

**Seismic imaging of the crust and lithospheric
mantle structure along a profile from the west to
the east coast of India covering Dharwar craton**

विद्या वाचस्पति की

उपाधि की अपेक्षाओं की आंशिक पूर्ति में प्रस्तुत शोध प्रबंध

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2024

*I dedicate this thesis to my family,
teachers and well wishers.*

CERTIFICATE

Certified that the work incorporated in the thesis entitled ("**Seismic imaging of the crust and lithospheric mantle structure along a profile from the west to the east coast of India covering Dharwar craton**") submitted by **Dipak Kumar Chaubey (Roll no. 20163451)** was carried out by the candidate, under my supervision. The work presented here or any part of it has not been included in any other thesis submitted previously for the award of any degree or diploma from any other University or institution.



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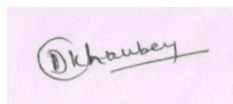
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Synopsis

Understanding the formation and evolution of continents is very crucial as it contains the long preserved evolutionary history of the Earth. Despite a number of studies, many issues remain unanswered or poorly investigated in terms of spatial and vertical resolution of the velocity structures, which includes the nature of Archean cratons having interacted with the tectonomagmatic events like mantle plumes that happened in the past. Their preservation, deformation over time and difference between the nature of Archean and Proterozoic terrains have been the subject of intense research over the past few decades. The detailed knowledge of the outer solid shell of the Earth, referred to as the lithosphere (consisting of crust and uppermost mantle) floating over the highly viscous (semi-solid) asthenosphere drifting with time, is of great importance as it can help us to understand the evolution of continents, crustal growth and crust-mantle interaction as well as the continental drift, reconstruction of continents in the past etc. This requires accurate modelling and understanding the structural and compositional variation of continental lithosphere using robust and reliable geophysical methods like seismology. This thesis aims to provide a well-constrained seismic structure of the crust and lithospheric mantle along a profile from the west to the east coast of South India. The E-W profile crosses through the mid-late Archean Dharwar craton having ~ 1.1 Ga old diamondiferous kimberlite fields and the Proterozoic Cuddapah basin. The western and eastern ends of the profiles separated from the Madagascar (~ 85 Ma) and the Eastern Gondwanaland (~ 130 Ma) respectively under the influence of mantle plumes. Our study area is an ideal natural laboratory to study the above issues with the seismological data having improved resolution.

The significantly improved shear wave velocity image of the lithosphere along a ~ 630 km long profile from the west to the east coast of South India across the Dharwar craton is mapped through the joint inversion of P-receiver functions and Rayleigh wave phase velocity dispersion (3-140 sec) having short periods data (3-30 s) at every ~ 10 km interval using ambient noise and long periods data (30-140 s) at every ~ 55 km using earthquake waveforms. The profile consists of 38 dense broadband seismic stations having 10-15 km spacing, recorded between January, 2012 and April, 2014. Common conversion point (CCP) stacking of the receiver function for the crust is also done along this profile.

Firstly, the shear velocity image of the crust through the joint inversion suggests flat Moho having 38 ± 2 km thick along the E-W profile and not showing any correlation of crustal structure with surface boundaries along the Dharwar representing homogenized crust. The average shear velocity in all the domains along the profile are in the range 3.70-3.75 km/s, suggesting the average composition of the crust to be felsic to intermediate. Along the profile, the upper crust is almost absent and can possibly be interpreted as upliftment and erosion. The middle and lower crust beneath the WDC and the domain to its west (Coast-WG) are ~20-22 km and ~8-10 km respectively while the middle and lower crust beneath the EDC domain are ~26 km and ~4 km showing more mafic crust beneath the WDC and the domain to its west than the EDC domain. This suggests the WDC and the domain to its west have been affected by the past magmatic/mantle plume event like Marion plume (~85 Ma). CB and EG domains show similar middle and lower crust thicknesses like EDC having more felsic behavior than the WDC and the domain to its west and also showing more felsic nature than the global average. Presence of a thin lower crust (~4 km) beneath the EDC would lead to sharp velocity contrast at the crust–mantle boundary. Our velocity image in the western and eastern end doesn't show a signature of extensive underplating due to India-Madagascar separation (~85 Ma) and the Eastern Gondwanaland (~130 Ma) respectively. We only find correlation of crustal structure with surface boundaries beneath the Eastern Ghats (EG) and Cuddapah Basin (CB). Also, we observe the unusual low velocity layer in the upper crust at the boundary of CB and EG representing the signature of Nallamalai Fold Belt (NFB). We don't find any correlation of crustal thickness with age along the profile. On the basis of our results, we don't support the two stage subduction-accretion process for the crustal evolution in the Neoarchean Dharwar Craton (as suggested by previous studies) as we don't find any signatures of dipping layers. There is also a possibility that the tectonic deformations like folding, faulting, and crustal tilting initially created relief on the Moho boundary. However, these structural variations may subsequently got erased by ductile flow showing widespread horizontality (flat) of the Moho as suggested previously which was later questioned. We don't observe strong intra-crustal layering along the profile. So the receiver functions have minimal multiples, leading to improved resolution of the mantle structure and mapping of the Lithosphere-Asthenosphere Boundary (LAB).

Finally, we investigated the preservation, modification and possible destruction of the lithospheric mantle along the same ~630 km long dense seismic station profile. Along this

profile, we developed a present day high resolution shear velocity model of the lithospheric mantle by the joint inversion of receiver function and Rayleigh wave phase velocity dispersion and compared it with the lithospheric thickness of 180 ± 20 km derived from kimberlite xenoliths erupted at ~ 1.1 Ga. Using $V_s \sim 4.6$ km/s or $\delta V_s \sim 2\%$ faster with respect to the global velocity model AK135 as the proxy, except the western end of WDC, we observe thicker lithosphere (160-180 km) and weak/absence (highly viscous) of asthenosphere ($V_s \sim 4.55$ -4.6 km/s) in the Dharwar craton and Cuddapah Basin indicating marginally thinned lithosphere and remaining almost undisturbed and unaffected by the tectonomagmatic events and the weak asthenosphere suggests possible presence of tectosphere. The longevity of the Dharwar craton may possibly be due to this weak asthenosphere beneath it, shielding the lithosphere from the events. The lithospheric mantle of Coastal plain-Western Ghat-Western end of Dharwar and the Eastern Ghat parts show the imprint of India-Madagascar separation (~ 85 Ma) due to Marion plume and eastern Gondwanaland (~ 130 Ma) respectively having thicknesses 100-120 km and 80-90 km respectively showing extensive lithospheric thinning and strong low velocity asthenosphere ($V_s \leq 4.5$ km/s) due to the intrusion of hydrous melts/supercritical fluids and re-fertilization. Except Eastern Ghat, Most part of the shallow lithosphere along the profile is preserved/highly depleted ($V_s \geq 4.7$ km/s; $\delta V_s \geq 4\%$) possibly due to spinel-garnet peridotite transformation. The lower lithosphere along the profile is totally modified ($V_s < 4.7$ km/s). We observe a distinct signature of Mid Lithospheric Discontinuity (MLD) across the diamond-bearing kimberlite field between 100-120 km depth along the E-W profile possibly suggesting accumulation of seismically slow metasomatic minerals (rich in hydrous minerals, pyroxenes and/or carbonates) and the acquisition of fabric due to deformation during craton construction involving plume or collisional tectonics.

Dipak Kumar Chaubey

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Chapter 1 - Introduction

Earth is the only planet where life exists. Formed around 4.6 Ga ago, it consists of internal layers with varying density and composition. Earth is mainly divided into three layers: Crust, Mantle and Core. The crust and uppermost mantle constitute the outer solid shell of the Earth, referred to as lithosphere, where heat flow is mainly by conduction. It is underlain by a layer called asthenosphere in which heat flow occurs mainly by convection. The boundary between the two is called lithosphere-asthenosphere boundary (LAB). The solid lithosphere floats over the highly viscous (semi-solid) asthenosphere and drifts with time. The detailed knowledge of formation of lithosphere provides us the opportunity to understand the evolution of continents, crustal growth and crust-mantle interaction as well as the continental drift, reconstruction of continents in the past etc. A generalised schematic diagram showing the working of the shallow earth is presented below.

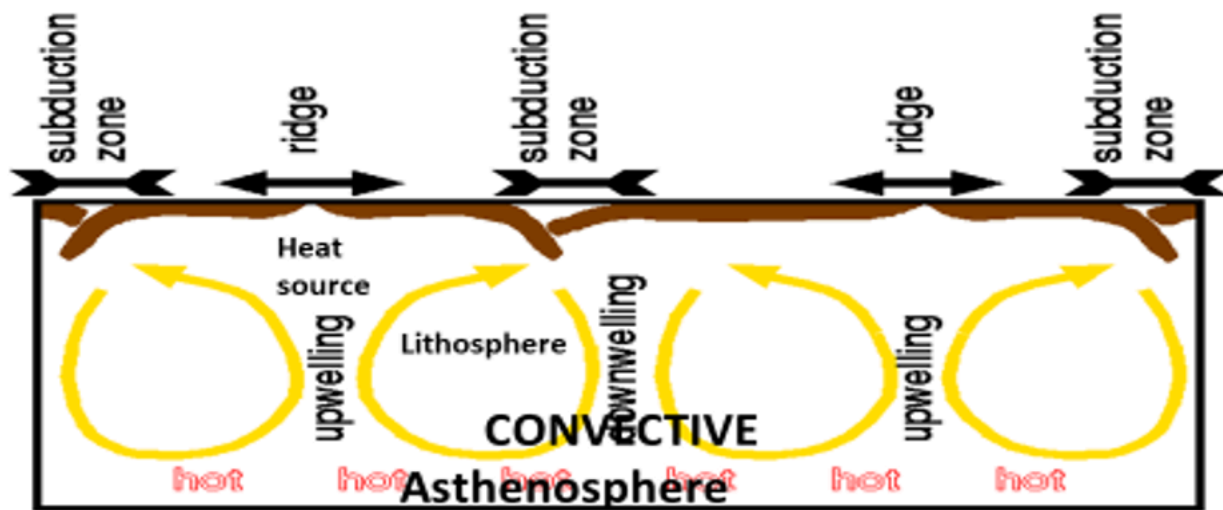


Fig. 1.1. A Generalised schematic diagram showing the working of the shallow earth.

The continental crust occupies ~41% of the Earth surface, of which only ~29% lies above the sea and is easily accessible for geological and geophysical mapping (Taylor & McLennan, 1985). Due to its longevity, the continents preserve the Earth's evolutionary history in its inherited internal structure. This necessitates an accurate modelling and understanding the structural and compositional variation of continental lithosphere using robust and reliable geophysical methods like seismology. Seismological tools use earthquake or chemical explosion

generated body and Surface waves to probe the P and S wave velocities that serve as proxy to image the material property inside the earth. We briefly discuss these below.

1.1 Continental Crust

Outermost layer of the Earth comprises the Continental/Oceanic crust. Continental crust includes the major continents, their margins, and several submerged microcontinents (Artemieva, 2002; Mooney, 2010) and constitutes only 0.40% of the Earth's mass (0.62% of the crust-mantle system) (Taylor & McLennan, 1985). It consists of upper (felsic), middle (intermediate) and lower crust (mafic) divided on the basis of seismic velocity, composition and temperature etc (Christensen & Mooney, 1995). Earlier, the continental crust was considered to have 2 layers: upper layer (granitic composition, $V_p = 6.0$ km/s) and lower layer (basaltic composition, $V_p = 6.5$ km/s) (Mueller, 1977; Mooney, 2010). In many continental regions, there exist boundary between upper granitic to lower basaltic crust, called Conrad discontinuity, first identified by Conrad (1925). This is, however, not well established globally. With newer seismological data emerging from investigations around the world (Christensen & Mooney, 1995; Rudnick & Fountain, 1995, Gao et al., 1998, Artemieva, 2002, Mooney, 2010), a three layer crustal model was evolved with the average thickness of 40 km, $V_p \sim 6.45$ km/s, $V_s \sim 3.65$ km/s and $\rho \sim 2.835$ gm/cc respectively. For an average 40 km thick continental crust: the upper, middle and lower crustal thicknesses are 11 km, 12 km and 17 km respectively. The upper crust ($V_p \sim 5.6$ - 6.3 km/s, $V_s \sim 2.8$ - 3.5 km/s, $\sigma \leq 0.26$), middle crust ($V_p \sim 6.4$ - 6.7 km/s, $V_s \sim 3.5$ - 3.8 km/s, $\sigma > 0.28$) and lower crust ($V_p \sim 6.8$ - 7.6 km/s, $V_s \sim 3.8$ - 4.3 km/s, $\sigma = 0.30$) contain granitic, tonalitic (rich in amphibole) and granulitic rocks corresponding to felsic, intermediate and mafic composition. The crust is defined as the outermost layer having $V_p < 7.6$ km/s and $V_s < 4.3$ km/s and boundary between crust and upper mantle is called Moho discontinuity.

1.2 Lithosphere and Asthenosphere -

Lithosphere is defined as the mechanically strong layer of the Earth which includes the crust and, in general, the shallow mantle (Artemieva, 2011; Lee et al., 2011; Rychert et al., 2020). Seismically, it is a high-velocity layer above the low-velocity zone (LVZ) in the upper mantle, presumably caused by partial melting. The thermal lithosphere is defined as the layer where conductive heat transfer takes place and its base has a constant isotherm of $\sim 1300^\circ\text{C}$ and can be

determined by the change from depleted (lithospheric) to undepleted (asthenospheric) composition. In terms of resistivity, the lithosphere is considered to be the highly resistive layer overlying conductive layer called asthenosphere, and the boundary is called Lithosphere-Asthenosphere Boundary (LAB). Lithosphere moves coherently with the earth's surface over the weaker asthenosphere which is the fundamental principle of plate tectonic theory. Estimation of the LAB or the lithosphere thickness is challenging and bit difficult as the high velocity layer (lithosphere) is overlain by low velocity layer (asthenosphere) and seismic wave propagation tends to steepen rather than turn back to the surface in the wide angle observations. The lithosphere-asthenosphere boundary (LAB) is delineated from surface wave derived V_{sv} (vertical shear wave velocity) models in seismology using various proxies, such as **(i)** The sharpest point of velocity drop beneath the high-velocity lid; **(ii)** The mean depth of the negative velocity gradient beneath the high-velocity lid; **(iii)** An iso-velocity surface, the depth where oceanic and continental velocity profiles converge (Rychert et al., 2020); and **(iv)** The top and bottom of the velocity decrease, assuming a gradual transition. But in many cratons, it is difficult to delineate LAB due to weak/lack of asthenosphere which means the velocity drop is very small or comparable to the surrounding layers suggesting possible presence of tectosphere (Jordan, 1979 & 1988), other domains show a strong asthenosphere/partial melt which happens mainly in proterozoic younger mobile belts.

1.3 Continent evolution models -

Earth is estimated to be 4.6 Ga old. The evolution of continents and the understanding of the operating tectonics and thermal processes is one of the most fundamental problems in geoscience. This has been the subject of research for a long time (details in Jordan, 1979 & 1988; Hawkesworth et al., 1990; Sleep, 2005; Lee et al., 2011; Pearson et al., 2021). The continents are formed primarily during the Archean (3.6-2.5 Ga) due to the specific thermal regime (Lee, 2006). Cratons, by definition, are the oldest parts of continents which have been more or less tectonically inert for at least a billion years. Cratons are characterized by a thick, depleted mantle lithosphere, which exhibits high shear wave velocity (V_s) and low heat flow. The melt depleted composition of the cratonic lithosphere provides positive chemical buoyancy that counteracts the negative thermal density of the cold mantle material. Additionally, the high viscosity of the lithospheric keel, in conjunction with the underlying low viscosity asthenospheric mantle,

minimizes the basal drag force acting on the cratonic root. This stability protects the craton against erosion from convective stresses and localized deformations, contributing to its longevity (Lenardic et al., 2003).

Craton formation can be discussed through three end-member scenarios, as illustrated in below figure 1.2. **(i)** the formation of cratons from the residue of melting within a single hot plume head, **(ii)** the formation through the accretion or stacking of oceanic lithosphere, and **(iii)** the formation by the thickening of arc lithosphere, as discussed in Lee (2006 and 2011).

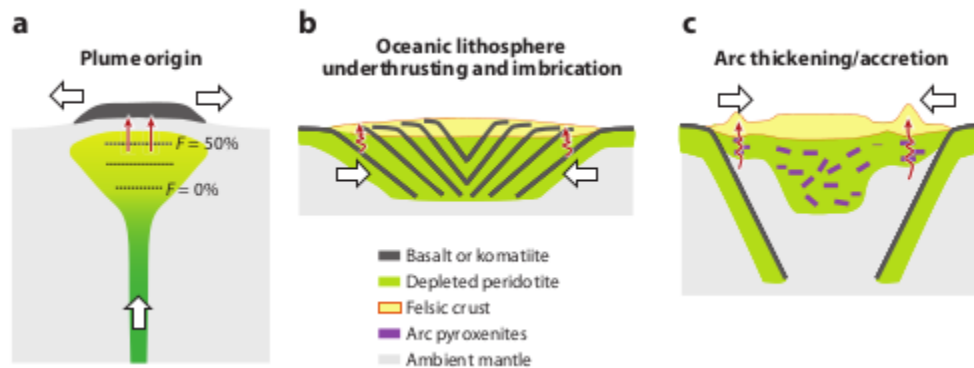


Fig. 1.2. End member models for craton formation. **(a)** Plume head (Vertical tectonics), **(b)** Oceanic lithosphere underthrusting and imbrication (Horizontal tectonics), **(c)** Arc thickening and accretion (Horizontal tectonics).

The first model for the craton formation requires high degrees of melting ($>30\%$) by both the scenarios: Either a more vigorously convecting and hotter Archean mantle or the existence of larger, hotter thermal upwellings, like plumes from the Archean (Figure 1.2a). In order to achieve greater than 30% melting in the pressure range of 3-7 GPa (the pressure range from which the cratonic mantle derives), a potential temperature significantly greater than 1650°C is needed, which exceeds the potential temperatures of the hottest modern plumes ($1400\text{-}1600^{\circ}\text{C}$) and is 200°C higher than the temperatures of the modern mantle beneath oceanic ridges ($1250\text{-}1450^{\circ}\text{C}$) (Herzberg, 2004; Lee, 2006 and 2011). Another aspect of the plume hypothesis posits that the residual mantle column should be stratified, resulting from decompression melting, where the

shallowest mantle is the most depleted (highest Mg#) (McKenzie, 1984; Lee, 2006 and 2011). The high levels of melt extraction (over 30%) necessitate a higher amount of ultramafic magma, such as komatiite (Lee, 2006), which is predominantly found in the Archean.

The second model of craton formation is accretion and stacking of oceanic lithosphere (Helmstaedt and Schulze, 1989). This hypothesis gets motivation from eclogite xenoliths interpreted as oceanic crust protoliths and from seismic studies from Slave craton showing presence of dipping reflectors (Bostock, 1998). The main problem with this hypothesis is the subducting feature of the oceanic lithosphere due to negative buoyancy with cooling. However, the mantle is hotter and vigorously convecting in Archean ($> 200^{\circ}\text{C}$), a thicker and more buoyant oceanic crust is formed.

The third model is formation of thick and highly melt-depleted mantle by thickening of island arcs. This hypothesis is preferred by Jordan (1988) for the formation of the tectosphere. This is different from the previous hypothesis as the arc lithosphere may be buoyant unlike the oceanic lithosphere, which subducts typically.

1.4 Geological Framework

We briefly discuss the framework of Indian geology (Meert et al., 2010). India is the amalgamation of different Precambrian geological domains divided on the basis of age, composition, degree of metamorphism interspersed by major younger lineaments or weak zones. There are five major cratons like Dharwar, Bastar, Aravalli, Singhbhum, Bundelkhand (Fig. 1.3). The younger geological domains include the southern Granulite terrain, The Mawar terrain, Eastern Ghat, Chhota Nagpur plateau, Cuddapah basin, the Deccan traps. The study region (see Figure 1.3 marked in black) encompasses a W-E linear domain from the west to the east coast covering Western ghat, Dharwar craton, Cuddapah basin and Eastern ghat. Details are discussed below:

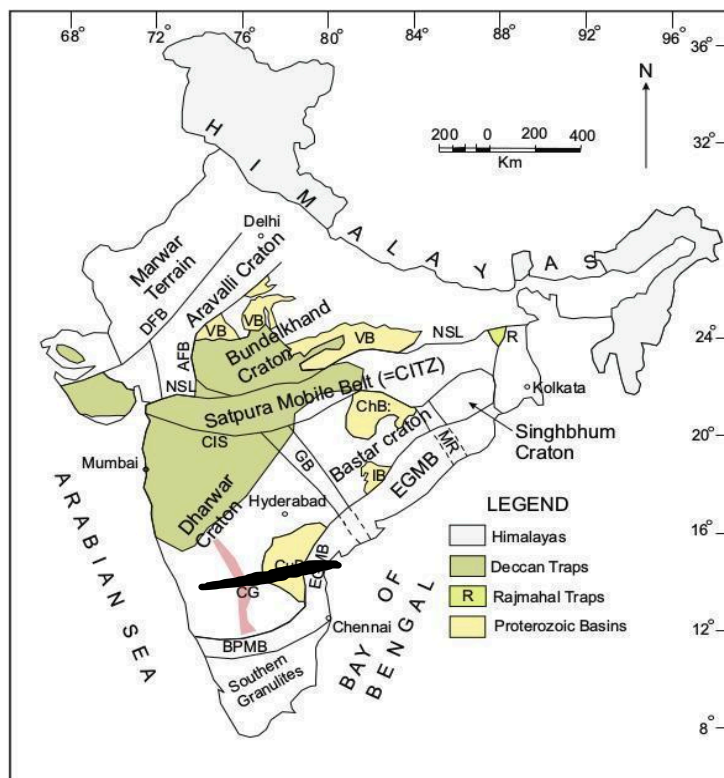


Fig. 1.3. Simplified geological map of India Craton map (Meert et al., 2010) with the study region marked with thick black.

We discuss below the geological framework of the region covered in this study (Figure 1.4). **Dharwar Craton** is the composite mosaic of Archean blocks situated in the southern part of the India, shares its north-western boundary with Deccan Volcanic Province of Cretaceous-Tertiary age, north-eastern boundary with the Bastar Craton, southern boundary with the proterozoic Southern Granulite Terrain (SGT), western boundary with the Western Ghats (WG), eastern boundary with the Proterozoic Eastern Ghats Mobile belts (EGMB). Dharwar craton contains tonalite-trondhjemite-granodiorite (TTG) gneisses (widely called Peninsular Gneisses), greenstone belts and K-rich granites. Dharwar craton is cut by a number of NNW-SSE to N-S trending faults, shear zones and schist belts. There is a change in metamorphic grade from greenschist to amphibolite in the north to granulite in the southern part of the Dharwar craton having E-W isograd lines (Naqvi & Rogers, 1987). TTG is formed due to the melting of mafic crust in the presence of water and other volatiles and last stage differentiation in three phases having ages of 3.4 Ga, 3.3-3.2 Ga and 3.0-2.9 Ga (Meen et al., 1992). Greenstones are the product of volcano-sedimentary sequences that have undergone metamorphism of basalts (Naqvi

& Rogers, 1987). Dharwar Craton is divided into the Western (WDC) and Eastern (EDC) blocks, separated by the Chitradurga Shear Zone (CSZ). The division is based on age and nature of the basement rock and the abundance and nature of greenstone sequences along with the degree of melting and metamorphism (Naqvi & Rogers, 1987).

Western Dharwar Craton (WDC) consists of mainly of 3.4-2.9 Ga TTG gneisses, volcano-sedimentary greenstone sequences (Sargur and Dharwar Supergroup) and K-rich granites (Naqvi & Rogers, 1987; Jayananda et al., 2018). WDC is marked by a number of N-S shear zones like Chitradurga Schist Belt (CSB) or Chitradurga Shear Zone (CSZ), Baba Budhan Shear Zone (BDSZ) and Balehonnur Shear Zone (BHSZ) in a sequence almost parallel from east to west respectively. The Sargur group (3.2-3.0 Ga) is a product of either an accretionary process in an oceanic trench environment or plume setting (Jayananda et al., 2008, 2018). Dharwar Supergroup (named by Foote in 1882) rests unconformably over the Sargur group and consists of Babubudhan and Chitradurga group separated by an unconformity. The metamorphic grade of the WDC increases from greenschist in the north to amphibolite to granulite facies in the south corresponding to the paleo-pressure increase from 3-4 kbar in the north to 9-10 kbar (35 km paleodepth) in the south respectively. Two stages of cratonisation are proposed in WDC: first at around 3.0 Ga due to a thermal event followed by the emplacement of K-rich plutons at around 2.6 Ga .

Eastern Dharwar Craton (EDC) consists of two episodes of magmatism represented by emplacement of TTG gneisses and K-rich granites during 2.75-2.6 Ga and 2.56-2.52 Ga, coeval with calc-alkaline volcanism in the greenstone belts (schist belts). The metamorphism of WDC is of low-T and high-P type, and EDC is of high-T and low-P type. It is due to the abundance of younger granites that seem to have supplied advective heat in the EDC but the younger granites are almost absent in the WDC (Sharma, 2009). The WDC and EDC formed as a coherent block through accretion at around 2.5 Ga. Subsequently, intense mafic dyke swarm events occurred during 2.4-2.1 Ga contemporaneous with similar events happening from different parts of the World, like Finland, Greenland, North America, Antarctica and Australia (French and Heaman, 2010).

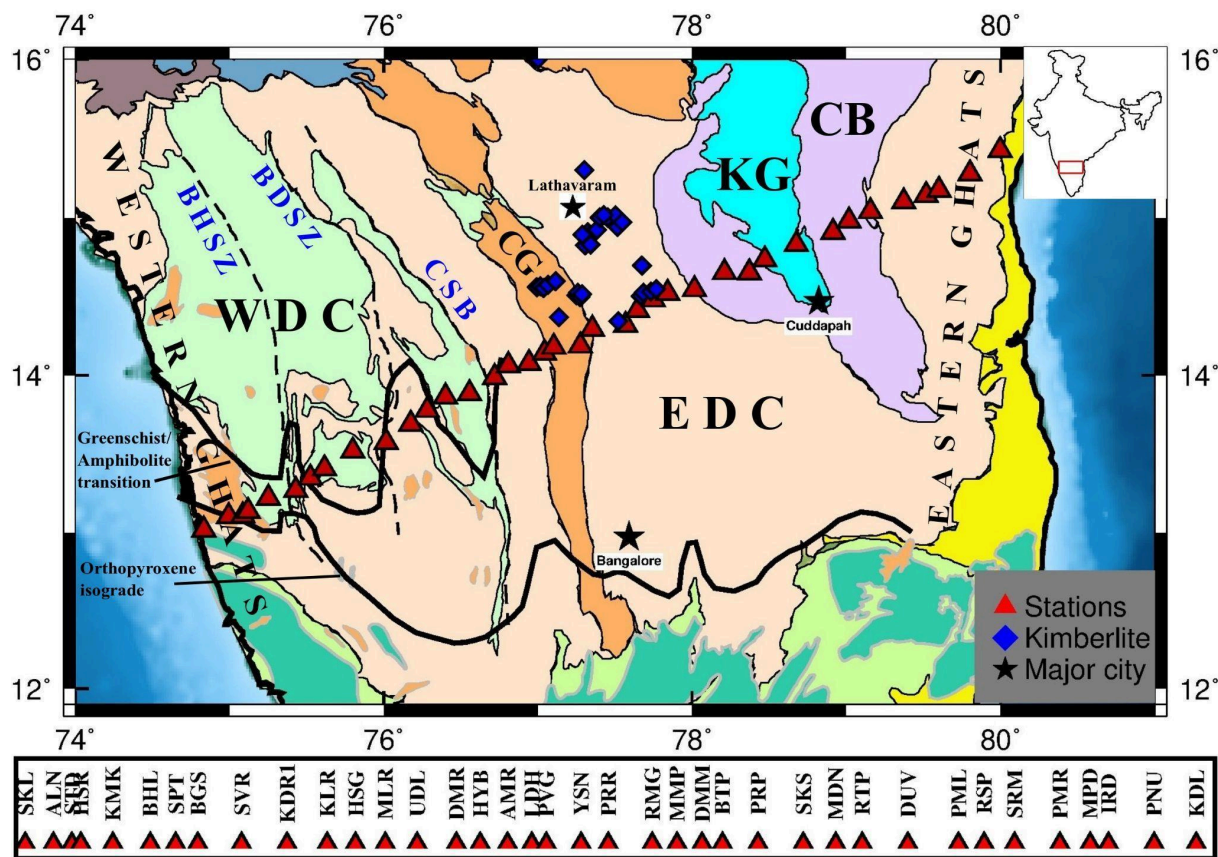


Fig. 1.4. Geological map along with 38 broadband seismometers (shown in red triangle) used in this thesis and approximate location of kimberlite pipes (shown in blue diamond). WDC: Western Dharwar Craton; EDC: Eastern Dharwar Craton; CSB: Chitradurga Schist Belt; CG: Closepet Granite; CB: Cuddapah Basin; KG: Krishna-Godavari Basin.

There are three basins (Cuddapah, Bhima and Kaladgi) formed during 1.8-0.8 Ga in the Dharwar Craton (DC) and thus have more than a billion years of evolutionary history. The mafic dyke swarm activity (magmatic activity) around ~1.8 Ga is related to the formation of Cuddapah basin in the SW part of the EDC (Anand et al., 2003). The Bhima and Kaladgi basins were formed at the NW boundary of the Dharwar Craton (DC). The Eastern Ghats was formed and evolved when the EDC and eastern Antarctica collided during ~1.8 Ga (Saha & Sain, 2019). The kimberlitic volcanism happened around ~1.1 Ga. Dharwar craton as part of India has been affected by Marion and Reunion mantle plume activities at 85 Ma and 65 Ma due to which India separated from Madagascar and Seychelles respectively.

1.4.1 Kimberlites and eastern Dharwar craton

Kimberlites are ultrabasic, poor in SiO_2 , low Al, low Na igneous rocks rich in CO_2 and H_2O which originate from the mantle mainly due to intraplate alkaline magmatism (Pearson et al., 2019). Majority of kimberlitic volcanism (>80%) are younger (200-50 Ma) and are linked with a global period of enhanced short-lived mantle plume activity, continental rifting and formation of ocean basins (Kumar et al., 2007; Pearson et al., 2019). The general common mineral assemblage in kimberlites include olivine, spinel, perovskite, Mg-Ilmenite, phlogopite, apatite and carbonate along with crust and mantle derived xenocrysts. The mantle xenoliths and the minerals present in the kimberlites can provide an estimation of the depth of origin of such magmas. Diamond-bearing kimberlite shows the existence of thick cratonic lithosphere at the time of their eruption (Tharimena et al., 2017; Foley et al., 2019; Pearson et al., 2019). The eastern Dharwar craton (EDC) in India has been subjected to diamond-bearing kimberlite volcanism at 1100 Ma (Ganguly and Bhattacharya, 1987; Kumar et al., 1993) suggesting thick lithospheric roots at the time of the eruption. There are four kimberlite fields located in south India mainly in EDC: (1) Wajrakarur Kimberlite Field (WKF) in Anantapur district, Andhra Pradesh (2) Narayanpet Kimberlite Field (NKF) in Mahabubnagar district, Andhra Pradesh and Gulbarga district, Karnataka (3) Raichur Kimberlite Field (RKF) in Raichur district, Karnataka, and (4) Tungabhadra Kimberlite Field (TKF) in Raichur district, Karnataka and Mahabubnagar District, Andhra Pradesh.

Systematic studies involving petrography, geochemistry, and thermo-barometry on the kimberlite drill core samples from the first discovered WKF cluster show that the xenoliths range in composition from pyroxenite to dunite (Ganguly & Bhattacharya, 1987). Additionally, they comprise garnet-harzburgite, lherzolites, wehrlites, and olivine-clinopyroxene, along with spinel and a considerable presence of phlogopite. The $\text{Mg}/(\text{Mg} + \text{Fe})$ ratio of olivines found in the EDC kimberlite xenoliths falls within a narrow range of 0.91-0.94, decreasing to 0.82 in the kimberlite matrix, indicating a variation in the mantle composition. According to Ganguly and Bhattacharya (1987), there is a positive correlation between $\text{Cr}/(\text{Cr} + \text{Al})$ and $\text{Mg}/(\text{Mg} + \text{Fe})$ in the kimberlite xenoliths, suggesting that mantle fertility progressively increases with depth. Their other noteworthy observations regarding the mantle thermal gradient, derived from the P-T conditions of the coexisting minerals in the xenoliths, indicate (i) ultrabasic and eclogite extrusions from

depths of 70–150 km and 100–185 km, respectively, and (ii) a lithospheric thickness of 185 km during the mid-Proterozoic period at the time of kimberlite eruption, similar to the African cratons (Griffin et al., 2009b). Further validation of these conclusions has come from subsequent studies by Griffin et al. (2009a) and Shaikh et al. (2020). An analysis of the vertical distribution of clinopyroxene and garnet xenocrysts, along with data from some rare mantle xenoliths, identifies two distinct compositional layers in the lithosphere beneath the kimberlite zone: a depleted lherzolite layer at depths of 80–145 km and a re-fertilized and Ti-metasomatized peridotite layer at depths of 160–190 km (Shaikh et al., 2020). The zone between 145 and 160 km depth is marked by a notable scarcity of garnet.

1.5 Seismic studies along the E-W Profile

1.5.1 Results from DSS crustal studies

Deep Seismic Sounding (DSS) research commenced in India in 1972 as part of an Indo-Soviet collaboration agreement. This initiative involved a 600 km long profile oriented roughly from ENE to WSW, situated near 14° N latitude, stretching from Kavali on the eastern coast to Udupi on the western coast of India (Figure 1.4). Figure 1.5 shows the raw reflection data and the interpreted section along the profile. Along this profile, Kaila et al. (1979) and Roy Chowdhury and Hargraves (1981) inferred 17 crustal scale faults having 15 major deep faults and 2 major low angle thrusts. The profile has shown a large number of reflectors from about 2 km to about 50 km depth. In addition, there are 5 faults/thrusts extending to intermediate depths. Most of them offset the Moho upto ~7 km in the Dharwar craton. These faults are interpreted based on reflections and refractions from shallow, intermediate and deep boundaries, and diffractions. Later, Gibbs (1986) questioned the validity of such a large number of steeply dipping faults. Kaila and Krishna (1992) utilized seismic reflection/refraction data obtained from the Kavali-Udupi profile spanning the WDC/EDC. Their model depicted an upper crust characterized by a P-wave velocity of 6.4 km/s, a lower crust velocity of 6.7 km/s, and a variable Moho depth ranging from 34 km in the east to 41 km in the west (Kaila & Krishna, 1992).

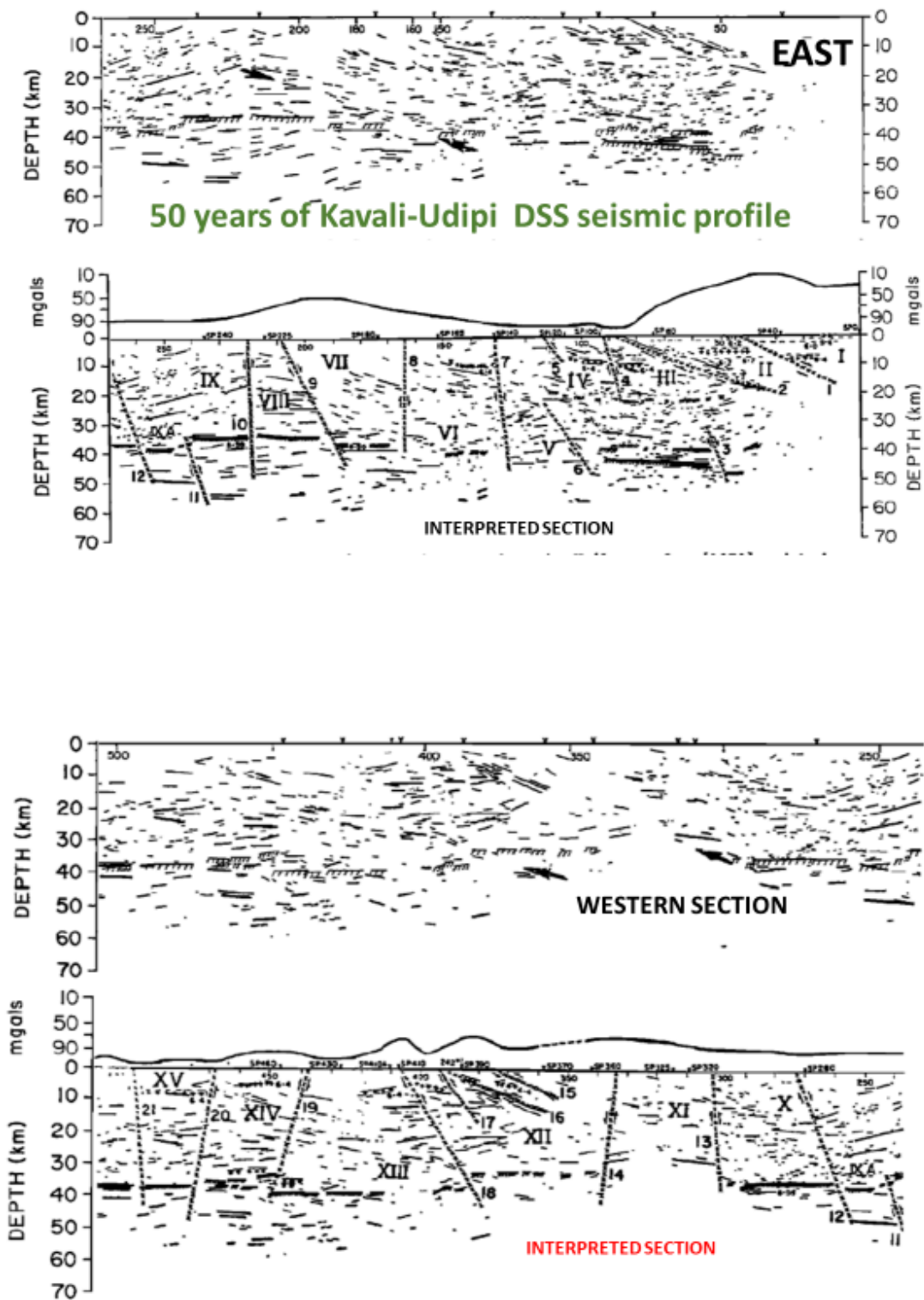


Fig. 1.5. The raw reflection data of the eastern and western sections and their interpreted sections along the Kavali-Udupi DSS seismic profile are shown in the figure above (Kaila et al., 1979).

Mall et al. (2008) investigated the eastern section of the profile covering the Cuddapah basin. They have estimated 15-20 km thick magmatic underplating and 40-44 km thick crust beneath southwestern part of the Cuddapah Basin and 32-36 km thick crust beneath EDC. On the western part of profile, Mall et al., 2012 has investigated the P-velocity from Udipi to Closepet Granite (CG) down to 60 km depth and found high velocity magmatic layer (7.30 km/s) at the base of the crust beneath WDC due to cooling and crystallization of magma ponded in the lower crust and crust gets thicker (~42-45 km) and an unusually low velocity layer (LVL) (7.40 km/s) just below the Moho at a depth range of 48–60 km in the uppermost mantle and interpreted this LVL as chemical alteration caused by mantle metasomatism.

Mandal et al. (2018) proposed that the western Dharwar Craton (WDC) features a relatively simple crust with a horizontal Moho, while the eastern Dharwar Craton (EDC) exhibits a more complex, dipping Moho that reflects a collision history, a thrust fault, and a crustal root. These diverse Moho characteristics indicate that the Dharwar Craton is not formed from a single process but a two-stage subduction-accretion process involved in crustal development during the Neoproterozoic era, accompanied by crust-mantle detachment. The dipping seismic reflections originating from the lower crust and extending into the mantle can be interpreted as a 2.5 Ga old suture that is linked to mantle subduction, implying that plate tectonics were active during the Neoproterozoic period over the Indian shield.

1.5.2 Broadband Seismology studies:

During 2012-14, a broadband seismic experiment coinciding with the DSS profile was performed using 38 seismic stations at ~15-20 intervals. Their detailed location is provided in Table 1.1. These stations were located densely in South India in an almost linear profile covering Konkan plain, Western Ghats (WG), Western Dharwar Craton (WDC), Chitradurga Schist Belts (CSB), Closepet Granite (CG), Eastern Dharwar Craton (EDC), Cuddapah Basin (CB) and Eastern Ghats (EG). The distribution of these 38 stations is shown in figure 1.4. The continuous waveforms were recorded at 50 Hz samples by seismometers that include the REFTEK 130 data acquisition system and Guralp CMG 3T/3ESP broadband sensors. Saikia et al. (2016) performed inversion of receiver function, modeling through the different approaches like H-Vp/Vs stacking, common conversion point (CCP) stacking and found Moho offset by upto ~8 km suggesting

differential uplift of Western Ghats (WG) as a consequence of India–Madagascar separation, variation of Moho depth between 34–41 km. Along the same profile, Saikia et al. (2017) has found ~40 km and ~35 km thick crust beneath WDC, EDC respectively, the average thickness of the upper, middle and lower crust is ~ 4 km, 12 ± 4 km and 24 ± 4 km respectively. A high velocity (HVL, $V_s > 4$ km/s) above the Moho is 3 ± 1 km and 11 ± 3 km below WDC, EDC, and CB respectively, while an unusually thick section of 18 ± 2 km can be found beneath the western coast of India and the surrounding region, representing underplated material that formed from the India-Madagascar rifting event that occurred around 85 Ma.

1.6 Lithospheric studies of Dharwar Craton

The upper mantle structure of Dharwar Craton (DC) has been explored by a variety of geophysical tools, such as seismological studies which consists of surface waves and receiver functions (Mitra et al., 2006; Kiselev et al., 2008; Bodin et al., 2014; Borah et al., 2014; Maurya et al., 2016; Saha et al., 2020; Mullick et al., 2022), magnetotelluric (MT) experiments (Abdul Azeez et al., 2015; Pratap et al., 2018; Malleswari et al., 2019), gravity data (Kumar et al., 2013; Vasanthi & Santosh, 2022) and heat flow observations (Negi et al., 1986; Roy & Rao, 2000; Roy & Mareschal, 2011). We discuss here the results based on the above studies.

The Lithosphere-Asthenosphere Boundary (LAB) is estimated at a depth of 160–200 km beneath the WDC, 120–185 km beneath the EDC based on the surface wave studies (Mitra et al., 2006; Borah et al., 2014; Maurya et al., 2016; Saha et al., 2020; Krueger et al., 2021). There is a large variation of lithospheric thickness among different methods due to their uneven sampling, resolution and sensitivity to composition, temperature, anisotropy etc. Also there is a significant variation of lithospheric thickness by the surface wave methods depending on the definition of the low velocity value below the high velocity layers (Rychert et al., 2020). Based on the negative gradient of S-receiver functions at about 10s, Kumar et al. (2007) and Mukherjee et al. (2022) suggested a ~100 km thin lithosphere beneath the Dharwar craton, similar to the significantly eroded North China, Wyoming, Yangtze, Omolon cratons (Pearson et al., 2021). However, no formal velocity inversion of the receiver function data was performed. This was supported by the earlier few heat flow data (Negi et al., 1986). Later the negative phase of the S-receiver function was termed as mid-lithospheric discontinuity instead (Bodin et al., 2014;

Krueger et al., 2021). The thin lithosphere model is seemingly incompatible with i) the presence of a cold mantle as seen in the low surface heat flux of 30 mWm^{-2} in the western to 40 mWm^{-2} in the eastern part of the EDC (Roy & Rao, 2000), ii) almost flat topography of 500-600 m over the EDC and iii) evidence for the positive converted phase in the receiver function due to the velocity increase at a depth of $\sim 90\text{-}100 \text{ km}$ (Jagadeesh & Rai, 2008; Kiselev et al., 2008; Mandal et al., 2012; Chaudhury et al., 2021).

The 3-D shear wave velocity model beneath the Dharwar Craton, derived from fundamental mode Rayleigh wave phase velocities within the 20-90 s period range, indicates a presence of anomalously high velocities at depths of 50-100 km (Borah et al., 2014). A notable high shear wave velocity, reaching up to 4.9 km/s, is found in the upper mantle beneath the northwestern region of the Dharwar Craton, which decreases towards the southern and eastern parts of the EDC. At depths below 100 km, the shear velocity under the Dharwar Craton is close to the global average. A more recent investigation (Saha et al., 2020) points to significant differences in lithospheric layering among the three Precambrian regions: WDC, EDC, and SGT, with the uppermost mantle in the EDC exhibiting an unusually high velocity of 4.7–4.8 km/s at depths of 70–120 km, which then decreases to 4.4–4.5 km/s at a depth of 200 km. Inversion of the Rayleigh wave group velocity dispersion map 10-100 s created at 1° grid intervals from ambient noise and earthquakes data was performed by Saha et al. (2020). Their study shows at least 160 km thick lithosphere beneath the EDC with a $V_s > 4.7 \text{ km/s}$ layer at a depth of 80-130 km. Subsequently, an improved velocity model of the Dharwar craton and the adjoining regions of South India (Mullick et al., 2022) created using the Rayleigh wave phase velocity dispersion data (30-140 s) with a better horizontal resolution of 0.5° shows lithospheric thickness variability from 150-200 km beneath EDC. Mitra et al. (2006) estimated a $\sim 155 \text{ km}$ thick seismic lithosphere estimated by inter-station Rayleigh wave phase velocity inversion.

Variability in lithospheric thickness estimated by different geophysical tools: 150-180 km using joint inversion of P and S-receiver function (Kiselev et al., 2008), 150-200 km beneath the station HYB in Dharwar craton through joint inversion of receiver function and global surface wave data (Bodin et al., 2014), $\sim 170 \text{ km}$ based on surface wave velocity inversion at a resolution of 1° from the global studies (Pasyanos et al., 2014) and 155-180 km from the S_p receiver function analyzed at station HYB, located in Dharwar craton (Krueger et al., 2021). Low

resolution $\sim 2^{\circ}$ - 3° surface wave data (McKenzie et al. 2015; Maurya et al., 2016, Priestley et al., 2018) and satellite gravity data (Vasanthi & Santosh, 2022) indicate ~ 130 km thick lithosphere but Kumar et al. (2013) computed 165-180 km thick lithosphere beneath the EDC by the integrated modeling of gravity, geoid and heat flow.

MT studies suggest a thick lithosphere of 180-200 km beneath the Dharwar craton (Abdul Azeez et al., 2015; Pratap et al., 2018; Malleswari et al., 2019). The surface heat flux varies from over Dharwar craton in the range 25 - 40 mWm^{-2} having 25 - 32 mWm^{-2} in the WDC and ~ 40 mWm^{-2} in the EDC (Roy & Rao, 2000; Roy & Mareschal, 2011) and the corresponding mantle heat flow contribution in the Dharwar craton is 11 - 16 mWm^{-2} (Ray et al., 2003). Seismic anisotropy along the E-W profile has been inferred by analyzing the core-refracted shear waves, such as SK(K)S and PK(K)S (Saikia et al., 2019). The delay time (δt), dependent on the thickness and strength of anisotropic layers, varies from 1.1 s in the western edge of EDC to 0.85 s for stations sitting on the kimberlite domain in the EDC. Assuming a 3% anisotropy in the lithosphere with an average V_s of 4.6 km/s, the observed time delay would correspond to an anisotropic layer of thickness 130 - 160 km. Malleswari et al. (2019) conducted a magnetotelluric study of the lithospheric structure beneath the Dharwar Craton and found a nearly vertical conductive anomaly in the upper mantle beneath the Chitradurga Shear Zone (CSZ), which indicates the Archean suture between the western and eastern segments of the Dharwar Craton. Although the craton is preserved having ~ 200 km thick lithosphere (highly resistive), a significant part of the lithosphere below the western Dharwar is modified due to plume-continental lithosphere interactions during the Cretaceous-Tertiary period. Thin (~ 120 km) lithosphere is estimated beneath the EGMB having a moderately conductive feature interpreted as the remnant of the Proterozoic collision process between the Indian landmass and East Antarctica. Arjun et al. (2022) has done the joint modeling of the teleseismic travel-time residuals, Bouguer gravity, and surface topography across the Western Ghats (WG) and Dharwar craton and found the WDC is (~ 20 - 30 km) thicker than under the EDC and crustal thickness of the WDC is thicker than the EDC.

Table 1.1 - Operational Details of Broadband Seismometers located in South India.

S.No.	Station Code	Latitude (°N)	Longitude (°E)	Elevation (m)	Operation Period (year/month)
01.	SKL	13.016	74.833	40	2012/04-2013/10
02.	ALN	13.102	74.995	98	2012/04-2013/12
03.	STD	13.107	75.086	94	2012/11-2013/12
04.	HSR	13.134	75.120	100	2013/02-2013/12
05.	KMK	13.219	75.252	841	2012/04-2013/10
06.	BHL	13.268	75.430	805	2012/04-2013/12
07.	SPT	13.347	75.525	750	2013/02-2013/12
08.	BGS	13.407	75.616	996	2012/04-2013/10
09.	SVR	13.519	75.799	1260	2012/07-2013/12
10.	KDR	13.574	76.016	774	2012/04-2013/12
11.	KLR	13.690	76.175	706	2012/04-2013/08
12.	HSG	13.778	76.282	723	2013/02-2013/12
13.	MLR	13.864	76.400	668	2012/04-2013/12
14.	UDL	13.884	76.556	648	2012/04-2013/10
15.	DMR	13.987	76.719	589	2012/04-2013/12
16.	HYB	14.063	76.808	608	2013/02-2013/09
17.	AMR	14.077	76.942	605	2012/04-2013/10

18.	LDH	14.142	77.042	620	2013/02-2013/11
19.	PVG	14.180	77.102	694	2012/04-2013/09

20.	YSN	14.187	77.275	601	2013/02-2013/10
21.	PRR	14.294	77.352	570	2012/04-2014/01
22.	RMG	14.318	77.568	453	2012/04-2014/01
23.	MMP	14.414	77.645	390	2013/02-2014/01
24.	DMM	14.479	77.748	350	2012/07-2014/01
25.	BTP	14.522	77.840	321	2013/02-2014/01
26.	PRP	14.544	78.014	343	2012/09-2014/01
27.	SKS	14.653	78.209	245	2012/09-2014/01
28.	MDN	14.652	78.370	217	2013/02-2014/01
29.	RTP	14.734	78.471	171	2012/09-2014/01
30.	DUV	14.830	78.675	185	2012/09-2014/01
31.	PML	14.908	78.913	167	2012/10-2014/01
32.	RSP	14.980	79.016	105	2013/02-2014/01
33.	SRM	15.038	79.155	222	2012/10-2014/01
34.	PMR	15.100	79.372	123	2012/10-2014/01
35.	MPD	15.141	79.516	105	2013/02-2014/01
36.	IRD	15.174	79.599	108	2012/10-2014/01
37.	PNU	15.277	79.801	28	2012/09-2014/01

38.	KDL	15.425	79.996	16	2013/10-2014/01
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1.7 Thesis objectives and plan

In this thesis, we have attempted to answer the questions related to structure, composition and evolution of the Earth's crust and lithosphere and its variability among geological domains of different ages from Archean to Proterozoic along the E-W profile using shear wave velocity (Vs) derived from modelling receiver function and surface wave dispersion. The E-W profile crosses through the mid-late Archean Dharwar craton hosting the 1.1Ga old diamondiferous kimberlite fields and the Proterozoic Cuddapah basin. The west and east ends of the profiles separated from the Madagascar and the Eastern Gondwanaland respectively under the influence of mantle plumes. Some of the key issues being answered in this thesis include the nature of the Archean cratons – their preservation and deformation comparing their lithospheric structure with those with the kimberlite domains, difference between Archean and Proterozoic terrains and influence of mantle plumes.

Despite a number of studies along this profile most of the issues discussed above remain unanswered or poorly investigated in terms of spatial and vertical resolution of the velocity structures. To create a quality velocity image we generated improved surface wave dispersion (SWD) data at shorter periods and combined them with earlier longer period data. The SWD data were jointly inverted with similar resolution receiver function data.

Chapter 1 discusses the global framework of crust and lithosphere-asthenosphere system, geological framework of the region, details of data used and the key scientific questions. This is followed by **Chapter 2** dealing with methodology, analysis, modelling/ inversion approach and results from joint inversion. These results are discussed in terms of crust structure and composition, and lithosphere-asthenosphere structure along the profile in **Chapter 3**. Finally, **Chapter 4** summarises results and proposes newer improved modelling approaches to be followed to model data along the profile.

Chapter 2 - Velocity modelling and imaging

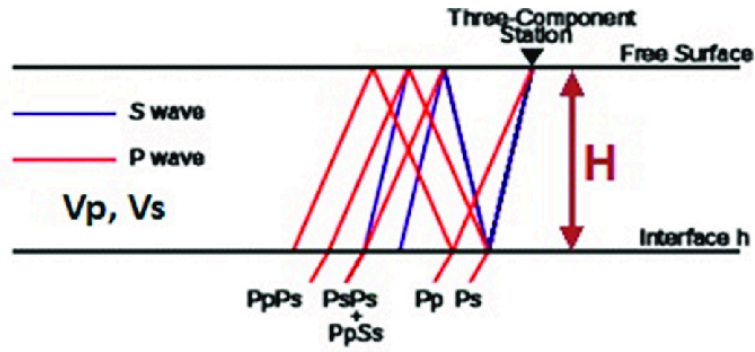
2.0 Data and Methodology

In this chapter, we use two approaches to map the velocity structure. First, a joint inversion of receiver functions with the Rayleigh wave phase velocity dispersion data (3-140 s) to compute velocity variation with depth. The phase velocity dispersion at shorter periods (3-30 s) is computed from the ambient noise tomography at ~ 10 km interval, while the longer period (30-140 s) dispersion at ~ 55 km node is extracted from an earlier work using the earthquake tomography (Mullick et al., 2022). The receiver function dataset utilized in this study was computed from teleseismic waveforms recorded at 38 broadband seismic stations between January, 2012 and April, 2014. These waveforms recorded ground vibration at a sampling rate of 50 Hz using seismographs equipped with the REFTEK data acquisition system and Guralp CMG 3T/3ESP broadband sensors. Details about the station network can be found in Chapter 1 (see Figure 1.4 and Table 1.1 and 2.1). Theoretically, Receiver function is an impulse response of a layered earth for a near vertical incident P or S wave (Langston, 1979). The method is used to compute the crust and upper-mantle structures from teleseismic waveforms recorded at three-component seismic stations (Langston, 1979; Ammon et al., 1990; Ammon, 1991). Radial and transverse receiver functions are generated by deconvolution of vertical components from the radial and transverse components of the seismogram, which helps in isolating the site effects from other information carried in a teleseismic P wave. The observed dispersion data allows for imaging the Earth to a depth of 300 km, with maximum sensitivity for the maximum period extending down to approximately ~ 220 km (see figure 2.21). The approach followed assumes a horizontally layered structure below the recording station. Secondly, to account for spatial sampling of the receiver function time series, we follow the Common Conversion Point (CCP) approach that helps in 2-D or 3-D imaging of the earth structure (Dueker and Sheehan, 1998). The CCP method does not provide any estimate of the velocity of the layer. Details of the receiver function and surface wave dispersion methodology are given in the subsequent sections.

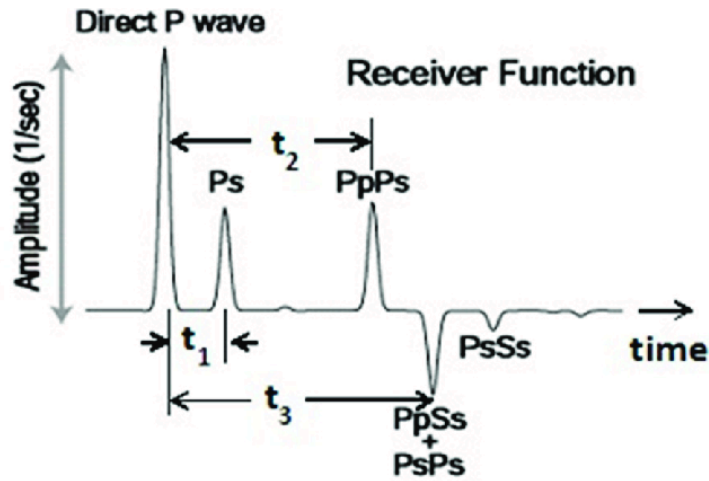
2.1 Receiver function -

Receiver function (RF) is a method to image the earth structure beneath the seismic station or receiver. It is the time series computed from a 3-component seismogram which provides the relative response of the earth structure below the receiver as shown in Fig. 2.1 (Langston, 1979; Ammon, 1990 & 1991). When the teleseismic P wave and its coda travels and interacts with the discontinuity, a part of the energy remains as P wave and a part of the energy converts to S wave which is called converted Ps wave which is followed by reverberations (PpPs, PsPs+PpSs, PsSs). The time delay and relative amplitude between direct P wave and converted Ps wave provides the information of the depth and magnitude (sharp or gradational) of the discontinuity respectively (Rondenay, 2009). The amplitude of the direct P wave is sensitive to near surface structure, converted Ps wave amplitude is sensitive to relative S velocity contrast, and the reverberations or multiples are sensitive to both relative S velocity and density contrasts (Julia, 2007). The teleseismic waves recorded contain the source, propagation effects along with the converted waves and their reverberations or multiples. To extract or isolate the converted waves and their reverberations or multiples from the source and propagation effects, deconvolution of radial or transverse components by the vertical component is done and the resulting time series, called receiver function, is approximately the impulse response of the Earth structure beneath the receiver. So it is a composite of direct P wave, converted Ps wave and reverberations or multiples (PpPs, PsPs+PpSs, PsSs) (Ammon, 1990 & 1991).

Teleseismic waves are used instead of local/ regional earthquake waveforms for the following reasons: **(a)** The incoming P wave and its coda are well separated from the other primary incident waves as the time delay between the phases increases with distance (Rondenay, 2009). **(b)** the polarization direction and ray parameter are almost constant since the teleseismic wavefronts are approximated to be the planar (Rondenay, 2009), **(c)** The epicentral distance range of the teleseismic waves taken for receiver function computation is $30^\circ - 95^\circ$ which corresponds to incident angle of P waves of $20^\circ \pm 10^\circ$ at Moho and $10^\circ \pm 5^\circ$ at the surface which is nearly vertical (Edme, 2007). So the different phases are well separated in radial (converted Ps wave) and vertical components (direct P wave). The epicentral distance range of 30° - 95° is used to avoid triplication from MTZ and complication from Core-Mantle boundary.



(a)



(b)

Fig. 2.1. (a) Ray diagram of different phases for an incoming P wave, (b) Radial receiver function time series.

2.2 Rotation of the components of seismogram -

For accurate representation and estimation of teleseismic energy recorded in the seismic station, we rotate the three orthogonal (N-S, E-W and vertical) components to the radial component (motion towards or away the direction of the source of seismic wave), transverse component (perpendicular to the radial component) and vertical remains the same (Rondenay, 2009).

$$\begin{bmatrix} R \\ T \\ Z \end{bmatrix} = \begin{bmatrix} -\cos \theta & -\sin \theta & 0 \\ \sin \theta & -\cos \theta & 0 \\ 0 & 0 & 1 \end{bmatrix} \begin{bmatrix} N \\ E \\ Z \end{bmatrix}$$

where E, N and Z are East-West, North-South, and vertical components respectively; R and T components are radial and transverse components; and θ is back azimuth of the incident wave. The rotated radial and transverse components are used to compute receiver functions which will be discussed in the next section.

2.3 Mathematical derivation of Receiver function -

In time domain, the three component seismogram recorded in the form of response D in the seismic station or receiver due to the teleseismic waveform is given by (Langston, 1979)

$$D_V(t) = I(t) * S(t) * E_V(t)$$

$$D_R(t) = I(t) * S(t) * E_R(t) \quad (1)$$

$$D_T(t) = I(t) * S(t) * E_T(t)$$

The response D recorded in the receiver is the convolution of effective source time function S(t) of the incident wave, instrument response I(t) and impulse response of the local earth structure E(t). The subscripts V, R, T denote vertical, radial and transverse components respectively.

Transforming equation (1) into the frequency domain,

$$D_V(\omega) = I(\omega) S(\omega) E_V(\omega) \quad (a)$$

$$D_R(\omega) = I(\omega) S(\omega) E_R(\omega) \quad (b) \quad (2)$$

$$D_T(\omega) = I(\omega) S(\omega) E_T(\omega) \quad (c)$$

Dividing 2(b) and 2(c) by 2(a), we get -

$$\begin{aligned} S_R(\omega) &= \frac{E_R(\omega)}{E_V(\omega)} = \frac{D_R(\omega)}{D_V(\omega)} \\ S_T(\omega) &= \frac{E_T(\omega)}{E_V(\omega)} = \frac{D_T(\omega)}{D_V(\omega)} \end{aligned} \quad (3)$$

where $S_R(\omega)$ and $S_T(\omega)$ are Radial and Transverse receiver functions respectively. In the time domain, $S_R(t)$ and $S_T(t)$ are obtained by deconvolving the radial and transverse components by the vertical component.

If $E_V(t) \approx \delta(t)$, then

$$S_R(\omega) \approx E_R(\omega) \text{ and } S_T(\omega) \approx E_T(\omega) \quad (4)$$

It means, when the vertical component is Dirac delta function then the radial or transverse receiver function becomes equivalent to the radial or transverse component of the seismogram (Langston, 1979). This deconvolution in equation (3) becomes unstable when there is random noise in the data and the denominator of equation (3) can become zero.

To avoid this problem, a parameter called water level c is introduced and the equation (3) can be rewritten as

$$S_R(\omega) = \frac{D_R(\omega) D_V^*(\omega)}{\Phi(\omega)} \quad (5)$$

where $D_V^*(\omega)$ is the complex conjugate of $D_V(\omega)$ and $\Phi(\omega) = \max \{[D_V(\omega) D_V^*(\omega), c \max [D_V(\omega) D_V^*(\omega)]]\}$ and c is water level parameter. The transform of Gaussian ($G(\omega)$) is multiplied to exclude the high frequency signals.

$$S_R(\omega) = \frac{D_R(\omega) D_V(\omega)}{\Phi(\omega)} G(\omega) \quad (6)$$

where $G(\omega) = A e^{-\omega^2/4a^2}$ and 'a' controls the width of the Gaussian pulse. Parameter A normalizes the Gaussian function to unit amplitude in the time domain.

The water level parameter c is a factor representing trade off amplitude resolution with arrival time resolution. If $c = 0$, the deconvolution provides the best arrival time resolution and if $c = 1$, the deconvolution provides the least square estimate of arrival amplitude resolution. In case of receiver function, the relative amplitude of phases in the seismogram with better possible resolution of arrival time is needed, so parameter c should be as small as possible.

2.4 Iterative deconvolution technique -

Iterative deconvolution is a time domain technique, originally developed to estimate large earthquake source time functions (Kikuchi & Kanamori, 1982) but later being adopted for the receiver function computation (Ligorria & Ammon, 1999). The receiver functions can be interpreted as a linear combination of Dirac delta function $\delta(t)$, shifted in time, each of them related to a different ray reverberating through the layers beneath the recording station.

Mathematically, receiver function signal can be written as:

$$RF(t) = \sum_{i=1}^n a_i \delta(t-t_i) \quad (7)$$

The amplitude of the Dirac delta function at time t_i , represented as a_i , is acquired during the i^{th} iteration. The iterative deconvolution method operates through a least square minimization process, aiming to reduce the variance between the observed horizontal seismogram (specifically, the radial component in the receiver function) and the predicted signal which is generated by convolving an iteratively updated spike train with the corresponding vertical component seismogram. This approach relies on utilizing the cross-correlation function to estimate both the lag and amplitude of the individual spikes constituting the final receiver function.

There are following steps being done at each iteration i for computing iterative deconvolution:

- (i) The cross-correlation of the radial component ($R^{(i-1)}$) with the vertical component (V) is analyzed, considering that $R^{(0)}$ represents the original radial seismogram.
- (ii) Locating the time t_i corresponding to the maximum amplitude a_i' of the cross-correlogram.
- (iii) Determine the scalar coefficient a_i that optimally fits the time-shifted vertical component with $R^{(i-1)}$ by minimizing the following quantity:

$$C = |R^{(i)} - a_i V * \delta(t - t_i)|^2 \quad (8)$$

where $*$ represents the convolution operator and a_i is defined as: $a_i' / (V.V)$, with the dot indicating the zero time cross-correlation.

- (iv) Incorporate a Dirac delta function with amplitude a_i at time t_i into the current receiver function.
- (v) Compute the residual $R(i)$ from $R(i-1)$ using the formula:

$$R^{(i)} = R^{(i-1)} - a_i V * a_i V * \delta(t - t_i) \quad (9)$$

- (vi) Return to the next iteration. This process concludes when the misfit between the receiver function convolved with the vertical component and the radial component, along with added spikes, becomes negligible, or when a significant portion of the radial component (R) has been diminished.

2.5 Computation of receiver function -

- (i) For RF computation, earthquake waveforms having magnitude ($M_w \geq 5.5$) and epicentral distance range 30° - 95° are selected. The epicentral distance range of 30° - 95° is used to avoid triplication from Mantle Transition Zone (MTZ) and complication from Core-Mantle boundary. Events are checked visually and only events having a good signal-to-noise ratio (SNR) are

selected for further processing. The teleseismic earthquake waveform is shown in the figure 2.2 below.

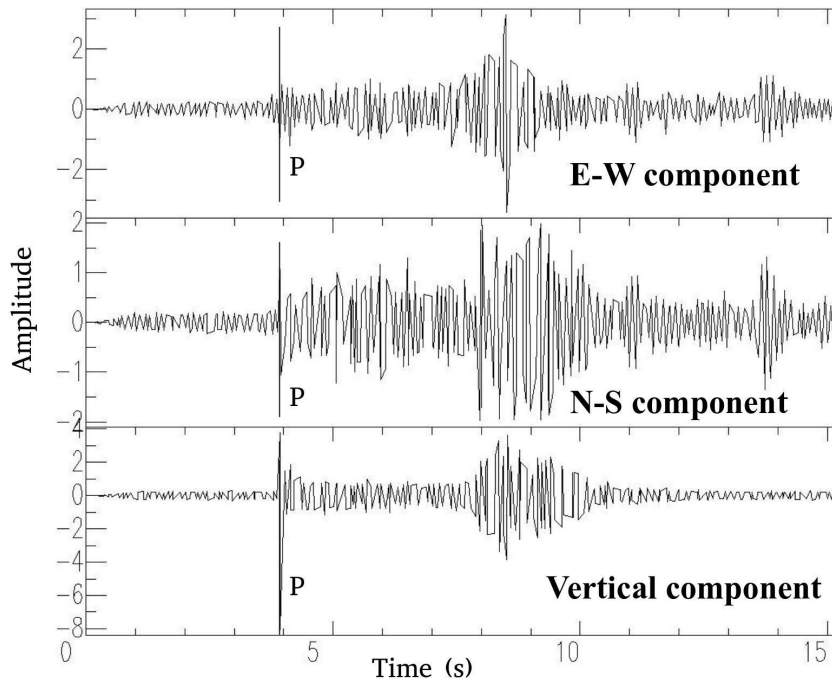


Fig. 2.2. Earthquake waveforms having magnitude ($M_w \geq 5.5$) and epicentral distance range 30° - 95° selected for RF computation.

(ii) The selected seismograms are then filtered using Butterworth high pass filter with a corner frequency of 0.02 Hz with the number of poles and passes set to 2. The P wave arrivals on the seismogram are marked with respect to the IASP91 velocity model (Kennett & Engdahl, 1991) using Taup-Tool-kit (Crotwell et al., 1999). The data traces are then cut to a length of 150 sec (30 sec prior and 120 sec post P wave arrival) and the mean and trend are removed from the seismograms.

(iii) Horizontal components (N-S and E-W) are rotated to great circle arc (gcarc) to obtain radial (perpendicular to vertical and directed towards the source and records SV energy) and transverse component (perpendicular to both vertical and radial component and records SH energy). The rotated waveform is shown in the figure 2.3 below.

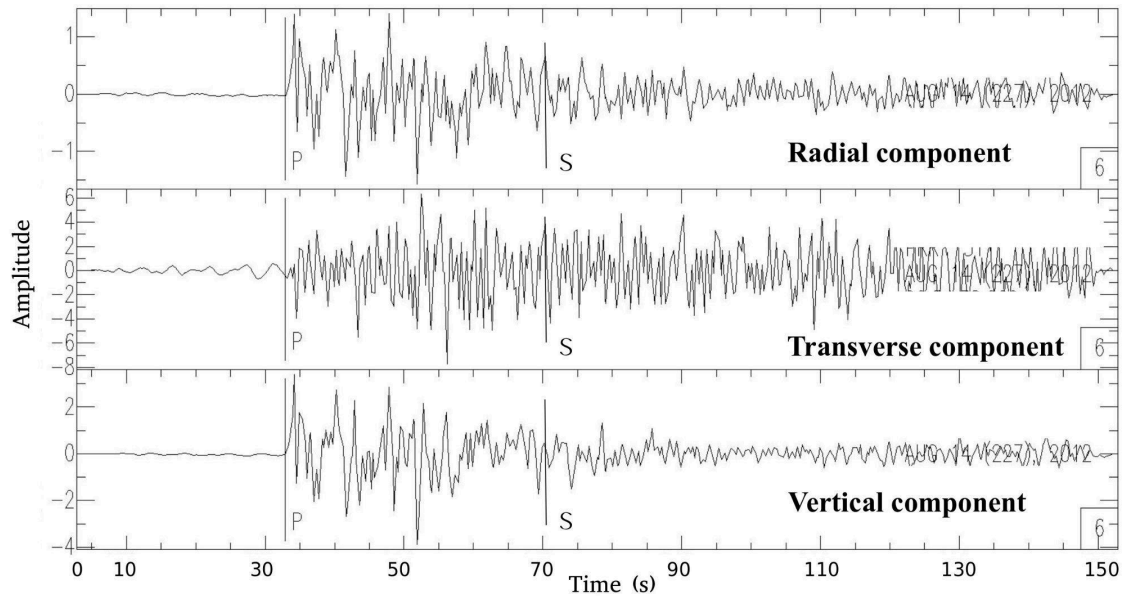


Fig. 2.3. The above horizontal components (N-S and E-W) are rotated to the great circle arc to obtain radial and transverse components. Vertical component remains the same.

(iv) Receiver functions are computed using iterative deconvolution (Ligorria & Ammon, 1999) of the radial and transverse component with the vertical component and a low pass filter called Gaussian width (Gw) (2 times of cut-off frequency) is multiplied to filter out the high frequency content. The computed RF is shown in the figure 2.4.

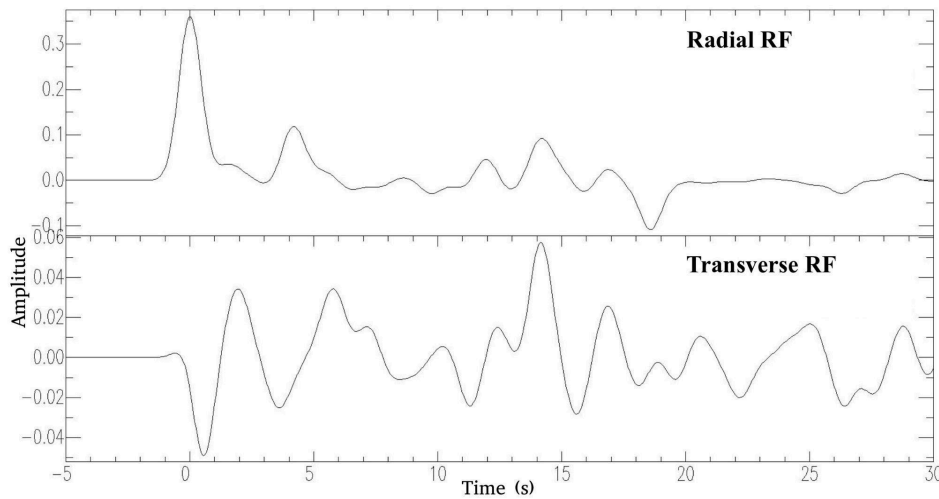


Fig. 2.4. The computed radial and transverse receiver functions (RFs) from the deconvolution of above radial and transverse component waveforms respectively with the vertical component.

(v) In the final step, the computed RFs having fit $< 80\%$ of the original waveform are excluded and selected RFs were visually checked to get very good quality RF waveforms for further analysis and synthetic tests. Fig. 2.5 shows the representative station BTP, recorded 143 events of magnitude ($M_w \geq 5.5$) and epicentral distance range 30° - 95° out of which 40 events were finally selected having acceptable RFs. The final selected RFs for the representative stations BTP and their variation with the epicentral distance and azimuth are presented in (Fig. 2.6).

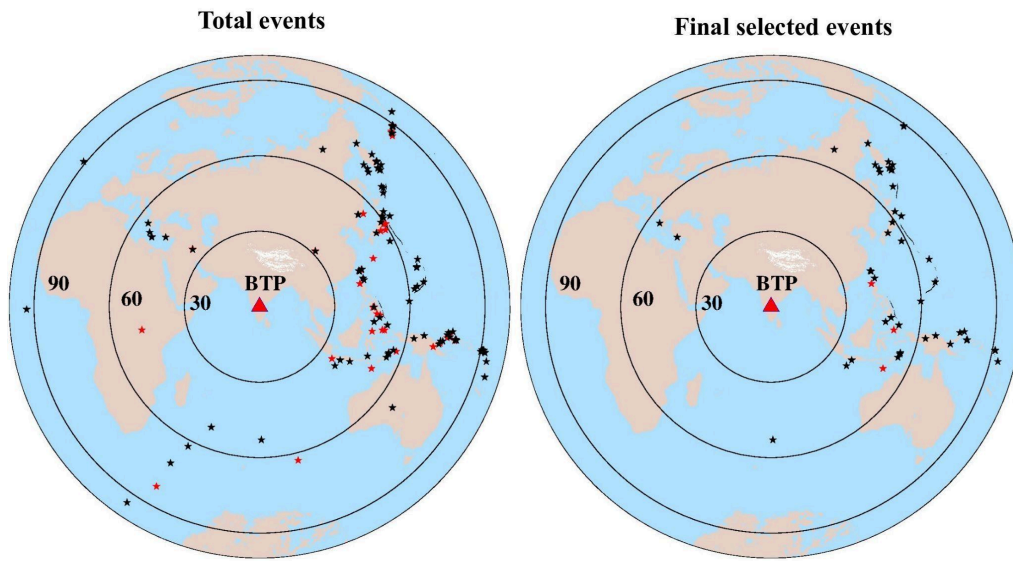


Fig. 2.5. (Left) Global distribution of earthquake waveforms having magnitude ($M_w \geq 5.5$) recorded at station BTP and **(Right)** finally selected events used for analysis and modeling of the receiver function.

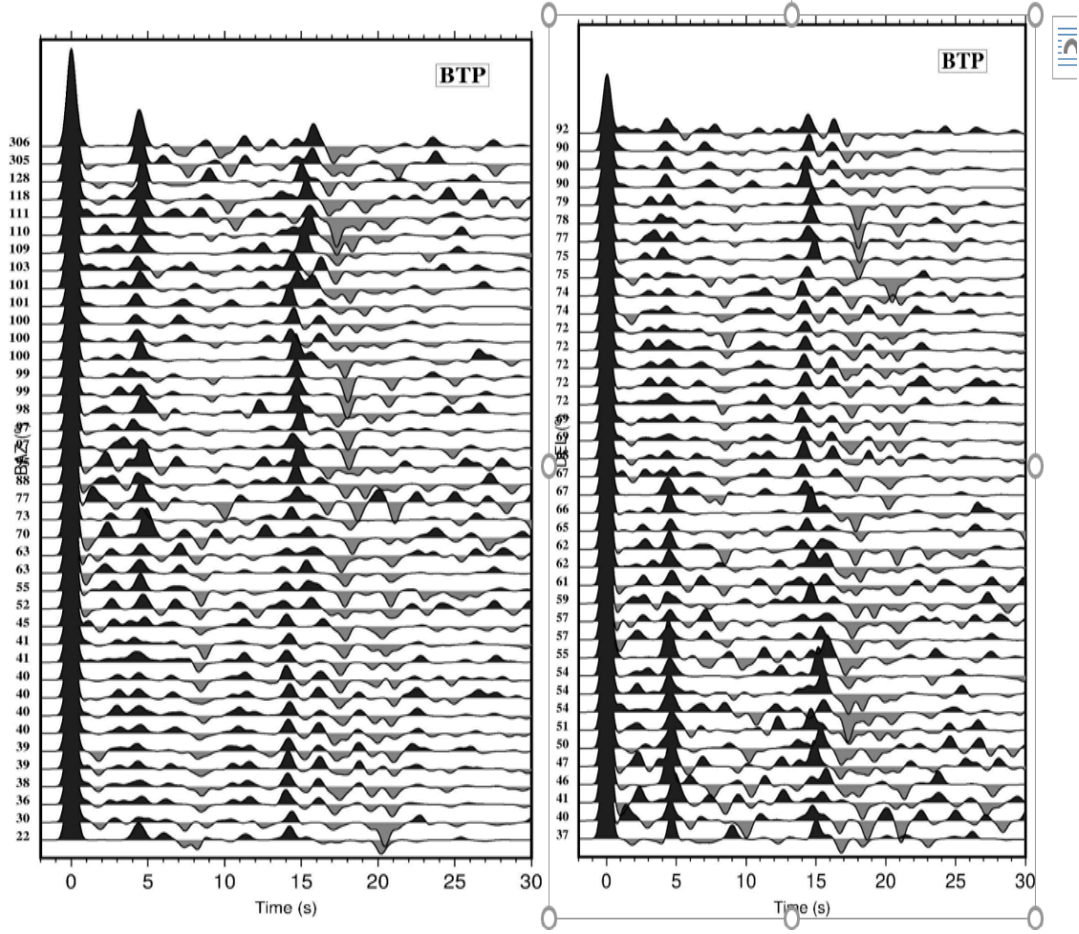


Fig. 2.6. Azimuthal and epicentral variation of the receiver function at a representative station BTP.

To enhance signal-to-noise ratio **RFs** were stacked in the narrow epicentral distance-back azimuth bins ($\pm 10^\circ$). The choice of narrow distance and azimuth bin avoids any significant Ps delay in the bins and any inhomogeneities below the stations. If N no. of RFs are stacked, then SNR increases by a factor of $\sqrt{N}$ (Owens, 1984). These stacked RFs are used with surface wave dispersion for joint inversion. Representative RF and their stacks for BTP in different bins are shown in figure 2.7. For all other stations, it is shown in Appendix figure S2.1.

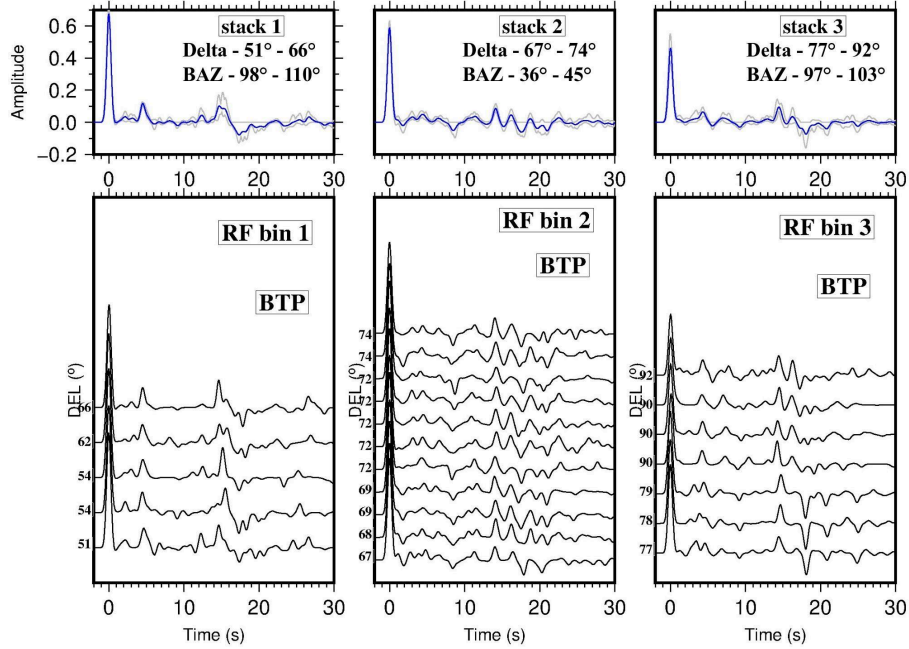


Fig. 2.7. individual RFs of station BTP in three distinct epicentral distance and azimuth ranges and their corresponding stacks.

2.5.1 Resolution capability of receiver function -

We can estimate the vertical resolution or the minimum thickness of the layer ($\lambda/4$) that RFs can resolve for the given Gw. For example, if Gaussian filter width (Gw) is 2.5 which corresponds to the cut-off frequency of ~ 1.2 Hz and an average crustal S-wave velocity (V_s) is 3.7 km/s, the corresponding wavelength (λ) is ~ 3.3 km (using $v = f\lambda$) and since the vertical resolution is $\lambda/4$ (Sheriff & Geldart, 1995), so the layer thickness > 0.8 km is resolvable.

The horizontal resolution of the receiver function depends on the radius of the first Fresnel zone of the incident P wave (Sheriff & Geldart, 1995). The radius of the Fresnel zone is defined

by $R = (H + \lambda/2)^2 - H^2$ and is the quantity to represent the region sampled by the energy of wavelength λ and the interface at a depth H .

Table 2.1. The no. of RFs for each station with the station locations along the E-W Profile is shown in the table below:

S.No.	Station Code	Latitude (°N)	Longitude (°E)	Elevation (m)	Operation Period (year/month)	No. of RFs
01.	SKL	13.016	74.833	40	2012/04-2013/10	42
02.	ALN	13.102	74.995	98	2012/04-2013/12	39
03.	STD	13.107	75.086	94	2012/11-2013/12	41
04.	HSR	13.134	75.120	100	2013/02-2013/12	34
05.	KMK	13.219	75.252	841	2012/04-2013/10	27
06.	BHL	13.268	75.430	805	2012/04-2013/12	27
07.	SPT	13.347	75.525	750	2013/02-2013/12	48
08.	BGS	13.407	75.616	996	2012/04-2013/10	25
09.	SVR	13.519	75.799	1260	2012/07-2013/12	34
10.	KDR	13.574	76.016	774	2012/04-2013/12	78
11.	KLR	13.690	76.175	706	2012/04-2013/08	53
12.	HSG	13.778	76.282	723	2013/02-2013/12	52
13.	MLR	13.864	76.400	668	2012/04-2013/12	41
14.	UDL	13.884	76.556	648	2012/04-2013/10	47
15.	DMR	13.987	76.719	589	2012/04-2013/12	75
16.	HYB	14.063	76.808	608	2013/02-2013/09	27
17.	AMR	14.077	76.942	605	2012/04-2013/10	16
18.	LDH	14.142	77.042	620	2013/02-2013/11	13
19.	PVG	14.180	77.102	694	2012/04-2013/09	30
20.	YSN	14.187	77.275	601	2013/02-2013/10	20
21.	PRR	14.294	77.352	570	2012/04-2014/01	60
22.	RMG	14.318	77.568	453	2012/04-2014/01	83

23.	MMP	14.414	77.645	390	2013/02-2014/01	26
24.	DMM	14.479	77.748	350	2012/07-2014/01	17
25.	BTP	14.522	77.840	321	2013/02-2014/01	40
26.	PRP	14.544	78.014	343	2012/09-2014/01	26
27.	SKS	14.653	78.209	245	2012/09-2014/01	8
28.	MDN	14.652	78.370	217	2013/02-2014/01	13
29.	RTP	14.734	78.471	171	2012/09-2014/01	67
30.	DUV	14.830	78.675	185	2012/09-2014/01	55
31.	PML	14.908	78.913	167	2012/10-2014/01	36
32.	RSP	14.980	79.016	105	2013/02-2014/01	33
33.	SRM	15.038	79.155	222	2012/10-2014/01	65
34.	PMR	15.100	79.372	123	2012/10-2014/01	44
35.	MPD	15.141	79.516	105	2013/02-2014/01	46
36.	IRD	15.174	79.599	108	2012/10-2014/01	42
37.	PNU	15.277	79.801	28	2012/09-2014/01	43
38.	KDL	15.425	79.996	16	2013/10-2014/01	35

2.6 Common Conversion Point (CCP) Imaging

To obtain a more detailed image of the crust and upper mantle discontinuities like the Moho geometry, Lithosphere-Asthenosphere Boundary (LAB) and Mantle Transition Zone (MTZ) having 410 km and 660 km discontinuities and to investigate their lateral variability, depth migration of receiver functions is performed using the Common Conversion Point (CCP) method (Dueker and Sheehan, 1998). Assuming a horizontally layered medium, the CCP stacking approach involves tracing each ray path, back-projecting the waveform along the theoretical ray path, and bin stacking the waveforms to estimate the average depth of the interface. However, CCP stacking is not sensitive to small-scale interface variations and tends to produce a smooth image.

To investigate the crust and upper mantle discontinuities, the Common Conversion Point (CCP) stacking approach (Dueker and Sheehan, 1998) is employed, involving the following steps:

- (i) Computation of receiver functions, each with a waveform fit equal and greater or equal to 80% at Gaussian filter width (Gw) of 2.5 during deconvolution.
- (ii) Determination of the travel time and piercing point for every P-to-S conversion down to a depth of 60 km, using a 2 km interval for a known local velocity model with the TauP toolkit (Crotwell et al., 1999).
- (iii) Calculation of the theoretical Pds -P arrival time at each depth (d), which corresponds to a specific time. Amplitude values associated with each timing are extracted from the receiver function at various depths.
- (iv) Division of the study area into rectangular or square bins based on their piercing points at each depth. The piercing point is defined as the location at depth where P-to-S conversion occurs. The distribution of piercing points at depths of 20 km, 40 km and 60 km are shown in the next chapter on crust. To create a depth image of the discontinuity, the amplitudes from individual receiver functions that penetrate a specific area are summed (stacked) using:

$$A(d) = \frac{1}{N} \sum_{i=1}^N A_i(T_{Pds}^i),$$

Here, A(d) represents the stacking amplitude for the discontinuity at depth d, while N denotes the number of receiver functions that intersect at that specific depth. $A_i(T_{Pds}^i)$ refers to the amplitude of the i th receiver function at the time T_{Pds}^i . T_{Pds}^i is the moveout time for the i th receiver function associated with a discontinuity, calculated using the following:

$$T_{Pds} = \int_{-d}^0 (\sqrt{V_s(z)^2 - p^2} - \sqrt{V_p(z)^2 - p^2}) dz,$$

In this context, p represents the ray parameter for the P wave, while $V_p(z)$ and $V_s(z)$ denote the P and S wave velocities at a given depth z . By integrating data from multiple stations, noise generated by topographic scattering can be minimized, thereby enhancing the signal-to-noise ratio (Dueker and Sheehan, 1998). For this study, rectangular bins having size $0.2^\circ \times 0.2^\circ$ with

0.1° overlapping along the E-W Profile are utilized. The receiver functions are then stacked to get the amplitude for each bin. It is required that each bin contains a minimum of 8-10 traces. Fortunately, in our study, the seismic stations are densely spaced having 10-15 km spacing and each station having enough RFs to get the high resolution CCP stacked image of the crust along the E-W profile. The piercing points and the resulting CCP stacked waveform image along the profile are shown in the next chapter on crust (Figure 3.4 and 3.5 of chapter 3 on crust).

2.7 Velocity inversion

2.7.1 Surface waves -

Most broadband seismograms are primarily dominated by the longer period waves that arrive after the P and S waves (Figure 2.2). These waves, known as surface waves, propagate along the Earth's surface. Surface waves are defined as waves that are generated in media with a free surface and propagate parallel to that surface. They are produced by energy that reaches the free surface from incident body waves and are typically the most prominent arrivals recorded at teleseismic distances. There are two main types of surface waves: Rayleigh waves and Love waves. In laterally homogeneous materials, Rayleigh waves are polarized radially and can occur at any free surface, whereas Love waves are polarized transversely and necessitate an increase in velocity with depth. Because of their dispersive characteristics, these waves offer significant information about the composition of the Earth's crust and upper mantle. Dispersion is characterized by the existence of two types of velocities: Group velocity (U) and Phase velocity (c), both of which depend on frequency (ω) or wave number (k). The group velocity and phase velocity can be written as:

$$U = d\omega/dk \quad \text{and} \quad c = \omega/k$$

And the relation between the two is:

$$U = c/(1 - kdc/d\omega)$$

In this thesis, we have used the phase velocity dispersion to construct velocity structures.

To accurately map the velocity structure of both the crust and lithosphere, a continuous phase velocity generated from Rayleigh waves from shorter to the longer periods are needed (Salaün et al., 2012). The shorter period dispersion is computed from ambient noise while for the longer period, earthquakes/events are used. In traditional deep Earth exploration, we analyze the waveforms from earthquakes recorded across a network of seismic stations. The spatial resolution of the experimental domain in these cases relies on the illumination of the region, which is influenced by the geometry of the earthquake sources and recording locations. In absence of uniform azimuthal distribution of the earthquake around the region of interest, the quality of Earth image is affected. Given the anisotropic distribution of earthquake sources, three-dimensional imaging of the shallow Earth often remains incomplete. Furthermore, conventional earthquake tomography limits detailed mapping of shallow crustal structures because waveforms recorded far from source regions lose high-frequency information due to intrinsic attenuation and scattering. To address this issue, we utilize ambient seismic noise generated by natural phenomena such as wind, ocean waves, rock fracturing, and human activities. These waves continuously propagate through the Earth and contain valuable information about its subsurface within their complex wave field. Ambient seismic waves are naturally occurring and provide a constant source of energy; however, they are typically considered unusable since they are not impulsive and are often removed from seismic data.

Ambient noise consists of continuous vibrations generated by low-amplitude sources that occur randomly (Larose et al., 2008). At higher frequencies (above 1 Hz), this noise is primarily produced by human activities such as transportation and industrial processes. Below 1 Hz, the sources of noise include natural phenomena like wind, storms, and ocean waves. For periods ranging from 1 to 20 seconds, these ambient vibrations are referred to as microseismic noise (or microseisms). Seismic noise is typically observed in two distinct periods: the primary (~15 seconds) and secondary (~7.5 seconds) microseisms. The primary microseism is thought to be associated with the direct forcing of strong ocean waves (Hasselmann, 1963), while the secondary microseism arises from the nonlinear interaction between two primary waves that have the same frequency but travel in opposite directions. The precise source locations of secondary microseisms remain a topic of debate (Yang and Ritzwoller, 2008). At very low frequencies, where fundamental modes of the Earth are observed (periods ranging from 100 to 1000 seconds), ambient noise is referred to as "hum". Across all frequencies, this noise is

generated by surface sources, which means the signals extracted from the noise are expected to predominantly consist of surface waves. In this study, we focus exclusively on the vertical component of seismographs, which limits our analysis to Rayleigh waves only.

When analyzed over sufficiently long periods, background seismic noise includes surface waves propagating in multiple directions. As a result, it contains valuable information about surface wave propagation and, consequently, the elastic structure of the crust and uppermost mantle. This information can be extracted by calculating the two-point cross-correlation between noise records from each pair of stations. The connection between the cross-correlation and the Green function of the wave propagating between the station pair is well established (e.g., Weaver and Lobkis, 2005; Snieder, 2004). The application of background noise to extract Green functions has been successfully implemented in various fields, including helioseismology, ultrasonics, exploration seismology, and marine acoustics. This method offers several advantages over conventional approaches: it does not require any external sources, provides uniform illumination of the study area, can be utilized in regions without earthquakes, and is highly cost-effective (Shapiro et al., 2005; Bensen et al., 2008; Ritzwoller et al., 2011).

2.7.2 Seismic Interferometry -

Interferometry is a phenomenon that involves the interference of pairs of signals to obtain important information about the subsurface. In contrast to many conventional tomographic imaging methods that have been used for a long time, the use of ambient noise for imaging the Earth is a relatively new technique (Wapenaar, 2003, 2004; Campillo and Paul, 2008; Wapenaar et al., 2011). The core concept of seismic interferometry is that the Green's function between two seismic stations can be approximated by cross-correlating a long time series of ambient noise recorded at those stations. This Green's function can be interpreted as the seismogram obtained at one location due to an impulsive or instantaneous source of energy at another location. Imagine two receivers located at r_1 and r_2 , enveloped by energy sources situated on an arbitrary boundary S (Figure 2.8). The wavefield generated by each source propagates through the medium within S and is captured by both receivers. The recorded signals at these two receivers are then cross-correlated. When the cross-correlation of all sources is summed, the energy traveling along the path will constructively interfere, while energy not following that path will destructively interfere. Consequently, the Green's function obtained between r_1 and r_2 will simulate the

situation as if one of the receivers were actually a source (Figure 2.8) (Wapenaar, 2003 and 2004). In the case of random noise, a surface S exists that encompasses all noise sources, and since these sources may activate simultaneously or overlap in time, their recorded signals at both receivers are effectively summed together in a natural manner. Snieder (2004) demonstrated that seismic sources located around the extension of the inter-receiver path contribute most significantly to constructing the interferometric Green's function, indicating that a complete boundary of sources is not necessary to approximate the inter-receiver Green's function. Given that strong sources of seismic noise are typically confined to locations within the Earth's crust and that surface waves travel along interfaces between different layers in the crust and upper mantle, the surface wave component of the inter-receiver Green's function appears distinctly in seismograms constructed through seismic interferometry.

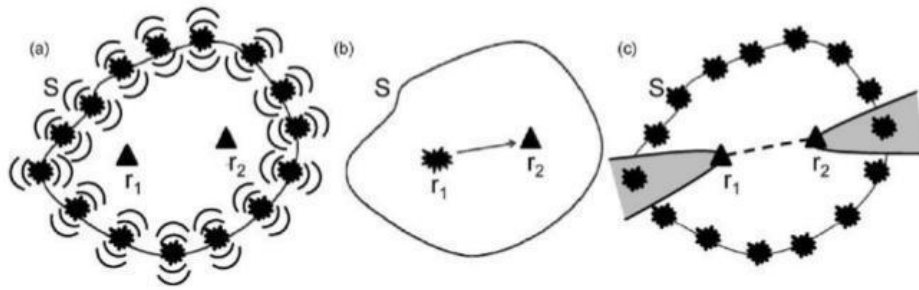


Fig. 2.8. (a) Two receivers (represented by triangles) are enclosed by a boundary S of sources, each of which emits a wavefield into both the interior and exterior of S . (b) The seismic interferometry method transforms one of the receivers (r_1) into a virtual source, allowing for the generation of a real seismogram. (c) Sources located within the shaded regions contribute the most significantly to the computation of the Green's function (Nicolson et al., 2012).

Green's function of a medium between two locations, A and B , illustrates the response captured at point A when an impulsive source is activated at point B . In a completely stochastic wavefield, the cross-correlation of the signals gathered between these two points approaches the complete Green's function of the medium, which includes all modes of reflection, scattering, and propagation (Weaver & Lobkis, 2005). If the ambient noise sources are evenly distributed and recorded by sensors located at points A (ϕ_A) and B (ϕ_B), then taking the time derivative of the

correlation provides the precise Green's function of the medium, as though the source were situated at either A or B (Wapenaar, 2004; Weaver and Lobkis, 2005; Larose et al., 2008):

$$\delta_{\tau} C_{AB}(\tau) = \delta_{\tau} \int \varphi_A(t) \varphi_B(t + \tau) dt \propto G^+(A, B, \tau) - G^-(A, B, -\tau)$$

By cross-correlation of vertical components of ambient noise from stations located between approximately 100 km to more than 2000 km apart, Shapiro and Campillo (2004) demonstrated the feasibility of extracting fundamental mode Rayleigh waves with periods ranging from less than 10 seconds to more than 100 seconds.

2.7.3 Ambient noise data processing

In this analysis, the data is processed following Bensen et al. (2007). One day (24 hours) vertical component data from a total of 38 stations are first corrected for instrument response, mean, trend. The waveforms are further decimated at 10 samples per second and are band-passed between 1-50 s. We apply time-domain normalization (i.e., running absolute mean) to the processed waveforms to remove the effect of earthquakes. Additionally, spectral normalization is applied to flatten the response in the frequency domain. Subsequently, the single-day data for a given pair of stations are cross correlated and stacked to produce Empirical Green's Functions (EGF; e.g. Shapiro & Campillo, 2004; Shapiro et al., 2005). Fig. 2.9a provides the stacked correlations with increasing distances. Fundamental mode Rayleigh wave phase velocity measurements for each EGFs are performed using the Frequency-Time Analysis (FTAN) method (Levshin & Ritzwoller, 2001). Phase velocity measurements are obtained in the period range of 3-50 seconds (Fig. 2.9b).

Selection of reliable dispersion measurements is important before any further processing, such as tomography, is performed (Bensen et al., 2007). First, we examine individual dispersion measurements manually to make sure that they lie within the expected range. Additionally, dispersion measurements with SNR > 10 are selected. Traditionally, the maximum observed period in the dispersion for a given station pair is selected only if the inter-station distance is within three times of the wavelength (Bensen et al., 2007). However, Luo et al. (2015) showed that inter-station distance within single wavelength is also reliable. Recently, Xie et al. (2023) showed that at least for the horizontal components the three-wavelength criteria should be

followed. Application of the three-wavelength criterion reduces the number of ray paths at longer periods (Fig. 2.10). To arrive at a conclusion, we performed resolution analysis for both i.e., single-wavelength (1λ) and three-wavelength (3λ). For this, we designed a synthetic checkerboard model with alternating negative and positive anomalies of two sizes:

(i) 10 km-sized anomalies separated by 10 km from each other.

(ii) 20 km-sized anomalies separated by 20 km from each other (Fig. 2.11). Forward modeling of dispersion curves is made for all ray paths (Fig. 2.9b) and subsequently inverted to recover the input anomalies. This process is performed using linear, least-squared based tomography, called the Fast Marching Surface Tomography (FMST), of Rawlinson & Sambridge (2005). The recovered patterns of the 1λ criteria is shown in Fig. 2.12, which shows that 10 km-sized anomalies are mainly recovered at periods ≤ 40 seconds. However, the 20 km-sized anomalies are reasonably recovered up to the maximum period of 50 seconds. For the 3λ case, shown in Fig. 2.13, the recovery for 20 km sized anomalies are reliable up to 30 seconds only and suffer from strong smearing beyond 30s. This is expected as the number of ray paths reduces significantly beyond 30 s (Fig. 2.10) due to strict criteria of 3λ . In order to maintain only robust measurements, we preferred the 3λ criteria and chopped off all the dispersion measurements beyond 30 s for further analysis.

The phase velocity measurements as discussed earlier are used to produce a 2-D phase velocity map following the FMST method (Rawlinson & Sambridge, 2005) at a grid-interval of 20 km in the period range of 3 to 30 seconds, which is sensitive up to a maximum depth of 60 km (Fig. 2.14). Fig. 2.15 shows a comparison of phase velocity tomography for the 3λ and 1λ criteria, which shows that the difference is minimal $\sim 1\%$ at shorter periods (≤ 30 s) and $\sim 2\%$ at periods more than 30 s.

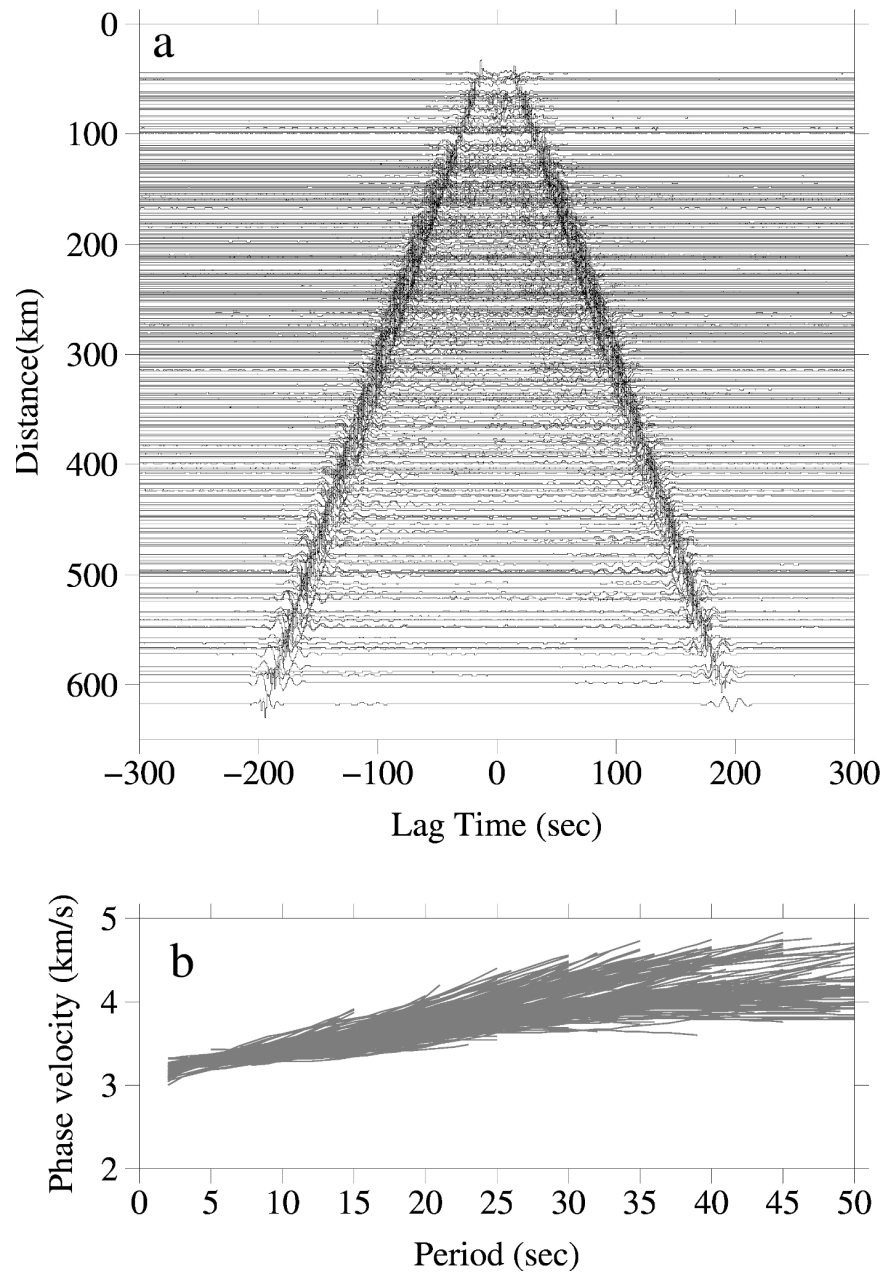


Fig. 2.9. (a) stacked correlations and, **(b)** phase velocity dispersion measurements.

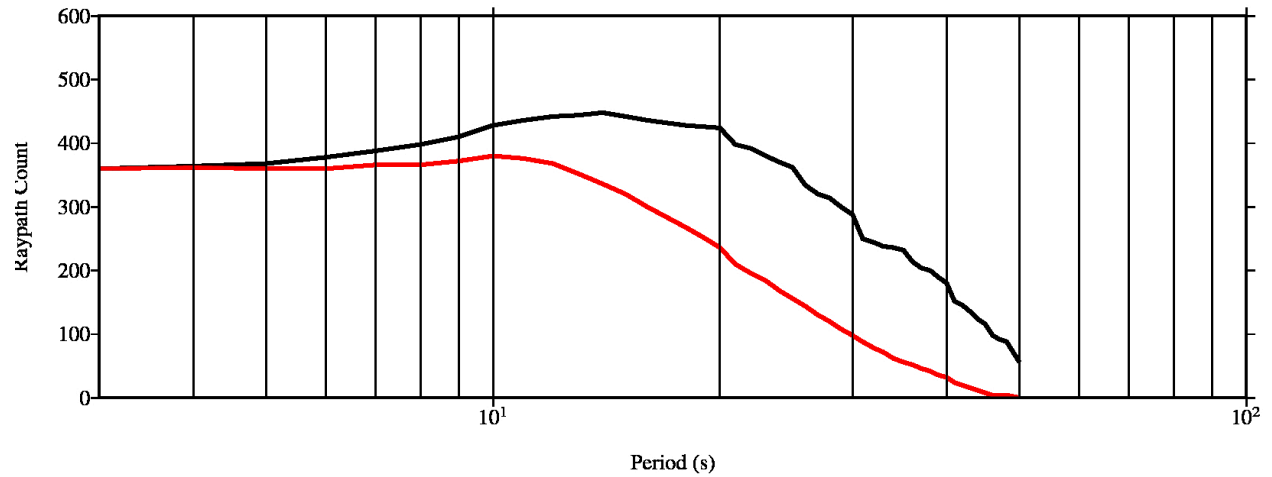


Fig. 2.10. Raypath count at different period after considering 3λ (red color) vs 1λ (black color).

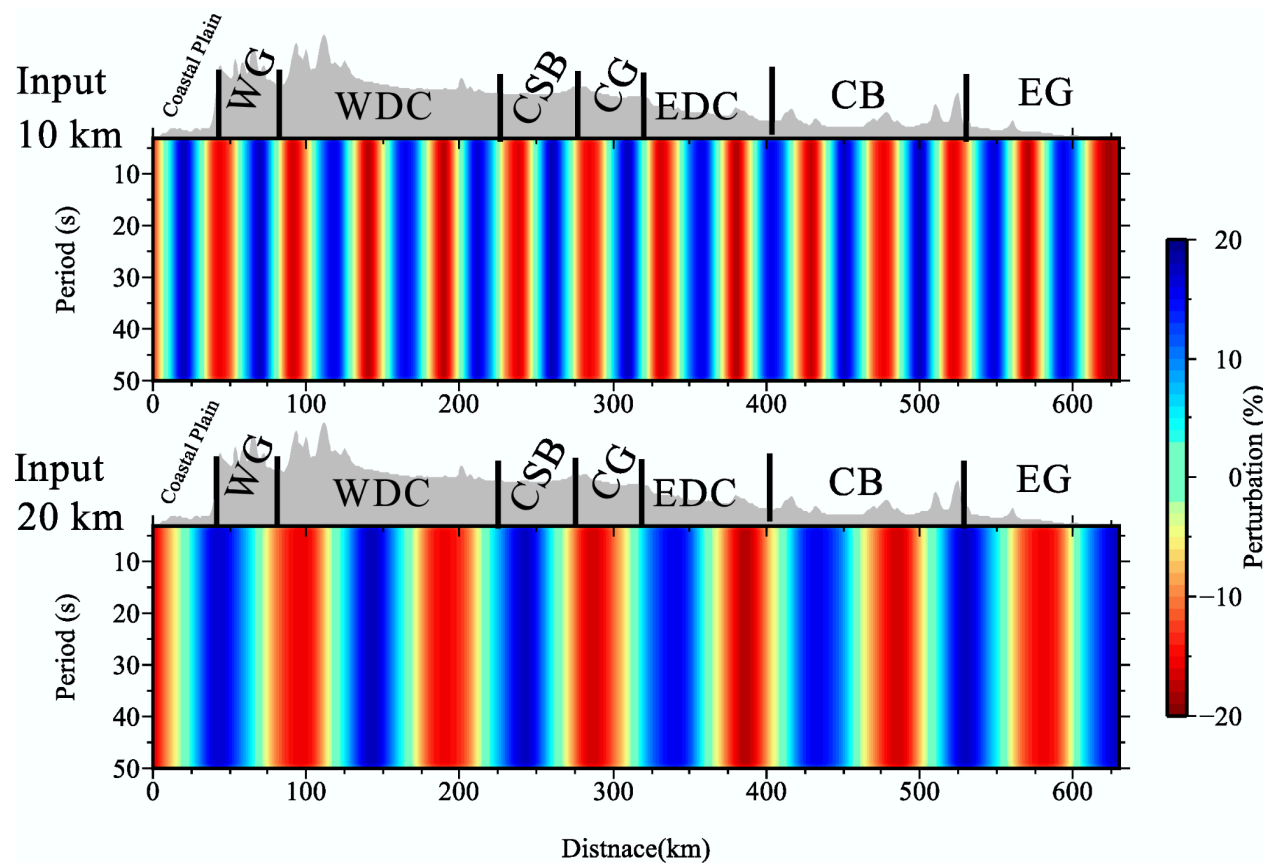


Fig. 2.11. Input checker board at 10 km (top) and 20 km (bottom).

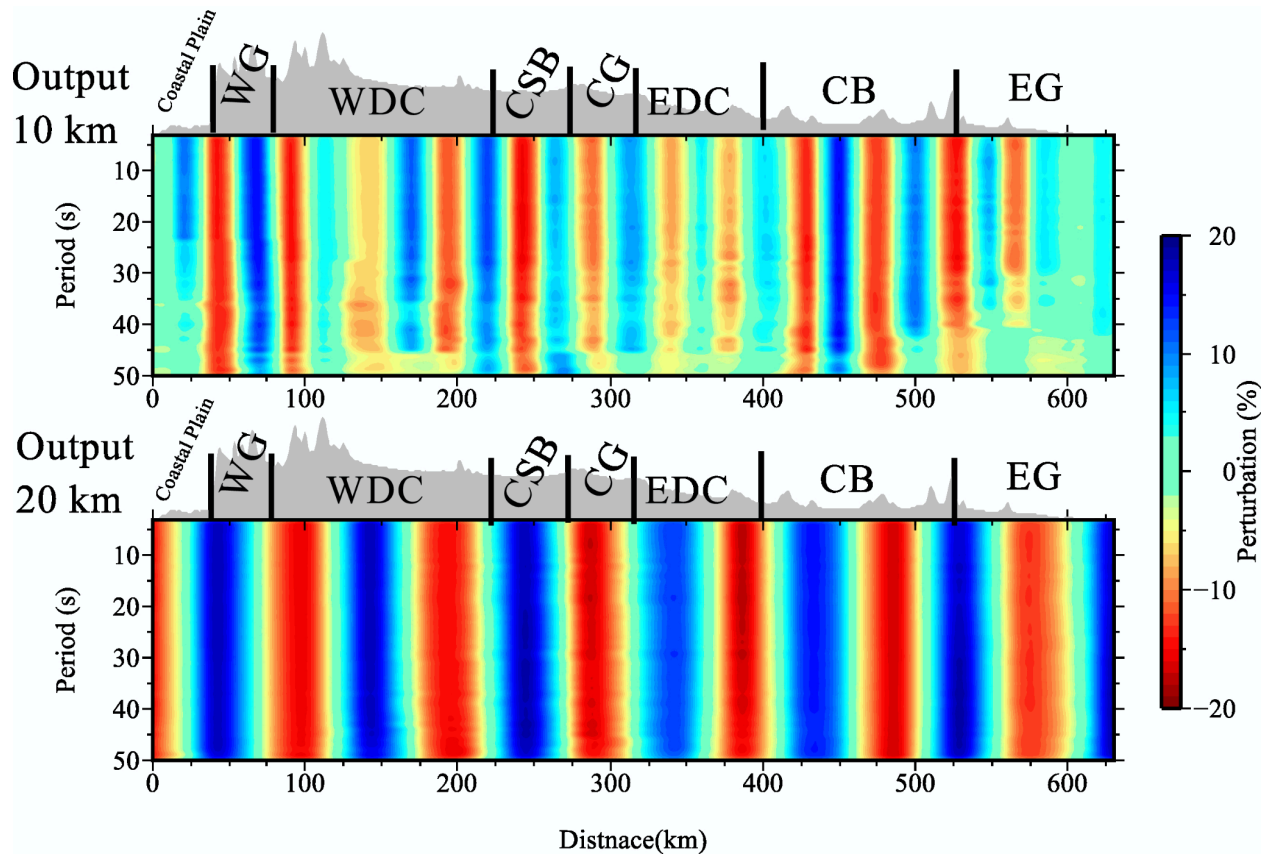


Fig. 2.12. Output checkerboard model at 10 km and 20 km considering 1λ .

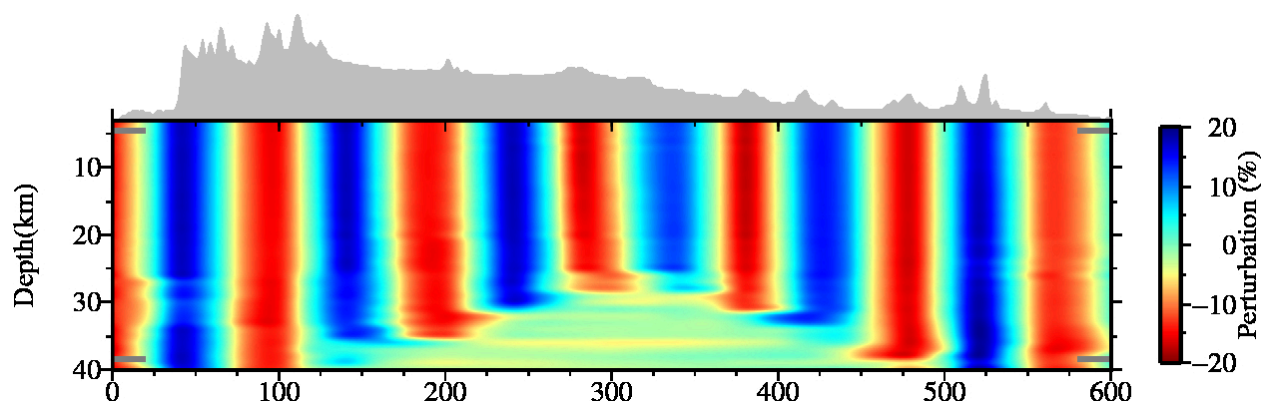


Fig. 2.13. Output checkerboard model at 20 km considering 3λ .

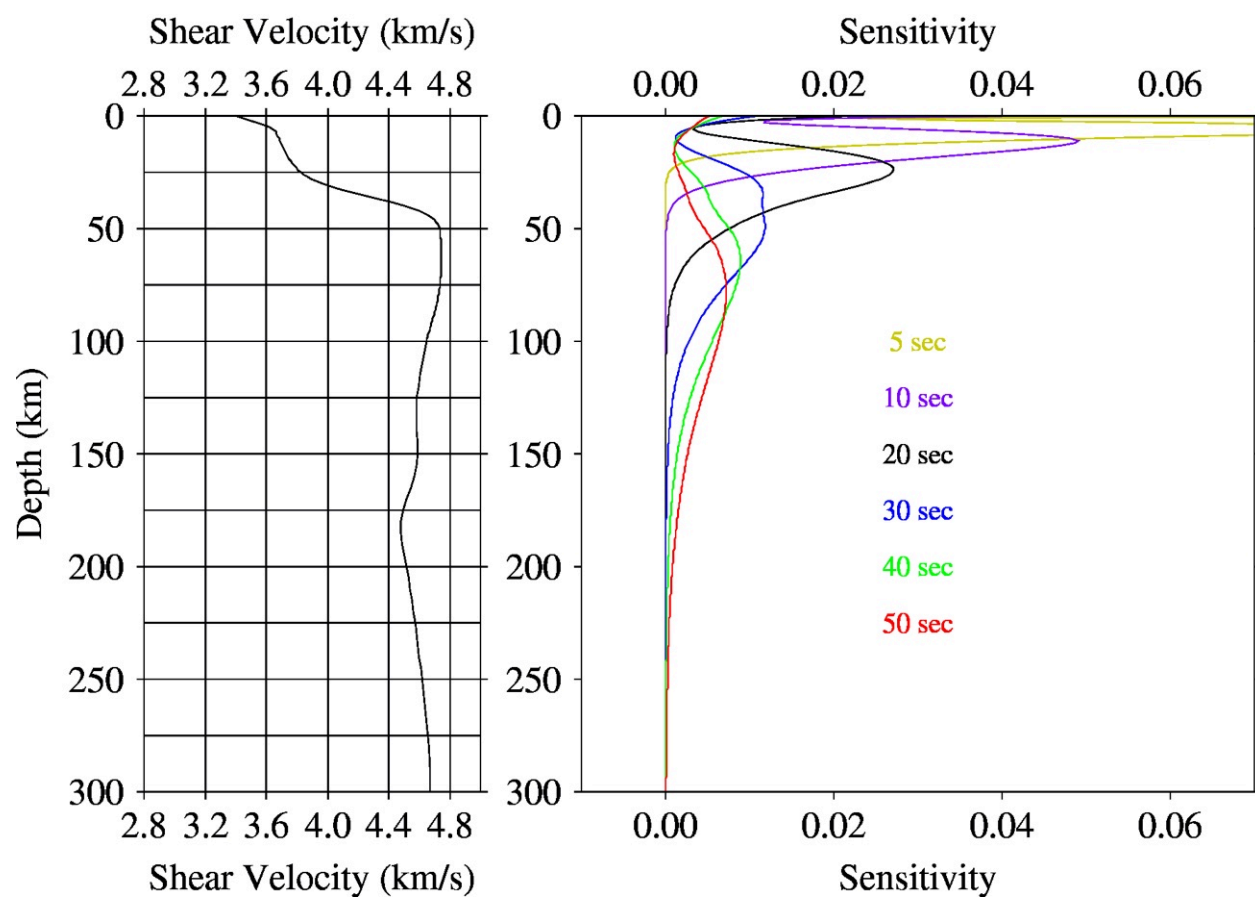


Fig. 2.14. Sensitivity kernels for Rayleigh waves.

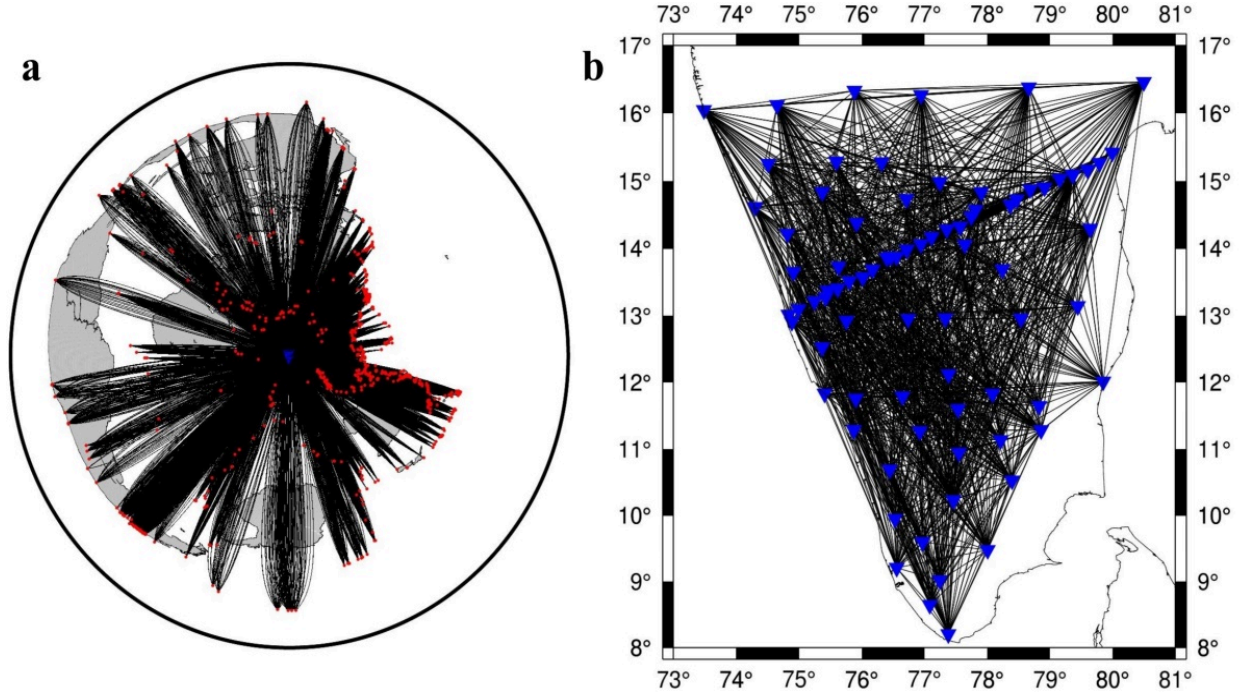


Fig. 2.16. a) Ray diagram connecting earthquake epicenters and seismic stations, b) Station distribution and inter-station links between the stations for earthquake surface wave tomography (Mullick et. al., 2022).

Seismic surface waves are one of the important tools to probe the deep structure of the Earth (Jin et al., 2015). Their energy, as a result of geometrical spreading, spreads two-dimensionally and decays from the source approximately as inverse of distance r . As a result, at a large distance from the source, these waves are prominent in seismograms. Its dispersion feature, having velocity variation with different periods, provides sensitivity to different depths (Jin et al., 2015).

Surface wave tomography is a two-step inversion scheme:

(i) 2-D phase or group velocity maps for different periods are estimated through tomographic inversion at each geographical point of interest using arrival time as data and phase or group velocity as model parameters (Zhang et al., 2018; Barone et al., 2021).

$$t = \int (1/V_x) dl = \int P_x dl$$

(1)

Where t is travel time (observed data), V_x and P_x are group/phase velocity and group/phase slowness (model parameters) at the position $x = (x_1, x_2, x_3)$ and dl is an infinitesimal segment of the path over which the integration is performed (Barone et al., 2021).

(ii) Group/phase velocity at different frequencies is then inverted to get velocity variation with depth at different points of interest.

2.8.1 Tomographic Problem -

In a 3D space, The above tomographic problem Eq. (1) is highly non-linear as the wave path depends on the phase velocity distribution in space which is by definition unknown. So the above forward problem Eq. (1) can be reformulated in terms of travel time perturbations -

$$\delta t = t - t_o = \int P(x) dl - \int P_o(x) dl \quad (2)$$

Where $P_o(x)$ is group/phase slowness computed from a reference model which is not too far from the real one. So Eq. (2) can be re-expressed as -

$$\delta t = \int \delta P(x) dl \quad (3)$$

Which can further be written as

$$Gm = d \quad (4)$$

Where m is the model vector having s unknown model parameters, d is the data vector having t observations and G is $s \times t$ matrix, whose G_{ij} element, has the length of the raypath at each j^{th} model element for the i th observation.

The least square solution to the Eq. (4) can be written as

$$\mathbf{m} = (\mathbf{G} \mathbf{G})^{-1} \mathbf{G} \mathbf{d} \quad (5)$$

Tomographic problems are generally ill-posed and need regularization for stability of the solution. Thus existing tomographic problems require reference model and regularization. Jin & Gaherty (2015) has built the cross-correlation technique based on generalized seismological data functional (GSDF) analysis of Gee & Jordan (1992) which utilizes the cross-correlation between observed and synthetic waveforms to quantify phase and amplitude variation. This is the first application of GSDF in which an observed waveform from a nearby station at S_2 is substituted as synthetic. Station S_1 has an observed waveform.

The first step is to apply a window function W_S on waveform at station S_2 that includes a primary surface wave and its coda as shown in figure 2.17. So the cross-correlogram $C(t)$ between S_1 and $W_S S_2$, defined as:

$$C(t) = S_1 \star W_S S_2 \quad (6)$$

$C(t)$ provides the delay or lag information of all coherent signals, having dominant energy (center frequency) typically around the period of 30 s for teleseismic Rayleigh waves.

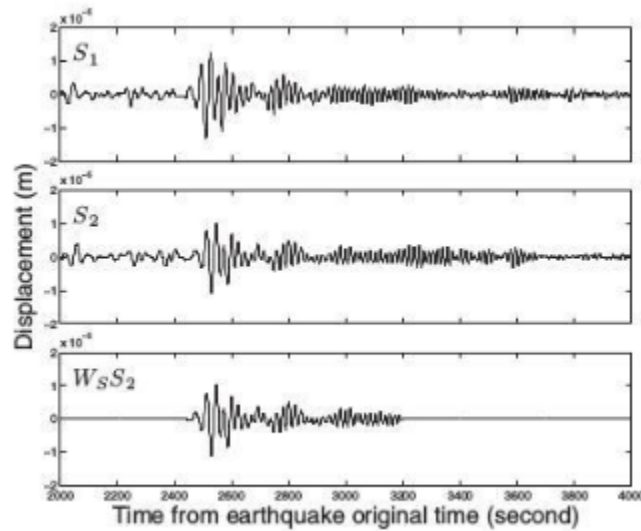


Fig. 2.17. Two waveforms recorded at stations S_1 and S_2 having waveform at station S_2 windowed.

Further isolation of the signal is done by windowing the cross-correlogram $C(t)$ by W_c to get $W_c C(t)$. Then the signal of interest in frequency domain is obtained by convolving $W_c C(t)$ with a sequence of Gaussian, narrow band filters, getting $F(\omega_i) \star W_c C(t)$ for each center frequency ω_i . These filtered cross-correlograms $F(\omega_i) \star W_c C(t)$ is the source of information for all frequency dependent group and phase delays between two stations as shown in Fig. 2.18. The narrow band filtered cross-correlation is approximated by a five-parameter wavelet which is the product of a Gaussian envelope and a cosine function (Gee & Jordan, 1992):

$$F(\omega_i) \star W_c C(t) \approx A G_a[\sigma(t - t_g)] \cos[\omega(t - t_p)] \quad (7)$$

Where t_g and t_p represent the group and phase delays between the stations as functions of frequency, respectively. G_a denotes the Gaussian function, A is a positive scaling factor, σ represents the half-bandwidth, and ω indicates the central frequency of the narrow-band waveform.

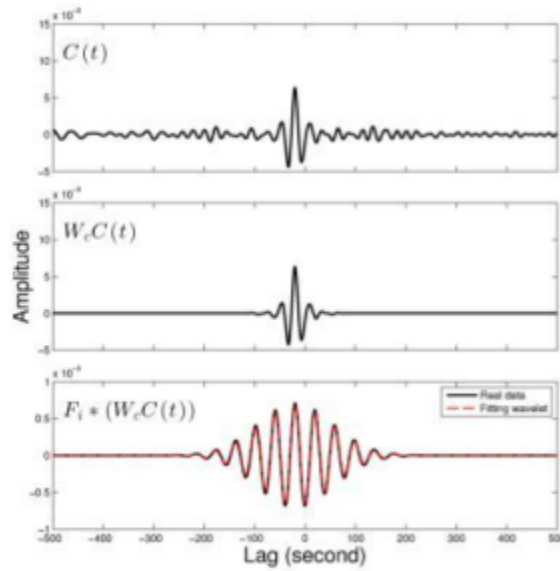


Fig. 2.18. Cross-correlograms windowed and filtered.

The window function W_s applied initially may also introduce bias which also contribute group and phase delay, if S_2 and $W_s S_2$ are not identical within the window of interest. So

$$C'(t) = S_2 \star W_s S_2, \quad (8)$$

$$F_i * W_c C'(t) \approx A' G_a [\sigma^i(t - t_g')] \cos[\omega^i(t - t_p')] \quad (9)$$

So from Eq. (7 and 9), we calculate a final bias-corrected delay times

$$\delta\tau_p = t_p - t_p' \quad (10)$$

$$\delta\tau_g = t_g - t_g' \quad (11)$$

Window function (W_c) also introduces a bias caused by non-coincidence of the center of the window function with the actual group delay at each frequency. This bias is negligible when taken between two nearby stations.

The phase delay time $\delta\tau_p$ from Eq. (10), can be used to estimate the phase velocity by the Eikonal equation

$$1/c(\vec{r}) = \left| \nabla \tau(\vec{r}) \right| \quad (12)$$

Where $c(\vec{r})$ is apparent phase velocity and $\tau(\vec{r})$ is phase travel time. Eq. (12) can be described in the integral form of the vector field as:

$$\Delta\tau_p = \int S(\vec{r}) \cdot d\vec{r} \quad (13)$$

Where $S(\vec{r})$ is the slowness vector and $\vec{r}$ is the spherical path connecting two stations as shown in Fig. 2.19.

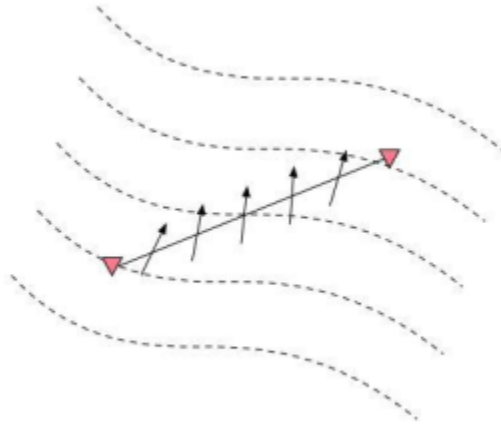


Fig. 2.19. Spherical path connecting two stations where dash lines illustrate wavefronts, red triangles are stations, black line is an inter-station link, and black arrows are slowness vectors.

The bias between apparent and structural phase velocity can be corrected by adding amplitude measurements into the inversion, using Helmholtz equation:

$$1/c(\bar{r}) = 1/c^s(\bar{r}) - \nabla^2 A(\bar{r})/A(\bar{r})\omega^2 \quad (14)$$

Where $c(r)$ is the structural phase velocity, $A(\bar{r})$ is the amplitude field and Laplacian term corrects for the influence of non-plane wave propagation on the apparent phase velocity $c(\bar{r})$.

Besides the general procedure of accepting good seismograms with above a certain threshold SNR, there are two additional components of data selection which are needed for automation of large datasets processing.

- (i) The selection of window function W_s should be wide enough to isolate fundamental-mode energy of the surface waves as well as to include the arrival of maximum amplitude of all frequencies and narrow enough to eliminate the higher-mode phases and body waves.
- (ii) Coherence at a certain frequency between waveforms of nearby stations is defined as:

$$\gamma^2 = A_{12}^2 / A'_{11} A'_{22} \quad (15)$$

Where A_{12} is the amplitude of narrow band cross-correlation wavelet and A'_{11} , A'_{22} are amplitudes of narrow band autocorrelation wavelets of the two stations. Actually A'_{11} , A'_{22} can be estimated by $C'(t)$ from Eq. (8). This coherence can be set to a certain threshold value, let's say, 0.5 below which the measurements are excluded.

(iii) Linearity between phase delays and difference in epicentral distance is followed. We discard all the measurements with misfit more than 10s relative to the linear prediction as shown in Fig. 2.20 below.

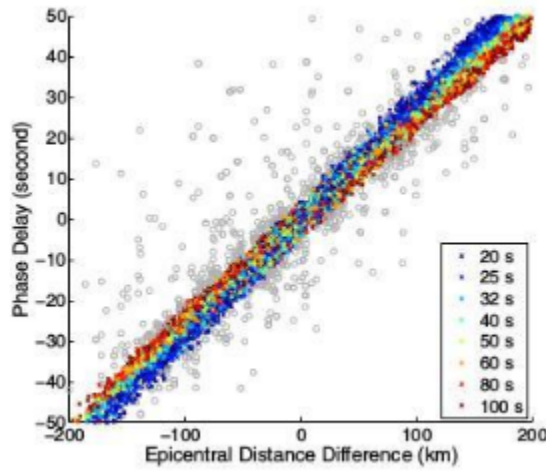


Fig. 2.20. Linearity between phase delays and difference in epicentral distance where different color dots represent measurements for different frequencies and gray color represent poor measurements that can be discarded.

2.8.2 Comparison with FTAN and Eikonal and Helmholtz tomography

(i) In this method, Cross-correlation of waveforms suppresses the influence of random noise, not being coherent among stations. It provides measurement having more robustness for the seismogram with relatively low SNR.

(ii) In this method, we measure differential phase between stations instead of absolute phase in FTAN method which applies Hilbert transform to single-station waveforms to retrieve absolute phase values. This differential phase is directly inverted for slowness vector instead of

reconstructing traveltime surface $\tau(\vec{r})$ and then estimating its gradient to obtain apparent phase velocity $c(\vec{r})$ from Eikonal equation.

Although being a better technique, measurements from this method can be erroneous. The possible sources of error are: station time-shift or an abnormal amplification problem, azimuthal anisotropy and overtone interference but these sources of error are small and can be avoided.

2.8.3 Surface wave dispersion data -

Surface waves have been extensively utilized to reconstruct regional velocity structures. To effectively delineate the velocity structure of the crust and lithospheric mantle, it is essential to obtain a continuous phase velocity across both short and long periods (Salaün et al., 2012). We integrated phase velocity dispersion data discussed in section 2.7.3 and Mullick et al. (2022) for periods ranging from 3 to 140 seconds. The phase velocity dispersion for shorter periods (3–30 seconds) was derived from ambient noise tomography results by discussed in section 2.7.3, following methodologies established by Shapiro and Campillo (2004), Shapiro et al. (2005), Bensen et al. (2007), and Bensen et al. (2008). For longer periods (30-140 seconds), we utilized Rayleigh wave phase dispersion from Mullick et al. (2022), which was computed using earthquake surface wave tomography via the Automated Surface Wave Measuring System (ASWMS) method developed by Jin and Gaherty (2015). This analysis was conducted at a spatial resolution of 55 km, utilizing data from 846 teleseismic earthquakes with magnitudes greater than 5.5, recorded across 85 broadband stations (**Fig. 2.16**). Rayleigh wave phase velocity dispersion map for longer periods for South India derived from Mullick et al., 2022 is shown in Fig 2.22. For shorter periods, the Rayleigh wave dispersion is done only along the E-W profile (Fig. 1.4) which is discussed in detail in the section 2.7.3.

The quality and horizontal resolution of phase velocity maps derived from ambient noise tomography tend to decline at longer periods beyond 30 seconds, while the data at this threshold represents the minimum for the velocity map provided by Mullick et al. (2022). The phase velocity maps generated through ambient noise and earthquake tomography reflect varying depths of resolution and scales of inhomogeneity within the subsurface. Notably, Das and Rai (2017) focused on crustal characteristics, whereas Mullick et al. (2022) concentrated on upper mantle inhomogeneities. For an average continental crust thickness of 35 km, phase dispersion

around 30 seconds predominantly reflects contributions from depths of approximately 20–40 km, corresponding to the lower crust, with contributions decreasing for the upper mantle (**Fig. 2.21**). At a period of 40 seconds, the phase velocity primarily originates from the uppermost mantle at depths of 50–70 km. Given the significant variability in lower crust velocities and Moho depths throughout South India, we anticipate notable lateral variations in phase velocity at 30 seconds compared to those at 40 seconds. Similar discrepancies in phase velocities obtained through different methodologies have been documented in other studies (Fig. 5d of Porritt et al., 2015; Pourpoint et al., 2018). The phase velocity dispersions having 50, 100 and 140 seconds (Fig. 2.21) have their maximum sensitivities around the depths 80 km, 120 km and 220 km respectively. The maximum period dispersion data having 140 seconds used in this thesis can reliably image the depth down to 300 km.

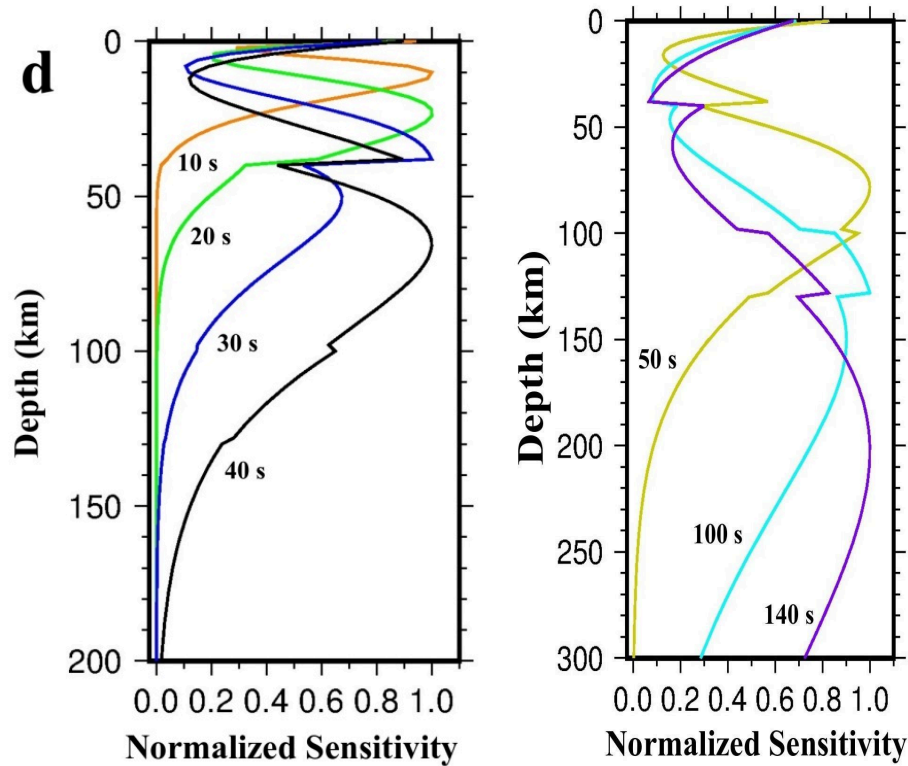


Fig. 2.21. Sensitivity kernels showing probing depth of Rayleigh wave at different time periods 10s, 20s, 30s, 40s, 50s, 100s and 140s for IASP91 earth model.

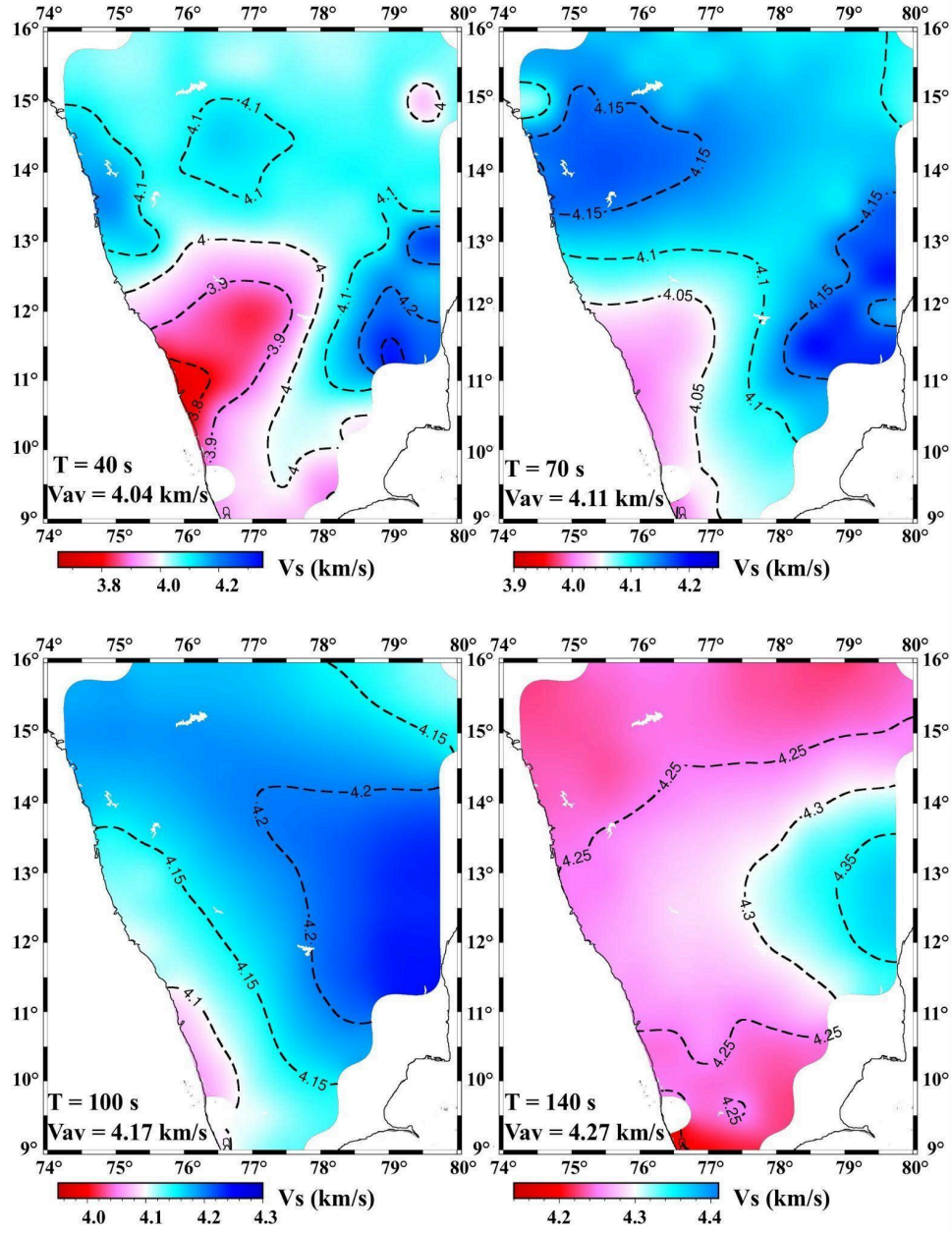


Fig. 2.22. Rayleigh wave phase velocity maps of E-W profile in South India at different time periods. Data for short periods are shown in the above ambient noise analysis section and data for periods (40 s, 70 s, 100 s, 140 s) derived from earthquake surface wave tomography (Mullick et al., 2022) are shown here.

2.9 Joint inversion -

The shear wave velocity (V_s) of each station is modeled by joint inversion of receiver function and surface wave phase velocity dispersion data from ambient noise and the earthquake that provides the dispersion in the period range 3-140 s . Receiver function is sensitive to velocity contrast across the layers, the surface waves provide the absolute velocity (background velocity) of these layers. Both datasets complement each other, their joint inversion reduces the non-uniqueness in the solution and provides better constraints and resolution in the final model. Additional benefit of adding receiver function is the high frequency content by which better depth resolution (vertical resolution) is achieved.

2.9.1 Methodology -

The joint inversion is performed by an iterative damped least square scheme (Julia et al., 2000). The velocity model is expected to minimize the objective function given by,

$$S = \frac{(1-p)}{N_r} \sum_{i=0}^{N_r} \left(\frac{O_{ri} - P_{ri}}{\sigma_{ri}} \right)^2 + \frac{p}{N_s} \sum_{j=0}^{N_s} \left(\frac{O_{sj} - P_{sj}}{\sigma_{sj}} \right)^2 \quad (16)$$

Where N_r and N_s are the total number of receiver functions and the surface wave dispersion data respectively. The parameter p determines the relative weighting between phase velocity and RF to the solution (Julia et al., 2000). It varies from 0 for only RF to 1 for only SWD. O_{ri} and P_{ri} are the observed and predicted receiver function at time t_i ; O_{sj} and P_{sj} are the j^{th} observed and predicted surface wave dispersion; σ_{ri} and σ_{sj} are the standard errors in each data set respectively. Since the uncertainties (σ) are squared, they contribute to the weighting of each datum more than p . To help stabilize the solution in the linearized inversion, a damping parameter is incorporated and least square minimization of above objective function (S) is done.

Joint inversion of RF and phase velocity dispersion is done having weight (p) = 0.1 and damp (θ) = 0.5. The selection of weight and damping parameters is done in the following way (Fig. 2.23):

- (a) For selection of damping parameter (θ), L-curve is plotted and the optimum damp value ($\theta = 0.5$) having a good trade-off between model roughness and RMS residual is selected. excess damping value may result in excess smoothing which is avoided here.
- (b) For the selection of optimum weight (p), examination of weight from 0 to 1 is done. To provide more weightage to RFs, for better smoothness and better fit of predicted RFs to the observed RFs, optimum weight (p) = 0.1 is finally selected.

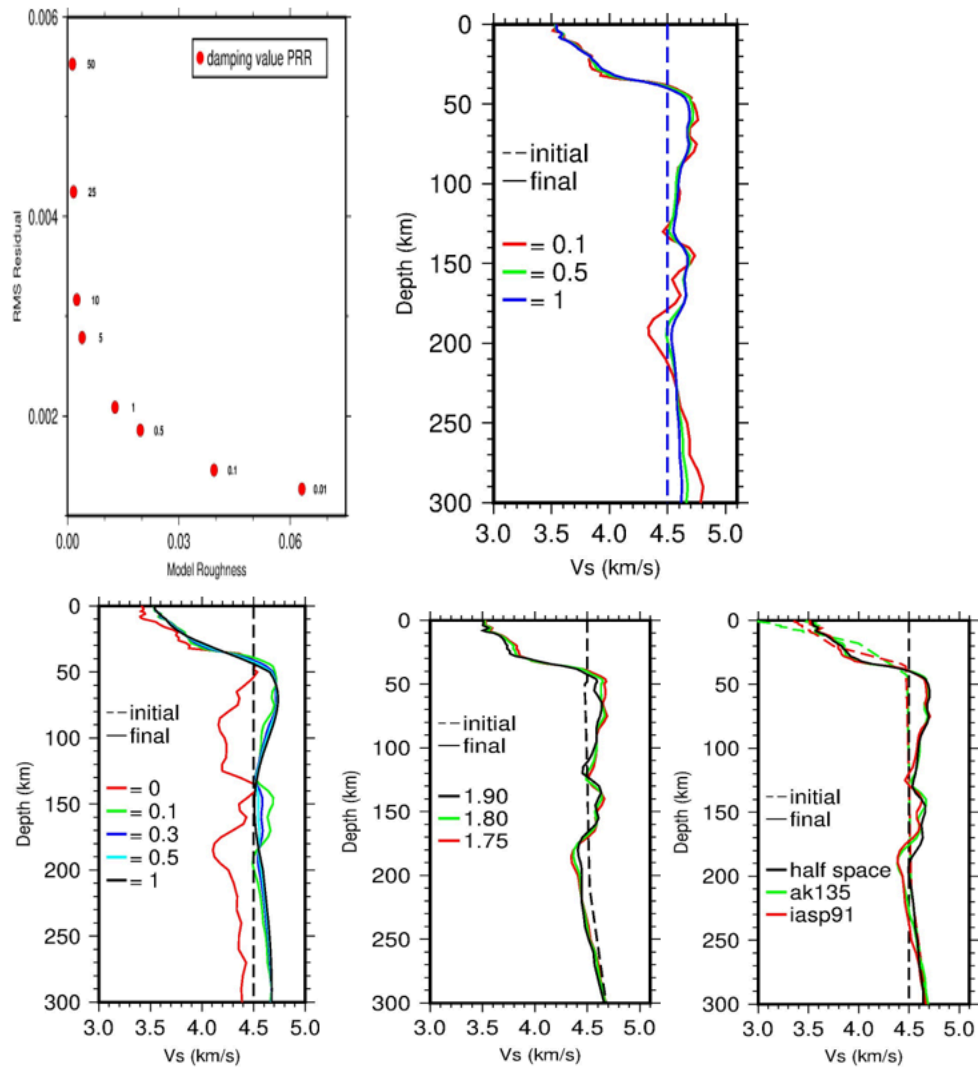


Fig. 2.23. Determination of optimal parameters for joint inversion of RF and SWD. **(Top left)** L-curve for optimal selection of damping parameter, **(Top right)** influence of damping parameter on the inversion, **(Bottom left)** effect of relative weighting parameter between RF and SWD; $p =$

0 is RF only and $p = 1$ is SWD only, **(Bottom middle)** effect of V_p/V_s on inversion, and **(Bottom right)** effect of using different initial velocity models on inversion.

After selection of optimum weight (p) and damping parameter (θ), dependence of jointly inverted final model with the initial model is to be checked. Since the V_p/V_s of the initial model is fixed, joint inversion having an initial model of different V_p/V_s (1.73-1.80) is done. Also different global velocity models (AK135 and IASP91) as initial models are taken for the inversion. Finally no significant variation of the final velocity model with different initial models is found. For simplicity, we have chosen the half space ($V_s = 4.5$ km/s) velocity model as the initial model for joint inversion and analysis finally.

For the joint inversion having layer thicknesses we used a half space ($V_s = 4.5$ km/s) velocity model with thickness: 1 km for 1-10 km depth, 2 km for 10-50 km depth, 5 km for 50-200 km depth, 10 km for 200-400 km depth. The RFs were stacked in the narrow epicentral distance-back azimuth range ($\pm 10^\circ$). To include the effect of error and to create bound in the velocity models, $\pm 5\%$ random error is added in each time period of the dispersion creating 100 samples for each station. The joint inversion of stacked RFs and the 100 samples of error incorporated dispersion is done to get 100 samples of final velocity models for each station. The mean of these 100 samples of final velocity models is taken as the representative model having bound for each station. Although the joint inversion is done upto 400 km depth but since the maximum period of dispersion is 140 s having most reliability at around 210 km depth and reliability decreases below it, The inversion model is most reliable to a depth of 210 km corresponding to $1/3$ of the maximum wavelength of dispersion data at period 140 s. So we have taken the velocity model only upto 300 km depth as reliable for final results. The velocity models resulting from joint inversion for the representative station and other stations are shown in Fig. 2.24 and S2.2 respectively, revealing a significant negative velocity gradient, peaking at depths of approximately 100–140 km for stations within the EDC and CB. This negative velocity corresponds to a depth associated with multiples (PpPs, PsPs, PpPs) observed at around 20 seconds in the receiver function (RF), attributed to signals reflected from the Moho (Fig.2.25b). To confirm that the modeled low velocity is not an artifact, we conducted a synthetic test using a model that features the same low velocity at depths of 100–130 km beneath a 35 km thick crust with a velocity of 3.7 km/s (Fig. 2.25a). The RF generated for this model displays a positive

phase at 4.5 seconds corresponding to Moho conversion, a negative phase at 11 seconds indicating the low-velocity layer, a positive phase at 14 seconds due to multiples from the Moho (PpPs), and a final negative phase at 21 seconds representing the second multiple from the Moho (Fig. 2.25b). We then performed an inversion using the theoretically computed RF and surface wave dispersion data (Fig. 2.25b, c), treating these as observations with a half-space initial model. The inversion accurately reconstructs the low velocity at depths of 100–130 km (Fig. 2.25a). The resulting velocity models for all other stations along the E-W profile from joint inversion down to 60 km depth and 300 km depths are shown in supplementary Fig. S2.

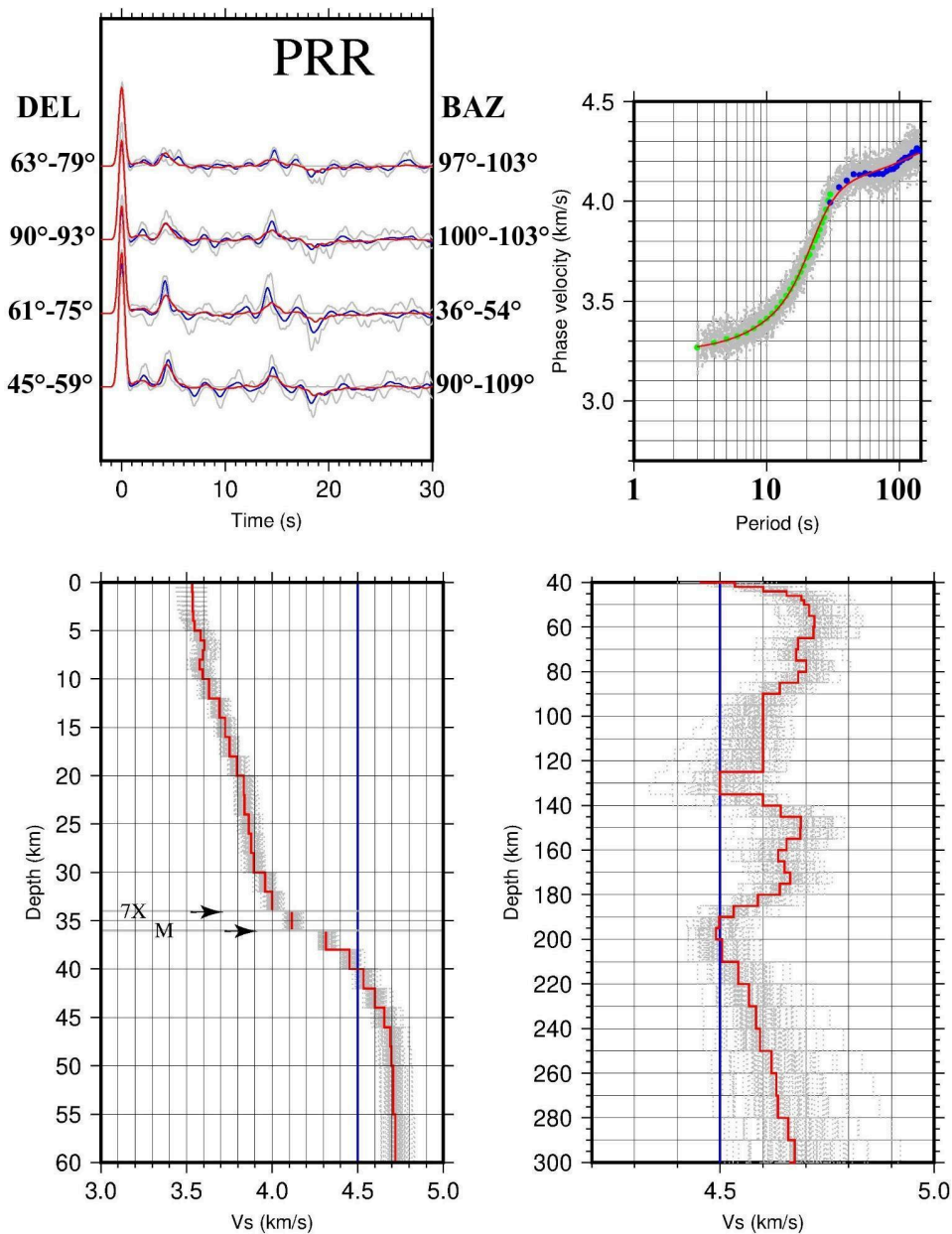


Fig. 2.24. Joint inversion result at station PRR; **(Top left)** upper and lower bound of stacked RFs (grey) along with their mean (blue) and the synthetic RF computed with the inverted model (red), **(Top right)** 100 random Rayleigh wave dispersion curves within 5% error bounds (grey) of observed phase velocity dispersion curve (green and blue) and the synthetic dispersion curve computed with the inverted model (red), **(Bottom left and right)** best 100 solutions (grey) from a set of 300 models (the red line represents the average of the 100 best solutions).

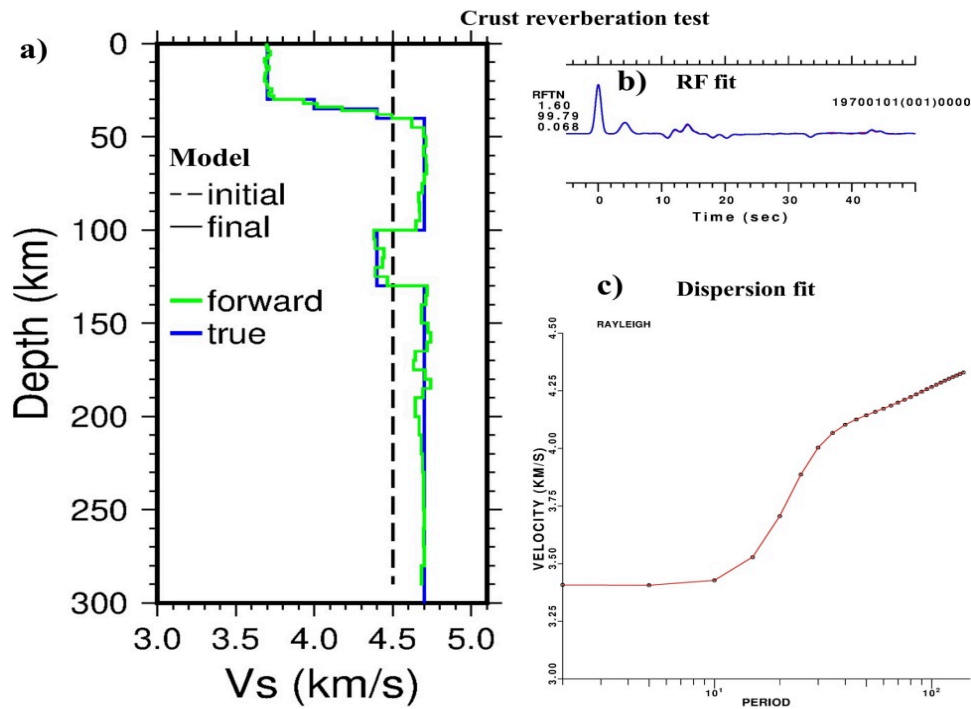
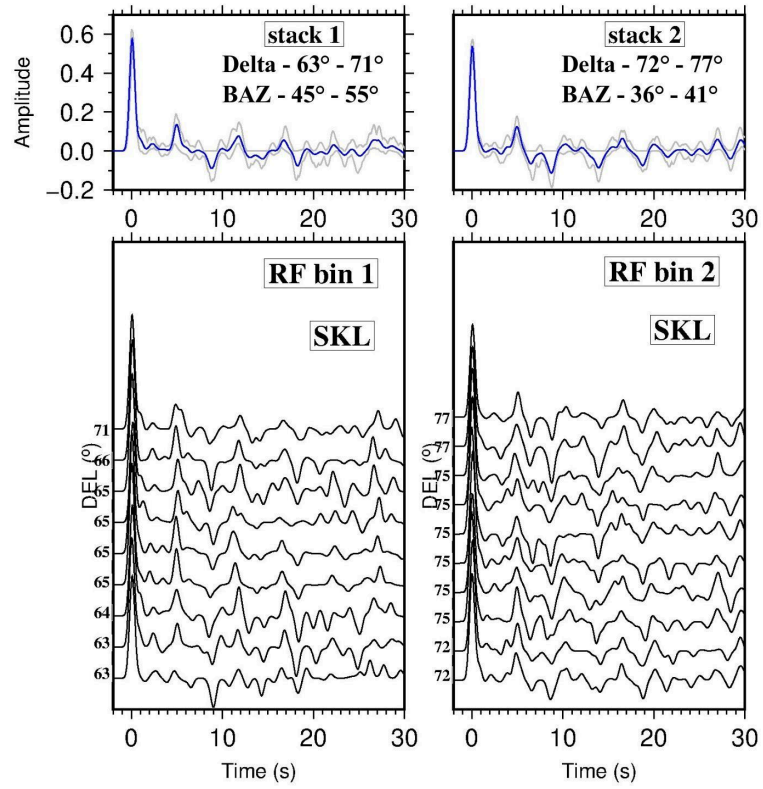
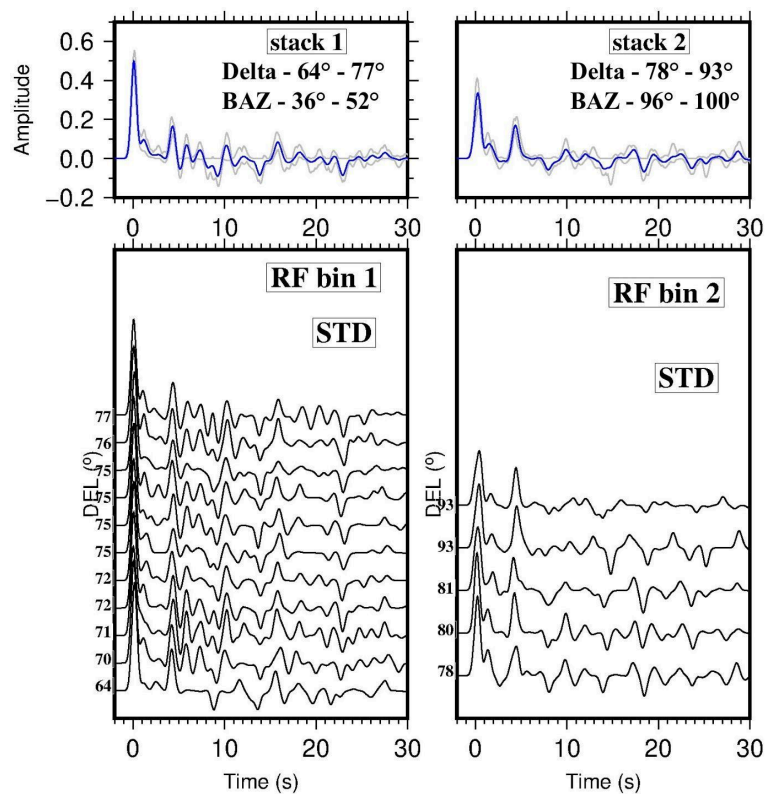
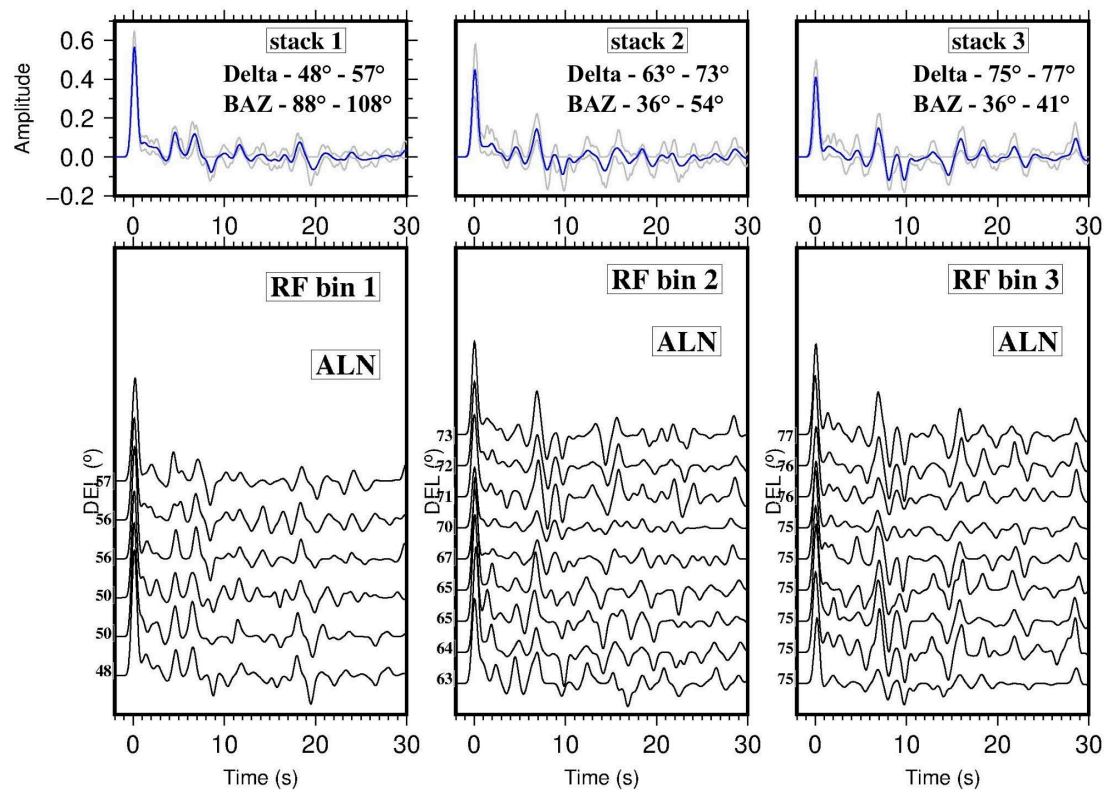
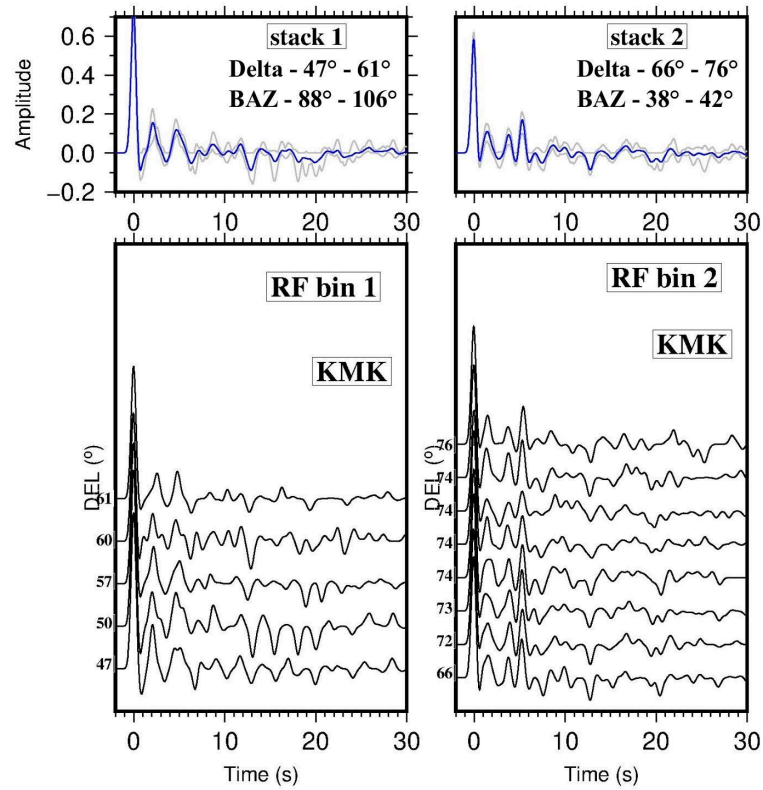
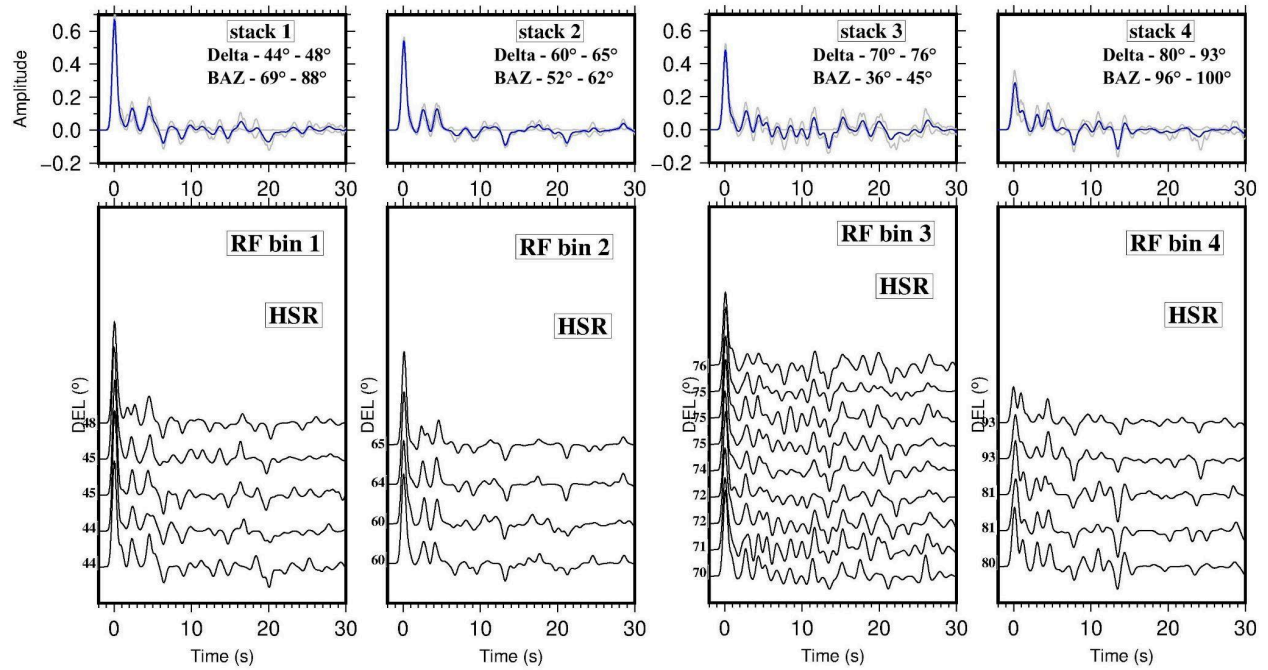


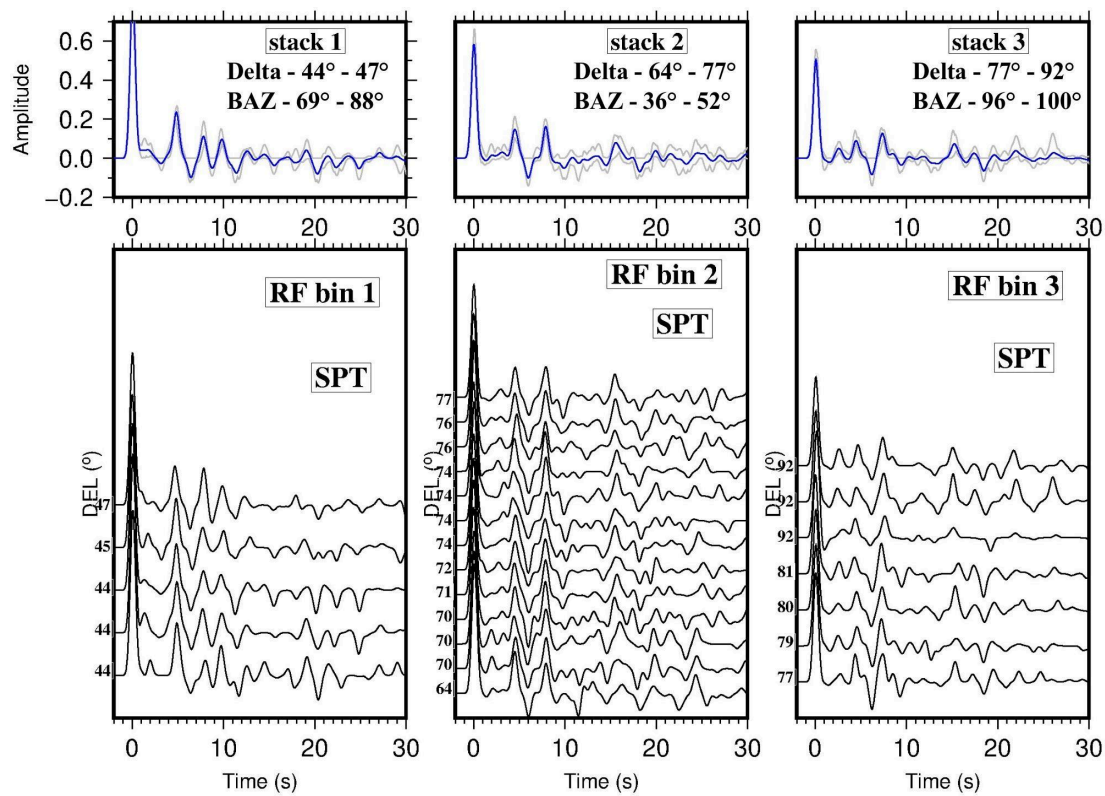
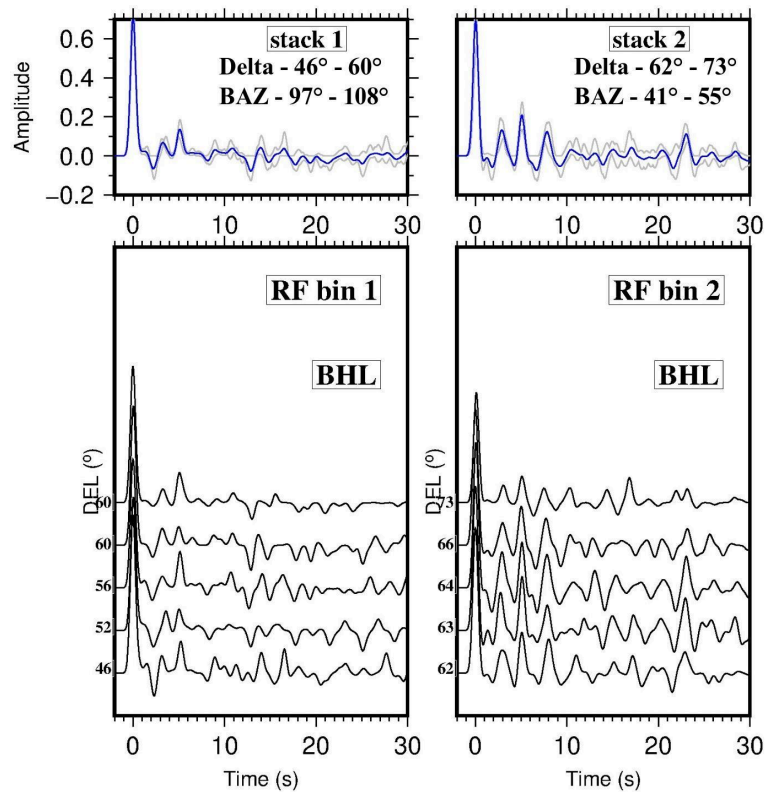
Fig. 2.25. Synthetic test to demonstrate that no significant influence of negative phase in RF (due to Moho multiples) on the inverted velocity model. Theoretical RF **(b)** and dispersion curve **(c)** are created for an Earth model **(a)** with a 35 km thick crust and a low velocity layer at 100-130 km depth. The RF shows multiple with negative phase at 20s. Inversion of RF and dispersion data using a half space model with $V_s = 4.5$ km/s **(a)** correctly reproduces the model with low velocity ($V_s \sim 4.4$ km/s) at 100-130 km **(a)**.

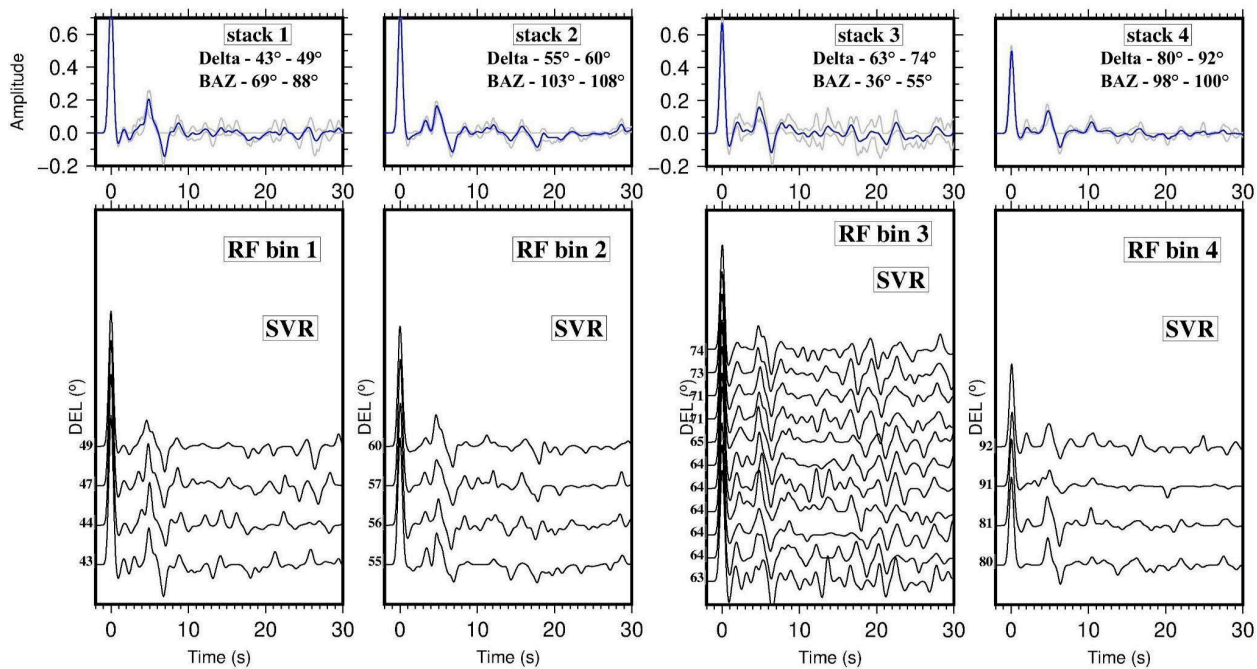
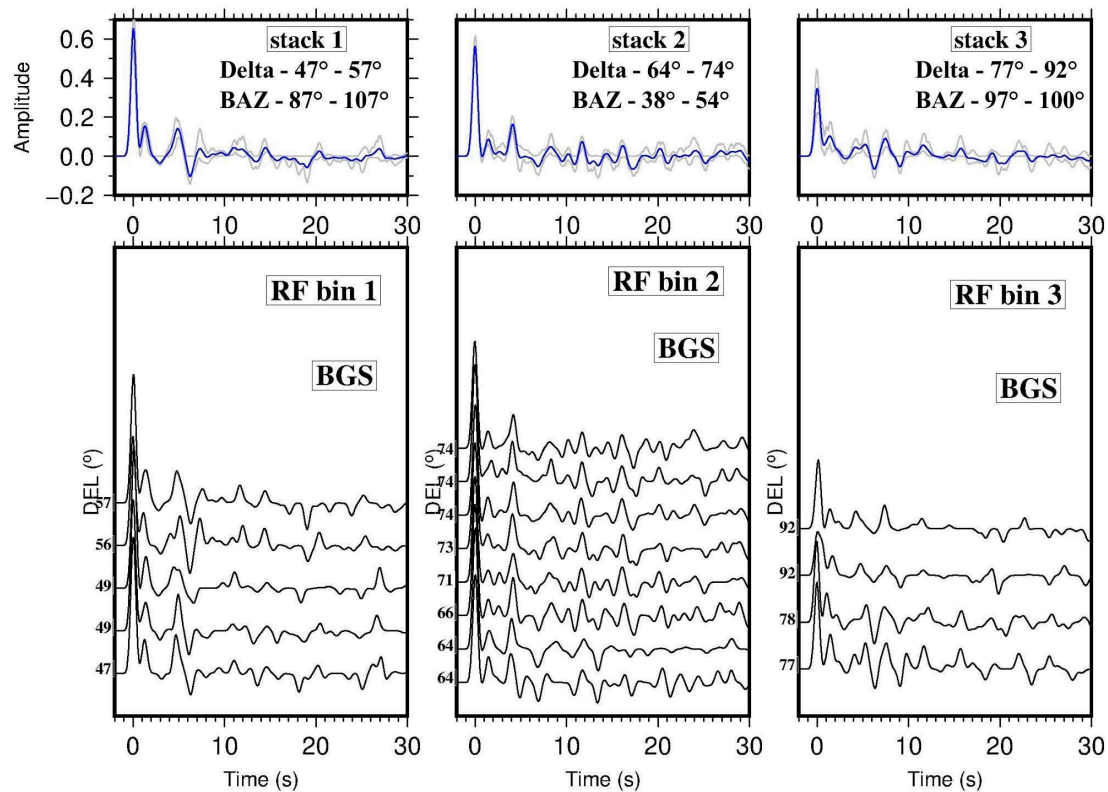
Appendix S2.1 - RF bins and stacks -

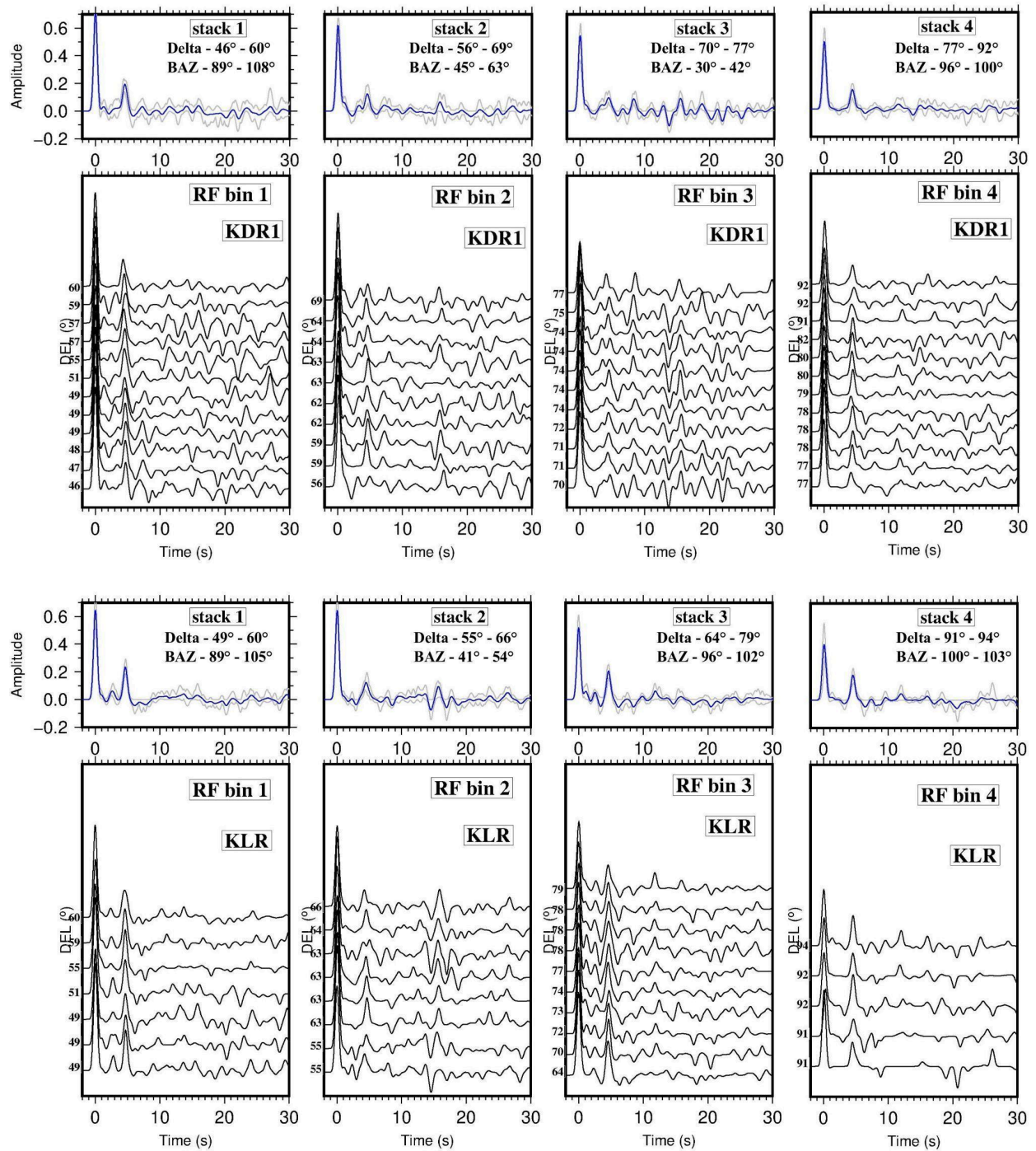


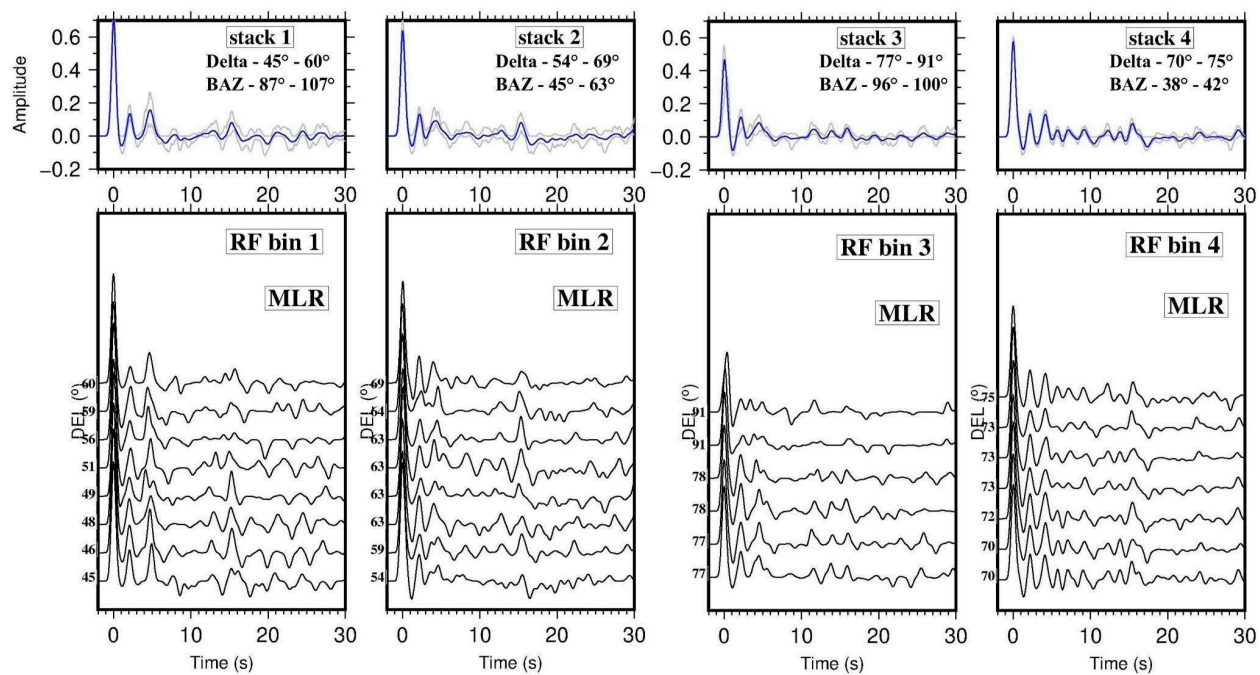
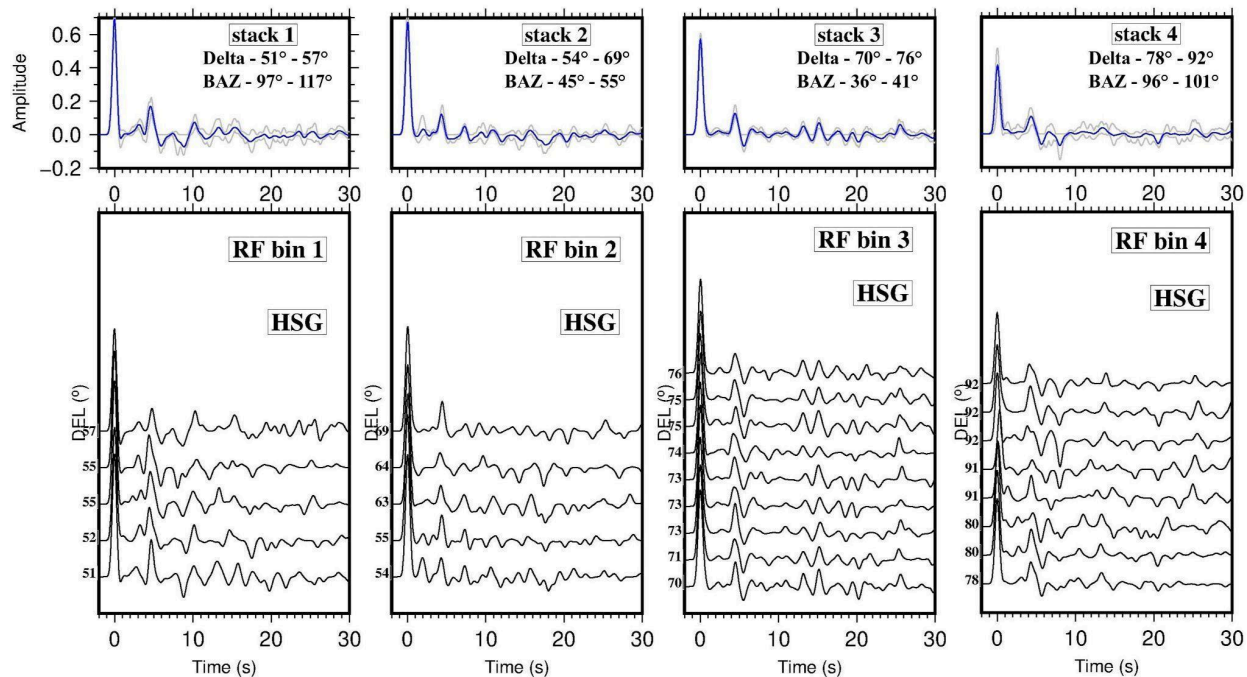


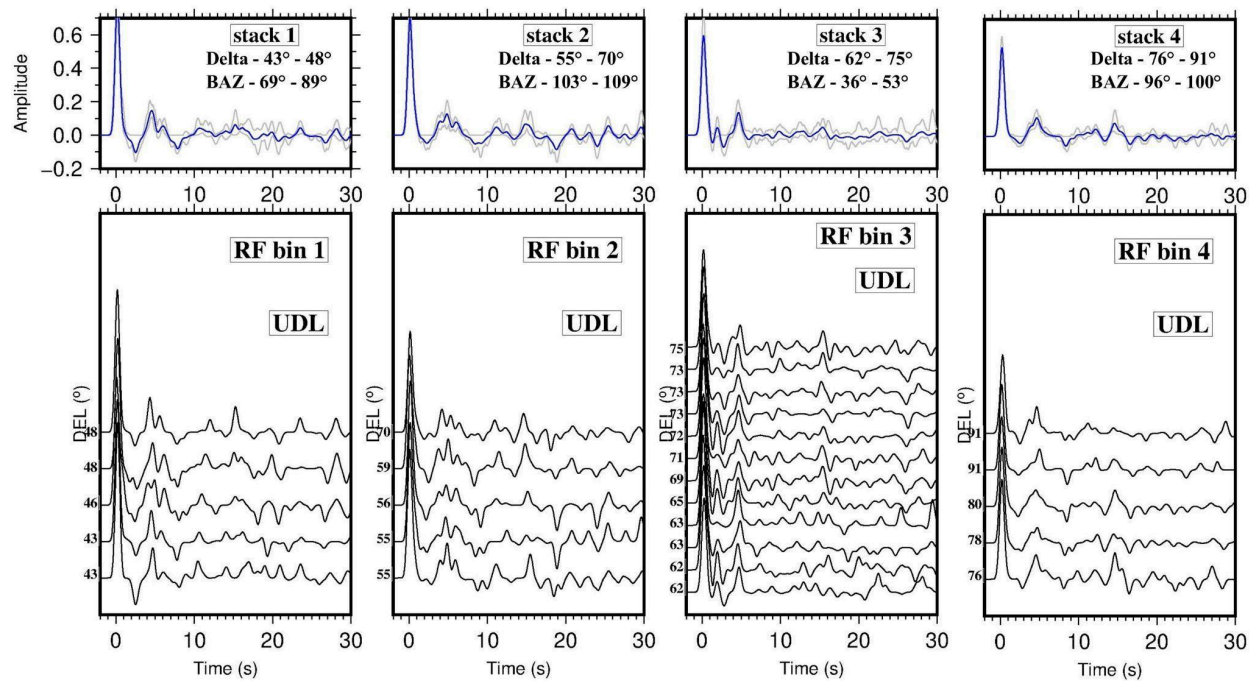


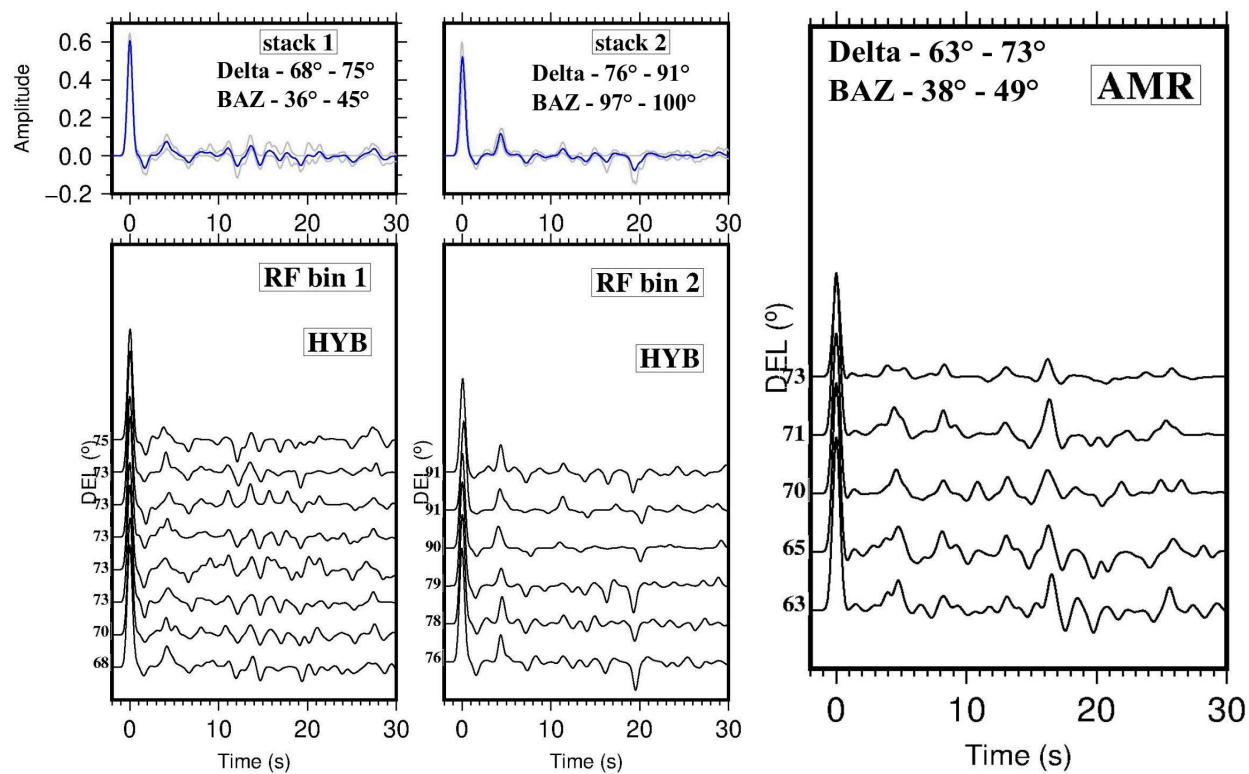
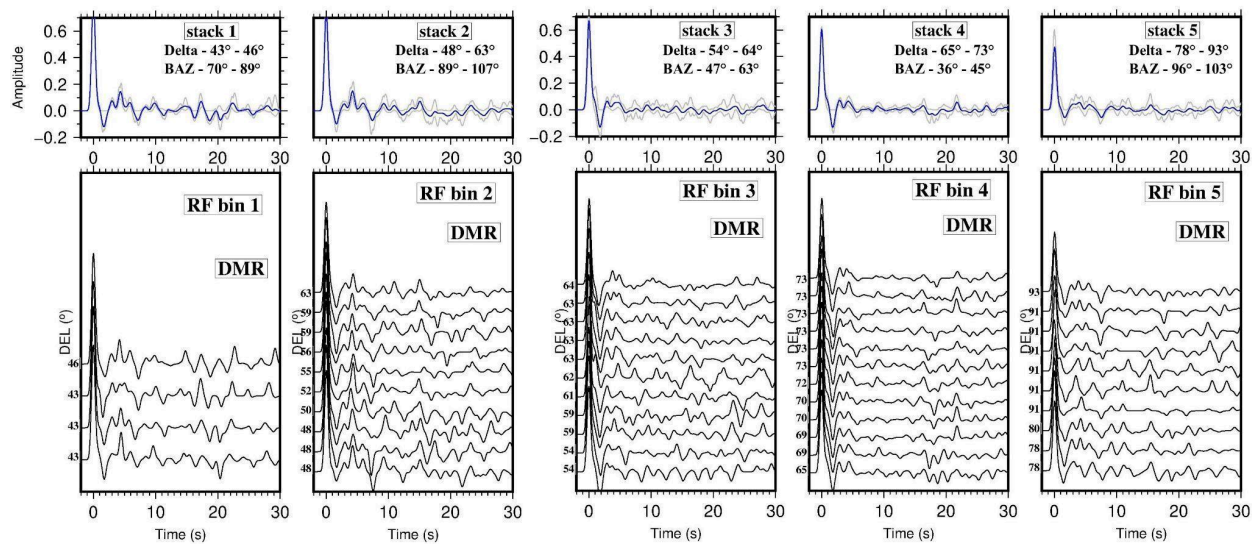


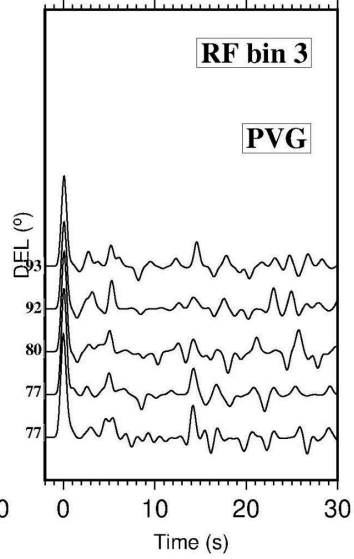
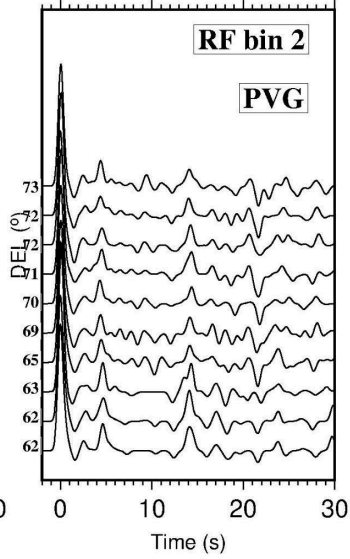
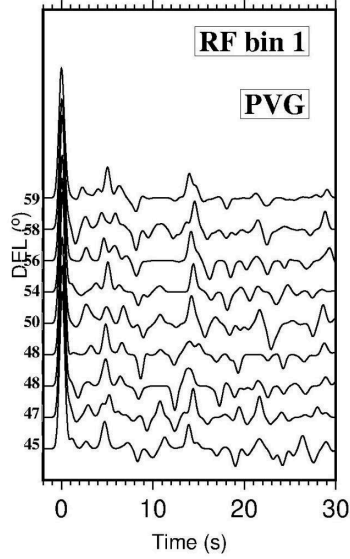
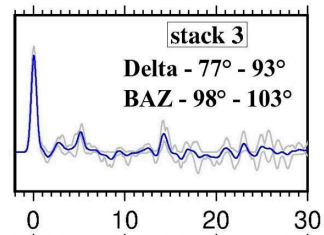
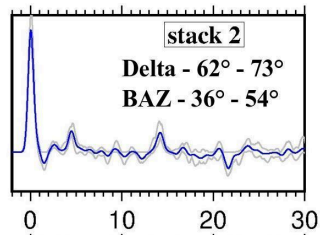
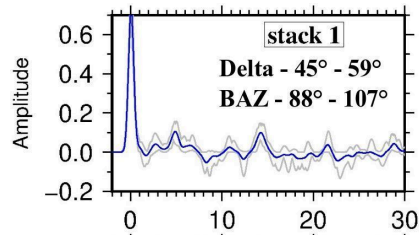
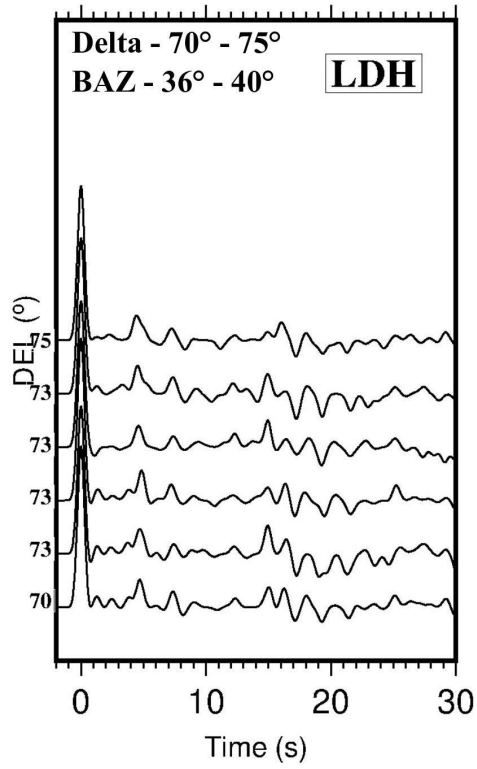


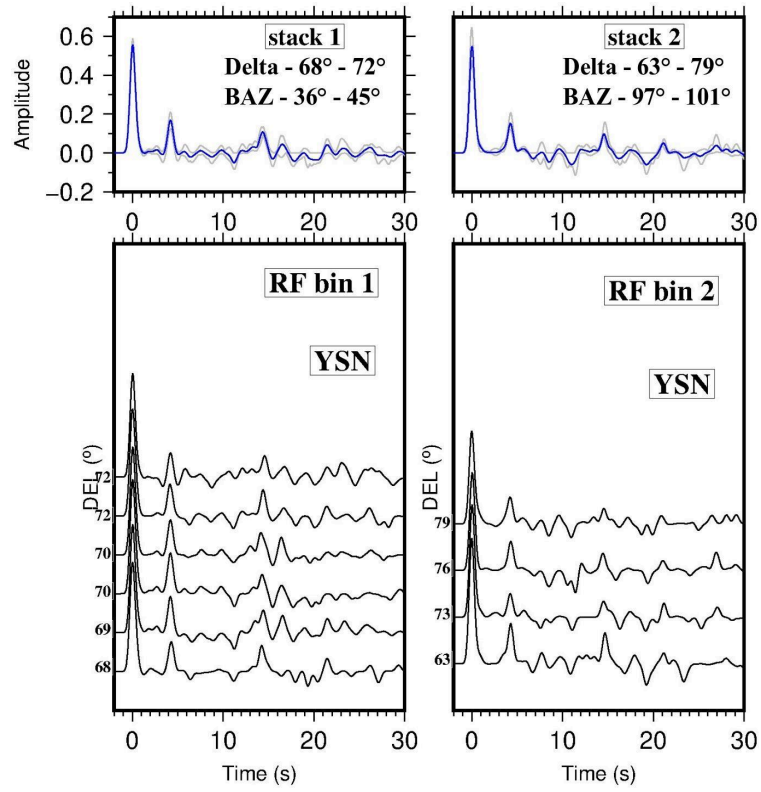


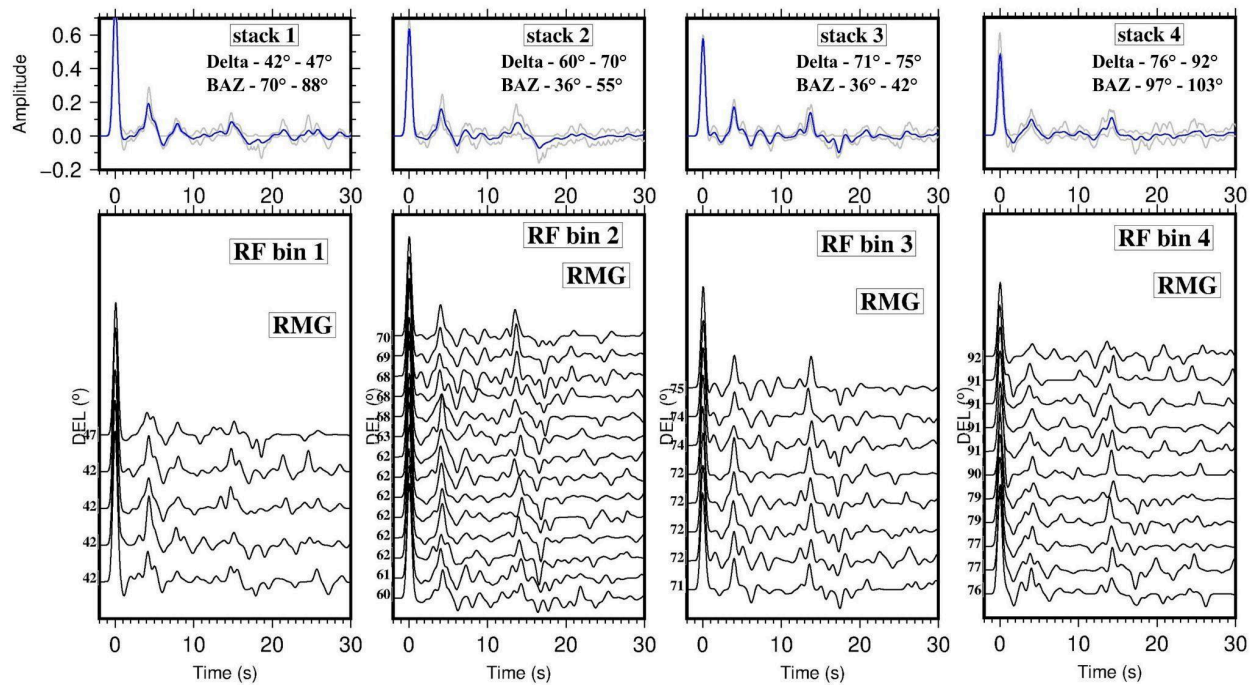
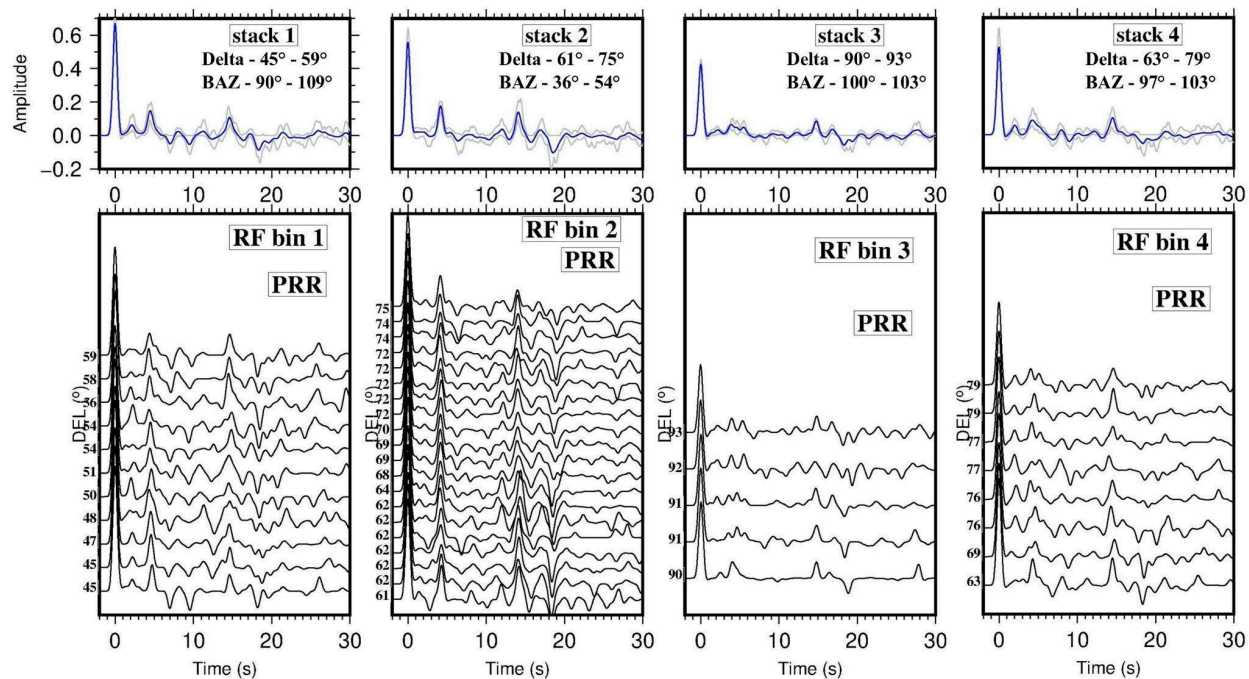


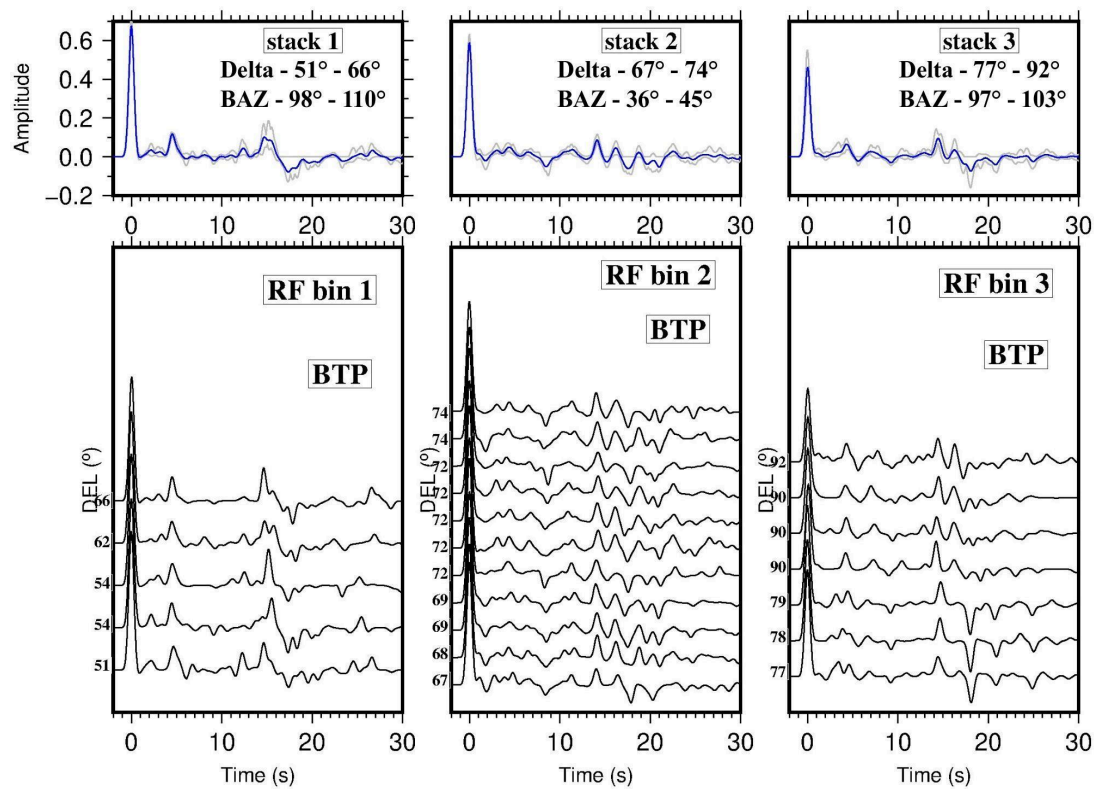
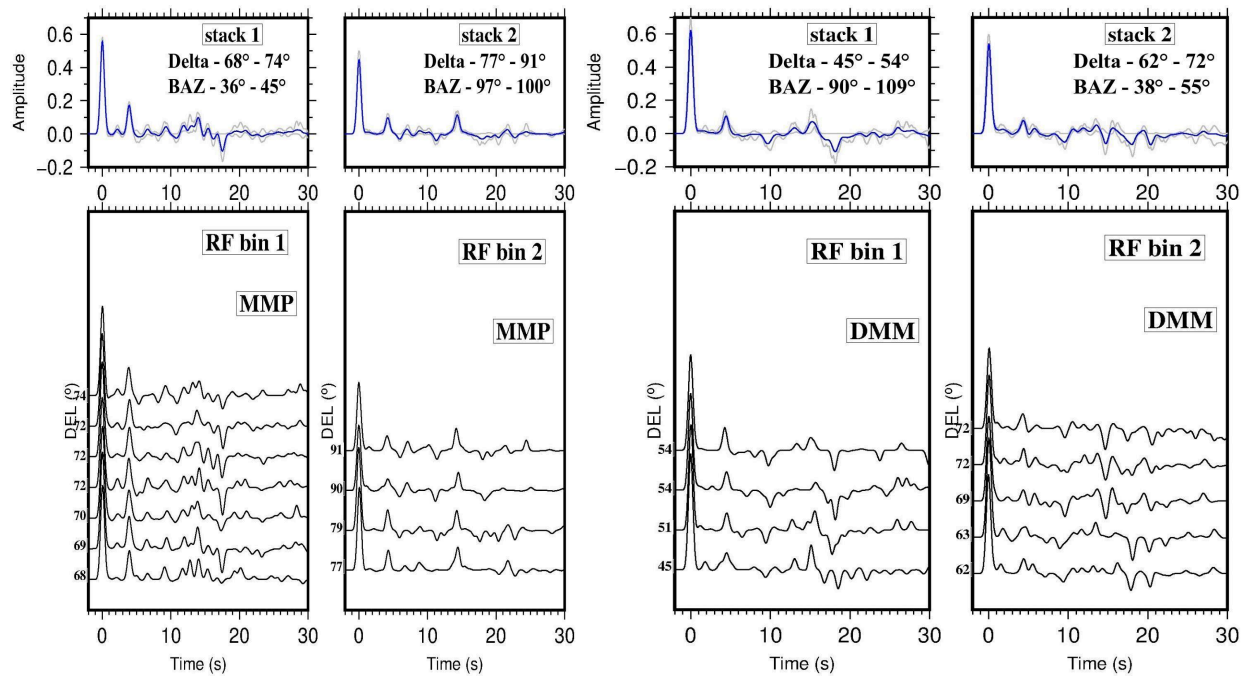


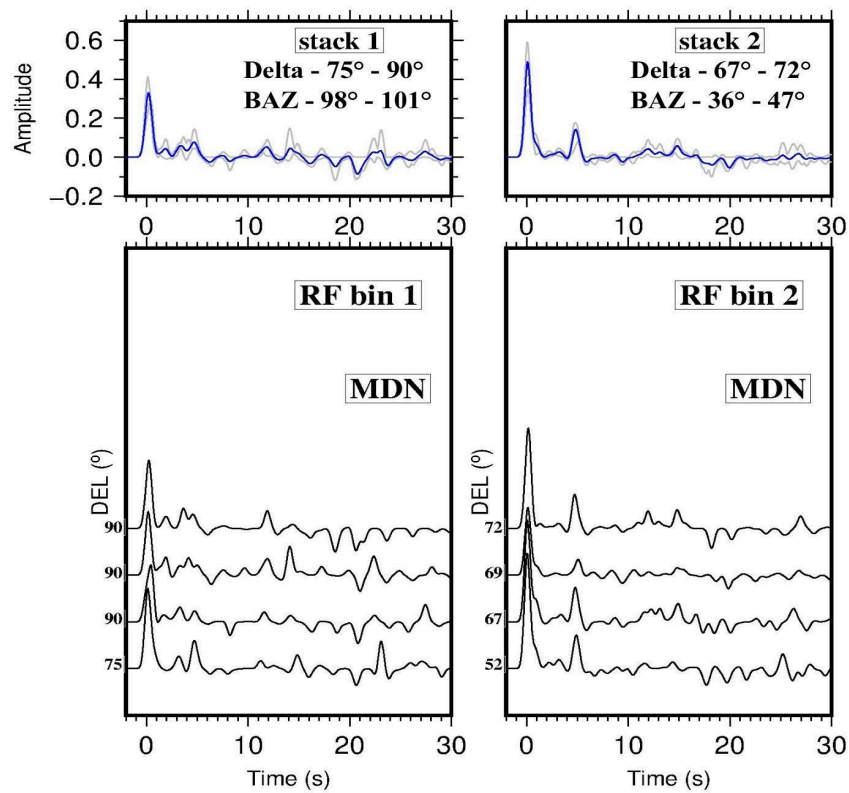
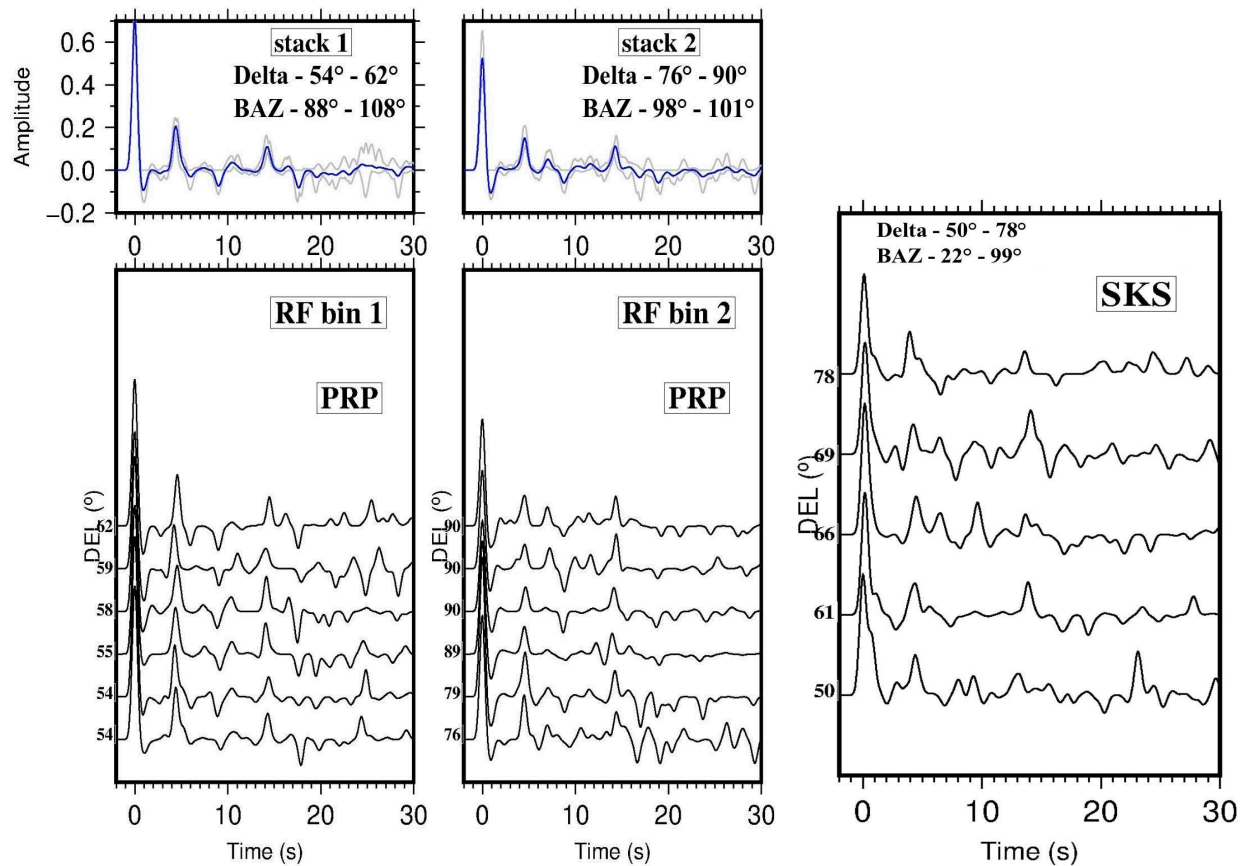


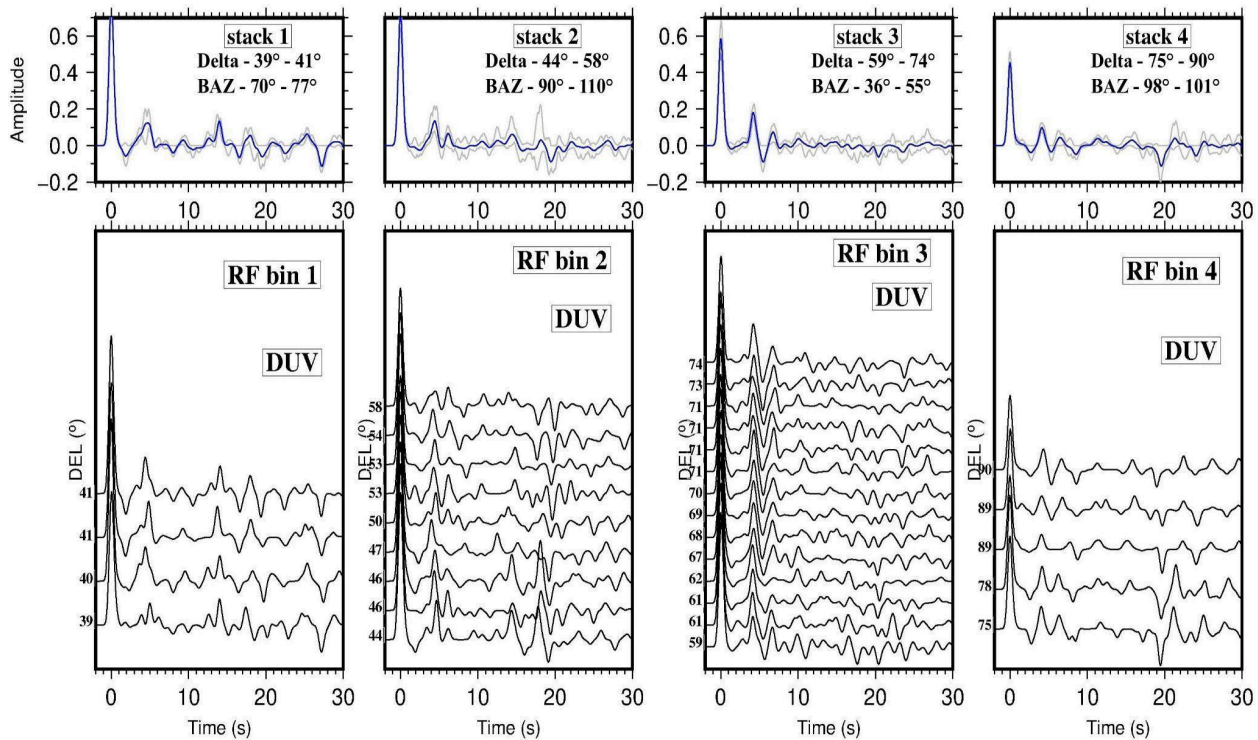
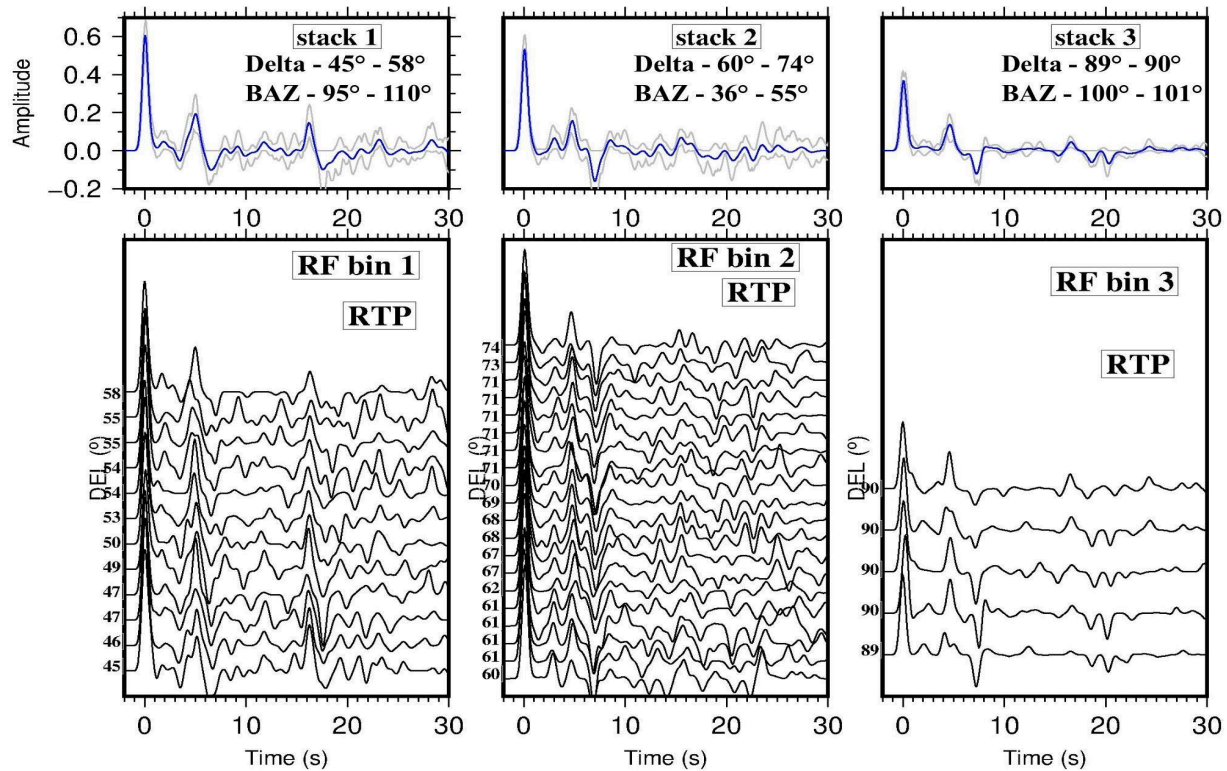


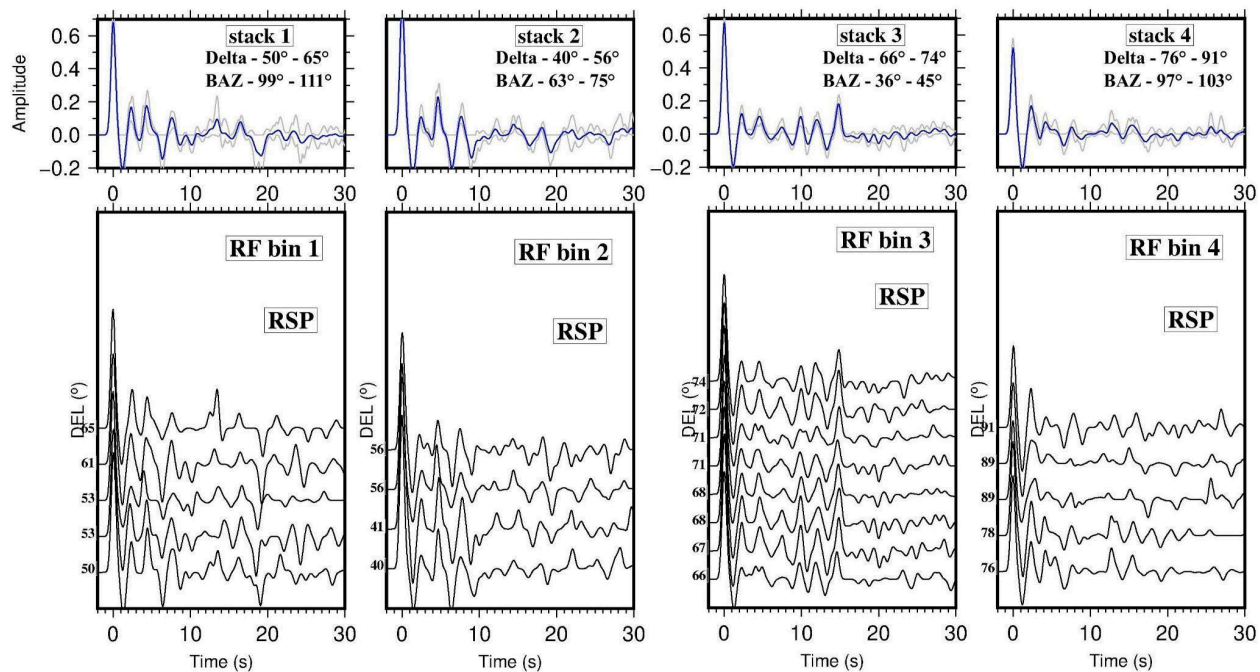
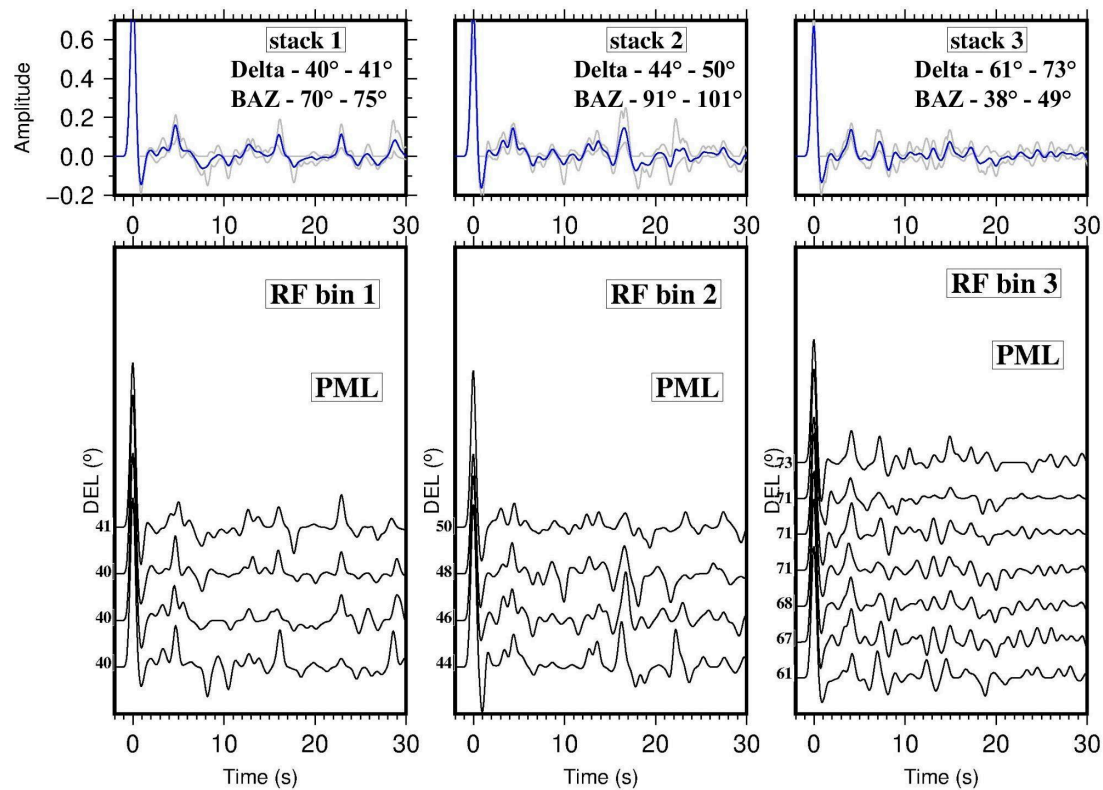


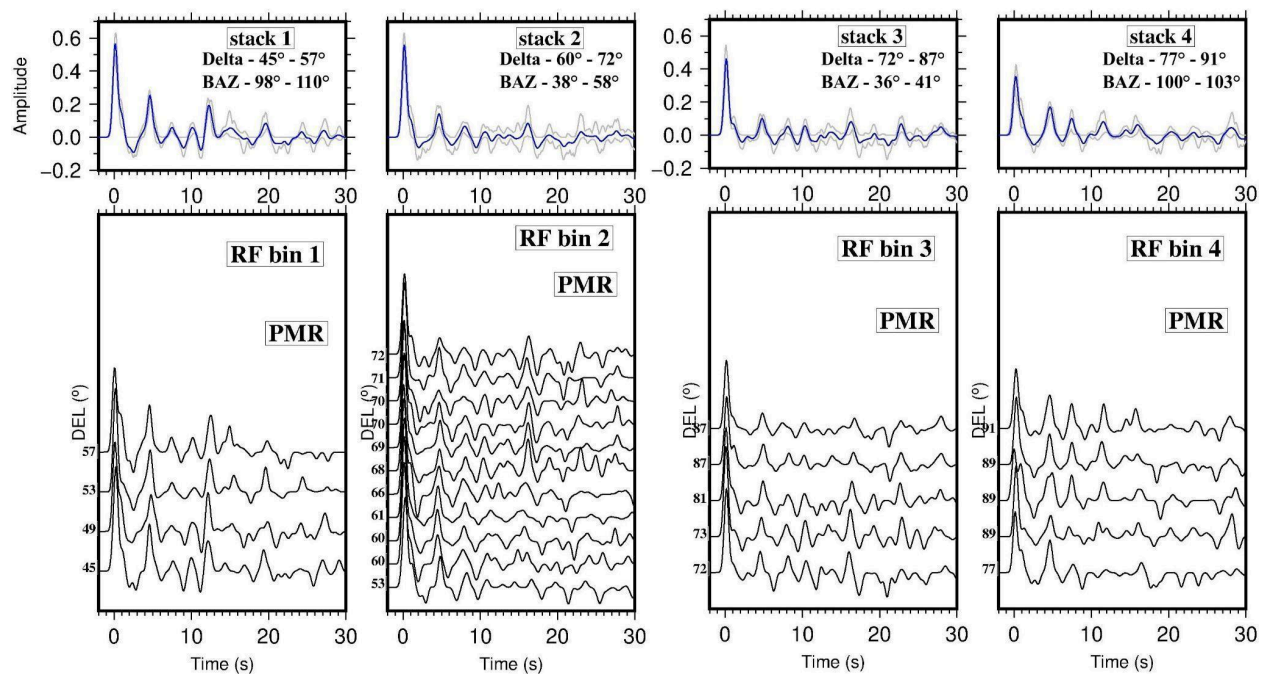
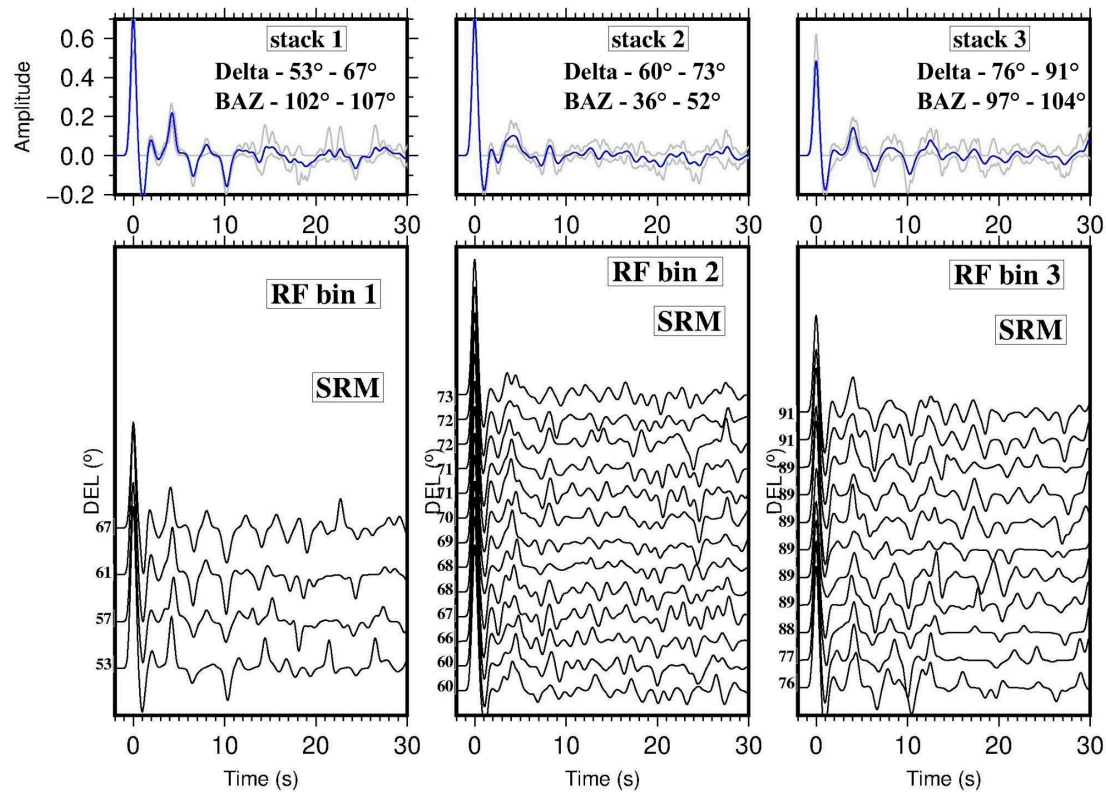


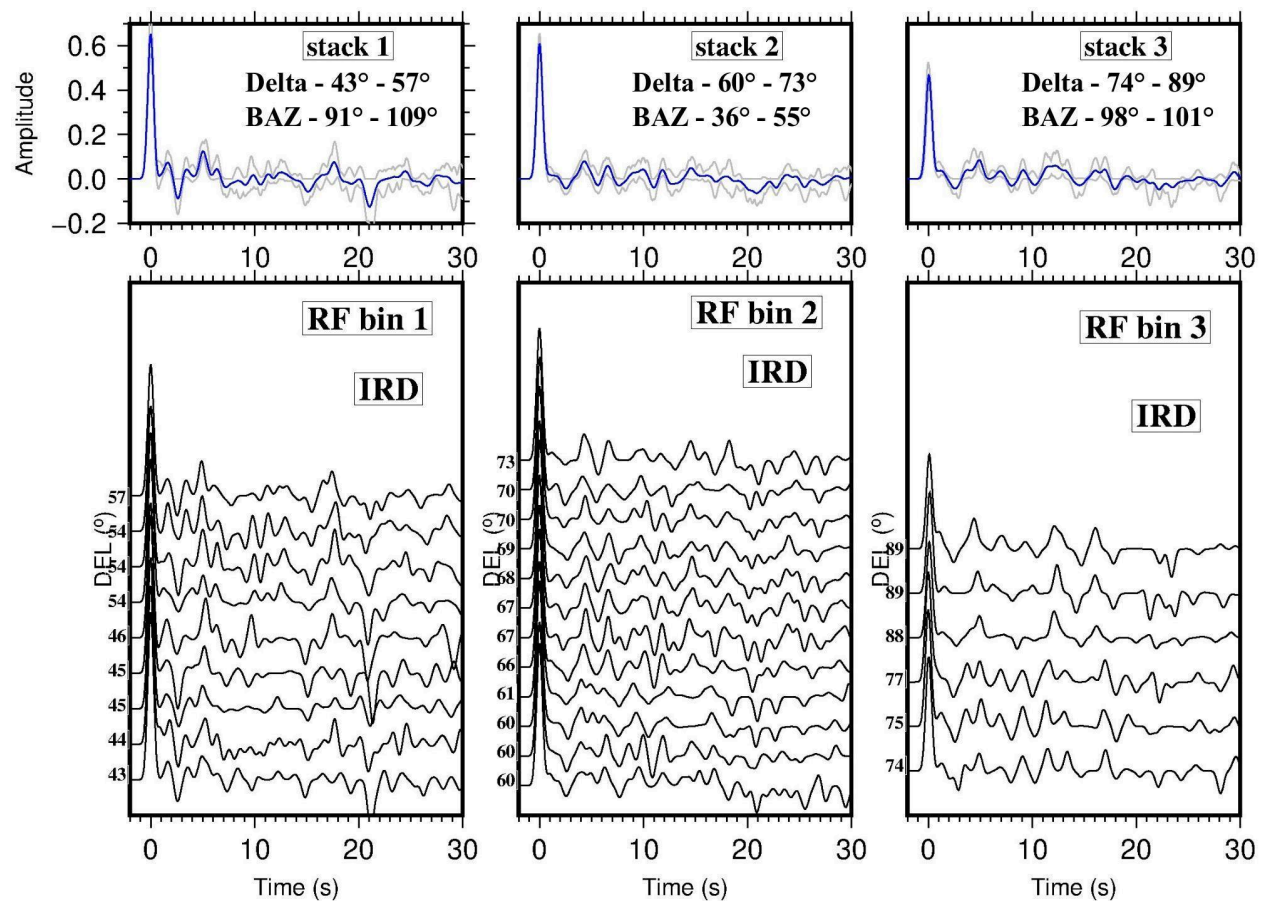
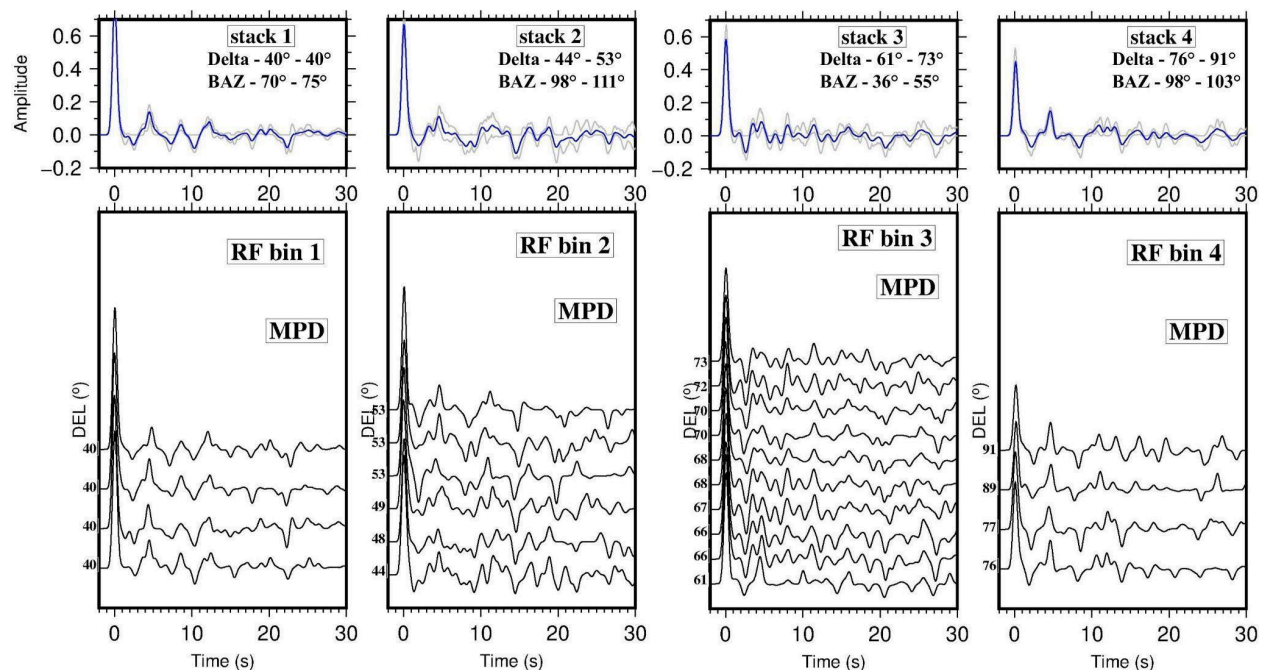












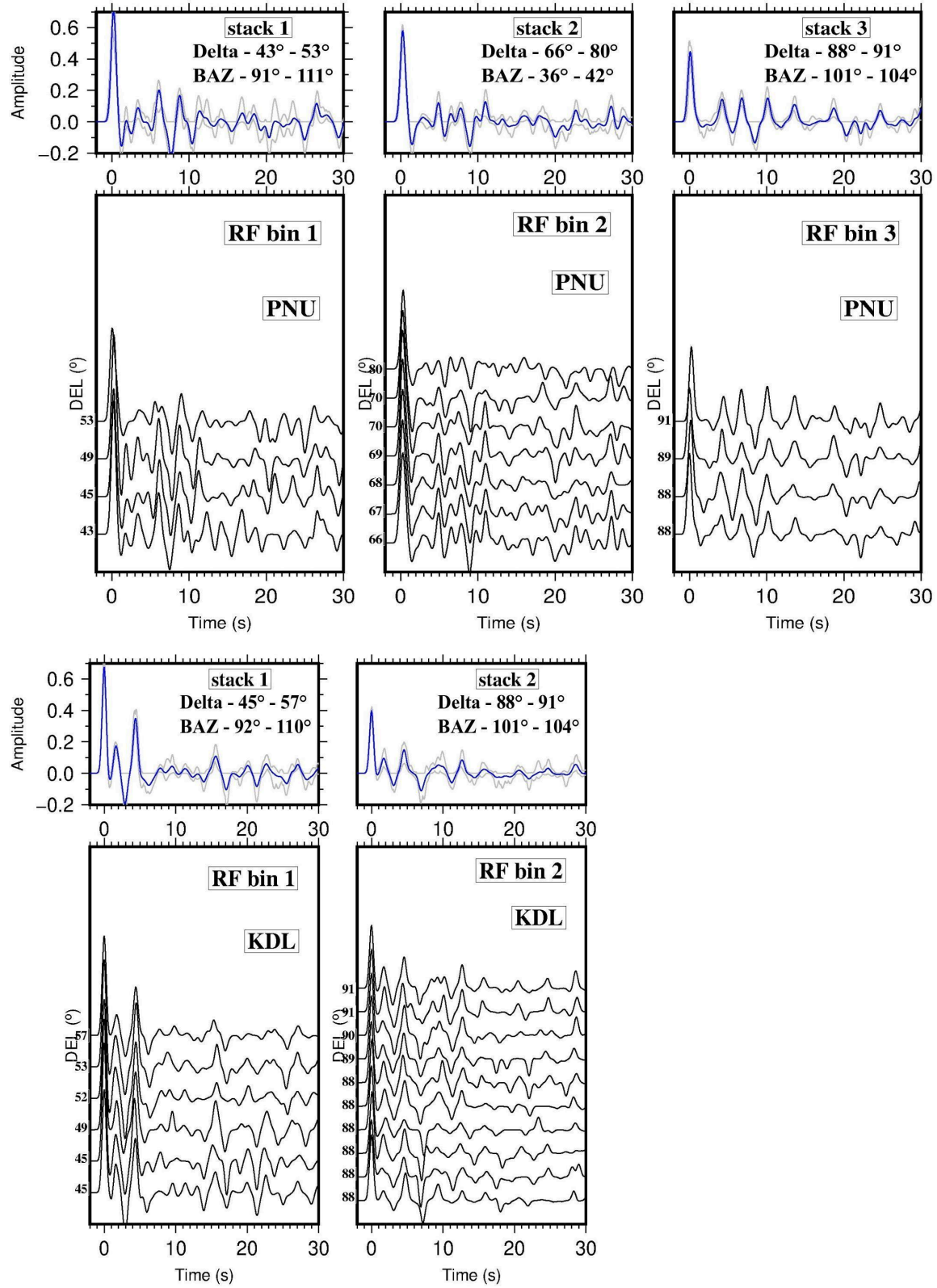
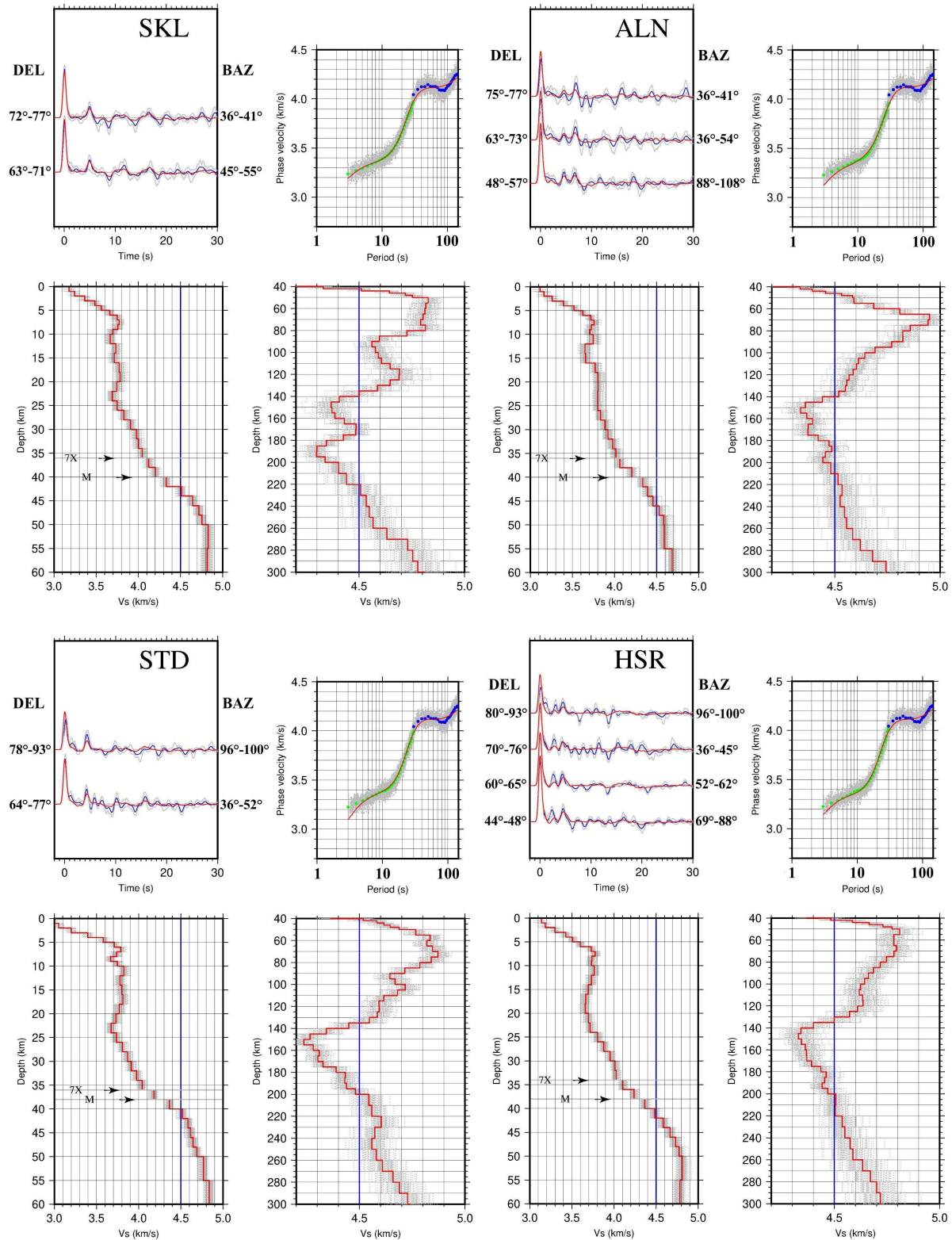
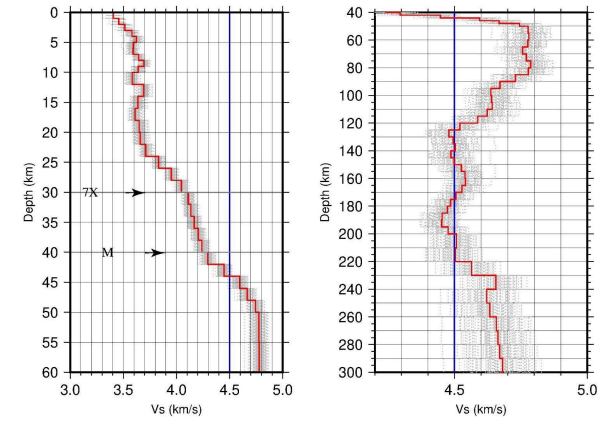
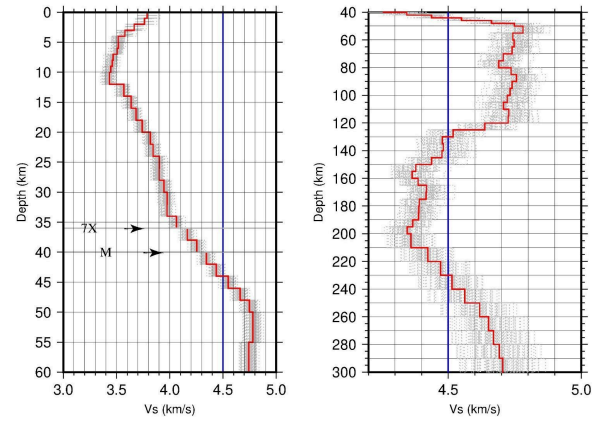
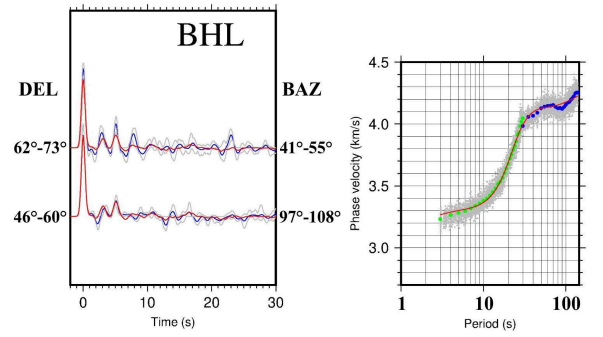
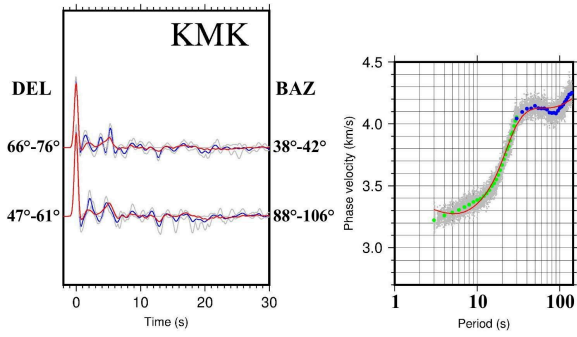
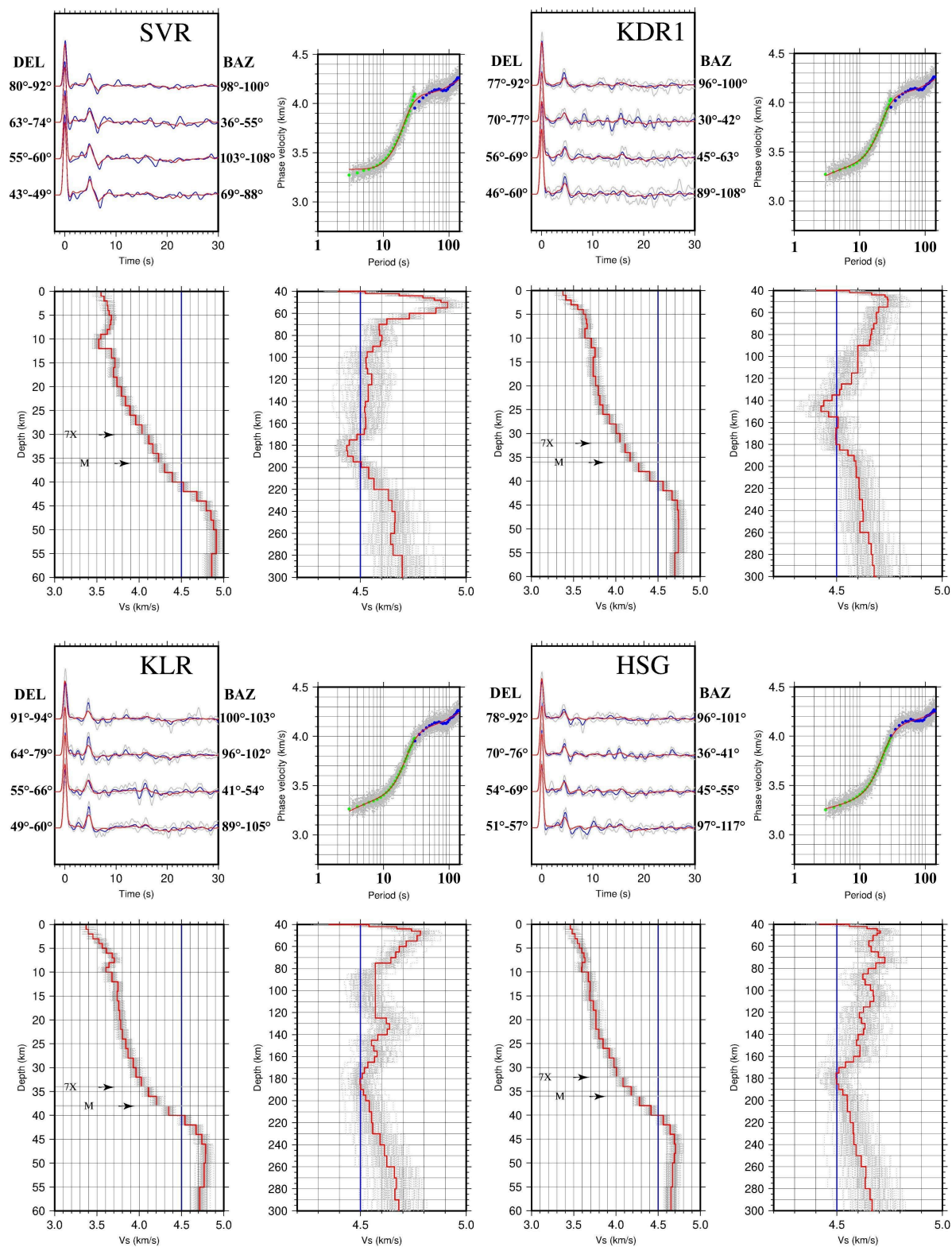


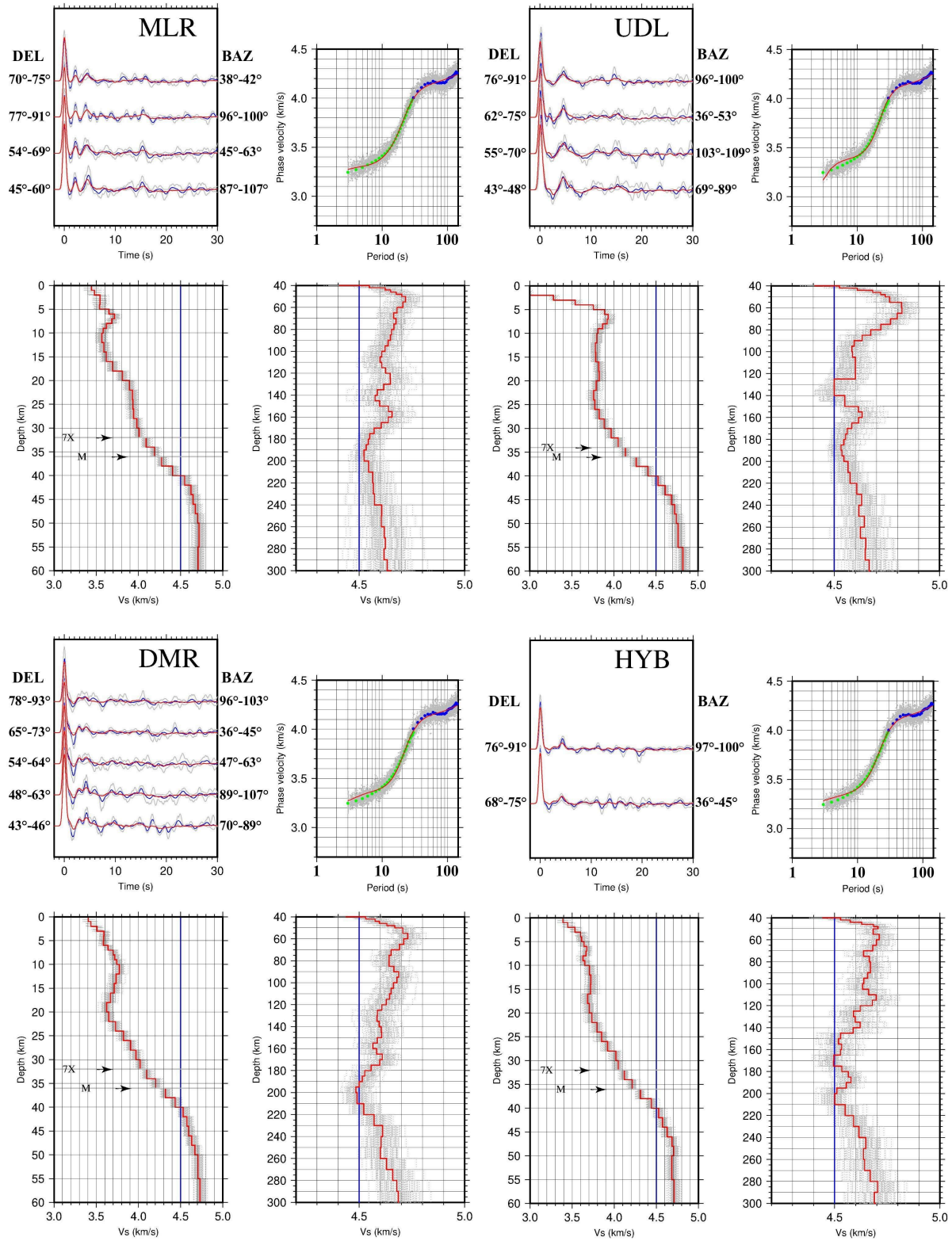
Fig. S2.1. RFs in specific delta, azimuth range and their stack for all the stations along the E-W profile.

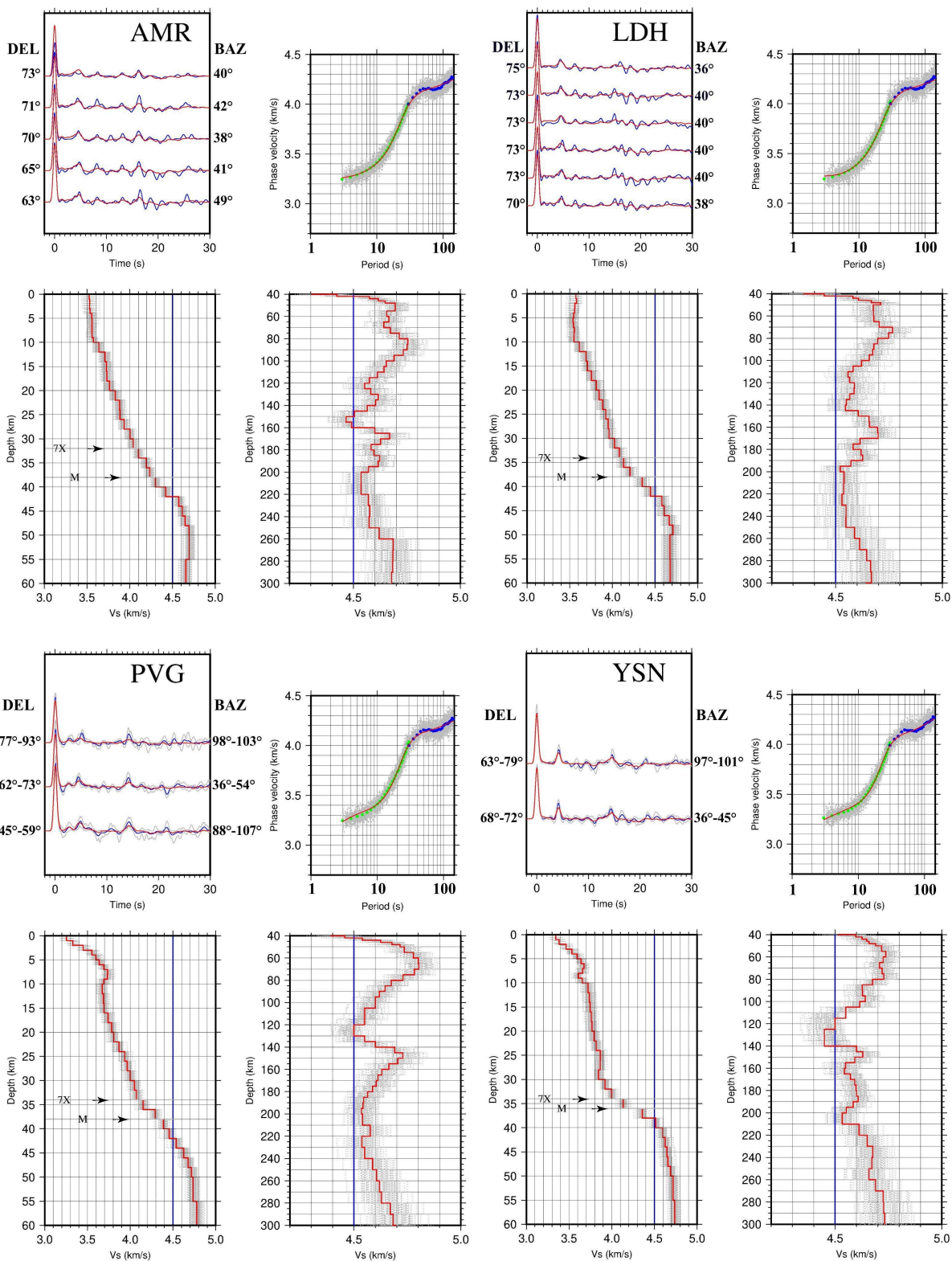
Appendix S2.2 - RF stacks, dispersion and final shear velocity models -

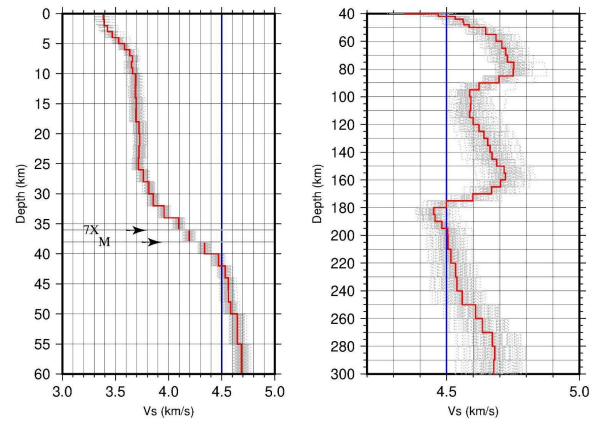
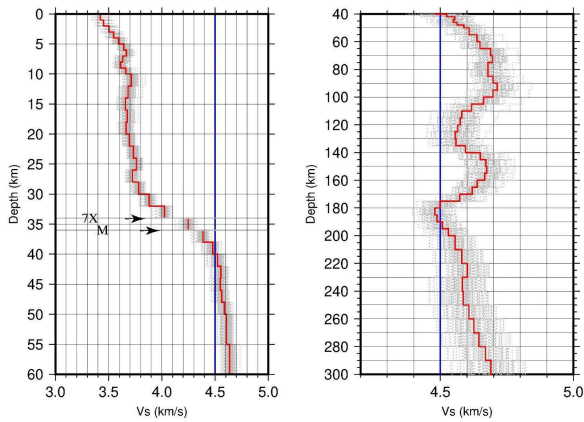
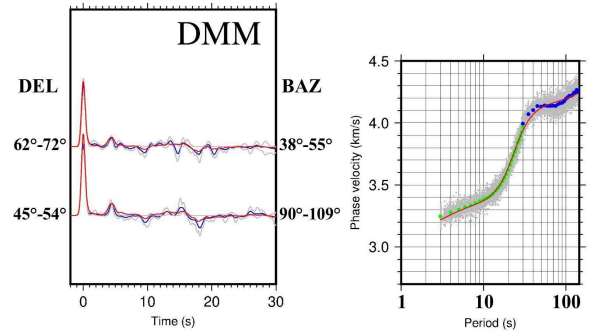
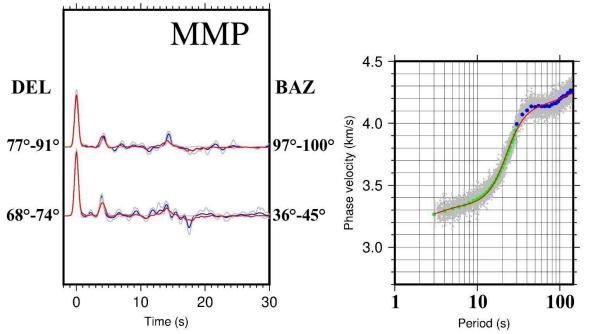
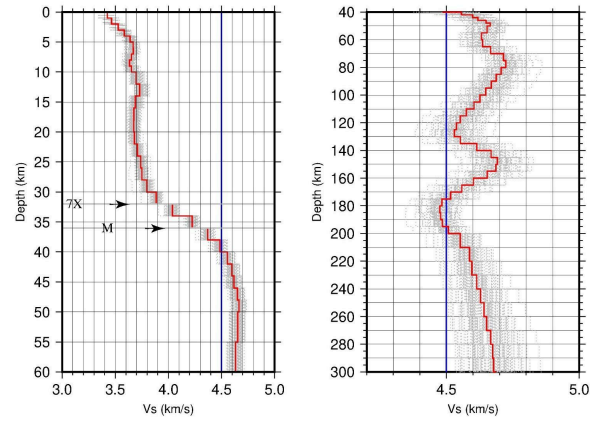
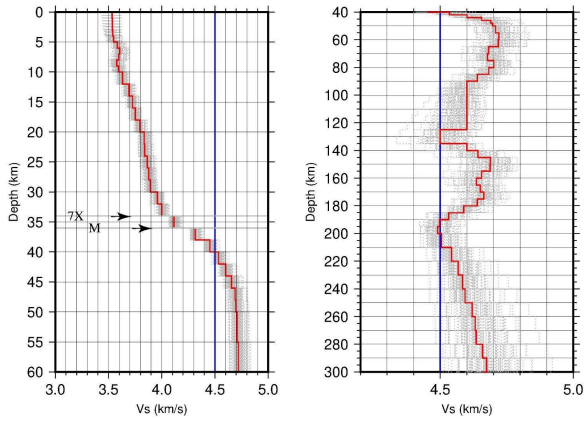
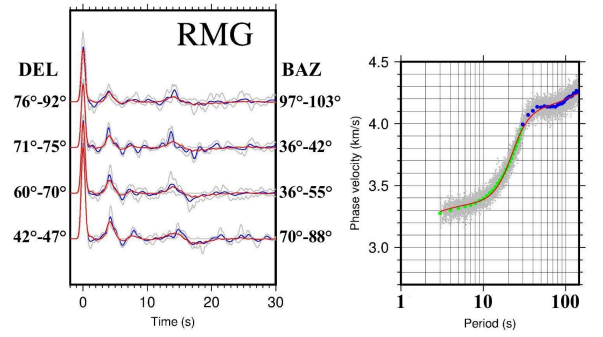
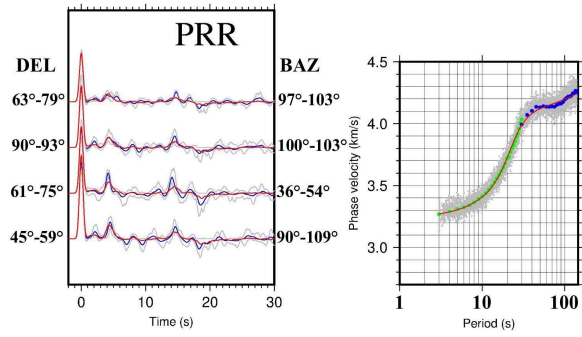


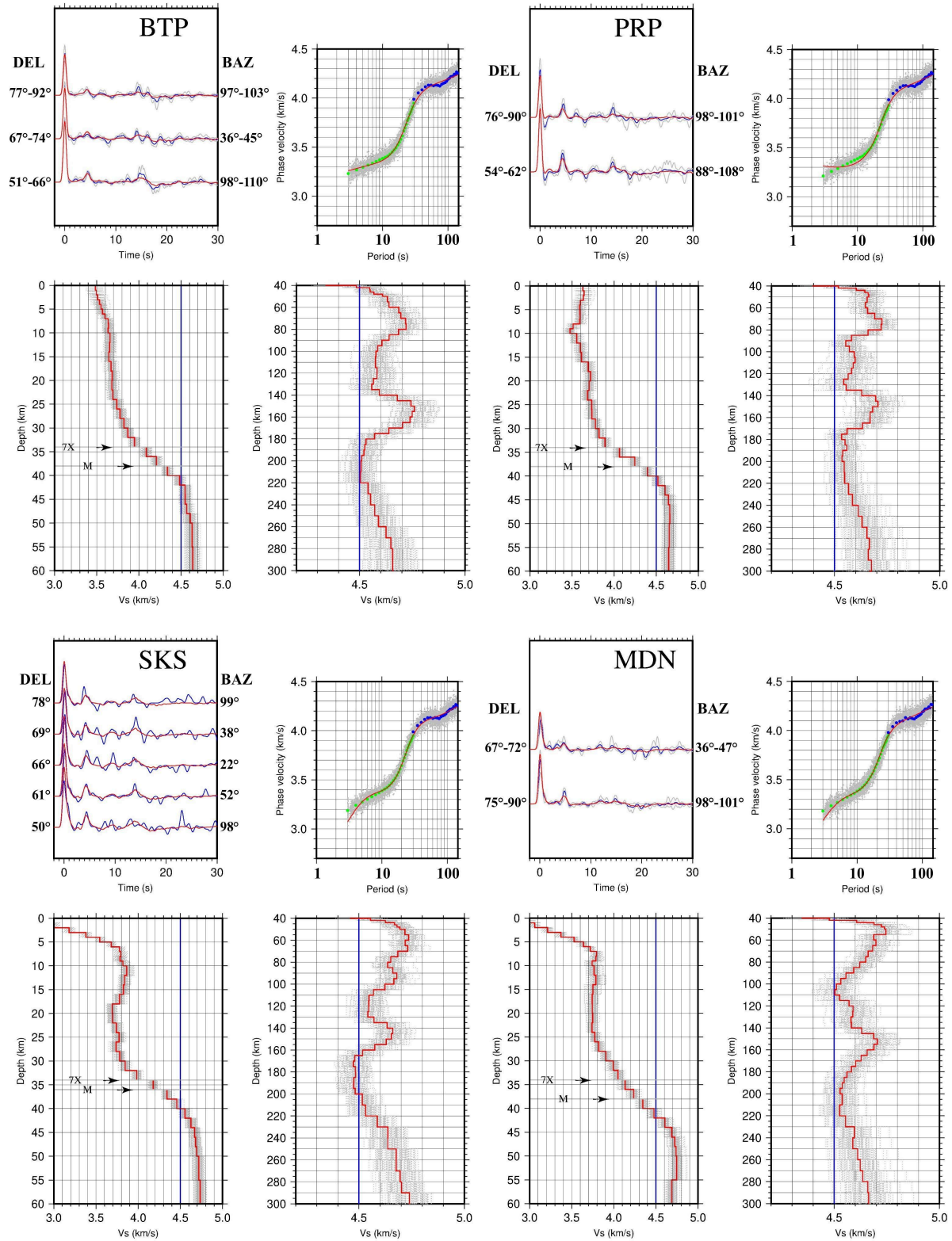


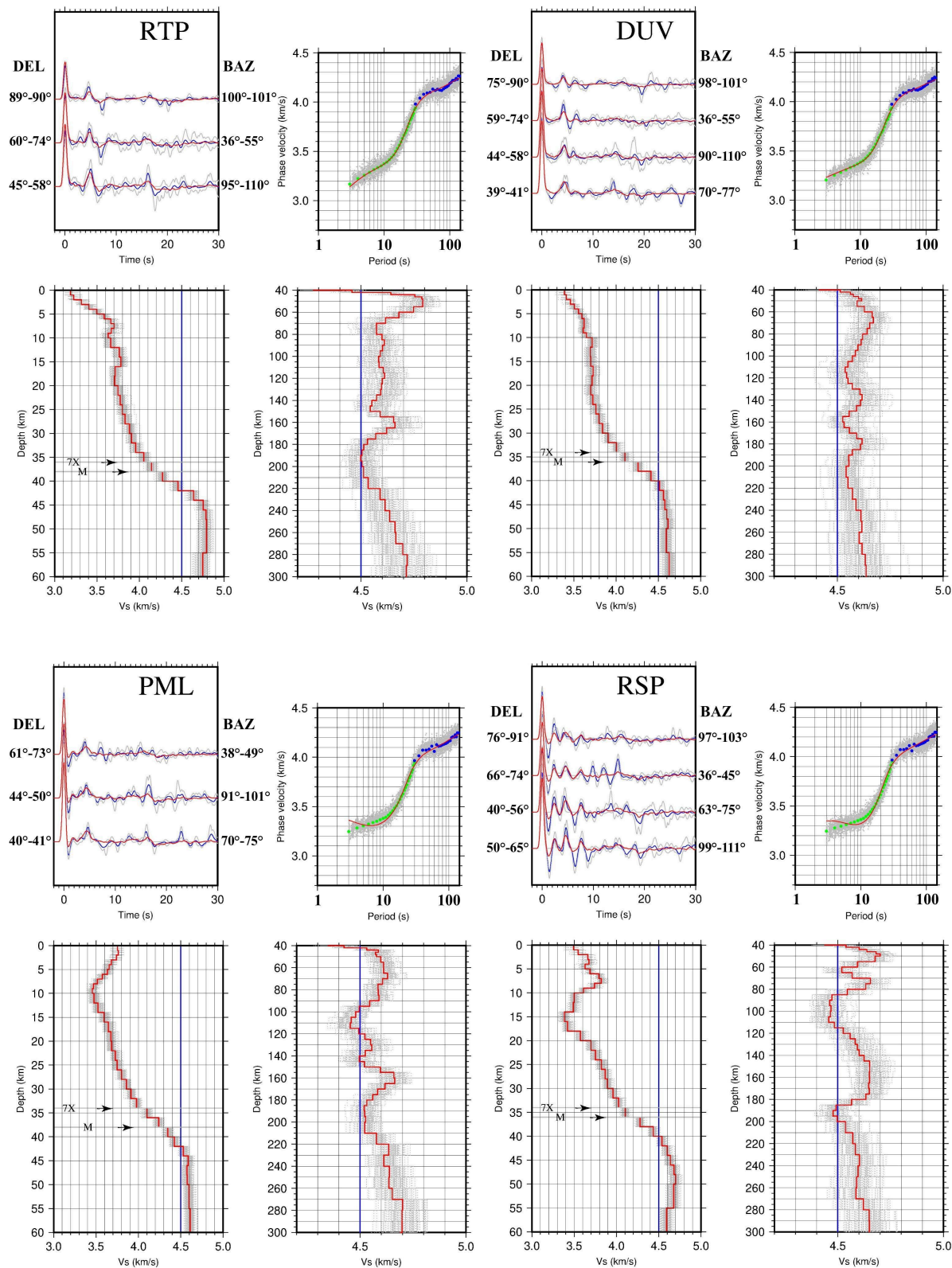


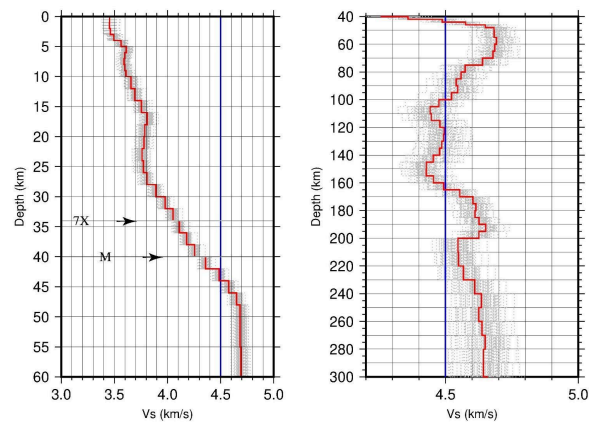
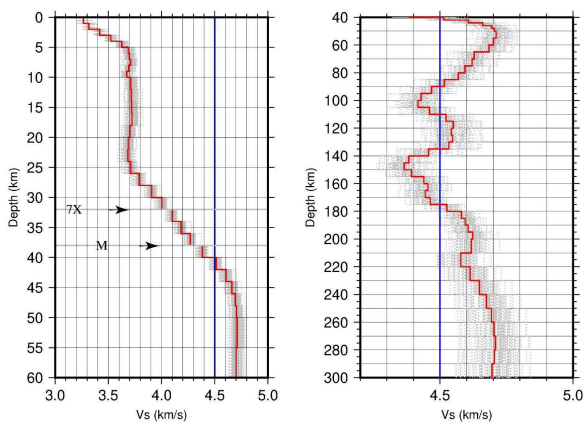
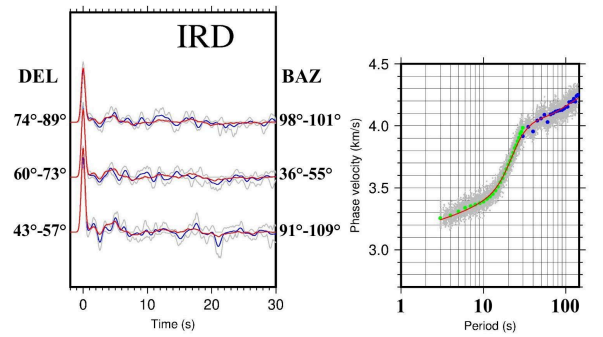
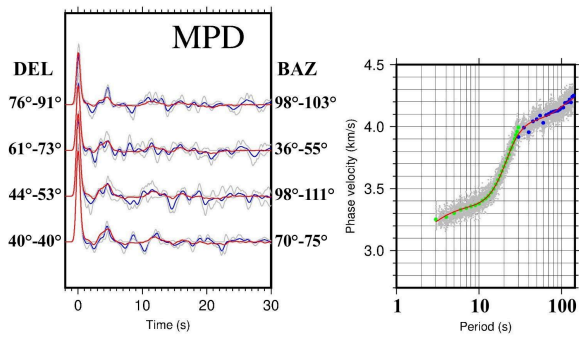
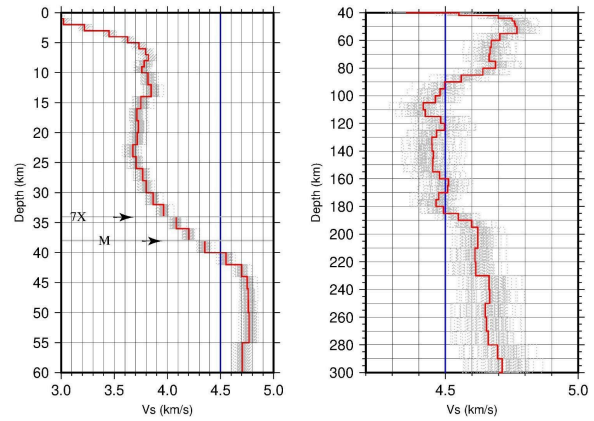
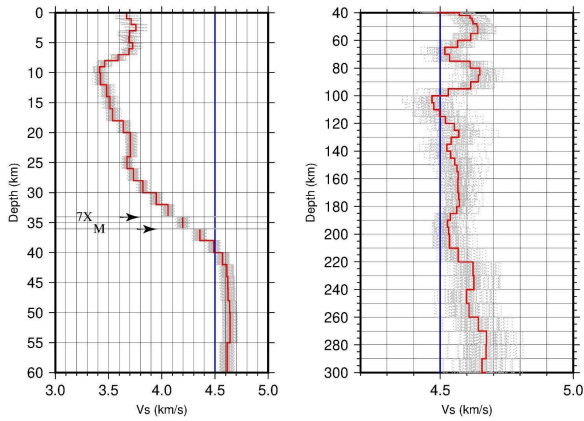
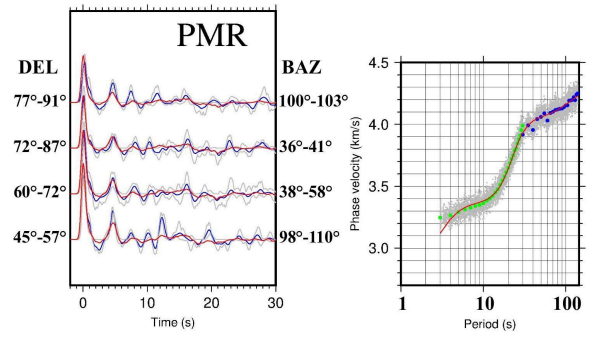
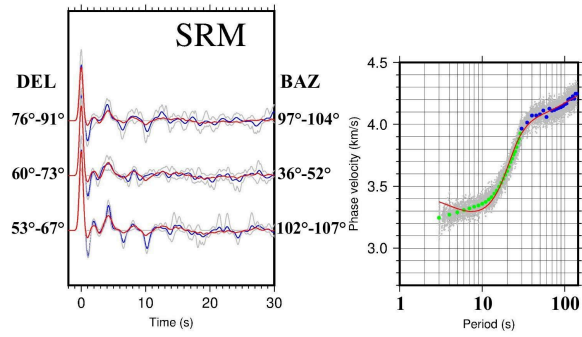












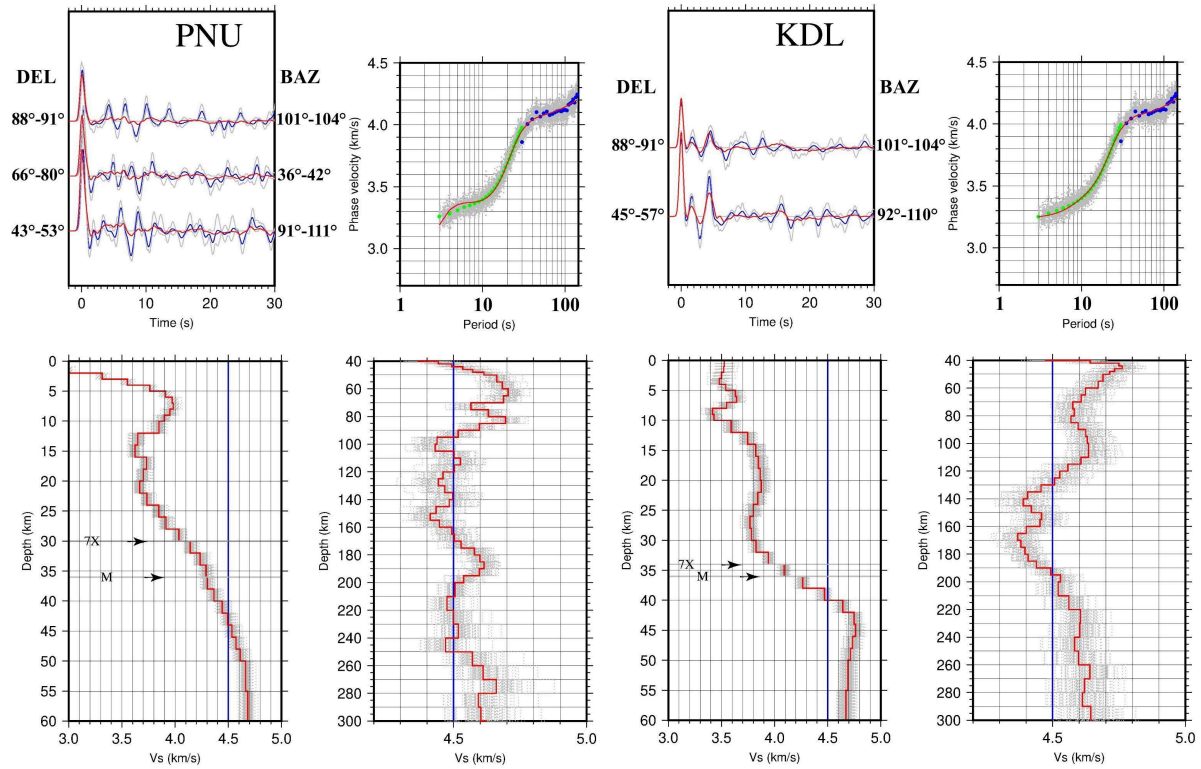


Fig. S2.2. Joint inversion results for all seismic stations along the profile **a)** upper and lower bounds of stacked RFs (grey) along with their mean (blue) and the synthetic RF computed with the inverted model (red), **b)** 100 random Rayleigh wave dispersion curves within 5% error bounds (grey) of observed phase velocity dispersion curve (red), **c)** best 100 solutions (grey) from a set of 300 models (the red line represents the average of the 100 best solutions).

Chapter 3 - Results and Discussion

This chapter presents the results pertaining to the crust and upper mantle velocity structure along the profile as discussed in the Chapter 2, in the global and regional context.

3.1 Global view of the continental crust -

The Continental crust having a large variation of thickness (30-70 km) is the repository for all the natural resources and serves as a record of crustal evolution since early Archean. It is vertically divided into upper, middle, and lower crust based on the chemical composition and the seismic wave velocity (Christensen and Mooney, 1995; Rudnick and Fountain 1995; Christensen, 1996). Global compilations of seismic velocity structure with depth reveal considerable variability in the composition of the continental crust across different tectonic and thermal environments. The upper crust is well understood due to its surface exposure studied through various direct methods that examine its structure and composition, the middle and lower crust remain poorly investigated and understood. There are significant variations in their composition interpreted using seismological and petrological data. The shear velocity model (V_s) for the global Precambrian crust (Fig. 3.1) is derived from the compilation of the global P-wave velocity model (Christensen and Mooney, 1995) taking the average V_p/V_s ratio for the crust calculated from the crustal petrology model (Christensen, 1996). The upper, middle, lower and lowermost crust have characteristic shear wave velocities of 2.9-3.5 km/s, 3.5-3.8 km/s, 3.8-4.0 km/s and 4.0-4.3 km/s respectively (Christensen, 1996; Rudnick and Gao, 2003). Rudnick and Fountain (1995) and Rudnick and Gao (2003) argued that $V_p > 7.0$ km/s ($V_s \geq 4.0$ km/s) correspond to mafic rocks, which are characteristic of the lowermost part of the crust in many Precambrian terrains. Such high-velocity layers are found in various geological contexts, including shields, platforms, continental arcs, and rift zones. The thickness of such mafic layers has previously been used to distinguish between Archean and Proterozoic terrains and as an important indicator of tectonic processes.

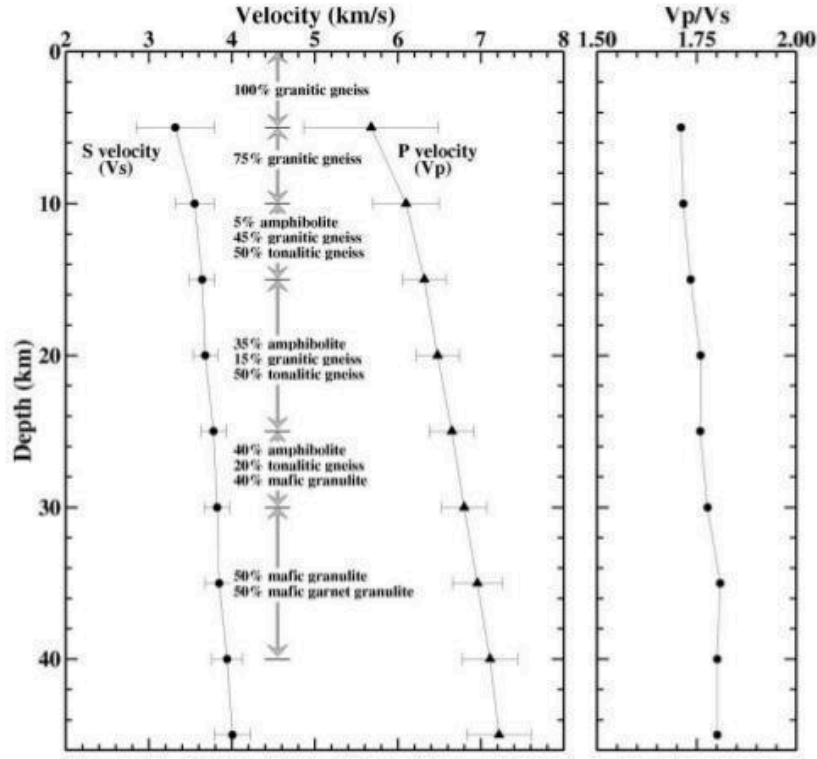


Fig. 3.1. The shear velocity model illustrating the global Precambrian crust, derived from a compilation of the global P wave velocity model (Christensen and Mooney, 1995). This model utilizes the average V_p/V_s ratio for the crust, calculated based on a crustal petrology model (Christensen, 1996).

3.2 The crust velocity along the profile -

The significantly improved shear wave velocity (V_s) image of the crust along a ~630 km long profile from the west to the east coast of South India across the Dharwar craton is mapped through the joint inversion of P-receiver functions and Rayleigh wave phase velocity dispersion (3-140 sec) (Julia et al., 2000). The profile consists of 38 dense broadband seismic stations having 10-15 km spacing, covering the Archean Dharwar craton, Proterozoic Cuddapah basin, Western Ghats and Eastern Ghats. Details of the station network are presented in chapter 1 (Figure 1.4 and Table 1.1). The joint inversion of receiver function (RF) and surface wave dispersion (SWD) and their resulting velocity models for the crust and lithospheric mantle for representative stations PRR in Fig. 3.2. For all other stations, it is shown in Fig. S2.2 (Chapter 2) We study the properties of the crust in this section.

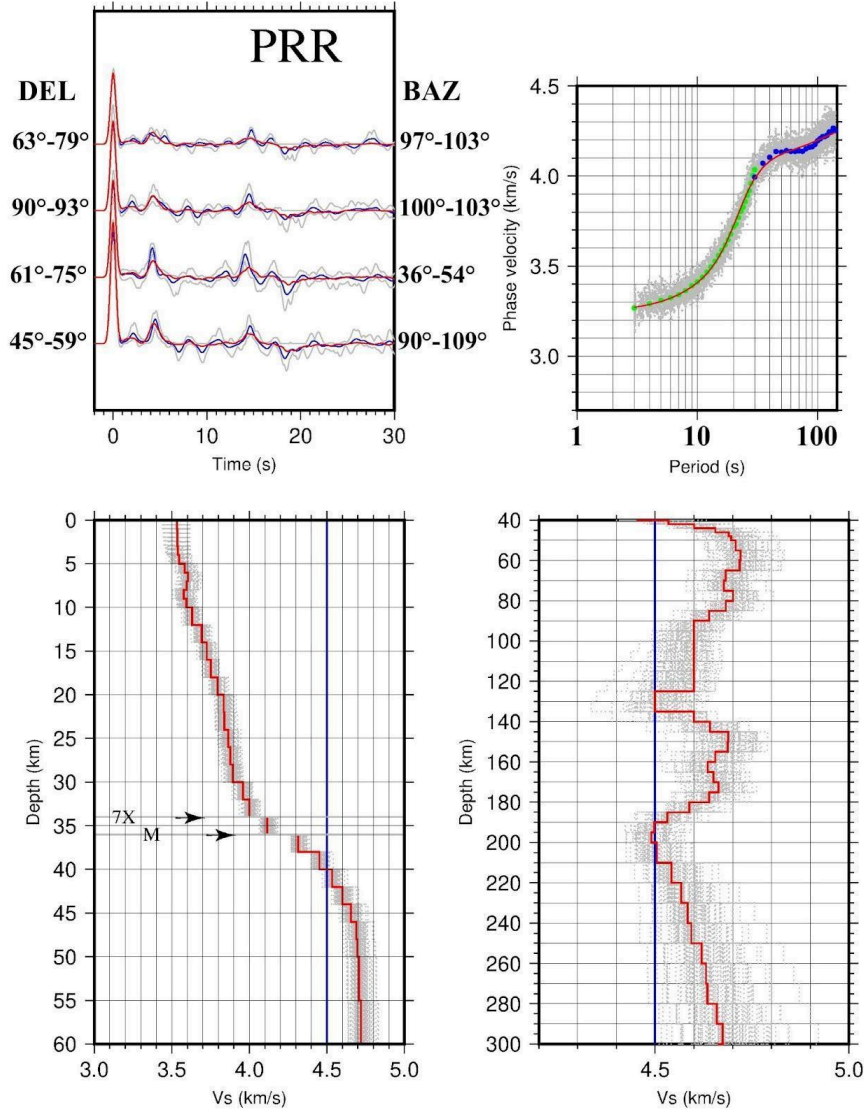


Fig. 3.2. Joint inversion result at station PRR; **(Top left)** upper and lower bound of stacked RFs (grey) along with their mean (blue) and the synthetic RF computed with the inverted model (red), **(Top right)** 100 random Rayleigh wave dispersion curves within 5% error bounds (grey) of observed phase velocity dispersion curve (green and blue) and the synthetic dispersion curve computed with the inverted model (red), **(Bottom left and right)** best 100 solutions (grey) from a set of 300 models (the red line represents the average of the 100 best solutions). The Moho (M) is marked at a depth where $V_s > 4.3$ km/s. 7X corresponds to velocity V_p of 7 km/s or $V_s > 4$ km/s. The depth difference between M and 7X represents the thickness of the underplated layer.

The velocity (V_s) at individual stations obtained from joint inversion modeling are interpolated over a 15×15 km grid to create a 2-D crustal velocity structure along the profile along with the surface topography of the area (**Fig. 3.3**). The Moho is seismologically defined as the depth at which $V_p > 7.6$ km/s or $V_s > 4.3$ km/s (Steinhart, 1967; Meissner, 1986) and the petrological Moho is defined as the base of the geological crust, the boundary between non-peridotitic mafic crustal rocks and olivine-dominated ultramafic mantle rocks, irrespective of seismic velocity. The continental crust consists of sediments, gneisses, granite, granodiorite, gabbro, amphibolite, and granulite and the upper mantle consists predominantly of peridotites and other ultramafic rocks. Petrological Moho does not necessarily coincide with the seismological Moho (Meissner, 1986; Thybo et al., 2013). Except the Coast-WG domain having Moho depth in the range 38-40 km, other domains show Moho depth range 36-38 km. The Moho depth is consistent and matching with the Common Conversion Point stacking (CCP) image (Dueker and Sheehan, 1998) using Gaussian Width (GW) ~ 2.5 for V_p/V_s of 1.73 (Fig. 3.4). The change in V_p/V_s does not alter the Moho depth. The piercing points for Ps converted phase at 20 km, 40 km and 60 km depths are **shown in the figure 3.5**. All geological domains along the profile show crustal thickness almost equal to the global average. Along the profile (**Fig. 3.3**), the contribution of the middle crust thickness out of total crustal thickness is maximum ($> 50\%$) and the average velocity in all the domains in the range 3.70-3.75 km/s, suggest the average composition of the crust to be felsic to intermediate.

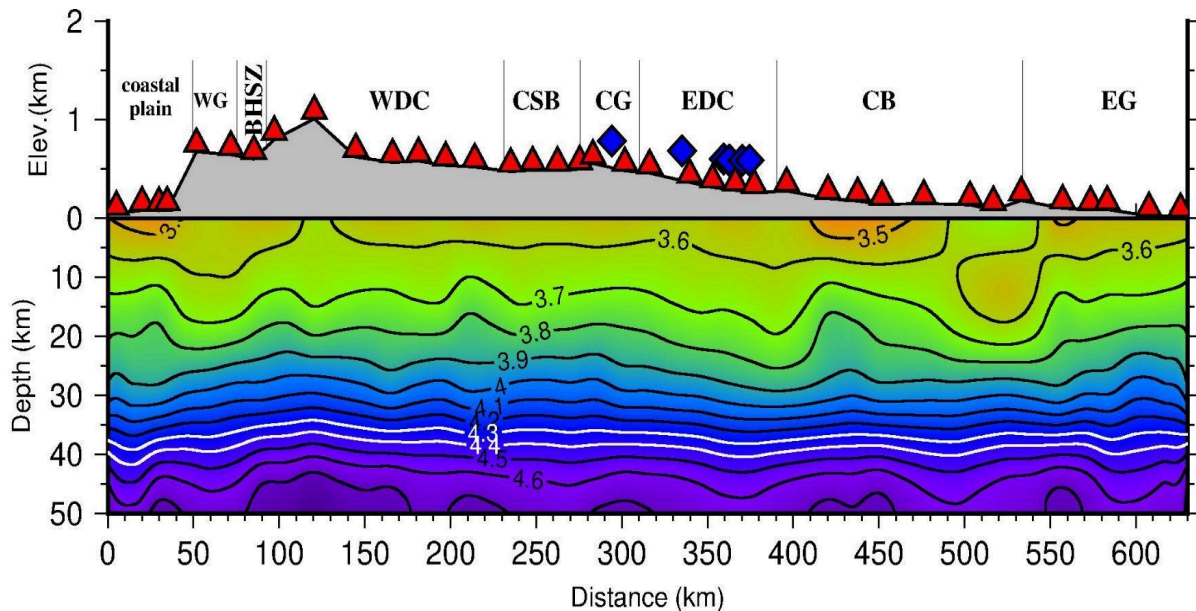


Fig. 3.3. Velocity-depth section along with the surface topography (grey color) of the crust along the E-W profile from the west to the east coast covering Dharwar craton in the depth of 0-50 km in terms of absolute shear velocity (V_s).

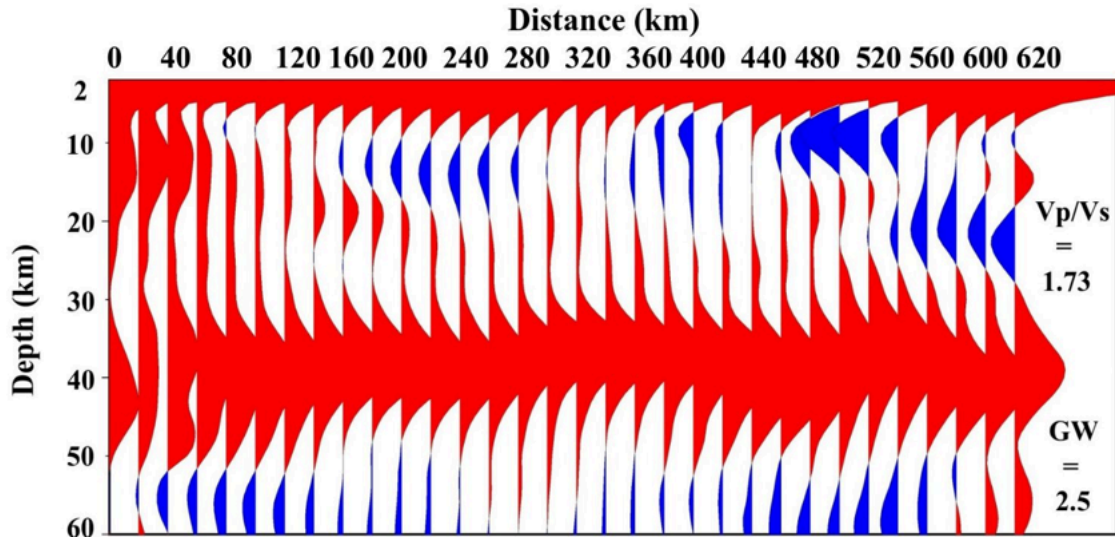


Fig. 3.4. The CCP amplitude waveform image.

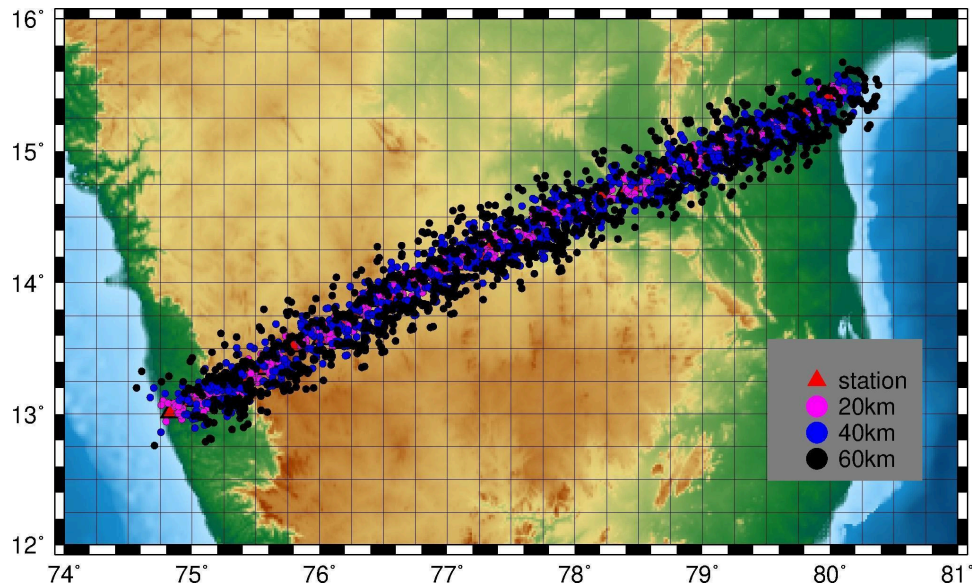


Fig. 3.5. The piercing points map for Ps converted phase at 20 km, 40 km and 60 km depths.

The upper crust is almost similar in all the domains and is almost absent/thinner 2-3 km thickness. This is no surprise due to 8-10 km erosion level of the rocks along the profile which are amphibolite facies corresponding to 2-3 kb pressure. The middle and lower crust beneath the Coast-WG and WDC-CG domains are 20-22 km and 8-10 km respectively while the middle and lower crust beneath the EDC domain are ~26 km and ~4 km showing more mafic crust beneath the Coast-WG and WDC-CG domains than the EDC domain. CB and EG domains show similar middle and lower crust thicknesses like EDC having more felsic behavior than the Coast-WG and WDC-CG domains. We don't observe strong intra-crustal layering. So the RFs have minimal multiples, leading to improved resolution of the mantle structure and mapping of Lithosphere-Asthenosphere Boundary (LAB) which is discussed in section 3.3.

3.2.1 Compressional wave velocity (Vp), Density (ρ) and Shear modulus (μ) image along the E-W profile -

The computation of compressional wave velocity (Vp), density (ρ) and shear modulus (μ) along the E-W profile is derived from the shear velocity (Vs) produced by the joint inversion of RF and SWD. We have followed the empirical regression relationship equation (3.1) to compute Vp from Vs and then equation 3.2 to compute density (ρ) from Vp (Brocher, 2005).

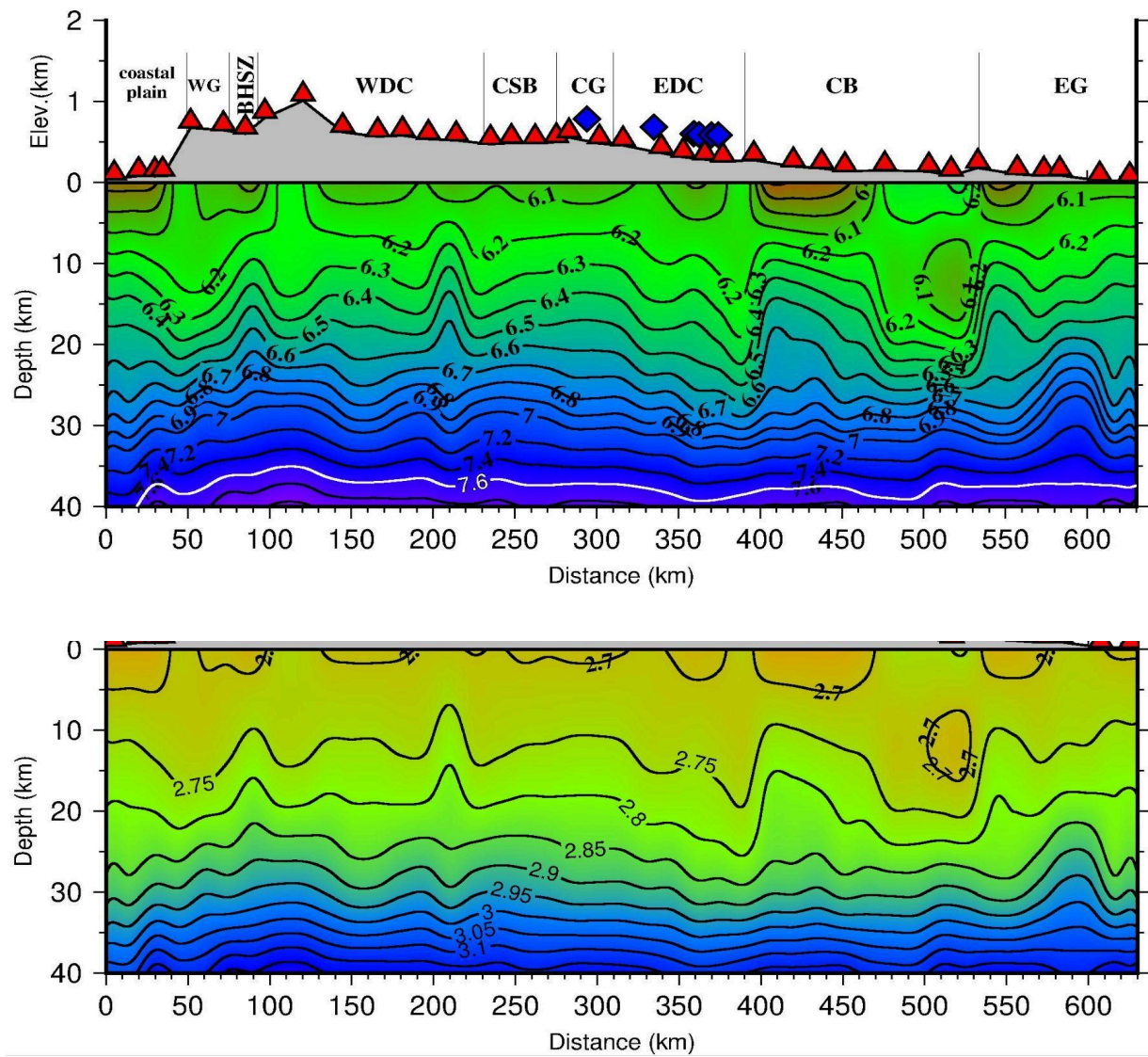
$$Vp \text{ (km/s)} = 0.9409 + 2.0947Vs - 0.8206 Vs^2 + 0.2683Vs^3 - 0.0251Vs^4 \quad (3.1)$$

$$\rho \text{ (gm/cc)} = 1.6612Vp - 0.4721Vp^2 + 0.0671Vp^3 - 0.0043Vp^4 + 0.000106Vp^5 \quad (3.2)$$

$$\text{Shear modulus } (\mu) = Vs^2 \rho \quad (3.3)$$

The Vp, density and shear modulus image (**Fig. 3.6**) show the flat Moho and well stratified crustal layers consistent with the Vs image along the profile. Vp ~ 7.6 km/s, ρ ~ 3.2 gm/cc and μ ~ 60-64 GPa are considered as the Moho globally on the basis of which we have found 38 ± 2 km crustal thickness which is consistent with the Vs image along the E-W profile. At the WG-WDC boundary, the same depression in the middle crust is seen while at the EDC-CB boundary, a local upwarping type or dipping type feature of middle crust can be seen beneath CB which may possibly be the signature of upwelling of mafic material from lower to the middle crust or some crustal reworking. At the CB-EG boundary, an unusual same low velocity layer (Vp < 6.1 km/s, ρ < 2.7 gm/cc, μ < 41 GPa) can be seen dipping towards CB from EG which

may possibly be the signature of Nallamalai Fold Belt (NFB) as shown in geological map (Fig. 1.4).



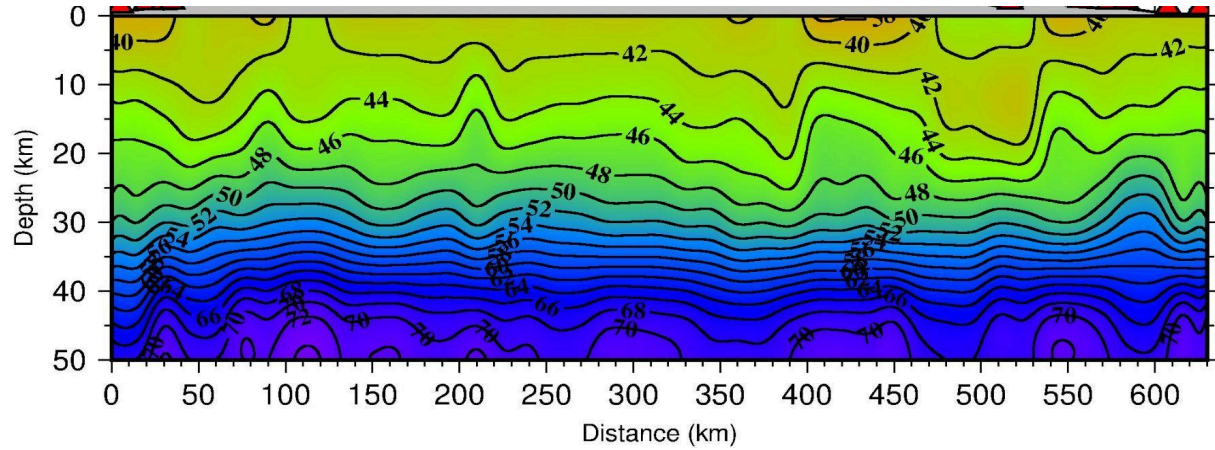


Fig. 3.6. (Top) Compressional wave velocity (V_p) image computed from the shear velocity (V_s) along the E-W profile. **(Middle)** Density (ρ) image computed from the compressional wave velocity (V_p) along the E-W profile by above equation 1 from Brocher, 2005. **(Bottom)** Shear modulus (μ) image.

3.2.2 Comparison with global velocity models and nature of the deeper part of the crust -

On the basis of age and geology, we have divided the whole E-W Profile into 5 domains: Coastal Plain-Western Ghat (Coast-WG), Western Dharwar Craton-Closepet Granite (WDC-CG), Eastern Dharwar Craton (EDC), Cuddapah Basin (CB) and Eastern Ghat (EG). Comparing the velocity characteristics of the Archean and Proterozoic crust of the Dharwar Craton with the global averages (**Fig. 3.7**), we observe that All domains have Moho depth in the range 36-40 km almost equal to the global average. In our results, We consider $V_s \sim 4.3-4.4$ km/s ($V_p \sim 7.6$ km/s and $V_p/V_s \sim 1.77$) as the Moho (Christensen, 1996). Coast-WG, WDC-CG and EG domains have velocity structure similar and equal to global average while the EDC and CB domains show lower velocity than the global velocity model in the deeper structure in the depth ranges 15-33 km and 20-33 km respectively suggesting felsic middle and lower crust. On the basis of compilation of the global P-wave velocity model and crustal petrology model (Christensen and Mooney, 1995; Christensen, 1996), the upper, middle, lower and lowermost crust (magmatic underplating) have the shear velocity ranges: 2.9-3.5 km/s, 3.5-3.8 km/s, 3.8-4.0 km/s and 4.0-4.3 km/s respectively (Christensen, 1996; Rudnick and Gao, 2003).

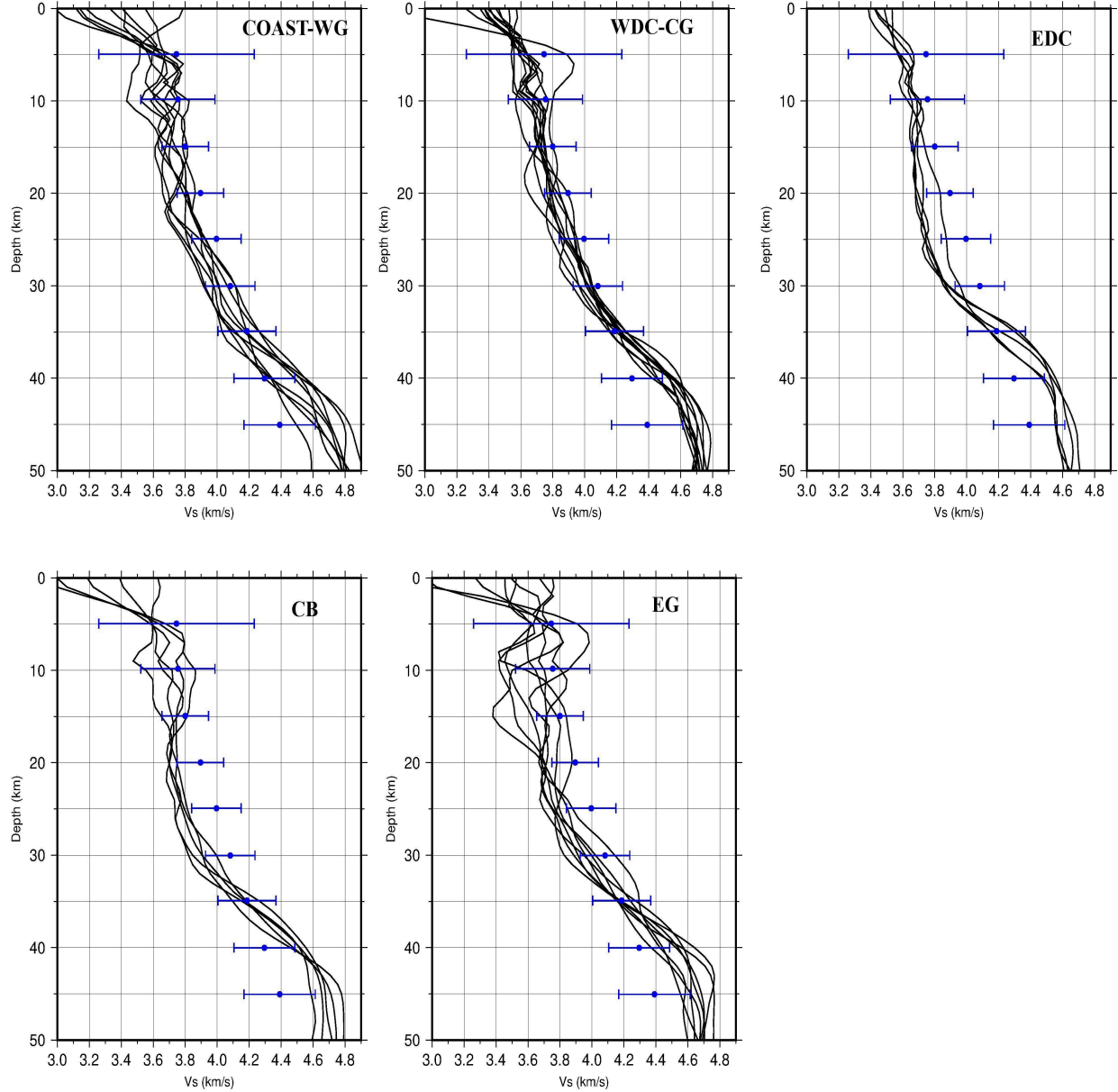


Fig. 3.7. Comparison of the velocity characteristics of the Dharwar craton and other domains with those derived for global average (Christensen and Mooney, 1995; Christensen, 1996).

On the basis of these velocity ranges, the upper crust is almost absent/very thinner 2-3 km thick along the entire domains suggesting the signature of either upliftment and erosion as our study area falls under the Greenschist-Amphibolite transition zone marked in the geological map (Fig. 1.4). Coast-WG and WDC-CG domains have middle and lower crust thicknesses in the range 20-22 km and 8-10 km respectively while EDC domain shows middle and lower crust thicknesses in the range 25-26 km and ~4 km respectively. CB domain shows anomalous

behavior having locally upwarped crustal columns beneath few stations having thinner middle crust ~ 20 km and thicker middle crust in other locations beneath the CB having 23-25 km thickness. The EG domain also shows a 23-25 km thick middle crust. EDC and EG domains both have thinner lower crust of ~ 4 km thickness while CB shows 6-8 km thick lower crust. EDC and CB domains show lowermost crust (HVL) thickness 4-6 km while the other domains show 6-8 km thickness. If we combine lower and lowermost crust thicknesses for all the domains, then EDC domains show thinnest layer (~ 8 -10 km), CB and EG domains have ~ 10 -12 km thickness and Coast-WG and WDC-CG domains show 14-16 km thickness suggesting EDC to be most felsic of all the domains and Coast-WG and WDC-CG domains to be most mafic of all the domains. This suggests the Coast-WG and WDC-CG domains have been affected by the past magmatic/mantle plume event like Marion plume (~ 85 Ma). Presence of a thin lower crust beneath the EDC would lead to sharp velocity contrast at the crust–mantle boundary. In contrast, the thicker lower crust of Coast-WG and WDC-CG domains would have more gradation in the velocity and lower impedance contrast at the crust–mantle boundary. EDC shows more felsic behavior than the WDC-CG domain as inferred by previous studies also. Our velocity image in the western and eastern end doesn't show a signature of extensive underplating due to India-Madagascar separation and multiple tectonic events happened in the east coast respectively.

3.2.3 Comparison with DSS Results in Dharwar craton -

The velocity structure along the E-W profile are compared with the results from DSS crustal studies done previously. The V_p model results from Udupi to CG (shown in geological map fig. 3.8) having ~ 250 km profile length are shown below in the figure 3.9 (Mall et al., 2012) are compared with our V_p results derived from Brocher, 2005 for the same profile length ~ 250 km down to 40 km depth as shown in figure 3.10 below.

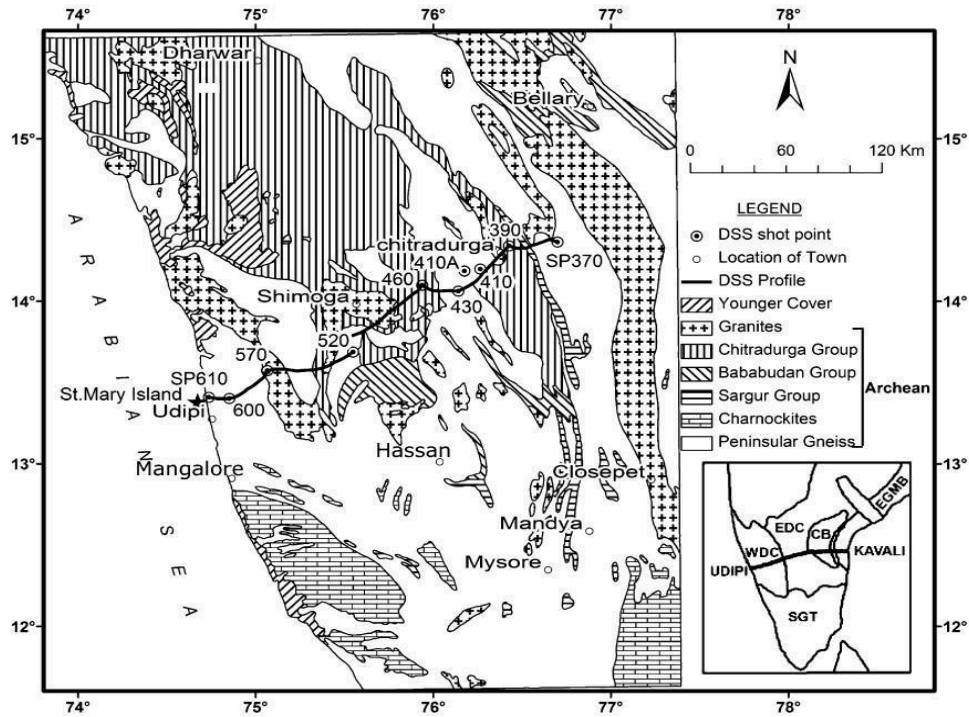


Fig. 3.8. Location map of western segment of Kavali-Udipi DSS profile having 250 km length (Mall et al., 2012). Labels represent Western Dharwar craton (WDC), Eastern Dharwar craton (EDC), Cuddapah basin (CB), and southern Granulite terrain (SGT). Other details are discussed in Mall et al., 2012.

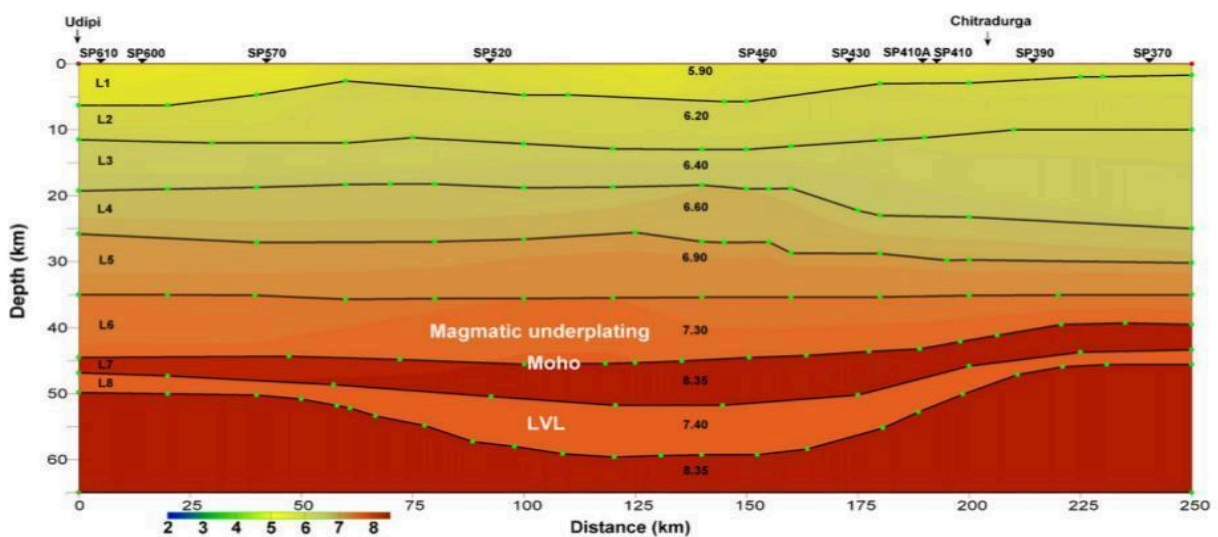


Fig. 3.9. Vp model along the Udupi to CG (shown in geological map Fig. 3.8) from Mall et al., 2012. The numbers show the average layer velocities (km/s). More details are discussed in Mall et al., 2012.

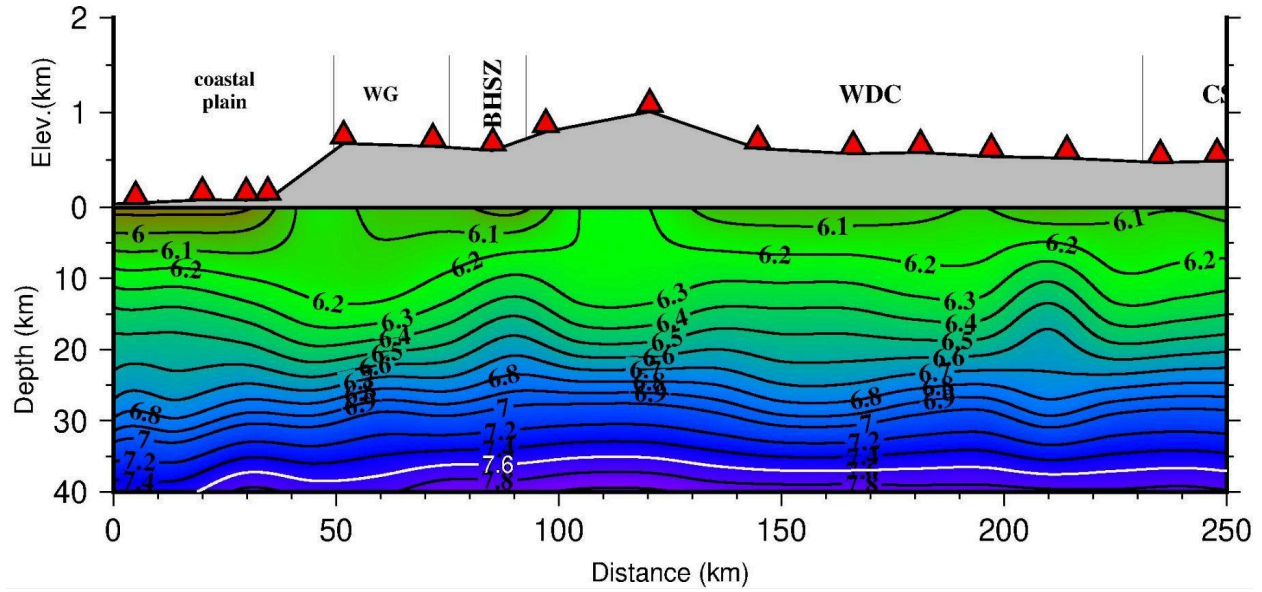


Fig. 3.10. Crust P-velocity image from Udupi to CG (shown in geological map Fig. 3.8) having ~250 km profile length compared with figure 3.9 (Mall et al., 2012).

Our results suggests the high velocity magmatic layer to be around 4-5 km thick and 36-40 km thick Moho along the ~250 km long profile while Mall et al. (2012) found high velocity magmatic layer ($V_p \sim 7.30$ km/s) at the base of the crust beneath WDC to be around 10 km thick except the CSB having ~5 km thick underplated layer and ~42-45 km thick Moho which is ~4-5 km thicker than the Moho thickness from our results. We didn't find any LVL just below the Moho. Instead, we found HVL which is shown in the above figure 3.10 and figure 3.15.

Also, our results of the crust study from Chikmagalur to Perur (shown in geological map Fig. 3.11) having ~200 km profile length (Fig. 3.13) are compared with the study of Mandal et al., 2018 (Fig. 3.12). On the basis of the seismic reflectivity pattern, Mandal et al., 2018 has suggested the subduction-accretion process of two stages for the crustal evolution in the Neoproterozoic Dharwar Craton having simple crust with subhorizontal Moho of the WDC and

complex dipping Moho in the EDC having signature of collision, a thrust fault and a crustal root but in case of our results, except at the EDC-CB boundary showing local eastward dipping layer or local upwarping, our results show flat Moho (not showing any dipping Moho) having ~36-38 km thickness along the ~200 km long profile from Chikmagalur to Perur covering WDC and EDC. On the basis of our results, we don't support the subduction-accretion process of two stages for the crustal evolution in the Neoproterozoic Dharwar Craton as we don't find any signatures of dipping layers. There is also a possibility that the tectonic deformations like folding, faulting, and crustal tilting initially created relief on the Moho boundary. However, these structural variations may subsequently got erased by ductile flow showing widespread horizontality (flat) of the Moho as suggested by Roy Chowdhury and Hargraves, 1981 but these studies were later questioned by Gibbs (1986) discussed in detail in the Chapter 1.

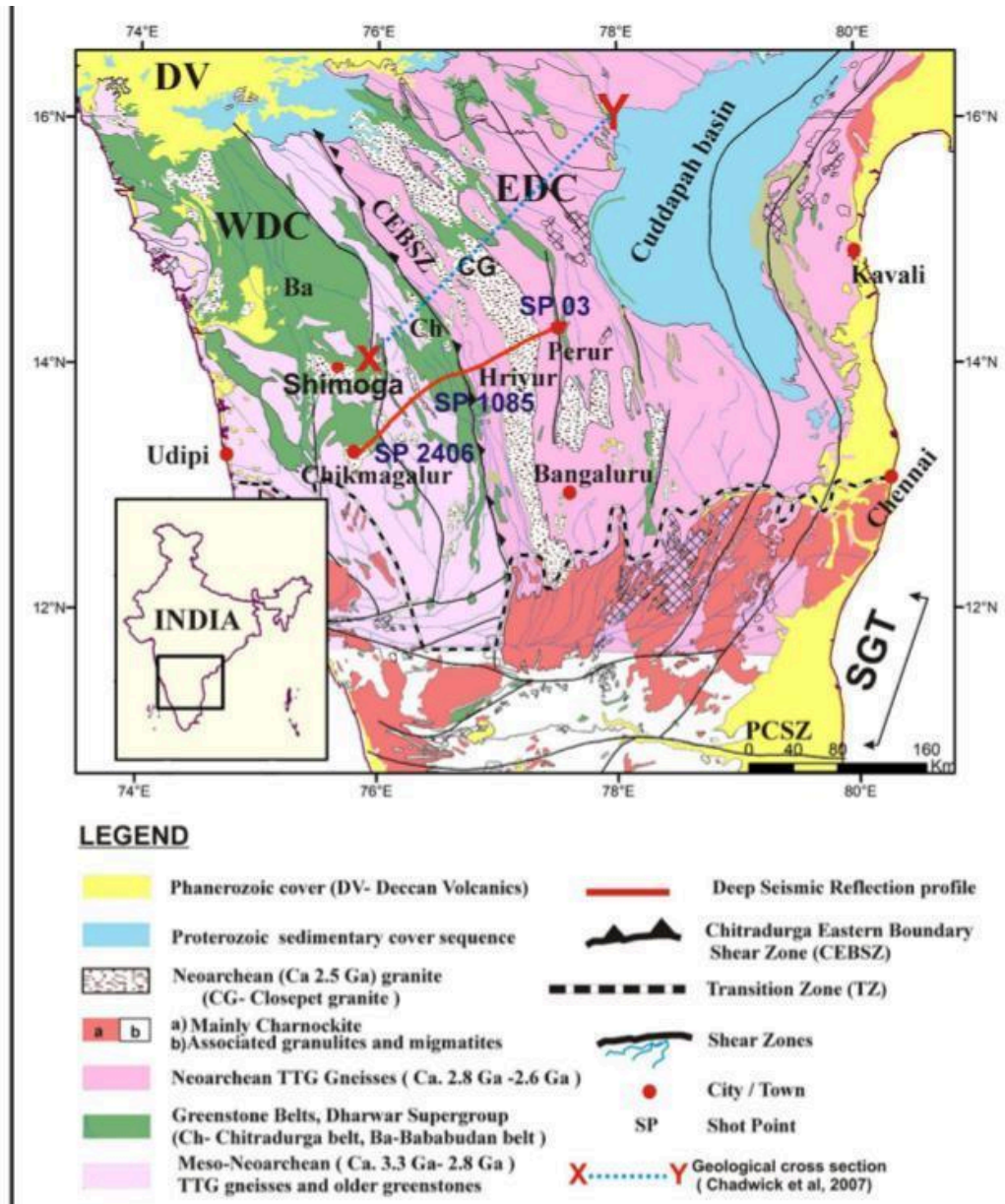


Fig. 3.11. A geological map showcasing southern India is presented, with the Perur-Chikmagalur seismic reflection profile indicated on it. The numbers denote the locations of the Shot Points utilized in the collection of reflection data. WDC represents Western Dharwar Craton; EDC represents Eastern Dharwar Craton and SGT represents Southern Granulite Terrain; and PCSZ indicates the Palghat Cauvery Shear zone (Mandal et al., 2018 and the references therein).

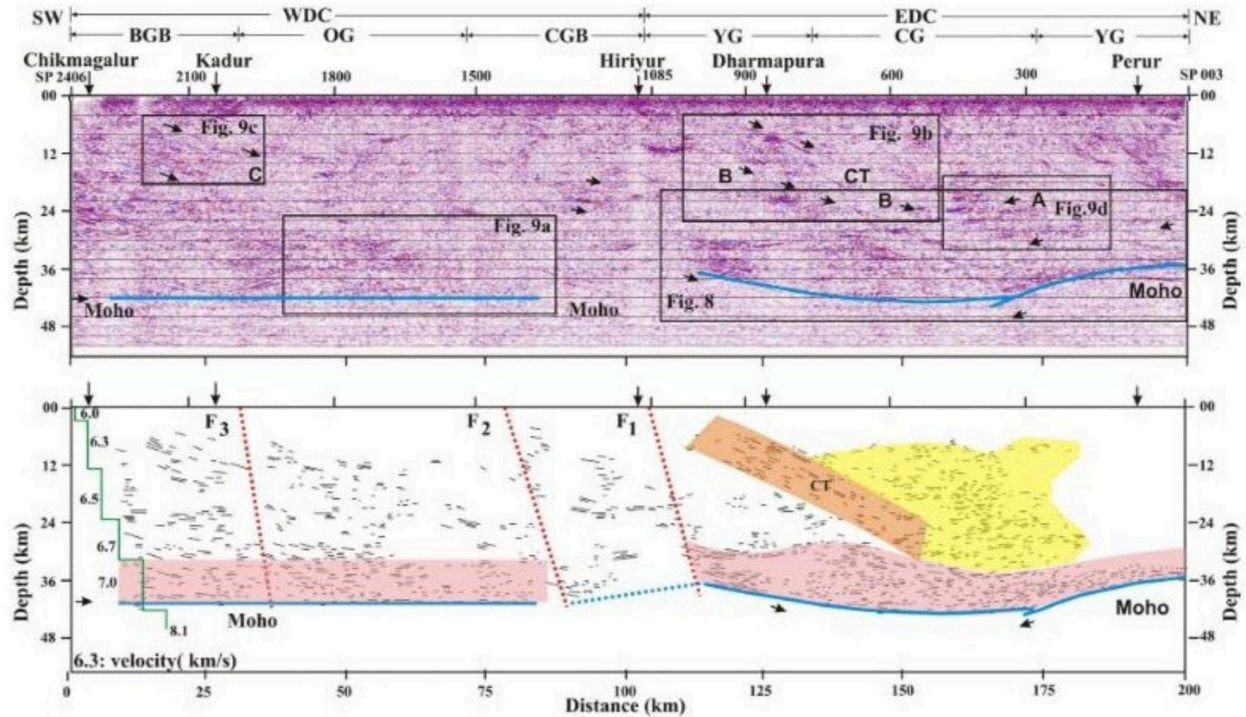


Fig. 3.12. (Top) A depth-migrated seismic section of the Perur-Chikmagalur profile is presented along with its interpretation. A bright subhorizontal reflective band located at a depth of 40 km corresponds to the Moho in WDC. There are variations in the Moho signature between WDC and EDC. A SW dipping reflective band has been labeled as 'A', while NE dipping reflective bands are also indicated are marked as 'B and C' (Mandal et al., 2018); **(Bottom)** Line drawing derived from prominent reflections from the seismic depth-migrated section, along with the velocity model derived from coincident refraction/wide-angle reflection data (1-D velocity model is after Vijaya Rao et al. 2015b). The lower-crustal subhorizontal reflection band (orange colour) coincides with the high-velocity (7.1 km/s) lower-crustal layer. Steeply dipping faults (F₁ - F₃) are inferred based on the differences in reflectivity on either side. They coincide with the tectonic boundaries of the region. Yellow colour represents the geometry of the Closepet granite and red colour corresponds to major thrust faults of the region. More details are provided in Mandal et al., 2018.

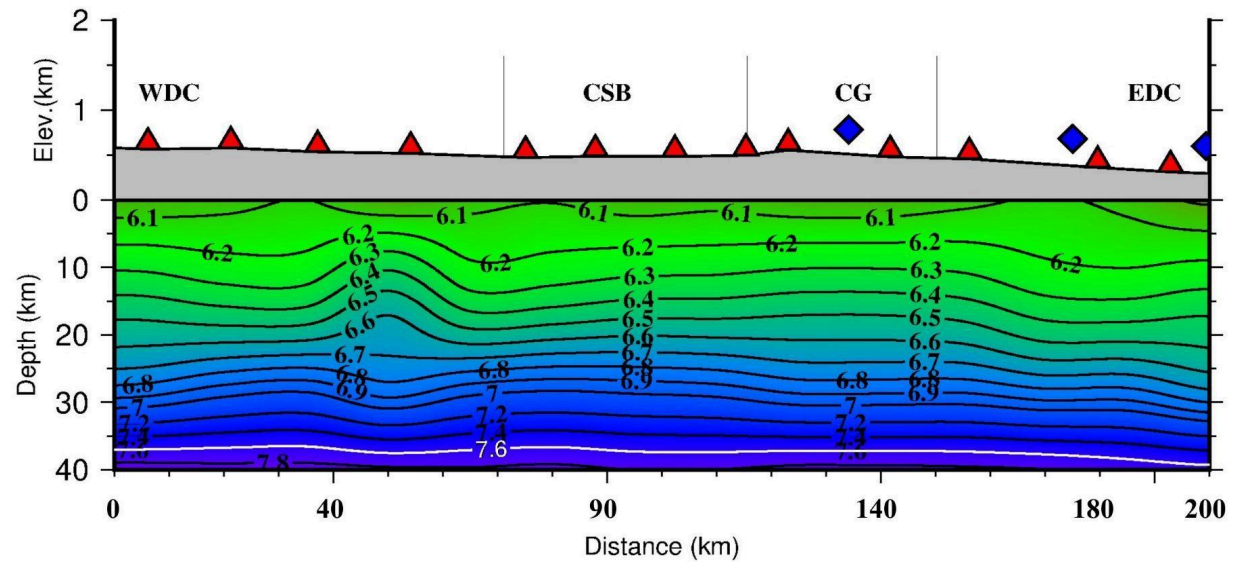


Fig. 3.13. Crust P- velocity image from Chikmagalur to Perur (shown in geological map Fig. 3.11) having ~200 km profile length compared with figure 3.12 (Mandal et al., 2018).

3.2.4 Distinct features of Cuddapah Basin and Eastern Ghats –

The Cuddapah Basin (CB) in India has been the subject of extensive research to understand its complex geological history and crustal architecture. A detailed 2D seismic study, along with geological and geophysical data, across the Proterozoic Cuddapah Basin in southern India shows an upwarping of layers in the middle crust beneath the Cuddapah Basin supported with: (i) The strong positive oval shaped gravity anomaly of about 55 mGal surrounded by a trough of gravity lows (NGRI, 1978; Krishna Brahmam et al., 1986 and references therein; Vasanthi and Mallick, 2005; Figure 9 of Mall et al., 2008), (ii) Widespread intrusive activity on a large scale. (iii) High conductivity anomaly attributed to a basic/ultrabasic source (Naganjaneyulu and Harinarayana, 2004). These geophysical signatures are interpreted as the expression of a possible thermal anomaly.

Mantle plumes, even in smaller volcanic regions, are typically local having 100-200 km width and contain a column of hot upwelling rocks. In typical scenarios, plumes that exhibit temperatures exceeding the background by 100-200 °C inducing regional uplift along with significant extension (White and McKenzie, 1989), which facilitates the substantial injection of volcanic material into the lower crust which laterally spreads upon encountering a pronounced

viscosity gradient (Meissner, 1986). The cooling of this magma leads to the development of a uniform layer composed of mafic residues and stratified melt products at the base of the crust which gets differentiated on the basis of density with the geological time and either gets delaminated or preserved beneath the Moho as highly dense mafic layer (Kay and Kay, 1991).

The CB has undergone episodic magmatism between 1.8-1.0 Gac (Anand et al., 2003). As a result of this, the crustal column beneath the Cuddapah basin has become increasingly mafic and dense, leading to significantly elevated crustal velocities. The magmatic underplating (HVL having $V_s \sim 4.0-4.3$ km/s) at the base of the crust has a thicker layer (6-8 km) beneath this column than the other locations beneath the Cuddapah Basin having ~ 4 km thickness. The crustal layers are compensated together in such a way to maintain a constant crustal thickness of 36-38 km beneath the CB suggesting flow type feature (visco-elastic flow).

The eastern margin of the Cuddapah Basin is marked by eastward-dipping thrust faults, with the Eastern Ghats (EG) thrusting upwards against the basin (Fault 2, Fig. 11 of Kaila et al., 1979). The juxtaposition of the obducted EG and the Cuddapah Basin, two geologically distinct terrains, supports the notion of a Proterozoic collision, which is suggested to have contributed to the continental growth during that era (Kroner, 1991).

Table 3.1 - Details of the broadband seismograph network along the E-W Profile covering Dharwar craton, having station location details, Moho depth, and the thickness of the upper crust, middle crust, lower crust, and high-velocity layer (HVL having $V_s \sim 4.0-4.3$ km/s).

S. No.	Station code	Lat (°N)	Long (°E)	Elev. (m)	Moho depth (km)	Upper crust (km)	Middle crust (km)	Lower crust (km)	HVL (km)
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Coastal plain-Western Ghat (WG)

1	SKL	13.016	74.813	40	40	3	23	8	6
2	ALN	13.102	74.995	98	40	4	22	8	6
3	STD	13.107	75.086	94	38	4	22	8	4

4	HSR	13.134	75.120	100	38	4	20	8	6
5	KMK	13.219	75.252	814	40	0	20	12	8
6	BHL	13.268	75.430	805	40	2	22	4	12

Western Dharwar Craton (WDC)

7	SPT	13.347	75.525	750	38	3	17	10	8
8	BGS	13.407	75.616	996	36	4	12	12	8
9	SVR	13.519	75.799	1260	36	0	20	8	8
10	KDR	13.574	76.016	774	36	2	20	6	8
11	KLR	13.690	76.175	706	38	3	19	10	6
12	HSG	13.778	76.282	723	36	2	20	8	6
13	MLR	13.864	76.400	668	36	1	17	12	6
14	UDL	13.884	76.556	648	36	3	21	6	6

Chitradurga Schist Belt (CSB)-Closepet Granite (CG)

15	DMR	13.987	76.719	589	36	2	22	6	6
16	HYB	14.063	76.808	608	36	2	20	6	8
17	AMR	14.077	76.942	605	38	0	18	10	10
18	LDH	14.142	77.042	620	38	0	18	12	8
19	PVG	14.180	77.102	694	36	3	17	8	8
20	YSN	14.187	77.275	601	36	3	17	12	4

Eastern Dharwar Craton (EDC)

21	PRR	14.294	77.352	570	36	0	18	14	4
22	RMG	14.318	77.568	453	36	2	26	4	4

23	MMP	14.414	77.645	390	36	2	26	4	4
24	DMM	14.479	77.748	350	38	3	25	4	6
25	BTP	14.522	77.840	321	38	1	25	6	6

Cuddapah Basin (CB)

26	PRP	14.544	78.014	343	38	0	28	6	4
27	SKS	14.653	78.209	245	36	4	24	4	4
28	MDN	14.652	78.370	217	38	4	20	6	8
29	RTP	14.734	78.471	171	38	4	20	8	6
30	DUV	14.830	78.675	185	36	3	23	6	4
31	PML	14.908	78.913	167	38	0	26	6	6
32	RSP	14.980	79.016	105	36	0	22	10	4
33	SRM	15.038	79.155	222	36	0	28	2	6

Eastern Ghat (EG)

34	PMR	15.100	79.372	123	38	3	25	4	6
35	MPD	15.141	79.516	105	36	3	23	4	6
36	IRD	15.174	79.599	108	38	3	23	4	8
37	PNU	15.277	79.801	28	36	3	21	4	8
38	KDL	15.425	79.996	16	36	0	28	4	4

3.3. The state of lithosphere- asthenosphere

Understanding the mechanism of evolution of continents comprising Archean Cratons flanked by younger mobile belts through time is challenging. Cratons are the oldest parts of continents having cold, thick, high velocity, depleted lithosphere which remained stable for billions of years

(Lee et al., 2011; Pearson et al., 2021) and contains the record of previous geological events. Archean Dharwar Craton is a part of the fast-moving Indian plate, and has undergone many tectonomagmatic events like mafic dyke swarms (~2.3 Ga), Cuddapah basin formation (~1.8 Ga), Kimberlitic volcanism (~1.1 Ga and ~90 Ma), the eastern end rifted from eastern Gondwanaland (~130 Ma) while the western end of the Dharwar craton has been subjected to the interaction with the Marion mantle plume (~85 Ma) and rifting from Madagascar. It is an excellent location to test hypotheses related to the preservation or possible modification of the lithosphere (from Archean to the recent) due to its interaction with different tectonomagmatic events. To understand the nature and quantify the erosion of the lithosphere of Dharwar craton and the surrounding younger mobile belts, its present-day architecture is investigated using seismic imaging and compared with those derived from diamond-bearing kimberlite proxies from only available Wajrakarur kimberlite field for the existence of thick cratonic lithosphere at the time of their eruption.

We investigate the preservation, modification and possible destruction of the lithosphere using significantly improved shear wave velocity (V_s) image along a ~630 km long profile from the west to the east coast of South India across the Dharwar craton. (as shown in map in **figure 1.4**). Details like computation of receiver function, surface wave dispersion, joint inversion and their analysis are given in chapter 2.

3.3.1 Rayleigh wave dispersion and nature of intra-crustal layering -

We analyze our average Rayleigh wave dispersion data for Dharwar craton in comparison with the average for South India (Mitra et al., 2006), global model AK135 (Kennett et al., 1995) and dispersion measurements from other cratons like the Kaapvaal, the Slave and the Yilgarn (Pedersen et al., 2009) in **Fig. 3.14**. Our findings indicate that **(i)** Our phase velocity dispersion aligns and correlates well with Mitra et al. (2006), **(ii)** Beyond 30 s, the Dharwar phase velocity is higher by 0.1 km/s (2–4%) compared to AK135, **(iii)** Dharwar phase velocity lower by 0.05–0.1 km/s (2–4%) compared to Kaapvaal, Slave, Yilgarn cratons for the period 60 to 140 s. In conclusion, the lithosphere of South India exhibits shear wave velocities that are either higher than or comparable to those of many other cratons at shallower depths, while being lower to those cratons in the lower lithosphere. Also the dispersion is being combined with the receiver function (RF) giving more weightage (90%) to the RFs and 10% weightage to the phase velocity

dispersion. The RFs provide additional constraint and more robustness to the shear wave velocity results beneath the stations.

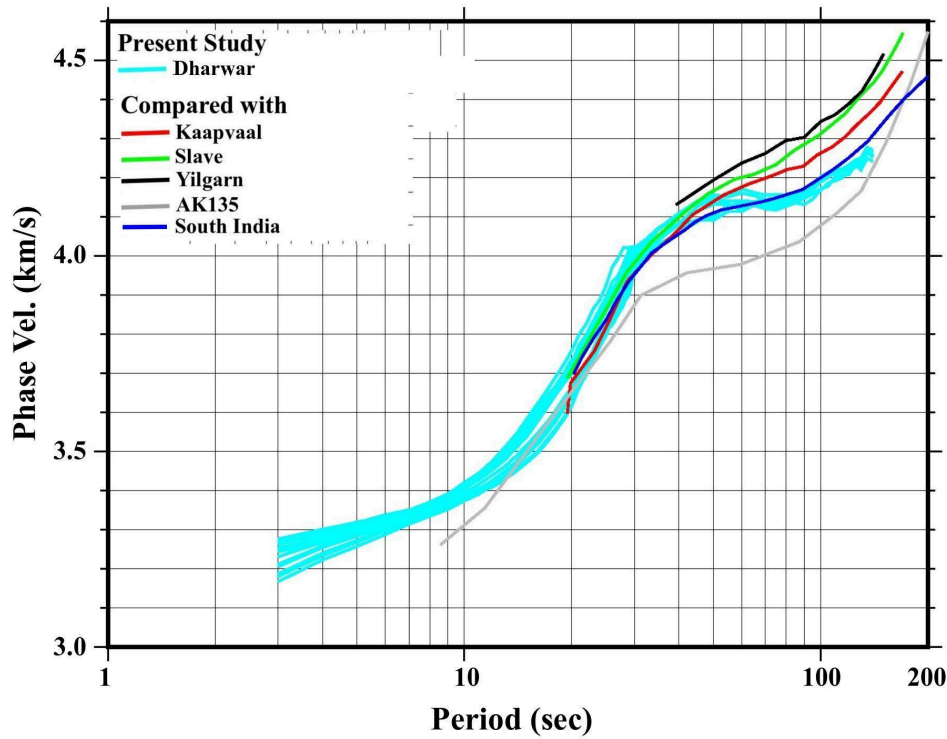
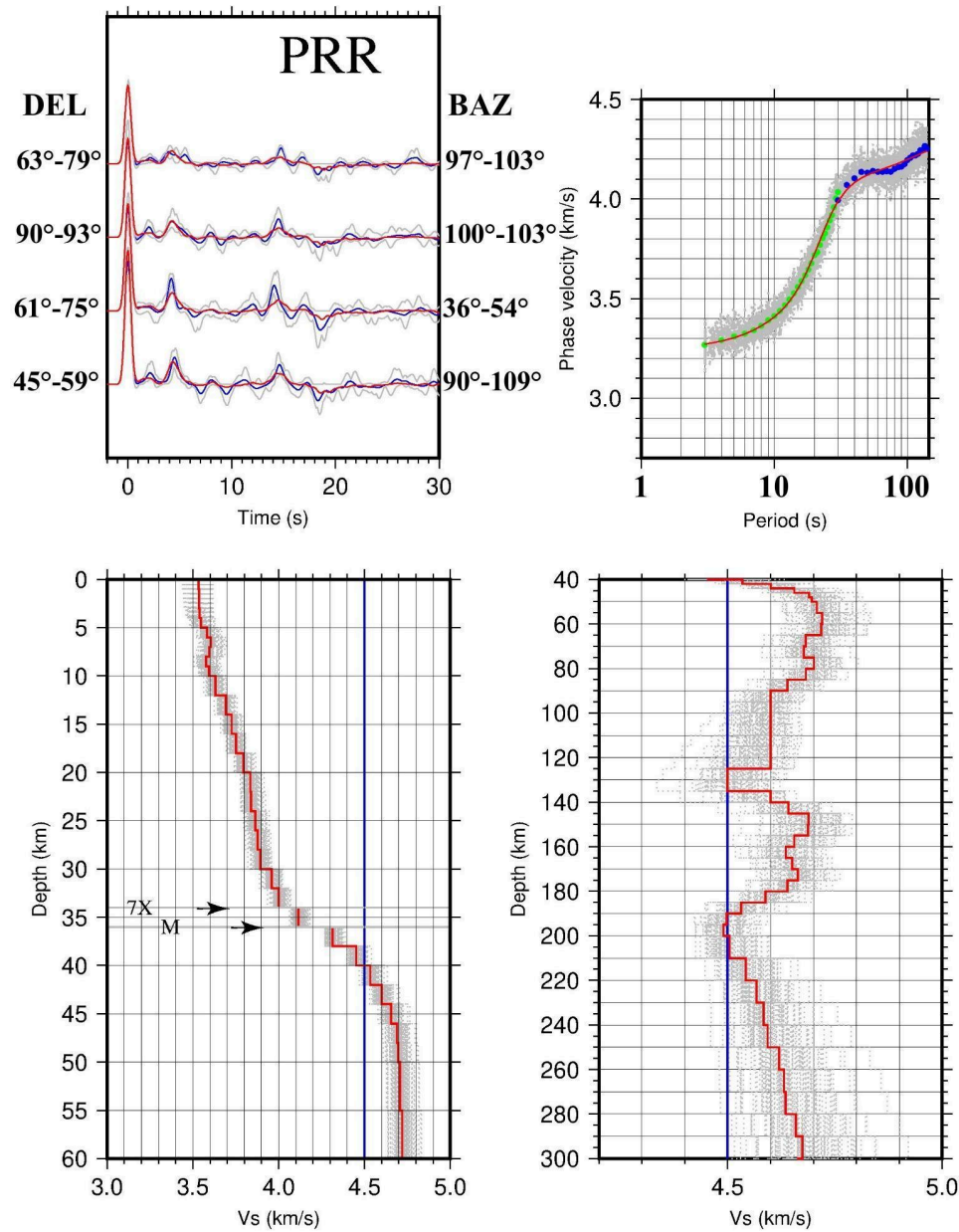


Fig. 3.14. Rayleigh wave phase velocity dispersion data in the time period 3–140 s in the Dharwar along the E-W profile (Fig. 1.4). The dispersion curves are compared with the average model for South India (Mitra et al., 2006), AK135 (Kennett et al., 1995) and three cratons Kaapvaal, Slave and Yilgarn (Pedersen et al., 2009).

The joint inversion of receiver function and phase velocity dispersion provides the shear wave velocity model for the representative station PRR in Fig. 3.2. For all other stations, it is shown in Fig. S2.2 (Chapter 2).



Same Fig. 3.2. Joint inversion result at station PRR; **(Top left)** upper and lower bound of stacked RFs (grey) along with their mean (blue) and the synthetic RF computed with the inverted model (red), **(Top right)** 100 random Rayleigh wave dispersion curves within 5% error bounds (grey) of observed phase velocity dispersion curve (green and blue) and the synthetic dispersion curve computed with the inverted model (red), **(Bottom left and right)** best 100 solutions (grey) from a set of 300 models (the red line represents the average of the 100 best solutions).

3.3.2 Velocity profile for the uppermost mantle along the profile -

The velocity models at individual stations are interpolated to generate a smooth shear wave velocity for the uppermost mantle (**Fig. 3.15 top**) along the profile. The upper mantle velocities are presented in terms of their absolute value (**Fig. 3.15 top**) as well as the percentage deviation from the global velocity model AK135 (**Fig. 3.15 bottom**). In terms of velocity perturbation, shear wave velocity (V_s) of 4.6 km/s and 4.7 km/s correspond to velocity perturbation (δV_s) of 2% and 4% respectively and the uppermost mantle layer of $V_s \geq 4.7$ km/s ($\delta V_s \geq 4\%$) is the highly depleted cratonic core which remains undisturbed and preserved from the tectonomagmatic events happened at different times (Artemieva, 2009; Darbyshire et al., 2013; Porritt et al., 2015; Celli et al., 2020).

Below Moho having shear wave velocity (V_s) ~ 4.3 km/s (**Fig. 3.3 & 3.15 top**), V_s along the E-W Profile increases progressively to 4.6 km/s at around 45 km depth and reaches to a maximum of 4.75 km/s having depleted/preserved layer ($V_s \geq 4.7$ km/s and $\delta V_s \geq 4\%$) at shallow depth (between 45-100 km depth) being thickest (~ 60 km thick) beneath the Coast-WG domain while getting thinner from WDC to half of the CB (30-40 km thick) while the remaining eastern part of the CB and EG has depleted layer having 10-20 km thickness. Below these layer, the shear wave velocity progressively decreases to 4.6 km/s ($\delta V_s \sim 2\%$) between 100-120 km depth beneath Coast-WG and western half of the WDC while it gets progressive deeper towards east having 4.6 km/s between 120-190 km depth till the half of the CB being deepest (160-180 km depth) beneath the EDC and gets drastically shallower as it reaches from CB to EG between 80-90 km depth. The shear wave velocity further decreases to 4.5-4.4 km/s beneath the Coast-WG domain, western part of the WDC; and the EG between 140-220 km and 80-170 km depths respectively showing strong low velocity partial melt (asthenosphere). The shear wave velocity reaches minimum ~ 4.55 to 4.6 km/s from the eastern part of the WDC till the half of the CB covering EDC showing very weak asthenosphere/highly viscous melt. This can be interpreted as the possible presence of tectosphere (Jordan, 1978 & 1979). $V_s \sim 4.6$ km/s or $\delta V_s \sim 2\%$ faster with respect to the global velocity model AK135 is being used as the proxy to estimate the lithospheric thickness or the mark for the Lithosphere-Asthenosphere Boundary (LAB).

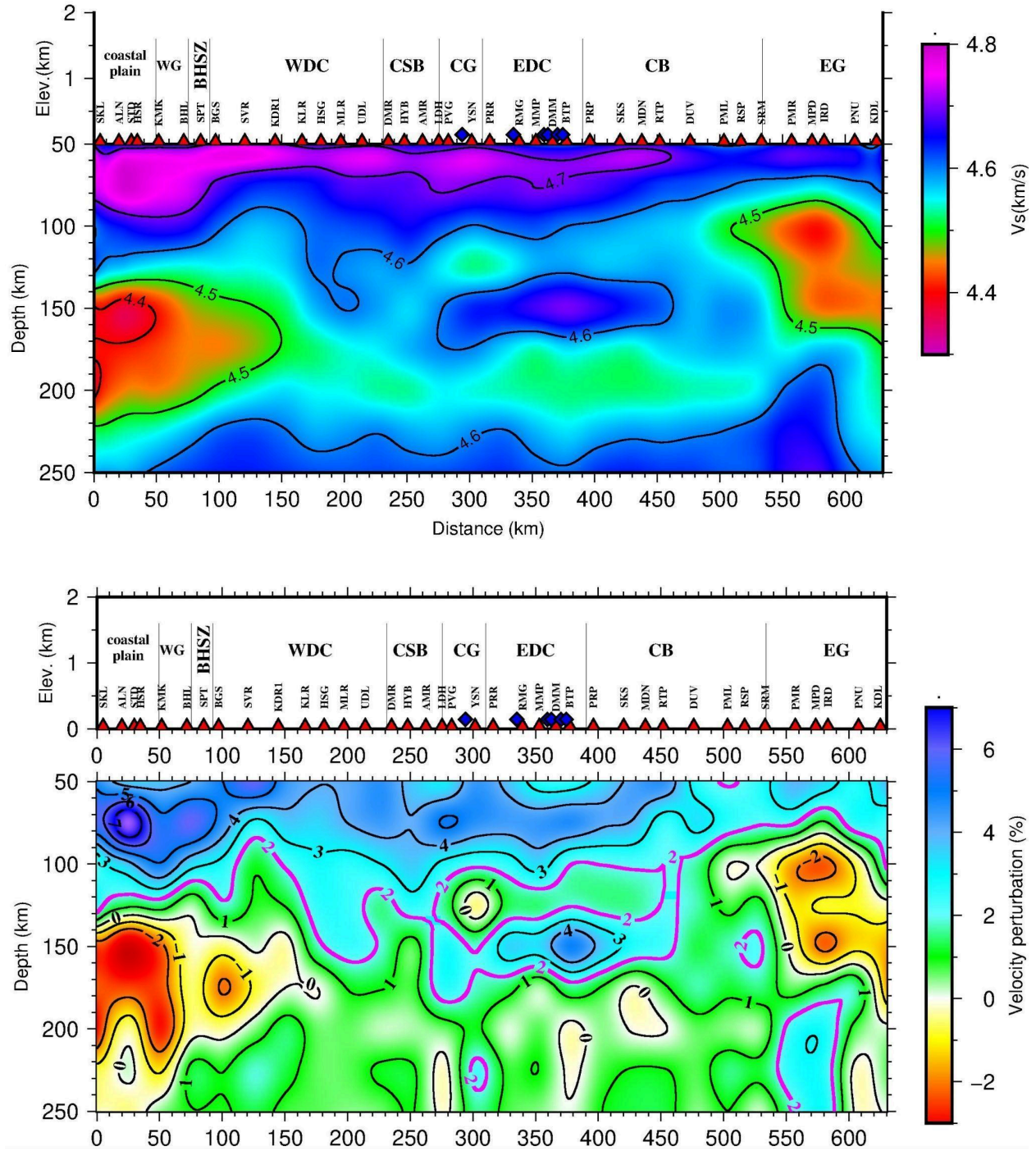


Fig. 3.15. Velocity-depth section of the uppermost mantle along the E-W profile (station locations map shown in Fig. 1.4) in the depth range of 50-250 km in terms of absolute shear velocity (**Top**) and shear velocity perturbation relative to AK135 (**Bottom**).

3.3.3 Comparison with global craton average, composition and geotherm -

We compare upper mantle velocity models of the stations beneath the Dharwar with the other cratons (Lebedev et al., 2009) and AK135 in Fig. 3.16 (Left). In the uppermost mantle, shear velocity beneath the Dharwar is ≥ 4.7 km/s at depth 45–70 km, which is higher than that of many cratons and becomes almost equal within the global craton range in the depth 60–100 km (Lebedev et al., 2009) (Fig. 3.16 Left). Between 100–150 km depth, the shear velocity average beneath the Dharwar is slightly less than the global craton model and in comparison, at > 150 km depth, the velocity is equal to or within the global craton model. The lower velocity bound shows the velocity beneath the western half of the Western Dharwar partially affected due to the Marion plume (~ 85 Ma). Otherwise the remaining part of the Dharwar is almost equal to the global craton model. We have taken velocity models from three representative domains: shear velocity (V_s) average with bounds beneath the Dharwar, one from western domain (Coast-WG) and the other from eastern domain (EG). The western domain shows shear velocity ≥ 4.7 km/s at depths 45–100 km, which is higher than that of many cratons (Lebedev et al., 2009) and shows V_s between 120–200 km depth being lower with respect to the global reaching ~ 4.3 km/s while the eastern domain shows lower V_s between 70–170 km depth reaching ~ 4.4 km/s. The low velocity ≤ 4.5 km/s here suggests strong partial melt/supercritical fluid due to the different tectonic events that happened in the past. These velocity characteristics are discussed using theoretical models computed for varying compositions (Fig. 3.16 Middle) and thermal gradients in the Earth (Fig. 3.16 Right).

The velocity-depth model computed for lherzolitic to hydrated peridotite mantle composition (Bruneton et al., 2004) predicts a maximum V_s of 4.8 km/s beneath the Moho at 50 km that progressively decreases (4.65–4.75 km/s) with depth due to gradual increase of temperature. For a moderately metasomatized peridotite which contains phlogopite or amphibole, the V_s at 50 km depth is 4.5–4.65 km/s, progressively decreasing to 4.3–4.4 km/s. The average velocity models beneath the Dharwar match well with the theoretically predicted velocity for garnet lherzolite/websterite composition for the whole upper mantle and its shear velocity lower bound just touching the amphibole peridotite (Olivine-59%) composition on certain depth range in the lower lithosphere suggesting weak/lack of asthenosphere or very weak low velocity zone (LVZ). While the representative station in the western end of the Dharwar matches well with the Garnet

harzburgite/wehrlite composition between 50-130 km depth, its minimum velocity at a depth of 130–220 km could correspond to the asthenosphere and the eastern end of the Dharwar matches well with the Garnet Wehrlite between 50-80 km depth below which its minimum velocity at a depth of 80–170 km could correspond to the asthenosphere. So except the Dharwar having weak/lack of asthenosphere suggesting possible presence of tectosphere, other domains show a strong asthenosphere/partial melt.

In Fig. 3.16 (Right), we compare the observed shear velocity beneath the three representative domains with thermal gradients corresponding to a surface heat flux of 30, 40, and 50 mWm^{-2} . In the whole upper mantle, all the models compare to the theoretically computed velocities for surface flux 30–40 mWm^{-2} as observed over the Dharwar (30 mWm^{-2} for WDC and 40 mWm^{-2} for EDC). While the western and eastern end representative velocity models match with the theoretically computed velocities for surface flux 30–40 mWm^{-2} between 50-130 km depth for western end and 50 mWm^{-2} below 80 km depth for eastern end respectively.

While most velocity models based on different mantle compositions predict a decrease in velocity with depth, the domains covering the Dharwar craton along the E-W profile, as well as many other cratons, show an increase in V_s with depth in the shallow mantle. This phenomenon is attributed to the phase change from spinel peridotite to garnet peridotite at around 80 km depth (Hales, 1969). In the presence of Cr, this phase transition occurs at a higher pressure and over a broader depth interval (O'Reilly and Griffin, 2006; Lebedev et al., 2009). Ziberna et al. (2013) demonstrated that for all possible upper mantle compositions, the spinel-to-garnet mineral transformation in a divariant (P-T) field is gradational. The transitions occur from a spinel-only field to a spinel + garnet coexistence field to a garnet-only field, within a pressure range of 1 to 6 GPa and a temperature range of 600° to 1400°C.

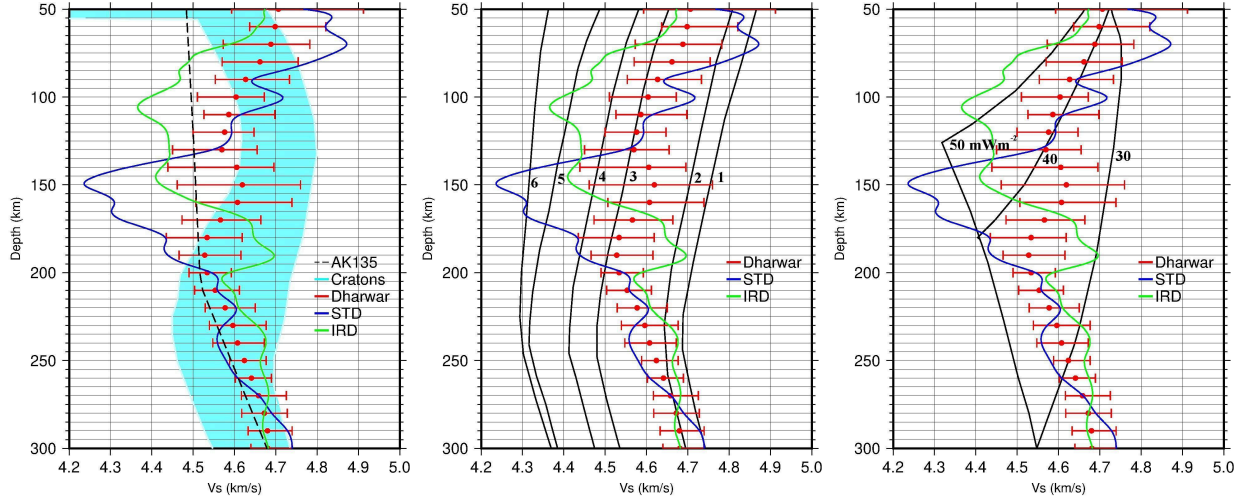


Fig. 3.16. Shear velocity-depth models (average with the bounds) for the Dharwar and at selected stations from western and eastern ends compared with **(Left)** velocity model for selected cratons of the world along with IASP91. The shaded region show summary profile of isotropic-average shear speed V_s beneath cratons showing best-fitting profiles from all the cratonic locations (Lebedev et al., 2009), **(Middle)** theoretically computed velocity models for different mantle compositions; 1- Garnet-Spinel harzburgite (Olivine-70%), 2-Garnet lherzolite (Olivine-64.2%), 3-Garnet wehrlite (Olivine-75.6%), 4- Amphibole peridotite (Olivine-59%) and 5-Amphibole peridotite (Olivine-50%) zone (Bruneton et al., 2004) and **(Right)** theoretical shear wave velocity-depth model for continental thermal gradient corresponding to surface heat flow of 30, 40 and 50 mWm^{-2} assuming undepleted peridotite composition of the mantle lithosphere (Goes et al., 2020).

The properties of rocks that influence the seismic wave velocity including elastic moduli, viscosity and strength are primarily determined by pressure, temperature, lithology (composition), and water content (H_2O) (Karato, 2008). Among these factors, temperature and composition are the most significant. However, the relative contributions of these key factors to seismic anomalies in the continental lithospheric mantle remain a topic of debate (Forte & Claire Perry, 2000; Godey et al., 2004; Artemieva, 2009). The geotherms for surface heat flow are observed in WDC ($\sim 30 \text{ mW/m}^2$) and EDC ($\sim 40 \text{ mW/m}^2$). The minimum seismic velocity is anticipated to coincide with the depth at which the conductive geotherm cuts (intersects) the mantle adiabat. (Sleep, 2003; McKenzie et al., 2005; Michaut et al., 2007). It is evident that the velocity-depth variation in our study area cannot be solely attributed to thermal variations. The

thickness and composition of the lithosphere have significant implications for the longevity of cratons.

3.3.4 Lithospheric Thickness Variability -

Determining the lithosphere thickness in cratonic regions using velocity variation with depth is often challenging and ambiguous due to the slow temperature increase with depth while its relatively easier but still challenging to estimate the lithospheric thickness of the younger mobile belt/terrains having relatively larger negative velocity gradient in the lower lithosphere. The vertical shear wave velocity (V_{sv}) is more sensitive to the temperature variations than to compositional changes. This property makes V_{sv} a useful parameter for mapping the thickness of the lithosphere (Jordan, 1979; Lee, 2003; Priestley & McKenzie, 2006; Schutt & Lesher, 2006). In cratons, the lithosphere-asthenosphere boundary (LAB) is expected to be marked by only a subtle change in the temperature gradient and seismic velocity with depth, in contrast to the well-defined sharp low-velocity boundary in younger terrains (Eaton et al., 2009; Fischer et al., 2010). The lithosphere-asthenosphere boundary (LAB) is delineated using various proxies derived from surface wave models of V_{sv} (vertical shear wave velocity), including (i) the point where the velocity drop is most pronounced underneath the high-velocity lid; (ii) the average depth of the negative velocity gradient located beneath the high-velocity lid; (iii) a specific iso-velocity surface, where the velocity profiles of oceanic and continental regions converge (Rychert et al., 2020); and (iv) the upper and lower boundaries of the velocity decrease, presuming a gradual transition. The poor velocity contrast at the LAB in cratons leads to a low amplitude negative pulse in the P-receiver function, which can be masked by multiples from the Moho. Alternatively, S-to-P conversion in the receiver function is a better tool for mapping the LAB because it avoids interference from Moho multiples/conversion with the LAB signature (Fischer et al., 2010). However, the signal-to-noise ratio for the Sp phase is generally lower than those for Ps. The S receiver function requires a significantly large number of waveforms and careful analysis to extract a stable conversion and avoid side lobes from the Moho in the processed data (Kind et al., 2020; Krueger et al., 2021). Due to the short-term deployments and limited data, we avoided analyzing the Sp signal for LAB computation. Instead, surface heat flow data has been used to map the depth to the potential temperature of 1300°C, representing the LAB beneath the continent (Artemieva, 2009 and 2011).

We defined the lithosphere-asthenosphere boundary (LAB) at a depth where the velocity is 2% greater than that of the global Earth model AK135, aligning with a negative conversion in the SdP CCP (Artemieva, 2009; Darbyshire et al., 2013; Porritt et al., 2015). This depth corresponds to a shear wave velocity (V_s) of approximately 4.6 km/s. It is important to note that 4.5 km/s represents the average shear wave velocity of global Earth models at depths between 100 and 200 km, as well as the minimum velocity observed in the global craton model (Lebedev et al., 2009). The mantle, composed of peridotite, exhibits a shear wave velocity of 4.56 km/s at a depth of 150 km (Bruneton et al., 2004), which gradually decreases to 4.5 km/s at 200 km (Debayle and Ricard, 2012). Utilizing the 2% velocity criterion, we delineate the LAB at depths 160-180 km underlain by a weak melt zone across the Dharwar (except the western end) and the CB, as indicated by the thick pink line in the lower part of Fig. 3.15 bottom. In contrast to the Dharwar and CB, the lithosphere beneath the Coast-WG domain and the EG is relatively thinner having 100-120 km and 80-90 km respectively.

We also employed an alternative method that uses the depth to the minimum velocity as a proxy for the lithosphere-asthenosphere boundary (LAB), as suggested by Goes et al. (2012). This approach offers a practical definition of the thermal thickness of the lithosphere, even in cases of transitional lithosphere-asthenosphere boundaries. The thermal thickness is determined at the depth where the conductive geotherm intersects with the mantle adiabat (Sleep, 2003; McKenzie et al., 2005; Michaut et al., 2007). For steady-state geotherms, this intersection corresponds to a minimum in seismic velocity (Goes et al., 2020). Based on this criterion, we mapped the thermal LAB at depths of 150–180 km beneath the Dharwar Craton and around 100–120 km beneath the Cuddapah basin. The initial lithospheric thickness was at least 120 km beneath the Cuddapah basin as suggested by Anand et al., 2003 which means the thickness can be more than 120 km beneath the CB and may be underlain by the lithospheric part of the EDC beneath the CB or else the lithosphere may get reocratonized or healed with the time (Liu et al., 2021) by the different tectonic events happened in the east. Numerical modeling results by Liu et al. (2021) indicate that mantle plumes can effectively fill areas of thin lithosphere by upwelling highly melt-depleted, high-pressure mantle residues. The LAB boundary along the E-W profile is indicated in Figure 3.15 bottom **as a thick dashed line in purple**. Maximum thermal lithospheric thickness of ~180 km (Negi et al., 1986) to more than 200 km (Gupta et al., 1991) indicates the existence of deep continental lithospheric root beneath the Dharwar Craton (Niraj

Kumar et al., 2013). The seismic lithosphere thickness mapped beneath the CB varies from 80 km in contact with the EG and reaches 170-180 km thick in contact with the EDC. So less than half of the CB in contact with the EG is affected by different tectonomagmatic events that happened in the eastern margin in the past. The thermal LAB for the Coastal plain-WG and EG are estimated to be 90-130 km and 90-120 km respectively.

By applying the aforementioned definition of the lithosphere-asthenosphere boundary (LAB) of the Dharwar, we interpret the 160-180 km thick lithosphere beneath the Dharwar except the western end. This correlates well with: i) a range of 150–180 km determined using Sp conversion at the nearby HYB station (Krueger et al., 2021); ii) a lithospheric thickness of 155–170 km reported by Mitra et al. (2006), Pasyanos et al. (2014), and McKenzie et al. (2015) beneath the Dharwar; iii) a thickness of 130–190 km derived from SS precursors and the deepest origins of diamonds (Tharimena et al., 2017); iv) a global lithospheric thickness of 150–250 km beneath cratons based on cluster analysis of seismic tomographic models (Schaeffer & Lebedev, 2015) or lithospheric thermal properties (Artemieva, 2006); and v) lithospheric thickness of 160–190 km has been inferred from xenoliths (rock fragments) found in the diamond-bearing kimberlite field (Ganguly & Bhattacharya, 1987; Shaikh et al., 2020). Also, our findings are consistent with the global observations indicating a lithospheric thickness of 150–200 km over cratons, as determined by heat flow and surface wave studies (Nataf & Ricard, 1996).

3.3.5 Layered composition of the lithosphere and identifying Archean core -

The cratonic mantle is stabilized and survived through a long time due to positive buoyancy caused by extensive depletion and high viscosity. Jordan (1988) suggested that the stabilization of cratonic roots is attributed to their composition. This idea is reinforced by findings from many subsequent studies indicating that the Archean cratonic mantle is highly viscous and positively buoyant due to significant depleted basaltic components (Kelley et al., 2003). The depletion decreases (in other words, fertility increases) with increase in depth due to temperature, influx of a variety of fluids such as water, CO₂, silicate, carbonatite and sulfide melts etc. This highly hydrated peridotite represents the metasomatized mantle. The representative velocity models from Dharwar and its western and eastern end are compared with the velocity-depth models computed from the selected xenoliths from eastern Finland and hydrated peridotites (Bruneton et al., 2004) as well as compared with the global craton model (Lebedev et al., 2009). The

velocity-depth models computed for the upper mantle having harzburgite/lherzolite to hydrated peridotite composition. The models predicted range from 4.8 km/s below the Moho at 50 km to 4.65–4.75 km/s with the increase in depth due to increase in temperature to the base of the lithosphere (Bruneton et al., 2004). For a moderately metasomatized peridotite composition which includes phlogopite or amphibole, the V_s at 50 km depth ranges from 4.5 to 4.65 km/s, also decreasing with depth. Arndt and Goldstein (1989) suggested a composition that includes olivine, clinopyroxene, and garnet restite, resulting in a shear wave velocity (V_s) of 4.65 km/s at a depth of 50 km and 4.55 km/s at a depth of 250 km. The average velocity models beneath the Dharwar match well with the theoretically predicted velocity for garnet lherzolite/websterite composition for the whole upper mantle and its shear velocity lower bound just reaching the amphibole peridotite (Olivine-59%) composition on certain depth range in the lower lithosphere suggesting weak/lack of asthenosphere or very weak low velocity zone (LVZ). While the representative station in the western end of the Dharwar matches well with the garnet harzburgite/wehrlite composition between 50-130 km depth, its minimum velocity at a depth of 130–220 km could correspond to the asthenosphere and the eastern end of the Dharwar matches well with the garnet wehrlite between 50-80 km depth below which its minimum velocity at a depth of 80–170 km could correspond to the asthenosphere. So except the Dharwar and CB, other domains show a strong asthenosphere suggesting possible presence of tectosphere may have deeper roots. Our inference of a compositionally layered lithosphere is also consistent with the findings of Tainton and McKenzie (1994) utilized the trace element composition of kimberlites and other alkaline rocks to demonstrate that their source regions had initially been depleted by approximately 20% melting in the garnet stability field and subsequently enriched by a metasomatic melt rich in light rare earth elements (REE) and other incompatible elements.

The Proterozoic Cuddapah Basin (CB) underwent considerable magmatic activity approximately 1.8 billion years ago, aligning with its developmental phase (Chalapathi Rao et al., 2004). Analyses of rare earth elements (REE) indicate that the initial melts of the Cuddapah were produced through about 10–15% partial melting of a lherzolite mantle source. The mechanical boundary layer thickness is estimated to be 70 km, with an initial lithospheric thickness of at least 120 km (Anand et al., 2003), which correlates with the high seismic wave velocities observed in the region.

Almost all the domains show seismic velocity with the exception of a more pronounced 4.7 km/s layer in the shallow lithosphere having the thinnest layer beneath the EG. We deduce the highest-velocity cores (and consequently, the lowest temperature regions) of the cratonic mantle lithosphere by identifying a depth of positive 4% V_s anomaly (Celli et al., 2020), which corresponds to a velocity of 4.675 km/s, as illustrated in Fig. 3.15 (bottom). This core of the mantle lithosphere extends down to around 100 km depth below the Coast-WG domain having ~60 km thickness which gets thinner having ~30-40 km thickness from WDC to western half of the CB and thinnest having ~10-20 km thickness beneath the EG. This high velocity layer potentially indicates a pristine unmodified cratonic structure. These values are close to the highest shear velocities reported globally for cratons (Hirsch et al., 2015; Garber et al., 2018; Ravenna et al., 2018).

The compositional variations inferred from mantle xenoliths found in kimberlites typically do not alter the velocity by more than 1%. Seismological studies that account for the radial anisotropy of minerals indicate that the high V_s (approximately 4.7 km/s) observed beneath cratons cannot be solely attributed to peridotite in the depth range of approximately 100–170 km (Hirsch et al., 2015). Instead, an increased presence of garnet or eclogite in the mantle is necessary to account for such high velocities (Garber et al., 2018). Petrological modeling suggests that the high V_s (> 4.7 km/s) found in cratonic cores may require a small proportion of eclogite ($<20\%$) and diamonds (2%) within the cratonic mantle lithosphere (Garber et al., 2018). The high velocities ($V_s > 4.7$ km/s) in the upper lithosphere along the profile (except EG) may be attributed to a harzburgite-dunite composition.

The lower lithosphere (defined as the layer between the depth where $V_s < 4.7$ km/s and the LAB) exhibits distinct variations from west to east coast covering Dharwar craton. The thickness of this layer is thinner beneath the Coast-WG and western part of the WDC domain and gets thicker towards the east until it gets maximum thickness beneath the EDC-CB domain and again getting drastically thinner beneath the EG. Surface heat flow in this region varies from 30 mW/m² beneath the WDC to 40 mW/m² beneath the EDC and is higher (> 40 mW/m²) in the western (WG) and eastern ends (EG) (Roy and Rao, 2000). A seismic tomography study by Rai et al. (1992) indicates a thin lithosphere in the Eastern Ghats (EG). Surface heat flow measurements in the EG range from 64 mW/m² in the north to 104 mW/m² in the south (Rao and

Rao, 1983). This heat flow is higher in the EG compared to neighboring Archaean cratons, which show values of 30–40 mW/m² in the Dharwar craton (Roy and Rao, 2000) and 52–64 mW/m² in the Bastar craton (Gupta et al., 1993).

Xenoliths reveal that the upper portion of the cratonic lithosphere, both globally and in the Dharwar craton, is more depleted than the deeper roots (Ganguly and Bhattacharya, 1987; Griffin et al., 2009; Shaikh et al., 2020). This observation is consistent with models of thermal structure, petrology, and geoid data (Afonso et al., 2008). Geophysicists often interpret lateral variations in seismic velocity as indicative of temperature differences; however, up to 20% of the observed range in S-wave velocity at depths of 100–175 km in the continental mantle can be attributed to compositional variations (Deen et al., 2006). The metasomatic refertilization of the lithospheric mantle increases its density and reduces its seismic velocity due to the influx of carbonatic fluids and the elevated thermal gradient associated with heat-producing elements (K, U, Th) that accompany metasomatic enrichment but here beneath the Dharwar craton, the metasomatic refertilization is almost absent or may have highly viscous fluids whose velocity is comparable with the surrounding layers. The metasomatic refertilization is more prominent and stronger beneath the WG and EG respectively having extensive lithospheric thinning due to the above mentioned mechanisms.

3.3.6 Dharwar lithosphere - preserved or modified -

By employing a joint inversion of surface wave phase velocity dispersion and receiver function data, we have determined the present day lithospheric thickness beneath the Dharwar and CB to be 160–180 km, except its western end in contact with WG affected by Marion plume (~85 Ma). This measurement is within the range of 160–200 km lithosphere thickness estimated from kimberlite xenoliths in the region (Ganguly and Bhattacharya, 1987; Griffin et al., 2009; Shaikh et al., 2020). Regarding composition, the presence of rock types such as peridotite, lherzolite, harzburgite, wehrlite, dunite, and eclogite in the kimberlite xenoliths (Ganguly and Bhattacharya, 1987; Patel et al., 2009) suggests that the lithosphere resembles the current harzburgite-dunite composition of the upper lithosphere beneath the Dharwar, as indicated by Vs values exceeding 4.7 km/s in our velocity model. The thickest layer (~60 km) having high Vs values (> 4.7 km/s) in the upper lithosphere is observed beneath the WG indicating pristine Archean lithosphere and/or iron and other incompatible elements being leached out which is

supported by iron and other economic deposits present in this region while this layer ($V_s > 4.7$ km/s) is thinnest beneath the EG suggesting most of the lithosphere is modified and/or affected by the tectonic events happened in the past in this region. In the lower lithosphere (the lowermost ~90-110 km thick layer), shear velocity decreases from 4.7 km/s to 4.6 km/s, suggesting depth-dependent metasomatism (Eeken et al., 2018) or modification of the lithosphere, a pattern observed in most cratons, albeit with varying magnitudes. This lower lithosphere is thickest beneath the EDC and CB suggesting metasomatism (having lesser amount of melt %) associated with different tectonomagmatic events that happened in the EDC and CB. The lower lithosphere is thinnest in both the ends (WG and EG). The upper lithosphere experiences only thermally related velocity perturbations, while the velocity in the lower lithosphere may also be influenced by metasomatism. Furthermore, we find evidence of a shallow mantle lithosphere discontinuity (possibly MLD) beneath the Dharwar and CB which is discussed in detail in the subsequent sections. The differences observed in the lower lithosphere are attributed to different degrees of metasomatism and the presence of fluids resulting from different tectonomagmatic events of different degrees.

Using xenolith samples from the Wajrakarur kimberlite field, Ganguly and Bhattacharya (1987) identified a systematic positive correlation between $\text{Cr}/(\text{Cr} + \text{Al})$ and $\text{Mg}/(\text{Mg} + \text{Fe})$ ratios in garnet, indicating progressive partial melting of the xenoliths with depth. They also noted a consistent relationship between the degree of refractory nature and the inferred depth of origin of the xenoliths, utilizing the variation of $\text{Cr}/(\text{Cr} + \text{Al})$ as a diagnostic tool. Their findings suggest that the refractory component of garnet decreases with increasing depth, implying that the mantle becomes more fertile at greater depths, which may explain the observed reduction in velocity ($V_s < 4.6$ km/s) in the lower lithosphere. Pintér et al. (2022) provide a comprehensive explanation for the presence of a metasomatized zone in the cratonic lower lithosphere, based on experiments where oxidized, hydrous, carbonate-rich melts react with reduced, depleted peridotite at 5 GPa pressure. They argue that the thick mantle lithosphere beneath cratons consists of strongly reduced rocks that have interacted with oxidized melts. These low-silica, incipient melts are rich in CO_2 and H_2O , and they react with surrounding rocks to form an enriched zone at the base of the lithosphere, which serves as the source region for many diamonds but these melt % is either less and/or more viscous having velocity ($V_s \sim 4.55\text{-}4.6$ km/s) comparable to the surrounding

layers. WG having minimum $V_s \sim 4.3$ km/s at 150-170 km depth has more melt % and/or less viscous melt than the EG having minimum $V_s \sim 4.4$ km/s at 100-140 km depth.

We critically evaluate the findings regarding the thickness and composition of the Dharwar craton in light of potential alterations that may have occurred due to interactions with the multiple tectonomagmatic events from mafic dyke swarms (~ 2.3 Ga), Cuddapah Basin formation (~ 1.8 Ga), kimberlite volcanism (~ 1.1 Ga & 90 Ma), Marion plume (~ 85 Ma) and Reunion plume (~ 65 Ma) and the different tectonic events (~ 130 Ma) in the Eastern Ghats region. Analysis of a global compilation of seismic wave velocity and its modeling in relation to lithospheric thickness (Steinberger and Becker, 2018; Pearson et al., 2021) indicates that cratons typically have a lithospheric thickness ranging from 150 to 255 km, covering about 60% of the continent. Many ancient cratons, including the North China Craton, Southern Africa, and the North Atlantic Craton, have experienced partial removal of thick continental lithosphere, which has been associated with thermo-mechanical erosion caused by mantle plumes. Our study suggests that the western end of the Dharwar connected to the WG is also affected by the Marion plume (~ 85 Ma) and significant thinning has happened while over the past ~ 1.1 Ga, the remaining part of WDC, EDC and CB show the lithospheric thickness slightly decreased from 160 to 200 km inferred from xenolith studies to 160-180 km as determined by seismology. We investigate the possible causes for this change. Numerical models of lithosphere-plume interaction (Wang et al., 2015) demonstrate significant variations in lithospheric destruction based on the composition and strength of the lithospheric root. These models indicate that the thick cratonic roots, stable for billions of years, are not significantly impacted by the mechanical erosion of a plume, and that a low degree of metasomatism does not necessarily alter the dynamic stability of the entire continental root. These findings are consistent with those of Thoraval et al. (2006) and Sleep et al. (2003), who proposed that the ancient continental lithosphere has thinned by 30 km and 12 km, respectively, due to mantle plumes. However, if ancient continents have a punctuated history of metasomatic enrichment, plumes can remove the lithospheric root by approximately 100 km within tens of millions of years, as observed in the North China, Omolon, Wyoming, Yangtze, and North Atlantic cratons. In the WDC, except the western end, there have been no significant tectonic activities reported since cratonization and mafic dyke swarms at ~ 2.3 Ga, and the region lacks a thick metasomatic zone (V_s in the range 4.6-4.7 km/s) in the lower lithosphere while the EDC has undergone different tectonomagmatic

events from the cratonization (~2.5 Ga), Cuddapah Basin formation (~1.8 Ga) to the kimberlite volcanism (~1.1 Ga and ~90 Ma). This may explain why the EDC has experienced minimal effects from interactions with mantle plumes over the last 130 Ma following its separation from eastern Gondwanaland. Beneath the kimberlite domain, there is slightly thick but weak metasomatism at the base of the lithosphere, which, as previously discussed, would not disrupt the dynamic stability of the continental root. This supports the notion that the Indian continent has been drifting with its deep cratonic root (Alvarez, 1990; Yoshida and Yoshizawa, 2021).

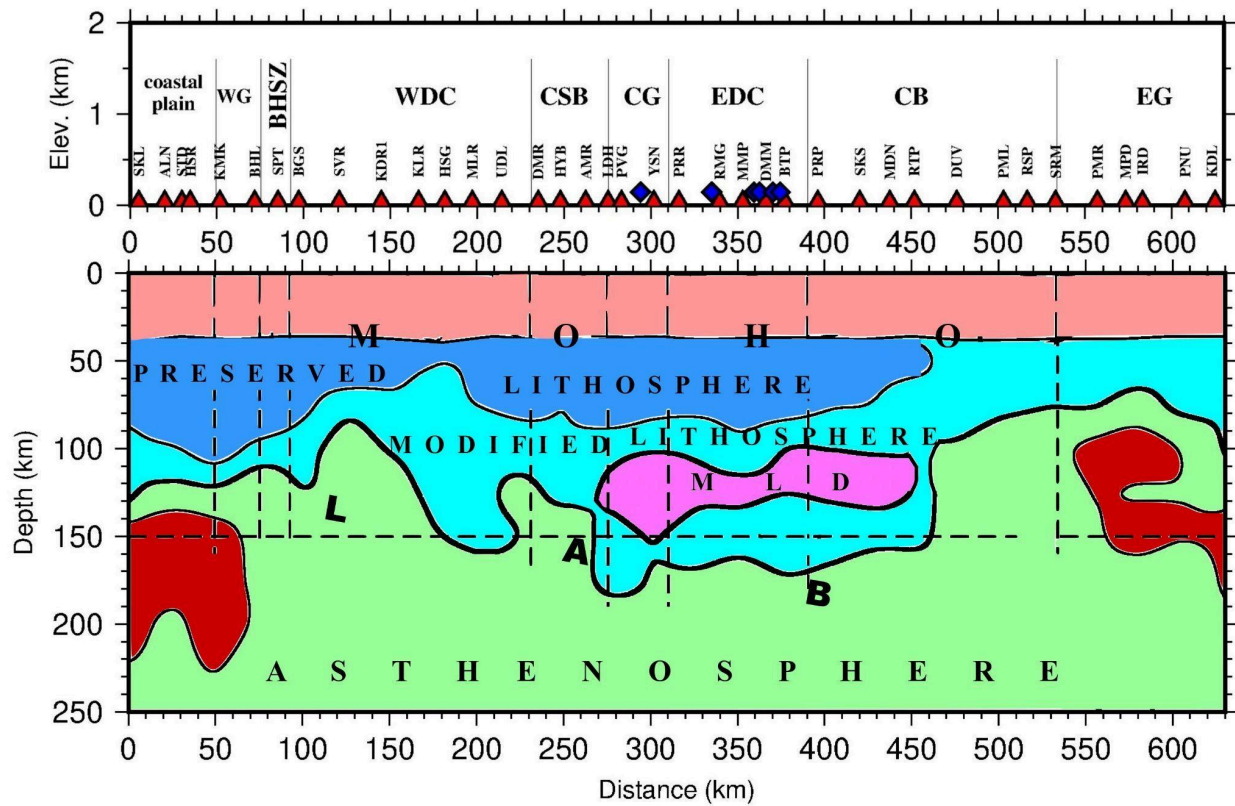


Fig. 3.17. Interpretive geological model along the E-W profile derived from the shear velocity-depth model along the profile presented in Fig. 1.4. LAB at a depth corresponding to velocity 2% faster (~4.6 km/s) than the global earth model AK135 is shown as a thick black line.

3.3.7 Mid lithospheric discontinuity beneath the Kimberlite field -

While the negative velocity gradients located beneath stable cratons at depths of approximately 170 to 250 km are typically understood to represent the lithosphere-asthenosphere boundary (LAB), seismic discontinuities that have a thickness of around 30 to 40 km and are

mainly found between 80 and 120 km depth beneath continental lithosphere (Rychert and Shearer, 2009; Thybo, 2006) have been interpreted as MLDs (Fischer et al., 2010; Ford et al., 2010; Selway et al., 2015; Yuan and Romanowicz, 2010). In certain instances, cratonic MLDs are continuous, while in other cases, they are truncated, suggesting the presence of trans-lithospheric boundaries that reflect terrane sutures at depth (Courtier et al., 2010). The MLD is attributed to **(1)** existence of partial melt (Thybo, 2006), **(2)** layering in seismic anisotropy (Yuan and Romanowicz, 2010), **(3)** compositional layering resulting from a decrease in bulk Mg# with depth (Karato et al., 2015; Selway et al., 2015), **(4)** elastically accommodated grain boundary sliding (EAGBS) (Karato et al., 2015), and **(5)** presence of hydrous or carbonate mineral phases formed through metasomatism (Selway et al., 2015; Saha et al., 2020).

We schematically show the cartoon figure of nature of the lithosphere and MLD in the diamondiferous kimberlite field (Fig. 3.17). Our result shows a 160-180 km thick lithosphere beneath the kimberlite field with the distinct cratonic lithosphere mantle ($V_s \sim 4.8\text{-}4.6$ km/s). We observe a distinct signature of MLD across the diamond-bearing kimberlite field between 100-120 km depth below the diamondiferous kimberlite field along the E-W profile (Fig. 3.15 and 3.17). Due to the cold and stable nature of cratonic roots, it is improbable that there are present-day global melts at MLD depths. The favored explanations focus on the accumulation of seismically slow metasomatic minerals, which are abundant in hydrous minerals, pyroxenes, and/or carbonates, as well as the development of fabric resulting from deformation during the construction of cratons influenced by plume or collisional tectonics (Aulbach et al., 2017; Zhang et al., 2024).

Chapter 4 - Conclusions

In this thesis, attempt has been made to answer the questions related to structure, composition and evolution of the Earth's crust and the lithospheric mantle and its variability among geological domains of different ages from Archean to Proterozoic along a profile using shear wave velocity (V_s) as a proxy derived from the joint inversion of surface wave phase velocity dispersion with improved resolution and P-receiver function. Our study area is a ~630 km long E-W profile has 38 dense broadband seismic stations having 10-15 km spacing which covers mid-late Archean Dharwar craton having ~ 1.1 Ga old diamondiferous kimberlite fields and the Proterozoic Cuddapah basin. The western and eastern ends of the profile are affected by the mantle plumes at ~85 Ma and ~130 Ma respectively. So, it is an ideal natural laboratory to study the above issues with the seismological data having significantly improved resolution.

Some of the issues discussed in the thesis include:

- (i) The relationship among the crustal characteristics like Moho depth, crustal composition, age, and surface geology, the similarity or diversity in the nature of the middle and lower crust across different terrains of varying ages, any signatures of extensive underplating due to the influence of mantle plumes that happened in the past and crust-mantle interaction.
- (ii) The nature of the Archean cratons – their preservation, modification and possible destruction and comparing their lithospheric structure with those with the kimberlite domains derived from xenoliths, difference between Archean and Proterozoic terrains and influence of mantle plumes.

Despite a number of studies along this profile, these issues remained unanswered or poorly investigated in terms of spatial and vertical resolution of the velocity structures. In this thesis, we generated a significantly improved shear wave velocity image of the crust and lithospheric mantle using improved surface wave dispersion (SWD) data at shorter periods having data points at every ~10 km interval and combined them with the longer period data at every ~55 km interval along the profile. Shear wave velocity down to 400 km depth along the profile is computed by the joint inversion of P-receiver function and Rayleigh wave phase velocity dispersion having maximum period at 140 s corresponding most reliability at around 210 km depth and reliability decreases below it. So, we have taken the velocity model only upto 300 km depth as reliable for final results to image the lithosphere. Common conversion point (CCP)

stacking of the receiver function for the crust is also done along this profile to provide additional constraints in the result. The V_p , density (ρ) and shear modulus (μ) for the V_s along the profile are computed using the empirical relationship of Brocher (2005) discussed in chapter 3.

Nature of the crust - Depth and seismic wave velocity are the key parameters to define the composition of the crust and the underlying mantle. Along the E-W profile, we observe flat Moho (38 ± 2 km thickness) having average velocity (V_s) in the range 3.70-3.75 km/s in all the domains of the profile, suggesting the average composition of the crust to be felsic to intermediate. The Moho depth along the profile is consistent with the CCP image (Dueker and Sheehan, 1998) using Gaussian Width (GW) ~ 2.5 for V_p/V_s of 1.73 which is unchanged with varying the V_p/V_s . We don't find any correlation of crustal structure with surface boundaries across the Dharwar suggesting homogenized crust.

The E-W profile lies over the greenschist-amphibolite facies transition zone. So, the upper crust is almost absent (~ 2 -3 km thickness) upliftment and erosion level of 8-10 km of the amphibolite facies rocks corresponding to 2-3 kb pressure. The EDC shows a more felsic composition (having middle and lower crust ~ 26 km and ~ 4 km thickness respectively) than the Coast-WG and WDC-CG domains (having middle and lower crust 20-22 km and 8-10 km thickness respectively) and also than the global average. This indicates that the Coast-WG and WDC-CG domains have been affected by the past mantle plume event like Marion plume (~ 85 Ma). Presence of a thin lower crust beneath the EDC leads to sharp velocity contrast at the crust-mantle boundary which is in contrast with the thicker lower crust of Coast-WG and WDC-CG domains having more gradation in the velocity and lower impedance contrast at the crust-mantle boundary.

EDC-like behavior in terms of middle and lower crust thickness is observed beneath the CB and EG domains. At the EDC-CB boundary, we observe a local upwarping or dipping type feature of middle crust beneath CB which may possibly be the signature of upwelling of mafic material from lower to the middle crust or some crustal reworking. At the CB-EG boundary, an unusual same low velocity layer ($V_p < 6.1$ km/s, $V_s < 3.6$ km/s, $\rho < 2.7$ gm/cc, $\mu < 41$ GPa) can be seen dipping towards CB from EG which may possibly be the signature of Nallamalai Fold Belt (NFB). We don't find a signature of extensive underplating in the western and eastern end due to

India-Madagascar separation (~85 Ma) and the Eastern Gondwanaland (~130 Ma) respectively and also don't observe any correlation of crustal thickness with age along the profile.

Except for some local upwarping or dipping type features of the middle crust beneath the CB, there are no signatures of dipping layers. So, we don't support the two stage subduction-accretion process for the crustal evolution in the Neoarchean Dharwar Craton as suggested by previous studies along the profile. It may also be possible that tectonic deformations like folding, faulting, and crustal tilting initially created relief on the Moho boundary but later got erased by ductile flow showing widespread horizontality (flat) of the Moho as suggested previously by Roy Chowdhury and Hargraves, 1981 which was later questioned by Gibbs (1986). Along the profile, there is no strong intra-crustal layering leading to minimum RF multiples which provided improved resolution of the mantle structure and mapping of the Lithosphere-Asthenosphere Boundary (LAB).

Nature of the lithospheric mantle - The velocity models at individual stations down to 300 km depth are interpolated to generate a smooth shear wave velocity for the uppermost mantle along the profile. Using $V_s \sim 4.6$ km/s or $\delta V_s \sim 2\%$ faster with respect to the global velocity model AK135 as the proxy to demarcate the Lithosphere-Asthenosphere Boundary (LAB), we observe 100-120 km thick lithosphere beneath Coast-WG and western half of the WDC which gets progressive deeper towards east till the half of the CB being deepest (160-180 km depth) beneath the EDC and gets drastically shallower as it reaches from CB to EG between 80-90 km depth.

The lithospheric thickness 160-180 km beneath the EDC and CB when compared with the lithospheric thickness of 180 ± 20 km derived from kimberlite xenoliths erupted at ~1.1 Ga (Ganguly and Bhattacharya, 1987) suggests marginally thinned lithosphere which remained almost undisturbed and unaffected by the tectonomagmatic events happened. This thick lithosphere is underlain by weak/absence (highly viscous) of the asthenosphere ($V_s \sim 4.55$ -4.6 km/s) suggesting possible presence of tectosphere (Jordan, 1978 and 1979). This weak (highly viscous) asthenosphere possibly shields the lithosphere from the plume events resulting in longevity of the Dharwar craton.

The 100-120 km and 80-90 km thick lithosphere beneath the Coastal plain-Western Ghat-Western end of Dharwar and the Eastern Ghat parts show the imprint of India-Madagascar

separation (~85 Ma) due to Marion plume and eastern Gondwanaland (~130 Ma) respectively. The low shear wave velocity ($V_s \leq 4.5$ km/s) is shown beneath the Coast-WG domain, western part of the WDC; and the EG between 140-220 km and 80-170 km depths respectively suggesting strong low velocity partial melt (asthenosphere) due to the intrusion of hydrous melts/supercritical fluids and re-fertilization.

We observe an increase in V_s with depth in the shallow mantle beneath the domains covering the Dharwar craton along the E-W profile possibly suggesting preserved/highly depleted ($V_s \geq 4.7$ km/s; $\delta V_s \geq 4$ %) due to the phase change from spinel peridotite to garnet peridotite at around 80 km depth (Hales, 1969). In the WDC, except the western end, no significant tectonic activities reported since cratonization and mafic dyke swarms at ~2.3 Ga, and the region lacks a thick metasomatic zone (V_s in the range 4.6-4.7 km/s) in the lower lithosphere while the EDC has undergone different tectonomagmatic events from the cratonization (~2.5 Ga), Cuddapah Basin formation (~1.8 Ga) to the kimberlite volcanism (~1.1 Ga and ~90 Ma). This may explain why the EDC has experienced minimal effects from interactions with mantle plumes over the last 130 Ma following its separation from eastern Gondwanaland. Beneath the kimberlite domain, there is slightly thick but weak metasomatism at the base of the lithosphere, which, as previously discussed, would not disrupt the dynamic stability of the continental root. This supports the notion that the Indian continent has been drifting with its deep cratonic root (Alvarez, 1990; Yoshida and Yoshizawa, 2021).

Our result shows a distinct signature of MLD across the diamond-bearing kimberlite field between 100-120 km depth below the diamondiferous kimberlite field along the E-W profile possibly explained by accumulation of seismically slow metasomatic minerals (rich in hydrous minerals, pyroxenes and/or carbonates) and the acquisition of fabric due to deformation during craton construction involving plume or collisional tectonics (Aulbach et al., 2017; Zhang et al., 2024).

Future Aspects from this thesis - In this thesis, we have done high resolution seismic imaging of the lithosphere along the E-W profile and have found many interesting and significant results to understand the structure, composition and evolution of Dharwar Craton and its adjacent areas but more improved inversion techniques can be used. In this thesis, we have used the joint inversion technique which provides a single solution for each inversion and has more human

interference in choosing the modeling and inversion parameters. So, probabilistic inversion like Bayesian providing all possible solutions can be the better tool and can be tried in future which is almost data driven in choosing the modeling and inversion parameters. Also, joint modeling of high frequency P receiver functions and autocorrelograms can be done to enhance higher frequency content in the data to map any possible major deep faults, low angle thrusts and reflectors etc. as found in the previous DSS study (Kaila et al., 1979). Despite numerous studies on Precambrian lithospheric formation and evolution, the deep thermochemical structure of the Dharwar craton and the adjacent areas remains poorly understood. So, investigation of the upper mantle beneath the region can be done by jointly inverting Rayleigh wave phase velocity dispersion data, elevation, geoid height and surface heat flow, using a probabilistic inversion to obtain a 3D model of composition, density and temperature in this area. The study of thermochemical structure of the Dharwar craton and the adjacent areas will have many benefits -

- (i)** High-resolution thermochemical models can inform exploration strategies by highlighting regions with favorable conditions for mineralization or hydrocarbon preservation etc.
- (ii)** Insights from Earth's cratonic thermochemical structure can be used to interpret the evolution of other terrestrial planets, informing comparative planetology and the search for planetary stability elsewhere.
- (iii)** The thermochemical structure can help assess geohazards in terms of localization of seismicity and magmatism. For instance, metasomatized or thermally weakened cratonic margins may be more prone to intraplate earthquakes or volcanic activity.

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