A review of crust and upper mantle structure beneath the India 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 (V_s) is low in the Himalaya-Tibet collision zone compared to Indian shield. Major crustal features are: (a) the Eastern Dharwar Craton has a thinner and simpler crustal structure 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 show 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

30 Asia. The mantle transition zone beneath India is predominantly normal with some
31 clear variations in the Himalayan region (early arrivals) and Southwest Deccan Volcanic
32 Province and Southern Granulite Terrain (delayed arrivals). No clear patterns on in-
33 fluence on the mantle transition zone discontinuities can be associated with lithospheric
34 thickness. Over 1000 anisotropic splitting parameters from SKS/SKKS phases and 231
35 using direct S waves are available from various studies. The shear-wave splitting results
36 clearly show the dominance of absolute-plate-motion related strain of a highly anisotropic
37 Indian lithospheric mantle with delay times between the split S phases close to 1 s. There
38 are still many parts of India where there is, at best, limited information on the charac-
39 ter of the crust and the mantle beneath. It is to be hoped that further installations of
40 permanent and temporary stations will fill these gaps and improve understanding of the
41 geodynamic environment of the Indian subcontinent.

42 *Keywords:* Moho; Mantle Transition Zone; Lithosphere-Asthenosphere Boundary;
43 Deep Seismic Sounding; Receiver Functions; Indian Shield

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64 1. Introduction

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

72 In the century since the detection of the Mohorovičić discontinuity (Mohorovičić,
73 1910) from earthquake observations, both controlled source and passive seismic studies
74 have made impressive advances in understanding the nature of the crust and uppermost
75 mantle (Prodehl et al., 2013). Multiple facets of seismic wave propagation can be brought
76 to bear on the structure of the Earth’s interior, and help to resolve the key issues related
77 to evolution and nature of the continental crust and upper mantle. To date there have
78 been only limited attempts to provide a full picture of the Indian crust and upper mantle.
79 There have been reviews of heat flow (Roy and Rao, 2000) and deep seismic sounding
80 studies (Kaila and Krishna, 1992; Reddy and Rao, 2013). However, the full range of
81 available information on the crust and upper mantle available from passive source studies
82 have not previously been exploited.

83 The foundation stones of seismology in India were laid by the pioneering works of
84 Dr. T. Oldham and Dr. R. D. Oldham, the father-son duo. The great Shillong earth-
85 quake of 12th June, 1897 is well documented and reported in the works of R. D. Oldham
86 (Oldham, 1899). This deadly Shillong earthquake achieved the maximum intensity XII
87 on MM scale (Richter, 1958), and provided the impetus for a series of initiatives to
88 install seismographs in India to monitor earthquakes. The first few installations were
89 made of Milne’s self registering seismographs in Alipore (Calcutta, now Kolkata), Colaba
90 (Bombay, now Mumbai) and Madras (now Chennai) (Tandon, 1992). An Omori-Ewing
91 seismograph was installed in Simla as a response to the great Kangra earthquake of

92 5th April, 1905. In the years from 1929 to 1930, the country was equipped with a few
 93 more Milne-Shaw seismographs, initially installed at Colaba observatory Mumbai, then
 94 Bombay) and later at few more places in Agra, Calcutta, Hyderabad and Kodaikanal.
 95 In the early 1960's five World Wide Standard Seismograph network (WWSSN) stations
 96 were installed at various places across the country following the recommendations of
 97 Berkner (1959). After the devastating Latur earthquake of September 30th, 1993 the
 98 Indian Meteorological Department upgraded ten of its observatories to the standard of
 99 Global Seismograph Network, and later complemented this network with 14 more broad-
 100 band stations during 1999-2000. At present the Indian Meteorological Department runs
 101 nearly 80 seismic stations in the national network, supplemented by various temporary
 102 networks operated by other organizations. Temporary and permanent networks in dif-
 103 ferent parts of India have been operated by the National Geophysical Research Institute,
 104 Indian Institute of Technology Bombay, Wadia Institute of Himalayan Geology, Tezpur
 105 University and the Institute of Seismological Research. The National Geophysical Re-
 106 search Institute has established more than 200 broadband seismic stations at various
 107 points of time, and so plays a major role in passive source seismology in India.

108 Deep seismic probing of Indian crust, started in 1972 with refraction/wide-angle re-
 109 flection work, but subsequently was dominated from the early nineties by deep seismic
 110 reflection. A good deal has been achieved (Kaila and Krishna, 1992), with more than
 111 10,000 km of profiles carried out in various experiments using controlled sources. A major
 112 supplementary source of information on Indian structure comes from the use of seismic
 113 receiver functions exploiting the recordings of distant earthquakes. Receiver functions
 114 provide a tool to map the Earth's response beneath a single three-component seismic sta-
 115 tion, and extract information on the seismic discontinuities at depth from the conversions
 116 and reverberations associated with the main seismic phases. The first receiver functions
 117 for the Indian region used data from the Hyderabad station (HYB) in India, using P-to-s
 118 converted waves (Gaur and Priestley, 1997). Since then the role of receiver functions
 119 in determining crust and upper mantle discontinuities (Moho, lithosphere-asthenosphere

120 boundary, mantle transition zone discontinuities 410 and 660) has been routine practice.
121 Further information comes from seismic anisotropic studies using SKS/SKKS phases and
122 heat flow that provide links to help understand both geodynamics and structure. The
123 present work presents as a complete picture of the Indian crust and upper mantle as pos-
124 sible, compiled from various sources with emphasis on passive source seismic datasets.
125 We synthesize results from seismic studies, heat flow and seismic anisotropy to develop a
126 comprehensive map of the properties of the crust and upper mantle beneath the Indian
127 subcontinent, with links into the Himalaya and Tibet to provide a wider perspective and
128 understanding of the whole region.

129 **2. Tectonic setting**

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

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

145 The other major rift systems are the Godavari Graben (GG) and the Mahanadi Rift
146 (MHR), which are passive in nature but which have left clear imprints on the surface.

147 The approximately east-west Narmada-Son Lineament (NSL), which originated in the
 148 Archean, is another prominent feature that divides various tectonic zones to the north
 149 and south (Meert et al., 2010). The southernmost part of the Indian peninsula is termed
 150 the Southern Granulite Terrain (SGT). This consists of high-grade granulites of late
 151 Archean to Neoproterozoic age, traversed by various shear zones (Meert et al., 2010).

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

161 The Indian craton is formed of a mozaic of a few smaller cratons (Taylor et al.,
 162 1984). These stable parts of the Indian shield include the Dharwar Craton, Bastar
 163 Craton, Singhbhum Craton (SC), Bundelkhand Craton (BUC) and Aravalli Craton (part
 164 of the Delhi-Aravalli Fold Belt, DAFB). These cratons are basically low to high-grade
 165 crystalline rocks formed by intense deformation and metamorphism in the Precambrian.
 166 The Dharwar Craton is separated into the Eastern Dharwar Craton (EDC) and Western
 167 Dharwar Craton (WDC) by the north-south trending Closepet Granite (CG) (Figure 1).

168 Northeast India under the influence of the Himalayan collision is a region of complex-
 169 ity, at one side are the Indo-Burmese Ranges and at the other are the collisional belts
 170 of the Himalaya, in between are the Shillong Plateau (SP) and the Mikir Hills (MKH).
 171 The Himalaya-Tibet collision zone formed due to collision of Indian and Eurasian plates,
 172 and is comprised of various blocks which are separated by suture zones (Indus-Tsangpo
 173 Suture Zone, ITSZ; Bangong-Nujiang Suture Zone, BNSZ). South of Indus-Tsangpo Su-
 174 ture Zone, lies the Himalaya, which is bounded by various north dipping Cenozoic faults

175 systems like Main Boundary Thrust (MBT) and Main Central Thrust (MCT). The re-
 176 gion between Indus-Tsangpo Suture Zone and Bangong-Nujiang Suture Zone, is known
 177 as the Lhasa terrane, adjacent to the Qiangtang terrane (Yin and Harrison, 2000),
 178 while the Qiantang terrane is situated between the Bangong-Nujiang Suture Zone and
 179 the Jinsha Suture Zone (JSZ) (Figure 1).

180 **3. Crust and crust-mantle boundary**

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

188 Many of the results for the Indian subcontinent come from crustal models extracted
 189 from receiver functions (Langston, 1979) beneath three-component seismic stations, from
 190 the Indian shield to the actively deforming regions of Himalaya and Tibet. Before we
 191 discuss the crustal models, we provide a brief discussion of receiver functions as a passive
 192 method, since this approach in recent decades has provided most of the information on
 193 the major discontinuities beneath a seismic station. The idea of exploiting teleseismic P
 194 waveforms for crustal information starts with Phinney (1964) who exploited the Fourier
 195 spectral amplitude ratios of these waves recorded at various World-Wide Standard Seis-
 196 mograph Network (WWSSN) stations. Receiver functions can be computed both in the
 197 frequency (Phinney, 1964; Kurita, 1973) and time domain (Jordan and Frazer, 1975;
 198 Burdick and Langston, 1977; Langston, 1979), with various advantages and limitations.
 199 Much of the work has followed (Langston, 1979), and used frequency domain deconvolu-
 200 tion procedure, where the numerically unstable spectral division is taken into account
 201 by applying water level (Clayton and Wiggins, 1976). Later a time domain inversion

202 technique was developed by Owens et al. (1984) and successfully implemented by vari-
 203 ous workers with certain modifications to obtain the one-dimensional velocity structure
 204 (Owens and Zandt, 1985; Owens, 1987; Kind et al., 1995). However, the issues of non-
 205 uniqueness and non-linearity in the inversion of receiver functions were demonstrated by
 206 (Ammon et al., 1990), with complex trade-offs between velocities above interfaces and
 207 the depth of apparent conversion. The methods have been developed significantly in re-
 208 cent decades. The weak conversions in individual receiver functions can be enhanced by
 209 applying moveout corrections and stacking receiver functions at a single station (Vinnik,
 210 1977; Kind and Vinnik, 1988). More complex structure can be considered by inverting
 211 waveforms for layer dip (Zhang and Langston, 1995), or even to recover both anisotropic
 212 and dipping effects (Levin and Park, 1997; Frederiksen and Bostock, 2000). However,
 213 these more complex schemes increase the non-uniqueness in the inversion of the receiver
 214 functions.

215 The first receiver functions were produced for India using data from Hyderabad sta-
 216 tion (HYB) located in Eastern Dharwar Craton (Gaur and Priestley, 1997), the only
 217 station in the International Federation of Digital Seismograph Networks (FDSN) oper-
 218 ating in India since 1989. Thereafter the exploitation of receiver functions to extract
 219 crustal discontinuities and Poisson’s ratio has become routine with every new installa-
 220 tion of broadband seismic station in India. Up until April 2014 data from ~ 442 seismic
 221 stations have been used to obtain estimates of crustal thickness and Poisson’s ratio for
 222 different geological provinces of the Indian sub-continent. Most of these profiles and
 223 seismic stations are focused on understanding the geometry of crustal discontinuities be-
 224 neath the Indian shield (Gaur and Priestley, 1997; Saul et al., 2000; Kumar et al., 2001;
 225 Sarkar et al., 2003; Rai et al., 2003, 2005; Gupta et al., 2003; Mohan and Kumar, 2004;
 226 Tiwari et al., 2006; Jagadeesh and Rai, 2008; Julià et al., 2009; Kayal et al., 2011; Kumar
 227 et al., 2012; Borah et al., 2014a) or north east India (Kumar et al., 2004; Ramesh et al.,
 228 2005; Mitra et al., 2005; Hazarika et al., 2012). Efforts are now being made by various
 229 workers in the Himalayan region along different seismic profiles in order to understand

the Moho geometry of the Indian plate beneath the Himalaya (Rai et al., 2006; Caldwell et al., 2013; Singh et al., 2010; Acton et al., 2011; Hazarika et al., 2013a).

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.

3.1. Average crustal velocities

Maps of average crustal velocity and Poisson's ratio from passive seismic studies are presented in Figure 2. The average velocity models employed are the end results of: (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 has 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 (Figure 3). The regions of the Himalaya and Tibet show significantly lower Vs ($\sim 3.57 \text{ km s}^{-1}$) than in the Indian shield ($\sim 3.7\text{--}3.75 \text{ km s}^{-1}$) (Figure 2). The Himalayan foredeep seismic stations, and stations in the Indo-Gangetic alluvium plains have an intermediate range with an average Vs of $\sim 3.64 \text{ km s}^{-1}$. Out of 323 seismic stations in the Himalaya and Tibet, 200 seismic stations have velocities $\leq 3.6 \text{ km s}^{-1}$, 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 (Stolk et al., 2013; Chen et al., 2010).

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 $>3.7 \text{ km s}^{-1}$. 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 is clear and obvious. Seismic stations on the west coast of India have somewhat lower crust shear wavespeed than in the shield (Figure 2b).

A more subtle difference appears between the Eastern Dharwar Craton ($V_s \sim 3.71 \text{ km s}^{-1}$) and the Western Dharwar Craton ($V_s \sim 3.73 \text{ km s}^{-1}$). The difference is almost negligible given the uncertainties (Figure 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 (Sarkar et al., 2003; Gupta et al., 2003; Borah et al., 2014a), however such a difference is restricted to the geographic patterns in the core central portions (Figure 2b) of Eastern and Western Dharwar Craton.

The Poisson's ratio is around normal (0.24-0.27) for significant parts of the Dharwar craton, Southern Granulite Terrain and basins like Cuddapah (Figure 2a). The somewhat lower Poisson's ratio for the Western Dharwar Craton (Sarkar et al., 2003; Gupta et al., 2003) than the Eastern Dharwar Craton, may possibly be due to a more felsic upper crust with Poisson's ratio <0.25 , while the lower portion of crust may have mafic velocities with a correspondingly higher Poisson's ratio. These inferences by Sarkar et al. (2003), are further validated by Kiselev et al. (2008) who jointly inverted P-to-s and S-to-p receiver functions.

The crustal composition for the Deccan Volcanic Province inferred from available

285 Poissons ratio information (~ 0.26), does not reflect its actual composition due to an
 286 uneven distribution of sampling. Most of the stations are close to the Western Ghats, or
 287 at the eastern most boundaries of the Deccan Volcanic Province. The DVP is classified
 288 as more felsic-to-intermediate in nature by Kumar et al. (2001). However, Kiselev et al.
 289 (2008) using P-to-s and S-to-p receiver functions, found the Deccan Volcanic Province
 290 composition to lie between the Eastern and Western Dharwar Cratons, with a layer of
 291 mafic velocities. A high Poisson's ratio is observed across the Narmada-Son lineament
 292 (~ 0.28) and the Western Ghats. The paleo-rift Narmada-Son Lineament is traced by
 293 two N-S profiles in different experiments (Rai et al., 2005; Jagadeesh and Rai, 2008).
 294 The crust beneath the Narmada-Son Lineament found to be more mafic in nature with
 295 high V_p/V_s ratios (1.84 to 1.91, Table S1) than its surroundings and is suggestive of a
 296 high-density mafic mass at depth that compensates the crustal root, also supported by
 297 the small topographic variation ($\sim 200\text{m}$) across the lineament (Rai et al., 2005).

298 In the Himalaya-Tibet collision zone, the average Poisson's ratio is 0.26, with strong
 299 variations over range from 0.16 to 0.32. A total of 107 stations show Poisson's ratio < 0.25 ,
 300 while 189 have a ratio ≥ 0.25 . The skew is clearly seen on the sides of the zone of high
 301 Poisson's ratio in Figure 2a. Examining the distribution based on nodal points (sampling
 302 of $0.25^\circ \times 0.25^\circ$), we find $\sim 80\%$ of the region has higher Poisson's ratio, with most centred
 303 around 0.26. With a sampling of $1^\circ \times 1^\circ$ the region with Poisson's ratio ≥ 0.25 is even
 304 greater ($\sim 92\%$). These results provides a better understanding of the region, in view of
 305 the uneven station distribution (Figure 2a). The high Poisson's ratio beneath most of
 306 the Tibet plateau suggests the presence of partial melts/fluids, as interpreted in various
 307 studies (Owens and Zandt, 1997; Li et al., 2006, 2009).

308 Figure 3 illustrates summary properties of the crust for specific geological regions: (a)
 309 Himalaya and Tibet, (b) the cratonic regions and Southern Granulite Terrain (c) the rift
 310 systems of the Narmada-son lineament and Godvari Graben and (d) the Deccan Volcanic
 311 Province. The strong contrasts between the orogenic belts of the Himalaya and Tibet
 312 and peninsula India is very evident in Figure 3. We can also recognise the area masked

313 by the Deccan Volcanic Province as having strong cratonic affinities.

314 3.2. Moho depth

315 Variations in the depth to the crust are large across the Indian region, ranging from
316 a minimum of 29 km to a maximum of 88 km in depth. The shallowest crust is for
317 stations close to the coastal plains of the peninsular India, and the deepest under the
318 highest mountain ranges of Himalaya. The crustal thickness variation presented in Figure
319 4 represents results from 615 seismic stations from various studies (Figure 4). The
320 regions of southern Indian shield and Himalaya-Tibet collision zone are well covered,
321 while gaps exist in central India, mainly in the Indo-Gangetic belts and Bastar and
322 Singbhum Cratons.

323 In the Dharwar Craton a clear cut difference is observed between the different com-
324 ponents. The Eastern Dharwar Craton exhibits a thinner crust, ~ 35 km, while in the
325 Western Dharwar Craton average crustal thickness estimates are ~ 45 km. An interesting
326 and unusual observation is a very thick crust, in the central portion of Western Dharwar
327 Craton (Gupta et al., 2003). One of the reasons suggested, is that the thickened crust
328 beneath a mid-Archaean (3.4 Ga) greenstone belt has not been subjected to any ma-
329 jor deformation, and may preserve an ancient crust of around 60 km thickness (Gupta
330 et al., 2003). The same seismic station (MTP, Table S1) has been reported with a crustal
331 thickness of 45 km (Julià et al., 2009), which is preferred for creating the map (Figure
332 4).

333 The crust of the Eastern Dharwar Craton with a sharper Moho is thinner and more
334 transparent than the Western Dharwar Craton, with an absence of any intracrustal layer
335 (Sarkar et al., 2003). The absence of any seismically distinct mafic cumulates overlying
336 the Moho and an intermediate to felsic nature of Eastern Dharwar crust, rather than
337 complex nature of Western Dharwar Craton crust was reported by almost all workers.
338 Receiver functions for the Western Dharwar Craton stations are more complex and show
339 a gradational transition from crust to mantle (Gupta et al., 2003; Sarkar et al., 2003).

340 The Southern Granulite Terrain, has a relatively thick crust (~ 42 km), close to that for
341 the Western Dharwar Craton.

342 An interesting observation is that the crustal thickness increases as we advance in
343 the Himalayan collision zone, stations in the Himalayan regions have crustal thickness in
344 the range 40–50 km, while the central Tibet exhibits a highly thickened crust (around
345 75 km). A few regions such as the Tarim Basin and the Qaidam Basin clearly stand out
346 from the Tibetan plateau with a thinner crust (Figure 4).

347 3.3. Heat flow

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

356 The regions with heat-flow values lower than the average value, 60 mWm^{-2} , are re-
357 garded as low-heat-flow and those with heat-flow above the average above are classified
358 as high-heat-flow (Figure 6). Continental heat flux exhibits lower value in stable regions
359 compared to the active provinces. Moreover there are some discrepancies observed in
360 correlation of surface heat flux with crustal thickness while comparing between tecton-
361 ically active region having high heat flux ($>65 \text{ mWm}^{-2}$) with anomalously thin (<30
362 km) or thick (>55 km) crust and stable region having relatively low surface heat flux
363 ($<65 \text{ mWm}^{-2}$) with little variations in crustal thickness (Mareschal and Jaupart, 2013).
364 It is also apparent in our case where a distinct correlation exists between high-heat-flow
365 and thick crust in the Himalayan collision zone as well as in Tibet. Certain parts of
366 peninsular India the Cambay basin, the Godavari Graben and some parts of the Singhb-
367 hum craton show higher heat flow, which does not link with their crustal thickness. The

368 thinner crust of the Qaidam Basin correlates well with a low heat flow value. Thick crust
369 mostly relates to high heat flow values, but no correlation has been found for normal
370 to thin crust. This may be due to the diversity in crustal compositions as well as the
371 evolution of the heat production in the crust (Mareschal and Jaupart, 2013).

372 *3.4. Comparison of Receiver functions and DSS*

373 We can supplement the passive seismic results in some areas with work using con-
374 trolled sources. Deep Seismic Sounding studies provide information through the full
375 thickness of the crust and are available for profiles extending over more than ten thou-
376 sand kilometres across the Indian landmass. The locations of the various controlled
377 source profiles are shown in Figure 7.

378 The crustal thickness picks been made across various Deep Seismic Sounding profiles
379 covering Indian shield and parts of Himalaya. The picks are made at regular intervals
380 from modelling results for the profiles 1-29 (Figure 7), with due allowance for the drastic
381 changes in the depth of the Moho. For Tibet, the crustal thickness are directly taken from
382 Deep Seismic Sounding and Deep Seismic Reflection profiles as provided in Xiaosong et al.
383 (2009). The station elevation is added to the specific profiles where the Moho estimates
384 are provided with respect to mean sea level. We then employ these crustal thickness
385 values in a similar way to Figure 4 to display the available thickness results from the
386 deep seismic sounding. The map of depth to crust (Figure 8) has sparse coverage, but
387 strong contrasts are evident. The thickened Tibetan crust, as compared to Indian shield
388 and intermediate depths in Tarim and Qaidam basins comes out clearly. Although the
389 controlled source results are somewhat limited, they provide very useful information to
390 fill in some of the gaps in the coverage of the crustal thickness from receiver functions.

391 We examine the variations and similarities in the results from Deep Seismic Sounding
392 and receiver functions. The overall pattern is of a good linear correlation, but there are
393 just a few outliers, notably in the Himalayan region where the differences reach 20 km
394 (Figure 9). These differences are likely to be due to varying influences from strong velocity
395 gradients in the different styles of analysis, as receiver functions are more sensitive to

396 discrete velocity jumps while refraction are to velocity gradients (Kennett and Salmon,
397 2012).

398 The good general compatibility between the results from passive and controlled source
399 seismic studies enables us to combine the two data sets and so produce a more complete
400 map of the depth to crust across the Indian region (Figure 10). The combined map using
401 data points from Deep Seismic Sounding and receiver functions (Figure 10), agrees well in
402 the areas already covered by passive seismic studies (Figure 4), but extends the coverage
403 into a number of important regions (Cambay Rift, Narmada-Son Lineament, Mahanadi
404 Rift, Himalaya and parts of Deccan Volcanic Province). The pattern of variation of
405 the depth to Moho across the Indian subcontinent is well defined, but shows substantial
406 differences from available global models such as CRUST1.0 (Laske et al., 2013), with
407 differences as large as 20 km in Tibet and along the east coast of peninsular India (Figure
408 11).

409 **4. Lithospheric mantle**

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

423 frequencies are lower for S, so depth resolution is reduced and gradient zones may appear
424 as apparent discontinuities.

425 *4.1. Seismic Wavespeed variations*

426 Results from surface wave tomography are very limited for peninsular India, because
427 only the HYB station and some stations in Sri Lanka and the Maldiv islands are available
428 for open access. The situation is much improved in the eastern Himalayan region because
429 of the many temporary deployments of seismometers in different parts of Tibet and
430 adjacent regions of China, but coverage diminishes to the west. Figure 2 of Pandey et al.
431 (2014) illustrates the very strong contrast in available path densities in the two regions
432 when using regional stations. Global studies (e.g. Debayle et al., 2005; Schaeffer and
433 Lebedev, 2013) can provide some improvement in coverage by exploiting longer paths
434 traversing India and its environs, but resolution remains limited.

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

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

448 Because most sampling in body-wave tomography is near vertical, except where re-
449 gional earthquakes can be exploited, vertical smearing of structure is very common.
450 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 localized 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 (Figure 12). Studies using S-p receiver functions (Kumar et al., 2007; Devi et al., 2011; Kumar et al., 2013b) cover much of India and give estimates in the range 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 80-155 km (Mitra et al., 2006; Bhattacharya et al., 2009). 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 (Figure 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 – 4.5 km s^{-1} beneath the Indian shield, much lower than found elsewhere beneath 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 presence of two westerly dipping interfaces at depths of 150 and 200 km. However, they refrain to reporting 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 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 (Mitra et al., 2006; Kiselev et al., 2008; 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 does 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

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

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

563 Concepts that have been employed to try to reconcile the results from tomography
 564 and receiver functions, are the differences between chemically and thermally defined litho-
 565 sphere, the presence of mid-lithospheric discontinuities and a depleted fertile lithosphere
 566 (O'Reilly and Griffin, 2010). Even within tomographic studies many differences arise
 567 depending on the data employed, particularly beneath the Indian shield (Singh et al.,
 568 2014; McKenzie and Priestley, 2008).

569 The Lehmann discontinuity, at depths around 220 ± 30 km beneath continents could
 570 be associated with the bottom of lithosphere, marking a transition to asthenospheric
 571 anisotropy (SH leading SV) related to present day flow (Gung et al., 2003). Such a
 572 scenario is present beneath India, where APM related asthenospheric anisotropy, with
 573 SH leading SV is reported using direct S and SKS/SKKS phases (Kumar and Singh, 2008;
 574 Saikia et al., 2010). Reports of the Lehmann discontinuity beneath India and Tibet are
 575 meagre, (Sharma and Ramesh, 2013; Kiselev et al., 2008; Kosarev et al., 2013; Heit et al.,
 576 2010; Oreshin et al., 2011; Wittlinger et al., 2004), but suggest a deeper boundary ($>$
 577 200 km) than that commonly inferred from S-to-p receiver functions. Beneath the stable
 578 parts of Indian shield, the Lehmann discontinuity may define the lithosphere boundary
 579 (Gung et al., 2003), revealed in body wave tomographic images.

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

583 *4.3. Effects of lithospheric thickness variation*

584 Various studies have proposed that the exceptionally fast motion of the Indian tectonic
 585 plate has been facilitated by the thinning of the Indian lithosphere (Negi et al., 1986;
 586 Kumar et al., 2007, 2013b). However body-wave tomographic studies do not suggest
 587 strong lithospheric thinning. The issue of the fast drift of the Indian tectonic plate is not
 588 that simple, and needs to be understood in terms of effects of various possible candidate
 589 mechanisms. Slab pull and ridge push models are considered to be possible mechanisms
 590 for the fastest drift of Indian tectonic plate (Copley et al., 2010; Capitanio et al., 2013).

591 The role of plumes has been also considered to facilitate the fast drift of Indian tectonic
 592 plate. A substantial push force due to the Réunion plume is thought to be responsible for
 593 the speed-up and subsequent slowdown of Indian plate (Capitanio et al., 2013), although
 594 the issues of the time period of anomalous Indian plate motion and of plume activity
 595 along with its impact on acceleration of India remains unsolved (Muller, 2011). The
 596 presence of the mantle conveyor belt deduced from mantle density distributions may
 597 also have played a role in the breakup and be responsible for the indentation of Indian
 598 and Arabian plates (Becker and Faccenna, 2011). The presence of such a conveyor belt
 599 may impact on the extent of slab pull and ridge push forces, which then contribute as a
 600 secondary driving mechanisms with mantle upwelling and active drag as primary (Becker
 601 and Faccenna, 2011).

602 *4.4. Seismic anisotropy*

603 The rapid movement of the Indian plate (with velocities reaching up to 20 cm/yr) and
 604 its quite variable surface geology makes it unique in comparison to other shield regions.
 605 In this context, seismic anisotropy parameters extracted from SKS/SKKS phases allow
 606 the investigation of a number of significant geodynamic questions: (a) do the distinct
 607 geologic provinces show disparate anisotropic character? (b) does the fast moving plate
 608 force deformation in the plate motion direction? or (c) is anisotropy from the past tec-
 609 tonic events frozen in the lithosphere? Although large data sets of seismic anisotropic
 610 parameters have been acquired from various Precambrian shields world wide, published
 611 measurements from the Indian shield are limited before 2006 (Singh et al., 2006). A few
 612 measurements exploiting analogue data, reveal dominance of Absolute Plate Motion re-
 613 lated strain (Ramesh et al., 1996). The results from a single seismic station SHIO located
 614 over Shillong plateau, show anisotropic directions parallel to the strike of mountain belts
 615 (Vinnik et al., 1992). Later on results using HYB and SHIO (Chen and Ozalaybey, 1998)
 616 and HYB (Barruol and Hoffmann, 1999) seismic stations are found to be null. Such ob-
 617 servations were interpreted in terms of the isotropic nature of Indian lithospheric mantle
 618 and onset of seismic anisotropy in Tibet north of ITSZ as a marker for northern limit of

619 Indian lithosphere (Chen and Ozalaybey, 1998). Now with a flood of seismic high quality
620 data sets in the recent decade it is possible to examine the results from SKS/SKKS and
621 direct-S phases for a complete picture of deformational nature of the Indian plate.

622 In the period from 2006 to 2013, more than 1000 individual splitting parameters
623 using SKS/SKKS and 231 using direct S waves have been presented in various studies
624 (Figure 13). The results using SKS/SKKS phases clearly reveal absolute plate motion
625 related strain as the dominant process beneath Indian plate (Kumar and Singh, 2008),
626 with a possibility of modified asthenospheric flow around lithospheric keels. The delay
627 times between the split S phases are ~ 1 s, as observed for continental shields globally.
628 The prominent trends observed are (a) NS to NE orientation of the fast axis in southern
629 part of peninsular India, (b) NE orientation in north to central India, (c) NE orientation
630 in Himalayan region east of the Sikkim Himalaya, (d) EW orientation parallel to the
631 mountain belts in northeast India, and e) in the Himalaya and Tibet collision zone the
632 orientation is roughly EW, parallel to major suture zones. Although attempts have
633 been made to characterize the seismic anisotropic character of the stable and actively
634 deforming regions of the Indian plate, a number of issues still remain to be addressed.
635 Since SKS splitting provides an integrated effect of both the crust and the mantle, the
636 crustal contribution to anisotropy needs to be isolated. Approaches that exploit the
637 splitting of Ps phases converted at the Moho to estimate the strength and orientation of
638 the fast axis, in a manner similar to SKS-splitting techniques, need to be adopted. The
639 utility of Ps phases in estimating crustal anisotropy has been demonstrated (McNamara
640 and Owens, 1993), although not widely applied. Studies of crustal anisotropy together
641 with those of the mantle should enable an improved understanding of the nature and
642 extent of coupling between the crust and mantle.

643 As is well known, SKS/SKKS phases are only sensitive to horizontal anisotropy.
644 Where expected deformation is complex as in the Andaman and Burmese arc regions,
645 analysis of splitting of S waves from deep earthquakes (to minimize the source side
646 anisotropy) and its modelling in terms of a dipping axis of symmetry can be undertaken.

647 However, separating the effects of mid-mantle and receiver side anisotropy is a challenge
648 that must be met when using direct-S waves. The results from Indian shield stations
649 using direct-S waves clearly brings out a close similarity with the SKS/SKKS results
650 (Saikia et al., 2010), as illustrated in Figure 12.

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

669 5. The 410 and 660 Mantle Discontinuities

670 Major discontinuities associated with mineralogical phase transitions are the man-
671 tle transition zone at 410 and 660 km. The properties of these discontinuities provide
672 important constraints on the thermal and compositional nature of the mantle. The min-
673 eralogical phase transition from olivine α -phase to the modified spinel β -phase at the 410

674 km discontinuity and the ringwoodite γ -phase to perovskite and magnesiowustite at the
 675 660 km discontinuity match well with the seismic results. Variations in the depths of these
 676 discontinuities in response to vertically oriented mantle temperatures are anti-correlated
 677 with positive Clapeyron slope ($\partial P/\partial T$) at the 410 km discontinuity and negative Clapey-
 678 ron slope at the 660 km discontinuity. The result is a thickened mantle transition zone in
 679 presence of trapped cold lithospheric slabs within the transition zone, and early returns
 680 from the 410 and 660 in the case the slabs or lithospheric roots lie well above the 410.
 681 The mantle transition zone beneath India is not akin to those observed globally beneath
 682 other Precambrian shields. Receiver functions results are frequently used to map the to-
 683 pography of 410 and 660 km discontinuities (Figure 14). These images for the behaviour
 684 at 410 and 660 km phase arrivals are produced based on the results of individual stations,
 685 rather than piercing points, except for those from the work of Singh and Kumar (2009).
 686 Thus, they represent a broad picture of the mantle beneath India and Tibet, rather than
 687 being suitable to map local variations.

688 *5.1. 410 km discontinuity*

689 The vertical passage time for seismic P waves to the 410 km discontinuity is close
 690 to that predicted by the *ak135* reference model beneath the key cratonic segments of
 691 the Indian shield: (Kumar and Mohan, 2005; Ramesh et al., 2005; Rai et al., 2009b;
 692 Kiselev et al., 2008; Oreshin et al., 2011; Kosarev et al., 2013; Kumar et al., 2013a).
 693 This 'normal' behaviour is rather different from that for other shield regions around the
 694 globe, where thick, fast, lithospheric roots have resulted in an elevated 410 discontinuity
 695 (Kumar et al., 2013a). Consistently early passage times, sometimes even close to ~ 2 s
 696 early, are seen for the 410 km discontinuity from the stations located in the Himalayan
 697 mountain belts and its foothills: (Oreshin et al., 2008, 2011; Kumar et al., 2013a; Kosarev
 698 et al., 2013). The continuity of early arrivals is also maintained for most parts of Ti-
 699 bet, south of the Bangong Nuijiang Suture Zone. The northwestern Deccan Volcanic
 700 Province and Southern Granulite Terrain show a delay in the passage times of ~ 1.5 s
 701 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 (Yuan et al., 1997; Kind et al., 2002). However, later studies argue in favour of detached lithospheric fragments within the transition zone itself (Singh and Kumar, 2009; Chen and Tseng, 2007).

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 (Wood, 1995; Smyth and Frost, 2002), (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 create 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 depths 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 discon-

730 tinity is also reported at few places beneath India (Oreshin et al., 2011; Kosarev et al.,
731 2013; Kumar et al., 2013a), but it remains to be seen how widespread is such a feature,
732 and what the full implications are for water in the transition zone.

733 5.2. 660 km discontinuity

734 Early vertical passage times compared with the *ak135* reference model (Kennett et al.,
735 1995), sometimes even close to ~ 2 s early are seen for the 660 km discontinuity for regions
736 of Western Tibet, Central India and few parts of Himalaya and Tibet. The early arrivals
737 of 660 near Himalayan collision belt may relate to present day subduction or the the
738 preserved keel of Indian shield (Kosarev et al., 2013), as they are well correlated with
739 the elevation of the 410 discontinuity. Apart from early arrivals or 410 and 660 in the
740 Himalaya, a thick transition zone in the northern part of Indian-subcontinent reflects
741 resting ground for Tethyan slabs within the transition zone itself (Singh and Kumar,
742 2009). Regions of the Southern Granulite Terrain and South West Deccan Volcanic
743 Province show delayed 660 arrivals, as also observed for 410. The delayed arrivals suggest
744 plume effects above 410 and 660 km, in those regions.

745 5.3. X-discontinuity

746 Hydration reactions in subduction zones, crystallographic transitions of pyroxene,
747 and the coesite to stishovite transition or exsolution of stishovite from clinopyroxenes
748 containing excess silica, have been used to explain discontinuities observed around 300
749 km depth (Williams and Revenaugh, 2005; Schmerr et al., 2013; Shen et al., 2014). The
750 intermittent nature of this feature may be due to lateral changes in the fraction of basalt
751 (Schmerr et al., 2013). The reports of this X-discontinuity (Revenaugh and Jordan,
752 1991) beneath India are few, with possible detection in northeast India (Ramesh et al.,
753 2005) and beneath the Hyderabad station HYB (Bodin et al., 2013). The X-discontinuity
754 inferred from common conversion point stacks of receiver functions in northeast India
755 has been interpreted in terms of melt migration and segregation at these depths (Ramesh
756 et al., 2005). SS precursors also trace such a discontinuity with a shear velocity contrast of

757 $\sim 2\%$ (Deuss and Woodhouse, 2002) beneath India. The presence of the X-discontinuity
758 has implications for the lateral variations and stratification of the mantle, and receiver
759 functions have considerable potential as a tool to track at this depth (Yuan et al., 1997;
760 Wittlinger et al., 2004; Ramesh et al., 2005; Bodin et al., 2013). Careful examination
761 of the seismic data from India may throw more light on the nature of X-discontinuity,
762 which may often be overlooked due to its intermittent nature and low shear impedance
763 contrast (Schmerr et al., 2013).

764 6. Summary

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

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

776 Beneath India the crust is quite variable in its depth beneath the surface and there
777 are indications also of significant changes in character (see supplementary Figure S2 and
778 S3). The cratonic regions generally have crustal thickness in the range 40-45 km, which
779 is typical for cratons in other parts of the world (cf. Kennett and Salmon, 2012). Though
780 there are limited zones, e.g. in the central portion of the Western Dharwar Craton, that
781 have rather thicker crust (more than 55 km). The Moho deepens rapidly as the Himalayan
782 collision zone is approached, and the crust reaches its maximum thickness beneath Tibet;
783 the crustal thickness exceeds 75 km in parts of the Lhasa terrain. The crustal thickness

784 map built from the passive seismic and deep seismic sounding studies shows significant
785 deviations from that predicted from global models such as CRUST1.0 (Laske et al.,
786 2013). Such differences are of potential significance whenever crustal corrections have to
787 be made to seismological results.

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

796 The shear-wave splitting results from body-wave studies show the dominance of
797 strain approximately in the direction of absolute plate motion, with an anisotropic Indian
798 mantle for which the delay times between the split S phases are close to 1 s. However,
799 the body wave results do not localise the anisotropic zone in depth. The global study by
800 Debayle et al. (2005) of azimuthal anisotropy using surface waves provides indications of
801 two-layer anisotropy beneath India with alignment to absolute plate motion at the base
802 of the lithosphere or in the asthenosphere beneath; though resolution was limited in their
803 study. The only other location with a similar configuration in this study is Australia, with
804 current rapid motion, which shows a distinct change in anisotropic character between 100
805 and 200 km depth.

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

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Figure 1: Simplified geological map of India and Tibet: SGT, Southern Granulite Terrain; CG, Closepet Granite; WDC, Western Dharwar craton; EDC, Eastern Dharwar craton; CB, Cuddapah Basin; DVP, Deccan volcanic province; NSL, Narmada-Son Lineament; GG, Godavari Graben; BC, Bastar craton; CTB, Chattishgarh Basin; VNB, Vindhyan Basin; BUC, Bundelkhand craton; DAFB, Delhi-Aravalli Fold Belt; SP, Shillong Plateau; MKH, Mikir Hills; MHR, Mahanadi Rift; CBR, Cambay Rift; MFT, Main Frontal Thrust; MBT, Main Boundary Thrust; MCT, Main Central Thrust; ITSZ, Indus-Tsangpo Suture Zone; BNSZ, Bangong-Nujiang Suture Zone; JSZ, Jinsa Suture Zone.

Figure 2: (a) The Poisson's ratio and (b) the average crustal velocity (V_s) from passive seismic studies at the stations marked with inverted triangles. The maps are produced using results from Saul et al. (2000); Kumar et al. (2001); Sarkar et al. (2003); Gupta et al. (2003); Kumar et al. (2004); Sherrington et al. (2004); Rai et al. (2003, 2005); Mitra et al. (2005); Li et al. (2006); Tiwari et al. (2006); Pathak et al. (2006); Mitra et al. (2008); Jagadeesh and Rai (2008); Li et al. (2009); Julià et al. (2009); Rai et al. (2009a); Xu et al. (2010); Bai et al. (2010); Chen et al. (2010); Singh et al. (2010); Wang et al. (2010); Acton et al. (2011); Kayal et al. (2011); Singh et al. (2012); Yue et al. (2012); Hazarika et al. (2013a); Xu et al. (2013b,a); Zha and Lei (2013); Borah et al. (2014a); Kumar and Mohan (2014). A total of 457 data points for the Poisson's ratio and 499 data points for V_s are employed (the specific values are presented in Supplementary Table S1). A search radius of 100 km has been used to initialize the gridding of the crustal data. A nearest neighbour interpolation scheme has been used to produce this and all the subsequent images in the paper (Wessel and Smith, 1998). As certain data points may have multiple estimates, the uncertainties in terms of standard deviation are represented in Figure S1.

Figure 3: Histograms for crustal parameters, crustal thickness, average Poisson's ratio, average P velocity, and average S velocity for specific geological regions. Each box is divided with distinct background colors (e.g., with a split at 0.25 for Poisson's ratio) to ease comparisons.

Figure 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 Saul et al. (2000); Kumar et al. (2001); Sarkar et al. (2003); Gupta et al. (2003); Kumar et al. (2004); Rai et al. (2003, 2005); Mitra et al. (2005); Tiwari et al. (2006); Pathak et al. (2006); Mitra et al. (2008); Jagadeesh and Rai (2008); Julià et al. (2009); Rai et al. (2009a); Singh et al. (2010); Acton et al. (2011); Kayal et al. (2011); Singh et al. (2012); Hazarika et al. (2013a); Borah et al. (2014a); Kumar and Mohan (2014); Li et al. (2014); Mandal (2012); Vinnik et al. (2007), supplemented by results from Tibet ((Sherrington et al., 2004) and (Li et al., 2014) and references therein).Figure 4 employs data points from broadband receiver functions, and the crustal surface is interpolated using a search radius of 100 km.

Figure 5: Heat flow values taken from the global heat flow data base of the International Heat Flow Commission (<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 crust (Figure 4)

Figure 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 $1 \times 1^\circ$. The histogram in the right panel shows heat flow estimates at particular nodes split into two regimes, < 60 and ≥ 60 mWm⁻².

Figure 7: Deep seismic sounding profiles over various geological provinces of the India regions. The crustal thickness values picked along profiles are shown by red rectangles at their actual position. The same values have been used to make the crustal image in Figure 7. The profile end points are shown by white circles. 1: Behera (2011); 2: Reddy and Rao (2003); 3: Kaila et al. (1979); 4: Kaila et al. (1987); 5: Kaila et al. (1990); 6: Kaila et al. (1990); 7: Kaila et al. (1981c); 8: Kaila et al. (1981c); 9: Behera et al. (2004); 10: Behera et al. (2004); 11: Behera et al. (2004); 12: Kaila et al. (1992); 13: Rao et al. (1999); 14: Kaila et al. (1992); 15: Rao et al. (1999); 16: Reddy et al. (2000); 17: Murty et al. (2008); 18: Kaila and Rao (1986); 19: Kaila et al. (1985); 20: Kaila et al. (1989); 21: Prasad and Rao (2006); 22: Kaila et al. (1981a); 23: Kaila et al. (1981b); 24: Kaila (1981); 25: Powers et al. (1998); 26: Prasad et al. (2011); 27: Prasad et al. (2011); 28: Powers (1996); 29: Powers et al. (1998); 30: Xiaosong et al. (2009) and references therein.

Figure 8: Map of crustal thickness, using published results from deep seismic sounding.

Figure 9: Correlation of crustal thickness estimates from Deep Seismic Sounding (DSS) vs RF(receiver functions) (left panel). Values are extracted at common nodes for which both crustal depth estimates are available, interpolated at $1 \times 1^\circ$. In the right panel the differences are plotted at particular node points, where color and symbol both represents the difference in the depth estimates.

Figure 10: Map of crustal thickness combining all available information from receiver functions and Deep Seismic Sounding.

Figure 11: The difference in thickness obtained from CRUST1.0 Laske et al. (2013) and using data from receiver functions and Deep Seismic Sounding. The negative and positive values show thin and thick crust respectively with respect to CRUST1.0.

Figure 12: (a) The lithospheric thickness estimates as observed in various receiver functions studies (Kumar et al. (2007); Oreshin et al. (2008); Patro and Sarma (2009); Rychert and Shearer (2009); Ramesh et al. (2010); Devi et al. (2011)). (b) lithospheric thickness estimates from surface wave tomography (Priestley and McKenzie, 2013).

Figure 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).

Figure 14: (a) The vertical passage time of seismic P waves to the 410 km discontinuity as reported by various studies: He et al. (2014); Kumar et al. (2013a); Kosarev et al. (2013); Oreshin et al. (2011, 2008); Kayal et al. (2011); Singh et al. (2009); Kiselev et al. (2008); Shen et al. (2008); Pathak et al. (2006); Kosarev et al. (1999); 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 Saul et al. (2000); Ramesh et al. (2005); Kumar and Mohan (2005); Singh et al. (2012); Sharma and Ramesh (2013). (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.