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

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The effective elastic thickness (EET) of the lithosphere is a measure of the lithosphere's ability to flex under long-term geological and topographic loads. It is often estimated through analyses of gravity and topographic data. The EET has a significant role in regulating the geodynamic evolution of both the continental and oceanic plates. Estimates of EET derived from geophysical data are consistent with rheological models in the oceanic regions. However, there are extensive debates on the estimates of 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 inconsistent results obtained using different datasets, e.g. the distribution of earthquakes, EET, gravity anomaly and rheology. This article discusses the evolution of these contrasting models and the critical necessity to resolve the model differences.

Keywords: Continental rheology, effective elastic thickness, flexural modelling, isostasy, lithosphere.

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, deposition of sediments in the basins and the erosion of plateaus, despite the large vertical strains they generate[1]. 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[2]. Therefore, understanding the properties of the lithosphere becomes vital in deciphering the pattern of 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[3].

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[4–7]. 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 the overlaying material remains constant throughout the Earth at a certain depth known as the depth of compensation (Figure 1). According to the Airy model, variations in the thickness of a uniform-density crust compensate for the excess loads[7]. In contrast, in the Pratt model, the necessary support is provided by the lateral variations in density of the crust or subcrustal mantle[5].

Both models predict gravity anomaly as the key observable sensitive to changes in mass distribution due to topographic loading and its associated compensation[3]. 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[3]. However, there are locations on the Earth (e.g. the Hawaiian Islands and Himalaya–Tibet region) where the Airy and Pratt models fail to account for the observed gravity anomalies adequately. This is because both models are highly idealized, so they do not account for the inherent rigidity of the lithosphere and only consider the state that the crust and subcrustal mantle would reach given a sufficiently long time.

Vening-Meinesz and Andries[8] used the formulation by Hertz[9] 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

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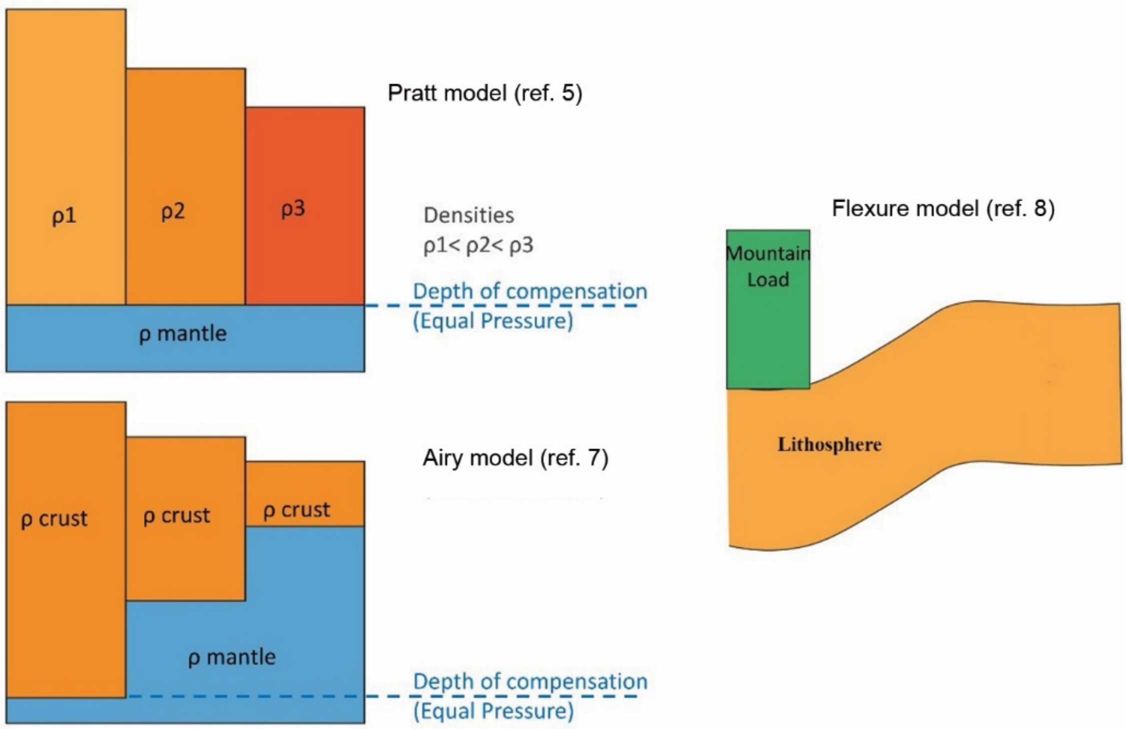

Figure 1. Schematic diagram illustrating the different models of isostasy. The Pratt model interprets high topography as a result of density changes within the crust. The airy model interprets high topography as a result of increased crustal thickness. According to the Flexure model, the lithosphere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

anomalies[8]. According to the Hertz model[9], 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[10]:

$$D = \frac{E \mathrm{T_e}^3}{12(1-\nu^2)},$$

where E is the Young's modulus, ν the Poisson's ratio and $\mathrm{T_e}$ is the effective elastic thickness (EET). D measures the deformability or stiffness of the plate. 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 EET, which is the thickness of the lithosphere with pure elastic rheology. EET generally does not represent a depth to any boundary within the lithosphere. It is only a mathematical analogue of the integrated strength of the lithosphere[11–13]. However, EET provides an adequate measure of the flexural rigidity of the lithosphere and thus possesses a true geological significance. Many subsequent studies concluded that the EET of the lithosphere controls most of the tectonics of the continents[14–16].

Knowledge of EET is crucial in determining how the lithosphere may respond to surface and subsurface loads. Furthermore, EET variations in the lithosphere could help explain some of the observed differences between the actual flattening of the Earth and that predicted by the hydrostatic theory[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[18,19]. EET can also scale the viscosity in thin, viscous sheet models that attempt to calculate stress and intraplate deformation due to plate boundary forces[20]. The lithospheric flexure also interacts with the atmosphere and asthenosphere, profoundly impacting landscape evolution and deep processes such as mantle convection[3]. However, significant regional variances exist in the currently estimated EET values worldwide. Low values of EET are usually associated with mid-ocean ridges and some continental rifts, while 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[3,21]. On the other hand, the oceanic lithosphere is thinner and more homogeneous than the continental lithosphere. Since it cools as it moves away from the mid-ocean ridge, the oceanic lithosphere is expected to gain elastic strength with age[22]. In conjunction with other geophysical parameters, the spatial variations in EET have been used to infer the detailed tectonic evolution of morphological features on the ocean floor[23]. Studies from subduction zones have shown that the EET of subducting plates is highest on the seaward side of the outer-rise region and decreases sharply near the trench-ward side[24–26].

Conventional methods of 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 EET is zero (i.e. the plate has no rigidity), the load is supported in hydrostatic equilibrium, according to models such as Pratt–Hayford and Airy–Heiskanen[10]. 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 EET.

The three most common methods of computing EET are: (i) forward modelling of gravity and topography in the spatial domain[27], (ii) inversion of the spectral properties of the gravity field and topography[28,29] and (iii) thermo-rheological modelling[30,31]. In the first method, a comparison of theoretical and observed gravity from flexural modelling is made in the space domain. In the second method, using Fourier or wavelet transforms, this comparison is made in the spatial frequency or wavenumber domain. The thermo-rheological modelling 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[32,33]. The Brace–Goetze failure envelope curves suggest that elastic strength increases with depth and then decreases according to the brittle and ductile deformation laws[32,33]. In the continental regions, the yield strength envelope is complex, and there may be more than one brittle and ductile layer depending upon the temperature and pressure profile of the region. Using data from experimental rock mechanics together with the temperature profile, it is possible to construct yield strength envelope profiles from which we can define the associated EET values. Suppose the estimated EET exceeds the crustal thickness or the brittle–ductile transition (BDT). This indicates that the elastic–ductile mantle also contributes to the strength of the lithosphere. However, this method significantly depends on assumed or laboratory-determined parameters such as composition, pore-fluid factor, elastic constants, strain rates, etc.

The spectral method of EET computation depends on derivatives of transfer functions of gravity and topography data in the spectral domain. The spectral approaches make it easy to determine the spatial and temporal variation of EET. The early spectral studies used the Bouguer admittance, Fourier transform periodogram method, and resulted in low EET values[34–37]. Later studies suggested these low values could result from subsurface loads uncorrelated with surface topographic loads[28]. They proposed using Bouguer coherence instead of Bouguer admittance because it is less sensitive to the surface-to-subsurface loading ratio. Using this approach in North America resulted in low and high EET values in the cratonic zones[38]. Later studies used new techniques, such as maximum entropy estimators[12] and multitapers[29,39]. McKenzie and Fairhead[29] suggested using free-air admittance rather than 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 erosion effects and thus overestimates the EET values. Pérez-Gussinyé and Watts[40] used Bouguer coherence and free-air admittance methods and found high values of 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[41,42], Australia[39] and Africa[43].

However, decades of flexural studies still need to arrive at a final consensus regarding these different models on the strength of the continental lithosphere. In the forward modelling approach, the gravity anomaly due to surface (topographic) load and flexural compensation associated with it are computed for varying values of EET and compared with the observed gravity anomaly[44,45]. The best approximation of 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 EET[46]. While forward modelling is a satisfactory way to estimate 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, EET is determined by calculating the transfer function between them as a function of wavelength and comparing it with the model predictions. This procedure assumes periodic or reflecting boundary conditions, making them inapplicable near plate boundaries with significant topography or tectonic loading conditions[47]. Supposing the mechanical problem is formulated identically in terms of the surface and subsurface loads, internal boundary conditions and the area chosen for modelling, we can expect both approaches to yield the same results. However, this is scarcely the case. EET from inverse spectral methods mostly disagrees with those from the forward modelling approach[10,15,16,29,48–55]. It is also possible that the two methods may be giving estimates of EET from different geological timespans – either

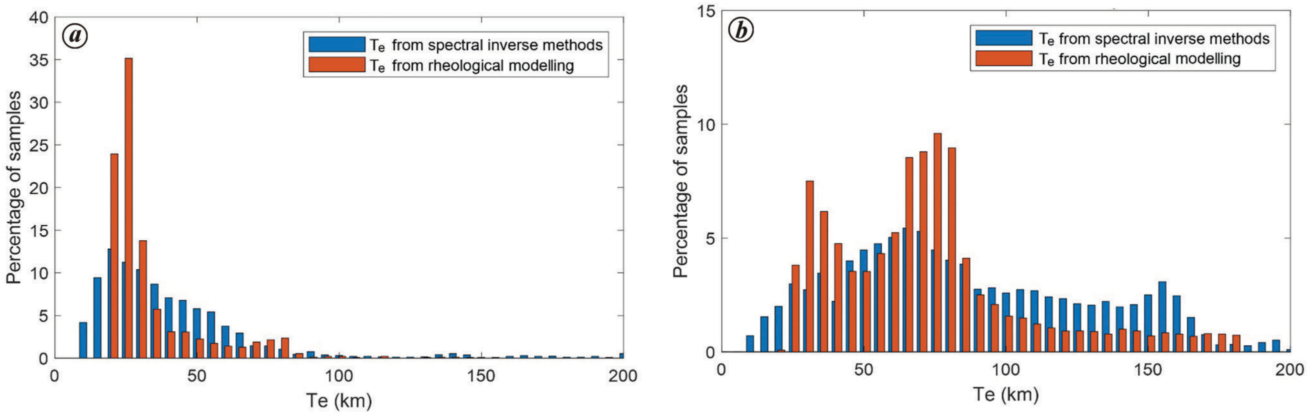

Figure 2. Histograms showing the distribution of the effective elastic thickness (EET) calculated using inverse spectral methods and through rheological modelling in (***a***) Phanerozoic orogens, and (***b***) Precambrian shields and platforms[56].

at the time of loading or the present day, depending on the geological age and tectonic setting (continent or ocean) of the region[3].

The thermo-rheological modelling approach also has been incapable of giving a final solution to the debate. Tesauro *et al.*[56] compared the global continental EET estimates from thermo-rheological parameters[21] with those from the Bouguer coherence deconvolution using wavelets of the spectral inversion method[57]. It was observed that the differences in EET values over Precambrian shields and platforms were high even though EET values in the Phanerozoic orogens were similar (Figure 2). The rheological EET was much less than the spectral EET in the Precambrian shields and platforms, with up to 50 km difference in certain regions. The difference in EET estimates may also be due to spectral and rheological calculation approaches (Figure 2). The relatively flat topography associated with some cratonic and basin areas increases the uncertainty of EET estimation using spectral inversion methods. The uncertainties of the rheological modelling depend on those of the thermal model used, and the uncertainties in thermal models increase considerably in cold Precambrian cratons[56]. Tesauro *et al.*[56] concede that since both methods can have relatively high uncertainties in the cratonic regions, estimating EET using them in such areas is challenging.

The inverse spectral technique is 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 gravity fields. In the early days, the periodogram estimation of admittance and coherence was the norm[57]. Nowadays, we have the maximum entropy, multitapers and wavelets at our disposal, using which we can estimate the admittance and coherence functions[39,58]. However, in a recent study, Simons and Olhede[59] have challenged the statistical validity of inverting ratio data to transform products and the estimation of 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 novel 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 EET in the geodynamic evolution of tectonic plates becomes evident, sophisticated techniques for measuring it from geophysical observables have emerged[28,29,60]. These techniques applied to the continental lithosphere often give varying results. Nearly all pioneering studies in this regime recovered low EET values in the continents ranging from 0 (Airy approximation) to tens of kilometres[60]. However, employment of the inverse spectral method to estimate EET using the Bouguer coherence and load deconvolution technique by Forsyth[28] resulted in obtaining very high EET values (100 km and more) in certain areas where low EET values were reported earlier. McKenzie and Fairhead[29] interpreted these findings as a result of the continental mantle lithosphere being more mechanically robust than previously understood. Their study[29] initiated a debate about the properties of the lithosphere, popularly dubbed in the literature as Jelly Sandwich versus Crème Brulee[15,47]. The three prominent areas of this ongoing debate are EET of the lithosphere, maximum earthquake depths within the lithosphere and rheological properties of the lithosphere[13,15,47,61–65].

The wide range of the observed EET values could be due to the considerable variance in composition, geothermal gradient and crustal thickness of the continental lithosphere[3]. Using theoretical modelling, Burov and Diament[11] demonstrated that a model in which a weak lower crust is sandwiched between a solid brittle–elastic upper crust and an elastic ductile mantle could account for the observed EET values over continents (Figure 3). Later, models with a strong crust and lithospheric mantle came to be known as 'Jelly Sandwich' models of lithospheric architecture. Maggi *et al.*[61] studied the depth distribution of earthquakes

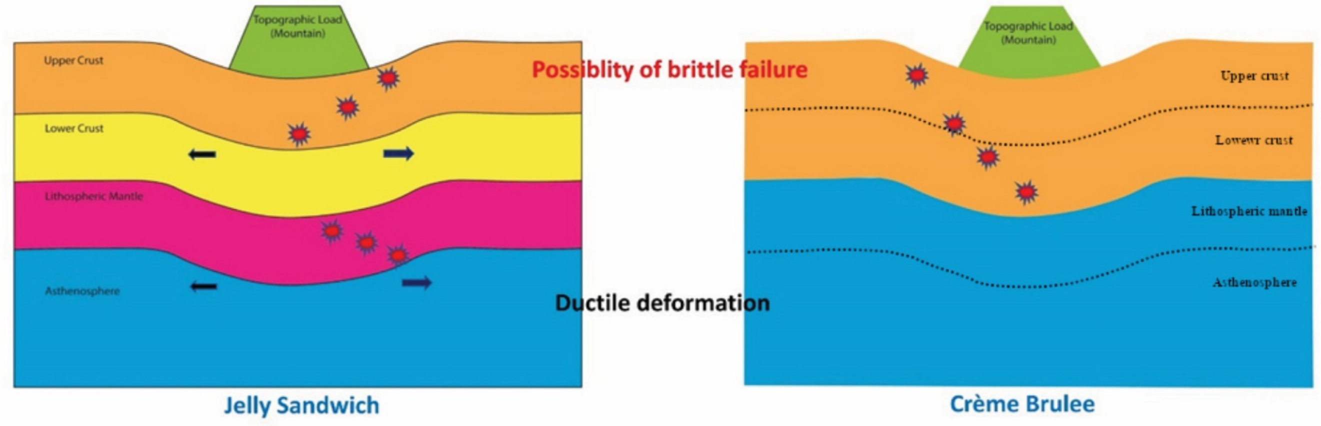

Figure 3. Schematic diagram illustrating the two competing models for long-term strength of the continental lithosphere. In the 'Jelly Sandwich' model[11], the brittle upper crust and the lithospheric mantle are strong and compensation for the surface loads occurs in the underlying asthenosphere. In the 'Crème Brulee' model[62], the strength is confined to the brittle crust and compensation for the topographic load is achieved in the lithospheric mantle.

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.*[61] and refuting the ideas of McKenzie and Fairhead[29], Jackson[62] proposed a model for lithospheric strength in which the crust is strong and the mantle is weak (Figure 3). Jackson's model was later dubbed in the literature as the Crème Brulee model of lithospheric architecture[62]. Since then, 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[15]. That being the case, several sophisticated techniques were developed for measuring EET from the observed gravity and topographic data. It is an often-contested case regarding which of these methods gives the best estimate of 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 Himalayan mountains were, in effect, only about one-third of what was expected[4,6].

The Himalaya–Tibet orogen is the most active continent–continent collisional belt on the Earth, which formed due to collision between the Indian and Eurasian plates over the past 60–50 Ma (refs 66, 67). Several models are proposed to address the sustenance mechanism of this orogen. They include the thrusting of the Indian Plate under Eurasia[68,69], southward subduction of the Asian lithosphere under Tibet[70], delamination of the thickened lithospheric mantle[71,72] and flow of the lower crust[73,74]. 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 groups. 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 the Himalaya evolution and sustenance (Figure 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 masses[75]. 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.*[76] modelled the Indian lithosphere as plunging beneath the Eurasian lithosphere (Figure 4 *a*). The model also showed a northward weakening of the strength of the Indian lithosphere (decrease in EET value). Later, several variations of the subduction model were proposed based on different geophysical studies[77–79], including opposite-facing subduction of both the Indian and Eurasian lithosphere and southward subduction of the Eurasian lithosphere. The origin of the underthrusting models dates back to Argand's[68] hypothesis. These models assume that the descending Indian lithosphere returns to a horizontal position immediately beneath the crust of the overlying Eurasian Plate (Figure 4 *b*). 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[67,80–82].

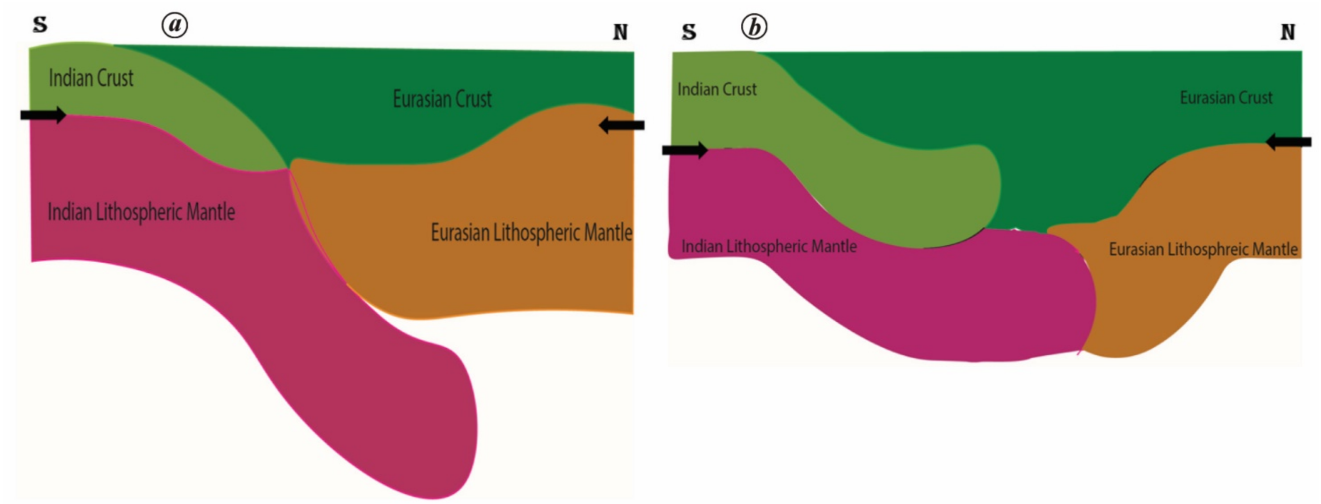

Figure 4. Schematic representation of the two main models of the Himalaya–Tibet orogeny. ***a***, The subduction model in which the Indian lithospheric mantle continues to subduct, but the crust is injected beneath the lower crust of the Eurasian plate[76,82]. ***b***, Underthrusting of the Indian lithosphere beneath the Eurasian lithosphere[67,68,80]. These models imply different types of lithospheric strength (EET) distributions. See text for more discussions.

These models also have significant implications for the mechanical strength of the lithosphere. A cold, stiff underthrusting Indian Plate beneath Tibet can create a robust lithospheric mantle, which could support the lithospheric loads[83,84]. 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, leading to a weakening of the Indian lithosphere and reduced EET values towards the north[76]. 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 flexural strength of the lithosphere is concentrated in the upper to middle crust[85]. Thus, measuring the EET of the region could play a defining role in understanding the mode, localization and sustenance of the lithospheric structure in and around Tibet. Since the inception of flexural modelling to assess EET values over the continental lithosphere, there have been many studies over the Himalaya–Tibet orogen. However, due to the differences in data and modelling tools, the estimated EET values were different even for the same regions[27,76,83,86–88].

Over the past few decades, numerous studies have used different methods to estimate EET distribution over various parts of the Indian continental lithosphere. Isostatic studies over the Indian shield generally use the intermediate to long-wavelength gravity anomalies produced by crustal thickness variations due to flexural bending[27,89,90]. The initial attempts to determine EET in cratonic regions of the Indian lithosphere were made by Karner and Watts[86], and Lyon-Caen and Molnar[84]. Using the forward modelling approach, they obtained the EET values of 80–110 km in the Ganges basin. The study by McKenzie and Fairhead[29] 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, a subsequent study by Handy and Brun[63] pointed out that EET could also exceed the seismogenic thickness. Rajesh *et al.*[91] delineated the apparent variations in EET in the India–Eurasia collision zones using multitaper flexure analysis. Rajesh and Mishra[92] characterized the tectonic provinces using transitional coherence wavelength analysis. Jordan and Watts[27] 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 EET under the Indian subcontinent reveal a significant variability across different regions, as EET depends on a region's rheology. Hence, significant variations in EET indicate corresponding changes in the rheology, possibly due to a reworked crust. Yadav and Tiwari[93] performed numerical simulations of present-day tectonic stress across the Indian subcontinent. They found a significant correlation between stress orientations within the Indian Plate and the spatial variation of EET.

The spatial variations of EET determined using different techniques also showed significant departure from each other in the Himalaya–Tibet orogen (Figure 5). For example, we compiled the north–south variation of EET computed by Chen *et al.*[94] and Hetényi *et al.*[31] (Figure 5 *b*). They have used the inverse spectral method of fan-wavelet transform and thermo-mechanical modelling respectively, to determine EET. Recently, Hetényi *et al.*[95] studied lateral variations in gravity and topography anomalies along the Himalayan arc. They concluded from their computed arc-parallel gravity anomaly (APGA) variations that the deep

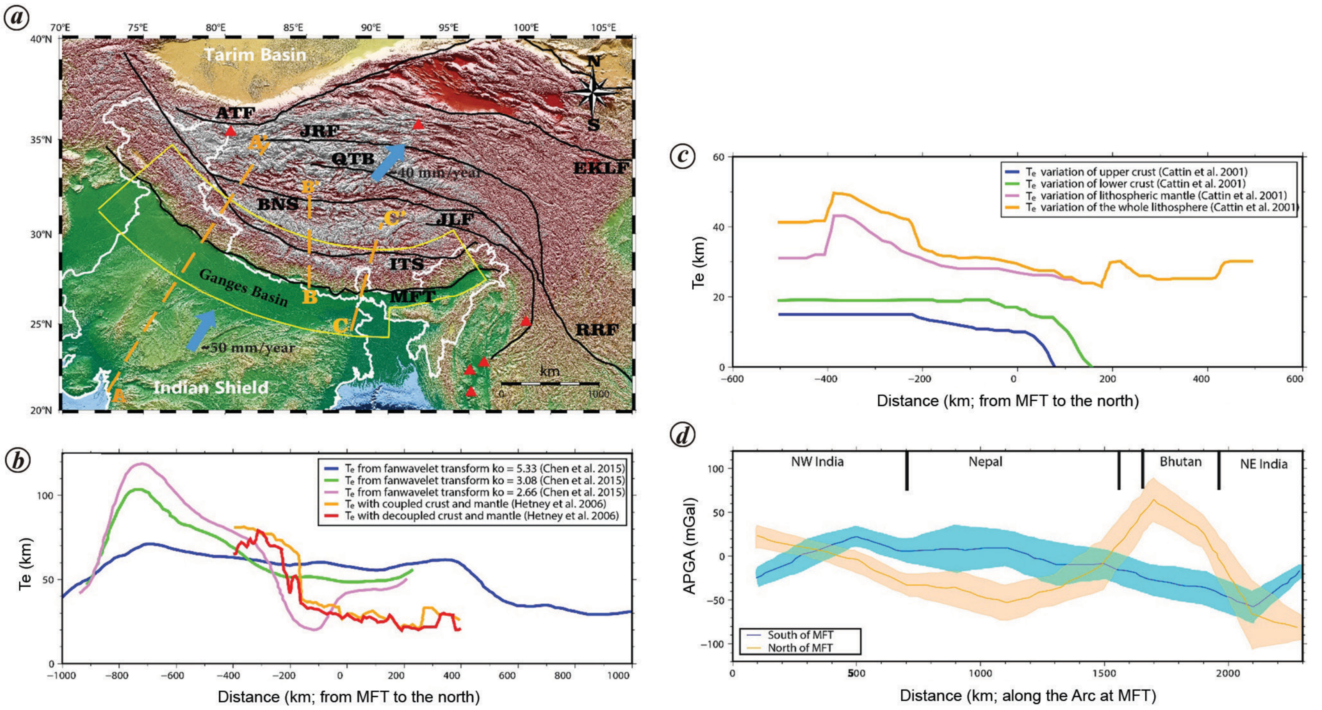

Figure 5. ***a***, Topography map of the Himalayan–Tibetan orogen showing locations of the region and profiles used to study the variation of EET in the orogeny. The orange dashed lines *AA′*, *BB′* and *CC′* represent the arc-normal profiles studied by Hetényi *et al.*[31], Chen *et al.*[94] and Tiwari *et al.*[96] respectively. The yellow box represents the arc-parallel study region of Hetényi *et al.*[95]. The major faults and sutures in the region (such as ATF, Altyn Tagh Fault; BNS, Bangong–Nujiang Suture; ITS, Indus–Tsangpo Suture; JLF, Jiali Fault; JRS, Jinsha River Suture and MFT, Main Frontal Thrust) are shown in (***a***) with black solid lines. ***b***, North–south variation of EET of the Indian lithosphere according to inverse spectral estimation by Chen *et al.*[94] and thermo-mechanical modelling by Hetényi *et al.*[31]. ***c***, North–south variation of EET of the upper crust, lower crust, lithospheric mantle and compact lithosphere according to the thermo-mechanical estimate by Cattin *et al.*[30]. ***d***, Arc-parallel gravity anomaly (APGA) on either side of the topographic front (MFT) is from Hetényi *et al.*[95]. Approximate regional boundaries are indicated as references. See text for details of EET estimates from the region as shown in this figure.

structure of the orogen has clear lateral boundaries. They identified four disparate flexural geometry segments: North East India, Bhutan, Nepal and Northwest India (Figure 5 *d*). The APGA pattern also suggests 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 EET along the strike of the Himalaya–Tibet orogen. However, except for one study[30], all the measured EET as the integrated strength of either the lithosphere or the upper crust alone, where the crust and lithospheric mantle are decoupled. Cattin *et al.*[30] measured the variation of EET from south to north separately for the upper crust, lower crust, lithospheric mantle and lithosphere as a whole, and obtained values of 30–60 km for the Indian lithosphere (Figure 5 *c*). Tiwari *et al.*[96] measured EET in the Sikkim Himalaya region and obtained a value of 50 km for the area. However, to determine whether the Himalayan orogen is sustained by a Jelly Sandwich or a Crème Brulee model for lithospheric architecture, we must first determine the topographic stress carried by the crust and lithospheric mantle. Even though Burov and Diament[11] had introduced a method of thermo-mechanical modelling to estimate EET for several lithospheric layers with constant Young's modulus, there have been only a few attempts to measure the amount of topographic load supported by the crust and lithospheric mantle individually. The vertical heterogeneity of the lithosphere, especially in regions like the Himalaya–Tibet orogen, is a significant drawback of this method that has limited its applications.

Since the rheological modelling 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 contributions of the crustal and lithospheric mantle layers towards supporting the topographic load and to measure the EET of the lithosphere from the *in situ* geophysical observables.

Joint modelling as a way forward

Since EET is a proxy for the integrated strength of the lithosphere, 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 EET and these attributes into the usual gravity and topography calculations[97]. Moreover, joint modelling 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 form the fundamental observables of the isostatic compensation, along with seismic velocity structures predicted by models with different EET values, can shed light on this ageing debate[98,99].

The main consequence of this ongoing debate has been to make most geologists and geophysicists suspect all estimates of EET from continental regions. However, since EET of the lithosphere probably controls much of the tectonics and geodynamics of continents, we could make greater use of the widely available measurements of gravity and topography data once this debate is settled. The integration of several geophysical measurements to better constrain the continental elastic thickness estimates can be a step in the right direction in this regard.

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