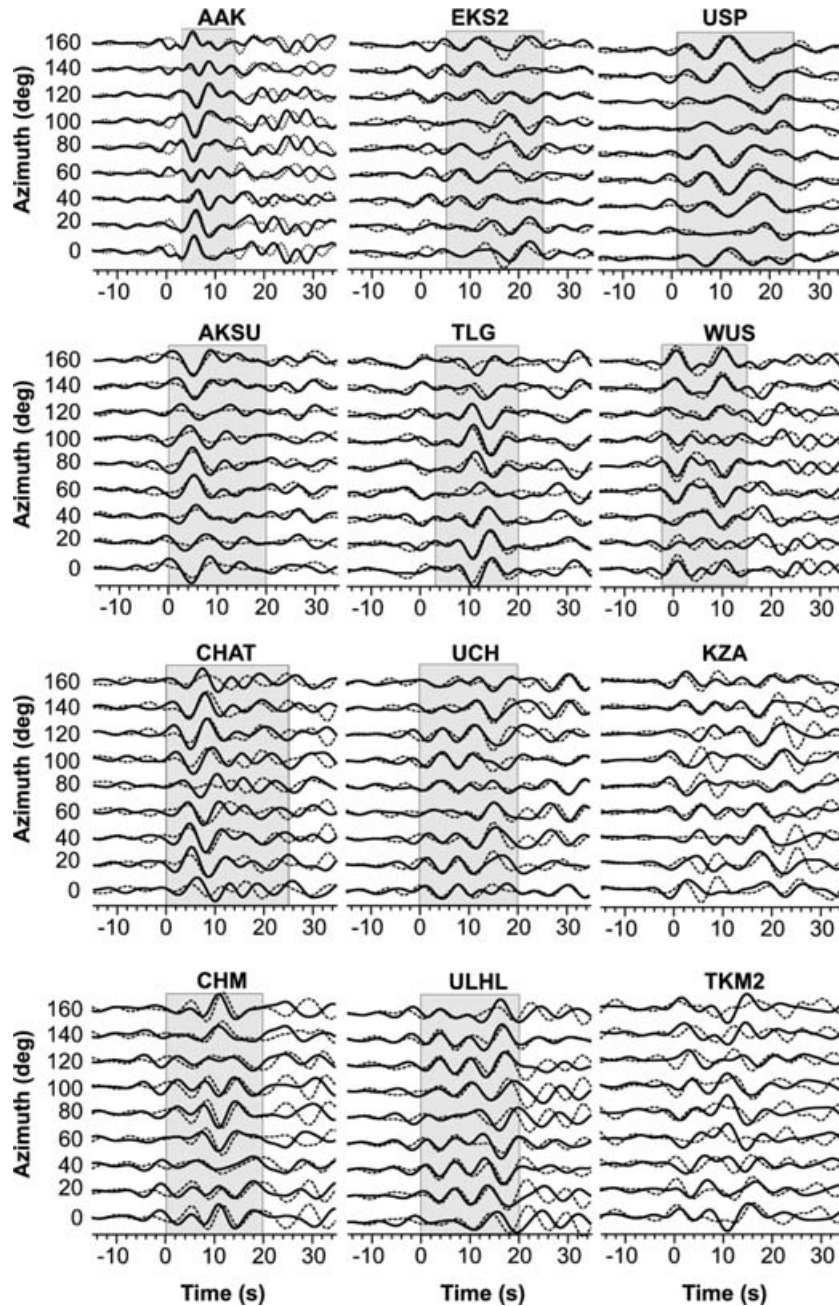


**Figure 10.**  $QF$  (solid lines) and  $TF$  (dash lines) components of 12 stations. Intervals of coherence between  $QF$  and  $TF$  are shown grey.  $QF$  and  $TF$  components of stations KZA and TKM2 are incoherent and shown for comparison.

receiver functions in the azimuth domain, the individual receiver functions were stacked in bins  $20^\circ$  wide. Secondary (reflected and converted) phases propagate with differing apparent velocities, depending on their depths of reflection/conversion, and generally the individual receiver functions should be stacked with appropriate move-out corrections. However, at the periods longer than 5 s these corrections are small and can be neglected. The average number of stacked receiver functions is around 10 per bin.

The  $Q$  and  $T$  components of the summary receiver functions of WUS are shown in Fig. 9. The resulting  $QF(t, \psi)$  and  $TF(t, \psi)$  (Fig. 9) are well correlated within the first 15 s. Similar correlation is observed in the data of other stations within the first 15–25 s (Fig. 10). The data for stations KZA and TKM2 of KNET network, where such correlation is absent, are shown in Fig. 10 for comparison. For each station we have a few good recordings of the  $SKS$  phase (see examples in Fig. 11). Most of them are in the back azimuths between  $90^\circ$  and  $105^\circ$ .

We try to obtain a good fit between the actual and synthetic waveforms of WUS with as few unknowns as possible. The data set of this station is the most complicated, because it requires anisotropy in the crust. At a number of other stations the crust is isotropic. The adopted isotropic model is a simplified version of the model, which

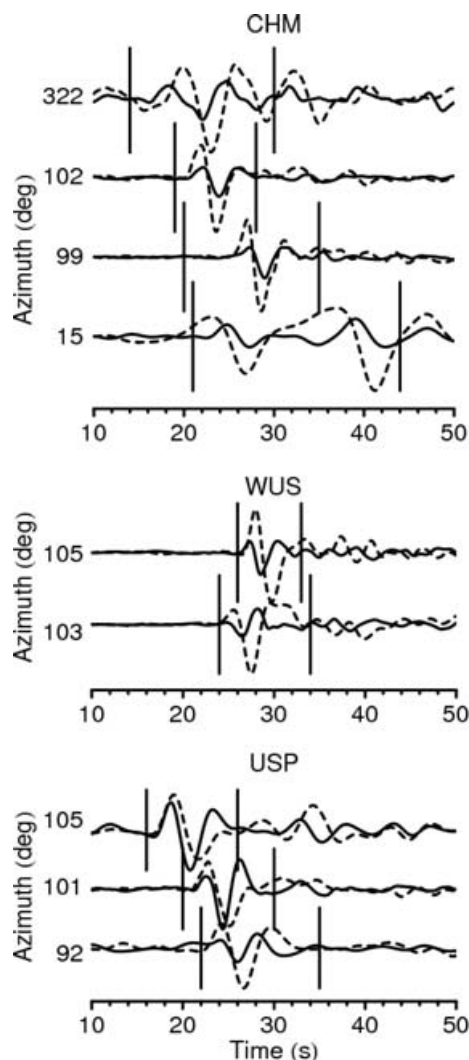


**Figure 11.** Examples of recordings of the  $SKS$  phase. The radial and transverse components are shown by dash and solid lines, respectively.

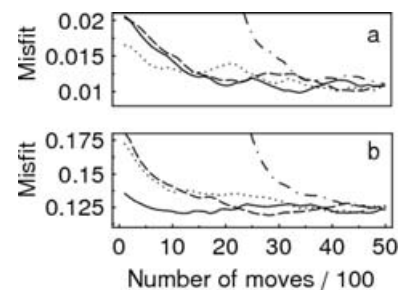


**Figure 12.** Inversion of the data for station WUS: misfit of  $SF$  (a) and  $T$  components of  $SKS$  (b) as functions of the number of moves.

was obtained by a joint inversion of the  $P$  and  $S$  receiver functions of this station and teleseismic travel time residuals of the  $P$  and  $S$  wave from the back azimuths between  $90^\circ$  and  $120^\circ$  (Vinnik *et al.* 2006). The crust consists of 4 layers; the thicknesses of the upper 3 layers are fixed; the thickness of the lowermost layer is allowed to vary between 16 and 22 km; the uppermost layer is isotropic. The upper mantle consists of 2 layers of unknown thickness, which can vary in the range from 20 km to 65 km (upper layer) and between 80 km and 185 km (lower layer). The other unknowns are  $\delta V_s/V_s$  and  $\alpha$  for the three crustal and two mantle layers. As a rule, the upper bounds for  $\delta V_s/V_s$  are 8 per cent in the crust and 6 per cent in the mantle. The values of  $\delta V_p/V_p$  and  $\eta$  are defined like in Section 3.

Temperature schedule for WUS consisted of two temperature iterations; we demonstrate only the 2<sup>nd</sup> iteration. During the 2<sup>nd</sup> iteration the cost functions reached sufficiently small values: 0.012 for  $SF$  and 0.12 for the  $T$  component of  $SKS$  (Fig. 12). We do not look for a better fit, because the related fine details of the actual waveforms are very likely caused by noise. The bounds for the unknowns in the 2<sup>nd</sup> iteration (Fig. 13) are narrower than the initial bounds in the first iteration. For example, in the crustal layers 2 and 3 the bounds for the azimuth of polarization of the fast wave are  $20^\circ$  and  $90^\circ$ , whereas the initial bounds were  $0^\circ$  and  $180^\circ$ . The search started from four randomly chosen models, all of which converged to nearly similar models. Approximation of the observed waveforms by the synthetics could be improved with a larger number of layers, but it would be impossible to get similar solutions for differing starting points. The search resulted in the data in Fig. 14 for  $SF(t, \psi)$  (a),  $T$  components of  $SKS$  (b) and models (c). In all acceptable models in Fig. 14 fast wave polarization azimuth in the upper subcrustal layer is around  $150^\circ$ ; in the lower layer it is around  $45^\circ$ . The depth of the boundary between the layers (90 km) is determined with the accuracy of several kilometres. We performed similar analysis assuming that the crust is isotropic. Then in the time interval corresponding to the crustal  $P_s$  phases the misfit of  $SF(t, \psi)$  is large, but, nevertheless, the values of the parameters of the mantle layers remain nearly the same.

The results for the other stations (Fig. 15) are of about the same quality, but anisotropic structures in the crust are either absent or relatively simple. The data can be explained by mantle models consisting of 2–3 anisotropic layers. The stations with 3 anisotropic layers are AAK, USP, TLG, EKS2 and ULHL, but at AAK, USP, TLG and EKS2 one layer is thin relative to the others ( $\sim 20$  km) and can be neglected. At ULHL the two upper layers can be combined into one. Then all mantle models can be characterized by the parameters of two layers. Most often, the depth of the boundary between the two layers is close to  $100 \pm 10$  km (CHM, USP, UCH, EKS2, AKSU, WUS, CHAT). The other parameters of the two-layer mantle models are summarized in Fig. 1.

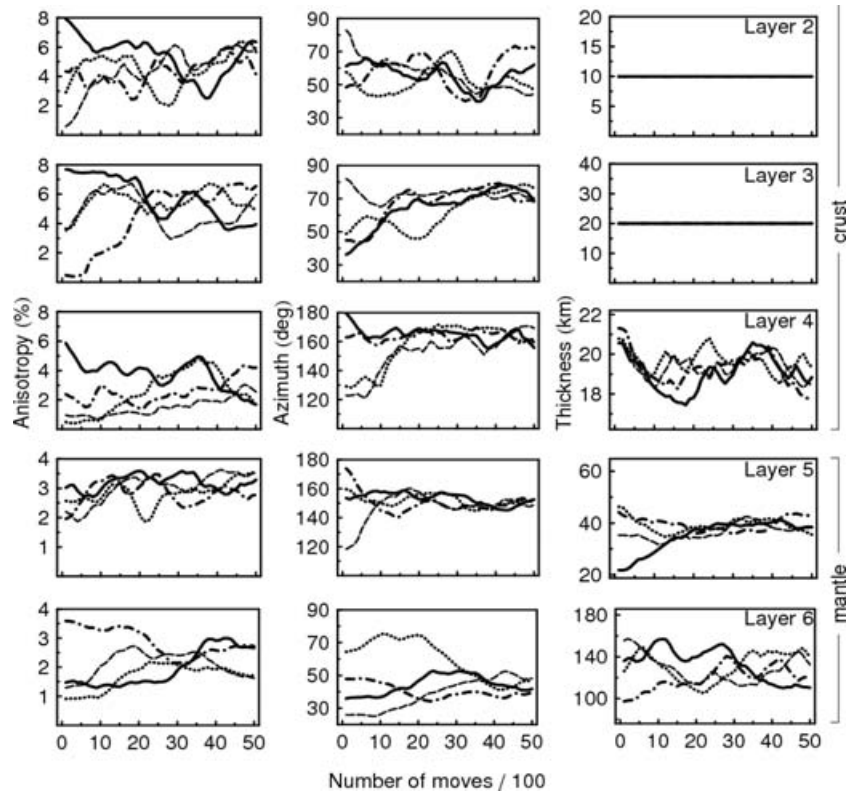


**Figure 13.** Unknowns of the model for station WUS as functions of the number of moves:  $\delta V_s/V_s$ , azimuth of fast axis and layer thickness are in the left, central and right column, respectively.

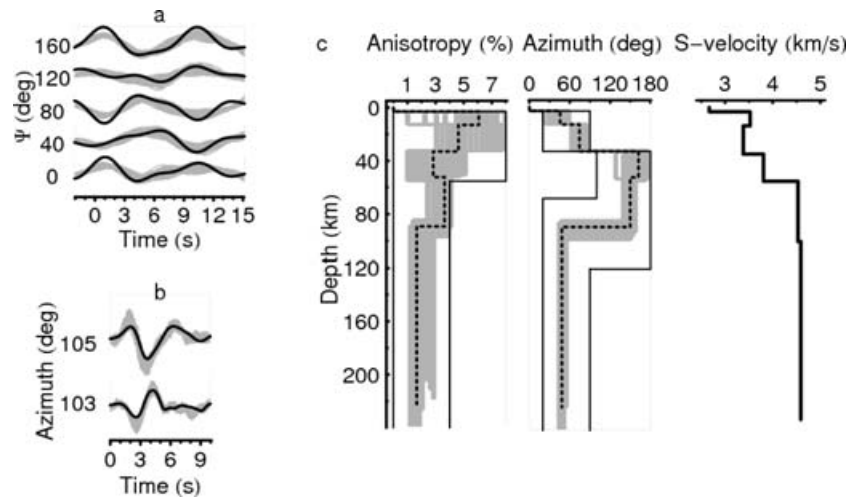


**Figure 14.** Results of inversion of the data for station WUS: a –  $SF(t, \psi)$  and the selected (corresponding to the cost less than 0.012 for  $SF$  and less than 0.12 for the  $T$  component of  $SKS$ )  $SF_{syn}(t, \psi)$  are shown black and grey, respectively; b – the same as (a), but for the  $T$  component of  $SKS$ ; c – selected models (grey); medians and *a priori* bounds for the search are shown by dash and solid lines, respectively; isotropic  $S$  velocity distribution is shown on the right.

Fast wave polarization directions in the upper layer at stations CHM and TLG are  $95^\circ$ . At the other stations the average azimuth of fast wave polarization is close to  $0^\circ$ . Magnitude of  $S$  wave anisotropy ( $\delta V_s/V_s$ ) varies in a range from 1 per cent to 5.5 per cent with the average around 3 per cent. In the lower layer, azimuth of fast wave polarization varies between  $30^\circ$  and  $90^\circ$  with the average of  $60^\circ$ . This average direction is practically parallel to the trend of the belt. Magnitude of  $S$  wave anisotropy in the lower layer displays

remarkable stability, and at all stations except WUS varies between 5 per cent and 6 per cent.

## 5 CONCLUSIONS

The method of our study is based not only on the  $SKS$  recordings but also on additional information, which is stored in the  $P$  receiver

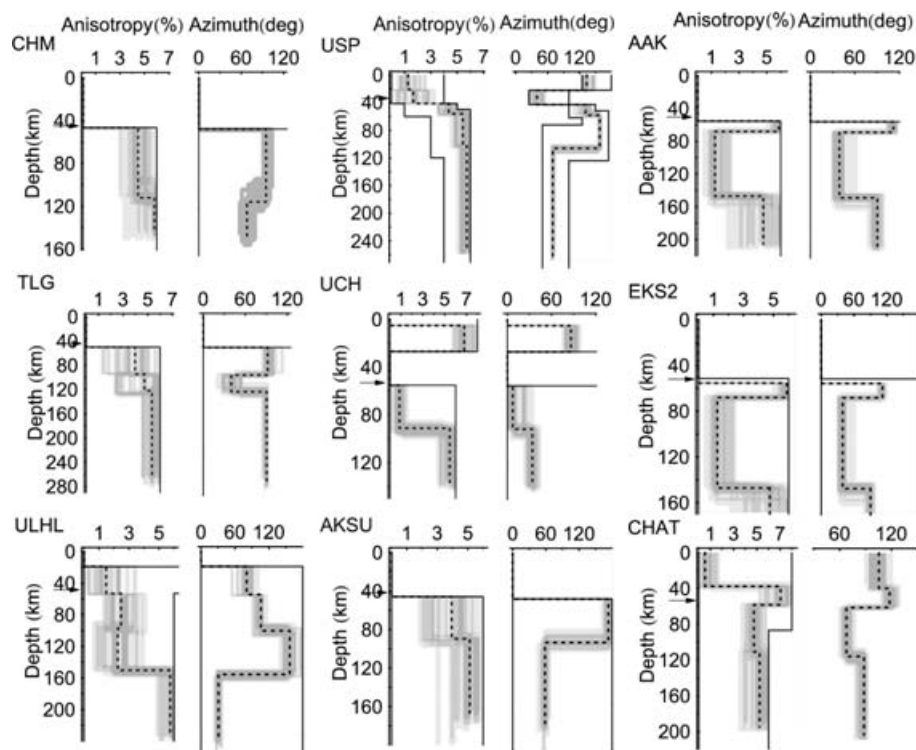


**Figure 15.** The same as in Fig. 14(c), but for the other stations and without isotropic  $S$  velocity models; the depths of the Moho are marked by arrows.

functions. Testing in Sections 2 and 3 demonstrates that this method is robust and able to resolve a few anisotropic layers in the crust and the upper mantle.

The results of application of this method to the data of 10 stations in the Tien Shan reveal a pronounced change in the parameters of anisotropy at a depth around 100 km. In the shallow mantle (at depth less than  $\sim 100$  km) anisotropy is relatively weak ( $\sim 3$  per cent) and polarization of the fast wave varies laterally in a wide range. At a larger depth,  $S$  wave anisotropy is about twice larger (usually, between 5 per cent and 6 per cent), and the azimuth of fast wave polarization is aligned with the trend of the belt (around 60 deg). These generalizations could not be obtained in the previous study of the Tien Shan (Vinnik *et al.* 2002), and they present a result of the improved methodology. The depth of 100 km roughly corresponds to the boundary between the high velocity mantle lid (seismic lithosphere) and the low velocity zone (seismic asthenosphere) beneath the Tien Shan (Oreshin *et al.* 2002a). Thus, seismic data reveal different patterns of seismic anisotropy in the mantle lithosphere and asthenosphere. The average heat flow in the Tien Shan is around 50 mW/m<sup>2</sup> (Duchkov *et al.* 2001), and the change of the pattern of anisotropy corresponds to the temperature of around 1000°C.

On theoretical grounds, in thrust belts, like the Tien Shan, deformations in the lithosphere can be described as a near-horizontal simple shear. This means that fast  $S$  wave should be polarized parallel to the shear direction that is normal to the belt. The  $S$  and  $SKS$  phases propagate in the direction, nearly normal to foliation planes in the rocks, and anisotropy should be relatively weak. This prediction is consistent with our results:  $S$  wave anisotropy in the mantle lithosphere is around 3 per cent, weak relative to the maximum value of around 8 per cent for a typical upper mantle composition (Tommasi *et al.* 1999). Large scatter of the fast wave polarization direction in this layer is a likely effect of errors of the estimates and of actual lateral variations. It is possible that at the relatively

low temperature in the lithosphere (on the order of several hundred degrees) a perfect lattice preferred orientation (LPO) requires larger deformations. If the deformations are large enough only in several narrow zones (similar to faults in the crust), part of the mantle lithosphere may preserve texture related to previous deformation episodes.

Fast wave polarization direction in the asthenosphere is aligned with the trend of the Tien Shan, in the azimuth of around 60°. This anisotropy is very likely an effect of horizontal uniaxial compression in the direction normal to the trend of the belt. Foliation planes corresponding to this deformation are vertical, facilitating the largest splitting in the  $SKS$  phase, and, in agreement with this prediction, anisotropy in the asthenosphere of the Tien Shan is twice larger than in the mantle lithosphere. At a high temperature in the asthenosphere (more than 1000°C), dynamic recrystallization accelerates LPO evolution (Ave Lallemant 1975; Tommasi *et al.* 1999), and significant anisotropy can be caused by smaller deformations. We note that the pattern of mantle anisotropy beneath and in the vicinity of the Tien Shan is not unique, and a similar two-layer structure is observed, for example, in Central Europe, where only anisotropy in the asthenosphere, at a depth exceeding 80 km, is clearly related to the Alpine orogeny (Farra & Vinnik 2000). A similar pattern is found beneath the Siberian platform (Oreshin *et al.* 2002b).

The alignment of the fast wave polarization direction in the  $SKS$  phase with the trends of collisional belts, contrary to the shear direction expected in the lithosphere, may lead to the conclusion that orogenic domains deforming by thrusting and crustal thickening are decoupled from the upper mantle (Tommasi *et al.* 1999). This conclusion depends on the implicit assumption that shear wave splitting in the  $SKS$  phase is an effect of wave propagation only in the mantle lithosphere. Our data demonstrate that this assumption is not correct, at least, for the Tien Shan. Similar studies may reveal deep fabric and patterns of mantle flow in other regions.

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