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*Seismic evidence of flow at the base
of the upper mantle*

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Seismic evidence of flow at the base of the upper mantle

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Abstract. A combined analysis of *SKS*, *SKKS* and *P660s* seismic phases detected in the records of BCAO/BNG seismic station in central Africa reveals that the waveform and amplitude of the *T* component of *P660s* deviate significantly from those predicted from the observations of shear wave splitting in *SKS* and *SKKS*. The inverse problem is investigated by using the anisotropic Thomson-Haskell code and a genetic algorithm. The discrepancy between *SKS/SKKS* and *P660s* can be explained by P-SH conversion at the boundaries of a weakly anisotropic layer 30 - 40 km thick lying atop 660 km discontinuity beneath BCAO/BNG. This anisotropy is probably induced by horizontal flow at the bottom of the upper mantle, in the transition zone.

Introduction

The form of convection operating in the mantle has been the subject of debate for over 20 years. Recent tomographic images of subduction zones favour whole-mantle convection models. There are seismic observations, however, which suggest that seismic azimuthal anisotropy can be present in a thin layer lying either above [Vinnik and Montagner, 1996] or on both sides [Vinnik and Kosarev, 1997] of 660 km discontinuity far outside subduction zones. Anisotropy in the mantle is commonly explained by finite strain [Karato, 1998], and azimuthal anisotropy is indicative of a flow along rather than across the discontinuity.

Indications of azimuthal anisotropy near 660 km depth have been obtained by comparing particle motions in *P660s* seismic phase (converted from *P* to *S* at 660 km discontinuity) with those predicted from the observations of *SKS* and *SKKS*, converted from *P* to *S* at the core mantle boundary. The discrepancies between the two kinds of particle motions were explained by conversion from *P* to *SH* at the boundaries of an anisotropic layer at the base of the upper mantle. However, a systematic comparison between the observations of *SKS/SKKS* and *P660s* has been conducted so far only at two seismic observatories: GRF in central Europe [Vinnik and Montagner, 1996] and OBN within the East-European platform [Vinnik and Kosarev, 1997]. The data are few because the seismic signal of interest can only be detected either with an array of broadband sensors or at a single station with extremely simple (plane-layered) subsurface structure and a low signal-generated noise in the time window of interest. In this paper, we present results for a station of such kind in

central Africa (4.4 S, 18.5 E; GDSN station BCAO from 1984 till 1990; Geoscope BNG from 1988 till present).

Observations of *SKS* and *SKKS*

To make observations of *SKS/SKKS* and *P660s* directly comparable, we processed *SKS/SKKS* at BNG with a technique [Vinnik *et al.*, 1989], similar to that applied to *P660s*. The idea is to standardize the records by deconvolving the radial (*R*) and transverse (*T*) components of *SKS/SKKS* by the *R* component and to stack the deconvolved components of many records with weights depending on back azimuths of the events. The weights W_i are defined as

$$W_i = \cos 2(\phi_i - \psi) / \sum_{i=1}^n \cos^2 2(\phi_i - \psi) \quad (1)$$

where ϕ_i is back azimuth of the *i*-th event, *n* is the number of events and ψ is a variable parameter. On the usual assumption of hexagonal anisotropy with a horizontal axis of symmetry, the maximum amplitudes of the stack are obtained for the values of ψ which differ from the fast direction of anisotropy by $\pm 45^\circ$. This, however, is not necessarily true for other symmetries. In general terms, there may be both $\cos(2\phi)$ and $\cos(\phi)$ components in the signal [Levin and Park, 1997].

Locations of the seismic events are shown in Figure 1a. In the selected records, signal/noise ratio in the *T* component, as a rule, is higher than 3. In some recordings, *SKS* and *SKKS* are well separated in time, and we stacked a total of 25 deconvolved waveforms (Figure 2a). The maximum amplitude of the stack is observed for $\psi = 75^\circ$. The corresponding azimuth of fast velocity ($30 \pm 10^\circ$) deviates from the value of $50 - 60^\circ$ obtained previously [Vinnik *et al.*, 1992] by using data of lower quality and a different data processing algorithm. The discrepancy is mainly an effect of a better quality and a larger quantity of the presently available data. The new estimate of the fast direction is close to the absolute plate motion direction, in agreement with the regularity previously noted in several other regions. In Figure 2b,

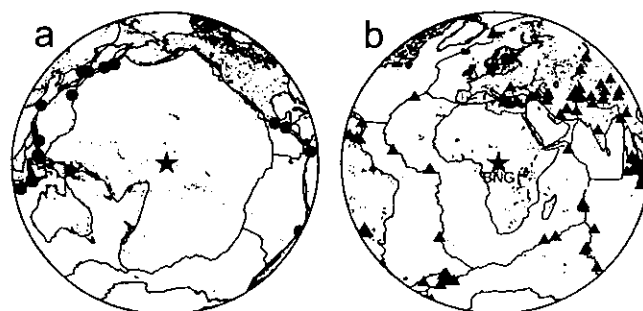


Figure 1. Locations of the events, used to analyse (a) *SKS/SKKS* and (b) *P660s*. In (a), the star is in the antipodal position relative to BNG/BCAO.

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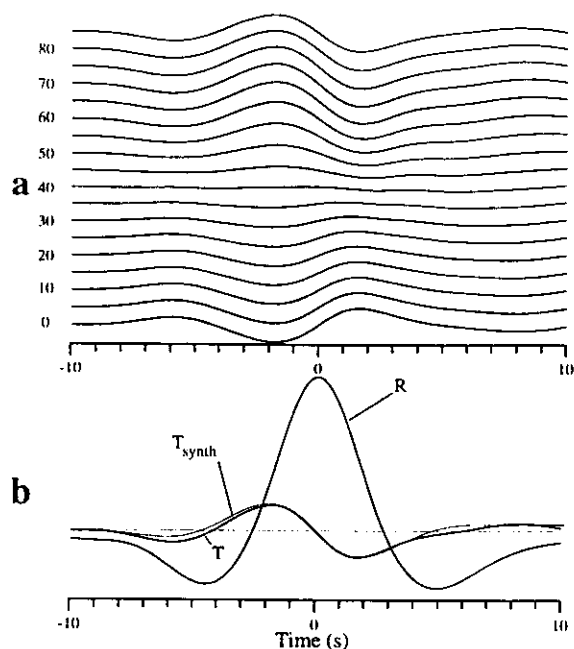


Figure 2. Effects of anisotropy in *SKS* and *SKKS*: a - *T* components deconvolved by the *R* components and stacked with weights; the values of ψ in degrees are shown on the left; b - Deconvolved and stacked without weighting *R* components, deconvolved and stacked *T* components for $\psi = 75^\circ$, and stacked synthetic *T* components for the same ψ .

the *T*-component stack for $\psi = 75^\circ$ is compared with the *R*-component stack and the related stack for the synthetic *T* components. The synthetics are calculated with the aid of Thomson-Haskell algorithm for plane waves in plane-layered anisotropic medium [Kosarev *et al.*, 1979], assuming hexagonal anisotropy with a horizontal axis of symmetry and using the *R* component signal in Figure 2b as an input. Amplitude of the *T*-synthetic is proportional to the lapse time δt between the split waves, and the assumed 0.6 s provides good fit between the observed and theoretical signals. In agreement with theoretical predictions, both theoretical and observed *T* component waveforms are close to the derivative of the *R* component waveform. The obtained estimate of δt is equal to that reported previously.

Observations of P660s

For the analysis of the *Ps* mantle phases, we have selected 97 low-noise records in a broad range of azimuth and distance (Figure 1b). The technique of data processing is similar to that described by Kosarev *et al.* [1984], Vinnik and Montagner [1996], and Vinnik and Kosarev [1997]. Each record is projected on the *P*, *SV* and *T* axes, and the *SV* and *T* components are deconvolved by the *P* component. The deconvolved *SV* and *T* components are weighted and stacked with moveout corrections. Figure 3a shows the broad-band *SV* stack. This stack is remarkable by very short periods (up to about 1-2 s) preserved in *P410s* and *P660s*. These short periods indicate, that a plane-layered model is a good approximation for the earth medium beneath the station. Otherwise, a pronounced topography on the discontinuities would destroy correlation between

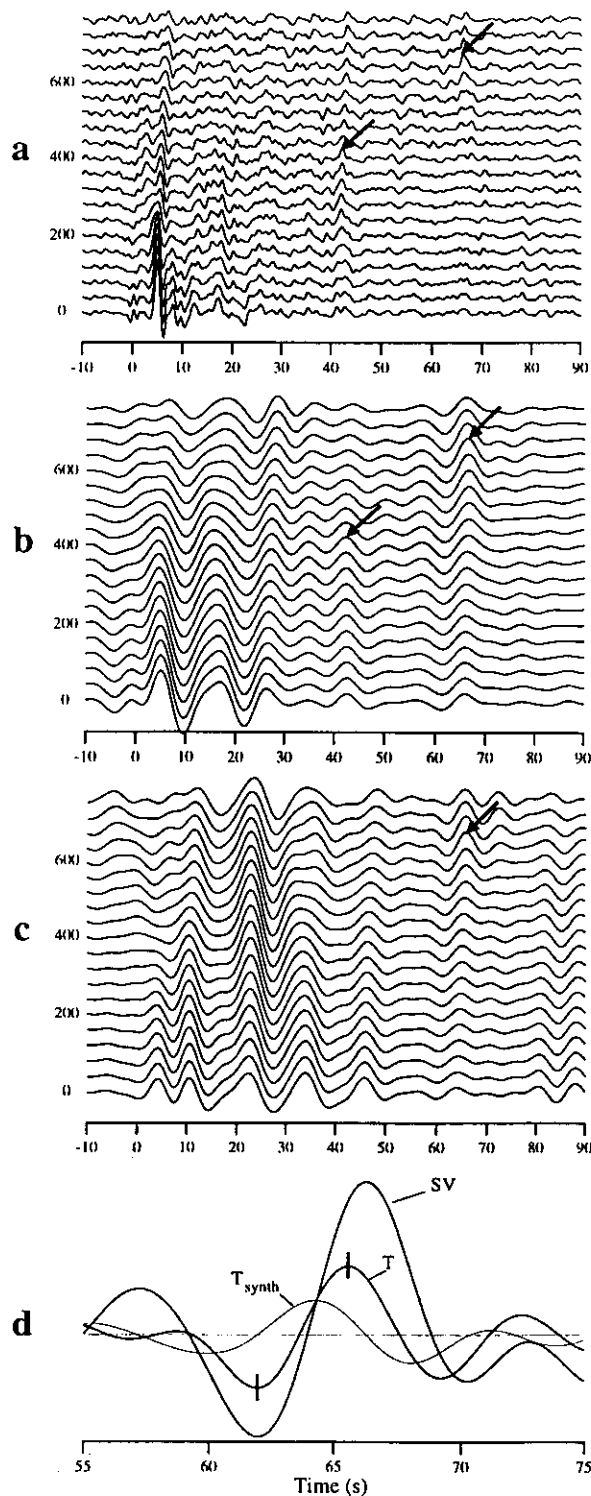


Figure 3. Results of stacking: broad-band *SV* without weighting (a), low-pass filtered *SV* without weighting (b), and low-pass filtered *T* for $\psi = 80^\circ$ (c). Each trace corresponds to a trial depth of conversion, indicated in km on the left. The detected mantle *Ps* phases are marked by arrows. d - *SV* and *T* components from (b) and (c) for 660 km trial depth. Standard errors of *T* are shown by vertical bars. The synthetic *T* component (T_{synth}) is calculated on the assumption that anisotropy with the parameters derived from observations of *SKS/SKKS* is in the uppermost mantle.

Table 1. Model for synthetics in Figure 4

Layer	H	V_p	V_s	ρ	$\delta V_s/V_s$	η	α°
1	30	8.50	4.70	3.45			
2	150	8.50	4.70	3.45	0.020	1.03	40
3	290	8.50	4.70	3.45			
4	40	8.50	4.70	3.45	0.015	1.03	140
5		9.15	5.00	3.80			

H is layer thickness in km. Parameters in the bottom line are for the halfspace.

the individual signals at short periods and would make such observations impossible. This simple structure is favourable for observing the effects of anisotropy in the T -component, but the signal-generated noise still is prohibitively strong at periods shorter than about 6 s. Therefore, for the subsequent analysis the records were low-pass filtered with a corner period at around 6 s. The records of SV and T were stacked with the same weights as T in SKS and $SKKS$.

Figure 3b shows the same stack of the SV component, as in Figure 3a, without weighting but low-pass filtered, and Figure 3c demonstrates the low-pass filtered T -component stack for $\psi = 80^\circ$. This value of ψ is close to the optimum value for SKS (Figure 2a). Amplitude of the T component at 0 s for this and other values of ψ is extremely small. This implies that side refraction of the recorded P waves is negligible, which confirms the hypothesis of plane-layered structure un-

derneath the station. In the SV stack in Figure 3b, a strong 660-km signal can be observed, a weaker 410-km signal, and crustal phases arriving earlier. In the stacked T component, a clear cut signal is present at a time of $P660s$. The largest amplitudes of this signal are observed at appropriate trial depths (around 660 km). This is a criterion to distinguish true Ps phases from artifacts of processing. The amplitudes of the other signal-looking phases in the T -component are in a sharp disagreement with their times. In particular, the largest amplitudes of the phases recorded at around 50 s and 80 s are observed at small trial depths, which is characteristic of multiple reflections.

In Figure 3d, both components of the 660-km signal from (b) and (c) are shown zoomed and superimposed. Standard errors of the T component are found with the aid of bootstrap resampling [Efron and Tibshirany, 1991]. The synthetic T component (T_{synth}) in Figure 3d is calculated for the anisotropic layer in the uppermost mantle, with the parameters derived from the $SKS/SKKS$ data. The synthetic is calculated with the same technique as in Figure 2b, but for a larger incidence angle characteristic of the $P660s$ phase. There are statistically significant differences between the synthetic and the real data: the observed T component signal is almost twice stronger, and its phase shift relative to the SV component is much smaller. The latter effect is clearly expressed in a similar time ($62.0 \pm 0.3s$) of the negative extrema in the two components of the observed signal, contrary to 2.0 s shift in the synthetic. The difference between the real T component and the synthetic can be evaluated especially easily at 62.0 s and 66.2 s, where the synthetic amplitude is 0. The amplitude of the residual thus evaluated differs from that of the observed T component by only 10%.

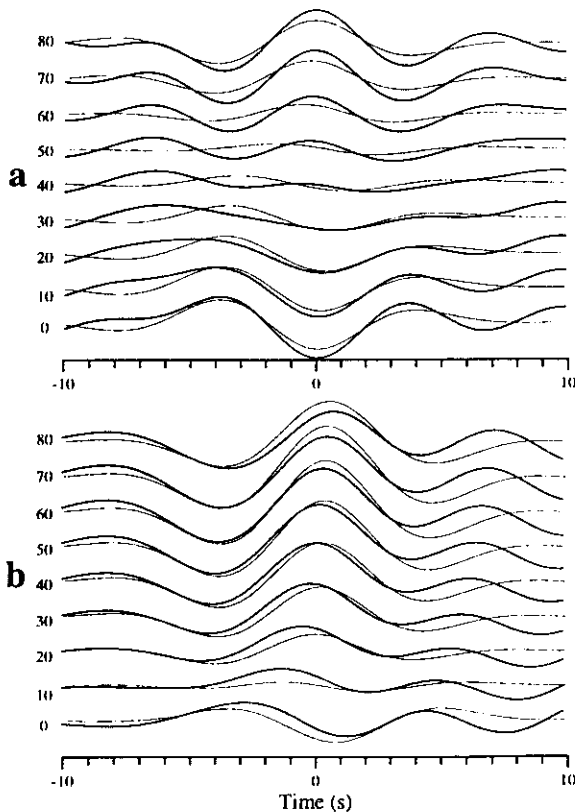


Figure 4. Observed $P660s$ (bold lines) for a trial depth of 660 km, and the related synthetics for model in Table 1, as functions of ψ : a - T ; b - SV . The values of ψ in degrees are shown on the left.

Inversion of $P660s$ particle motion

There are two conceivable explanations for the properties of the T -component of $P660s$. One possibility is $P-SH$ scattering at bumps and dips of 660 km discontinuity, but this is hard to reconcile with the results of stacking in Figure 3a. Moreover, the T component signal of the required strength can be generated by topography of the discontinuity with an amplitude of about ± 15 km [van der Lee et al., 1994]. In stable tectonic regions, such as central Africa, such large amplitudes are unlikely, and the related scattered signals should be smaller by an order of magnitude.

The other explanation, proposed in a few previous studies, is conversion from P to SH at the boundaries of an anisotropic layer near 660 km discontinuity. Variations of the T component signal at 660 km trial depth,

Table 2. Parameters found with a genetic algorithm

α_1°	$H(km)$	$\delta V_s/V_s$	η	α_2°
25	40	0.021	1.024	134
30	38	0.022	1.024	134
35	34	0.017	1.040	137
40	35	0.017	1.016	134

α_1 and α_2 are fast directions in the upper and the lower anisotropic layer. The other parameters are related to the lower layer.

to be modelled, are shown as a function of ψ in Figure 4a. A similar weighting and stacking has been conducted for the SV component (Figure 4b). Synthetic seismograms were calculated using Thomson-Haskell algorithm. It is assumed that anisotropy is hexagonal with a horizontal axis of symmetry. Isotropic component of the model is defined by velocities V_p , V_s and density ρ . Anisotropic component is defined by δV_p , δV_s , η , and α . Among them, only the difference between fast and slow S velocities δV_s , η (without clear physical meaning), and azimuth of fast velocity α are constrained by the observations. To avoid numerical noise, the model (Table 1) is made simple, consisting of a few layers. 660 km discontinuity is imitated by discontinuity between 4th layer and the half-space. Anisotropy is present in 2nd and 4th layers. The parameters of 2nd layer provide δt of 0.7 s for split $SKS/SKKS$. Apparent velocity is fixed at 12.4 km/s. This value is lower than the real values, but the related angle of incidence of the P wave at the discontinuity in the model is around 45° , close to the actual values. The V_s step at the discontinuity in the model (0.3 km/s) and the related amplitudes of the Ps phase (around 4% of the P wave amplitude) are also close to the actual values. The stack of the deconvolved P wave records was used as an input. The same back azimuths were adopted for the synthetics, as for the real data, and the synthetics were stacked with weights, like the real data. An acceptable set of the anisotropic parameters found by trial and error is specified in Table 1, and the related synthetics are superimposed on the real data in Figures 4a and 4b.

A search for other models was conducted with the aid of a genetic algorithm [Sambridge and Drijkoningen, 1992]. During the inversion, the fast direction in the upper layer (α_1) was allowed to vary between 20° and 40° . Anisotropic parameters of the 4th layer and its thickness H were allowed to vary in a broad range of values. The rms difference between the real data and synthetics was minimized adopting 0.5 and 1.0 weights for the SV and T components, respectively. A smaller weight was adopted for the SV component, because it is affected by both, isotropic and anisotropic parts of the wavefield, and its amplitude is generally twice larger. Four best sets of the variable parameters thus found are listed in Table 2. All of them are close to those found by trial and error.

Discussion

We have presented data indicating weak azimuthal anisotropy in a thin layer at the bottom of the upper mantle. This conclusion is in agreement with the results of a few other studies. Transverse isotropy at this depth in a global average model [Montagner and Kennett, 1996] can be a result of averaging laterally varying azimuthal anisotropy. Observations of splitting in the shear phases of deep events sometimes lead to a conclusion that the lower part of the mantle transition zone is isotropic [Fischer and Wiens, 1996]. However, if the anisotropic layer is so thin and anisotropy so weak, as in our models, its contribution to the lapse time between the split waves is less than 0.2 s, which is hard to detect with a conventional analysis of splitting. Moreover, this anisotropy can be confined to certain geographical regions rather than distributed continuously. Most gen-

erally, our results can be interpreted as indications of horizontal flow along 660 km discontinuity. Low viscosity in the corresponding depth range [Forte et al., 1993] could be among the main reasons for this. Unfortunately, the data are still too sparse to discuss the regularities in the direction of flow, its strength and geographical extent. In the future, mapping anisotropy in the transition zone may provide constraints on the mass flux between the upper and the lower mantle.

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