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Upper and lower mantle anisotropy inferred from comprehensive SKS and SKKS splitting measurements from India

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ABSTRACT

In this study, we investigate the upper mantle anisotropy beneath India using high quality SKS and SKKS waveforms from 382 teleseismic earthquakes recorded at 119 broadband seismic stations. In addition, we present evidence for anisotropy in the D″ layer beneath southeast Asia using SKS and SKKS splitting discrepancies on the same seismogram. During this exercise, we obtain 200 new splitting measurements from 35 stations recently deployed in the Indo-Gangetic plains (IGP), central India and northeast India. While the delay times between the fast and slow axes of anisotropy (δt) range from 0.3 to 1.7 s, the fast polarization azimuths (Φ) at a majority of stations in the IGP and central India coincide with the absolute plate motion of India implying shear at the base of the lithosphere as the dominant mechanism for forging anisotropy. However, stations in NE India reveal fast polarization azimuths mainly in the ENE–WSW direction suggestive of lithospheric strain induced by the ongoing Indo-Eurasian collision. Our analysis for D″ anisotropy yielded a total of 100 SKS–SKKS pairs, which can be categorized into those exhibiting (I) null measurements for one phase and significant splitting for the other phase, (II) null measurement for both the phases, (III) significant splitting for both the phases. A pair is considered to be anomalous if the splitting difference between SKS and SKKS is $\geqslant$0.5 s and the individual split time is $\geqslant$0.5 s. Using this criterion, we obtain 12 measurements under category III and 9 under category I that show a null measurement for SKS and large splitting for the SKKS phase. Further, we quantify the strength of the lower mantle anisotropy by correcting the SKKS measurement for the upper mantle anisotropy obtained by the SKS phase on the same seismogram. The SKS delay times are found to be consistently less than SKKS times, suggesting that the SKS phases do not capture the lower mantle anisotropy in comparison to their SKKS counterparts. Seven coherent measurements thus obtained reveal measurable D″ anisotropy, with fast polarization azimuths oriented mainly in the ENE–WSW direction. These results suggest presence of a large region of deformation in the lowermost mantle beneath southeast Asia. A possible model for anisotropy in these regions could be the presence of slab material that pounded upon the core mantle boundary (CMB) and is experiencing large shear deformation, resulting in lattice preferred orientation (LPO) of the lower mantle (Van der Hilst and Kárason, 1999; Long, 2009). The other possibility is the phase transformation from $MgSiO_3$ perovskite to a more stable post-perovskite phase under favorable conditions, which results in LPO of the lower mantle.

1. Introduction

Understanding the depth distribution and the causative mechanisms for seismic anisotropy enhances our knowledge of the past and present geodynamic processes in a tectonic region. Although most seismological studies reveal that anisotropy mainly exists in the upper mantle above the 600 km discontinuity, it is opined that the Lehmann discontinuity serves as a boundary between an anisotropic and an isotropic mantle (Revenaugh and Jordan, 1991), where pressure induced changes result in deformation of olivine from dislocation creep to diffusion creep (Karato and Li, 1992). However, there is growing evidence that anisotropy may exist even deeper under typical geothermal gradients (Jip et al., 1994) or in subducting slab regions (Fouch and Fischer, 1996). The presence of anisotropy within and below the transition zone was reported by Vinnik et al. (1996a), with possible explanations of a convective boundary layer between the upper and lower mantle or horizontal flow at the bottom of the upper mantle. Below the transition zone, many studies indicate an anisotropic D″ layer (e.g., Vinnik et al., 1989, 1995a; Kendall and Silver, 1996; Maupin et al., 2005), above the core–mantle boundary. The inner

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core was also found to be anisotropic owing to preferred orientation of anisotropic iron crystals, but reasons for such a development are still unclear (Song, 1997).

In the upper mantle, seismic anisotropy arises primarily from the strain-induced lattice-preferred orientation (LPO) of olivine, which is its dominant mineral (e.g., Nicolas and Christensen, 1987). The fast polarization direction of anisotropy (Φ) tends to align parallel to the *a*-axis of olivine. Anisotropy in the upper mantle may result from both past and current deformation. In cratonic regions, there is a debate whether the symmetry axis is parallel to past geological features due to anisotropy frozen in the lithosphere (Silver and Chan, 1991; Silver, 1996) or to the present day absolute plate motion (APM) as a result of asthenospheric flow Vinnik et al., 1992, 1995b, 1996b). Silver and Chan (1991) and Silver (1996) convincingly argue that the fast axis of anisotropy measured above older continental regions tends to be parallel to the surface structural features, suggesting that the anisotropy may have been formed at the same time as those features due to coherent lithospheric deformation from past orogenic episodes. However, Vinnik et al. (1992) point out that if the anisotropy frozen in the lithosphere gets annihilated at depths exceeding the 900 °C isotherm, as much as 5% shear wave anisotropy in the lithosphere would be required to explain the observed splitting measurements for older continents. A single hypothesis may not suffice to explain these splitting observations, necessitating contributions from both the lithosphere and asthenosphere (Bormann et al., 1996; Savage, 1999; Fouch et al., 2000; Fouch and Rondenay, 2006).

The first shear-wave splitting measurements in the Indian continent suggest anisotropy closely aligned with the APM of the Indian plate (Ramesh and Prakasam, 1995; Ramesh et al., 1996). Subsequently, null measurements found at stations HYB and SHL were interpreted in terms of an isotropic Indian plate (Chen and Ozalaybey, 1998; Barruol and Hoffmann, 1999). Later studies that utilized larger data sets sampling diverse geological provinces of the Indian continental shield established dominance of plate motion related strain (Singh et al., 2006, 2007; Kumar and Singh, 2008; Heintz et al., 2009; Kumar et al., 2010; Saikia et al., 2010; Mandal, 2011; Roy et al., 2012) (see Kumar and Singh, 2010, for review). Further, these studies also brought out evidence of fossilized anisotropy in the proximity of shear zones in the southern granulite terrain and the Cuddapah basin (Kumar and Singh, 2008; Roy et al., 2012). The N–S orientations along the western coast of India are related to the edge flow associated with a transition from thicker to a thinner lithosphere (Kumar and Singh, 2008). Within the Himalaya and its foredeep, the source of anisotropy is attributed to lithospheric deformation due to finite strain induced by collision, rather than fossilized anisotropy (Singh et al., 2006; Heintz et al., 2009). Very recently, an exhaustive analysis of 471 splitting measurements from 50 stations situated in the northwestern Deccan Volcanic Province (DVP) brought out two distinct layers of anisotropy, a stronger one oriented in the APM direction and the weaker one in the E–W direction associated with strain related to large scale magmatism in this plume affected region (Rao et al., 2013).

Numerous studies within the Tibetan plateau located to the North of Himalaya suggest a change in the mode of deformation within the lithosphere (McNamara et al., 1994; Hirn et al., 1995). McNamara et al. (1994) observe a northward increase in the delay times along a north-south oriented array of eleven broadband stations with the fast polarization azimuths progressively varying from NE to E–W direction, parallel to the surface geological features that deform in response to the India–Eurasia collision. The variation in polarization azimuths in the Tibetan–Himalayan collision zone can be classified into three patterns (Hirn et al., 1995). To the South, on the Indian plate where the collision effects are expected to be small, the anisotropy is orientated NE, parallel to the plate motion. Within the Tethyan Himalaya, the orientations are aligned with the NNW differential motion between India and Tibet, probably due to the imbrication of lithosphere, resulting in crustal thickening and elevation. To the North of the Indus–Yarlung suture (IYS), the observed S-wave delays and large splitting are presumably related to the elevated temperatures in the crust and mantle. The source of anisotropy in the Tibetan plateau is dominated by the finite strain induced by the left lateral shearing of the Tibetan mantle lithosphere, owing to the India–Eurasia collision (Davis et al., 1997; Holt, 2000). Huang et al. (2000) observe null splitting beneath the Himalaya, IYS and the southernmost Lhasa terrane, and relate it to the existence of an LPO fabric with a subvertical axis of symmetry or lack of a coherent LPO fabric in the underthrusting Indian lithosphere beneath Tibet. The latter possibility was reiterated by the finding of an isotropic Indian plate (Chen and Ozalaybey, 1998). Small delay times or null splitting are observed around the IYS, southern part of central Tibet and also in the region where Tibet is underlain by the Indian mantle lithosphere (Zhao et al., 2010). The Indian lithosphere is located below the entire western Tibet, while in the central and eastern Tibet it is located beneath the IYS (Li et al., 2008).

Various studies reveal that the D″ layer, which is a ~200–300 km thick layer above the core–mantle boundary (CMB) (Weber et al., 1996; Wysession et al., 1998; Garnero, 2000), is seismically highly heterogeneous compared with the lower mantle above it (Mitchell and Helmberger, 1973; Lay and Helmberger, 1983; Young and Lay, 1990; Ritsema et al., 1997). The first global tomographic model of the Earth's lower mantle obtained using P-wave travel times (Dziewonski et al., 1977; Dziewonski, 1984; Inoue et al., 1990) does reveal velocity anomalies in excess of 1% at the CMB, indicative of the D″ region. However, this was not observed globally. Also, it is observed that an anomalous D″ layer mostly occurs in faster than average velocity regions, away from the Ultra-Low Velocity Zones (ULVZs) (Williams et al., 1998). The ULVZs are regions with 30% reduction in S-wave velocity near the base of the CMB having a thickness of 5–20 km (Helmberger et al., 2000; Thorne and Garnero, 2004). These are interpreted as evidence for partial melts and can be detected by examining the strength of scattering of the shear waves that sample the CMB (Williams et al., 1998; Vidale and Hedlin, 1998). Several tomographic models consistently show pervasive low velocity anomalies in the D″ under the central Pacific and western Africa, and high velocity anomalies under the Pacific rim, Eurasia and Caribbean (Li and Romanowicz, 1996; Masters et al., 1996; Grand et al., 1997; Bijwaard and Spakman, 1999; Van der Hilst and Kárason, 1999; Garnero, 2000). While the high velocity anomalies are interpreted in terms of subducted slabs, the low velocity zones (LVZs) are explained in terms of upwellings or superplumes, thermal anomalies and chemical heterogeneity within the D″ (see Garnero, 2000). Another tomographic model utilizing diffracted P-wave data in addition to the direct P-phases (Zhao, 2001, 2004) brings out a good correlation between the distribution of slow anomalies at the CMB beneath central Pacific and Africa with the hotspots on the surface or within the mantle, suggesting that most of the strong mantle plumes under the hotspots originate from the CMB (Zhao, 2004). Several geophysical studies provide evidence for the presence of compositionally distinct mantle domains in the bottom 1000 km of the lower mantle, which is also proposed for eastern Asia (Van der Hilst and Kárason, 1999). This is perhaps related to local iron enrichment and transformation of silicate-to-oxide, indicating a complex mantle convection (Van der Hilst and Kárason, 1999). The absence of anisotropy in the lower mantle is due to deformation by superplasticity, with the exception of the D″ layer (Karato et al., 1995). This was also observed through seismological observations (e.g., Meade et al., 1995; Panning and Romanowicz, 2004). Lateral velocity variations and lowermost mantle anisotropy