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Earth and Planetary Science Letters 237 (2005) 252–263

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Mantle discontinuities beneath the Deccan volcanic province

M. Ravi Kumar [a,*], G. Mohan [b]

[a]National Geophysical Research Institute, Uppal Road, Hyderabad - 500 007, India
[b]Department of Earth Sciences, Indian Institute of Technology Bombay, Powai, Mumbai - 400 076, India

Received 15 December 2004; received in revised form 20 April 2005; accepted 20 June 2005
Available online 27 July 2005
Editor: V. Courtillot

Abstract

Mantle discontinuities beneath the Deccan volcanic province (DVP) are imaged using about 900 seismograms from a network of six broadband stations deployed in the DVP along a 350 km long north–south profile, paralleling the west coast of India. Teleseismic receiver function analysis of the data enabled distinct identification of the Moho, 410 and 660 km discontinuities. Presence of sub-Moho low velocity zones with velocity reductions in the range of 0.1–0.4 km/s have been brought out through modeling of receiver function stacks at individual stations. The arrival times of P410s and P660s are normal, as predicted by the IASP91 model, and the mantle transition zone thickness is 252 km, which is close to the global average. This study reveals that the upper mantle beneath DVP, south of the Narmada rift does not exhibit any anomalous signatures associated with a mantle plume. A shield-like lithosphere, coupled with the presence of an upper mantle low velocity zone confined to shallow depths beneath DVP, imposes constraints on the mechanisms feasible for Deccan volcanism. The present study does not favour the plume impact/plume incubation models, instead necessitates searching for clues offshore west coast of India and exploring alternate mechanisms.

Keywords: Deccan volcanic province; upper mantle structure; receiver functions

1. Introduction

The Deccan volcanic province (DVP) of India (Fig. 1) is a classic example of continental flood basalts (CFB) that belongs to the genre of large igneous provinces which represent the most voluminous volcanic events on earth. The eruption of the Deccan basalts 65 Ma ago, is understood to have taken place rapidly in less than a million years [1], spanning an areal extent of about 0.5 M km^2 on the continent alone. The Deccan volcanism is commonly attributed to the upwelling of a deep mantle plume beneath the northerly drifting Indian subcontinent in the late Cretaceous [2]. The time progressive chain of volcanic ridges including the Laccadives–Maldives–Chagos ridges and Mascarene plateau in the Indian ocean are believed to represent the hotspot track linking the DVP to the Reunion hotspot [3]. The

* Corresponding author.
E-mail addresses: mravi@ngri.res.in (M. Ravi Kumar), gmohan@iitb.ac.in (G. Mohan).

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doi:10.1016/j.epsl.2005.06.034

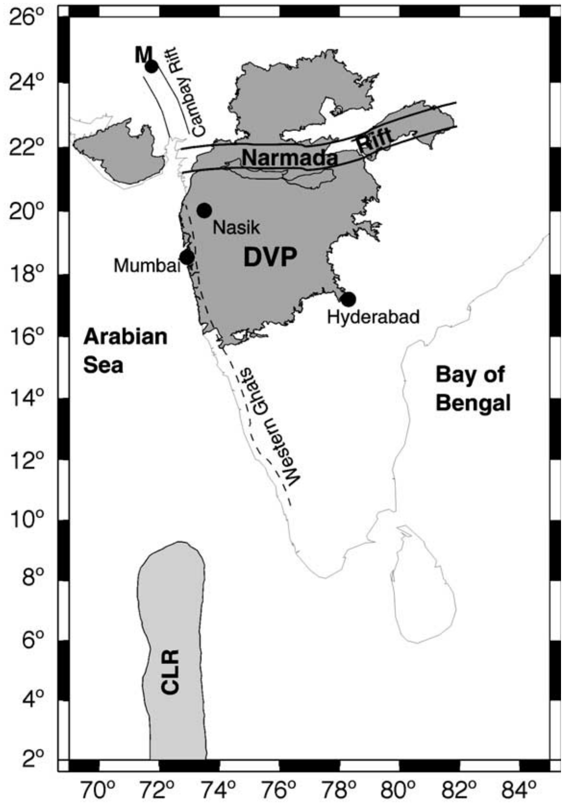

Fig. 1. Map of the Indian subcontinent demarcating the Deccan volcanic province (DVP) along with the major tectonic features. The abbreviation used M: Mundwara alkaline complexes; CLR: Chagos Laccadive Ridge.

alkaline magmatism at Mundwara in north Cambay, further north of the present outcrop of the DVP, is interpreted to represent the earliest signatures of the mantle plume [4]. However, a pre-Deccan marine phase of early Reunion hotspot was suggested [5] based on cretaceous volcanic rocks in the south Tethyan suture zone of Pakistan. Most theories proposed for the evolution of the flood basalts were summarized by Courtillot et al. [6] to fall into the category of active models viz., plume initiated [3] or passive models viz., rifting due to regional plate tectonic stresses [7] while others propose a combination of both models [6,8]. Alternative non-plume models invoke the edge driven mechanism, a near surface process arising due to discontinuity in the thickness of the lithosphere [9]. However, the proposed mechanisms ranging from plume to non-plume models remain debatable [10]. While young volcanic provinces like Hawaii and Iceland reflect the thermal and compositional imprints of the mantle source in terms of seismic wave speeds, (eg. [11]) such evidences over CFBs are largely restricted to a few regions like Parana [12] and Columbia river basalts [13]. Possible signatures of a mantle plume in DVP in terms of a low seismic wavespeed anomaly in the upper mantle beneath north Cambay which extends from shallow depths down to a more extensive low velocity zone below 200 km were mapped through teleseismic P-wave tomography [14]. This seismic anomaly is seen to link to the Narmada rift and possibly extends further south beneath the Western Ghats, the main effusive phase in the DVP, where the tomograms were not well resolved. Earlier, tomographic studies using teleseismic data recorded by a portable analog network deployed south of the Narmada rift, indicated a high velocity root beneath the DVP extending down to 300 km, with marginal evidence of upper mantle low velocities in the westernmost part of DVP coinciding with the west coast rift [15]. However, an important aspect that remains to be addressed is the mantle layering and the spatial disposition of mantle discontinuities – the products of the evolutionary process of the earth, which places vital constraints on the mechanisms feasible for the evolution of the DVP.

While discontinuities at several depths in the mantle are reported in the literature [16], only the primary discontinuities (410 and 660 km) are globally observed while the second order discontinuities eg., Hales, Gutenberg, Lehmann are weak and observed regionally. The 410 and 660 km discontinuities are commonly interpreted to be the result of phase transformations of olivine to high pressure forms producing sharp gradients in the earth's density and velocity structure [17]. The phase boundaries thought to produce P410s and P660s have positive and negative Clayperon slopes, respectively, and thus, variations in mantle transition zone thickness can be interpreted as variations in thermal anomalies [18]. These discontinuities which are sensitive to abrupt changes in composition, mineralogy, temperature and mantle fabric, are diagnostic of the physical and chemical state of the mantle. In the recent past, the receiver function technique using P to s converted phases, emerged as an effective tool to image the signatures of mantle plumes by mapping the disposition of the upper mantle discontinuities, as in the case of Hawaii [11,19]. Although DVP is not centered over an active hotspot, it is expected to retain and reflect the thermal/chemical perturbations caused due to plume–lithosphere interaction in terms of variation in the lithospheric thickness and/or variations in the topography of the mantle discontinuities. For instance, a low velocity anomaly of chemical origin extending down to 600 km, anchored to the asthenosphere has been reported for the 135 Ma Parana [12], which is much older than the 65 Ma DVP. The seismic tomographic studies in DVP done earlier [14,15] map the smooth parameter variations but not seismic discontinuities which provide additional information on the nature of the upper mantle. The present study is the first experiment in India utilizing an array of broadband seismic stations to search for the imprints of Deccan volcanism by imaging the mantle discontinuities beneath the DVP using receiver function analysis.

2. Study region

The study area (Fig. 1) encompasses the western segment of the Deccan volcanic province, south of the Narmada rift, comprising of the spectacular and high rising (average elevation 1 km) Western Ghats which represent the main effusive phase of the Deccan basalts. The combined stratigraphic thickness of the predominantly tholeiitic lava flows of the Western Ghats is a maximum of 3 km which gradually thins to a few meters to the south and east at the periphery of the DVP [20]. There is no significant age difference between the lava flows from the base to the top of the Western Ghats and the mean age of the entire Western Ghats sequence is 67.5 Ma [21]. Courtillot et al. [22] emphasise that the flood basalts were erupted in a short time, centered on chron 29 R and cover only three magnetic chrons and thus amount to no more than a million years, with a large part of the lava volume concentrated over a short period of time overlapping the Cretaceous–Tertiary boundary. These eruptions are understood to have originated on the continent at the northern end of the Western Ghats, northeast of Mumbai (Fig. 1), with the formation of a volcanic ridge running south along the Western Ghats [23]. The study region is bounded on the north by the east–west trending Precambrian Narmada rift system. The Arabian sea to the West comprises of stretched continental crust intruded by volcanic bodies related to the Deccan traps [24]. Courtillot et al. [6] suggest that rifting took place along the east west trending Gop rift (in Arabian sea) and the Narmada rift which subsequently failed and was replaced by the active north–western Indian ocean ridge along the western coast of India.

3. Data and methodology

In this study, the P-to-s conversions are used to image the mantle discontinuities down to a depth of 700 km beneath the DVP, south of the Narmada rift, using teleseismic data from broadband stations along a 350 km long profile paralleling the west coast of India (Fig. 2). The data comprises of approximately 900 3-component seismograms from six broadband stations. Of these, two permanent stations (KARD, PUNE) are being operated by the India Meterological Department (IMD) since 1997 and two temporary stations were operated in a campaign mode at 4 locations (MULG, MPAD, VARE and ALMN) by the Indian Institute of Technology Bombay (IITB). The broadband station at MULG was operated for three and a half years since 1999 while the stations at MPAD and VARE were operated for one year each during 2001–2003. Station ALMN was in operation for only six months and produced only a small amount of data. Teleseismic events in the distance range 30°–100°, mostly from the eastern azimuths, and a few (about 63 events) from the western azimuths, are used.

The approach adopted for computation of receiver functions (RFs) involves rotation of the Z, N and E components into a ray coordinate system using optimized back azimuth and incidence angles [25]. This decomposes the wave field into its P, SV and SH components. The converted phases are then isolated from the P-coda by deconvolving the P from the SV component by simple spectral division using a water level stabilization. In this study, a Gaussian filter centered around 0.5 Hz is used to enhance the signals from the upper mantle discontinuities while using a 2 Hz filter for those within the crust. To compare the RFs at different slowness values and to distinguish

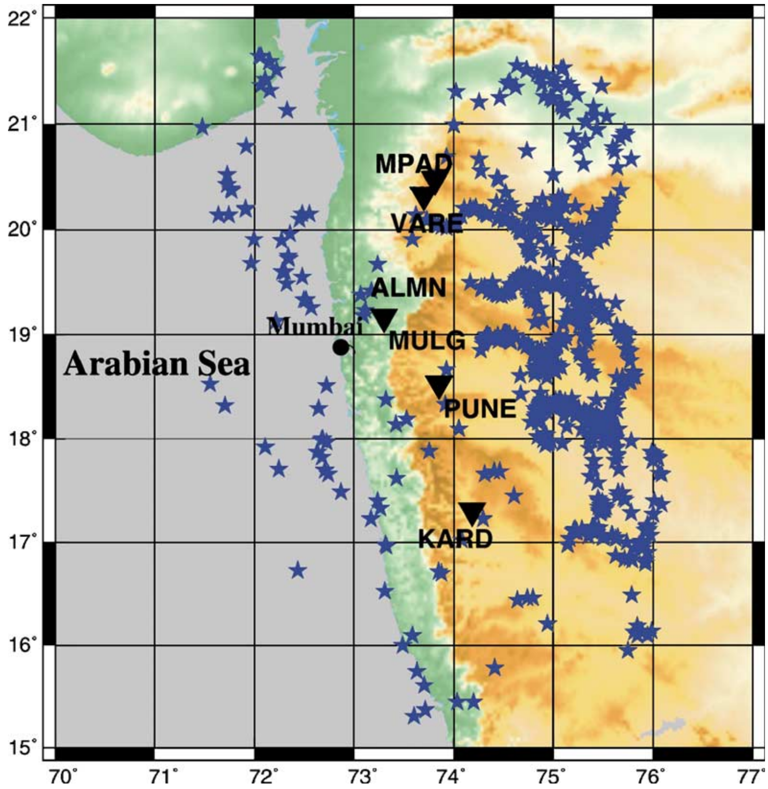

Fig. 2. Map showing stations used in the study (inverted triangles). Stars indicate the piercing point locations of the P-to-s converted phases at a 520 km depth.

multiples from converted phases, a moveout correction is applied separately for the converted phases and multiples which is made for a reference slowness of 6.4 s/°. This reference slowness is chosen since it corresponds to an epicentral distance of about 67° in the IASP91 model, which is close to the mean epicentral distance of the events used for RF analysis (30° to 100°). The times of the receiver function samples are corrected to the reference distance by either stretching the receiver functions corresponding to distances larger than the reference distance or compressing those which are smaller.

4. Waveform modeling

The receiver functions depicting the response of the crust and the uppermost mantle, moveout corrected for the converted phases and stacked in narrow slowness bins are shown in Fig. 3 for all the stations. The conversion from the Moho (Pms) can be traced as a sharp phase close to 4.3 s at most stations, parallel to the slowness axis, while the corresponding multiples (Ppms and Psms) are seen as inclined features. The Moho at station ALMN appears too feeble due to the small number of data used. The crustal thicknesses and Poisson's ratios determined earlier at PUNE, KARD and MULG are 36 ± 2.5 km and 0.26 ± 0.01 [26,27]. Importantly, the crust is simple, devoid of any prominent intracrustal layers, felsic to intermediate in composition with no evidence of underplating to testify the effect of volcanism. A negative phase that corresponds to a distinct low velocity zone (LVZ) in the shallow upper mantle follows the Pms phase at most stations with varying amplitudes. This negative phase is prominent at stations VARE, PUNE and MULG and is weak at others (Fig. 3).

In order to quantify the crustal structure and constrain the seismic velocities in the LVZs, the

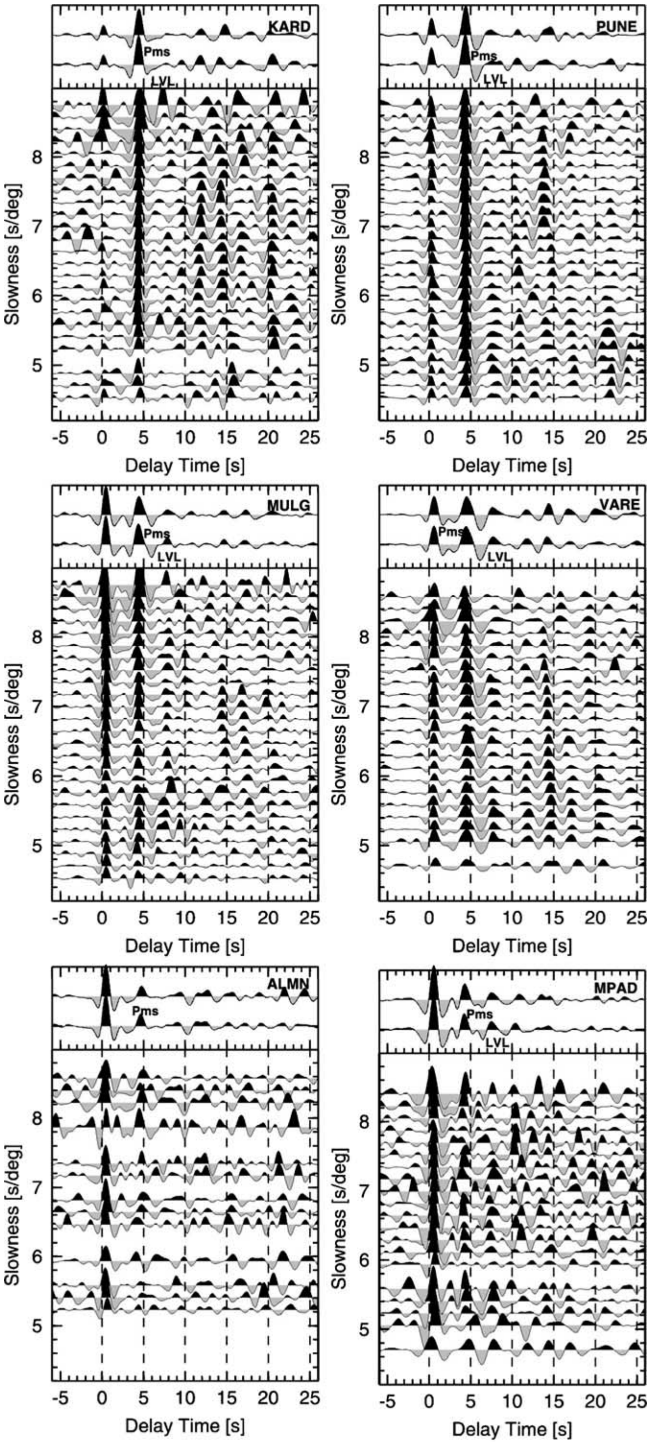

Fig. 3. Receiver functions for all the stations moveout corrected for the converted phases and stacked in narrow slowness bins. For each section, the top frame contains two summation traces. The bottom one corresponds to the stack of the RFs moveout corrected for converted phase and the top one corresponds to the stack of the RFs moveout corrected for multiples.

stacks of the moveout-corrected traces at each station are inverted. Since the receiver functions are only sensitive to the velocity contrasts across discontinuities, the inversion procedure is constrained by the near surface velocities used for rotating the receiver functions into the LQT system and the Poisson's ratio estimates from our earlier publications [26,27]. The models (Fig. 4) that best fit the observed receiver function stacks (Fig. 5) indicate a fairly uniform crustal thickness close to 36 km beneath all the stations. Commensurate with the observed sharp Moho conversions (Pms) the velocity jump across the Moho is largest for stations VARE, KARD, PUNE and MPAD. Similarly, the velocity contrasts needed to explain the sub-Moho LVZs vary across the stations. While a contrast of nearly 0.4 km/s is required to fit the LVZ at VARE, a velocity reduction of 0.2 km/s matches the observed RFs at PUNE, MULG and MPAD. At KARD, the order of velocity contrast is only about 0.1 km/s. In view of the paucity of data, we did not invert the data for ALMN. From the above modeling it appears that presence of a LVZ beneath the DVP cannot be ruled out. Similar LVZ was also identified through travel times and relative amplitude modeling of a seismic record section from the Koyna DSS profile near KARD, at relatively shallow depths of about 56 km, which were interpreted to be associated with zones of weakness and lower viscosity [28] probably unrelated to the origin of DVP.

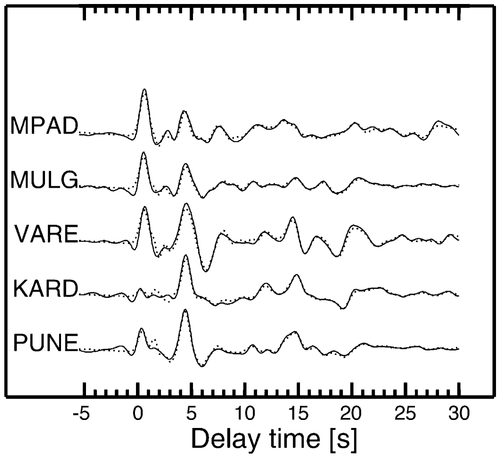

Fig. 5. Waveforms fit between the observed receiver function stacks (solid) and the synthetics (dotted) generated by the models corresponding to each station.

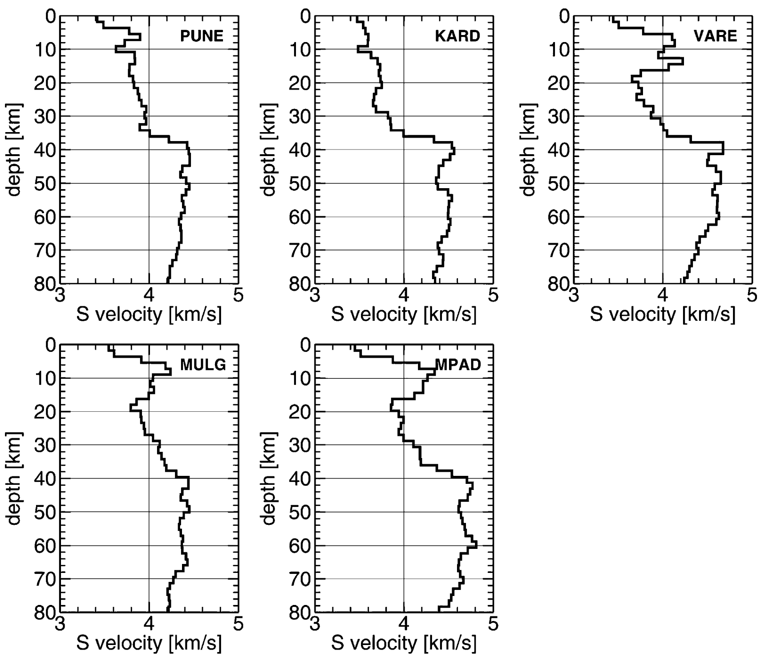

Fig. 4. Models obtained by inversion of the stacked receiver functions at each station.

5. Imaging of the upper mantle

To facilitate detection of signals especially from the deeper interfaces, the moveout-corrected RFs from all the stations are combined, binned and stacked in narrow slowness bins. The composite image of the receiver functions as a function of slowness, corresponding to the upper mantle beneath DVP is shown in Fig. 6. This image clearly traces the conversions from 410 and 660 km as distinct features linearly aligned at 44.4 and 68.3 s, respectively. These times are similar to those predicted by IASP91 model for the standard earth. The difference in arrivals of these phases is 23.9 s, close to the global average of 24 s. The errors in the arrival times of the P410s and P660s obtained from summation of all the receiver function traces have been estimated using the bootstrap resampling technique [29,30]. In this scheme, a single receiver function stack is generated by summation of those traces decided by a random number generator where the number of traces for summation remains the same as the total number of RFs. This procedure is repeated 200 times, by stacking a totally different sequence of randomly generated trace numbers. The mean and standard deviation of the times of the converted phases are then calculated from the values picked from these 200 realisations. The errors in the estimated times were found to be ± 0.1 s through this bootstrapping procedure.

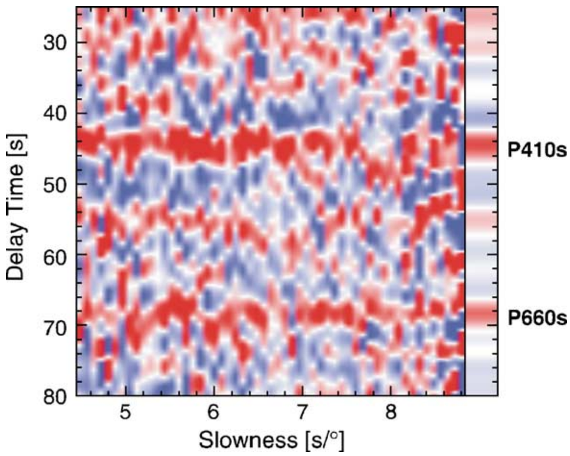

Fig. 6. Composite image of the SV response of the upper mantle utilizing moveout-corrected receiver functions from all stations. The 410 and 660 phases are labelled next to the summation stacks on the right.

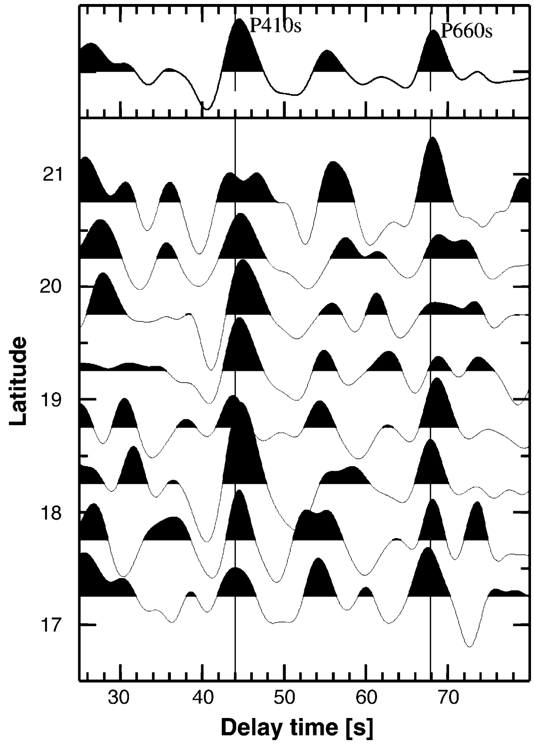

Fig. 7. Moveout-corrected receiver functions sorted from South to North based on piercing points at a 520 km depth and stacked, averaged within 0.5° bins.

To study the spatial variations of the 410 and 660 km discontinuities beneath the array the receiver functions are binned based on the locations of the piercing points at 520 km depth. These piercing points for all the events (Fig. 2) are obtained by ray tracing using the IASP91 model based on the slowness of the P-wave. A depth of 520 km was chosen, since it is almost midway between the 410 and 660 km discontinuities. The location of the piercing points at 410 (660) km depth is 118 (216) km offset from the station location for an event at an epicentral distance of 67°. The receiver functions whose piercing points lie in a 2° (longitude) × 0.5° (latitude) are binned after moveout correction and then stacked. These stacked RFs sorted from south to north from 17–21° N are shown in Fig. 7 with the cumulative stack of all RFs on the top. This plot indicates little variability in the arrivals of the P410s and P660s phases across the 450 km stretch from south to north indicating a relatively flat topography of both the discontinuities. Such a two dimensional sampling geometry minimizes the effects of any variable discontinuity structure perpendicular to the array. The piercing point plot provides a much higher resolution of mantle discontinuities in comparison with the single station stacks.

6. Discussion

The mechanisms suggested for the generation of the enormous melt volumes observed on CFBs commonly invoke the plume model. A widely accepted model is the passive model [7] based on the physics of partial melting at rift zones which suggests rifting of the continental lithosphere over a thermal anomaly caused due to a mantle plume and subsequent outpouring of basalts by decompressional melting of hot asthenospheric mantle. Alternatively, in the starting plume model [8], basalts are formed, again, by decompression melting at shallow depths within the mushrooming head of an anomalously hot mantle plume. Presence of a large plume head at such shallow depths implicitly assumes a thin lithosphere to overlie it. The plume incubation model [31] suggests that the plume head resides at the base of the lithosphere, where melting occurs over a period of time during which the lithosphere is gradually thinned from below, with subsequent outpouring of basalts. A common requirement for the rapid effusion of large volumes of flood basalts would therefore be a thermal anomaly coupled with a thin lithosphere. Seismic evidence for the interaction of a mantle plume with a lithosphere in terms of thermal and/or chemical alterations is manifested as extensive low velocity zones in the upper mantle [32]. The best known example is the upper mantle low velocity anomaly of chemical origin, extending down to the mantle transition zone beneath Parana [12,16]. The changes in the velocity characteristics of the upper mantle due to a mantle plume above the transition zone are usually reflected in terms of delays of the P410s and P660s phases. This topographic variation is interpreted to be due to thermal anomalies in the upper mantle but at short scale lengths the velocity gradients associated with phase transformations of the garnet-pyroxene component of the mantle [33], chemical layering [33] amongst other interpretations are also suggested as sources of discontinuity topography.

7. Sub-Moho low velocity zone (LVZ)

The presence of a shallow upper mantle LVZ assumes significance in the context of Deccan volcanism, since LVZs are attributed to thermal/chemical anomalies possibly related to the source of magmatism. Shallow sub-Moho low velocity zones are observed beneath young volcanic provinces like Arabia [34] and Basin and Range provinces [35]. However, unlike the DVP, both Basin and Range provinces and Arabia, are young volcanic provinces, where the thermal anomaly is more dominant. A large thermal/chemical anomaly associated with Deccan volcanism would tend to affect a large region more or less uniformly. However, it is observed that the velocity reductions are strong at some stations eg., VARE and relatively weak at other stations in DVP. Further, this LVZ has little influence on the timings of the P400s and P660s phases implying that it does not have a deep extension into the mantle and is confined to the shallow upper mantle as also suggested by Krishna et al. [28]. These observations are more consistent with local upper mantle heterogeneity than with any large thermal/chemical anomaly. Low velocities within a zone indicate presence of magma or fluids, since even small amounts of melts or fluids (<1–2%) can decrease the seismic velocity substantially [36]. Presnell and Gudfinnsson [37] find that LVZs can be caused by melting due to the presence of carbonate which reduces the solidus temperature by >300 °C at pressures >1.9 GPa corresponding to depth ranges of the observed LVZs. Thus, melting of locally occurring carbonated lherzolite and carbonated eclogite in the upper mantle would result in variable local reductions in velocities [37]. Thus, a velocity reduction due to partial melting can be due to several causes and does not necessarily require elevated temperatures. In light of the above, it is preferred to attribute the origin of the observed LVZs to local causatives rather than to a large scale process that is required to facilitate eruption of large volumes of basalts seen in the DVP.

8. 410 and 660 km discontinuities

The present study reveals that the arrival times of the converted phases from the primary mantle discontinuities beneath the DVP are close to the global average at 44.4 and 68.3 s, respectively. These times translate to depths of 413 and 665 km using the IASP91 velocity model. In view of non-availability of an upper mantle velocity model suitable for the study region, the IASP91 model seems to be a viable alternative. Although a thick cold lithosphere is inferred from teleseismic tomographic studies [38] they do not provide information about the velocity–depth function in the study region. Additionally, the modeled sub-Moho low velocity zone does not appear strong enough to influence the average depths derived from the P410s and P660s conversion times close to typical shield values. Therefore our depth determinations of the 410 and 660 km discontinuities though not absolute are not likely to differ substantially from those in the standard earth models.

Travel time modeling of primary, underside and free surface multiple reflections sparsely sampling tectonically diverse regions of the Indian shield tentatively reveal presence of the upper mantle discontinuities at average depths of 430 and 680 km [39]. This study however, does not mention about values that are specific to regions like the DVP. Receiver function analysis of data from the GEOSCOPE station at Hyderabad (HYB) on the Eastern Dharwar Craton, image the 410 and 660 km discontinuities at depths of 409 and 659 km, respectively [40]. These values compare well with those from the DVP. A global study of the upper mantle discontinuites reveals that these discontinuities show strong correlation and are affected by the mantle heterogeneities between the Moho and the 410 km discontinuity [41]. According to the Deccan plume hypothesis [14] the region of lowered seismic velocities extending to the mantle transition zone represent the initial conduit through which plume material forced its way through the lithosophere beneath the Cambay rift, thus implying that the initial plume head needs to have laminated to the Indian lithosphere during the latter's passage over the Reunion hotspot. This upper mantle low velocity anomaly of thermal origin in north Cambay as evidenced by the high heat flow values in the Cambay rift, suggested to extend further south of Narmada rift into the present study region should have manifested as deepening of both the 410 and 660 km discontinuities. There are several reports of anomalies in the upper mantle above the mantle transition zone (MTZ) which have an effect on the topography of the 410 km discontinuity. For example, an excellent correlation between the track of the Yellowstone hotspot (YHT) and the discontinuity structure is the 20 km depression in the 410 km discontinuity from 415 km beneath the NW margin of the YHT to 435 km beneath the easternmost extent of the Basin and Range faulting [42]. Similarly, a delay of >1.5 s is observed for both the upper mantle discontinuities beneath Arabia [34] which is underlain by an extensive zone of low velocities down to the MTZ [43]. Such variations in the depths to the upper mantle discontinuities is not observed beneath DVP. Lack of significant variations of delay times of P410s and P660s in DVP and their similarity with the values from HYB on the Eastern Dharwar craton suggests absence of anomalous mantle structure related to volcanism. These findings are at variance from those observed in other flood basaltic provinces by Vinnick and Farra [44] who report a super deep low velocity layer in the upper mantle due to a strong reduction in S velocity at around 360 km and also at depths of 280–300 km in the eastern segment of the Kaapvaal craton in southern Africa and a similar feature in the Tunguska basin of the Siberian platform. Earlier, a depression in the 410 km discontinuity observed beneath southern Africa through the P-to-s converted phases was interpreted to be due to the presence of a low S velocity immediately above the discontinuity [45]. However, other studies [46] do not find any such evidence of an LVZ beneath southern Africa and on the contrary find that the P410s and P660s are earlier than predicted by the IASP91 model suggesting that the upper mantle velocities are higher by about 3.5% than the IASP91 model. Such anomalous time excursions – either early or delayed arrivals with respect to the IASPEI model – are not observed beneath DVP. The present high resolution studies probing down to a 660 km depth indicate an unperturbed upper mantle beneath DVP. This study also complements and supplements the earlier shallow upper mantle investigations through body- and surface-wave tomographic studies [15,47].

9. Mantle transition zone (MTZ)

Imaging the upper mantle discontinuities also provides useful information about the thickness of the mantle transition zone which provides a measure of the mantle temperature. The MTZ thickness is primarily associated with the temperature within MTZ, which if low results in an increase in velocity and consequently due to the positive and negative clapeyron slopes at the top and bottom of the MTZ, the thickness increases and vice versa. Chevrot et al. [41] also noted that any correlation between the properties of the upper mantle at depths <410 km and the thickness of the MTZ is absent. Hence, unusually high/low temperatures in the transition zone (TZ) should therefore result in thinning/thickening of the TZ. The classic Yellowstone continental hotspot which is underlain by low velocities in the upper 200 km does not exhibit any variation in the thickness of the MTZ indicating that the source is shallow and not a deep mantle plume [48]. Similarly, Liu et al. [16] report that thinning of the MTZ is not observed beneath the Parana basaltic province inspite of detectable presence of a low velocity anomaly (<1% $\delta V_p/V_p$) of 300 km dimension at depths close to 500 km [12]. This observed anomaly is explained invoking a thermal anomaly of 200 °C in combination with compositional effect. However, based on a normal MTZ beneath Parana, Liu et al. [16] conclude that either the thermal anomaly does not extend into the transition zone or alternatively that the observed anomaly is dominantly compositional in origin rather than being primarily thermal. In DVP, the difference in arrivals of the P410s and P660s phases is 23.9 s, which translates to a MTZ thickness of 252 km which is close to the global average. A normal MTZ therefore suggests absence of anomalous temperature within the MTZ and the normal 410 and 660 times argue for lack of anomalous upper mantle structure close to these depths. This observation further supports our arguments that the delineated LVZ beneath DVP is neither uniformly strong nor extends deep into the mantle to influence the MTZ. These results seem to suggest that the upper mantle and the MTZ beneath the DVP are largely unperturbed though embedded with a subcrustal low velocity zone.

10. Conclusions

Overall, the upper mantle architecture of the DVP does not indicate anomalous mantle structure that could be linked to the origin of the Deccan flood basalts. Absence of variations in the 410 and 660 km mantle discontinuities implies that the upper mantle in this region is devoid of any significant thermal or compositional anomalies associated with the interaction of the Indian lithosphere with a mantle plume head. Absence of upper mantle anomalies in this region is intriguing since the Indian plate is presumed to have drifted over the Reunion hotspot which is identified to be originating from the deepest part of the lower mantle at the D'' layer [22]. Further, a large instantaneous plume head upwelling beneath the Indian plate should have resulted in alterations in the lithospheric configuration. The present day heat regime in the DVP, averaging 44 mW/m^2, does not reflect any thermal transients associated with the Deccan volcanism [49]. The present study reveals that the DVP is underlain by a lithosphere which is neither anomalously warm nor thin, with a normal MTZ. These results, coupled with the presence of an upper mantle LVZ confined to shallow depths beneath the DVP, impose constraints on the mechanisms suggested for Deccan volcanism. Absence of clinching evidences in the continental lithosphere supporting the plume impact/incubation models in the DVP necessitates searching for clues offshore west coast of India and also exploring the feasibility of alternate mechanisms.

Acknowledgements

We thank two anonymous reviewers and Dr. Courtillot for their extremely useful suggestions which helped in improving the manuscript. We thank Dr. D.S. Ramesh for his critical review and Dr. J. Saul for his programs. The financial support provided by the State Government of Maharashtra is gratefully acknowledged. We thank India Meteorological Department for providing the data for two stations.

References

[1] V. Courtillot, J. Besse, D. Vandamme, R. Montigny, J.J. Jaeger, H. Cappetta, Deccan flood basalts at the Cretaceous/Tertiary boundary? Earth Planet. Sci. Lett. 80 (1986) 361–374.

[2] W.J. Morgan, Deep mantle convection plumes and plate motions, Bull. Am. Assoc. Pet. Geol. 56 (1972) 203–213.

[3] M.A. Richards, R.A. Duncan, V. Courtillot, Flood basalts and hot spot tracks: plume heads and tails, Science 246 (1989) 103–107.
[4] A.R. Basu, P.R. Renne, D.K. Dasgupta, F. Teichmann, R.J. Poreda, Early and late alkali igneous pulses and a high ^{3}He origin for the Deccan flood basalts, Science 261 (1993) 902–906.
[5] J.J. Mahoney, R.A. Duncan, W. Khan, E. Gnos, G.R. McCormick, Cretaceous volcanic rocks of the south Tethyan suture zone, Pakistan: implications for the Reunion hotspot and Deccan traps, Earth Planet. Sci. Lett. 203 (2002) 295–310.
[6] V. Courtillot, C. Jaupart, I. Manighetti, P. Tapponier, J. Besse, On causal links between flood basalts and continental breakup, Earth Planet. Sci. Lett. 166 (1999) 177–195.
[7] R.S. White, D.P. McKenzie, Magmatism at rift zones; the generation of volcanic continental margins and flood basalts, J. Geophys. Res. 94 (1989) 7685–7729.
[8] L.H. Campbell, R.W. Griffiths, Implications of mantle plume structure for the evolution of flood basalts, Earth Planet. Sci. Lett. 99 (1990) 79–93.
[9] S.D. King, D.L. Anderson, An alternate mechanism of flood basalt formation, Earth Planet. Sci. Lett. 136 (1995) 269–279.
[10] H. Sheth, A historical approach to continental flood basalt volcanism: insights into pre-volcanic rifting, sedimentation and early alkaline magmatism, Earth Planet. Sci. Lett. 168 (1999) 19–26.
[11] X. Li, R. Kind, K. Priestley, S.V. Sobolev, F. Tilmann, X. Yuan, M. Weber, Mapping the Hawaiian plume conduit with converted seismic waves, Nature 405 (2000) 938–941.
[12] J.C. VanDecar, D.E. James, M. Assumpcao, Seismic evidence for a fossil mantle plume beneath South America and implications for plate driving forces, Nature 378 (1995) 25–31.
[13] R.D. Catchings, W.D. Mooney, Crustal structure of the Columbian plateau, evidence for continental rifting, J. Geophys. Res. 93 (1988) 459–474.
[14] B.L.N. Kennett, S. Widiyantoro, A low seismic wavespeed anomaly beneath northwestern India: a seismic signature of the Deccan plume? Earth Planet. Sci. Lett. 165 (1999) 145–155.
[15] D.S. Ramesh, D. Srinagesh, S.S. Rai, K.S. Prakasam, V.K. Gaur, High-velocity anomaly under the Deccan Volcanic Province, Phys. Earth Planet. Sci. 77 (1993) 285–296.
[16] K.H. Liu, S. Gao, P.G. Silver, Y. Zhang, Mantle layering across central South America, J. Geophys. Res. 108 (2003) 2510, doi:10.1029/2002JB002208.
[17] C.R. Bina, G. Hellfrich, Phase transition Clapeyron slopes and transition zone seismic discontinuity topography, J. Geophys. Res. 99 (1994) 15853–15860.
[18] S. Lebedev, S. Chevrot, R. Van der Hilst, Seismic evidence for olivine phase change at the 410 and 660 km discontinuities, Science 296 (2002) 1300–1302.
[19] X. Li, R. Kind, X. Yuan, I. Wolbern, W. Hanka, Rejuvenation of the lithosphere by the Hawaiian plume, Nature 427 (2004) 827–829.
[20] K.L. Kaila, Mapping the thickness of Deccan trap flow in India from DSS studies and inferences about a hidden Mesozoic basin in the Narmada–Tapti region, in: K.V. Subba rao (Ed.), Deccan Flood Basalts, Mem. Geol. Soc. Ind., vol. 10, 1988, pp. 91–116.
[21] R.A. Duncan, D.G. Pyle, Rapid eruption of the Deccan flood basalts at the Cretaceous/Tertiary boundary, Nature 333 (1988) 841–843.
[22] V. Courtillot, A. Davaille, J. Besse, J. Stock, Three distinct types of hotspots in the Earth's mantle, Earth Planet. Sci. Lett. 205 (2003) 295–308.
[23] P.R. Hooper, The timing of crustal extension and the eruption of continental flood basalts, Nature 345 (1990) 246–249.
[24] J.A. Malod, L. Droz, B. Mustafa Kemal, P. Patriat, Early spreading and continental to oceanic basement transition beneath the Indus deep-sea fan: northeastern Arabian Sea, Mar. Geol. 141 (1997) 221–235.
[25] L.P. Vinnik, Detection of waves converted from P to SV in the mantle, Phys. Earth Planet. Inter. 15 (1977) 294–303.
[26] M.R. Kumar, J. Saul, D. Sarkar, R. Kind, A.K. Shukla, Crustal structure of the Indian shield: new constraints from teleseismic receiver functions, Geophys. Res. Lett. 28 (2001) 1339–1342.
[27] G. Mohan, M.R. Kumar, Seismological constraints on the structure and composition of western Deccan volcanic province from converted phases, Geophys. Res. Lett. 31 (2004) L02601, doi:10.1029/2003GL018920.
[28] V.G. Krishna, K.L. Kaila, P.R. Reddy, Low velocity layers in the subcrustal lithosphere beneath the Deccan Trap region of western India, Phys. Earth Planet. Inter. 103 (1991) 621–637.
[29] B. Efron, R. Tibshirani, Bootstrap methods for standard errors, confidence intervals, and other measures of statistical accuracy, Stat. Sci. 1 (1986) 54–77.
[30] K.G. Dueker, A.F. Sheehan, Mantle discontinuity structure beneath the Colorado Rocky Mountains and High Plains, J. Geophys. Res. 103 (1998) 7153–7169.
[31] R.W. Kent, M. Storey, A.D. Saunders, Large igneous provinces: sites of plume impact or plume incubation? Geology 20 (1992) 891–894.
[32] D.L. Anderson, T. Tanimoto, Y.-S. Zhang, Plate tectonics and hotspots: the third dimension, Science 256 (1992) 1645–1651.
[33] D.L. Anderson, Theory of the Earth, Elsevier, New York, 1989.
[34] M.R. Kumar, D.S. Ramesh, J. Saul, D. Sarkar, R. Kind, Crustal structure and upper mantle stratigraphy of the Arabian shield, Geophys. Res. Lett. 29 (8) (2002), doi:10.1029/2001GL014530.
[35] H.M. Benz, I. McCarthy, Evidence for an upper mantle low velocity zone beneath the southern Basin and Range–Colarado plateau transition zone, Geophys. Res. Lett. 21 (1994) 509–512.
[36] G.M. Mavko, Velocity and attenuation in partially melted rocks, J. Geophys. Res. 85 (1980) 5173–5189.
[37] D.C. Presnell, G.H. Gudfinnsson, Carbonate rich melts in the oceanic low velocity zone and deep mantle, in: G.R. Foulger, J.H. Natland, D.C. Presnell, D.L. Anderson (Eds.), Plates, plumes and paradigms, Geol. Soc. Am., Spec. Pap. 388 (in press).
[38] H.M. Iyer, V.K. Gaur, S.S. Rai, D.S. Ramesh, C.V.R. Rao, D. Srinagesh, K. Suryaprakasam, High velocity anomaly beneath

the Deccan volcanic province, evidence from seismic tomography, Proceed. Indian Academy of Sciences, vol. 98, 1989, pp. 31–60.

[39] V.G. Krishna, Propagation of regional seismic phases in the Indian shield: constraints on crustal and upper mantle velocity models, B.S.S.A. 94 (1) (2004) 29–43.

[40] J. Saul, M.R. Kumar, D. Sarkar, Lithospheric and upper mantle structure of the Indian Shield from teleseismic receiver functions, Geophys. Res. Lett. 27 (2000) 2357–2360.

[41] S. Chevrot, L. Vinnik, J.P. Montagner, Global scale analysis of the mantle Pds phases, J. Geophys. Res. 104 (1999) 20203–20219.

[42] K.G. Dueker, A.F. Sheehan, Mantle discontinuity structure from midpoint stacks of converted P to S waves across the Yellowstone hotspot track, J. Geophys. Res. 102 (1997) 8313–8327.

[43] L. Vinnik, M.R. Kumar, R. Kind, V. Farra, Super deep low velocity layer beneath the Arabian plate, Geophys. Res. Lett. 30 (7) (2003) 1415–1419, doi:10.1029/2002GL016590.

[44] L. Vinnik, V. Farra, Subcratonic low-velocity layer and flood basalts, Geophys. Res. Lett. 29 (2002), doi:10.1029/2001GL014064.

[45] L.P. Vinnik, R.W.E. Green, L.O. Nicolaysen, G.L. Kosarev, N.V. Petersen, Tectonophysics 262 (1996) 67–75.

[46] S.S. Gao, P.G. Silver, K.H. Liuand the Kaapwal Seismic Group, Mantle discontinuities beneath southern Africa, Geophys. Res. Lett. 29 (2002), doi:10.1029/2001GL013834.

[47] G. Mohan, S.S. Rai, G.F. Panza, Shear velocity structure of the laterally heterogeneous crust and uppermost mantle beneath the Indian region, Tectonophysics 277 (1997) 259–270.

[48] R.L. Christiansen, G.R. Foulger, J.R. Evans, Upper mantle origin of the Yellowstone hotspot, Geol. Soc. Amer. Bull. 114 (2002) 1245–1256.

[49] S. Roy, R.U.M. Rao, Heat flow in the Indian shield, J. Geophys. Res. 105 (2000) 25587–25604.