Effective elastic thickness of the continental lithosphere with particular reference to India – Eurasia collision system: a review

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Abstract

The effective elastic thickness (EET) of lithosphere represents a part or parts of lithospheric thickness that responds/respond to the long-term geological / topographical loads elastically and is often estimated through analyses of gravity and topographic data. The EET defines the resistance of the lithosphere to bend under vertical loads and has a significant role in regulating the geodynamic evolution of both the continental and oceanic plates. The estimates of the EET derived from geophysical data are consistent with rheological models in the oceanic regions. However, there are extensive debates on estimates of the EET and rheological models over the continental areas; differences are probably due to the complex structure and history of the continental plates. For instance, according to one model of continental rheology, popularly known as the ‘Jelly Sandwich’, the mechanical strength of the lithospheric plate is distributed
in the upper crust and the lithospheric mantle. In another model, dubbed as ‘Crème Brulee’, the lithospheric mantle is weak, and the mechanical strength of the lithosphere is limited to the upper portion of the crust. These model differences have arisen because of inconsistency in the results obtained using different data sets, e.g. the distribution of earthquakes, EET, gravity anomaly and rheology. This review article discusses the evolution of these opposing models and the critical necessity to resolve this issue.

**Keywords:** Effective elastic thickness, isostasy, flexural modeling, continental rheology
Introduction:

The lithosphere of the Earth is characterized by a relatively rigid outer layer that is situated above the ductile and weaker layer of the upper mantle known as the asthenosphere. It was introduced to explain the sustenance of the undulating topography on the Earth’s surface, such as the mountains, depositions of sediments in basins, and erosion of plateaus, despite the large vertical strains they generate. The idea was that the elastic strength of the lithosphere maintains the topographic / geological loads by accommodating the vertical strains produced by the surface and subsurface loads. Later, this concept played a significant role in developing the theory of plate tectonics. Therefore, understanding the properties of the lithosphere becomes vital in deciphering the pattern of the tectonic evolution and dynamics of the planet. The lithosphere generally tends to have higher average densities, cooler temperatures, and higher average seismic velocities than the asthenosphere.

Decades before the concept of the lithosphere was introduced, there were models of a light and rigid crust floating on a dense and fluid mantle. These were based on isostatic considerations, which suggest that the outermost layers of the Earth are in a state of hydrostatic equilibrium. According to the initial and most widely acknowledged models of floatation, the height attained by such a floating crustal block depends on its thickness and density. These models are known as Airy and Pratt models of isostasy, and they envision individual blocks of crust and mantle supporting the surface loads. The mass surplus at the surface, within a loaded block, is balanced by a mass deficit beneath. The pressure due to overlaying material remains constant throughout the Earth at some depth known as the depth of compensation (Fig. 1). According to Airy model, variations in the thickness of a uniform-density crust compensate for the excess loads. In contrast, in the Pratt model, the necessary support is provided by the lateral variations in density of either crust or subcrustal mantle.
Both models predict gravity anomaly as the key observable sensitive to changes in mass distribution due to topographic loading and its associated compensation. Later studies revealed that the Pratt model could account for the gravity anomalies associated with the lateral changes in density within the suboceanic crust and mantle. In contrast, the Airy model can describe the gravity anomalies associated with the subcontinental crustal structure. However, there are locations on the Earth (e.g., The Hawaiian Islands and Himalaya – Tibet region) where the Airy and Pratt models fails to account for the observed gravity anomalies adequately. This is because both the models are highly idealized in such a way that they do not account for the inherent rigidity of the lithosphere and only consider the state that the crust and the subcrustal mantle would approach given a sufficiently long time.

Vening Meinesz used the formulation by Hertz for the flexure of a thin elastic plate due to a concentrated point load to model the bending of the lithosphere under topographic loads. This illustrated that a model in which the topographic load is supported regionally rather than locally could better describe the observed gravity anomalies. According to the Hertz model, the extent of flexure of the elastic plate is controlled by the density difference between the underlying material and the material that infills the flexure, gravitational acceleration, and flexural rigidity of the plate \( D \), defined as:

\[
D = \frac{ETe^3}{12(1 - \nu^2)}
\]

Where, \( E \) is Young’s modulus, \( \nu \) is Poisson’s ratio, and EET is the effective elastic thickness. Flexural rigidity \( D \) measures the plate’s deformability or stiffness. Its magnitude determines the degree to which the elastic plate bends under applied loading. The lithosphere comprises materials of varying rheologies (elastic, plastic, viscous, etc.). As a result, flexural rigidity is primarily determined by the EET, which is the thickness of the lithosphere with pure elastic rheology. The EET generally does not represent a depth to any boundary within the lithosphere.
It is only a mathematical analog of the integrated strength of the lithosphere\textsuperscript{11–13}. However, the EET provides an adequate measure of the flexural rigidity of the lithosphere and thus possesses a true geological significance. Many subsequent investigations concluded that the effective elastic thickness of the lithosphere controls most of the tectonics of the continents\textsuperscript{14–16}.

The knowledge of the EET is crucial in determining how the lithosphere may respond to surface and subsurface loads. Furthermore, the EET variations in the lithosphere could help us explain some of the observed differences between the actual flattening of the Earth and the flattening predicted by hydrostatic theory\textsuperscript{17}. This is useful while studying intraplate deformation, lunar and solar tides, and the figure of the Earth. Some studies also suggest a relationship between the lunar semidiurnal component of the tidal gravity anomaly and EET\textsuperscript{18,19}. The EET can also be used to scale the viscosity in thin viscous sheet models that attempt to calculate stress and intraplate deformation due to plate boundary forces\textsuperscript{20}. The lithospheric flexure also interacts with the atmosphere and asthenosphere, profoundly impacting landscape evolution and deep processes such as mantle convection\textsuperscript{3}. However, significant regional variances exist in the currently estimated EET values worldwide. Low values of the EET are usually associated with mid-ocean ridges and some continental rifts, while the higher EET values typically correlate with the cratonic regions. The Phanerozoic orogenic belts are generally associated with low (~20 km) as well as high (~150 km) EET values\textsuperscript{3,21}.

\textbf{Conventional methods of the EET computation from flexural isostasy:}

In flexural isostatic studies, the observed topography and gravity anomalies are compared with the theoretical models to estimate the elastic properties of the lithosphere. The simplest scenario is a vertical surface load flexing the plate, with the magnitude of the resulting curvature determined by its EET. When the EET is zero (i.e., the plate has no rigidity), the load
is supported in hydrostatic equilibrium, according to models such as Pratt-Hayford or Airy-Heiskanen. However, as the EET or flexural rigidity of the plate increases, the load receives greater support from the internal stresses of the plate, leading to a decrease in plate curvature. This change in curvature causes variations in layered density structures within the lithosphere, which generate gravity anomalies that can be observed from the surface. Consequently, comparing the theoretical gravity generated from the flexural compensation models of the observed topography can yield an estimate of the EET.

The three most common methods of computing the EET are (i) forward modeling of gravity and topography in the spatial domain, (ii) inversion of the spectral properties of the gravity field and topography, and (iii) thermo-rheological modeling. In the first method, the comparison of the theoretical and observed gravity from the flexural modeling is carried out in the space domain. In the second method, using Fourier or wavelet transforms, this comparison is made in the spatial frequency or the wavenumber domain. The thermo-rheological modeling assumes that the strength of the lithosphere is limited by brittle failure, which depends on confining pressure in its uppermost part and ductile flow controlled by both confining pressure and temperature in its lowermost part. The Brace-Goetze failure envelope curves suggest that elastic strength increases with depth and then decreases according to the brittle and ductile deformation laws. In the continental regions, the yield strength envelope is complex, and there may be more than one brittle and ductile layer depending upon the region’s temperature and pressure profile. Using data from experimental rock mechanics together with a temperature profile, it is possible to construct yield strength envelope profiles from which we can define the associated EET values. Suppose the estimated EET is greater than the crustal thickness or the brittle-ductile transition (BDT). In that case, it indicates that the elastic-ductile mantle also contributes to the strength of the lithosphere.
However, this method heavily depends on assumed or laboratory-determined parameters such as the composition, pore fluid factor, elastic constants, strain rates, etc. The spectral method of the EET computation depends on derivatives of transfer functions of the gravity and topography data in the spectral domain. The spectral approaches made it easy to calculate the spatial and temporal variation of the EET. The early spectral studies used the Bouguer admittance, Fourier transform periodogram method, and produced the low EET values. Later investigations suggested that the low values could be the result of subsurface loads that were uncorrelated with surface topographic loads. They proposed using Bouguer coherence instead of Bouguer admittance because it is less sensitive to the surface-to-subsurface loading ratio. The use of this approach in North America shows that it produces both, low and high, EET values in the cratonic zones. Following investigations used new techniques, such as maximum entropy estimators and multitapers. McKenzie and Fairhead suggested using free-air admittance rather than the Bouguer coherence because, unlike subsurface loads, the surface topography is known and its gravity effect is not removed from the free-air gravity anomaly. They claimed that the Bouguer coherence approach is influenced by the erosion effects and thus overestimates the EET values. Perez-Gussinye and Watts used Bouguer coherence and free-air admittance methods, and found high values of the EET (>70 km) over the Archean and Early Proterozoic (<1.5 Ga) East European craton, and low values (10-45 km) over the flanking Caledonian, Variscan, and Alpine orogenic belts. Similar results were found for North and South America, Australia and Africa.

However, decades of flexural studies still need to arrive at a final consensus regarding these different models concerning the strength of the continental lithosphere. In the forward modeling approach, the gravity anomaly due to surface (topographic) load and the flexural compensation associated with it are computed for varying values of the EET and compared with the observed gravity anomaly. The best approximation of the EET for the region is then
estimated as the one that minimizes the difference between the observed and calculated gravity
anomalies and flexure surfaces. If the gravity anomaly and depth to basement data are
available, we get the most reliable estimates of the EET values. While forward modeling is
a satisfactory way to estimate the EET, the number of sites where information on both load and
basement depth are available is limited. The inverse method first converts the gravity and
topography data to the wavenumber domain using Fourier or wavelet transforms. Then the EET
is determined by calculating the transfer function between them as a function of wavelength
and then comparing it with the model’s predictions. This procedure assumes periodic or
reflecting boundary conditions, making them inapplicable near plate boundaries with
significant topography or tectonic loading conditions. Supposing their mechanical problem
is formulated identically in terms of the surface and subsurface loads, internal boundary
conditions, and the area chosen for modeling, we can expect both approaches to yield the same
results. However, it is scarcely the case. The elastic thickness estimates from inverse spectral
methods mostly disagree with those from the forward modeling approach. It is also
possible that the two methods may be giving estimates of the EET from different geological
timespans - either at the time of loading or the present day, depending on the geological age
and the tectonic setting (continent or ocean) of the region.

The thermo–rheological modeling approach also has been incapable of giving a final
solution to the debate. Tesauro et al. compared the global continental EET estimates from
thermo – rheological parameters with those from the Bouguer coherence deconvolution using
wavelets of the spectral inversion method. It is observed that the differences in the EET values
over Precambrian shields and platforms are pretty high even though the values of the EET in
the Phanerozoic orogens are very similar (Fig. 2). The rheological EET is much less than the
spectral EET in the Precambrian shields and platforms, with up to a 50 km difference in certain
places. The difference in the EET estimates may also be due to spectral and rheological
calculation approaches (Fig. 2). The relatively flat topography associated with some cratonic and basin areas increases the uncertainty of the EET estimation using spectral inversion methods. The uncertainties of the rheological modeling depend on the uncertainties of the thermal model used, and the uncertainties in thermal models increase considerably in cold Precambrian cratons. Tesauro et al. concede that since both methods can have relatively high uncertainties in the cratonic regions, estimating the EET using both methods in such areas is challenging.

The inverse spectral technique remains the one that has advanced considerably during the past half a century since its inception. There has been significant advancement, especially in the method of spectral estimation of the gravity fields. In the early days, the periodogram estimation of admittance and coherence was the norm. Nowadays, we have the maximum entropy, multitapers, and wavelets at our disposal through which we can estimate the admittance and coherence functions. However, in a recent study, Simons and Olhede challenge the statistical validity of inverting ratios data to transform products and estimation of the EET from such ill-posed observables. They suggest using a method based on the maximum likelihood estimation theory, which can give unbiased results with minimum variance. However, it remains to be seen how this new method applied to the different tectonic domains will contribute to the current debate on weak and strong mantle lithospheres.

The strong versus weak lithospheric mantle debate:

As the significance of the EET in the geodynamic evolution of tectonic plates became evident, sophisticated techniques for measuring the EET from geophysical observables emerged. These techniques applied to the continental lithosphere often gave varying results. Nearly all pioneering studies in this regime recovered the low EET values in the continents ranging from 0 (Airy approximation) to tens of kilometers. However, the inception
of an inverse spectral method to estimate the EET using the Bouguer coherence and load
deconvolution technique by Forsyth\textsuperscript{23} resulted in obtaining very high EET values (100 km and
more) in certain areas where the low EET values were reported earlier. McKenzie and
Fairhead\textsuperscript{24} interpreted these findings as a result of the continental mantle lithosphere being
more mechanically robust than previously understood. The paper\textsuperscript{24} initiated a debate about the
properties of the lithosphere, popularly dubbed in the literature as jelly sandwich versus crème
Brulee\textsuperscript{15,42}. The prominent three areas of this ongoing debate are the effective elastic thickness
of the lithosphere, maximum earthquake depths within the lithosphere, and the rheological
properties of the lithosphere\textsuperscript{13,15,42,58–62}.

The wide range of observed EET values can be due to the considerable variance in
composition, geothermal gradient, and crustal thickness of the continental lithosphere\textsuperscript{3}.
Through theoretical modeling, Burov and Diament\textsuperscript{11} demonstrated that a model in which a
weak lower crust sandwiched between a solid brittle-elastic upper crust and an elastic ductile
mantle (Fig. 3) could account for the observed EET values over continents. Later on, the models
with a strong crust and lithospheric mantle came to be known as ‘Jelly Sandwich’ models of
lithospheric architecture. Maggi et al.\textsuperscript{58} studied the depth distribution of earthquakes inside the
continental lithosphere and proposed that, unlike the oceanic mantle lithosphere, the
continental mantle lithosphere is almost aseismic. Based on the observations of Maggi et al.\textsuperscript{58}
and refuting the ideas of McKenzie and Fairhead\textsuperscript{24}; Jackson\textsuperscript{59} proposed a model for the
lithospheric strength in which the crust is strong and the mantle is weak (Fig. 3). Jackson’s
model\textsuperscript{59} was later dubbed in literature as the Crème Brulee model of lithospheric architecture.
Since then, the deviations of the measured gravity signal from the predictions of the flexure
models have been crucial in this debate concerning the appropriate rheological model for the
strength of the lithosphere. Later studies have shown that a ‘Jelly Sandwich’ type rheology is
more stable in a collisional tectonic setting than a ‘Crème Brulee’ type rheology\textsuperscript{15}. That being
the case, several sophisticated techniques were developed for measuring the EET from the observed gravity and topographic data. It is an often-contested case which of these methods gives the best estimate of the EET in a given geodynamic/tectonic regime.

**The Himalaya – Tibet orogen, the holy grail of isostatic studies:**

The Himalaya – Tibet orogen has a special place among isostatic studies. The concept of isostatic equilibrium emerged during the great triangulation survey of India from the discovery in the 1850s that measurement errors due to changes in the vertical deflection caused by the mass of the Himalaya mountains were in effect only about one–third of what was expected4,6.

The Himalaya – Tibet orogen (Fig. 5a) is the most active continent-continent collisional belt on the Earth, which formed due to the collision between the Indian and Eurasian plates over the past 60 – 50 Ma63,64. Several models are proposed to address the sustenance mechanism of this orogen. These include the thrusting of the Indian plate under Eurasia65,66, southward subduction of the Asian lithosphere under Tibet67, delamination of the thickened lithospheric mantle68,69, and the flow of the lower crust70,71. Such models are numerous and often represent competing schools of thought. However, from a broader viewpoint, we can classify the available models for the evolution and sustenance of the Himalaya – Tibet orogen into four. These are subduction, underthrusting, diffuse thickening, and channel flow. However, the EET of the region can impose significant constraints on the subduction and underthrusting models of Himalaya evolution and sustenance (Fig. 4). Initial studies of the region based on surface geology indicated that the Indus Tsangpo Suture is the possible boundary between the Eurasian and Indian continental masses72. Then geophysical measurements helped constrain the subsurface structures beneath the mountain ranges, and several models describing the subsurface structure beneath the orogen were proposed. Based
on observations from gravity anomalies, Jin et al.,\textsuperscript{73} modeled the Indian lithosphere as plunging beneath the Eurasian lithosphere (Fig. 4a). Their models also showed a northward weakening of the strength of the Indian lithosphere (decrease in the EET values). Later, several variations of the subduction model were proposed based on different geophysical investigations\textsuperscript{74–76}, including opposite-facing subduction of both Indian and Eurasian lithosphere and southward subduction of the Eurasian lithosphere. The origin of the underthrusting models dates back to Argand’s hypothesis\textsuperscript{65}. These models assume that the descending Indian lithosphere returns to a horizontal position immediately beneath the crust of the overlying Eurasian plate (Fig. 4b).

The variations of this class of models involve differences in the extent of the horizontal advance of the underthrusting lithosphere and whether the underthrusting lithosphere stops horizontally or descends at some point. However, all the underthrusting models show the Indian lithosphere underthrusting beneath the Eurasian lithosphere and never the contrary\textsuperscript{64,77–79}.

These models have severe implications for the mechanical strength of the lithosphere also. A cold, stiff underthrusting Indian Plate beneath Tibet can create a robust lithospheric mantle there, which could support the lithospheric loads\textsuperscript{80,81}. However, if the lithospheric mantle and crust are decoupled, the continent-continent plate boundary where the Indian plate collides with the Eurasian plate can develop a subduction-type mechanism. This is because the lithospheric mantle can descend to significant depths beneath the Tibetan Plateau in such a scenario. This led to a weakening of the Indian lithosphere and reduced EET values towards the north\textsuperscript{73}. The Channel flow class of models suggests a decoupling of the upper crust from the lithospheric mantle due to ductile flow near the base of the crust. This implies that much of the lithosphere’s flexural strength is concentrated in the upper to the middle crust\textsuperscript{82}. Thus, measuring the effective elastic thickness of the region could play a defining role in understanding the mode, localization, and sustenance of lithospheric structure in and around Tibet. Since the inception of flexural modeling to assess the EET values over the continental
lithosphere, there have been many studies over the Himalaya – Tibet orogen. However, due to the differences in data and modeling tools, the estimated EET values were different even for the same regions\textsuperscript{22,73,80,83–85}. Over the past few decades, different studies used different methods to estimate the EET distribution over various parts of the Indian continental lithosphere. Isostatic studies over the Indian shield\textsuperscript{22,86,87} generally use the intermediate to long-wavelength gravity anomalies produced by the crustal thickness variations due to flexural bending. The initial attempts to determine the EET in cratonic regions of the Indian lithosphere were made by Karner and Watts\textsuperscript{83} and Lyon-Caen and Molnar\textsuperscript{81}. Using the forward modeling approach, they obtained the EET values of 80 km to 110 km in the Ganges basin. The studies by McKenzie and Fairhead\textsuperscript{24} using the free air admittance method yielded a lower EET value of 24 km, which readily correlates with the seismogenic thickness from the region. However, the subsequent studies by Handy and Brun\textsuperscript{60} pointed out that the EET could also exceed the seismogenic thickness. Rajesh et al.\textsuperscript{88} delineated the apparent variations in the EET in the India-Eurasia collision zones using Multitaper flexure analysis. In another study, Rajesh and Mishra\textsuperscript{89} characterized the tectonic provinces using the transitional coherence wavelength analysis. Jordan and Watts\textsuperscript{22} employed flexural and gravity modelling (both forward and inverse) techniques to analyse the India-Eurasia collision zones and obtain the spatially varying EET structures ranging from 0 to 125 km. The available estimates of the EET under the Indian subcontinent reveal a significant variability across different regions, as the EET depends on a region's rheology. Hence, significant variations in the EET indicate corresponding changes in the rheology, possibly due to a reworked crust. Yadav and Tiwari\textsuperscript{100} conducted numerical simulations of present-day tectonic stress across the Indian subcontinent. They found a significant correlation between the stress orientations within the Indian plate and the spatial variation of the EET.
The spatial variation of the EET computed using different techniques also shows significant departures from each other. For example, we compile the North-South variation of the EET computed by Chen et al.,\textsuperscript{91} and Hetenyi et al.,\textsuperscript{26} (Fig. 5b). They used the inverse spectral method of fan wavelet transform and thermo–mechanical modeling, respectively, to calculate the EET. Recently, Hetenyi et al.,\textsuperscript{92} investigated lateral variations in gravity and topography anomalies along the Himalayan arc. They concluded from their computed Arc-parallel Gravity Anomaly (APGA) variations that the orogen’s deep structure has clear lateral boundaries. They identified four disparate flexural geometry segments: NE India, Bhutan, Nepal, and NW India (Fig. 5c). The APGA pattern also suggested that while Nepal and NE India begin to flex farther south of the topographic front and disappear beneath the Himalaya at a relatively lower angle, NW India and Bhutan begin to flex closer to the topographic front and dip at a steeper angle.

These studies suggest possible variations in the effective elastic thickness (EET) along the strike of the Himalaya-Tibet orogen. However, except for one study\textsuperscript{25}, all the measured EET as the integrated strength of the lithosphere or the upper crust alone where the crust and the lithospheric mantle are decoupled. Cattin et al.\textsuperscript{25} measured the variation of the EET from south to north separately for the upper crust, lower crust, lithospheric mantle, and lithosphere as a whole and obtained values of 30 to 60 km of effective elastic thickness for the Indian lithosphere (Fig. 5c). Tiwari et al.,\textsuperscript{93} measured the effective elastic thickness in the Sikkim Himalaya region. They obtained an EET value of 50 km for the area. However, to determine whether the Himalayan orogen is sustained by a Jelly sandwich or a Crème Brûlée model for lithospheric architecture, we shall first determine the amount of topographic stress carried by the crust and lithospheric mantle parts of the lithosphere. Even though Burov and Diament\textsuperscript{11} had introduced a method of thermomechanical modeling to estimate the EET for several lithospheric layers with constant Young’s Modulus, there have been very few attempts to measure the amount of topographic load born by the crust and lithospheric mantle individually.
The vertical heterogeneity of the lithosphere, especially in regions like the Himalaya-Tibet orogen, was a significant drawback of this method that limited its application. Recently, Tesauro et al.\textsuperscript{21}, derived new equations to calculate the EET from the lithospheric strength distribution, considering different Young’s modulus for each lithospheric layer. When the plate consists of \( n \) decoupled layers:

\[
Te = (\sum \Delta h_i^3)^{1/3}
\]

Where, \( \Delta h_i \) is the elastic thickness of the \( i \)th layer. Tesauro et al.\textsuperscript{21} also individually calculated the amount of topographic load supported by the crust and lithospheric mantle. However, Bellas and Zhong\textsuperscript{94} suggested that in settings where the competent upper crust is much thinner than the mantle lithosphere (\( Te1 \ll Te2 \)), \( Te = \Delta h_1 \) is a more accurate measurement of the regional EET.

Since the rheological modeling heavily depends on several assumed and laboratory-derived parameters, the uncertainty in the estimated EET is high. Therefore, there is a need for a method to calculate the individual contribution of the crustal and lithospheric mantle layers towards supporting the topographic load and to measure the effective elastic thickness (EET) of the lithosphere from the in situ geophysical observables.

‘Joint modeling’ as a way forward:

Since the EET is a proxy for the lithosphere’s integrated strength, comparing it to other physical parameters representing its thermal and mechanical structure is also relevant. The seismic (surface and body wave) velocities reflect temperature and rheologic composition. It may be advantageous to incorporate the relationship between the EET and these attributes into the usual gravity and topography calculations\textsuperscript{95}. Moreover, joint modeling or inversion of multiple physical properties is a well-recommended method to overcome the inherent non-uniqueness of the geophysical interpretations. Thus, a combined analysis of the gravity and
topographic data, which forms the fundamental observables of the isostatic compensation, along with seismic velocity structures predicted by models with different EET, can shed some light on this aging debate⁹⁶,⁹⁷.

The main consequence of this ongoing debate had been to cause most geologists and geophysicists to hold under suspicion all estimates of the EET from continental regions. However, since the effective elastic thickness of the lithosphere probably controls much of the tectonics and geodynamics of continents, we would be able to make much greater use of the widely available measurements of gravity and topography data once this debate can be settled. This can be a step in the right direction toward the integration of several geophysical measurements to better constrain the continental elastic thickness estimates.

Acknowledgments

This work is a part of Ph.D. thesis of VHA. We are very grateful to the editor and two reviewers for their exhaustive reviews and critical comments. VHA thanks Junior Research Fellowship from University Grants Commission and the project fellowship from CSIR project MLP-FBR-003(AM). The CSIR-NGRI reference number of the manuscript is NGRI/LIB/2023/Pub-???.

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**Figure Legends**

**Figure 1:** Schematic diagram illustrating the different models of isostasy. Pratt model interprets high topography as a result of density changes within the crust itself. Airy model interprets the high topography as a result of increased crustal thickness. According to Flexure model, the lithosphere acts like an elastic beam that spreads the load of the topographic load due to its inherent elastic strength.

**Figure 2:** Histograms showing the distribution of the EET calculated using inverse spectral methods and through rheological modeling in (a) Phanerozoic orogens and (b) Precambrian shields and platforms.

**Figure 3:** Schematic diagram illustrating the two competing models for the long-term strength of the continental lithosphere. In the “jelly sandwich” model, the brittle upper crust and the lithospheric mantle are strong and the compensation for the surface loads occurs in the underlying asthenosphere. In the “crème Brûlée” model, the strength is confined to the brittle crust and the compensation for the topographic load is achieved in the lithospheric mantle itself.

**Figure 4:** Schematic representation of the four main models of the Himalaya-Tibet orogeny. (a) The subduction model in which the Indian lithospheric mantle continues to subduct, but the crust is injected beneath the lower crust of the Eurasian plate (b) The underthrusting of the Indian lithosphere beneath the Eurasian lithosphere. These models imply different types of lithospheric strength (EET) distributions. Please see the text for more discussion.

**Figure 5:** (a) Topography map of the Himalayan – Tibetan orogen showing locations of the region and profiles used to study the variation in the EET in the Himalayan – Tibetan orogeny. The orange dashed lines AA’, BB’ and CC’ represent the arc-normal profiles studied by Hetenyi et al., Chen et al., and Tiwari et al. respectively. The yellow box represents the arc-parallel study region of Hetenyi et al. The major faults and sutures in the region (such as
ATD - Altyn Tagh Fault; BNS - Bangong-Nujiang Suture; ITS - Indus-Tsangpo Suture; JLF - Jiali Fault; JRS - Jinsha River Suture; MFT - Main Frontal Thrust) are shown in (a) with black solid lines. (b) The north south variation of the effective elastic thickness of the Indian lithosphere according to inverse spectral estimation by Chen et al. and thermo-mechanical modeling by Hetenyi et al. (c) The north south variation of the effective elastic thickness of the upper crust, lower crust, lithospheric mantle and the compact lithosphere according to thermo-mechanical estimate by Cattin et al. (d) Arc-Parallel Gravity Anomaly (APGA) on either side of the topographic front (MFT) from Hetenyi et al. Approximate regional boundaries are indicated as references. Kindly see the text for the details of the EET estimates from the region as shown in this figure.
Figure 1: Schematic diagram illustrating the different models of isostasy. Pratt model interprets high topography as a result of density changes within the crust itself. Airy model interprets the high topography as a result of increased crustal thickness. According to Flexure model, the lithosphere acts like an elastic beam that spreads the load of the topographic load due to its inherent elastic strength.
Figure 2: Histograms showing the distribution of the EET calculated using inverse spectral methods and through rheological modeling in (a) Phanerozoic orogens and (b) Precambrian shields and platforms\textsuperscript{51}.
Figure 3: Schematic diagram illustrating the two competing models for the long-term strength of the continental lithosphere. In the “jelly sandwich” model, the brittle upper crust and the lithospheric mantle are strong and the compensation for the surface loads occurs in the underlying asthenosphere. In the “crème Brulee” model, the strength is confined to the brittle crust and the compensation for the topographic load is achieved in the lithospheric mantle itself.
Figure 4: Schematic representation of the four main models of the Himalaya-Tibet orogeny. (a) The subduction model in which the Indian lithospheric mantle continues to subduct, but the crust is injected beneath the lower crust of the Eurasian plate\textsuperscript{73,79} (b) The underthrusting model of the Indian lithosphere beneath the Eurasian lithosphere\textsuperscript{64,65,77}. These models imply different types of lithospheric strength (EET) distributions. Please see the text for more discussion.
Figure 5: (a) Topography map of the Himalayan – Tibetan orogen showing locations of the region and profiles used to study the variation in the EET in the Himalayan – Tibetan orogeny.
The orange dashed lines AA’, BB’ and CC’ represent the arc-normal profiles studied by Hetenyi et al.\textsuperscript{26}, Chen et al.\textsuperscript{91}, and Tiwari et al.\textsuperscript{93} respectively. The yellow box represents the arc-parallel study region of Hetenyi et al.\textsuperscript{92}. The major faults and sutures in the region (such as ATD - Altyn Tagh Fault; BNS - Bangong-Nujiang Suture; ITS - Indus-Tsangpo Suture; JLF - Jiali Fault; JRS - Jinsha River Suture; MFT - Main Frontal Thrust) are shown in (a) with black solid lines. (b) The north south variation of the effective elastic thickness of the Indian lithosphere according to inverse spectral estimation by Chen et al.\textsuperscript{91} and thermo-mechanical modeling by Hetenyi et al.\textsuperscript{26}. (c) The north south variation of the effective elastic thickness of the upper crust, lower crust, lithospheric mantle and the compact lithosphere according to thermo-mechanical estimate by Cattin et al.\textsuperscript{25} (d) Arc-Parallel Gravity Anomaly (APGA) on either side of the topographic front (MFT) from Hetenyi et al.\textsuperscript{92}. Approximate regional boundaries are indicated as references. Kindly see the text for the details of the EET estimates from the region as shown in this figure.