Invited Review

A review of crust and upper mantle structure beneath the Indian subcontinent

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ABSTRACT

This review presents an account of the variations in crustal and upper mantle structure beneath the Indian subcontinent and its environs, with emphasis on passive seismic results supplemented by results using controlled seismic sources. Receiver function results from more than 600 seismic stations, and over 10,000 km of deep seismic profiles have been exploited to produce maps of average crustal velocities and thickness across the region. The crustal thickness varies from 29 km at the southern tip of India to 88 km under the Himalayan collision zone, and the patterns of variation show significant deviations from the predictions of global models. The average crustal shear velocity (Vs) is low in the Himalaya–Tibet collision zone compared to Indian shield. Major crustal features are as follows: (a) the Eastern Dharwar Craton has a thinner and simpler crustal structure crust than the Western Dharwar Craton, (b) Himalayan crustal thickness picks clearly follow a trend with elevation, (c) the rift zones of the Godavari graben and Narmada–Son Lineament show deeper depths of crust than their surroundings, and (d) most of the Indian cratonic fragments, Bundelkhand, Bhandara and Singhbhum, show thick crust in comparison to the Eastern Dharwar Craton. Heat flow and crustal thickness estimates do not show any positive correlations for India. Estimates of the thickness of the lithosphere show large inconsistencies among various techniques not only in terms of thickness but also in the nature of the transition to the asthenosphere (gradual or sharp). The lithosphere beneath India shows signs of attrition and preservation in different regions, with a highly heterogeneous nature, and does not appear to have been thinned on broader scale during India’s rapid motion north towards Asia. The mantle transition zone beneath India is predominantly normal with some clear variations in the Himalayan region (early arrivals) and Southwest Deccan volcanic province and Southern granulite terrain (delayed arrivals). No clear patterns on influence on the mantle transition zone discontinuities can be associated with lithospheric thickness. Over 1000 anisotropic splitting parameters from SKS/SKKS phases and 139 using direct S waves are available from various studies. The shear-wave splitting results clearly show the dominance of absolute-plate-motion related strain of a highly anisotropic Indian lithospheric mantle with delay times between the split S phases close to 1 s. There are still many parts of India where there is, at best, limited information on the character of the crust and the mantle beneath. It is to be hoped that further installations of permanent and temporary stations will fill these gaps and improve understanding of the geodynamic environment of the Indian subcontinent.

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http://dx.doi.org/10.1016/j.tecto.2015.01.007
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1. Introduction

The Indian subcontinent is formed of a mosaic of various Precambrian tectonic provinces, with stable shields in peninsular India to actively deforming collision belts in the Himalaya, and has experienced extensive volcanism and rifting. India lies on a fast moving plate and has covered a large distance since its separation from the other components of Gondwana (ca 130 Ma). The influence of the fast drift on the stability of cratons, and removal of lithospheric roots are key issues which are much debated (Kumar et al., 2007), but as yet are not fully understood.

In the century since the detection of the Mohorovičić discontinuity (Mohorovičić, 1910) from earthquake observations, both controlled source and passive seismic studies have made impressive advances in understanding the nature of the crust and uppermost mantle (Prodehl et al., 2013). Multiple facets of seismic wave propagation can be brought to bear on the structure of the Earth’s interior, and help to resolve the key issues related to evolution and nature of the continental crust and upper mantle. To date there have been only limited attempts to provide a full picture of the Indian crust and upper mantle. There have been reviews of heat flow (Roy and Rao, 2000) and deep seismic sounding studies (Kaila and Krishna, 1992; Reddy and Rao, 2013). However, the full range of available information on the crust and upper mantle available from passive source studies have not previously been exploited.

The foundation stones of seismology in India were laid by the pioneering works of Dr. T. Oldham and Dr. R. D. Oldham, the father-son duo. The great Shillong earthquake of 12th June, 1897 is well documented and reported in the works of Oldham (1899). This deadly Shillong earthquake achieved the maximum intensity XI on MM scale (Richter, 1958), and provided the impetus for a series of initiatives to install seismographs in India to monitor earthquakes. The first few installations were made of Milne’s self registering seismographs in Alipore (Calcutta, now Kolkata), Colaba (Bombay, now Mumbai) and Madras (now Chennai) (Tandon, 1992). An Omori-Ewing seismograph was installed in Simla as a response to the great Kangra earthquake of 5th April, 1905. In the years from 1929 to 1930, the country was equipped with a few more Milne-Shaw seismographs, initially installed at Colaba observatory Mumbai, then Bombay and later at few more places in Agra, Calcutta, Hyderabad and Kodaikanal. In the early 1960s five World Wide Standard Seismograph Network (WWSSN) stations were installed at various places across the country following the recommendations of Berkner (1959). After the devastating Latur earthquake of September 30th, 1993 the India Meteorological Department upgraded ten of its observatories to the standard of Global Seismograph Network, and later complemented this network with 14 more broadband stations during 1999–2000. At present the India Meteorological Department runs nearly 80 seismic stations in the national network, supplemented by various temporary networks operated by other organizations. Temporary and permanent networks in different parts of India have been operated by the National Geophysical Research Institute, Indian Institute of Technology Bombay, Wadia Institute of Himalayan Geology, Tezpur University and the Institute of Seismological Research. The National Geophysical Research Institute has established more than 200 broadband seismic stations at various points of time, and so plays a major role in passive source seismology in India.

Deep seismic probing of Indian crust, started in 1972 with refraction/wide-angle reflection work, but subsequently was dominated from the early nineties by deep seismic reflection. A good deal has been achieved (Kaila and Krishna, 1992), with more than 10,000 km of profiles carried out in various experiments using controlled sources. A major supplementary source of information on Indian structure comes from the use of seismic receiver functions exploiting the recordings of distant earthquakes. Receiver functions provide a tool to map the Earth’s response beneath a single three-component seismic station, and extract information on the seismic discontinuities at depth from the conversions and reverberations associated with the main seismic phases. The first receiver functions for the Indian region used data from the Hyderabad station (HYB) in India, using P-to-s converted waves (Gaur and Priestley, 1997). Since then the role of receiver functions in determining crust and upper mantle discontinuities (Moho, lithosphere–asthenosphere boundary, mantle transition zone discontinuities 410 and 660) has been routine practice. Further information comes from seismic anisotropic studies using SKS/SSKS phases and heat flow that provide links to help understand both geodynamics and structure. The present work presents as a complete picture of the Indian crust and upper mantle as possible, compiled from various sources with emphasis on passive source seismic datasets. We synthesize results from seismic studies, heat flow and seismic anisotropy to develop a comprehensive map of the properties of the crust and upper mantle beneath the Indian subcontinent, with links into the Himalaya and Tibet to provide a wider perspective and understanding of the whole region.

2. Tectonic setting

The major tectonic units of peninsular India comprise Precambrian terranes (Fig. 1). A vast region in between the peninsula and the actively deforming regions of Himalaya and Tibet is covered by quaternary sediments. These sediments, mainly of Himalayan origin, form the Indo-Gangetic plains with very thick sedimentary deposits (≥8 km).

The western central portion of India is overlain by flood basalts known as the Deccan Traps or the Deccan Volcanic Province (DVP). The Indian plate has crossed over various hotspots (Réunion, Krozet, Kerguelen and Marion) in its rapid transit to the north. The passage over the Réunion hot spot (Chenet et al., 2007) has led to a major volcanic event, which resulted in creation of the Deccan Traps. The flood basalts are of considerable thickness (>1.5 km) and cover a region of more than 500,000 km². Recent results from Deep Scientific Drilling in the Koyana region provide direct estimates of a 931 m thick basaltic layer followed by a paleoregolith of thickness of 4 m (Rao et al., 2013). The Cambay Rift (CBR) divides the Deccan Traps into two distinct units, one in the northwest and the other in the southwest. The Cambay Rift is filled with tertiary sediments, and is interpreted as a failed rift formed due to extensional tectonics.

The other major rift systems are the Godavari Graben (GG) and the Mahanadi Rift (MHR), which are passive in nature but which have left
clear imprints on the surface. The approximately east–west Narmada–Son Lineament (NSL), which originated in the Archean, is another prominent feature that divides various tectonic zones to the north and south (Meert et al., 2010). The southernmost part of the Indian peninsula is termed the Southern Granulite Terrain (SGT). This consists of high-grade granulites of late Archean to Neoproterozoic age, traversed by various shear zones (Meert et al., 2010).

The Indian landmass is host to a number of Proterozoic basins with significant sedimentation namely the Vindhyan Basin (VNB), Chattishgarh Basin (CTB), Godavari Graben and Cuddapah Basin (CB) related to various collisional events. The Godavari Graben and Cuddapah Basin, are of considerable importance as they throw light on the collision events related to Antarctica and India since the Proterozoic. The southern margin of the Singhbhum craton and the eastern margin of the Bastar Craton (BC) are girdled by the Eastern Ghats Belt (EGT), a Proterozoic granulite terrane widely considered to have formed during orogenic collisions between eastern India and east Antarctica (Dobmeier and Raith, 2003).

The Indian craton is formed of a mosaic of a few smaller cratons (Taylor et al., 1984). These stable parts of the Indian shield include the Dharwar Craton, Bastar Craton, Singhbhum Craton (SC), Bundelkhand Craton (BUC) and Aravalli Craton (part of the Delhi–Aravalli Fold Belt, DAFB). These cratons are basically low to high-grade crystalline rocks formed by intense deformation and metamorphism in the Precambrian. The Dharwar Craton is separated into the Eastern Dharwar Craton (EDC) and Western Dharwar Craton (WDC) by the north–south trending Closepet Granite (CG) (Fig. 1).

Northeast India under the influence of the Himalayan collision is a region of complexity, at one side are the Indo-Burmese Ranges and at the other are the collisional belts of the Himalaya, in between are the Shillong Plateau (SP) and the Mikir Hills (MKH). The Himalaya–Tibet collision zone formed due to collision of Indian and Eurasian plates, and is comprised of various blocks which are separated by suture zones (Indus–Tsangpo Suture Zone, ITSZ; Bangong–Nujiang Suture Zone, BNSZ). South of Indus–Tsangpo Suture Zone, lies the Himalaya, which is bounded by various north dipping Cenozoic faults systems.
like the Main Boundary Thrust (MBT) and Main Central Thrust (MCT). The region between Indus–Tsango Suture Zone and Bangong–Nujiang Suture Zone, is known as the Lhasa terrane, adjacent to the Quiangtang terrane (Yin and Harrison, 2000), while the Quiangtang terrane is situated between the Bangong–Nujiang Suture Zone and the Jinsha Suture Zone (JSZ) (Fig. 1).

3. Crust and crust–mantle boundary

The seismological definition of the Moho is linked to the rapid rise in seismic wavespeeds between the crust and the mantle (Prodehl et al., 2013). The transition from crust to mantle is not always sharp, and there can be differences in the interpretation of the seismological and petrological definitions of the base of the crust (O’Reilly and Griffin, 2013). However, seismological results provide the most comprehensive coverage of the nature of the crust and its boundary with the mantle including constraints on the variation of physical properties, such as density and seismic wavespeed, with depth.

Many of the results for the Indian subcontinent come from crustal models extracted from receiver functions (Langston, 1979) beneath three-component seismic stations, from the Indian shield to the actively deforming regions of Himalaya and Tibet. Before we discuss the crustal models, we provide a brief discussion of receiver functions as a passive method, since this approach in recent decades has provided most of the information on the major discontinuities beneath a seismic station. The idea of exploiting teleseismic P wavesforms for crustal information starts with Phinney (1964) who exploited the Fourier spectral amplitude ratios of these waves recorded at various World-Wide Standard Seismograph Network (WWSSN) stations. Receiver functions can be computed both in the frequency (Kurita, 1973; Phinney, 1964) and time domain (Burdick and Langston, 1977; Jordan and Frazer, 1975; Langston, 1979), with various advantages and limitations. Much of the work has followed (Langston, 1979), and used frequency domain deconvolution procedure, where the numerically unstable spectral division is taken into account by applying water level (Clayton and Wiggins, 1976). Later a time domain inversion technique was developed by Owens et al. (1984) and successfully implemented by various workers with certain modifications to obtain the one-dimensional velocity structure (Kind et al., 1995; Owens, 1987; Owens and Zandt, 1985). However, the issues of non-uniqueness and non-linearity in the inversion of receiver functions were demonstrated by Ammon et al. (1990), with complex trade-offs between velocities above interfaces and the depth of apparent conversion. The methods have been developed significantly in recent decades. The weak conversions in individual receiver functions can be enhanced by applying move-out corrections and stacking receiver functions at a single station (Kind and Vinnik, 1988; Vinnik, 1977). More complex structure can be considered by inverting waveforms for layer dip (Zhang and Langston, 1995), or even to recover both anisotropic and dipping effects (Frederiksen and Bostock, 2000; Levin and Park, 1997). However, these more complex schemes increase the non-uniqueness in the inversion of the receiver functions.

The first receiver functions were produced for India using data from Hyderabad station (HYB) located in Eastern Dharwar Craton (Gaur and Priestley, 1997), the only station in the International Federation of Digital Seismograph Networks (FDSN) operating in India since 1989. Thereafter the exploitation of receiver functions to extract crustal discontinuities and Poisson’s ratio has become routine with every new installation of broadband seismic station in India. Up until April 2014 data from ~442 seismic stations have been used to obtain estimates of crustal thickness and Poisson’s ratio for different geological provinces of the Indian sub-continent. Most of these profiles and seismic stations are focused on understanding the geometry of crustal discontinuities beneath the Indian shield (Borah et al., 2014a; Gaur and Priestley, 1997; Gupta et al., 2003; Jagadeesh and Rai, 2008; Julià et al., 2009; Kayal et al., 2011; Kumar et al., 2001, 2012; Mohan and Kumar, 2004; Rai et al., 2003, 2005; Sarkar et al., 2003; Saul et al., 2000; Tiwari et al., 2006) or north east India (Hazarika et al., 2012; Kumar et al., 2004; Mitra et al., 2005; Ramesh et al., 2005). Efforts are now being made by various workers in the Himalayan region along different seismic profiles in order to understand the Moho geometry of the Indian plate beneath the Himalaya (Acton et al., 2011; Caldwell et al., 2013; Hazarika et al., 2013a; Rai et al., 2006; Singh et al., 2010).

In this review we attempt to summarize the full set of results based on thickness and average velocities of the crust for the Indian subcontinent, exploiting passive seismic studies and deep seismic sounding (DSS), and links to other information such as Heat flow. We include crustal information for Tibet from the compilation of Li et al. (2014) to provide a more complete picture of the collisional structures to the north. We present a number of maps of different aspects of crustal structure. These have been created using a nearest neighbour interpolation scheme with a search radius of 100 km (Wessel and Smith, 1998). Where discrepant values of properties occur in close proximity we have carefully assessed the data and included only the most likely results. All values of the crustal parameters are reported in Supplementary material (Table S1), and excluded data are specifically marked.

3.1. Average crustal velocities

Maps of average crustal velocity and Poisson’s ratio from passive seismic studies are presented in Fig. 2. The average velocity models employed are the end results of the following: (a) 1-D inversion schemes mostly on receiver function stacks for a particular seismic station, (b) use of the approach of Zhu and Kanamori (2000) for average Poisson’s ratio and Moho depth, and (c) a few instances where data have been inverted for velocities incorporating both dipping and anisotropic effects (Sherrington et al., 2004; Singh et al., 2010). In most cases, the emphasis is on Vs rather than Vp. Frequently Vp is fixed from previous results of deep seismic sounding, standard velocity models, or seismic tomography.

There are significant variations in the average crustal velocities and Poisson’s ratio across the Indian subcontinent, with some very clear trends (Fig. 3). The regions of the Himalaya and Tibet show significantly lower Vs (~3.57 km s⁻¹) than in the Indian shield (~3.7–3.75 km s⁻¹) (Fig. 2). The Himalayan foredeep seismic stations, and stations in the Indo-Gangetic alluvium plains have an intermediate range with an average Vs of ~3.64 km s⁻¹. Out of 323 seismic stations in the Himalaya and Tibet, 200 seismic stations have velocities ≤3.6 km s⁻¹, and 86 even lower than 3.5 km s⁻¹. Although the stations are not uniformly distributed, the results are significant and clearly indicate a low velocity crust in the collision zone. The observed low velocities are characteristics of orogens (Christensen and Mooney, 1995), and most likely represent a felsic-to-intermediate nature for the crust. Such low shear wavespeeds are also reported in earlier compilations for Tibet and surrounding regions (Chen et al., 2010; Stolc et al., 2013).

In northeast India, the seismic stations located over the Shillong Plateau (centred around 92° longitude and 25.5° latitude) can be distinguished from the neighbouring regions by relatively high shear velocities >3.7 km s⁻¹. The shield-like character of Shillong Plateau has been reported by various workers (Kumar et al., 2004; Mitra et al., 2005). In the Indian shield itself, of the 120 seismic stations which fall in the regions of Dharwar Craton, Deccan Volcanic Province, Bundelkhand Craton, Singhbhum Craton, Bastar Craton and Southern Granulite Terrain, only 25 show velocities less than 3.6 km s⁻¹. The observed high velocities in the regions of Indian shield in comparison to the Himalaya are clear and obvious. Seismic stations on the west coast of India have somewhat lower crust shear wavespeed than in the shield (Fig. 2b).

A more subtle difference appears between the Eastern Dharwar Craton (Vs ~ 3.71 km s⁻¹) and the Western Dharwar Craton (Vs ~ 3.73 km s⁻¹). The difference is almost negligible given the uncertainties (Fig. S1). A higher velocity crust beneath the Western Dharwar Craton and a lower velocity crust beneath Eastern Dharwar Craton is reported in a number of studies (Borah et al., 2014a; Gupta
In the Himalaya–Tibet collision zone, the average Poisson’s ratio is 0.26, with strong variations over range from 0.16 to 0.32. A total of 107 stations show Poisson’s ratio < 0.25, while 189 have a ratio ≥ 0.25. The skew is clearly seen on the sides of the zone of high Poisson’s ratio in Fig. 2a. Examining the distribution based on nodal points (sampling of 0.25° × 0.25°), we find ~80% of the region has higher Poisson’s ratio, with most centred around 0.26. With a sampling of 1° × 1° the region with Poisson’s ratio ≥ 0.25 is even greater (~92%). These results provide a better understanding of the region, in view of the uneven station distribution (Fig. 2a). The high Poisson’s ratio beneath most of the Tibet plateau suggests the presence of partial melts/liquids, as interpreted in various studies (Li et al., 2006, 2009; Owens and Zandt, 1997).

Fig. 3 illustrates summary properties of the crust for specific geological regions: (a) Himalaya and Tibet, (b) the cratonic regions and Southern Granulite Terrain, (c) the rift systems of the Narmada–Son Lineament and Godvarī Graben and (d) the Deccan Volcanic Province. The strong contrasts between the orogenic belts of the Himalaya and Tibet and peninsula India is very evident in Fig. 3. We can also recognise the area masked by the Deccan Volcanic Province as having strong cratonic affinities.

3.2. Moho depth

Variations in the depth to the crust are large across the Indian region, ranging from a minimum of 29 km to a maximum of 88 km in depth. The shallowest crust is for stations close to the coastal plains of the peninsular India, and the deepest under the highest mountain ranges of Himalaya. The crustal thickness variation presented in Fig. 4 represents results from 615 seismic stations from various studies (Fig. 4). The regions of southern Indian shield and Himalaya–Tibet collision zone...
are well covered, while gaps exist in central India, mainly in the Indo-Gangetic belts and Bastar and Singhbhum Cratons.

In the Dharwar Craton a clear cut difference is observed between the different components. The Eastern Dharwar Craton exhibits a thinner crust, ~35 km, while in the Western Dharwar Craton average crustal thickness estimates are ~45 km. An interesting and unusual observation is a very thick crust, in the central portion of Western Dharwar Craton (Gupta et al., 2003). One of the reasons suggested, is that the thickened crust beneath a mid-Archaean (3.4 Ga) greenstone belt has not been subjected to any major deformation, and may preserve an ancient crust of around 60 km thickness (Gupta et al., 2003). The same seismic station (MTP, Table S1) has been reported with a crustal thickness of 45 km (Julià et al., 2009), which is preferred for creating the map (Fig. 4).

The crust of the Eastern Dharwar Craton with a sharper Moho is thinner and more transparent than the Western Dharwar Craton, with an absence of any intracrustal layer (Sarkar et al., 2003).
of any seismically distinct mafic cumulates overlying the Moho and an intermediate to felsic nature of Eastern Dharwar crust, rather than complex nature of Western Dharwar Craton crust was reported by almost all workers. Receiver functions for the Western Dharwar Craton stations are more complex and show a gradational transition from crust to mantle (Gupta et al., 2003; Sarkar et al., 2003). The Southern Granulite Terrain, has a relatively thick crust (~42 km), close to that for the Western Dharwar Craton.

An interesting observation is that the crustal thickness increases as we advance in the Himalayan collision zone, stations in the Himalayan regions have crustal thickness in the range of 40–50 km, while the central Tibet exhibits a highly thickened crust (around 75 km). A few regions such as the Tarim Basin and the Qaidam Basin clearly stand out from the Tibetan plateau with a thinner crust (Fig. 4).

3.3. Heat flow

Ongoing deformation and the presence of large scale inhomogeneity across regions are well reflected in the geothermal heat distribution for India and its environs. Previously, several studies have established a positive correlation between heat flow and crustal thickness (e.g., Bodri and Bodri, 1985), while many others have shown no obvious relationship exists between these two parameters (e.g., Mareschal and Jaupart, 2013). Using heat flow results, from the global data base of the International Heat Flow Commission, with additional new measurements (Nagaraju et al., 2012), we examine the correlation with crustal thickness for India and Tibet.

The regions with heat-flow values lower than the average value, ~60 mW m⁻², are regarded as low-heat-flow and those with heat-flow above the average above are classified as high-heat-flow (Fig. 6). Continental heat flow exhibits lower values in stable regions compared to active provinces. Moreover there are some discrepancies observed in correlation of surface heat flux with crustal thickness when comparing between tec-tonically active regions having high heat flux (~65 mW m⁻²) with anomalously thin (~30 km) or thick (~55 km) crust and stable regions having relatively low surface heat flux (~65 mW m⁻²) with little variations in crustal thickness (Mareschal and Jaupart, 2013). This is also apparent in our case where a distinct correlation exists between high-heat-flow and thick crust in the Himalayan collision zone as well as in Tibet. Certain parts of peninsular India the Cambay basin, the Godavari Graben and

Fig. 4. Crustal thickness map produced using estimates from receiver functions. A search radius of 100 km has been used to initialize the gridding with a nearest neighbour interpolation scheme, also employed in the subsequent figures. A total of 834 data points from 615 stations have been used to create the image. The location of the broadband seismic stations are marked by inverted triangles, colour coded by the original depth estimates. The data sets for the Indian region are from Acton et al. (2011), Borah et al. (2014a), Gupta et al. (2003), Hazarika et al. (2013a), Jagadeesh and Rai (2008), Julà et al. (2009), Kayal et al. (2011), Kumar and Mohan (2014), Kumar et al. (2001), Kumar et al. (2004), Li et al. (2014), Mandal (2012), Mitra et al. (2005), Mitra et al. (2008), Pathak et al. (2006), Rai et al. (2003, 2005, 2009a), Sarkar et al. (2003), Saul et al. (2000), Singh et al. (2010, 2012), Tiwari et al. (2006), and Vinnik et al. (2007), supplemented by results from Tibet (Li et al., 2014; Sherrington et al., 2004 and references therein). This figure employs data points from broadband receiver functions, and the crustal surface is interpolated using a search radius of 100 km.
some parts of the Singhbhum craton show higher heat flow, which does not link with their crustal thickness. The thinner crust of the Qaidam Basin correlates well with a low heat flow value (Fig. 5). Thick crust mostly relates to high heat flow values, but no correlation has been found for normal to thin crust. This may be due to the diversity in crustal compositions as well as the evolution of the heat production in the crust (Mareschal and Jaupart, 2013).

3.4. Comparison of receiver functions and DSS

We can supplement the passive seismic results in some areas with work using controlled sources. Deep seismic sounding studies provide information through the full thickness of the crust and are available for profiles extending over more than 10,000 km across the Indian landmass. The locations of the various controlled source profiles are shown in Fig. 7.

The crustal thickness picks been made across various deep seismic sounding profiles covering Indian shield and parts of Himalaya. The picks are made at regular intervals from modelling results for the profiles 1–29 (Fig. 7), with due allowance for the drastic changes in the depth of the Moho. For Tibet, the crustal thicknesses are directly taken from deep seismic sounding and deep seismic reflection profiles as provided in Xiaosong et al. (2009). The station elevation is added to the specific profiles where the Moho estimates are provided with respect to mean sea level. We then employ these crustal thickness values in a similar way to Fig. 4 to display the available thickness results from the deep seismic sounding. The map of depth to crust (Fig. 8) has sparse coverage, but strong contrasts are evident. The thickened Tibetan crust, as compared to Indian shield and intermediate depths in Tarim and Qaidam basins comes out clearly. Although the controlled source results are somewhat limited, they provide very useful information to fill in some of the gaps in the coverage of the crustal thickness from receiver functions.

We examine the variations and similarities in the results from deep seismic sounding and receiver functions. The overall pattern is of a good linear correlation, but there are just a few outliers, notably in the Himalayan region where the differences reach 20 km (Fig. 9). These differences are likely to be due to varying influences from strong velocity gradients in the different styles of analysis. Receiver functions are more sensitive to discrete velocity jumps, while refraction models have greater influence from velocity gradients (Kennett and Salmon, 2012).

The good general compatibility between the results from passive and controlled source seismic studies enables us to combine the two data sets and so produce a more complete map of the depth to crust across the Indian region (Fig. 10). The combined map using data points from deep seismic sounding and receiver functions (Fig. 10), agrees well

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**Fig. 5.** Heat flow values taken from the global heat flow data base of the International Heat Flow Commission ([http://www.heatflow.und.edu](http://www.heatflow.und.edu)) with additional values from Nagaraju et al. (2012). The individual data points are shown by diamonds. Heat flow data from 112 observations have been used to create the map, with the same interpolation scheme as used for the crust (Fig. 4).
Fig. 6. Heat flow vs crustal thickness for the Indian region (left panel). The values are extracted at common nodes for which both heat flow and crustal thickness estimates are available, interpolated at 1 × 1°. The histogram in the right panel shows heat flow estimates at particular nodes split into two regimes, <60 and ≥60 mW m⁻².

in the areas already covered by passive seismic studies (Fig. 4), but extends the coverage into a number of important regions (Cambay Rift, Narmada–Son Lineament, Mahanadi Rift, Himalaya and parts of Deccan Volcanic Province). The pattern of variation of the depth to Moho across the Indian subcontinent is well defined, but shows substantial differences from available global models such as CRUST1.0 (Laske et al., 2013), with
4. Lithospheric mantle

There are two major tools to investigate the mantle component of the lithosphere. The exploitation of the fundamental and higher modes of surface waves in tomography can determine the broad scale patterns of seismic wavespeed variation in 3-D (Priestley and Tilmann, 2009). The surface waves provide dominantly horizontal sampling, with depth information through the character of the different modes. Horizontal resolution is not better than 250 km with about 40 km resolution in depth. This is sufficient to provide indications of the base of the lithosphere, but no details. Alternatively body wave tomography and long-duration receiver functions can provide constraints with near vertical sampling, since they exploit distant earthquakes (Rawlinson et al., 2010). For incident P waves the conversions associated with discontinuities in the lithospheric mantle overlap with reverberations in the crust, and so can be difficult to disentangle. On the other hand the p converted waves from incident S waves arrive before the main phase and are unaffected by such reverberations (Kumar et al., 2007). However, the available frequencies are lower for S, so depth resolution is reduced and gradient zones may appear as apparent discontinuities.

4.1. Seismic wavespeed variations

Results from surface wave tomography are very limited for peninsular India, because only the HYB station and some stations in Sri Lanka and the Maldive islands are available for open access. The situation is much improved in the eastern Himalayan region because of the many temporary deployments of seismometers in different parts of Tibet and adjacent regions of China, but coverage diminishes to the west. Figure 2 of Pandey et al. (2014) illustrates the very strong contrast in available path densities in the two regions when using regional stations. Global studies (e.g. Debayle et al., 2011; Schaeffer and Lebedev, 2013) can provide some improvement in coverage by exploiting longer paths traversing India and its environs, but resolution remains limited.

The surface wave tomographic results suggest a slight contrast between the southern tip of India and the zones to the north. The detailed study of Rayleigh wave dispersion by Borah et al. (2014b) is focussed on the Dharwar craton, but extends somewhat into the Southern Granulite Terrane. The detailed 3-D shear wavespeed model suggests that the contrast in the lithosphere occurs somewhat south of the surface transition.

A contrast in the mantle lithosphere at the southern tip of India was presented in the P wavespeed model of Kennett and Widiantoro (1999) derived from regional body-wave tomography, though the available resolution was poor. The recent body-wave tomography study by Singh et al. (2014) was able to exploit a larger, and better distributed,
suite of stations and thereby achieve nearly uniform resolution at 2 × 2° of both P and S wavespeeds across peninsular India. The results of Singh et al. (2014) indicate distinct contrasts at about the 1% level in wavespeed between the southern tip and the cratonic zone to the north in the top 200 km of the mantle.

Because most sampling in body-wave tomography is near vertical, except where regional earthquakes can be exploited, vertical smearing of structure is very common. Nevertheless, some indications can be extracted on the relative thickness of different lithospheric blocks identified by lateral variations in seismic wavespeed. The body-wave tomographic results of Singh et al. (2014) indicate varying thickness of lithospheric roots beneath India and Tibet, most likely arising from various geodynamic events that affect the preservation and attrition of the lithosphere at various epochs.

4.2. Lithospheric discontinuities and transitions

The lithosphere is a thermal boundary layer which is mechanically strong zone due to colder temperatures near the Earth’s surface and is underlain by a weaker zone — the asthenosphere. The transition between the two elements is an important component of the Earth system since it accommodates differential motion between tectonic plates and underlying mantle. But, depending on what class of geophysical and geochemical information are examined there are many different ways of assessing the base of the lithosphere. A succinct summary is provided in Chapter 1 of Artemieva (2011). Geochemical and isotopic evidence suggest that the lithospheric mantle is largely depleted in character due to the extraction of basaltic components. The lithospheric mantle beneath the continents shows a trend of increasing metasomatic refertilization for younger ages, a phenomenon attributed to episodes of melt and fluid infiltration from the underlying asthenosphere (O’Reilly and Griffin, 2010). The asthenospheric mantle is regarded as much more fertile due to the recycling of crustal components during subduction. Thus, in addition to being a mechanical–thermal boundary, the lithosphere–asthenosphere transition also marks a major break in the chemical composition of the subcontinental mantle (O’Reilly and Griffin, 2010).

In some places the lithosphere–asthenosphere boundary (LAB) is sharp enough to be viewed as a discontinuity, but in many places beneath cratonic lithosphere the transition is gradual (see e.g., Yoshizawa, 2014). Other potential discontinuities in cratonic lithosphere are the Hales discontinuity (HD) and the mid-lithospheric discontinuity (MLD). The Hales discontinuity at 80–100 km depth has been associated with either depth localised anisotropy (Bostock, 1998) or phase change from spinel
to garnet peridotite (Hales, 1969) and visible by wide-angle seismic reflection with a positive velocity jump (Ayarza et al., 2010). The MLD is reported as having a decrease in seismic wavespeed within continental interiors, and has been attributed to alteration by melt or metasomatism (Ford et al., 2010) or a change in anisotropy (Yuan and Romanowicz, 2010). We attempt to analyse the results obtained from various studies in the Indian region that see HD, MLD and the LAB reported at almost similar depths.

There is a wide range of estimates for the thickness of the lithosphere using different techniques (Fig. 12). Studies using S–p receiver functions (Devi et al., 2011; Kumar et al., 2007, 2013b) cover much of India and give estimates in the range of 70–140 km. At HYB station (Rychert and Shearer, 2009) using P-to-s receiver functions estimate ~60 km. From surface wave dispersion estimates range from 80 to 155 km (Bhattacharya et al., 2009; Mitra et al., 2006). McKenzie and Priestley (2008) have used surface wave tomography with empirical relations to convert shear wave velocity to depth, and their results indicate variations from ~100–280 km from south to north (Fig. 12b).

Joint inversion of P-to-s and S-to-p receiver functions (Kiselev et al., 2008), Dharwar craton; Oreshin et al. (2008), western Himalaya and Ladakh; (Oreshin et al., 2011), Indian shield, western Himalaya and Ladakh) have revealed low shear velocities 4.4 km s⁻¹ beneath the Precambrian shields. The authors suggest metasomatic alteration of the high velocity keel by some recent tectonic event. Using P-to-s and S-to-p receiver functions from four seismic stations beneath the Dharwar craton (Ramesh et al., 2010) argued in favour of a thick lithosphere on the lines of the idea of a tectosphere (Jordan, 1988). This interpretation is based on the presence of two westerly dipping interfaces at depths of 150 and 200 km. However, they refrain to report the low-velocity-zone observed at around 10 s as a possible LAB. The high shear velocities of 4.7 km s⁻¹ in the depth range of 100–150 km beneath the western Himalaya is suggested to be a recovery of a shield-like structure (Oreshin et al., 2011).

Not only is there considerable variation in the estimates of depth to the base of the lithosphere from different methods, but there are also very different reports on the sharpness of the transition and the velocity change across the LAB, particularly in the old cratonic regions. Kumar et al. (2007, 2013b) describe the LAB as a sharp boundary, while other authors do not report the LAB as a rapid transition (Kiselev et al., 2008; Mitra et al., 2006; Oreshin et al., 2008, 2011).

Beneath the oceanic lithosphere Kawakatsu et al. (2009) has used observations of S-to-p receiver functions at 0.25 Hz to infer a very sharp change at the LAB, associated with partial melting in lenses in the asthenosphere. This model has been also invoked by Kumar et al. (2013b), as a possible explanation for a sharp continental transition. However, such models including partial melt have limitations, and may not be concordant with all aspects of the geodynamics. For example, Karato (2012) suggests anelastic relaxation caused by elastically accommodated grain boundary sliding can explain large velocity drops at a mid-lithospheric discontinuity (MLD), while there would be less contrast across the LAB. The presence of a MLD (Yuan and Romanowicz, 2010) provides an alternative explanation to the thin lithosphere beneath cratons reported in various studies of P-to-s and S-to-p receiver functions (Kumar et al., 2007; Rychert and Shearer, 2009). Beneath India, Kumar et al. (2013b) argued that their observations of thin lithosphere do represent the LAB, not the MLD; they interpret another reflection at much shallower depths at a few stations as the MLD beneath India.

Discrepancies remain between inferences made using different methods. For example, Mitra et al. (2006) have reported the LAB as a gradational boundary layer beneath Indian shield from Rayleigh wave...
phase-velocity measurements. However, in another study (Kiselev et al., 2008) show that synthetic seismograms using the velocity models of Mitra et al. (2006) do not reflect a layer which can be matched with actual S–p receiver functions. The difficulties arise because the results from analysis of surface waves naturally favour rather smooth variations in depth, since minor fluctuations cannot be resolved. Yet, it is precisely such rapid variations that would give rise to signals that could be interpreted as discontinuities in the analysis of receiver functions.

Hales (1969) reported a seismic discontinuity at a depth of 80–90 km south of Lake Superior from the analysis of seismic refraction profiles. Since then, there has been on-going debate on its nature, the cause of its existence, and whether it is a global phenomenon or just a regional feature. The first report of the Hales discontinuity beneath India was made by Saul et al. (2000) for the station HYB, located in the north-eastern part of Dharwar craton, at a depth of 90 km. They were uncertain about the origin of the observed energy on the transverse (SH) component, and attributed it to an anisotropic layer, taking a clue from the results of Bostock (1998). Elsewhere the presence of the H-discontinuity has been proposed from south to north across India, using data from a few stations (Jagadeesh and Rai, 2008). Such an interpretation is supported by velocity models presented by Mitra et al. (2006), who reported an increase in S-velocity from 4.52 to 4.77 km s\(^{-1}\) at depths of near 75 km. Positive velocity jumps are also reported for certain stations in the Western Dharwar Craton, Deccan Volcanic Province and for station HYB (Kiselev et al., 2008). In the S velocity profiles for azimuths around 100° a distinct positive discontinuity was reported in the depth interval of 80 to 100 km (80 km for HYB, 120 km for Western Dharwar Craton and 100 km for Deccan Volcanic Province), however this feature is missing for azimuths between 30° and 50° (Kiselev et al., 2008). To establish the existence of the H-discontinuity beneath India is a challenge because of the limited back-azimuthal coverage of the data sets available for use in such studies. A point of concern is the reports of HD and LAB at almost similar depths beneath India. For one of the most thoroughly studied seismic stations HYB, we have inferences of the LAB at 99–101 km depth from S–p receiver functions (Kumar et al., 2007, 2013b) and HD at depths of 90 km from P-to-s receiver functions (Saul et al., 2000). Reconciling such results is not at all easy, and often anisotropy is invoked. Clear analysis of such lithospheric discontinuities needs joint application of multiple techniques if more reliable results are to be obtained.

Body-wave tomographic results do not provide direct estimates of lithospheric depth, but reveal thick high velocity perturbations observed at depths greater than 200 km (Singh et al., 2014), while receiver function studies infer strong variations at much shallower depths (60–140 km, Kumar et al., 2013b; Rychert and Shearer, 2009).

Concepts that have been employed to try to reconcile the results from tomography and receiver functions, are the differences between chemically and thermally defined lithosphere, the presence of mid-lithospheric discontinuities and a depleted fertile lithosphere (O’Reilly and Griffin, 2010). Even within tomographic studies many differences arise depending on the data employed, particularly beneath the Indian shield (McKenzie and Priestley, 2008; Singh et al., 2014).

The Lehmann discontinuity, at depths around 220 ± 30 km beneath continents could be associated with the bottom of lithosphere, marking a transition to asthenospheric anisotropy (SH leading SV) related to present day flow (Gung et al., 2003). Such a scenario is present beneath India, where APM related asthenospheric anisotropy, with SH leading SV is reported using direct S and SKS/SSK phases (Kumar and Singh, 2008; Saikia et al., 2010). Reports of the Lehmann discontinuity beneath India and Tibet are meagre (Heit et al., 2010; Kiselev et al., 2008; Kosarev et al., 2013; Oreshin et al., 2011; Sharma and Ramesh, 2013; Wittlinger et al., 2004), but suggest a deeper boundary (>200 km) than that commonly inferred from S-to-p receiver functions. Beneath the stable parts of Indian shield, the Lehmann discontinuity may define the lithosphere boundary (Gung et al., 2003), revealed in body wave tomographic images.

The conspicuous and intermittent nature of this discontinuity (Gu et al., 2001) and discrepancies to associate Lehmann as a boundary between anisotropic and isotropic medium (Vinnik et al., 2005), remain as ongoing issues.

4.3. Effects of lithospheric thickness variation

Various studies have proposed that the exceptionally fast motion of the Indian tectonic plate has been facilitated by the thinning of the Indian lithosphere (Kumar et al., 2007, 2013b; Negi et al., 1986). However body-wave tomographic studies do not suggest strong lithospheric thinning. The issue of the fast drift of the Indian tectonic plate is not that simple, and needs to be understood in terms of effects of various possible candidate mechanisms. Slab pull and ridge push models are considered to be possible mechanisms for the fastest drift of Indian tectonic plate (Capitanio et al., 2013; Copley et al., 2010). The role of plumes has been also considered to facilitate the fast drift of Indian tectonic plate. A substantial push force due to the Réunion plume is thought to be responsible for the speed-up and subsequent slowdown of Indian plate (Capitanio et al., 2013), although the issues of the time period of anomalous Indian plate motion and of plume activity along with its impact on acceleration of India remains unsolved (Muller, 2011). The presence of the mantle conveyor belt deduced from mantle density distributions may also have played a role in the breakup and be responsible for the indentation of Indian and Arabian plates (Becker and Facenna, 2011). The presence of such a conveyor belt may impact on the extent of slab pull and ridge push forces, which then contribute as a secondary driving mechanisms with mantle upwelling and active drag as primary (Becker and Facenna, 2011).

4.4. Seismic anisotropy

The rapid movement of the Indian plate (with velocities reaching up to 20 cm/year) and its quite variable surface geology makes it unique in comparison to other shield regions. In this context, seismic anisotropy parameters extracted from SKS/SSK phases allow the investigation of a number of significant geodynamic questions as follows: (a) do the distinct geologic provinces show disparate anisotropic character? (b) Does the fast moving plate force deformation in the plate motion direction? Or (c) is anisotropy from the past tectonic events frozen in the lithosphere? Although large data sets of seismic anisotropic parameters have been acquired from various Precambrian shields worldwide, published measurements from the Indian shield are limited before 2006 (Singh et al., 2006). A few measurements exploiting analogue data, reveal dominance of Absolute Plate Motion related strain (Ramesh et al., 1996). The results from a single seismic station SHIO located over Shillong plateau, show anisotropic directions parallel to the strike of mountain belts (Vinnik et al., 1992). Later on results using HYB and SHIO (Chen and Ozalaybey, 1998) and HYB (Barroul and Hoffmann, 1999) seismic stations are found to be null. Such observations were interpreted in terms of the isotropic nature of Indian lithospheric mantle and onset of seismic anisotropy in Tibet north of ITSZ as a marker for northern limit of Indian lithosphere (Chen and Ozalaybey, 1998). Now with a flood of seismic high quality data sets in the recent decade it is possible to examine the results from SKS/SSKs and direct-S phases for a complete picture of deformational nature of the Indian plate.

In the period from 2006 to 2013, more than 1000 individual splitting parameters using SKS/SSKs and 139 using direct S waves have been presented in various studies (Fig. 13). The results using SKS/SSKs phases clearly reveal absolute plate motion related strain as the dominant process beneath Indian plate (Kumar and Singh, 2008), with a possibility
of modified asthenospheric flow around lithospheric keels. The delay times between the split S phases are ~1 s, as observed for continental shields globally. The prominent trends observed are (a) NS to NE orientation of the fast axis in southern part of peninsular India, (b) NE orientation in north to central India, (c) NE orientation in Himalayan region east of the Sikkim Himalaya, (d) EW orientation parallel to the mountain belts in northeast India, and (e) in the Himalaya and Tibet collision zone the orientation is roughly EW, parallel to major suture zones. Although attempts have been made to characterise the seismic anisotropic character of the stable and actively deforming regions of the Indian plate, a number of issues still remain to be addressed. Since SKS splitting provides an integrated effect of both the crust and the mantle, the crustal contribution to anisotropy needs to be isolated. Approaches that exploit the splitting of Ps phases converted at the Moho to estimate the strength and orientation of the fast axis, in a manner similar to SKS-splitting techniques, need to be adopted. The utility of Ps phases in estimating crustal anisotropy has been demonstrated (McNamara and Owens, 1993), although not widely applied. Studies of crustal anisotropy together with those of the mantle should enable an improved understanding of the nature and extent of coupling between the crust and mantle.

As is well known, SKS/SKKS phases are only sensitive to horizontal anisotropy. Where expected deformation is complex as in the Andaman and Burmese arc regions, analysis of splitting of S waves from deep earthquakes (to minimize the source side anisotropy) and its modelling in terms of a dipping axis of symmetry can be undertaken. However, separating the effects of mid-mantle and receiver side anisotropy is a challenge that must be met when using direct-S waves. The results from Indian shield stations using direct-S waves clearly brings out a close similarity with the SKS/SKKS results (Saikia et al., 2010), as illustrated in Fig. 13.

SKS splitting measurements that do not directly identify the origin of anisotropy whether anisotropy resides in the lithosphere or in the sublithospheric mantle is an open issue. To be able to resolve this debate, there is a need to evolve to strategies that isolate the Ps conversions from deeper anisotropic layers and model them. Since the interference of reverberations from shallow layers and conversions from deeper layers make this task difficult, effective techniques to suppress the multiples need to be developed. Techniques like joint inversion of Ps and SKS phases need to be extended to deal with 2-D situations, which are expected in plate boundary regions. Further, joint inversion of P and S receiver functions (that is recently emerging as a powerful tool to map the upper most mantle stratigraphy using Sp conversions) to model the dip and anisotropic effects can be a big step forward. Synthetic tests demonstrate that the Sp conversions are less sensitive to anisotropy, and more importantly separate the converted and multiple phases to naturally lie on either-side of the parent S-wave. Moreover, conversions from the lesser used phases like PP and PKP can be used to fill the back-azimuthal gaps resulting from the uneven earthquake source distribution and the short periods of operation of temporary seismic stations. Lastly, data from long running permanent stations hold the key to validate newer methodologies and understand the complex inner workings of the dynamic Earth system.

5. The 410 and 660 mantle discontinuities

Major discontinuities associated with mineralogical phase transitions are the mantle transition zone at 410 and 660 km. The properties of these discontinuities provide important constraints on the thermal and compositional nature of the mantle. The mineralogical phase transition from olivine α-phase to the modified spinel β-phase at the 410 km discontinuity and the ringwoodite γ-phase to perovskite and magnesiowustite at the 660 km discontinuity match well with the seismic results. Variations in the depths of these discontinuities in response to vertically oriented mantle temperatures are anti-correlated with positive Clapeyron slope (∂P/∂T) at the 410 km discontinuity and negative Clapeyron slope at the 660 km discontinuity. The result is a thickened mantle transition zone in the presence of trapped cold lithospheric slabs within the transition zone, and early returns from the 410 and 660 in the case the slabs or lithospheric roots lie well above the 410. The mantle transition zone beneath India is not akin to those observed globally beneath other Precambrian shields. Receiver function results are frequently used to map the topography of 410 and 660 km discontinuities (Fig. 14). These images for the behaviour at 410 and 660 km phase arrivals are produced based on the results of individual stations, rather than piercing points, except for those from the work of Singh and Kumar (2009). Thus, they represent a broad picture of the mantle beneath India and Tibet, rather than being suitable to map local variations.

Fig. 13. Individual shear wave splitting observations: (a) SKS/SKKS waves: Singh et al. (2006) — northeast Himalaya; Singh et al. (2007) — Sikkim Himalaya; Kumar and Singh (2008) — Indian shield; Oreshin et al. (2008) — western Himalaya; Heintz et al. (2009) — Indian shield; Kumar et al. (2010) — Godavari Graben; Mandal (2011) — northwest India; Madhusudhan Rao et al. (2013) — northwest Deccan Volcanic Province; and Hazarika et al. (2013b) — northeast India. (b) Direct-S waves: Roy et al. (2012) — eastern Dharwar craton; Saikia et al. (2010) — Indian shield. The splitting results for Tibet are taken from the compilations of Wüstefeld et al. (2008).
5.1. 410 km discontinuity

The vertical passage time for seismic P waves to the 410 km discontinuity is close to that predicted by the *ak135* reference model beneath the key cratonic segments of the Indian shield: (Kisielev et al., 2008; Kosarev et al., 2013; Kumar and Mohan, 2005; Kumar et al., 2013a; Oreshin et al., 2011; Rai et al., 2009b; Ramesh et al., 2005). This 'normal' behaviour is rather different from that for other shield regions around the globe, where thick, fast, lithospheric roots have resulted in an elevated 410 discontinuity (Kumar et al., 2013a). Consistently early passage times, sometimes even close to ~2 s early, are seen for the 410 km discontinuity from the stations located in the Himalayan mountain belts and its foothills: (Kosarev et al., 2013; Kumar et al., 2013a; Oreshin et al., 2008, 2011). The continuity of early arrivals is also maintained for most parts of Tibet, south of the Bangong Nuijiang Suture Zone. The northwestern Deccan Volcanic Province and Southern Granulite Terrain show a delay in the passage times of ~1.5 s compared to stable parts of Indian shield, as both regions are influenced by plumes. The northwestern Deccan Volcanic Province shows the strong imprints of Deccan volcanism, and the Southern Granulite Terrain has been affected by Marion plume responsible for separation of Madagascar from India (Kumar et al., 2013a). The other areas which show delayed 410 arrivals are regions close to the eastern coast of India.

In Tibet across various profiles, continuity of the 410 and 660 discontinuities was reported precluding the idea that mantle transition zone is resting ground for detached lithospheric fragments (Kind et al., 2002; Yuan et al., 1997). However, later studies argue in favour of detached lithospheric fragments within the transition zone itself (Chen and Tseng, 2007; Singh and Kumar, 2009).

Olivine and its polymorphs (wadsleyite, β - spinel and ringwoodite, γ - spinel) can incorporate small amounts of water into their nominally anhydrous crystal structures. This makes the mantle transition zone a potential water reservoir. There are a number of implications of the effect of water on transition zone discontinuities mainly (a) an elevation of the 410 by changing the thermodynamic stability fields of olivine polymorphs (Smyth and Frost, 2002; Wood, 1995), (b) broadening of transition zone by 20 to 25 km (Wood, 1995), (c) frequency dependence of amplitudes (van der Meijde et al., 2003), (d) a sharp 520 km discontinuity if present (Inoue et al., 1998), and (e) hydrous melting can creates low velocity layers atop the 410 with lesser water content above and below (Huang et al., 2005). Beneath India, there is no compelling evidence for the existence of the 520 km discontinuity (Kumar et al., 2013a). There are indications of a seismic discontinuity at a depth of ~475 km beneath Ladakh and Karakoram north of the Indus Suture Zone (Rai et al., 2009b), these were interpreted as arising from the presence of remnants of subducted Indian slab as the depth is well above 520 km. There has so far been little analysis of the frequency dependence of amplitudes of receiver functions for Indian data sets. Singh and Kumar (2009) have not found any frequency dependence in northeast India close to Himalayan mountain belts, which appears to rule out possibility of water within the mantle transition zone in this area. A low velocity layer atop the 410 discontinuity

![Fig. 14.](image-url)

(a) The vertical passage time of seismic P waves to the 410 km discontinuity as reported by various studies (He et al., 2014; Kayal et al., 2011; Kisielev et al., 2008; Kosarev et al., 1999, 2013; Kumar et al., 2013a; Oreshin et al., 2008, 2011; Pathak et al., 2006; Shen et al., 2008; Singh and Kumar, 2009; Yuan et al., 1997). Different symbols are used to indicate each set of results. The results of Rai et al. (2009b) are not shown here for consistency, as they are based on piercing points over very large areas. The normal behaviour for 410 (+) is seen in the studies by Kumar and Mohan (2005), Ramesh et al. (2005), Saul et al. (2000), Sharma and Ramesh (2013), and Singh et al. (2012). (b) Same as (a) for 660 km discontinuity. The only exception is the 660 km discontinuity reported by Sharma and Ramesh (2013) which is not plotted due to its sporadic nature.
is also reported at few places beneath India (Kosarev et al., 2013; Kumar et al., 2013a; Oreshin et al., 2011), but it remains to be seen how widespread is such a feature, and what the full implications are for water in the transition zone.

5.2. 660 km discontinuity

Early vertical passage times compared with the ak135 reference model (Kennett et al., 1995), sometimes even close to ~2 s early are seen for the 660 km discontinuity for regions of Western Tibet, Central India and few parts of Himalaya and Tibet. The early arrivals of 660 near Himalayan collision belt may relate to present day subduction or the preserved keel of Indian shield (Kosarev et al., 2013), as they are well correlated with the elevation of the 410 discontinuity. Apart from early arrivals or 410 and 660 in the Himalaya, a thick transition zone in the northern part of Indian-subcontinent reflects resting ground for Tethyan slabs within the transition zone itself (Singh and Kumar, 2009). Regions of the Southern Granulite Terrain and South West Deccan Volcanic Province show delayed 660 arrivals, as also observed for 410. The delayed arrivals suggest plume effects above 410 and 660 km, in those regions.

5.3. X-discontinuity

Hydration reactions in subduction zones, crystallographic transitions of pyroxene, and the coesite to stishovite transition or exsolution of stishovite from clinopyroxenes containing excess silica, have been used to explain discontinuities observed around 300 km depth (Schmerr et al., 2013; Shen et al., 2014; Williams and Revenaugh, 2005). The intermittent nature of this feature may be due to lateral changes in the fraction of basalt (Schmerr et al., 2013). The reports of this X-discontinuity (Revenaugh and Jordan, 1991) beneath India are few, with possible detection in northeast India (Ramesh et al., 2005) and beneath the Hyderabad station HYB (Bodin et al., 2013). The X-discontinuity inferred from common conversion point stacks of receiver functions in northeast India has been interpreted in terms of melt migration and segregation at these depths (Ramesh et al., 2005). SS precursors also trace such a discontinuity with a shear velocity contrast of ~2% (Deuss and Woodhouse, 2002) beneath India. The presence of the X-discontinuity has implications for the lateral variations and stratification of the mantle, and receiver functions have considerable potential as a tool to track at this depth (Bodin et al., 2013; Ramesh et al., 2005; Wittlinger et al., 2004; Yuan et al., 1997). Careful examination of the seismic data from India may throw more light on the nature of X-discontinuity, which may often be overlooked due to its intermittent nature and low shear impedance contrast (Schmerr et al., 2013).

6. Summary

With the aid of the growing body of results from passive seismic studies we have been able to provide a summary description of the crustal properties of much of the Indian subcontinent, and also to characterise the upper mantle beneath. Geographic coverage of peninsular India is good, but information is still limited in the more northern parts of the cratons, the region of the Indo-Gangetic plains and parts of the Himalayan foothills.

More detailed crustal information is currently available only from the deep seismic sounding coverage, which though extensive is geographically restricted. The exploitation of seismic noise through ambient noise tomography should provide means for augmenting information on crustal structure on a broad scale in a similar way to that used for the AuSREM crustal model in Australia (Kennett and Salmon, 2012). However, this will not be a straightforward task because of rather restrictive data practices by many agencies.

Beneath India the crust is quite variable in its depth beneath the surface and there are indications also of significant changes in character (see Figs. S2 and S3). The cratonic regions generally have crustal thickness in the range of 40–45 km, which is typical for cratons in other parts of the world (cf. Kennett and Salmon, 2012). Though there are limited zones, e.g. in the central portion of the Western Dharwar Craton, that have rather thicker crust (more than 55 km). The Moho deepens rapidly as the Himalayan collision zone is approached, and the crust reaches its maximum thickness beneath Tibet; the crustal thickness exceeds 75 km in parts of the Lhassa terrain. The crustal thickness map built from the passive seismic and deep seismic sounding studies shows significant deviations from that predicted from global models such as CRUST1.0 (Laske et al., 2013). Such differences are of potential significance when ever crustal corrections have to be made to seismological results.

Although it is difficult to extract a consistent definition of lithospheric thickness from the range of studies that have been undertaken in the Indian region, a number of clear trends emerge. The lithosphere tends to thicken from south to north. The lithosphere beneath the cratonic areas is somewhat thinner than in similar locations around the globe. This has been ascribed to the effects of the rapid motion of India after splitting from Gondwana, but is in contrast with the currently most rapidly moving continent, Australia, where lithospheric thickness estimates exceed 220 km beneath the Archean Yilgarn craton (Kennett and Salmon, 2012).

The shear-wave splitting results from body-wave studies show the dominance of strain approximately in the direction of absolute plate motion, with an anisotropic Indian mantle for which the delay times between the split S phases are close to 1 s. However, the body wave results do not localise the anisotropic zone in depth. The global study by Debayle et al. (2011) of azimuthal anisotropy using surface waves provides indications of two-layer anisotropy beneath India with alignment to absolute plate motion at the base of the lithosphere or in the asthenosphere beneath; though resolution was limited in their study. The only other location with a similar configuration in this study is Australia, with current rapid motion, which shows a distinct change in anisotropic character between 100 and 200 km depth.

The influence of crustal structure is significant for many tomographic studies, both in body wave tomography and in the frequency band used in surface wave tomography to study the lithospheric mantle and the asthenosphere below. Our results for the Indian subcontinent indicate the need for improved crustal models built from direct observations, rather than assigned by analogy with regions with a similar geological environment. This will be an ongoing task.

Acknowledgements

We thank M. Ravi Kumar and R. K. Chadha for the useful discussions at different points of time. This study has been supported by a grant from the Ministry of Earth Sciences (MoES), IITKGP/CKH. We thank Stewart Fishwick and an anonymous reviewer for their valuable comments and suggestions which helped us to significantly improve the manuscript.

Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.tecto.2015.01.007.

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