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Gravity Anomalies, Isostasy and Density Structure of the Indian Continental Lithosphere

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Gravity anomalies across the Indian region depict most of the geological and tectonic domains of the Indian continental lithosphere, which evolved through Archean cratonic nucleation, Proterozoic accretion, Phanerozoic India-Eurasia plate convergence, and modification through many thermal perturbations and rifting. Integrated analysis of gravity and geoid anomalies together with topographic and heat flow data led to deciphering the mechanism of isostatic compensation of topographic and geological loads, lithospheric structure, and composition. This study discusses the nature of gravity (free-air, Bouguer and Isostatic) and geoid anomalies in relation to the topography, geology, and tectonics, and presents a lithospheric density model across the peninsular India and Himalaya. Southern peninsular Indian region shows relatively low Bouguer gravity anomalies compared to the northern region. The mobile belts are generally observed to have relatively higher Bouguer gravity anomalies, e.g., Eastern Ghats Mobile Belt compared to the shield regions. The gravity lows are observed over topographic features like the Western Ghats and Himalaya, while some of the topographic highs like Aravalli show positive gravity anomaly. The Indian Ocean Geoid Low varies from -82 m over Dharwar Craton to -98 m over the Southern Granulite Terrain and finally reaches a significant low of -106 m in the Indian Ocean. Flexural isostatic compensation with variable Effective Elastic Thickness (EET) ~10 km to 50 km prevails over the stable continental region. The lithospheric thickness varies from 80 km along the coastal region to 120-130 km beneath the Saurashtra Plateau, the Southern Granulite Terrain, and the Eastern Indian Shield, and reaches to more than 200 km under the Himalayan orogenic belt in the north. From Dharwar Craton to Bundelkhand Craton in central

India, the lithospheric thickness varies between 160 and 180 km.

Introduction

The Indian subcontinent comprises the Precambrian cratonic elements, fold belts, Paleogene continental collision zones with significant Phanerozoic sedimentation in the frontal basins towards the north and is punctuated by numerous thermal perturbations and magmatic pulses (Radhakrishna and Naqvi, 1986; Sharma, 2009; Valdiya, 2016; Chetty, 2017; Roy and Purohit, 2018 and references therein). The formation of early crust, initiation of plate tectonics, and sighting of hidden natural resources still pose a great challenge in the Earth Sciences and are often debated based on geophysical studies of the continental lithosphere (Yuan and Romanowicz, 2019). Despite the extensive earlier work and ongoing research on the above subject, there is still a lot to be discussed on the nature of the deep lithospheric structure and the support mechanism for the elevated plateaus and the orogenic belts of peninsular India and Himalaya. Notably, the depth to the lithosphere-asthenosphere boundary (LAB) remains uncertain despite a wide range of the geophysical studies carried out in the recent past. Partly because the findings of the geophysical proxies are not matching, leading to significant variations in resolution and sometimes inadequately understood by the proxies of specific geophysical methods (Eaton et al., 2009; Artemieva, 2011).

The relation between topography and gravity anomalies over the plateaus and orogenic belts provide insights on the buoyancy and/or the mechanical strength of the lithosphere supporting the topographic load (Watts, 2001; Fischer, 2002). Thus, the gravity anomalies in combination with topography data assist in determining the status of the lithosphere under the premise of isostatic equilibrium, which can be interpreted as a consequence of previous procedures through which the lithosphere obtained its real architecture (Lachenbruch and Morgan, 1990). Density anomalies may be associated with compositional variations in rock materials, as is frequently experienced between upper and lower crustal rocks and between crustal and mantle rocks, but also due to temperature-dependent parameters that may represent lithospheric thickness variations (Zeyen et al., 2005). The combination of gravity and geoidal anomalies can help to distinguish between mass in homogeneities in the mantle and those confined to the crust (Turcotte and Schubert, 2014).

In this paper, we discuss variations of gravity and geoid anomalies,

their causes, and the subsurface density structure that enhances our knowledge of the deeper regions of the Indian continental lithosphere. The 2-D lithospheric density models which cut across two representative regions, one passing the Precambrian Southern Indian shield and the other crossing the Cenozoic Himalayan orogenic belt are presented for comprehending a coherent picture of the Indian lithospheric structure of orogenic belts that governed lithospheric growth and built up the topography by the force of buoyancy.

Tectonic setting of Indian Subcontinent

The Indian subcontinent encompasses the mountainous province of Himalaya girdling the northern border, the almost flat expanse of the Indo-Gangetic Plains in the middle, and the uplands and plateaus of the Peninsular Indian shield (Fig. 1). The Peninsular Indian shield, also known as the southern Indian shield, is a composite of Mid-

Archaean and Neo-Proterozoic ancestral terrestrial domains (Radhakrishna and Naqvi, 1986). Most of these associated tectonic domains are bounded by significant shear zone structures, some of which are sutures (Chetty, 2017). Competitive geodynamic scenarios with components of “complete orogenic cycle” (Nelson, 1992; Leech, 2001) have been suggested to explain the tectonic and magmatic evolution of the Southern Indian shield (Singh et al., 2006; Santosh et al., 2009 and references therein). Most of these scenarios depend on models of the current-day lithospheric structure beneath the region.

The Himalayan orogen is considered to be the result of continued convergence between the Indian and Eurasian continents ensuing their initial collision approximately 50–65 Myr ago (Molnar and Tapponnier, 1975; Yin and Harrison, 2000; Yin, 2006 and references therein). It was later suggested that the dense, adjacent Indian continental lithosphere collided with Asia around 40 Ma, 34 Ma, or even 25–20 Ma (Agius and Lebedev, 2013 and references therein).

The Himalayan mountain range was produced by the convergence of India’s strong cratonic lithosphere and Tibet’s weaker lithosphere (Hatzfeld and Molnar, 2010). The three-primary end-member lithospheric convergence models are proposed viz lithospheric under-thrusting, subduction, and viscous thickening. According to the under-thrusting model, the Indian lithosphere is located right below the Tibetan lithosphere, with no asthenospheric window between them. The lithospheric subduction model describes the lithosphere’s sliding into the asthenosphere (the Tibetan lithosphere above and the Indian lithosphere below are divided by an asthenosphere layer). Under the third end-member Tibet model, lithospheric convergence is viscous lithospheric thickening (Agius and Lebedev, 2013 and references therein). The lithosphere of the mantle thickens and remains intact in one of the proposed scenarios, while the thickened lithosphere becomes destabilized in another situation, with parts of it convectively removed and sinking into the deep mantle (Jiménez-Munt et al., 2008; Tunini et al., 2016).

The sedimentary fills of the Ganga Basin are a continuation of the Siwaliks, while the younger sediments are derived from the Himalaya. The late Cenozoic deposits of the Ganga Basin lap onto the Indian shield to the south, and they increase in thickness, reaching a maximum of 6–8 km against the Himalayan front (e.g., Lyon-Caen and Molnar, 1985; Manglik et al., 2015). The northern border of the depression is clearly defined, while the southern border (often referred to as the hinge zone) is diffuse and highly irregular. It separates the northern Himalayan foreland basin from the southern Indian craton highlands of the peninsula. Burbank et al. (1996) observed that a drainage

Figure 1. Topography map of the Indian shield, Himalaya, and Tibet regions (SRTM 90m elevation data <ftp://edcsgs9.cr.usgs.gov/pub/data/srtm>). The major geological boundaries on the Indian shield and Tibet are shown by solid black lines. Abbreviations used are: ADFB: Aravalli-Delhi Fold Belt, EGMB: Eastern Ghats Mobile Belt, BC: Bastar Craton, BUC: Bundelkhand Craton, EDC: Eastern Dharwar Craton; WDC: Western Dharwar Craton; SC: Singhbhum Craton, SGT: Southern Granulite Terrain; DVP: Deccan volcanic province, GG: Godavari Graben, VNB: Vindhyan Basin, MFT: Main Frontal Thrust, MBT: Main Boundary Thrust, MCT: Main Central Thrust, ITSZ: Indo-Tsangpo Suture Zone, BNSZ: Bangong-Nujiang Suture Zone. The green triangle shows the distribution of broadband seismic stations along the two profiles AB and CD used for the 2-D lithospheric modelling.

system flows directly into the hinge zone in the eastern Ganges, approximately 200 kilometers from the Himalayan front. They attribute this pattern to the erosion-dominated Himalayan formation in Plio-Pleistocene times, which caused isostatic uplift of both the Himalayan range and southward progradation of large alluvial fans and pushed the river far away from the mountain front. Lyon-Caen and Molnar (1985) suggested that a hinge line ~200 km from the Himalayan thrust front has migrated steadily southward at a rate of ~15 mm / yr based on successive onlapping in the depression.

The Indian lithosphere thus offers an outstanding opportunity to study the interplay between surface and subsurface structures and their mutual dependencies during its geodynamic evolution. Despite centuries of extensive research, the most fundamental characteristics of the lithospheric setup under the Indian landmass is debatable. Whether the lithospheric mantle under the Indian shield is thick and intact, or the thickened lithosphere is destabilized, with parts of it convectively removed and sinking into the mantle?

The gravity anomalies

Gravity and geoid anomalies provide insights into buried rocks. Most of the Indian landmass has been covered with high-quality ground gravity data, and a series of gravity maps are published using 3-minute arc interval data extracted from all the gravity data made available by different geoscientific organizations of India (RGMI, 2006). These maps are amalgamated with gravity data from WGM2012 Earth's gravity model, which is a combination of satellite as well as terrestrial measurements (Bonvalot et al., 2012) to prepare a unified gravity anomaly map of India and adjoining regions. Good coherency between the terrestrial gravity (RGMI, 2006) and WGM 2012 Earth's gravity anomalies for wavelengths >50 km (Fig. 2) suggests that the global gravity model filtered to wavelength >50 km can be used for combining the two data sets. Merging the grided terrestrial data with the filtered WGM2012 Earth's gravity data of adjoining regions, the free-air, and complete Bouguer anomaly

Figure 2. Comparison of the power spectrum of terrestrial gravity data of NGRI, and satellite gravity data (Bonvalot et al., 2012).

maps are generated (Fig. 3a, b). The Airy-Heiskanen isostatic anomaly map (Fig. 3c) of the entire region is adopted from the WGM2012 global gravity grids, which have been computed by means of a spherical harmonic approach with crustal thickness as 30 km (Bonvalot et al., 2012). This paper primarily discusses the continental-scale gravity anomalies and the corresponding lithospheric density structures; thus, detailed description of specific local gravity anomalies in these maps have not been elaborated.

Figure 3. (a)

Figure 3. (b)

Figure 3. Gravity anomaly maps (a) free-air gravity anomaly, (b) complete Bouguer anomaly, and (c) The Airy-Heiskanen Isostatic anomaly (Bonvalot et al., 2012) (in mGal) over the Indian shield and adjoining regions.

The gravity anomaly maps exhibit alignments/trends parallel to major structural patterns of the Indian subcontinent such as the NNW-SSE Dharwarian trend of South India, NE-SW Eastern Ghat trend parallel to the east coast of South India, NNE-SSW Aravalli-Delhi trend of North-Western India, ENE-WSW Satpura trend of Central India and the Himalayan trend (Qureshy and Warsi, 1980a; Verma and Subrahmanyam, 1984; Mishra et al., 2008; Rama Rao, 2017). Besides these linear trends, there are several prominent gravity 'highs' and 'lows' associated with local geological features. The largely negative free-air anomaly and complete Bouguer anomaly (-100 to -380 mGal) over the Himalayas are linked to the crustal thickening caused by the collision of Indian and Eurasian plates. Short-wavelength relative free-air anomaly highs of +10 to +50 mGal along the Main Boundary and Main Central Thrust and Indus-Tsangpo Suture Zone might suggest high-density lower crustal rocks, mafic/ultramafic rocks associated along with them. The Western Himalayan front also shows a higher free air anomaly of a limited extent that indicates an under-compensated crust and/or a high-density body in the crust (Mishra et al., 2008). Further south, the Ganga foredeep basin shows negative free-air, Bouguer as well as an isostatic anomaly (<-100 mGal) indicating an over compensated/or a low-density body in the crust in the foreland basins. This is usually the case with the present-day orogenic belts where the frontal part is over compensated while their interior is nearly compensated with free-air anomaly close to zero, as in the case of Tibet (Mishra et al., 2008). The effect of the subducting Indian slab on the negative gravity anomaly, at least partly, can not be ruled out.

In shield regions, the complete Bouguer gravity low (-30 mGal) over the Bundelkhand craton might be related to thick granites of slightly lower density compared to average crustal density

(2.67 g/cm³). The trend of the encompassing gravity high is related to the higher density Bijawar group of rocks (Qureshy and Warsi, 1980b). The large-wave length low Bouguer gravity over the South Indian shield, which contains several short-wave length gravity lows and highs, is a regional feature associated with a source deep in the crust (Tiwari et al., 2001). Some of the short-wave length gravity highs coincide well with the schistose rocks and lows with the granitic intrusions in this region (Krishna Brahman and Kanungo, 1976). As we move westwards, the western Rajasthan is characterized by semi-circular gravity highs (<80 mGal), indicating the high-density intrusive rocks at depths beneath the Malani Igneous Suite and/or extension of Deccan Volcanic Province (Raval and Veeraswamy, 2003; Ravikumar et al., 2013b). Similar pockets of positive gravity anomaly in the Kutchchh and Saurashtra regions and their southward extension all along the west coast, reaching a maximum value of +60 mGal near Bombay, indicate the locus of the Deccan plume path (Negi et al., 1992; Chandrasekhar et al., 2002). Similarly, the gravity highs of Singhbhum craton have been attributed to Dalma-Dhanjori-Simlipal volcanic (Verma, 1985).

Positive gravity anomalies over the fold belts indicate an uplifted block and/or high-density material at the bottom of the crust (Qureshy, 1981; Verma and Subrahmanyam, 1984; Mishra et al., 2000, 2008). The Aravalli-Delhi Fold Belt shows gravity highs both in free-air anomaly (<50 mGal) and complete Bouguer anomaly (<10 mGal) with high elevation. A positive correlation between free-air anomaly, Bouguer anomaly and elevation suggests a lack of isostatic compensation. Gravity highs are possibly caused by a shallow crust with high-density underplated materials (Mishra et al., 2000; Mishra, 2006). The Satpura Fold Belt, characterized by gravity high both in free-air (<70 mGal) and complete Bouguer (<10 mGal) anomaly with rising elevation also indicates lack of isostatic compensation and gravity anomalies are caused by block upliftment and/or high-density body in the crust (Qureshi, 1971; Mishra, 1992; Verma and Banerjee, 1992; Singh and Meissner, 1995; Singh, 1998, 2002). Rajsekhar and Mishra (2008) have suggested its extension further eastward up to Shillong Plateau, which has under gone a history of vertical uplift since Cretaceous with an average plateau elevation of about 1 km (Verma and Mukhopadhyay, 1977). The Moho beneath the Shillong Plateau is at a shallower depth of about 35 km and attributed to reverse faulting due to 'pop-up' tectonics (Nayak et al., 2008) and also to magmatic underplating fed by mantle-plume related tectonics (Kent et al., 1992; Bakshi, 1995). The north-south elongated Bouguer anomaly of 0 to 25 mGal over the Rajmahal Traps attains its peak amplitude over the shield edge with an average wavelength of 100 km in an east-west direction. This long-wavelength nature of the anomaly argues for a source at some deeper level, possibly at the base of the crust (Singh et al., 2004a).

The parallelism between the gravity contours and the structural grains of the Eastern Ghats Mobile Belt with a strong gravity gradient along its western boundary suggests a faulted contact resulting in a close juxtaposition of contrasting crustal blocks (Subrahmanyam 1978, 1983; Singh and Mishra, 2002; Niraj Kumar et al., 2004; Singh et al., 2004b; Mishra, 2006). The intervening gravity lows are attributed to low-density Gondwana and Vindhyan sediments. The gravity highs adjacent to Godavari Gondwana Basin suggest high-density rocks along shoulders, typical of extensional tectonics associated with continental rift basins (Mishra et al., 1999). The linear belt of negative isostatic anomaly ranging from -40 to -70 mGal along the west coast of India seems to be governed by the topography as it

extends even outside the Western Ghats (Subrahmanyam and Verma, 1980) and possibly indicates a thicker than normal crust beneath the region (Dubey and Tiwari, 2018). This feature may be due to an undissipated crustal root in the mantle and/or the local decrease in upper mantle density, probably caused by the large mass transfer during the Deccan magmatism (Qureshy, 1971; Singh and Mall, 1998; Tiwari et al., 2001).

Geoid Anomaly

The hybrid geopotential model EGM2008 uses long-wavelength data from satellite and short-wavelength data from available terrestrial gravity and provides reasonably good information over the region of small geoidal anomalies (Pavlis et al., 2012). The most significant Geoid anomaly on the Earth is “the Indian Ocean Geoid Low (IOGL)” located just south of the Indian peninsula (Fig. 4a). The value of this geoid anomaly varies from -82 m over Dharwar Craton to -98 m over the Southern Granulite Terrain with a steep gradient between the two domains (Carrion et al., 2009). It spans over the Indian Ocean and is dominated by a significant low of -106 meters to the south of Sri Lanka. The mass anomalies contributing to the degree 2-3 field are inferred to be the result of topography at the core-mantle boundary, whereas the degree 4-10 geoid anomaly appears to lie mostly in the lower mantle depths (Bowin, 1983, 1991). The source of IOGL may have the shape of a vertical column at one or more locations. Probably, one of the locations of mass deficiency is at the core-mantle boundary (Hide and Malin, 1970; Negi et al., 1987; Padma Rao et al., 2017), and the other within the mantle (Kaula, 1972). According to Ihnen and Whitcomb (1983), the IOGL is caused by a sag in the crustal boundary whereas Marsh (1979), and Mishra et al. (2004) opined as

due to low-density rocks in the upper mantle (Mishra, 2014; Mishra and Ravikumar, 2012). The IOGL is caused by a low-density anomaly stretching between the depth of 300 km and ~900 km in the northern Indian Ocean (Ghosh et al., 2017; Reiss et al., 2017) while Chase (1979) suggested approximately 1200 km. Moberly and Khan (1969) and Kaula (1970) have linked the IOGL to the absence of asthenospheric return flows or less viscous asthenospheric flows as a result of lithospheric plate movements at ridges or deep trenches.

The topography corrected geoid undulation in the Himalaya is small, but still significant, compared to other mountainous regions (Banerjee et al., 1999). The geoid anomaly over the Himalaya-Tibet region suggests the existence of a deeper density distribution plausibly caused by the subducted high-density Tethys oceanic slab beneath the region (Jiménez-Munt et al., 2008). The steep geoid gradients on both the Himalayan front and the northern margin, reaching between 20 and 30 m under the plateau, show a marked thinning of the lithospheric mantle (Robert et al., 2015). Late uplift in the tectonic evolution of the Tibetan plateau, the widespread extension, and the associated magmatism was attributed to convective removal and replacement of the lower part of the lithospheric mantle by hotter and lighter asthenosphere (Jiménez-Munt et al., 2008).

Geoid above degree and order 20 is normally attributed to near-surface mass density anomalies and between degree 10 and 20 to mass density anomalies at intermediate depths (Christou et al., 1989). The long-wavelength geoidal undulations are due to deeper sources, and filtering the degree 8-10 of spherical harmonics probably retains the effects of density anomalies shallower than ~400 km depth. This filtered/residual geoid anomaly map (Fig. 4b) shows a significant high in the Central Indian plateau region, and its causative source is a matter of debate. It is quite likely that the local low over the Ganges

Figure 4(a). The geoid height (in meters) of the Indian Shield, Himalaya, and Tibet regions, taken from the EGM2008 global model (<http://www.agu.org/pubs/crossref/2012/2011JB008916.shtml>), and (b) Residual Geoid undulations with spherical harmonics up to degree and order eight removed.

and high over the Central Indian plateau region indicate flexure in the Indian plate (Bilham et al., 2003). The filtered Geoid anomaly map is used by Niraj Kumar et al. (2013, 2014) and Singh et al. (2015b) to model the lithospheric density structure of the southern and eastern Indian shield, respectively. Following a similar approach, the lithospheric density structure of the Himalayan orogenic region and adjoining Tibetan Plateau is modelled by Jiménez-Munt et al. (2008), Robert et al. (2015) and Tunini et al. (2016).

Isostasy

The surface elevation, the density difference between the lower crust and the upper mantle, the configuration of the crustal root, and rigidity of the lithospheric plate are closely linked through the phenomenon of isostasy and best reflected in the gravity data. Uplift of the earth's surface is often accompanied by thickening of continental crust, and buoyancy of these deep crustal roots supports the topography (Fischer, 2002). Even if the lithosphere is mechanically very stiff, a buoyant crustal root and/or mantle inflow would complement post-tectonic erosion of the topographic mass. However, if the lithosphere is quite weak, the net mass change would be zero over time, maintaining local isostatic balance (Fischer, 2002).

Local Isostasy

In general, substantial non-zero Airy-Heiskanen isostatic anomalies persist in the active continental collision regions, whereas shield regions are normally characterized by compensated topography. Positive bias in isostatic anomalies linked with the active orogeny is dynamically maintained by the stresses involved with the plate strength and movement. When subduction stops, the under-compensated topography will subside as isostatic equilibrium is achieved in a brief characteristic time compared to the orogenic time scale (Gratton, 1989). The Western Himalayan front shows a positive isostatic anomaly of a limited extent that indicates an under-compensated crust implying a smaller crustal thickness compared to isostatically balanced crust or high-density body in the crust or lithosphere (Mishra et al., 2008). In shield areas, the topography of short-wavelength is retained due to the resistance of the lithosphere; hence, there would be hardly any undulation of the Moho interface. However, if the lithosphere's mechanical strength is weak, the root response of the regional topography is the manifestation of (1) the Moho configuration (Woollard, 1959; Subba Rao, 2002), and/or (2) the deep crustal and upper mantle density distributions supporting the topography in a way that is consistent with the Archimedes principles (Martinez, 1994a; Menke, 1999).

Flexural Isostasy

Part of the lithosphere that behaves elastically over geological time is termed as Effective Elastic Thickness (EET) and is often referred to as a measure of the strength of the lithosphere. The larger the EET, the stronger is the lithosphere, causing regional isostasy, whereas a weak lithosphere with zero EET leads to local isostasy (Watts, 2001). Flexural isostasy is essentially a regional effect due to the rigidity of the lithospheric plate, and generally shows long-wavelength gravity anomalies (> 200 km).

Isostatic studies in India indicate that the variation in crustal thickness due to regional isostatic compensation results in the medium

to long-wavelength anomalies of gravity of > 250 km (Tiwari and Mishra, 1999, 2008; Jordan and Watts, 2005). The EET calculated for distinct regions differs extensively between 10 and 50 km using spectral methods, reflecting a reworked crust in the Indian continent (Tiwari and Mishra, 1999, 2008). Stephen et al. (2003) used the multi-taper technique and obtained low EET values of 11–16 km through separate prolate spheroidal Slepian sequences in the South Indian shield. EET varies from 18 to 26 km, with an average of 23 km over the North Indian shield and 12 to 16 km with an average of 14 km over the South Indian shield (Rajesh and Mishra, 2004). The average admittance function computed across the Himalayas and Tibet shows an EET of 50 km (Rajesh and Mishra, 2003). The EET structure of the Indian shield derived from isotropic fan wavelet methodology documents spatial variations of lithospheric deformation in different tectonic provinces. The thinned, attenuated lithosphere beneath Peninsular India has a mechanically weak strength (< 30 km). Average EET values (40–50 km) for the Central Indian Tectonic Zone, the Bastar Craton, and the northern part of the Eastern Ghats Mobile Belt are suggestive of stable Indian lithosphere (Ratheesh Kumar et al., 2014). McKenzie & Fairhead (1997) suggested an EET of ~ 40 km south of the Main Frontal Thrust while Ratheesh-Kumar et al. (2014) reported an anomalously high EET (60–85 km) to the north of the Narmada–Son Lineament, primarily in NW Himalaya, and the northern Aravalli and Bundelkhand Cratons. The EET of Central Himalaya region to Sikkim is estimated as $\sim 50 \pm 10$ km (Tiwari et al., 2006; Berthet et al., 2013) while the extreme Western and Eastern Himalaya including Bhutan (25 km) seem to be different from the central Himalaya (Hammer et al., 2013). The northwards decrease of EET from 60–80 to 20–30 km as Indian plate flexes down beneath Himalaya and Tibet is due to the thermal and flexural weakening (Jin et al., 1996; Cattin et al., 2001; Jordan and Watts, 2005; Hetényi et al., 2006; Tiwari et al., 2006, 2010). It may be mentioned here that the EET has been studied using different techniques that produce different values sometime based on the methodologies used. There are uncertainty studies that argue reasonably the difference in absolute values of the EET between different methods. The obtained values of EET in the Indian shield are indeed low when compared to the values obtained from other shield regions world over (Rajesh et al., 2012).

The isostatic state of the South Indian shield of comparatively small EET is believed to be over compensated essentially because the mean elevation of the region is lower than that anticipated by Airy-Heiskanen model (Mishra et al., 2004). The solution to this mystery was put forward by postulating the decrease in the contrast of crust-mantle density, trying to reduce the buoyancy that counter acting the isostatic response of the shield to denudation (Niraj Kumar et al., 2011). However, a single solution such as lowered buoyancy may not probably be sufficient for the entire shield of South India. Instead, the isostatic adjustment process must incorporate lateral density variations both within the crust and the upper mantle (Niraj Kumar et al., 2011; Paul et al., 2018).

Lithospheric structure of India using Potential Fields

The Airy-Heiskanen model of isostatic compensation only partly compensates the surface topographic masses. To fit the external gravitational potential induced by the surface topography, the Pratt-Hayford concept of compensation has also to be considered. This

means that the topographic masses are compensated throughout the Earth's lithosphere (Martinec, 1994b). The complete Bouguer anomaly over India consists of three major components: (i) The contribution of the Ganga Basin foreland sediments that are reflected in the long-wavelength component related to their broad spatial distribution, (ii) effect associated with the variability of Moho, and (iii) the crustal sources including shallow/exposed sources represented by the short-wavelength (<250 km) component of the gravity field (Mishra et al., 2012; Tiwari et al., 2014). The removal of these three components from the observed field provides the gravity effect of the lithospheric mantle sources. Modeling of this long-wavelength gravity anomaly provides minimum lithospheric thickness of 120 km beneath Southern Granulite Terrain and reaches 140–155 km under the EDC and 155–175 km underneath the Satpura Mobile Belt, rising to 155 km beneath Vindhyan and Ganga basins due to the lithospheric bending of the Indian plate underneath the Himalayas. It increases finally to 190 km along the Himalaya front (Tiwari et al., 2006, 2013). In the Hindu Kush–Pamir section of the Himalayan collision zone, the lithosphere lies between 140 and 160 km, where the Indian plate is under-thrusting the Asian one (Tiwari et al., 2009) while the Asian lithosphere lies between 140 and 250 km (Tiwari et al., 2014). Intense seismicity to depths of 100–300 km beneath the Pamir is attributed to the fast subduction of the Indian lithosphere in this region (Tiwari et al., 2014). Spectral analysis of both the Bouguer as well as the Geoid anomalies resulted in a similar lithospheric thickness in the region (Ravikumar et al., 2013a).

Incorporating gravity and geoid anomalies and topography data with the seismic and geothermal record, lithospheric modeling along a couple of geo-transects passing through the Himalayan orogenic belt (Jiménez-Munt et al., 2008; Tunini et al., 2016) and the southern Indian shield (Niraj Kumar et al., 2013) has recently been accomplished. In the Western Himalaya, the depth of the lithosphere–asthenosphere boundary increases from 220 km beneath the Himalayan foreland basin to 295 km underneath the Kunlun Shan and then shallows to approximately 230 km under the Tarim basin (Tunini et al., 2016). In the Central Himalaya, the lithospheric depth increases from 150 km beneath the Indo-Gangetic belt to ~160 km beneath the Main Central Thrust. It rises suddenly to 240 km below the Tethyan Himalayas and reaches a maximum depth of approximately 260 km below the southern Tibetan plateau. The lithosphere thins sharply towards north to about 100 km below the central and northern Tibetan Plateau (Jiménez-Munt et al., 2008). A 3-D lithospheric thickness model in the southern and western parts of the Tibetan plateau shows approximately 200–240 km (Robert et al., 2015). This thick Tibetan lithosphere corresponds to the northern boundary of the under-thrusted Indian lithosphere below the Tibetan Plateau.

The lithospheric thickness of the southern Indian shield differs considerably across the three profiles from ~70 to 100 km below the adjacent oceans to ~130 to 135 km under the southern block of Southern Granulite Terrain, including Sri Lanka, and gradually rises to ~165 to 180 km underneath the northern block of Southern Granulite Terrain and Dharwar Craton (Niraj Kumar et al., 2013). The 3-D lithospheric structure of the southern Indian shield and Singhbhum Craton is derived from joint inversion of free-air gravity and geoid anomalies and topography data (Niraj Kumar et al., 2014; Singh et al., 2015b). The LAB is located at a depth of 70–120 km under oceanic regions and 150–180 km below the Dharwar Craton and the Northern block of Southern Granulite Terrain. A remarkably

thinned lithosphere of 130 km close to Bangalore in the Eastern Dharwar Craton, 140 km beneath the Southern block of Southern Granulite Terrain, and 130 km in Sri Lanka are reported (Niraj Kumar et al., 2014). The LAB beneath the Singhbhum Craton is at a depth of about 130–140 km. In the regions of the Bastar Craton and Bengal Basin, the LAB dips to approximately 155 km depth. The confluence of Mahanadi and Damodar Gondwana basins toward the north-west and the foreland Ganga Basin toward the north is characterized by a deeper LAB lying at a depth of over 170 and 200 km, respectively (Singh et al., 2015b).

In view of the foregoing, a 2-D lithospheric density cross-section is generated along two profiles in two representative lithological units (Profile A-B in the southern Indian shield, and C-D across the Himalayan orogenic belt; Fig. 1) applying a finite-element method that uses the lithospheric mantle density as a function of temperature and a constant density in the asthenosphere (Zeyen and Fernández, 1994; Zeyen et al., 2005). Relatively sufficient available information enabled us to reasonably fix the initial crustal structure and properties along the profile (Reddy and Rao, 2013; Singh et al., 2015a, 2017). The modelling approach assumes local isostasy, however, contraining seismic results reduce the influences in density model of such assumption but increases mismatch in the short wavelength topography model. Further changes in crustal thickness and densities were minor and restricted to the uncertainty range (5–10% of the value). Consequently, we varied the depth of the lithosphere to fit the elevation data and gravity and geoid anomalies.

2-D Modelling across the Peninsular India

The LAB is situated at depths of 80–120 km beneath the oceanic areas, 140 km under the southern block of Southern Granulite Terrain, and 150–180 km below the northern block of Southern Granulite Terrain and reduces to about 165 km beneath the Dharwar Craton. In the southern segment of the transect, we used a very similar crustal model to that of Niraj Kumar et al. (2013), who considered a 3-layer crust with a gentle up warp beneath the region whereas we used a better controlled thicker lower crust (Singh et al., 2017). Subsequently, the LAB depths produced in our model are 10–15 km deeper underneath the South Indian Margin and Southern Granulite Terrain; however, it is about 10–20 km shallower under the northern Karnataka Plateau region (Fig. 5). Anyhow, this might be considered as minimum uncertainty for LAB depth modeling. The 130 km thin lithosphere in the eastern Dharwar Craton close to Bangalore delineated by Niraj Kumar et al. (2013, 2014) is conspicuously missing in the present model. The modeled topography in this region is also less than that of the actual topography. It is quite likely that the actual topography may have been supported by an elastic plate or formed by a deep process (hot mantle). The former possibility is, however, ruled out by the low EET existing in the southern Indian shield. These disparities in the LAB-depth and topography may be attributed to a small difference in the calculated mantle thermal conductivity and the lateral density inhomogeneity of the crust. Signature of density inhomogeneity is seen in the average crustal shear wave velocity distribution, which is lower beneath the Bangalore (Bora et al., 2014).

2-D Modelling across the Himalayan Orogeny

The deep structure of the Indian-Eurasian collision zone and the

Figure 5. Final modelling results along with the observed and calculated response across profile AB. In 5(a–d), continuous lines correspond to the calculated response of the initial model, and dots with uncertainty bars correspond to measured data. In 5(b), blue line, and symbols correspond to free-air gravity anomalies, green ones to complete Bouguer anomalies. The lithospheric model is presented in 5(e), and 5(f) shows a blow-up of the crustal structure of our model with densities and related dashed lines showing the available Moho depths from different seismological models.

mechanism supporting the Himalayan Orogenic belt are under debate. Applying the same finite-element method and using crustal constraints from Banerjee and Prakash (2003), Tiwari et al. (2006), Jiménez-Munt et al. (2008), Chamoli et al. (2011), Ansari et al. (2014), Tunini et al. (2016), and Singh et al. (2017), a 2-D density cross-section of the crust and upper mantle is produced along this profile (Fig. 6). Our findings suggest that the height of the Tibetan Plateau as a whole cannot be supported isostatically only by ~65 km dense crust. The elevated topography, gravity, geoid, and crustal temperatures need a thin and warm lithosphere under the plateau region. Interestingly, the lithosphere thins up to a depth of ~150 km beneath the southern Tibetan Plateau. The depth of LAB increases from 160 km beneath the Himalayan foreland basin to >250 km under the Main Central Thrust. Further north, the LAB reduces to ~150 km beneath the Tethyan Himalaya. The existing density models indicate a noticeable change in lithospheric thickness below the Himalayan Orogenic belt, with the lithosphere–asthenosphere boundary to be found at ~260 km depth in the Western Himalaya (Tunini et al., 2016) that changes to ~250 km in the Central Himalaya and reaches to 200 km in the Eastern Nepal Himalaya (Jiménez-Munt et al., 2008).

Summary

The 2-D lithospheric density model across the two distinct geological domains helps to understand the deep lithospheric density structures beneath the South Indian Shield and the Himalayan orogenic belt. Integrating the seismic tomographic images, the modelled profile across the Himalayan orogenic belt delineates the existing location of the Indian lithospheric mantle under the thrust regime of the Himalaya and Tibetan Plateau. The virtually steady lithospheric thickness of the Indian plate (~150 km) increases sharply to about 180 km north of the Main Boundary Thrust, reaching a peak of 260 km below the southern block of Lhasa. The abrupt change in the thickness of the lithosphere further north indicates hot asthenospheric mantle under the Tibetan Plateau (where the lithospheric mantle is only 100 km thick) as compared to the Indian region i.e., colder asthenosphere beneath the Tethys Himalaya (Jiménez-Munt et al., 2008). A low lower crustal S-wave velocity and the absence of an S-wave velocity gradient in the mantle below northern Tibet indicate higher temperatures below north-eastern Tibet than in southern Tibet (Artemieva, 2006). The presence of basaltic volcanism of different

Figure 6. Final modelling results along with observed and calculated response along the profile CD. For further explanations, see Fig. 5.

ages, and the exceptionally small P-velocities in the lower crust and upper mantle match nicely with the notion of a hot, attenuated and soft lithosphere (Singh et al., 2017 and references therein). The decline in flexural wavelength from west to east is correlated with weakening rheology of the mantle (Hammer et al., 2013). Four Himalayan sub-segments, namely NE India, Bhutan, Nepal & India until Dehradun, and NW India have different flexural geometry (Hetényi et al., 2016).

The density structure of the Indian lithosphere is mapped using continental-scale topography, gravity and geoid data sets following the principle of local isostasy. As already mentioned, the southern Indian shield's crustal isostatic compensation is inadequate to justify findings and requires differences in density to be present in either the crust or in the deeper lithosphere. More than two-thirds of the Indian continent is underlain by lithosphere of over 150 km thick, signifying that the Proterozoic lithospheric root does exist today. The thinned lithosphere underlying the Saurashtra Region, Southern Granulite Terrain, together with Sri Lanka in the south and Eastern Indian Shield of Peninsular India, is characterized by a typical body wave velocity distribution indicating unusual mantle conditions (Singh et al., 2014; Maurya et al., 2016). The heterogeneous and uneven lithospheric thickness has controlled many features of Indian geology with a major influence on how the continent has responded to tectonic forces related to Gondwana breakup and Pan-African to K-T boundary thermal perturbations. In Himalaya, it provides control on topography and landscape, seismicity, and the evolution of the frontal basin. After the

Indo-Tsangpo Suture zone, a sudden decrease of lithosphere may be the imprint of predominant mantle isostasy beneath the Tibetan plateau which is different from the negative isostasy due to thickening of the lithosphere beneath the Himalayan mountains and the consequence of this will be thick crust beneath Himalaya and thin-crust beneath Tibet for the similar topography (Molnar et al., 1993).

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