Lithosphere Structure and upper mantle characteristics below the Bay of Bengal

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SUMMARY:

The oceanic lithosphere in the Bay of Bengal (BOB) formed 80 - 120 Ma following the breakup of eastern Gondwanaland. Since its formation, it has been affected by the emplacement of two long N-S trending linear aseismic ridges (85°E and Ninetyeast) and by the loading of ca. 20-km of sediments of the Bengal Fan. Here, we present the results of a combined spatial and spectral domain analysis of residual geoid, bathymetry and gravity data constrained by seismic reflection and refraction data. Self-consistent geoid and gravity modeling defined by temperature-dependent mantle densities along a N-S transect in the BOB region revealed that the depth to the Lithosphere-Asthenosphere boundary (LAB) deepens steeply from 77 km in the south to 127 km in north, with the greater thickness being anomalously thick compared to the lithosphere of similar-age beneath the Pacific Ocean. The Geoid-Topography Ratio (GTR) analysis of the 85°E and Ninetyeast ridges indicate that they are compensated at shallow depths. Effective elastic thickness (Te) estimates obtained through admittance/ coherence analysis as well as the flexural modeling along these ridges led to the conclusions: i) 85°E Ridge was emplaced in off-ridge environment (Te=10-15 km); ii) the higher Te values of ~ 25 km over the Afanasy Nikitin Seamount (ANS) reflect the secondary emplacement of the seamount peaks in off-ridge environment, iii) that the emplacement of the Ninetyeast Ridge north of 2°N occurred in an off-ridge environment as indicated by higher Te values (25-30 km). Furthermore, the admittance analysis of geoid and bathymetry revealed that the admittance signatures at wavelengths > 800 km are compensated by processes related to upper mantle convection.

Key Words: Lithospheric structure; Admittance; Geoid-Topography Ratio; Flexural modeling; Aseismic ridges; Upper mantle; Bay of Bengal

1. INTRODUCTION

Combined analyses and modeling of gravity/ geoid anomalies together with bathymetry data provide key information on thermal structure, mass distributions and evolution of the oceanic lithosphere and upper mantle (Parsons and Daly, 1983; Sandwell and McKenzie, 1989; Black and McAdoo, 1988). While gravity anomalies are more sensitive to crustal variations, the geoid can better detect deeper mantle density variations (Zeyen and Fernandez, 1994; Zeyen et al., 2005; Afonso et al., 2008) and our combined approach permits simultaneous modelling of crustal structure and the Lithosphere-Asthenosphere Boundary (LAB). The LAB is of fundamental importance in understanding the mantle dynamic processes as it separates the cold, rigid, conducting lithosphere from the underlying weaker, hot, convective asthenosphere. Seismological studies have revealed that the LAB is sharp and identifiable in the case of younger continental lithosphere, whereas, below stable continents it is too diffused and/or heterogeneous to identify as a single coherent boundary (James, 2011). However, below the ocean basins, the LAB is generally related to an isotherm as its thickness is mainly controlled by thermal cooling with respect to oceanic lithosphere age (Parsons and Sclater, 1977), except in the regions affected by plumes or major tectonism. The spectral domain analysis of topography with gravity / geoid data provides valuable information for comparing thermo-mechanical models of the lithosphere and sub-lithosphere in the oceanic areas (McKenzie and Bowin, 1976; Black and McAdoo, 1988; Watts, 2001). Further, the spatial domain analysis of geoid and

topography data is also found to be useful for determining the deep structure beneath oceanic aseismic ridges and their mode of compensation (Sandwell and McKenzie, 1989).

Our focus of investigation is the Bay of Bengal (BOB), where the oceanic lithosphere mostly formed during 120 - 80 Ma (Krishna et al., 2009), and has been affected by the emplacement of two aseismic ridges (Fig. 1) as well as substantial sediment cover associated with the pre- and post- collisional sediments which can be up to 22 km in thickness (Curray, 1991). The BOB lithosphere evolved in two stages, the initial rifting and drifting of Indian subcontinent from East Antarctica (~130 Ma) followed by seafloor spreading, and subsequent breakup of the Elan Bank from the Indian continental margin and seafloor spreading (Gaina et al., 2007). This complex breakup history together with the northward movement of Indian subcontinent and its passage over major hotspots (Duncan et al., 1992), large-scale sedimentation processes, collision of India and Eurasia and the eastward subduction of the Indian plate may all have significantly impacted the lithosphere structure below the BOB. Previous investigations have been limited to deriving crustal structure based on seismic reflection and refraction data or through the integrated interpretation of gravity and magnetic data (Curray et al., 1982; Mukhopadhyay and Krishna, 1991; Gopala Rao et al., 1997; Ramana et al., 1997; Subrahmanyam et al., 1999; Krishna, 2003; Krishna et al., 2009; Radhakrishna et al., 2010, 2012); and have not provided constraints on structure, mechanical properties of lithosphere and sub-lithospheric processes, hence making our understanding of the evolution of the BOB lithosphere incomplete. Here, we focus on determining variability in LAB depth along and across the BOB, sub-lithospheric upper mantle characteristics, the compensation mechanism and mode of emplacement of the aseismic ridges on the oceanic lithosphere, through spatial and spectral domain analysis of the correlation between geoid / gravity and bathymetry, and by combined gravity-geoid modeling.

2. REGIONAL GEOLOGIC SETTING AND EVOLUTION

The lithosphere in the northeast Indian Ocean resulted from the breakup of eastern Gondwanaland in the early Cretaceous and evolved through three major phases of seafloor spreading; i) in a NW-SE direction until mid-Cretaceous, ii) N-S until the early Tertiary, and iii) subsequently in a NE-SW direction continuing until the present (Royer et al., 1989; Muller et al., 2000). While the N-S spreading regime can be easily recognized in the distal Bengal Fan (south of 6° N) as evidenced by well-developed E-W trending magnetic anomaly identifications offset by N-S oriented fracture zones (Fig. 1); the presence of thick sediments and lack of identifiable magnetic anomalies in the northern part of BOB prevents the recognition of NW-SE spreading regime identified elsewhere (e.g. Krishna et al., 2009). Recently, based on basement trends observed from large number of multi-channel reflection profiles, Radhakrishna et al. (2012) have mapped several NW-SE trending fracture zones in the northern part of the BOB. The BOB is also characterized by the presence of one of the largest submarine fans of the world which extends from 20° N to 7° 40' S over a length of 3000 km (Krishna et al., 2001a; Curray et al., 2003). The BOB has received sediments from different sources since the breakup of eastern Gondwanaland (~130 Ma): initially from the Indian peninsular rivers including the Mahanadi, Krishna-Godavari and Cauvery rivers until Oligocene time (Bastia et al., 2010). After the continental collision between India and Eurasia, the Ganges and Brahmaputra rivers have discharged a huge volume of sediments from the Himalayan mountain range. Seismic stratigraphic studies of the BOB sediments (Gopala Rao et al. 1997; Michael and Krishna, 2011) have recognized the unconformity separating the pre-collision sediments from the postcollision sediments and it has been ascribed an Eocene age, but the age of sediments at the base of the Bengal Fan is still uncertain. The sediment thickness estimated as over 22 km in offshore Bangladesh (Curray, 1991), decreases towards the south with a significant decrease in sediments thickness between 8° and 10°N, with also relative thinning over the 85°E and Ninetyeast ridges. These two linear longitudinal features divide the BOB into three major subbasins: Western, Central and Nicobar basins (Fig.1). While, emplacement of the Ninetyeast Ridge is known to be the result of the Kerguelen hotspot related volcanism (Frey and Weis, 1995); in the case of 85°E Ridge, the nature and source of plume as well as mode of its emplacement is not yet clear.

3. DATASETS USED

In this study, we carried out an integrated analysis of bathymetry, gravity, geoid, multichannel seismic reflection and refraction data to understand the lithospheric structure beneath the BOB. The bathymetry data from the ETOPO1 (Amante and Eakins, 2009) and free-air gravity data from the 1 minute grid DNSC08 global gravity database (Anderson and Knudsen, 2008) were used to prepare the bathymetry map (Fig. 2A) and free-air gravity anomaly map (Fig. 2B) of the BOB region. Sediment thickness data in the region (Curray, 1991; Divins, 2003; Radhakrishna et al., 2010) were integrated with the bathymetry to generate the basement depth map (Fig. 2C). The geoid grid for the BOB region was extracted from geoid data generated for the northern Indian Ocean by Sreejith et al. (2013). The regional multi-channel seismic reflection profiles (Curray et al., 1982; Curray, 1991; Gopala Rao et al., 1997; Krishna et al., 2001a; Michael and Krishna, 2011) and sonobuoy refraction data (Naini and Layden, 1977; Curray et al., 1982) within the study region are utilized for gravity-geoid modeling. Additionally, for the purpose of 2-D flexural analysis over the aseismic ridges, we used several short E-W seismic reflection profiles across the 85°E Ridge (Bastia et al., 2010; Radhakrishna et al., 2012; Krishna et al., 2014; Rao and Radhakrishna, 2014) and across the Ninetyeast Ridge (Curray et al., 1982; Gopala Rao et al., 1997; Subrahmanyam et al., 2008; Maurin and Rangin, 2009).

4. METHODOLOGY AND DATA ANALYSIS

4.1 Spectral Analysis of the Geoid

As the geoid data include the effects of all distributed masses within the earth, the contribution of deeper mantle density sources must be removed from the observed geoid data using a reference geoid model in order to separate the anomalies contributed by lithospheric and sub-lithospheric mass variations (Bowin, 1983; Sandwell and Renkin, 1988). We, therefore, generated three residual geoid anomaly maps from the full spectrum geoid (Fig. 3A) by subtracting different geoid contributions corresponding to the spherical harmonic coefficients of degree and order less than or equal to 10, 30 and 50 using the EGM2008 reference geoid. This procedure removes the contribution of the geoid of wavelengths > 4000 km, > 1334 km, and > 800 km respectively and give rise to different residual geoid anomalies. As suggested by Sandwell and Renkin (1988), while subtracting these different geoid contributions, the model coefficients are smoothly rolled over by applying a cosine taper at that particular degree and order to avoid sharp truncation effects. However, in order to check the reliability of the EGM2008 data, we made a spectral comparison with the latest DTU10 model (Anderson, 2010) and observed that the difference between the EGM2008 and DTU10 models is found mainly at shorter wavelengths (< 30 km) (Fig. 3B).

In general, long wavelength components of geoid data (> 4000 km) are sensitive to the density variations in the lower mantle (Bowin, 1983), whereas, shorter wavelength components (< 330 km) reflect the crustal thickness variations and lithosphere flexural effects (Marks and Sandwell, 1991). Three residual geoid anomalies were generated (Fig. 3C-E) for the BOB region in order to approximately understand geoid contributions from different depth regions. While, n=10 (λ < 4000 km) and n=30 (λ < 1333 km) degree residual geoid anomaly maps mostly reflect the sub-lithospheric mass variations, the n=50 (λ < 800 km) corresponds to sources within the lithosphere (Featherstone, 1997). Further, we prepared the band-pass filtered

residual geoid anomaly map (Fig. 3F) for degree (10 < n < 50) for the eastern Indian Ocean region as it reflects the mass variations within the upper mantle.

4.2 Geoid-Topography Ratio (GTR)

The geoid data are known to relate linearly to the topography between wavelengths 400 km and 4000 km (Sandwell and Renkin, 1988), therefore, the GTR analysis for these wavelengths can be used to understand the mode of compensation of topographic features such as swells, plateaus and aseismic ridges (Haxby and Turcotte, 1978; Sandwell and McKenzie, 1989; Coblentz et al., 2011). In general, low GTR values (0-2 m/km) indicate shallow level of compensation, whereas, the intermediate GTR values (2-6 m/km) suggest that the compensation takes place at deeper level by lithosphere thinning (Crough, 1978). Conversely, higher GTR values (> 6 m/km) indicate compensation below the lithosphere supported by convective stresses (McKenzie et al., 1980).

In the present study, we carried out GTR analysis along the 85°E and the Ninetyeast ridges at two wavelength scales. First, the band pass filtered (330 – 4000 km) maps of the bathymetry and geoid data were prepared (Figs. 4A and 4B) by considering the spherical harmonic coefficients for degrees between n=10 (< 4000 km) and n=120 (> 330 km). The scatter plots of these band-pass filtered geoid and the bathymetry (corrected for sediment load using the method of Sykes, 1996) have been prepared for twenty-two 4°×2° overlapping (1° overlap) blocks along these two ridges (see Fig. 4B for location of blocks). For each block, least-square regression lines are fitted and the correlation coefficient is estimated. While, the correlation coefficient for each block reveals goodness of fit, the slope of the regression line gives the GTR value for each block. Similarly, we analyzed the band pass filtered geoid and bathymetry data between 330 and 800 km wavelengths corresponding to degrees n = 50 (< 800 km) and n = 120 (>330 km). The computed GTR values for each block are plotted

longitudinally (centered at particular latitude) along both the 85°E and Ninetyeast ridges (Figs. 5A and 5B) are presented.

4.3 Admittance / Coherence analysis

Though the aforementioned GTR analysis provides valuable information on the nature of isostatic compensation, it would not be able to discriminate between different isostatic compensation models along the aseismic ridges. For this purpose, the relationship between geoid/ gravity and the topography in the spectral domain (admittance / coherence) is utilized for better understanding the isostatic response of the topographic loads, and for studies related to thermo-mechanical characterization of the oceanic lithosphere and/or sub-lithospheric upper mantle. Several researchers have effectively used this approach i) for evaluating isostatic response of the oceanic lithospheric using free-air gravity/ geoid and bathymetry admittance function (e.g., McKenzie and Bowin, 1976; Watts, 1978; Black and McAdoo, 1988); ii) for estimation of flexural rigidity of continental lithosphere through Bouguer gravity and topography coherence function (Forsyth, 1985; Lowry and Smith, 1994; McKenzie and Fairhead, 1997; Simons et al., 2000; Swain and Kirby, 2003). In the present study, we analyzed the admittance between bathymetry and residual geoid on two wavelength scales, one using degree-10 residual geoid for modeling upper mantle convection processes, and other using degree-50 residual geoid for understanding mode of isostatic compensation along the 85°E and Ninetyeast ridges. Further, as these two linear ridges in BOB are buried below thick sediments, we computed the Mantle Bouguer Anomalies (MBA) to further analyse the mode of isostatic compensation of these ridges. The MBA were computed by removing the gravity contributions of water, sediment and a 6 km uniformly thick oceanic crust from the free-air gravity anomalies using a normal oceanic crustal density of 2900 kg/m³. We then carried out MBA-bathymetry coherence analysis along these two ridges. Further details on the methodology are described below

4.3.1 Long wavelength (> 800 km) geoid-bathymetry Admittance

We computed the admittance of bathymetry and residual geoid for three overlapping blocks of size 13°×13° with a grid sampling of around 3'×3' covering the BOB region (Fig. 1). The large block size and the degree-10 residual geoid is purposely chosen in order to incorporate the wavelengths greater than 800 km in the spectral estimates that are attributable to upper mantle convection processes. The blocks were limited to 17°N latitude in the BOB region to avoid the effect of continental mantle signatures on the admittance computation. The method used for the calculation was adopted from Black and McAdoo (1988). For each block, 256 E-W and N-S profiles of bathymetry and the residual geoid (degree-10) data were extracted from the grid, and these profiles were stacked to obtain average E-W and N-S bathymetry and geoid profile data sets. For these averaged profiles, admittance was computed using power spectrum of the bathymetry and cross spectra of the bathymetry and geoid data using the formula (after McKenzie and Bowin, 1976)

Where, *k* is the wave number, Z(k) admittance between geoid and bathymetry; C(k) is cross spectrum of the geoid and bathymetry, $E_b(k)$ is power spectrum of the bathymetry.

We compare the observed admittance values with the theoretical admittance curves computed for a simplified mantle convection model assuming Newtonian convection in a uniformly viscous mantle layer, and neglecting the effect of inertia, self-gravitation and sphericity (Parsons and Daly, 1983). It is known that the observed geoid can be defined as a function of the deformation of the upper and lower surfaces of the convective layer and density structure imposed by the temperature distribution (T) within the convective layer (Black and McAdoo, 1988). The admittance values were computed assuming free slipping upper boundary and non-slipping rigid lower boundary conditions for two different temperature distribution

$$T \sim \frac{\sinh k^{|(1-z^{|})}}{\sin k^{|}} \dots \dots \dots (2)$$

$$T \sim \sin \pi z^{\dagger} \dots \dots (3)$$

Here, k' and z' represent non-dimensional wavenumber and depth of the convecting layer respectively. It is noticeable that model-2 produces higher amplitudes for theoretical admittance, and according to Parsons and Daly (1983) as the largest temperature variation occurs closer to the bottom boundary, the amplitude of transfer increases. The analytical expressions for the calculation of theoretical admittance curves for the above temperature distributions (after Parsons and Daly, 1983) and for different convective layer thicknesses (640 km and 400 km) were presented in detail in earlier publication (Black and McAdoo, 1988), and we do not reproduce here.

The computed E-W and N-S admittances were plotted along with the theoretical admittance curves as shown in Fig. 6. While, the E-W admittances show comparable amplitudes with the theoretical admittance curves derived for the mantle convection model (after Parsons and Daly, 1983; Black and McAdoo, 1988), the N-S admittances are significantly low, particularly at longer wavelengths. According to Black and McAdoo (1988), such low admittances (Fig. 6) are observed for the profiles oriented in the direction of plate motion (N-S in the case of BOB) due to weaker signal of convective rolls in the upper mantle along the plate motion.

4.3.2 Isostatic response study of the aseismic ridges

In order to estimate the Effective Elastic Thickness (Te) variation along the aseismic ridges in the BOB region (85°E and the Ninetyeast ridges), we computed the admittance of

bathymetry and the degree-50 residual geoid for eleven $4^{\circ} \times 4^{\circ}$ blocks (see Fig. 3E for location), with 2° overlap, along each of these ridges. For each block, we selected 88 E-W lines of bathymetry and residual geoid anomalies, and averaged them to define a profile representing the center of that block. For this averaged profile, admittance is computed using the formula given in (eq.1). The computed admittance values for each block in the flexural waveband (0.01 < k < 0.04) were compared with the theoretical admittance curves for different Te values derived based on the flexural model of isostasy (Sandwell and Renkin, 1988) as shown in Fig. 7A. Additionally, in order to incorporate sub-surface loads in the flexural response estimation (Forsyth, 1985), we carried out coherence analysis for the same blocks along both the ridges.

In the case of small-scale tectonic features, the admittance / coherence functions based on the periodogram method do not possess sufficient spectral resolution for robust estimation of Te (Lowry and Smith, 1994). In order to overcome this problem, Lowry and Smith (1994) suggested the Maximum Entropy Method (MEM) in which the spectrum of signal is estimated by extrapolating the signal's autocorrelation function (the inverse Fourier transform of the power spectrum) under the condition of it having maximum entropy. Here, for each block along these ridges, the observed coherence between MBA (see section 4.3) and bathymetry is computed using MEM and these were compared with the theoretical coherence curves (Fig. 7B) calculated for sub-surface to surface loading ratio f equal to 0.5 using the formula given by Forsyth (1985).

$$\gamma^{2} = \frac{(1 + (f/\xi(k))^{2}\phi(k))^{2}}{(1 + (f/\xi(k))^{2})(1 + f^{2}\phi(k)^{2})} \dots \dots \dots (4)$$

Where, $\phi(k) = 1 + \frac{Dk^4}{(\rho_c - \rho_w)g}$; $\xi(k) = 1 + \frac{Dk^4}{(\rho_m - \rho_c)g}$

 $\rho_{\rm w}$, $\rho_{\rm c}$, $\rho_{\rm m}$ represent density of water, crust and upper mantle respectively Flexural rigidity is $D = \frac{ET_e^3}{12(1-\nu^2)}$, E is Young's modulus, Te is effective elastic thickness, and $\nu = \text{Poisson's ratio.}$ The best-fit Te was chosen based on the least residual error of L_2 norm of observed minus theoretical coherence (Fig. 7B) within the flexural waveband (0.01 < k < 0.04). The choice of f equal to 0.5 is based on the earlier studies over the Ninetyeast Ridge in the southern part of the present study area (Grevemeyer and Flueh, 2000; Grevemeyer et al., 2001; Sreejith and Krishna, 2013). The Te values obtained from the admittance and coherence analysis along the 85°E and Ninetyeast ridges are given in Table 1.

4.4. 2-D flexural modeling

Te values obtained from the admittance / coherence analysis (Table 1) indicate the strength of the lithosphere due to all geological processes since the emplacement of the aseismic ridges/seamounts (Calmant et al., 1990; Watts, 2001). In order to analyze the Te at the time of emplacement of the loads, we further carried out flexural modeling along five multi-channel seismic profiles crossing over both the 85°E (S1-S5 in Fig. 3E) and Ninetyeast (P1-P5 in Fig. 3E) ridges. In order to reconstruct the topography of both the ridges at the time of emplacement, the overlying sediment layer was backstripped by assuming that the strength of the lithosphere remained constant since the emplacement. Accordingly, we considered a constant Te value for flexurally backstripping the sediment layer. In the case of Ninetyeast Ridge, we further removed the subduction related forebulge from the reconstructed topography. Here the maximum forebulge (w_b) has been estimated based on the method of Harris and Chapman (1994) by assuming a Te of 35 km for the subducted portion of the Indian plate (after Subrahmanyam et al., 2008) and it varies between 190-250 m at a distance of 100 km from the trench axis. Further, the ridge emplacement is modeled as a load on thin elastic plate to estimate the flexural Moho configuration. For the flexural calculation, we used a density of 2300 kg/m³ for the sediments (1.6-4.4 km/s) and 2650 kg/m³ for the ridge material (4.2-4.6 km/s) based on the seismic velocity data available for the 85°E Ridge (Rao and Radhakrishna, 2014) and Ninetyeast Ridge (Curray et al., 1982; Maurin and Rangin, 2009). A simplified density

configuration of 2900 kg/m³ for the oceanic crust (6.0-7.5 km/s) and 3300 kg/m³ for the upper mantle (7.9-8.1 km/s) were used from the available seismic refraction velocities in the BOB region (Curray et al., 1982). Finally, the backstripped sediment layer was added to the restored section to construct the present-day configuration of the ridge, and for the resultant crustal structure, the gravity and geoid response was calculated through forward modeling (Fig. 8A and 8B). The modeling was iteratively carried out for a range of Te values, and the best fit Te was obtained based on the minimum RMS error between the observed and calculated by matching both amplitude and wavelength of anomalies.

4.5. Geoid-Gravity modeling

The lithospheric structure was modeled along three regional transects AA' to CC' (two E-W, one N-S) in the BOB region (see Fig. 3C for profile locations). For this purpose, the gravity and geoid anomalies were extracted along these transects from the free-air gravity (Fig. 2B) and residual geoid anomaly (Fig. 3C) grids of the BOB. Here, we used degree-10 residual geoid to avoid the effects of deeper mantle density variations on the geoid. While transect AA' is a longitudinal section in the BOB cutting across the lithosphere of different ages (110-65 Ma) with variable sediment thickness, transects BB' and CC' run across the 85°E and Ninetyeast ridges and the intervening basins in the central and southern parts of BOB respectively. The sediment thickness along longitudinal transect AA', constructed from the results of Curray (1991) and Krishna et al. (2001a), gradually decreases from > 9 km in the north to about 2 km in the south. Curray (1991) divided the entire sediment column into pre-collision sequence (pre-Eccene) with velocities > 4.4 km/sec and post-collision sequence (post-Paleocene) with velocities range from 1.6 to 4.3 km/sec. We further considered the division of post-collision sediment sequence into two layers; i) upper layer with velocity 1.6-3.0 km/sec ii) lower layer with velocity 3.0-4.4 km/sec (Michael and Krishna, 2011). The transect BB' was constructed along the multi-channel seismic reflection profile (MAN-01) (Gopala Rao et al., 1997),

whereas, transect CC' is close to the NGDC single channel reflection profile and constructed from Radhakrishna et al. (2010). For the MAN-01 profile, we split the sedimentary sequence into three major layers: Present to upper-Miocene (1.7-3.0 km/sec), middle Miocene to upper Oligocene base (3.0-4.4 km/sec) comprising of post-collision sequences, and a lower layer of the Eocene pre-collision sequence (> 4.5 km/sec).

In the initial model, the geometry of the Moho was constructed either by utilizing the crustal velocity information (for the N-S transect AA') available from the published seismic refraction data (Curray et al., 1982) or adding 6 km thick normal oceanic crust to the basement (in the case of transects BB' and CC'). The Base of the lithosphere (LAB) was initially chosen based on the plate cooling model for the lithospheric ages in the region (Parsons and Sclater, 1977). Available seismic velocities along transects allowed us to consider the densities 2300 kg/m^3 for the upper sediment layer from present to upper Miocene, 2400 kg/m³ for the middle layer from middle Miocene to upper Oligocene base and 2550 kg/m³ for the pre-collision sediment layer. A two-layer oceanic crustal density configuration having 2700 kg/m³ for the upper crust and 2900 kg/m³ for the lower crust was considered (after Mukhopadhyay and Krishna, 1991). A generalized density configuration for both lithospheric mantle and underlying asthenosphere were chosen as 3300 kg/m³ and 3250 kg/m³ respectively. As the sediment and oceanic crustal densities are reasonably constrained by the seismic velocity data, we computed the gravity / geoid responses for different models by adjusting the geometry of Moho and LAB interface adopted for the initial model, and these responses were compared with the observed anomalies. The final Moho and LAB geometries were obtained by minimizing the discrepancies between the computed response and the observed anomalies through several iterations until a good fit was achieved. The gravity response of the model was calculated using the algorithm of Talwani et al. (1959) and the geoid response of the model was computed based on the algorithm of Chapman (1979) and Ayala et al. (1996). These algorithms are useful to compute the gravity / geoid response for any arbitrary shaped 2-D body by approximating the body to a 'n' sided polygon. The polygons are defined with the vertices in coordinate system and the respective responses (gravity / geoid) are calculated using the line integral method. Hence, these forward modeling methods are highly useful to compute the gravity / geoid responses in the case of complex geological structures.

Using the above methodology, we computed the best-fitting lithospheric structure along the three regional transects AA' – CC' in the BOB region (Fig. 9A-C). Additionally, as the N-S profile (AA') lies across oceanic lithosphere of age ranging from 65 to 110 Ma, we incorporated the temperature-dependent lithospheric mantle density structure defined by the global plate cooling models (Parsons and Sclater, 1977; Stein and Stein, 1992) while deriving the lithospheric structure.

5. RESULTS AND DISCUSSION

The wavelength filtering analysis of geoid anomalies in the BOB region (Fig. 3C-F) revealed that the amplitude of the degree-10 residual geoid varies from -10 m to +12 m with a prominent high parallel to the Andaman subduction zone close to the Ninetyeast Ridge between 10°N to 5°S and a low corresponding to the rifted basins (Fig. 3C) offshore Eastern Continental Margin of India (ECMI). The degree-30 residual geoid map (Fig. 3D) reveals a distinct geoid high corresponding to the Ninetyeast Ridge between 2°N to 16°N. However, to the north of 4°N, where the Ninetyeast Ridge intersects with the Andaman subduction zone, the geoid anomalies of the ridge merge with the arc parallel anomalies and become indistinguishable. The signature of the Ninetyeast Ridge can be clearly identified in degree-50 residual geoid anomaly map (Fig. 3E), and is seen to extend up to 17°N as series of localized highs. Overlapping of long wavelength arc parallel geoid anomalies and short wavelength anomalies of the ridge at places indicate the interaction of the Ninetyeast Ridge compensation with the subduction process. The 85°E Ridge is weakly identifiable in the degree-30 residual geoid anomaly map (Fig. 3D),

whereas, on the degree-50 residual geoid anomaly map (Fig. 3E), the ridge distinctly appear as a belt of geoid lows in the north, and as geoid highs in the region south of 5°N to the Afanasy-Nikitin Seamount (ANS) location. It is to be noted that morphologically, the northern part of the ridge (north of 5°N) is buried below the thick Bengal Fan sediments, whereas; towards south the ridge is intermittently exposed on the seafloor with a thin sediment veneer. Further, the ridge is also wider having double peaked morphology in the north (Rao and Radhakrishna, 2014). These characteristic structural / geometrical variation and the differences in sediment thickness along the ridge between north and south is reflected on the gravity / geoid anomalies in north and south. West of the 85°E Ridge in the northern BOB, a localized high at 13° N is also observed in the degree-50 residual geoid anomaly map (Fig. 3E).

The band-pass filtered residual geoid anomaly map (10 < n < 50) reveals strong bi-polar signature of the geoid field with a belt of positive anomalies in the Andaman-Sumatra arc region and a negative field parallel to the ECMI with a transition zone in the central part of BOB (Fig. 3F). Subduction zones in general are characterized by long-wavelength (n=4-9) positive geoid anomalies due to the cold upper mantle slabs, particularly in the case of subduction zones that are associated with the deeper penetration of subducted slabs to the lower mantle, and an effective mantle viscosity change (Hager, 1984; King, 2002). The Andaman-Sumatra arc forms part of the Sunda subduction zone in the eastern Indian Ocean, along which the Indian and Australian plates together subduct below the Eurasian plate. Seismicity and other geophysical data along this subduction zone demonstrates that maximum depth of the down going lithospheric slab varies from 600-800 km below Java, 250-350 km below Sumatra and ca. 200 km beneath the Andaman arc region (Hayes et al., 2012). The long-wavelength residual geoid anomalies in the eastern Indian Ocean region did not reveal significant geoid signature associated with the Sunda subduction zone. However, in the band-pass filtered (10 < n < 50) residual geoid anomaly map (Fig. 3F), a geoid high zone is observed between the Ninetyeast

Ridge and the Sumatra trench, which could be related to mass heterogeneities within the upper mantle below the region. Elsewhere, the low geoid anomalies observed in the northern BOB and along the ECMI (Fig. 3F) are attributed to gradual transition of oceanic lithospheric mantle of BOB to more continental affinity beneath the eastern Indian shield.

5.1 Upper mantle characteristics

As described in section 4.3.1, the geoid-bathymetry admittance at wavelengths > 800 km provides some insights on thermal structure of the upper mantle beneath the BOB. The theoretical admittance curves derived for two convective layer thicknesses (400 km and 640 km) defined by different temperature distribution models (1 and 2 described in section 4.3.1) are compared with the observed admittance values (Fig. 6A) at wavelengths greater than 800 km (k < 0.008). As the spatial extent of study area in the BOB is confined to $13^{\circ} \times 13^{\circ}$ region, the observed admittance is available at only one wavenumber that falls within the wavelength band of the mantle convection model. The peak value of the observed admittance at this wavenumber (k~0.004) is very high (>8 m/km) and closely matches with the theoretical admittance curve computed for a 640 km thick convective layer with the temperature distribution defined by model-2, which means that maximum temperature variation lies at the bottom of the convecting layer. However, for higher wavenumbers k > 0.01, the observed admittance values are not related to the mantle convection processes, therefore they are not comparable with the theoretical curves derived for convection model. The admittance values at these higher wavenumbers could be ascribed to the compensations related to flexure of the BOB lithosphere. It is evident from fig. 6A that observed admittance values fall within the pure flexure curves at the higher wavenumbers. In order to bring these two processes into a single unified model, Black and McAdoo (1988) presented theoretical admittance due to a combination of convective loading and flexural loading of the lithosphere following the relations developed by Forsyth (1985). Accordingly, we computed theoretical admittance curves for combined flexure and convection model for different values of f, which is the ratio of convective load to the flexural load. For this purpose, we consider a 640 km thick convecting layer with temperature distribution defined by model-2 and an effective elastic thickness Te of 10 km for the BOB lithosphere. The resultant curves along with observed admittance values are presented in Fig. 6B. While, the admittance value at the lowest wavenumber (Fig. 6B) supports mantle convection, the admittance values at higher wavenumbers (k > 0.01) agree well with flexural compensation.

5.2 Mode of compensation below 85°E and the Ninetyeast ridges

Majority of the GTR values computed for two band-pass filtered geoid and topography datasets (Figs. 5A and 5B) along the 85°E and the Ninetyeast ridges range from 0 to 2 m/km at both wavelength scales, which indicate that these two ridges are compensated at shallow depths. However, at few places an exceptional GTR values are observed showing >2 m/km and negative values. For the 85°E Ridge, the GTR values are higher ~ 3-6 m/km between the equator and $6^{\circ}N$, > 4 m/km at $17^{\circ}N$ in the 10-120 degree dataset, and negative GTR values of -2 to -3 m/km between 9°N and 12°N in the 50-120 degree dataset (Fig. 5A). The blocks in the north at around 18°N show higher GTR values in 10-120 degree dataset as this part of the ridge is under the influence of continental edge effect of the adjoining Indian lithosphere. A correlation of degree-10 and degree-50 residual geoid anomalies of the ridge further reveals that the blocks between equator and 6°N falls in the region of stronger gradients in the degree-10 map. While, the GTR values for 50-120 degree dataset indicate compensation of the ridge at shallow level, the higher GTR values observed over these blocks in the 10-120 degree dataset could be due to the influence of deeper mass variations. On the other hand, between 9° and 12°N, the negative GTR values of -2 to -3 m/km observed in the 50-120 degree dataset changes to 0-2 m/km in the 10-120 degree dataset (Fig. 5A) and the scatter plots do not seem to indicate

any data artifacts at this location. A close look at the image enhanced gravity maps (Fig. 2 of Bastia et al., 2010) and the degree-50 residual geoid anomaly map (Fig. 3E present study) of the 85°E Ridge reveal minor breaks in the trend of the ridge and presence of localized body west of the ridge around 12-13°N. It is not clear whether the negative GTR values in 50-120 dataset could be ascribed to these shallow structural heterogeneities. For the Ninetyeast Ridge, the 10-120 degree dataset gave rise to slightly higher GTR values of 3 m/km between 6°N and 8°N and negative GTR values between 9° and 13°N with the lowest value of -2 m/km at 11°N (Fig. 5B). The higher GTR values between 6°N to 8°N correlate with the prominent geoid high zone observed in the degree-10 residual geoid map and could be due to deeper mass heterogeneities. The negative GTR values observed between 9° and 13°N in the 10-120 degree dataset become close to 0 m/km in the 50-120 degree dataset. As the Ninetyeast Ridge between 9° and 13°N is closer to the Andaman trench, we believe that the negative GTR values observed in the longer wavelengths could be due to complex interaction of ridge subduction below the Andaman-Sumatra arc.

The residual geoid-bathymetry admittance analysis and MBA-bathymetry coherence analysis carried out along the 85°E and Ninetyeast ridges provide further insights with regard to the nature of compensation as well as mode of emplacement (on ridge/ off ridge). The Te values obtained from both the methods, in general, show similar variations along both the ridges and over the ANS (Table-1). In selective blocks, the observed admittance values (in the diagnostic waveband) show large scatter and do not match with the theoretical curves (e.g., Block-9 on 85°E Ridge and Block- 6 on Ninenyeast Ridge, see Fig. 7A), hence, the Te value becomes less reliable for these blocks. This could be due to lack of ridge expression yielding low coherence between bathymetry and geoid/gravity in the case of 85°E Ridge, or due to close proximity of the trench around 8°N in the case of Ninetyeast Ridge (see Figs. 2 & 3E). However, the Te was considered from the coherence analysis (Fig. 7B) in these blocks. As stated earlier, for the

coherence analysis, we considered a value of 0.5 for the loading ratio f as the modeled f values elsewhere along the Ninetyeast Ridge range from 0.5 - 0.7 (Grevemeyer and Flueh, 2000; Grevemeyer et al., 2001) and 0.3 - 0.6 (Sreejith and Krishna, 2013).

Further, Te values at the time of emplacement of these ridges were obtained through flexural modeling along five seismic profiles crossing the ridges. The study reveals Te values of 10-15 km for the 85°E Ridge, 20 km for the ANS and 25-30 km for the Ninetyeast Ridge. The crustal structure modeled along these profiles (as described in section 4.4) for the best-fit Te values are presented in Figs. 8A and 8B. In order to check the sensitivity of Te on the choice of upper mantle density, we computed flexural response for a range of upper mantle densities from 3200 to 3400 kg/m³ and found that variation in Te is not significant (< 2 km). While, the admittance/ coherence analysis as well as the flexural modeling give rise to similar Te values for the 85°E Ridge; for the Ninetyeast Ridge (north of 10°N), the Te values obtained through flexural modeling are relatively higher than from the admittance/ coherence analysis. This difference in Te values is ascribed to subsequent modification of the ridge topography by subduction related forebulge effect of the Andaman-Sumatra arc system in this region.

The present Te values obtained from both 85°E and Ninetyeast ridges are combined with the Te estimates published for the 85°E Ridge (Sreejith et al., 2011) and for the Ninetyeast Ridge (Sreejith and Krishna, 2013) to provide a comprehensive picture on the Te variation along the ridges (see Figs. 8A and 8B). It is notable that for the 85°E Ridge the estimated Te values range from 10 to 15 km except in the area of ANS, where comparatively higher Te values of 20 km is observed. However, along the Ninetyeast Ridge, the Te values are about 20-30 km for the ridge north of 2°N latitude. The Te values extracted from the global Te estimates presented by Watts et al. (2006) for the Ninetyeast Ridge in the study area in general show higher Te values (>25 km) for the ridge north of 2°N and around 7-18 km between equator and 6°S. We observed that Te values obtained in this study are in general agreement with the global Te values (Watts et al., 2006), however, our estimates are better constrained due to the availability of regional seismic data for the ridge north of 10°N which is progressively buried below the Bengal Fan sediments.

5.3 Lithospheric Structure

LAB depths derived from combined modeling of gravity and residual geoid anomalies along profile AA' (Fig. 9A) range from 122 km in the north to 73 km in the south. Further, the degree-10 residual geoid anomalies (Fig. 3C) along this profile reveal that the regional geoid high observed in the southern BOB turns into a regional low towards northern BOB. This change in both residual geoid and gravity anomalies is explained by the deepening of LAB towards north on profile AA' and the geometry of upper crustal layers in the model further indicates the extent of sediment depocenter of the Bengal Fan.

In order to understand the effect of temperature-dependent mantle densities at LAB depths, we further modeled the lithosphere geometry along profile AA' taking into consideration of two different thermal structures, the GDH1 (Stein and Stein, 1992) and the PSM (Parsons and Sclater, 1977) for the BOB lithosphere. The isotherms defined by these two thermal models were utilized to constrain the initial density layering for the lithospheric upper mantle and this geometry was changed iteratively to fit with the observed geoid and gravity anomalies along the profile. The final modeled geometry (see Fig. 9A) reveals that depth to the LAB varies from 77 to 127 km for the GDH1, while the LAB depths derived from the PSM model are systematically deeper by 15 - 20 km than GDH1. We observed that the modelled LAB geometry for the PSM matches well with the LAB values retrieved from the global grid of upper mantle structure PM_v2_2012 (Priestley and McKenzie, 2013) except for the difference in lateral placement of their deepening. On the other hand, the modeled LAB values from the GDH1 agree well with the estimates of LAB based on the available seismological data in the BOB (Singh, 1990; Mitra et al., 2011; Bhattacharya et al., 2013). Hence, we consider the

interesting to note that the modeled LAB depths (127-77 km) below BOB are significantly deeper than the LAB values derived in the Pacific Ocean lithosphere (Rychert et al., 2011; Rychert et al., 2012 and references there in) of similar age (65-110 Ma). This observation would be consistent with the presence of colder upper mantle below the BOB, as earlier suugested by Brune et al. (1992), causing the deepening of isotherms at the LAB level. Alternatively, the deeper LAB below the northernmost BOB may indicate a greater continental affinity of the BOB lithosphere. The modeled LAB geometry along two E-W profiles, profile BB' (Fig. 9B) and profile CC' (Fig. 9C), which reveal that LAB is almost flat below the 85°E Ridge, whereas minor localized thinning is observed below the Ninetyeast Ridge. This eastward thinning of the lithosphere observed in both the profiles could be ascribed either to the plume related thinning below the Ninetyeast Ridge (Davies, 1994) or due to the interaction of the ridge with the Andaman-Sumatra subduction zone in this region. However, resolving these two competing

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processes needs further detailed investigation.

Generally, in plume-affected basins such as the northern Atlantic, high amplitude geoid anomalies are associated with shallow basement depths (Jung and Rabinowitz, 1986). In contrast, presence of deeper basement depths and low amplitude geoid anomalies in the BOB region suggest that the mantle rheology is not significantly affected by mantle plume activities, despite the presence of two major hot spot traces on the seafloor - the 85°E Ridge and the Ninetyeast Ridge. Ito and Keken (2007) noticed that the rise of swells formed by the hot spot volcanism decreases with time and becomes undetectable beyond 80 Ma. As the volcanism leading to the emplacement of the 85°E and Ninetyeast ridges in the BOB are older (> 80 Ma), the swells associated with the initial volcanism will have subsided. Earlier derived crustal models show the presence of thicker crust with and without underplating below these ridges

modeled LAB along profile AA' based on GDH1 as a representative one for the BOB. It is

(Detrick and Watts, 1979; Mukhopadhyay and Krishna, 1995; Subrahmanyam et al., 1999; Grevemeyer et al., 2001; Krishna et al., 2001b; Krishna, 2003; Subrahmanyam et al., 2008; Radhakrishna et al., 2010). The present analysis revealed that both ridges, in general, are compensated at shallow depths. The Te values of 10-15 km estimated for the 85°E Ridge further confirms its emplacement in off-ridge environment. The consistently higher Te value of 20 km over the ANS is not in agreement with the low Te values reported by Paul et al. (1995), however, match very well with the Te values obtained by Watts et al. (2006) in this region. Recent geophysical studies and ⁴⁰Ar/³⁹Ar dates of the ANS (Krishna et al., 2014) suggested that the ANS was emplaced in two phases with the construction of main plateau in the initial phase around 80-73 Ma close to the spreading center, and emplacement of seamount highs in later stage at about 67 Ma. Hence, the higher Te values obtained over the ANS could be related to the later emplacement over a relatively stronger lithosphere. The Ninetyeast Ridge is also characterized by higher Te values of 20-30 km for the ridge segment north of 2°N, suggesting its emplacement in the intraplate setting. On the other hand, the scattered Te values observed by earlier investigators for the ridge south of 2°N revealed that this part of the ridge had evolved due to frequent ridge jumps in the vicinity of the Kerguelen hot spot during the rapid northward migration of the Wharton spreading ridge (Royer et al., 1991; Krishna et al., 1999; Sager et al., 2010; Krishna et al., 2012). The modeled lithosphere structure across these ridges suggests a lack of plume impact at the LAB, although there is minor eastward thinning beneath the Ninetyeast Ridge.

7. CONCLUSIONS

Forward as well as flexural modeling of geoid-gravity anomalies of the BOB have provided a new understanding on lithospheric structure and nature of emplacement of the 85°E Ridge, ANS and Ninetyeast Ridge. The major conclusions of the present study are described below:

- The seismically constrained geoid-gravity modeling defined by the temperature-dependent mantle densities based on the GDH1 model has revealed that depth of LAB is about 127 km for the 110 Ma age oceanic lithosphere in the northern BOB region, while it is about 77 km for the 65Ma age lithosphere in the south BOB. The deepening of LAB towards the north, and the geometry of upper crustal layers constrain the extent of sediment depocenter of the Bengal Fan. Comparison of LAB depths for the lithosphere of similar age in the Pacific Ocean suggested the presence of anomalously thick lithosphere beneath BOB.
- Spatial variations in GTR values as well as Te estimates along the 85°E Ridge, ANS and the Ninetyeast Ridge suggests variable isostatic mechanism beneath the ridges. While, the Ninetyeast Ridge could have been emplaced during the interaction of Kerguelen hot spot with the Wharton spreading ridge, present study reveals that the ridge north of 2°N was emplaced on a stronger lithosphere in an intraplate setting. The modeled lithospheric structure across these two ridges in the BOB suggests that they are dominantly crustal-scale features regionally compensated through flexure at shallow depths.
- Further, the admittance analysis between geoid and bathymetry had revealed that the admittance signatures at wavelengths > 800 km are compensated by an upper mantle convective layer below the BOB region.

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Table1 Estimates of Te obtained from the admittance (Geoid- Bathymetry), coherence (MBA-Bathymetry) analysis, and 2-D flexural modeling along the 85°E and Ninetyeast ridges. Refer Fig. 3E for location of blocks (B1-B11) and profiles (S1-S5 & P1-P5).

85 °E ridge & AfanasyNikitin Seamount					Ninetyeast ridge				
Block/ Profile name	Center latitude	Effective elastic thickness (Te) in km			Block/ Profile name	Center latitude	Effective elastic thickness (Te) in km		
		Admittance	Coherence (MBA)	Flexure modeling			Admittance	Coherence (MBA)	Flexure modeling
S1	17.25			15	P1	17.9			25
S2	16.5			10	P2	16.0			25
B1	16.0	7-17 (12)	12		B1	16.0	7-21 (14)	20	
S3	15.5			15	P3	15.0			25
B 2	14.0	12-20 (16)	18		B 2	14.0	7-20 (13)	20	
B3	12.0	12-26 (19)	18		P4	13.5			30
B4	10.0	19-21 (20)	22		B3	12.0	15-20 (17)	22	
B 5	8.0	17	12		P5	10.0			30
B6	6.0	8-18 (13)	18		B4	10.0	18-30 (24)	22	
B 7	4.0	7-25 (16)	18		B 5	8.0	20	18	
B8	2.0	25	28		B6	6.0		18	
B9	0		28		B 7	4.0	16-26 (21)	12	
B10	-2.0	20-36 (28)	22		B8	2.0	12-16 (14)	12	
S4	-3.0			20	B9	0.0	8-16 (12)	10	
B11	-4.0	25	20		B10	-2.0	10-20 (15)	14	
S5	-4.5			20	B11	-4.0	12-20 (16)	14	



Fig.1 Topographic map of the Bay of Bengal (BOB) and the contiguous regions. The present study area is outlined in red. Black rectangles (named blocks 1-3) are three overlapping areas of 13°×13° size selected for the long wavelength admittance analysis. The N-S oriented fracture zones (dotted lines) and the magnetic anomaly identifications (E-W trending short lines and the corresponding numbers) in the central Indian Ocean basin are adopted from Royer et al. (1989). Folded thrust front at the Burmese arc is adopted from Steckler et al. (2008). ASC–Abandoned Spreading Centre, ANS –Afanasy Nikitin Seamount. Inset- shows location of the BOB.



Fig. 2 (A) Bathymetry map from ETOPO1 (Amante and Eakins, 2009) (B) Satellite-derived free-air gravity anomalies from DNSC08 (Andersen and Knudsen, 2008) (C) Depth to the basement below BOB obtained by adding bathymetry and the sediment thickness grids (after Divins, 2003; Radhakrishna et al., 2010).



Fig.3 (A) Geoid map of the BOB (after Sreejith et al., 2013), (B) the power spectral density and coherence between the EGM2008 and DTU10 geoid models for spectral comparison, (C-F) the residual geoid maps of the BOB obtained after subtracting different long wavelength geoid contributions using smoothly truncated EGM2008 spherical harmonic coefficients (Pavlis et al, 2008) from the full spectrum geoid (after Sreejith et al., 2013). (C) degree-10 residual geoid anomaly map, (D) degree-30 residual geoid anomaly map, (E) degree-50 residual geoid anomaly map, (F) Band pass filtered residual geoid map between wavelengths corresponding to spherical harmonic coefficients of degree and order > 10 and < 50. Dashed lines in (C) represent location of the modeled lithospheric profiles. Trace of both 85° E and Ninetyeast ridges is shown on the residual geoid maps. Blocks 1 to 11 shown in (E) are selected for the admittance and coherence analysis along these ridges, and profiles marked (S1-S5) across 850E Ridge and (P1-P5) across the Ninetyeast Ridge are seismic lines used for the study of 2-D flexure modeling.



Fig.4 Band pass (330 - 4000 km) filtered maps of bathymetry and the geoid expanded from spherical harmonic coefficients of degree 10–120 with the edges attenuated by a cosine filter. Rectangles (in the right panel) shown along the 85°E and the Ninetyeast ridges represent various overlapping blocks (85ER1-22 and NER1-22) chosen for the Geoid-Topographic Ratio (GTR) analysis along the ridges. Details are discussed in the text.





Fig. 5A) Plot showing variation of computed GTR values for different blocks (shown in Fig. 4) along the 85°E Ridge (left panel).Both degree 10-120 (red) and degree 50-120 (blue) wavelength scales of residual geoid and bathymetry are used for this analysis. Representative scatter plots for two selected blocks are also presented. Further the along ridge variation of GTR values are also shown in plan view (right panel) for better perception



Fig. 6 Comparison of the observed (black dots indicate E-W admittances; red dots represents N-S admittances) and theoretical (green and blue lines) admittance between the geoid and bathymetry for three $13^{\circ} \times 13^{\circ}$ blocks (Fig. 1 for location) in the BOB region. A) Here, the theoretical admittance curves are computed for the mantle convection model involving different convective layer thicknesses (640 km and 400 km) and for two different temperature distributions within the convecting layer; model-1 (in eq. 2) imposes temperature variation at the top boundary (blue curves) and model-2 (in eq. 3) imposes temperature variation at the top boundary (blue curves) and model-2 (in eq. 3) imposes temperature variation at the bottom boundary (green curves) (after Black and McAdoo, 1988). The observed admittance values falling within the diagnostic waveband (0.004 < k < 0.008) are compared with the theoretical curves derived for mantle convection model. The theoretical admittance curves (thick black lines) for pure flexure model (after Sandwell and Renkin, 1988) also presented in the figure. B) Here, the theoretical admittance curves are computed for the combined convection and flexure model by incorporating combination of convective loading and flexural loading of the lithosphere for different values of 'f', which is ratio of convective load on the bottom to the flexural load on the top of the lithosphere (after Black and McAdoo, 1988). In this model, we considered thickness of the convecting layer 640 km with temperature structure defined by model-2, and effective elastic thickness (Te) of 10 km for the BOB region. Details are discussed in the text.



Fig. 7A Plots showing the observed Admittance values (dots) between bathymetry and residual geoid (degree-50) for selected blocks (refer Fig. 3E for locations of blocks) along the 85° E and Ninetyeast ridges. The theoretical admittance curves (solid lines) for Te values 5 to 35 km are presented for comparison with the observed admittance values in the flexural waveband ($0.01 \le k \le 0.04$).



Fig. 7B Plots of observed coherence (black dots) between MBA and bathymetry computed for selected blocks (refer Fig. 3E for locations of blocks) over 85° E and Ninetyeast ridges. The theoretical coherence curves (in red) computed for Te values 5, 10, 20, 30 km and a uniform f equals to 0.5 are plotted for comparison. The best-fit Te (dashed blue line) was chosen based on the least residual error of L2 norm between observed and theoretical coherence within the flexural waveband (0.01 < k < 0.04) indicated with vertical dashed lines.





Fig.8 2-D flexural modeling of selected profiles (A) S1-S5 across the 85°E Ridge and the ANS, (B) P1-P5 across the Ninetyeast Ridge. See Fig. 3E for profile location. For each profile, the observed gravity and residual geoid (degree-50) anomalies (continuous line) along with the calculated responses for different Te values and their RMS error (value in bracket) is shown in the upper panel, and the resultant crustal model for the best-fit Te is shown in the lower right most panel, the Te values obtained in this study and those from Sreejith et al. (2011) for the 85° E Ridge and from Sreejith and Krishna (2013) for the Ninetyeast Ridge are plotted longitudinally.







Fig.9 Lithospheric structure along three regional profiles (AA'-CC') in the BOB (see insert for profile location) based on constrained gravity-geoid modeling. In each of the models, the top and middle panel shows the residual geoid (degree-10) and gravity anomalies along with their calculated responses; while, the bottom panel shows the lithospheric structure obtained by adopting the density values shown in bracket. Crossing points with other profiles are shown as vertical dashed lines. A) N-S profile (AA') along the BOB covering the lithosphere of age between 110-65 Ma. The LAB geometry obtained for the temperature-dependent mantle densities referring to two different plate-cooling models, GDH1 (blue dashed line) and PSM (red dashed line) are also shown. Thick black line represents the LAB geometry from the PM_v2_2012 global shear wave velocity model of Priestley and McKenzie (2013). The numbers in the model indicate refraction velocities and the inverted triangle on the top of the model refer to approximate age of the lithosphere. B) & C) represent modeled lithospheric structure across the 85° E and the Ninetyeast ridges in the northern part of BOB (profile BB') and southern part of BOB (profile CC') respectively. Details of construction of these transects are discussed in the text.