Isostasy, equivalent elastic thickness, and inelastic rheology of continents and oceans

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ABSTRACT

The equivalent elastic thickness (EET) is used to estimate lithospheric strength expressed in response to loading by topography and subsurface loads. The data on EET allow comparisons between different plates and detection of thermal events. In oceans, the EET corresponds to the mechanical "core" of the lithosphere, i.e., a geotherm (400–600 °C). In continents, the EET has no relation to any depth. This has led to doubts in applicability of a unique approach to the continents and oceans, and in the utility of estimates of EET for continents. Rheological data suggest that most rocks are inelastic in the long term (>0.1 m.y.). This requires interpretation of the EET in terms of real rheology. We propose an analytical model that gives rheological interpretation of both the oceanic and continental EET. It also allows estimates of the mechanical thickness of the lithosphere. The EET depends upon three parameters: geotherm age, crustal thickness, and flexural plate curvature. Any one of these values can be estimated if the others are known. Comparisons of model predictions with the observed EET suggest that most continental plates have a weak lower crust, allowing mechanical decoupling between the upper crust and the mantle lithosphere. Such decoupling leads to strong reduction in the EET and thus can be easily detected. Flow of rocks in the weak lower crust may have a significant influence on the temporal evolution of relief (mountain building, erosion). Differences in the mechanical behavior of oceans and continents can be explained by domination of different parameters: geotherm age has a major control in the oceans, whereas in the continents crustal thickness is equally important. Additional local variations of the EET result from weakening by flexural stresses.

INTRODUCTION

The gravity effect of large-scale excess mass on Earth's surface is significantly reduced (compensated) by light "roots" at depth. This effect is called isostasy. The classic Pratt or Airy models of local isostasy assume that the outer part of the Earth is balanced hydrostatically and cannot support any deviatoric stress. However, geologic and geophysical observations point to high intraplate stresses. Therefore, if local isostasy holds for very large features (typically >1000 km) and/or for places where the lithosphere is weakened due to some specific conditions, for smaller features the rigidity and hence rheology of the lithosphere must be taken into account. Because the lithosphere consists of crust and mantle parts, the rheology of both must be considered.

On the geologic time scale, rheological properties of the lithosphere can only be obtained by studying long-term lithospheric response to the loading by topography and subsurface loads. These studies estimate the integrated strength exhibited by the lithosphere when it is flexed in response to loading or unloading. The effective elastic rigidity (D) or equivalent elastic thickness of the lithosphere (EET) are the units widely used to measure the lithospheric strength. The EET is defined through D as EET $= \sqrt[3]{12(1 - \nu^2)D/E}$, where $D = |MR|$, $M(x)$ is the flexural moment acting on the plate, $R(x)$ is the local radius of plate curvature, ν and E are the Poisson's ratio and Young's modulus of rock, respectively. The EET is the thickness of an imaginary elastic plate used to model the lithosphere. Estimations of the EET are made by fitting the geometry of a flexed theoretical elastic plate with the basement (oceans) and/or with the deflected Moho. The Airy model corresponds to a nonrigid plate (EET = 0). It is one end member of flexural modeling; the other (infinite EET) is the noncompensated case. The computations of EET use gravity and, sometimes, seismic data to delimit the geometry of the basement and Moho. Gravity studies might use forward modeling (e.g., Burov et al., 1994) or the interpretation of admittances or coherence functions (e.g., Poudjom et al., 1995).

From a rheological point of view, the oceanic lithosphere away from the ridges is more homogeneous than the continental one. The oceanic crust is thin and consists of basalts that are as strong as the mantle olivines. That is why it is considered as a single mechanical layer, the thickness of which is largely controlled by the geotherm. The oceanic geotherms are strongly age dependent: thus, a simple EET-age relation is true (Watts, 1978).

The continental crust is complex relative to its oceanic counterpart. It is six to ten times thicker, and its bottom (Moho) is much deeper (~40 km) and hence hotter (300–600 °C). Mechanical properties of the crustal rocks differ from the mantle olivines. Whereas the upper crust is cold and strong, the lower crust is hot and ductile, and it may easily flow at Moho temperatures. The mantle olivines can remain strong to twice the depth of the Moho. Therefore, the EET cannot be associated with a unique mechanical layer, and the continental EET does not correspond to any geotherm (McNutt et al., 1988; Burov and Diament, 1995). Indeed, continental EET estimates may be much less (by ~40%) than those inferred from geotherms, and large variations are observed for plates of the same thermotectonic age. Here we expose the significance of continental EET in terms of rheology, including possible crust-mantle decoupling (e.g., Meissner and Wever, 1988), and we discuss the geodynamic implications.

REALISTIC RHEOLOGY MODEL

The rheology data predict that at a constant strain rate ($\dot{\varepsilon}$) deformed rock may stay elastic only if the applied stress is below some strength called the yield stress $\sigma(y)$ (y is depth). At this stress it will undergo either brittle or ductile deformation. Ductile strength depends on $\dot{\varepsilon}$ and temperature $T(y)$ (power law); brittle strength depends (Byerlee's law) on pressure $P(y)$. Constitutive laws used to obtain yield-stress envelopes (YSE) (Goetze and Evans, 1979) are extrapolated from data of experimental rock mechanics (Fig. 1A). These laws are subject to large uncertainties because (1) the lithospheric conditions are only partly reproducible in the experiments and (2) the results are extrapolated from laboratory time and space scales to geologic scales.

Two unknown parameters are important to construct the YSE: the thickness of the strong upper crust, h_l, and the mechanical thickness of the lithosphere, h_m (Fig. 1). Uncertainties are too large (>100%, Fig. 1) to predict the YSE or h_l and h_m only from rheology laws and the geotherm. The slopes of the YSE are much less sensitive to variations in rheology laws than the predicted values of yield

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stress (Fig. 1A). Thus, if h_l and h_m are known a priori, the error in YSE may be reduced to 10%–20% (Fig. 1B). We can obtain h_l from various independent sources, such as seismological or mechanical thickness data. Estimation of h_m is much more difficult. We proposed a semianalytical approach based on a plate model with realistic YSE rheology (Burov and Diament, 1995). This approach allows us to derive h_m from the EET, or to predict the EET from the YSE. Here we consider implications for the continental and oceanic EET.

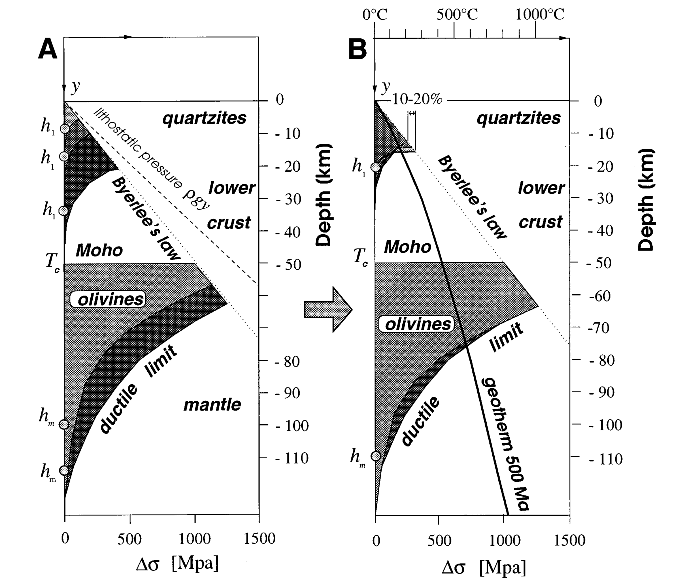

Figure 1. A: Typical continental yield-stress envelope $\Delta\sigma(y)$ (YSE) for thermal age 500 Ma and strain rate $\dot{\varepsilon} = 3 \times 10^{-15}$ s^{-1}. Quartz-controlled crust and olivine-controlled mantle. Crust: Light and medium gray areas correspond to quartz rheology (Kirby and Kronenberg, 1987), computed for "hot" (radiogenic heat) and "cold" (no radiogenic heat) geotherms, respectively. Dark gray area is for quartz rheology from Ranalli (1995). Mantle: Light area is for peridotite values with low activation energy (Meissner, 1995, personal commun.). Darker area is for common rheology (Kirby and Kronenberg, 1987). h_l and h_m are depths to bottoms of mechanical crust and mantle, respectively (depths at which yield strength is <1%–5% of lithostatic pressure ρgy). Note large differences between YSE proposed by different authors. B: If h_l and h_m are known a priori, uncertainties can be reduced to 10%–20%.

REDUCTION OF THE RIGIDITY BY STRESSES

Differential stress created by flexure of the lithosphere under surface topography and subsurface loads is not homogeneous. It is higher in places of larger curvature. The lithospheric strength is smaller close to the interfaces between the different lithological layers (upper, lower crust, mantle, Fig. 1), whereas the flexural strain should be maximum in those areas. There, flexural stress may easily exceed the yield strength, and the plate will be weakened due to relaxation of the stress either by brittle failure or by ductile flow (Fig. 2). Characteristic relaxation times of the stresses which are below the yield strength are very long, 10 to 100 m.y., depending on temperature and stress level. As a result, the EET of an inelastic plate is reduced in more "curved" areas and can be related to the radius of flexural curvature of the lithosphere (R). This is observed in oceanic domains, for example, beneath seamounts (McNutt, 1984). This also explains why some surface loads (e.g., mountain or large sedimentary deposits) in continents may appear "more locally" compensated (Cloetingh et al., 1982; Burov and Diament, 1995).

DECOUPLING AND FLOW OF MATERIAL IN THE LOWER CRUST

The mechanical response of sandwich-like continental plate, containing weak low-viscosity layers between more rigid ones, is very different from the behavior of a "single" layer oceanic plate (Fig. 2). First, the shear and horizontal stresses applied to one of the rigid cores (to the upper crust or upper mantle) will be only slightly transferred to another competent core(s). The approximate estimate for the EET (T_e) of a multilayered plate (McNutt et al., 1988; Burov and Diament, 1995) is:

$$T_e = \sqrt[3]{\sum_{i=1}^{n} \Delta h_i^3}, \tag{1}$$

where Δh_i is the thickness of the ith competent layer in the lithosphere (upper crust, middle crust, uppermost mantle, and so on). If the EET and the thickness of some layers are known from observations, they can be used to estimate h_m. For the typical two-layer case we have $T_e = \sqrt[3]{h_l^3 + h_2^3}$, with $h_2 = h_m - T_c$ (the thickness of the mechanical mantle). T_c is the total crustal thickness (see Fig. 1). Thus

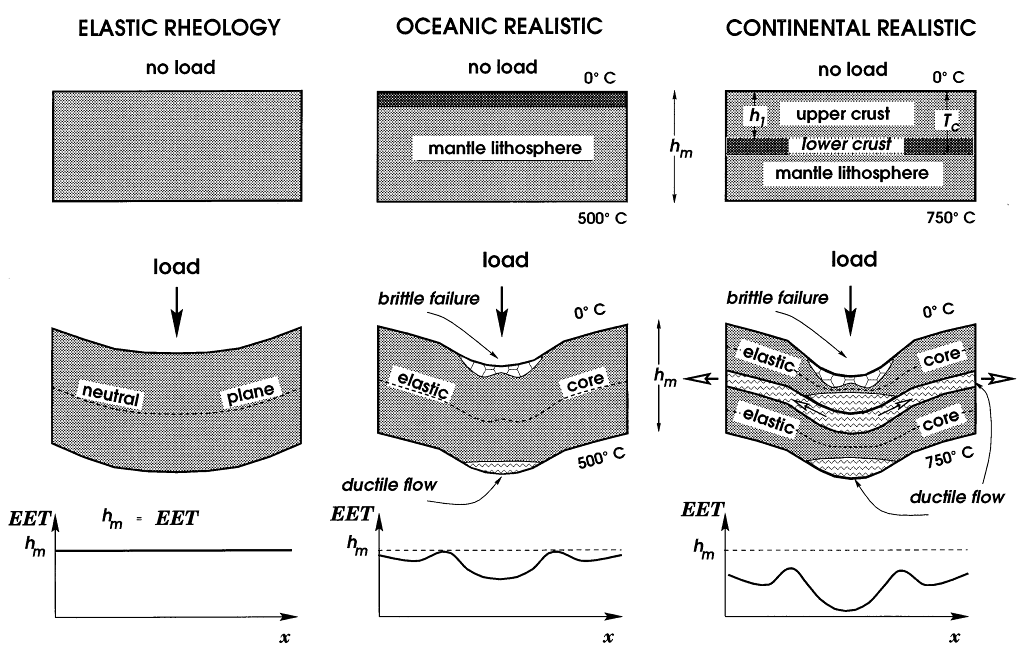

Figure 2. Difference between elastic and realistic rheology in oceanic and continental case (plates with same h_m). If no load is applied, plate is effectively elastic. If it is loaded and flexed, large areas of inelastic strain result in local weakening. Inelastic plate will be more flexed beneath load than elastic plate. In continents, weak lower crust eases mechanical decoupling between upper crust and mantle. This results in total weakening of plate (EET is less than for oceanic counterpart with same h_m). Flexure of continental plates may be associated with flow in lower crust. Down: horizontal variation of EET.

$$h_m = \sqrt[3]{(T_e^3 - h_l^3)} + T_c. \quad (2)$$

At small flexure h_l is usually 15–20 km (e.g., Cloetingh and Burov, 1996). In a thermally stable regime, the values of h_l and h_m are only slightly dependent on the crustal thickness T_c. However, h_2 is inversely related to T_c: the thicker the crust, the thinner the part of the strong mantle that falls in the depth interval between T_c and h_m (depth to 700–800 °C). Therefore, the thicker the crust, the weaker the lithosphere. This implies that in active orogenic belts, the lithosphere should become weaker with time due to thickening of the crust and to the above-mentioned effects of flexure. This is opposed to the effect of strengthening due to cooling.

WHAT CONTROLS EET?

For the case of two or three competent layers, the EET will be roughly equal to the thickness of the strongest layer, and not to the algebraic sum of the thicknesses ($h_l + h_2 + \ldots$). Thus, if the thickness of the upper crust (h_l) is equal to the thickness of the rigid mantle (h_2), the value of the EET will be around $(h_l + h_2)/2$. If the upper crust is thicker than the strong mantle (for thermally young lithosphere, <250 Ma), the EET will be largely controlled by the crustal strength (Kusznir and Karner, 1985) and will be equal to h_l. In turn, if the crust is weaker than the mantle (old and/or cold lithosphere, >400–700 Ma), the EET will be controlled only by the mantle strength and will be equal to h_2. Equation 1 is a simplified approximation originally derived in engineering for layered elastic materials. In the inelastic rheology, one needs to take into account the finite strength of the intermediate inelastic layers and the nonlinear dependence of $h_l, \ldots h_m$ on the local yield stress vertical gradient. Nevertheless, equation 1 is still a good approximation of the exact solution obtained using integration of the YSE (Fig. 3; Burov and Diament, 1995). Figure 3 (theoretical EET as a function of thermal age and Moho depth) reveals a strong EET dependence on the crustal thickness. If the crust is thick and thus sufficiently hot at the Moho depth, the upper crust and the uppermost mantle become mechanically decoupled, resulting in a large reduction of the EET. If the crust is thin and/or cold at Moho, no decoupling occurs and the lithosphere remains strong. This mode can be called "oceanic," because in this case the mechanical responses of the continental and oceanic lithospheres are similar. After ~500–700 m.y., the continental geotherm stabilizes, and the EET variations are related to variations of the crustal thickness, to local (e.g., flexural) and far-field tectonic stresses, and/or to large-scale thermal events. According to Figure 3, the averaged EET value for a >1000 Ma continental lithosphere with a 40–50-km-thick crust is around 65–75 km. This estimate corresponds to most of the estimates of EET obtained for old plates (e.g., Urals, Fennoscandia [Burov and Diament, 1995; Cloetingh and Burov, 1996]), except in some cratonic (very cold) areas.

There is a critical crustal thickness that defines whether the crust and mantle lithosphere can be coupled or decoupled (Fig. 3). Its asymptotic value for old lithosphere is 35 ± 5 km, that is, the average thickness of the continental crust (for the young lithosphere this critical thickness is much smaller, about 15–20 km, Fig. 3). This suggests the following conclusions. (1) Most of the continents should be in the decoupled mode (reduced EET and low-viscosity lower crust); 75% of observed EET is explained by decoupling (Burov and Diament, 1995). (2) It is possible that there is some kind of a temporal feedback between decoupling (lower-crustal flow) and crustal thickness.

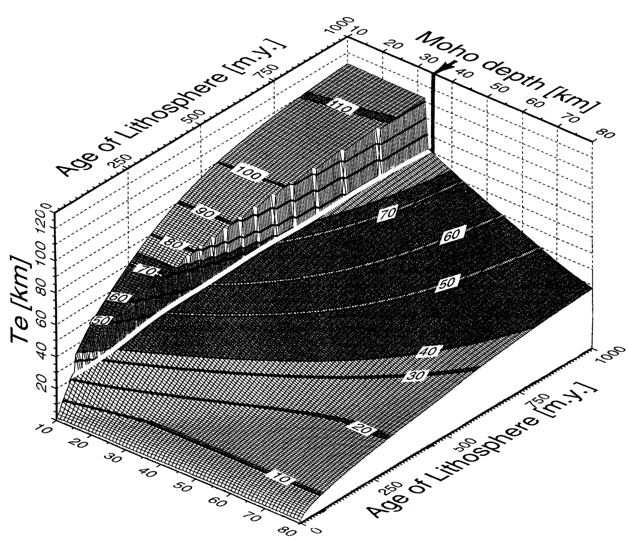

Figure 3. Theoretical equivalent elastic thickness (EET) (T_e)-thermal age-crustal thickness dependence computed for unloaded lithosphere with quartz-controlled upper crust and olivine-controlled mantle. Estimation of the mechanical thickness of the lithosphere (h_m) from EET: $h_m \approx \sqrt[3]{(T_e^3 - h_l^3)} + T_c \leq$ depth (600 °C ÷ 700 °C), where T_c is the crustal thickness, T_e is EET, h_l is thickness of the strong upper crust (~15 ÷ 20 km). Plot allows estimation of either effective thermal age from EET or EET from thermal age (geotherm). Crustal thickness is usually known. Note critical crustal thickness (white line) that defines decoupled (reduced EET, below the line) and coupled (above the line) regime.

GEODYNAMIC IMPLICATIONS

The low-viscosity channel near the Moho depth was inevitably formed when the lithosphere was young (i.e., weak and hot). After that, it could actively participate in crustal evolution, generally thickening. The viscosity of the lower crust, $\mu = \sigma/2\dot{\varepsilon}$, may be a few orders lower than that of the upper crust and of the strong mantle (Fig. 1; Bird, 1991). This suggests that the material may flow through the lower crust even under moderate pressure gradients associated with the slopes (α) of the surface relief (Bird, 1991; Burov and Avouac, 1994). Thickening of the crust due to shortening and flow increases the Moho temperature. Thus, when a crustal thickness of around 35 km is reached, the lower crust will be sufficiently hot (>400 °C) to remain weak. Flow in the lower crust can not only ease crustal thickening and mountain growth, but also prevent it when the topographic heights exceed certain values. This is because pressure gradients $\Delta p \approx \rho_c g \sin(\alpha\rho_c/\Delta\rho)$ created by topographic relief may result in extrusion of the low-viscosity material from the mountain root. The root will then be eroded and the crust thinned (Bird, 1991). This, however, will stop as soon as a balance with the compression is reached. Thus lower-crustal flow may regulate the thickness of the crust and topographic elevations. The EET estimates give a reliable control on the possibility of the lower-crustal flow, because the EET is two times smaller for a decoupled lithosphere than for a coupled one, whereas the usual error in EET estimation is about 20%.

It is generally assumed that compensation of the topography occurs at the Moho level, the response of the strong mantle lithosphere playing a leading role. This is true when changes of surface topography cannot be accommodated by flow in the lower crust. Because the rates of the tectonic uplift are typically 0.1–1 mm · yr^{-1} and thickness of the lower crust is between 10 and 35 km, the associated vertical strain rate should be 10^{-16}–10^{-15} s^{-1}. Continuity requires equal vertical and horizontal strain rates. Typical continental horizontal strain rates are about 10^{-15}–10^{-14} s^{-1}. Thus, in some cases, the mountain root can be dynamically sustained with a velocity comparable to the one of tectonic uplift, and some part of compensation will occur inside the crust. In this case, the mantle

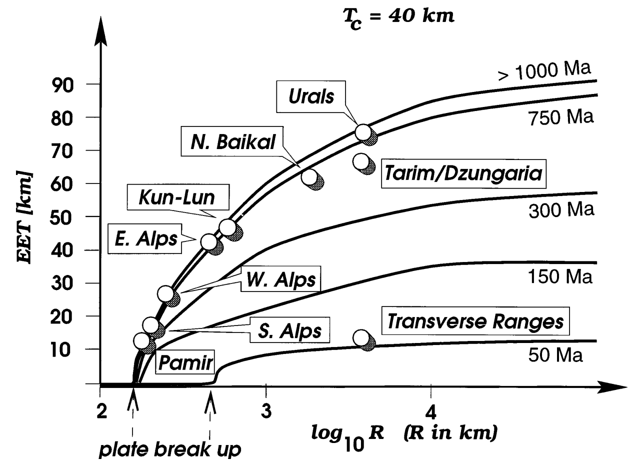

Figure 4. Dependence between radius of curvature (*R*) of lithosphere and equivalent elastic thickness (EET). Circles correspond to direct observations (after Burov and Diament, 1995).

lithosphere will be unloaded and the Moho may appear significantly flattened, resulting in overestimates of the EET.

The results shown in Figure 3 are obtained for moderate gradients of flexural stress (radius of plate curvature $R > 3000$ km). In collision zones, the lithosphere is much more flexed and R may drop to 250–500 km (McNutt et al., 1988; Ranalli, 1994; Burov and Diament, 1995). Associated flexural strains result in large stresses, σ_f (e.g., in the elastic "cores" $h_l, h_2, \ldots$ [see equation 1]). $\sigma_f \approx -7 \times 10^4$ MPa $\times\, y^*/R$, where y^* is the distance from a neutral plane of the core. At small R, σ_f may easily reach the local strength $\Delta\sigma(y^*)$. The elastic cores (and thus the EET) will be reduced. We computed theoretical EET-R dependence and compared it with the observations. The results (Fig. 4) show good agreement with theory, allowing to explain large variations of observed EET, e.g., in the Alps (Royden, 1993) or in the Tien Shan. Because the exact expression for EET-R dependence is too complicated (Burov and Diament, 1995), we replace it here with a simplified approximation:

$$T_e(R) = T_{emax} \times \left(\sqrt{1 - \sqrt{R_{min}/R}}\right)^{(1+\frac{1}{2}\, T_{emax}/T_{elim})}. \tag{3}$$

$T_{emax} = T_e(R_{max}, h_l, T_c, h_m)$ is the EET at small flexure (Fig. 3). It can be roughly estimated from equation 1 assuming maximum possible values for h_l and h_m (i.e., depths to ≈300 °C and ≈700 °C, respectively). $R_{min} \le R \le R_{max}$, where $R_{min} = 180$ km $\times$ [1 + 1.3 $\times (T_{emin}/T_{emax})^6$], $R_{max} \approx 10^4$ km, $T_{emin} = 15$ km, and $T_{elim} = 120$ km. If $R \le R_{min}$, then $T_e \le$ 5–10 km, which is almost below the sensitivity of EET estimates. At $R \approx R_{min}$, flexure leads to plate break ups and detachments of the crust from the mantle. Another consequence of EET dependence on the local tectonic stress (flexural or other) is that the EET in the sedimentary basins may even decrease with time due to increase of sedimentary load. This is contradictory to the conventional models (e.g., McKenzie, 1978) suggesting cooling and strengthening of the lithosphere in the post-rift period.

CONCLUSIONS

The EET is not a physical thickness, but it can be used to constrain the mechanical thickness of the lithosphere and to map its variations.

Strong deviations of the continental estimates of the EET from those predicted by thermal models can be explained by strength reduction due to crust-mantle decoupling and variations of crustal thickness. For the same total mechanical thickness, the EET is inversely dependent on the crustal thickness and local flexural and in-plane stresses. That is why plates of the same age may exhibit a large range of EET. The EET of a thermally young continental lithosphere is mainly controlled by the crustal strength, but the EET of an older lithosphere is roughly equal to the thickness of the mechanical mantle.

The coincidence of the asymptotic minimal crustal thickness (35 ± 5 km) required for decoupling with the average continental crustal thickness suggests the existence of a feedback between decoupling and crustal thickness. For example, flow through the lower crust may control the total crustal thickness to keep it sufficient for decoupling.

The flexural stresses caused by bending of the lithosphere beneath topographic load, due to slab pull and horizontal forces in the collision zones, can be responsible for local variations of the EET of the lithosphere. This explains, for example, EET variations in the southern, eastern, and northern Alps or in Central Asia, and why some features appear to be locally compensated in spite of being emplaced on a stiff plate.

By putting limits on the experimental rheological profiles and estimates of the mechanical thickness of the lithosphere and allowing us to see whether the crust is mechanically coupled with the mantle, EET estimates are a powerful tool for constraining the rheological properties of the lithosphere.

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