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Lithospheric strength and its relationship to the elastic and seismogenic layer thickness

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Abstract

Plate flexure is a phenomenon that describes how the lithosphere responds to long-term ($>10^5$ yr) geological loads. By comparing the flexure in the vicinity of ice, volcano, and sediment loads to predictions based on simple plate models it has been possible to estimate the effective elastic thickness of the lithosphere, T_e. In the oceans, T_e is the range 2–50 km and is determined mainly by plate and load age. The continents, in contrast, are characterised by T_e values of up to 80 km and greater. Rheological considerations based on data from experimental rock mechanics suggest that T_e reflects the *integrated* brittle, elastic and ductile strength of the lithosphere. T_e differs, therefore, from the seismogenic layer thickness, T_s, which is indicative of the depth to which anelastic deformation occurs as unstable frictional sliding. Despite differences in their time scales, T_e and T_s are similar in the oceans where loading reduces the initial mechanical thickness to values that generally coincide with the thickness of the brittle layer. They differ, however, in continents, which, unlike oceans, are characterised by a multi-layer rheology. As a result, $T_e \gg T_s$ in cratons, many convergent zones, and some rifts. Most rifts, however, are characterised by a low T_e that has been variously attributed to a young thermal age of the rifted lithosphere, thinning and heating at the time of rifting, and yielding due to post-rift sediment loading. Irrespective of their origin, the Wilson cycle makes it possible for low values to be inherited by foreland basins which, in turn, helps explain why similarities between T_e and T_s extend beyond rifts into other tectonic regions such as orogenic belts and, occasionally, the cratons themselves.

Keywords: lithosphere; flexure; elastic thickness; seismogenic layer; rheology; mantle

1. Introduction

Plate tectonics is based on the observation that the Earth's outermost layers, or lithosphere, can be divided into a number of plates which have remained rigid for long periods (i.e. $>10^5$ yr) of geological time. One manifestation of plate rigidity is the flexural response of the lithosphere to surface loads such as ice, volcanoes, and sediments, sub-surface loads such as those associated with tectonic emplacement and magmatic intrusions, and tectonic boundary loads such as slab pull and ridge push. By comparing the flexure observed in the vicinity of these loads to the predictions of simple elastic, plastic, and viscous

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plate models, it has been possible to determine the flexural rigidity of the lithosphere and its spatial and temporal variation.

Both continental and oceanic lithosphere are modified by flexure. The two types of lithosphere, however, have had a distinct physical and chemical evolution. As a result, they respond to long-term loads in fundamentally different ways. The oceanic lithosphere, for example, is relatively young, has a single-layer rheology and, as a consequence, its effective flexural rigidity (which is determined by the elastic thickness, T_e) depends mainly on thermal age [1]. The continental lithosphere, in contrast, is relatively old, has a multilayer rheology, and its flexural rigidity does not depend on a single controlling parameter such as thermal age, but is an integrated effect of a number of other factors that include the geotherm, crustal thickness, and composition [2].

Rock mechanics data suggest that the flexural rigidity of the lithosphere is determined by the brittle and ductile properties of the constitutive rocks that comprise it. Since brittle behaviour, as described by Byerlee's law, is only possible at relatively low confining pressures and depths, it is likely that this deformation type will determine the strength of only the uppermost part of the lithosphere [3–5]. Ductile strength, in contrast, only *slowly* increases with confining pressure, but strongly decreases with temperature [6]. The transition to brittle, elastic or ductile behaviour is mainly determined by the weakest rheology at the respective depth [4,5]. However, brittle frictional sliding is improbable at great depths (i.e. $>$50–70 km), irrespective of the ductile strength.

As first shown by Goetze and Evans [7], loads on the lithosphere are supported by a quasi-elastic layer that separates the brittle and ductile deformation fields. This competent layer is the strong portion of the lithosphere that is capable of supporting loads for long periods (i.e. $>10^5$ yr) of geological time. Watts et al. [8] and Burov and Diament [9] dubbed this layer the elastic 'core' and it has become popular to associate T_e with its thickness. It is important to point out, however, that the T_e revealed by flexure studies is the depth integral of the bending stress, which is not necessarily associated with a particular competent layer. Indeed, the brittle *and* ductile fields also contribute to the apparent strength of the lithosphere. In the oceans, the depth of the brittle–ductile transition (BDT) falls approximately mid-way in the elastic portion of the lithosphere. As a result, the brittle and ductile deformation fields are roughly equally involved in the support of loads. In the continents, there may be more than one BDT and so the relationship of the brittle and ductile fields to the elastic portion of the lithosphere is more complex.

Plate flexure studies [2,10,11] have shown that T_e corresponds to, or indeed exceeds, the strong portion of the lithosphere, the thickness of which ranges from 2 to 50 km for oceanic lithosphere and as high as 100+ km for the continental lithosphere.

Probably the best known manifestation of brittle deformation is seismicity data. It has been known for some time [12], for example, that continental earthquakes are confined to a seismogenic layer, T_s, in the uppermost brittle part of the crust. More recent studies, using improved hypocentre location techniques [13], have shown that in regions such as California, the Aegean, Tibet and the Zagros mountains, T_s is remarkably uniform in thickness and rarely exceeds 20 km. Seismicity extends to depths as great as 40 km in old shield areas such as the Tien Shan, the Indian shield, and in the vicinity of the East Africa and Lake Baikal rift systems, but is usually less intense than it is in the upper crust.

Rock mechanics data help constrain the mechanisms that control the distribution of earthquakes. These data suggest, for example, that despite its usual representation as pressure-dependent Byerlee behaviour, frictional sliding in continental rocks is controlled by other factors such as geotherm, strain rate, mineralogical inclusions, and fluid content. In general, frictional sliding is limited to a depth of $\sim$15 km (and $<$6–8 km depth on large faults). Beginning at this depth, a semi-brittle/semi-ductile strain rate-dependent plastic flow occurs, with a frictional component that is no longer significant at depths $>$40–50 km [14–16]. This helps explain why seismicity in the oceanic lithosphere is actually limited to this depth range. It also accounts for why earthquakes are so rare in the continental sub-crustal mantle. This is because these depths typically correspond to the lowermost part of the continental crust where ductile creep is dominant.

There are, of course, regions where earthquakes extend to depths >40 km, for example, in subduction zones. However, it is generally agreed [4,5] that deep seismicity is not related to frictional sliding, at least not in the form expressed by Byerlee's law, but to various other metastable mechanisms. These mechanisms are only weakly related to absolute rock strength. For example, it has been known for some time that unlike shallow (i.e. depths <50–70 km) earthquakes, deep earthquakes produce very few aftershocks. Shallow earthquakes, on the other hand, are characterised by numerous aftershocks that decay exponentially with time. This aftershock behaviour is a strong argument that the earthquake-generating mechanism differs between shallow and deep earthquakes.

Because both T_e and T_s reflect the strength of the lithosphere, Jackson and White [17] assumed that the two parameters are the same. While there *is* some similarity between T_e and T_s in the oceans [18], their relationship in the continents is unclear. Recently, Maggi et al. [19] suggested that the strength of the lithosphere is determined by the extent of the brittle deformation and therefore that T_s is the same as T_e. Hence, T_e, like T_s, should be low (<20 km), except perhaps in old shield regions where it might extend to depths as great as 40 km. Such an assertion, which implies that the crust, rather than the mantle, is mainly involved in the support of long-term loads, raises serious questions concerning the observations of flexure, the extrapolations of data based on experimental rock mechanics, and the results of analytical and numerical modelling of flexure.

The purpose of this paper is to consider the relationship between the elastic thickness, T_e, and the seismogenic layer, T_s. We begin by reviewing the results of flexure studies. The physical meaning of T_e and T_s is then discussed in the light of analytical and numerical models for plate flexure with both a single- and multi-layer rheology. We conclude that T_e and T_s reflect the fundamentally different ways that the Earth's lithosphere responds to loads of variable spatial and temporal scales and, therefore, while both parameters are indicative of strength, they are *not* the same.

2. Observations of flexure

There have been a number of studies that have focussed on the estimation of T_e and T_s. Fig. 1 shows a histogram plot of T_e derived from flexure studies and T_s as revealed by the depths of intraplate earthquakes. Although there are currently debates about the reliability of both T_e [20] and T_s [13] estimates, we believe that the distributions shown in Fig. 1 are representative and will not change significantly as more direct measures of the surfaces of flexure are used to constrain T_e and improved earthquake hypocentral location techniques are used to define T_s.

Fig. 1 shows that T_e and T_s are similar in the oceans with the large majority of values <50 km. Moreover, both T_e and T_s clearly exceed the average thickness of oceanic crust. Therefore, not only do earthquakes occur in the sub-oceanic mantle, but it is also involved in the support of long-term loads. The continents, by way of contrast, show a T_s that is confined to the crust and, interestingly, a T_e that sometimes exceeds T_s.

While T_e and T_s are similar in the oceans and in some continents, we believe that they have fundamentally different meanings. T_s reflects the

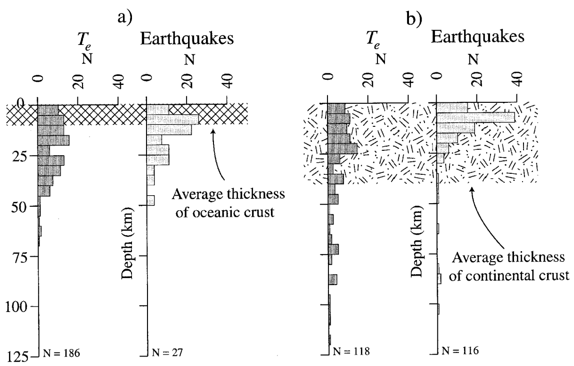

Fig. 1. Histograms of global compilations of elastic thickness, T_e, and seismogenic layer thickness, T_s. (a) Oceans. (b) Continents. The compilation is based on data in tables 6.1 and 6.2 of Watts [80], tables 1–6 of Chen and Molnar [12], and table 1 of Wiens and Stein [18].

strength of, or more precisely the stress level in, the uppermost brittle layer of the lithosphere while T_e is indicative of the strength of the elastic portion of the lithosphere that supports a particular load and, importantly, the brittle *and* ductile strength of the entire elastic layer. The difference is most obvious in unflexed lithosphere when T_s is zero and T_e takes on its highest possible value [2]. Differences persist into flexed lithosphere where T_e slowly decreases with increasing curvature and, hence, bending stress [11,21] while T_s simply reflects the local stress level. Yield strength envelope (YSE) considerations (e.g. Fig. 2) suggest that in the case of the oceans, with their single-layer rheology, the BDT just happens to fall approximately mid-way in the strong elastic portion of the lithosphere. Therefore, T_e and T_s may well be similar.

These conclusions are in accord with observations at deep-sea trench–outer rise systems. Because of the high curvatures that are experienced by the oceanic lithosphere as it approaches a trench, T_e is less than it would otherwise be on the basis of plate age because of yielding. Judge and McNutt [22], for example, have shown that in the high-curvature region of the seaward wall of the northern Chile trench (curvature $K = 1.3 \times 10^{-6}$ m^{-1}) T_e is 22 ± 2 km, which is less than the T_e of 34 km that these workers expected on the basis of the thermal age of the subducting oceanic lithosphere.

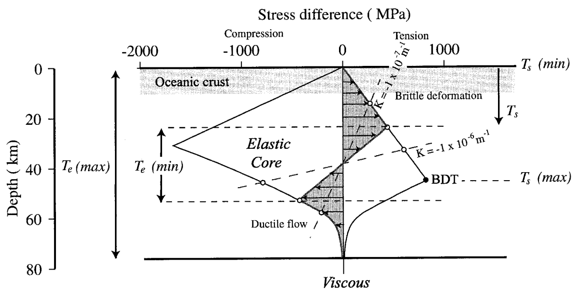

Fig. 2. T_e and T_s for an idealised case of flexure of the oceanic lithosphere. Thin solid line shows the YSE for 80 Ma oceanic lithosphere. The brittle behaviour is described by Byerlee's law, the ductile behaviour by an olivine power law (stress constant, A, of 7×10^{-14} Pa^{-3} s^{-1}; creep activation energy, Q, of 512 kJ/mol [81]), and the thermal structure by the cooling plate model [82]. Thick solid line shows the stress difference for a load which generates a moment, M, of 2.2×10^{17} N/m and curvature, K, of 5×10^{-6} m^{-1}. The figure shows that the load is supported partly by a strong elastic layer or 'core' and partly by the brittle and ductile strength of the lithosphere. The dashed lines show the cases of K of 1×10^{-7} m^{-1} and 1×10^{-6} m^{-1} which bracket the range of observed values at deep-sea trench–outer rise systems [7,11,22]. The figure shows that T_s corresponds to the depth of the intersection of the moment–curvature curve with the brittle deformation field, but could extend from the surface, T_s (min), to the BDT, T_s (max). T_e, in contrast, could extend from the thickness of the elastic 'core', T_e (min), to the thickness of the entire elastic plate, T_e (max). Both T_s and T_e depend on the moment generated by the load and, hence, the plate curvature.

Fig. 3 compares T_e and T_s at northern Chile as well as other different deep-sea trench–outer rise systems in the Pacific and Indian oceans. The figure shows that T_e and T_s are similar, at least for ages of the oceanic lithosphere ≤ 100 Ma. For greater ages, there is a suggestion of a 'threshold' beyond which T_e exceeds T_s. This is attributed to the fact that as the lithosphere ages, its strength increases, and so the curvature and, hence, the bending stress associated with the same load decreases. Therefore, T_e, which mainly reflects the integrated strength of the lithosphere, increases while T_s, which depends more directly on the stress level, decreases.

We have based the discussion thus far on a YSE that only considers the stresses generated in a flexed plate by bending. We have not, therefore, taken into account the effect of any in-plane stresses that act on the plate due, for example, to tectonic boundary loads. It is difficult to conceptualise T_e and T_s from a YSE in the presence of such stresses. However, we can say that T_s will be conditioned by the brittle failure due to these stresses and, as a consequence, will be even less related to the bending stress and, hence, the local T_e.

The oceanic T_e estimates plotted in Fig. 1 are based on more than 48 individual studies and so can be considered a reliable measure of the long-term ($> 10^5$ yr) flexural properties of oceanic lithosphere. The continental T_e estimates, in contrast, are based on fewer studies (16) and some authors [20] have recently questioned their validity, especially estimates that are in excess of 25 km.

The evidence for high continental T_e is based on forward models in which observations of flex-

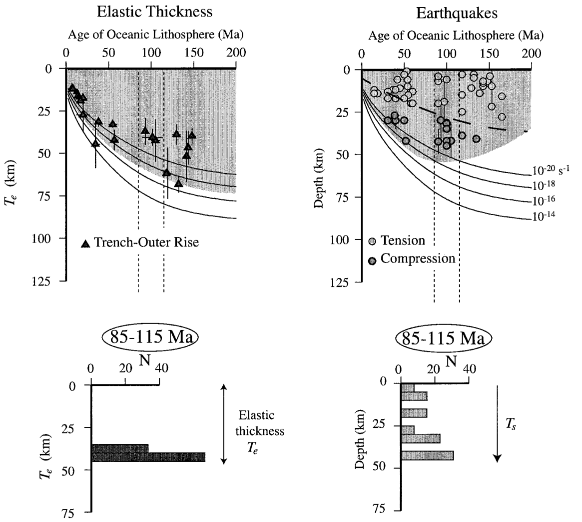

Fig. 3. Summary of T_e and T_s estimates for deep-sea trench–outer rise systems. (a) Plot of T_e and depth of earthquakes as a function of the age of the oceanic lithosphere approaching the trench. Data based on table 6.1 of Watts [80] and table 1 of Seno and Yamanaka [83]. The T_e estimates have been corrected for curvature. Solid lines show the YSE based on the same rheological structure as assumed Fig. 1, a stress difference of 10 MPa, and thermal ages of oceanic lithosphere of 0–200 Ma. (b) Summary of T_e and depth of earthquakes for age of oceanic lithosphere of 85–115 Ma. Note the similarity between T_e and the depth to which earthquakes extend.

ure are compared directly to predictions, and spectral, or inverse, methods in which the relationship between gravity anomalies and topography is analysed as a function of wavelength. Forward modelling of late-glacial rebound in North America, for example, reveals T_e values as high as 80–90 km [23]. These estimates are in accord with forward modelling of some of the foreland basins that form by flexure in front of migrating thrust and fold loads. The Ganges basin, for example, is associated with values that are in the range 80–90 km [24–27]. Similar high values have been reported from spectral studies of parts of Australia [28], Siberia [29], Brazil [30], North America [31], and Africa [32].

McKenzie and Fairhead [20], however, have criticised the results of spectral studies, especially those that are based on the Bouguer coherence technique which they consider upper bounds rather than true estimates. These workers suggest the use of the free-air admittance instead since this technique takes into account what are considered the only known loads: surface loads. There is persuasive geological evidence, however, that subsurface or buried loads are also present in the continents [33] and, possibly, the oceans [34]. Such loads, which include intra-crustal thrusts [35], dynamically induced viscous stresses [36], and buoyant magmatic underplate [37], contribute to both the gravity anomaly and the topography.

As Forsyth [33] has shown, however, increasing the ratio of buried to surface loading shifts the isostatic 'roll-over' in the free-air admittance to short wavelengths while increasing T_e shifts it to long wavelengths. It is always possible, therefore, to interpret a free-air admittance that has been attributed to surface loading of a lithosphere with low T_e by a model of combined surface and buried loading of a high-T_e lithosphere. McKenzie and Fairhead [20] are correct when they state that there is no advantage in using the Bouguer coherence over the free-air admittance *if* surface loads are the only ones that are operative. However, if buried loads are also present, then the coherence is better because it is less sensitive to the amount of buried loading and more sensitive to T_e than is the admittance [33].

McKenzie [38], in a recent paper, has taken into account buried loading. His focus, however, is on the buried loads that because of the modifying effects of erosion and sedimentation generate gravity anomalies, but no topography. The effect of such loads is to reduce the Bouguer coherence – in much the same way as would a surface load which is emplaced on a sufficiently rigid plate that there is no flexure and, hence, Bouguer anomaly. McKenzie [38] showed that in the region of the US Mid-Continent gravity 'high' the free-air coherence is nearly zero for wavelength, λ, <200 km suggesting there is no topography that might be caused by buried loading. The topography of the region is indeed subdued. However, it is by no means clear that erosion and sedimentation will *always* reduce a landscape to a flat plane. To the contrary, erosion and sedimentation are part of an isostatic cycle that involves uplift and subsidence, both of which would be expected to contribute to the gravity anomaly and the topography. The contributions may, because of the strength of the lithosphere, be small and so we agree with McKenzie [38] that it is always important to determine whether the coherence (free-air or Bouguer) is high enough to yield a reliable T_e estimate. This is now possible using techniques such as the maximum entropy method (MEM) [39,40] where spectral estimates are calculated in boxes that are moved step-wise across a study area.

Significantly for this discussion, spectral techniques are not the only means of estimating continental T_e. The large majority of these estimates are based on forward modelling methods in which observations are compared directly to the predictions of elastic plate models. Forward modelling has the advantage that it is not sensitive as to whether the free-air or Bouguer gravity anomaly is used. Indeed, many forward modelling studies are not based on the gravity anomaly at all, but on direct observations of the surfaces of flexure such as those derived from seismic reflection and refraction profile, stratigraphic, and deep well data. Where field geological data exist for buried loads (e.g. in the region of the US Appalachian gravity 'high'), they can be explicitly taken into account in forward modelling [41].

One way to verify the T_e derived from spectral analysis is to compare it to the results of forward modelling. Although only a few such comparisons have been carried out, they show [42–44] good agreement between the T_e estimated using the two different approaches. Armstrong and Watts [44], for example, used the MEM method of analysing Bouguer anomaly and topography data and showed that T_e in the southern Appalachians is in range 50–60 km. This is in agreement with the mean estimate of 57.5 km based on the forward modelling of profiles [41].

However, McKenzie and Fairhead [20] have also queried the high T_e values determined from forward modelling. They obtained, for example, a best fit T_e for the Ganges foreland basin of 42 km, which is about a factor of 2 smaller than the previous estimates by Lyon-Caen and Molnar [25] and Karner and Watts [24]. Their forward modelling followed a 'no-load' approach (D.P. McKenzie, written communication) in which only the flexure and gravity anomaly to one side of a load is used to derive T_e. This is different from the approach of Lyon-Caen and Molnar [25] and Karner and Watts [24] who used the topography to define the load. In this approach, the flexure and gravity anomaly to one side of and, importantly, beneath a load is used to estimate T_e.

Fig. 4 compares the T_e derived using the two approaches along a composite of five closely spaced profiles of the Ganges foreland basin. The figure shows that they yield similar results. T_e is high and $\sim$60 km. However, the minimum in the root mean square (RMS) error between observed and calculated Bouguer gravity anomaly is broad. This makes it difficult to resolve the actual value of T_e. We can say though that T_e is high since application of the same approaches to a profile of the Hawaiian ridge shows that if T_e was lower, as it is in this tectonic setting, then the minimum would be sharper.

These considerations of spectral and forward modelling results suggest there is no reason to query high continental T_e, especially those estimates that are based on the MEM technique where a coherence criterion is applied and forward modelling which takes into account the possibility of both surface and sub-surface loading. While a majority of the T_e estimates are, indeed, low, there are values of up to 80+ km. Hence, T_e in the continents can be significantly larger than the crustal thickness and, importantly, T_s.

Unfortunately, there are still only a few continental regions where T_e and T_s have been compared directly. The main reason is that seismicity in the continental plate interiors is limited in its spatial extent. Seismicity, first of all, implies some level of tectonic stress. These stresses would be expected to be higher in the oceans, where the displacement rates, and probably strain rates, are typically an order of magnitude higher than they are in the continents [45]. Moreover, stresses decrease away from plate boundary zones [46] and the tectonic strains may be insufficiently large in the slowly deforming interior regions of the cratons to cause seismicity. Indeed, Byerlee's law (e.g. Fig. 2) implies that in order to produce seismicity below depths of 25 km, a differential stress

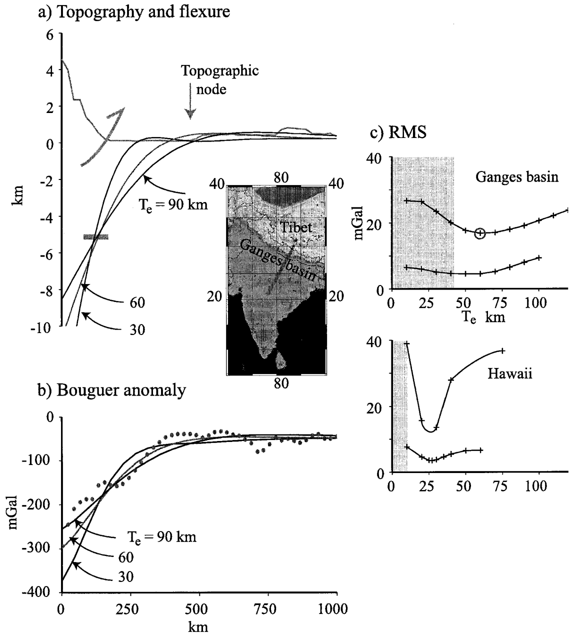

Fig. 4. Comparison of observed and calculated Bouguer gravity anomalies along a profile of the Ganges foreland basin, north-central India. (a) Topography and flexure. Thick line = topography. Thin lines show the calculated flexure due to surface (topographic) loading of a semi-infinite (i.e. broken) plate with a plate break at $x = 0$ and T_e of 30, 60 and 90 km. Thick grey horizontal bar shows the estimate of Burbank [84] for the depth to the base of the foreland basin sequence. Curved half-arrow indicates the approximate location of the main boundary thrust. (b) Bouguer anomaly. Solid dots = observed based on a 5×5 min grid of GETECH's South-East Asia Gravity Project (SEAGP) data (J.D. Fairhead, personal communication). Thin lines show the calculated anomaly based on the flexure profiles in panel a. (c) RMS difference between observed and calculated Bouguer anomalies. Upper plot = Ganges basin. Lower plot = Hawaiian Ridge. The two curves for each plot compare the RMS based on McKenzie's 'no load' (lowermost plot) case to the 'load' (uppermost plot) case for the Ganges basin [24,25] and the Hawaiian Ridge [1]. Note: the RMS in the 'no load' case is based only on the Bouguer anomaly to one side of the load whereas in the 'load' case the RMS is based on the Bouguer anomaly both beneath and to one side of the load.

on the order of 0.4 GPa (tension) to 1.4 GPa (compression) is required.

We concur with Maggi and co-workers [19] that continents, like oceans, show cases where $T_e \approx T_s$ and, indeed, where $T_e < T_s$. However, there are many cases of $T_e \gg T_s$. Two examples are Fennoscandia and the Baikal Rift. In Fennoscandia, the thickness of the seismogenic layer based on an earthquake data set that spans 1965–1995 and a magnitude threshold of 3.0 is 25–30 km whereas YSE considerations [47] and Bouguer coherence studies using both periodogram [48] and MEM [49] methods suggest a T_e that is locally in excess of 60 km. In the Baikal Rift, historical seismicity peaks at depths of 15–20 km [50]. This is significantly less than the 45–55 km, 30–50 km and 40–60 km reported [51–53] for T_e based on forward modelling of free-air and Bouguer gravity anomaly data along west–east topographic profiles of the rift.

3. Analytical and numerical models

Mechanical models that take into account both brittle and ductile deformation provide additional insight into the manner that the lithosphere responds to long-term loads. Moreover, they are capable of predicting the relationship that would be expected between parameters such as T_e and T_s.

Consider first the flexure of a thin elastic plate that overlies an inviscid fluid, a well-known problem in mechanical engineering. As has been shown by Turcotte and Schubert [54], among others, the thickness of such a plate, T_e(elastic) is given by:

$$T_e(\text{elastic}) = \left(\frac{M_{\text{elastic}} 12(1-\upsilon^2)}{EK}\right)^{1/3} \qquad (1)$$

where M_{elastic} = the bending moment, K = curvature, and E and υ are the Young's modulus and Poisson's ratio respectively. M_{elastic} and K are indicative of the total amount of flexure and, hence, are directly related to the stress and strain that accumulates in a deformed plate.

Theoretical considerations suggest that, with some modification, Eq. 1 is valid for *any* rheology (e.g. elastic, ductile, or viscous) that is capable of maintaining stresses over a sufficient (for a quasi-static approximation) amount of time. YSE considerations suggest, however, that only in the case of a single-layer, elastic rheology, high T_e, and small curvatures, will T_e(elastic) approach the actual mechanical thickness of the lithosphere. For a multi-layered structure, T_e is determined by the bending moment, M_{YSE}, which is the sum of all the moments associated with the brittle, elastic, and ductile parts of the YSE. The correspondence principle allows the behaviour of any competent plate (elastic, plastic, or viscous) to be related to that of an 'equivalent' elastic plate.

In oceanic lithosphere, the equivalent elastic thickness, T_e(YSE), that generates the bending moment M_{YSE} and has the same curvature of an elastic plate, K, is:

$$T_e(\mathrm{YSE}) = \left(\frac{M_{\mathrm{YSE}}12(1-\upsilon^2)}{EK}\right)^{1/3} = \left(\frac{M_{\mathrm{YSE}}}{D_0 K}\right)^{1/3} \quad (2)$$

where $D_0 = E/12(1-\upsilon^2)$. Since D_0 is determined by the elastic properties of the plate and M_{YSE} is determined by depth integration of the YSE, Eq. 2 can be used to calculate T_e(YSE) for different values of plate curvature, K. Again, T_e(YSE) is not the actual thickness of the plate. Rather, it is a 'condensed' thickness that reflects the 'integrated' strength of the flexed, competent, plate.

McAdoo et al. [55] used Eqs. 1 and 2 to calculate the ratio of T_e(YSE) to T_e(elastic) for a plate of thermal age 80 Ma, an olivine rheology, and a uniform strain rate of 10^{-14} s^{-1}. They showed that for low curvatures (i.e. $K < 10^{-8}$ m^{-1}) the ratio is 1, indicating little difference between the elastic thickness values. However, as curvature increases, the ratio decreases as T_e(YSE) decreases and the flexed plate yields. For $K = 10^{-6}$ m^{-1} the ratio is $\sim$0.5, indicating 50% yielding and a corresponding reduction in the elastic thickness.

The fact that the oceanic lithosphere yields can be understood in terms of simple mechanics. Ideal elastic materials support any stress level. In the case of real materials, however, stress levels are limited to the rock strength at the corresponding depth. Strains in a bending plate increase with distance from the neutral surface. Consequently, the upper and lowermost parts of a flexed plate are subject to higher strains and may experience brittle or ductile deformation as soon as the strain can no longer be supported elastically. These deformed regions constitute zones of mechanical weakness since the stress level there is lower than it would be if the material maintained its elastic behaviour and, importantly, is lower than it would be in the elastic portion of the plate that separates these two regions. The level of stress in the brittle and ductile region, however, is not zero. A load emplaced on the oceanic lithosphere will therefore be supported partly by the strength of the elastic 'core' and partly by the brittle and ductile strength of the plate. The significance of T_e that has been estimated in regions such as trenches where yielding occurs is that it reflects this combined, integrated, strength of the plate.

T_s has its own significance. Byerlee's law of frictional brittle failure, which characterises deformation in the uppermost part of the lithosphere, suggests that strength linearly increases with pressure and, hence, depth. According to this law, the minimal brittle rock strength in the absence of fluid pressure is at least 65% of the lithostatic pressure at the corresponding depth. Near the surface, the strength is $\sim$10–20 MPa while at depth it may theoretically reach a few GPa. For this reason, the Earth's uppermost layers will fail more easily than its lowermost layers even for the same amount of strain. In the oceans, where the crust is only $\sim$10 km thick, this leads to a concentration of seismicity in the crust and uppermost mantle. In the continents, however, where the crust is thicker, seismicity is concentrated in the uppermost parts of the crust, in contrast to the lowermost part of the crust and uppermost mantle that may never fail due to the lack of a sufficient stress. Indeed, at typical Moho depths (i.e. $\sim$35 km), the theoretical brittle rock strength reaches 0.6 GPa (tension) to 2.0 GPa (compression). To exceed these stresses by, for example, tectonic strain, a fibre force of at least $\sim 10^{13}$ N is required, which is an order of magnitude higher than any previous estimates of the body and plate dynamics forces for the whole lithosphere [56].

The significance of T_e in the continents is not as clear as it is in the oceans. This is because the continents may comprise more than one brittle layer that is de-coupled from an underlying layer by an intermediate ductile layer. As Burov and Diament [2] have shown for thermally 'young' continental lithosphere, a weak ductile layer in the lower crust does not allow bending stresses to be transferred between the strong brittle layers that 'sandwich' it and this leads to a mechanical de-coupling between them. Nevertheless, stress levels are reduced for the same amount of flexure and so each brittle layer is involved in the support of a load. T_e reflects the strength of each elastic layer and the combined strength of all the brittle and ductile layers. It is not simply a sum of the thickness of these layers (h_1, $h_2 ... h_n$), however, but is given by the following Kirchoff relation [2]:

$$T_e(\mathrm{YSE}) \approx (h_1^3 + h_2^3 + h_3^3 ...)^{1/3} = \left(\sum_{l=1}^{n} h_l^3\right)^{1/3} \quad (3)$$

In the case of equally strong layers ($h_1 = h_2 = h_3 ... = h$): $T_e(\mathrm{YSE}) \approx n^{1/3}h$, which yields $T_e \approx 1.25h$ for two strong layers ($n = 2$). The meaning of T_e(YSE) in the continents is now clearer. It reflects the integrated effect of *all* the competent layers that are involved in the support of a load, including the weak ones.

As in the case for the single-rheology-layer oceanic lithosphere, if the multi-layered continental lithosphere is subject to large loads, it flexes, and the curvature of the deformed plate, K, increases. T_e(YSE) is again a function of K and is given [2] [9] by:

$$T_e(\mathrm{YSE}) = T_e(\mathrm{elastic})C(K, T, h_1, h_2 ...) \quad (4)$$

where C is a function of the curvature, K, the thermal age, T, and the rheological structure. A precise analytical expression for C is bulky, although Burov and Diament [9] provide a first-order approximation for a 'typical' case of continental lithosphere with a mean crustal thickness of 35 km, a quartz-dominated crust, and an olivine-dominated mantle which they indicate is valid for $10^{-9} < K < 10^{-6}$ m^{-1}. T_e(YSE) then simplifies to:

$$T_e(\mathrm{YSE}) \approx T_e(\mathrm{elastic})$$
$$\left(1-(1-K/K_{\max})^{1/2}\right)^{(1/2+1/4(T_e(\mathrm{elastic})/T_e(\max)))} \quad (5)$$

where $K_{\max}$ (in m^{-1}) $= (180\times10^3\ (1+1.3T_e(\min)/T_e(\mathrm{elastic}))^6)^{-1}$, $T_e(\max) = 120$ km, $T_e(\min) = 15$ km, and T_e(elastic) is the initial elastic thickness prior to flexure and can be evaluated from Eq. 3.

The considerations above are based only on first-order approximations. It is therefore difficult to precisely track T_e and, particularly, T_s, both of which would be expected to change as a result of loading.

We show in Fig. 5, therefore, how T_e and T_s would be expected to change using the more precise analytical formulations of Burov and Diament [21]. The figure illustrates how the thicknesses of the brittle and ductile layers evolve with different loads and, hence, curvatures. On bending, brittle failure and, hence, the potential for seismicity preferentially develops in the uppermost part of the crust. The onset of brittle failure in the mantle is delayed, however, and does not occur until the amount of flexure and, hence, curvature is very large.

Observations of curvature in the region of large loads help constrain the strength of continental lithosphere. Curvatures range from 10^{-8} m^{-1} for the sub-Andean [41] to 5×10^{-7} m^{-1} for the West Taiwan [57] foreland basins. The highest curvatures are those reported by Kruse and Royden [58] of 4–5×10^{-6} m^{-1} for the Apennine and Dinaride forelands. Fig. 5 shows, however, that curvatures this high may still not be large enough to cause brittle failure in the sub-crustal mantle, *unless* the flexed plate is subject to an additional in-plane tectonic stress. In the case illustrated in Fig. 5, the tectonic stress that is required to cause failure in the sub-crustal mantle for a plate curvature of 7×10^{-7} m^{-1} is 0.7 GPa. This is already close to the maximum likely value for tectonic boundary loads [59], suggesting that brittle failure and, hence, earthquakes in the mantle will be rare. Instead, seismicity will be limited to the uppermost part of the crust where rocks fail by brittle deformation, irrespective of the stress level. This limit does not apply, of course, to T_e. For curvatures of up to 10^{-6} m^{-1}, Fig. 5 shows that T_e is always larger than T_s. Only for the highest curvatures will $T_e \rightarrow T_s$ and, interestingly, will the case that $T_s > T_e$ arise.

The YSE in Fig. 5 is based on a relatively weak dry quartz and quartz diorite continental crust. The question is raised, therefore, of how T_e and T_s would change if a weaker or stronger crust had been assumed. We know, for example, that fluids may reduce not only the frictional strength, but also the ductile strength of rocks by a significant

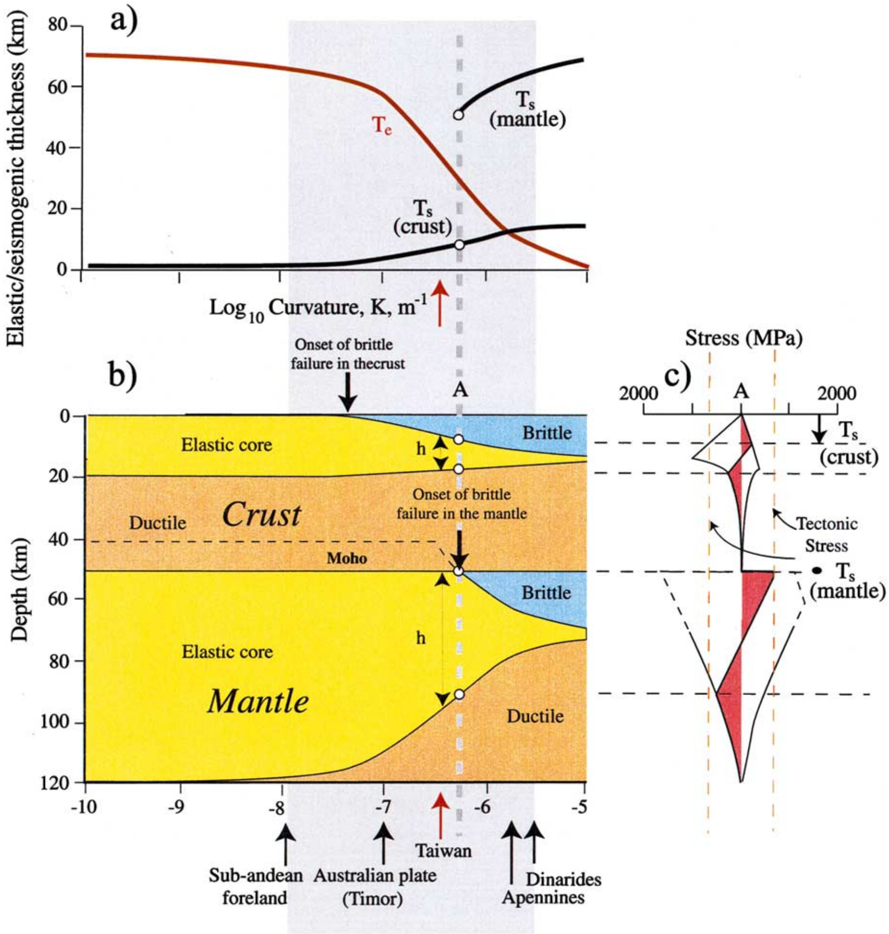

Fig. 5. T_e and T_s for a multi-layered continental lithosphere as a function of plate curvature, *K*. (a) T_e and T_s. Note that there are two main seismogenic layers: one in the uppermost crust and the other in the uppermost mantle. The two layers are separated by an aseismic region. (b) Evolution of the brittle and ductile deformation fields. The observed *K* values are from the sub-Andean [41], West Taiwan [57], Dinarides [85] and Apennines [58] foreland basins and the Australian plate in the Timor region [86]. (c) YSE showing the stress difference for a load that generates a curvature of $K = 7\times10^{-7}$ m^{-1}. The calculations are based on the semi-analytical models of Burov and Diament [21][2] and assume a thermal age of continental lithosphere of 400 Ma, a thermal thickness of 250 km, a mean crustal thickness of 50 km (i.e. an orogenic belt setting), a fixed background strain rate of 10^{-15} s^{-1}, a dry quartzite upper crust ($A = 1.26\times10^{-7}$ MPa$^{-2.7}$ s^{-1} and $Q = 134$ kJ/mol [81]), a dry quartz or quartz diorite lower crust ($A = 5.01\times10^{-15}$ Pa$^{-2.4}$ s^{-1}, $Q = 212$ kJ/mol [87]), an olivine mantle ($A = 7\times10^{-14}$ Pa^{-3} s^{-1}, $Q = 212$ kJ/mol), densities of the upper crust, lower crust, and mantle of 2650, 2900, and 3330 kg/m^3 respectively, thermal diffusivities of the upper crust, lower crust, and mantle of 8.3×10^{-7}, 6.7×10^{-7}, 8.75×10^{-7} m^2/s respectively, and a thermal conductivity of the upper crust, lower crust, and mantle of 2.5, 2.0, and 3.5 W/m/°K respectively. Elastic moduli, *E* and υ, equal 80 GPa and 0.25, respectively. Byerlee's rock friction law [3] is assumed for the brittle portions of the lithosphere. Note that cataclastic flow, pressure solution and some other 'atypical' mechanisms may significantly reduce the frictional rock strength and hence T_s.

amount. As a result, a rock may flow rather than break in response to bending stresses and T_e would be expected to decrease. The effect on T_s is less clear, but a reduction in brittle strength by fluids is not, by itself, the only possible explanation of seismicity in the mantle. As T_e decreases, curvature increases, and so the likelihood that bending stresses will exceed the brittle strength of mantle rocks and cause seismicity also increases. We also know that more mafic rocks, such as diabase, would have a higher ductile strength. As a result, the lower crust is less likely to flow and T_e would be expected to increase. However, we found that when the quartz diorite lower crust in Fig. 5 is replaced by diabase there is little change in T_e, so long as the lower crust is de-coupled from the mantle. This is because diabase is weaker than the mantle and so the mantle strength still dominates the rheology of the entire lithosphere. The main influence, we found, of a diabase lower crust is to T_s which can now extend from depths of 20–30 km, although not as deep as the Moho itself.

These considerations have implications for the relationship between T_s and T_e and, importantly, the strength of the mantle. In low-curvature regions, the mantle may be devoid of earthquakes, but still involved in the support of long-term loads. In high-curvature regions, however, the mantle may have earthquakes, but long-term loads *appear* to be supported by the crust. This apparent 'duality' of role is explicable in terms of a model in which the strength of the lithosphere resides more in the mantle than the crust. The presence or absence of seismicity depends on the curvature and stress level and so by itself is not an indicator that the mantle is weak or strong. The observations of flexure together with the results of analytical modelling, however, indicate that the mantle is strong and that, irrespective of curvature, the thickness of the strong competent layer in the mantle is always greater than that of its counterpart in the crust. This is consistent with the general observation that seismic instabilities may have little or no relation to the long-term strength of rocks that deform in a more stable brittle and ductile regime.

Fig. 6 summarises the expected relationship between T_s and curvature for different thermal ages, coupled and de-coupled continental lithosphere, and oceanic lithosphere. The circles show the maximum observed curvatures and, hence, the maximum likely value of T_s. In the de-coupled continental cases, T_s does not exceed 15 km, which corresponds well with observations. Moreover, there is no obvious correlation between T_e and T_s. T_e always exceeds T_s, irrespective of the thermal age and curvature. High T_e values limit the amount of curvature due to flexure and, hence, the ratio of T_e to T_s increases with thermal age. Interestingly, it is the coupled continental and oceanic lithosphere cases that have the potential to yield the highest values of T_s. The reason is that in both these cases, Byerlee's friction law extends, uninterrupted by a ductile lower crust, from the uppermost part of the crust into the underlying mantle. Observed curvatures, however, still limit T_s in the continents. In contrast, T_s in the oceans could extend to depths of ~30 km.

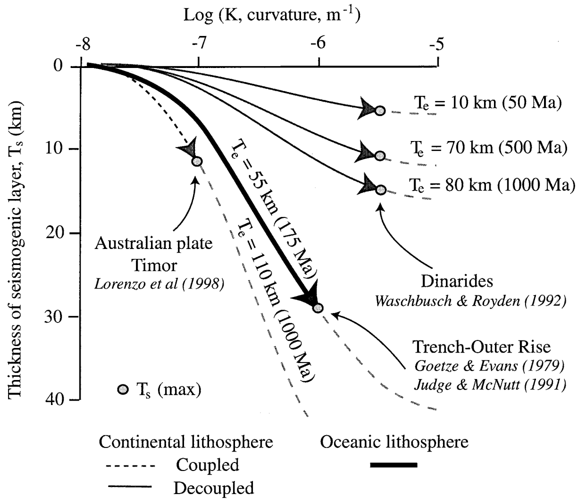

Fig. 6. The relationship between T_s and K for different thermal ages and coupled and de-coupled continental lithosphere. The filled circles show K_{max} and the corresponding T_s(max). Thin lines show de-coupled continental lithosphere of thermal age of 50, 500 and 1000 Ma. Dashed line shows the coupled case which is based on a diabase rather than a quartz diorite lower crust. It is assumed that the Dinarides are representative of the deformation of a de-coupled continental lithosphere while the Australian plate (Timor) is representative of the deformation of a coupled lithosphere. Thick line shows the case of old, strong, oceanic lithosphere as it approaches a deep-sea trench–outer rise system (thermal age = 175 Ma, T_e = 55 km).

The models discussed thus far are limited in that they are based on thin-plate theory and the small-strain approximation. Moreover, we have followed previous studies and assumed that ductile deformation is controlled by uniform, rather than varying, strain rate. We have therefore carried out numerical modelling using the finite-element code PAROVOZ [60]. This code solves the force, mass and energy balance equations for each element of a flexed plate and takes into account large strain deformation. It also allows brittle strain to be localised, thereby simulating faulting. Importantly, there is no limit on the strain rate field. The model accounts for brittle, elastic, and ductile deformation and is similar in formulation to that used by Burov and Molnar [61] and Burov et al. [62]. The main difference is that we consider surface rather than buried loads, as were assumed, for example, by Burov and Diament [2] in their study of orogenic loading.

Fig. 7 shows a numerical model for the flexure of continental lithosphere of thermal age 400 Myr

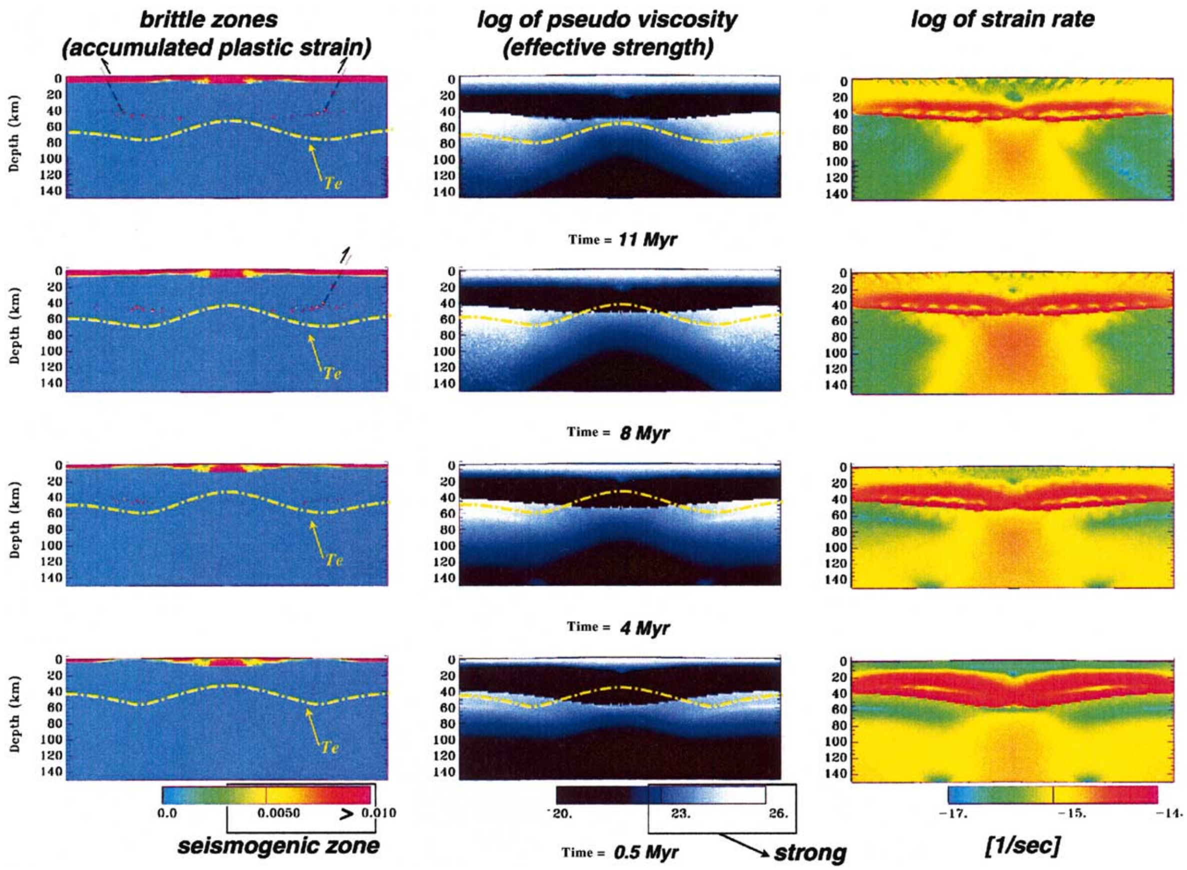

Fig. 7. 'Snapshots' of a numerical large-strain simulation of the response of the continental lithosphere to vertical loading, 0.5, 4, 8, and 11 Myr following the load emplacement. The rheological parameters, crustal structure, and the thermal structure are the same as assumed in Figs. 5 and 6. The figure shows the strain (left-hand panel), viscosity (middle panel), and strain rate (right-hand panel). The strain shows that the potentially active seismic zones are confined to the brittle uppermost part of the crust. The thickness of the seismic zone corresponds approximately to the depth of the BDT. The red dots at ∼40 km depth indicate a potential for seismicity in the lower crust. The presence or absence of seismicity in the lower crust would depend, of course, on the applicability of Byerlee's law to such great depths. The viscosity (as determined from the ratio of stress to strain rate) shows the evolution of weak (blue-black) and strong (blue-white) zones in the lithosphere. Those zones in which the viscosity exceeds 10^{23} Pa s can be considered effectively elastic and their thickness would correspond roughly to that of the strong competent layer. The resulting T_e is shown with yellow dashed line using the same vertical axes as depth. Note that there is no apparent correlation between T_e and the thickness of the uppermost brittle layer. The strain rate, which reaches its maximum values in the ductile lowermost part of the crust, shows that the continental lithosphere responds rapidly to loading.

by a gaussian-shaped load, 3 km high, 200 km wide, and uniform density of 2650 kg/m^3. The figure shows the evolution of the brittle layer, the effective strength of the lithosphere (defined here as the ratio of the deviatoric stress to the strain rate), and the strain rate. The thickness of the brittle layer and, hence, T_s remains small despite changes in stress levels as the lithosphere accommodates the load. In particular, T_s is always three to four times smaller than the thickness of the competent, high-effective-viscosity 'core' of the lithosphere that supports the load for long periods of time. We also note that the model predicts a localisation of brittle strain at the base of the crust as well as along other density and rheological boundaries. The localisation at the base of the crust is explained by a strain rate amplification due to intense ductile flow and strong shear deformation in the narrow low-viscosity zone near the interface between the relatively weak lower crust and relatively strong mantle. In this zone, the flow stress becomes comparable to the brittle strength, thereby promoting occasional brittle failure and, hence, seismicity in what is otherwise a ductile regime for loads of long duration.

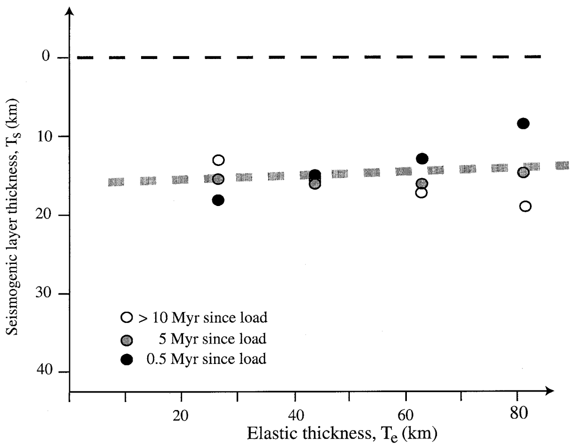

Fig. 8. Relationship between T_e and T_s derived from numerical experiments based on equilibrium equations, an explicit brittle–elastic–ductile rheology, and a variable strain rate. Other model parameters are as given in Fig. 4. The modeling assumes a gaussian-shaped orogenic load (200 km wide, 3 km high). Note that there is no apparent relationship between T_e and T_s. The time dependence of T_s arises because of a load-induced stress relaxation in the lithosphere.

The relationship between T_e and T_s predicted from numerical modelling for load ages of 0.5, 5 and >10 Myr, is illustrated in Fig. 8. T_s is the thickness of the uppermost brittle layer and T_e is determined from the effective flexural rigidity which, in turn, has been derived from a vertical integration of the stress field. The figure shows that T_s is less than 20 km, even for the case that $T_e > 80$ km. T_s varies with load-induced stress relaxation such that if there is any change in T_s, it is a decrease which increases even further the contrast with T_e.

4. Discussion

These considerations of flexure observations and model predictions have clarified, we believe, our understanding of the mechanical behaviour of the Earth's lithosphere. T_s, as defined by the depth of earthquakes, reflects the extent of faulting in the uppermost, competent, part of the lithosphere. The seismogenic layer potentially extends to the BDT. The level of stress, however, may not be sufficiently large to cause failure at such depths. Byerlee's law predicts that T_s in the continents is unlikely to exceed 20 km. Earthquakes may extend deeper, into the brittle part of the sub-crustal mantle on major faults, in regions of high curvature or high bending stress. In this case, there will be two seismogenic layers with an aseismic layer in between.

There is evidence from seismic reflection profile data that the continental Moho is sometimes offset by faults [63], although the significance of this observation is not entirely clear. Despite this, earthquakes in the continental sub-crustal mantle are rare. We interpret this as the consequence of the increase in brittle strength of rocks with confining pressure and the absence of a sufficiently large stress in these regions to overcome the strength of the mantle. This is not to claim that frictional sliding may not give way to a different, aseismic, mechanism at great depths. Rather, if brittle frictional sliding is possible in the region of the Moho, the stresses at these depths are too small to activate such sliding.

T_s reflects current stress levels in the lithosphere. We have shown (Fig. 5), from the observations of curvature and its associated bending stress, that continental T_s is likely to be limited to the uppermost 10–15 km of the crust. T_s may extend deeper, depending on the level of the bending and/or in-plane tectonic stress. Irrespective, we note that T_s correlates with a part of the geotherm that is strongly influenced by crustal heat production (up to 50% of the surface heat flow) [54]. Radiogenic heat production in granites, however, exponentially decreases with depth and becomes negligible at Moho depths. Continental T_s may therefore be limited, in some way, by crustal heat production.

In contrast to T_s, T_e reflects the *integrated* strength of the entire lithosphere. This includes a significant contribution from the aseismic part of the lithosphere. In oceanic lithosphere, the potential brittle zone extends to the BDT which may be as deep as 50 km. This is because there is no intermediate ductile layer that prevents stresses from being propagated into surrounding competent layers. As a result, the stresses generated by flexure accumulate locally and, if they exceed the confining pressure, cause earthquakes. In the continents, however, there are one or more ductile layers which may de-couple the competent parts of the lithosphere and cause smaller stresses for the same amount of flexure and, hence, curvature. Furthermore, the small flexures and long loading times suggest that most continental lithosphere will deform at rates that are significantly smaller than oceanic lithosphere, which further reduces stress levels. While the stresses generated by flexure are large enough to cause earthquakes in the uppermost brittle part of the continental crust, they may not be sufficient to overcome the brittle strength of the continental sub-crustal mantle. It is precisely because of this ‘threshold’, we believe, that mantle earthquakes are rare and continental T_e usually exceeds T_s.

Another, more obvious, difference between T_s and T_e concerns their time scales. T_s reflects weakness on historical time scales in the uppermost competent layer of the lithosphere. T_e, on the other hand, is indicative of the integrated strength of the lithosphere on time scales which exceed $\sim 10^5$ yr.

In the oceans, there is evidence [64] of a time dependence to T_e as the thickness of the lithosphere that supports a load thins from its short-term thickness to its long-term elastic thickness. Oceanic T_e studies suggest that thermal cooling which strengthens the lithosphere dominates over that of load-induced stress relaxation which weakens it, such that the mantle becomes increasingly more involved in the support of loads with thermal age. There is no evidence of a similar dominance in the continents. Nevertheless, the existence of high T_e values in old cratonic regions suggests that the thermal age of the continents plays some role. Certainly, we can say – with some confidence – that the sub-crustal mantle is an important contributor to the support of long-term loads in *both* the oceans and the continents.

While we have emphasised in this discussion the *differences* between T_e and T_s, there are tectonic settings where T_e is low and may, therefore, approach or, on occasions, be less than T_s. Most rift basins are associated with T_e values that are low when compared to that of surrounding terrains. Moreover, the T_e values are in accord with T_s which ranges from 5 to 15 km for most rifts [65]. This similarity does not indicate, however, that T_e and T_s are the same. Rather, the low T_e at rifts is a consequence of all the geological processes that have influenced these features through time. The low T_e has been variously attributed, for example, to a young initial thermal age of the rifted continental lithosphere, heating during rifting, and thermal blanketing and yielding due to post-rift sediment loading [66].

In some rifts, notably East Africa, there is evidence of earthquakes that extend to depths as great as 25–40 km in the crust and, hence, that $T_s > T_e$. It has been proposed [67] that the deep seismicity is indicative of a strong mafic lower crust. However, the brittle failure criterion for such a crust must be different from that associated with Byerlee’s law, which requires very high stresses at such depths. One possible contributor are fluids, which would reduce the brittle strength. Fluids, however, also lower the ductile strength. Another is temperature. The Basin and Range, for example, is associated with high heat flow which imply Moho temperatures of $\sim 600°C$.

There are only a few pyroxene-rich rocks, however, that would remain brittle at such high temperatures. These difficulties lead us to other explanations. Most data from experimental rock mechanics are based on simple monophase rocks (e.g. feldspars), yet we know that monophase rocks are stronger than polyphase ones. Consequently, it may be difficult to maintain high ductile strength in the complex geological terrains that are involved in extensional terrains. The brittle strength of the lower crust has therefore also to be low, in order to deform seismogenically.

One consequence of the Wilson cycle is that, as oceans open and close, rifts become the sites of compression. The T_e structure of rifts may therefore be 'inherited' during subsequent orogenic loading events. The reason for this is that the lithosphere is effectively elastic on long time scales such that the T_e acquired at the time of loading is 'frozen in' and will not change significantly with time. We know this to be the case for the oceans [8] and it is probably also the case for the continents. Although orogenic belts are associated with crustal thickening and high strains, both of which would weaken the lithosphere, they are characterised at one or both of their edges by fold and thrust loads that flex the flanking lithosphere not involved in orogeny, forming foreland basins. The stratigraphic 'architecture' of these basins may therefore reflect the T_e structure of the underlying rift basins. Indeed, low values of T_e have been reported for the Apennine and Dinaride foreland basins [58], which developed on the ancient Tethys margin, and the West Taiwan foreland basin [57], which developed on the present-day South China Sea margin. Other low values have been reported from forelands in the Western USA [40], Caspian Sea [68] and, interestingly, the Superior province in the Canadian shield [69].

Because of inheritance, low T_e values may extend beyond rifts into orogenic belts and, in some cases, to the interiors of the cratons themselves. Low T_e values may therefore appear widespread. They should not be construed, however, as indicating that the sub-crustal mantle is weak in the continents. To the contrary, we believe that the mantle beneath continents is strong. This is probably best manifest in wide foreland basins, such as the Ganges and Appalachian/Allegheny, where thrust and fold loads have been emplaced on high-T_e cratonic interiors, beyond regions that were extensively heated and thinned during rifting [41].

Other considerations suggest that the continents, despite appearing weak in some regions, are also capable of great strength. The loads associated with thrust and fold belts are among the largest to have been emplaced on the Earth's surface, exceeding oceanic volcanic loads, for example, by a factor of 2 or more. Numerical modelling [70–72] suggests that loads of this size are unlikely to be preserved for >1 Myr, *if* T_e is much less than 20 km. This is because loading of weak continental lithosphere would result in a strong ductile flow in the lower crust and, in the case of very low T_e, in the sub-crustal mantle. Such flow will eventually lead to a flattening of the Moho and the collapse of topography. High T_e, on the other hand, inhibits such flattening and collapse. Indeed, a strong sub-crustal mantle is required [73–76] to order to explain the great height of orogenic belts. Finally, the transmission of tectonic stresses over large distances (which by themselves may not be large enough to cause earthquakes), as confirmed by recent geodetic data and observations of large-scale continental folding [77,78], implies high integrated strength of most of the continental interiors.

5. Conclusions

1. The elastic thickness of the lithosphere, T_e, is in the range 2–50 km for oceans and up to 80 km and higher for continents.
2. The seismogenic layer thickness, T_s, is the range 0–40 km in the oceans. In the continents, T_s is generally in the range 0–25 km, but some earthquakes extend to greater depths.
3. T_e and T_s are similar in young (<100 Ma) oceans and in some rifts and orogenic belts.
4. The largest differences between T_e and T_s occur in cratons, in the old, cold, interior of the continents.
5. Mechanical models that take into account the

strength of the brittle, elastic, and ductile parts of the YSE closely reproduce the observed distribution of T_e and T_s in a wide range of tectonic settings.
6. Both the observations of flexure and the predictions of analytical and numerical models indicate that T_e is only weakly correlated with T_s.
7. T_s reflects the thickness of the uppermost weak layer that responds on historical time scales to stresses by faulting and earthquakes. T_e, in contrast, reflects the response of the lithosphere to long-term ($>10^5$ yr) geological loads.
8. T_s reflects the strength of the uppermost brittle layer of the lithosphere, or more precisely the stress level in this layer. T_e, however, reflects the integrated strength of the entire lithosphere, a significant component of which may be provided by an aseismic sub-crustal mantle.
9. The continental lithosphere, unlike its oceanic counterpart, is associated with a multi-layer rheology and small plate flexures. Both factors result in stress levels that are unlikely to approach the very high brittle strength limits below the Moho. The absence of deep mantle seismicity in the continents is therefore more a consequence of its strength, not its weakness.

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References

[1] A.B. Watts, An analysis of isostasy in the world's oceans: 1. Hawaiian-Emperor Seamount Chain, J. Geophys. Res. 83 (1978) 5989–6004.
[2] E.B. Burov, M. Diament, The effective elastic thickness (Te) of continental lithosphere: What does it really mean?, J. Geophys. Res. 100 (1995) 3895–3904.
[3] J.D. Byerlee, Friction of rocks, Pageophys 116 (1978) 615–626.
[4] S. Kirby, W. Durham, L. Stern, Mantle phase changes and deep-earthquake faulting in subducting lithosphere, Science 252 (1991) 216–225.
[5] S.H. Kirby, S. Stein, E.A. Okal, D.C. Rubie, Metastable mantle phase transformations and deep earthquakes in subducting oceanic lithosphere, Rev. Geophys. 34 (1996) 261–306.
[6] C. Goetze, The mechanisms of creep in olivine, Phil. Trans. R. Soc. London 288 (1978) 99–119.
[7] C. Goetze, B. Evans, Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics, Geophys. J. R. Astron. Soc. 59 (1979) 463–478.
[8] A.B. Watts, J.H. Bodine, M.S. Steckler, Observations of flexure and the state of stress in the oceanic lithosphere, J. Geophys. Res. 85 (1980) 6369–6376.
[9] E.B. Burov, M. Diament, Isostasy, equivalent elastic thickness, and inelastic rheology of continents and oceans, Geology 24 (1996) 419–422.
[10] J.H. Bodine, M.S. Steckler, A.B. Watts, Observations of flexure and the rheology of the oceanic lithosphere, J. Geophys. Res. 86 (1981) 3695–3707.
[11] M.K. McNutt, H.W. Menard, Constraints on yield strength in the oceanic lithosphere derived from observations of flexure, Geophys. J. R. Astron. Soc. 71 (1982) 363–394.
[12] W.P. Chen, P. Molnar, Focal depths of intracontinental and intraplate earthquakes and their implications for the thermal and mechanical properties of the lithosphere, J. Geophys. Res. 88 (1983) 4183–4214.
[13] A. Maggi, J.A. Jackson, K. Priestley, C. Baker, A re-assessment of focal depth distributions in southern Iran, the Tien Shan and northern India: do earthquakes occur in the continental mantle?, Geophys. J. Int. 143 (2000) 629–661.
[14] G. Ranalli, Rheology of the Earth, Chapman and Hall, London, 1995, 413 pp.
[15] F.M. Chester, A rheologic model for wet crust applied to strike-slip faults, J. Geophys. Res. 100 (1995) 13033–13044.
[16] B. Bos, C.J. Spiers, Frictional-viscous flow in phyllosilicate-bearing fault rock: Microphysical model and Implications for crustal strength profiles, J. Geophys. Res. 107 (2002) 10.1029/2001JB000301.
[17] J.A. Jackson, N.J. White, Normal faulting in the upper continental crust: observations from regions of active extension, J. Struct. Geol. 11 (1989) 15–32.
[18] D.A. Wiens, S. Stein, Intraplate seismicity and stresses in young oceanic lithosphere, J. Geophys. Res. 89 (1984) 11442–11464.
[19] A. Maggi, J.A. Jackson, D. McKenzie, K. Preistley,

Earthquake focal depths, effective elastic thickness, and the strength of the continental lithosphere, Geology 28 (2000) 495–498.
[20] D.P. McKenzie, D. Fairhead, Estimates of the effective elastic thickness of the continental lithosphere from Bouguer and free-air gravity anomalies, J. Geophys. Res. 102 (1997) 27523–27552.
[21] E.B. Burov, M. Diament, Flexure of the continental lithosphere with multilayered rheology, Geophys. J. Int. 109 (1992) 449–468.
[22] A.V. Judge, M.K. McNutt, The relationship between plate curvature and elastic plate thickness: A study of the Peru-Chile trench, J. Geophys. Res. 96 (1991) 16625–16640.
[23] R.I. Walcott, Isostatic response to loading of the crust in Canada, Can. J. Earth Sci. 7 (1970) 716–727.
[24] G.D. Karner, A.B. Watts, Gravity anomalies and flexure of the lithosphere at mountain ranges, J. Geophys. Res. 88 (1983) 10449–10477.
[25] H. Lyon-Caen, P. Molnar, Constraints on the structure of the Himalaya from an analysis of gravity anomalies and a flexural model of the lithosphere, J. Geophys. Res. 88 (1983) 8171–8191.
[26] A. Caporali, Gravity anomalies and the flexure of the lithosphere in the Karakoram, Pakistan, J. Geophys. Res. 100 (1995) 15075–15085.
[27] Y. Jin, M.K. McNutt, Y. Zhu, Evidence from gravity and topography data for folding in Tibet, Nature 371 (1994) 669–674.
[28] M.T. Zuber, T.D. Bechtel, D.W. Forsyth, Effective elastic thickness of the lithosphere and mechanisms of isostatic compensation in Australia, J. Geophys. Res. 94 (1989) 13919–13930.
[29] M.G. Kogan, J.D. Fairhead, G. Balmino, E.L. Makedonskii, Tectonic fabric and lithospheric strength of northern Eurasia based on gravity data, Geophys. Res. Lett. 21 (1994) 2653–2656.
[30] N. Ussami, N. Côgo de Sá, E.C. Molina, Gravity map of Brazil. 2. Regional and residual isostatic anomalies and their correlation with major tectonic provinces, J. Geophys. Res. 98 (1993) 2199–2208.
[31] T.D. Bechtel, D.W. Forsyth, V.L. Sharpton, R.A.F. Grieve, Variations in effective elastic thickness of the North American lithosphere, Nature 343 (1990) 636–638.
[32] R. Hartley, A.B. Watts, J.D. Fairhead, Isostasy of Africa, Earth Planet. Sci. Lett. 137 (1996) 1–18.
[33] D.W. Forsyth, Subsurface loading and estimates of the flexural rigidity of continental lithosphere, J. Geophys. Res. 90 (1985) 12623–12632.
[34] G. Ito, A. Taira, Compensation of the Ontong Java Plateau by surface and subsurface loading, J. Geophys. Res. 105 (2000) 11171–11183.
[35] W.E. Holt, T.A. Stern, Subduction, platform subsidence, and foreland thrust loading: The late Tertiary development of Taranaki Basin, New Zealand, Tectonics 13 (1994) 1068–1092.
[36] E.B. Burov, M.G. Kogan, H. Lyon-Caen, P. Molnar, Gravity anomalies, the deep structure, and dynamic processes beneath the Tien Shan, Earth Planet. Sci. Lett. 96 (1990) 367–383.
[37] A.B. Watts, J.D. Fairhead, Gravity anomalies and magmatism at the British Isles continental margin, J. Geol. Soc. London 154 (1997) 523–529.
[38] D. McKenzie, Estimating T_e in the presence of internal loads, J. Geophys. Res. (2003) in press.
[39] A.R. Lowry, R.B. Smith, Flexural rigidity of the Basin and Range-Colorado Plateau-Rocky Mountain transition from coherence analysis of gravity and topography, J. Geophys. Res. 99 (1994) 20123–20140.
[40] A.R. Lowry, R.B. Smith, Strength and rheology of the western U.S. Cordillera, J. Geophys. Res. 100 (1995) 17947–17963.
[41] J. Stewart, A.B. Watts, Gravity anomalies and spatial variations of flexural rigidity at mountain ranges, J. Geophys. Res. 102 (1997) 5327–5352.
[42] M.G. Kogan, M.K. McNutt, Isostasy in the USSR I: Admittance data, in: K. Fuchs, C. Froidevaux, Eds., Composition, Structure, and Dynamics of the Lithosphere-Asthenosphere System, Geodynamics Series 16, AGU, Washington, DC, 1987, pp. 301–307.
[43] M.K. McNutt, M. Kogan, Isostasy in the USSR II: Interpretation of admittance data, in: K. Fuchs, C. Froidevaux, Eds., Composition, Structure, and Dynamics of the Lithosphere-Asthenosphere System, Geodynamics Series 16, AGU, Washington, DC, 1987, pp. 309–327.
[44] G.D. Armstrong, A.B. Watts, Spatial variations in T_e in the southern Appalachians, eastern United States, J. Geophys. Res. 106 (2001) 22009–22026.
[45] C. DeMets, R.G. Gordon, D.F. Argus, S. Stein, Current plate motions, Geophys. J. Int. 101 (1990) 425–478.
[46] M.H.P. Bott, D.S. Dean, Stress diffusion from plate boundaries, Nature 243 (1973) 339–341.
[47] V. Pasquale, M. Verdoya, P. Chiozzi, Heat flux and seismicity in the Fennoscandian Shield, Phys. Earth Planet. Inter. 126 (2001) 147–162.
[48] Y.H. Poudjom-Djomani, J.D. Fairhead, W.L. Griffin, The flexural rigidity of Fennoscandia: reflection of the tectonothermal age of the lithospheric mantle, Earth Planet. Sci. Lett. 174 (1999) 139–154.
[49] G.D. Armstrong, A.B. Watts, Mapping flexural rigidity of the lithosphere in the eastern USA and mainland Europe, EGS Newsletter, Scientific Programme, Nice, 69 (2000).
[50] J. Déverchère, C. Petit, N. Gileva, N. Radziminovitch, V. Melnikova, V. San'kov, Depth distribution of earthquakes in the Baikal rift system and its implications for the rheology of the lithosphere, Geophys. J. Int. 146 (2001) 714–739.
[51] E.B. Burov, F. Houdry, M. Diament, J. Déverchère, A broken plate beneath the North Baikal rift zone revealed by gravity modelling, Geophys. Res. Lett. 21 (1994) 129–132.
[52] P. vanderBeek, Flank uplift and topography at the central

Baikal Rift (SE Siberia): A test of kinematic models for continental extension, Tectonics 16 (1997) 122–136.
[53] C. Petit, C. Ebinger, Flexure and mechanical behaviour of cratonic lithosphere: Gravity models of the East African and Baikal rifts, J. Geophys. Res. 105 (2000) 19151–19162.
[54] D.L. Turcotte, G. Schubert, Geodynamics: Applications of Continuum Physics to Geological Problems, John Wiley, New York, 1982,450 pp.
[55] D. McAdoo, C.F. Martin, S. Poulouse, Seasat observations of flexure: Evidence for a strong lithosphere, Tectonophysics 116 (1985) 209–222.
[56] P. Molnar, H. Lyon-Caen, Some simple physical aspects of the support, structure, and evolution of mountain belts, Geol. Soc. Am. Spec. Pap. 218 (1988) 179–207.
[57] A.T. Lin, A.B. Watts, Origin of the West Taiwan basin by orogenic loading and flexure of a rifted continental margin, J. Geophys. Res. 107 (2002) 10.1029/2001JB000669.
[58] S.E. Kruse, L.H. Royden, Bending and unbending of an elastic lithosphere: The Cenozoic history of the Apennine and Dinaride foredeep basins, Tectonics 13 (1994) 278–302.
[59] S. Cloetingh, E.B. Burov, Thermomechanical structure of European continental lithosphere: constraints from rheological profiles and EET estimates, Geophys. J. Int. 124 (1996) 695–723.
[60] A.N.B. Poliakov, P. Cundall, Y. Podladchikov, V. Laykhovsky, An explicit inertial method for the simulation of visco-elastic flow: An evaluation of elastic effects on diapiric flow in two- or three-layers models, in: D.B. Stone, S.K. Runcorn (Eds.), Flow and Creep in the Solar System: Observations, Modelling and Theory, Dynamic Modelling and Flow in the Earth and Planets Series, 1993, pp. 175–195.
[61] E.B. Burov, P. Molnar, Gravity anomalies over the Ferghana Valley (central Asia) and intracontinental deformation, J. Geophys. Res. 103 (1998) 18137–18152.
[62] E.B. Burov, F. Houdry, M. Diament, J. Déverchère, Large-scale crustal heterogeneities and lithospheric strength in cratons, Earth Planet. Sci. Lett. 164 (1998) 205–219.
[63] S. Klemperer, R. Hobbs, The BIRPS Atlas: Deep Seismic Reflection Profiles around the British Isles, Cambridge University Press, Cambridge, 1991.
[64] A.B. Watts, S. Zhong, Observations of flexure and the rheology of oceanic lithosphere, Geophys. J. Int. 142 (2000) 855–875.
[65] J. Jackson, T. Blenkinsop, The Malawi earthquake of March 10, 1989: Deep faulting within the East African rift system, Tectonics 12 (1993) 1131–1139.
[66] L.L. Lavier, M.S. Steckler, The effect of sedimentary cover on the flexural strength of continental lithosphere, Nature 389 (1997) 476–479.
[67] A. Foster, J. Jackson, Source parameters of large African earthquakes: implications for crustal rheology and regional kinematics, Geophys. J. Int. 134 (1998) 422–448.
[68] M.B. Allen, S. Jones, A. Ismail-Zadah, M. Simmons, L. Anderson, Onset of subduction as the cause of rapid Pliocene-Quaternary subsidence in the south Caspian basin, Geology 30 (2002) 775–778.
[69] J. Grotzinger, L. Royden, Elastic strength of the Slave craton at 1.9 Gyr and implications for the thermal evolution of the continents, Nature 347 (1990) 64–66.
[70] P. Bird, Lateral extrusion of lower crust from under high topography, in the isostatic limit, J. Geophys. Res. 96 (1991) 10275–10286.
[71] J. Gratton, Crustal shortening, root spreading, isostasy, and the growth of orogenic belts: a dimensional analysis, J. Geophys. Res. 94 (1989) 15627–15634.
[72] J.P. Avouac, E.B. Burov, Erosion as a driving mechanism of intracontinental mountain growth?, J. Geophys. Res. 101 (1996) 17747–17769.
[73] C. Beaumont, S. Ellis, A. Pfiffner, Dynamics of sediment subduction-accretion at convergent margins: short-term modes, long-term deformation, and tectonic implications, J. Geophys. Res. 104 (1999) 17573–17602.
[74] S. Ellis, P. Fullsack, C. Beaumont, Oblique convergence of the crust driven by basal forcing: implications for length-scales of deformation and strain partitioning in orogens, Geophys. J. Int. 120 (1995) 24–44.
[75] G.E. Batt, J. Braun, On the thermomechanical evolution of compressional orogens, Geophys. J. Int. 128 (1997) 364–382.
[76] A.I. Chemenda, M. Mattauer, A.N. Bokun, Continental subduction and a mechanism for exhumation of high-pressure metamorphic rocks: New modeling and field data from Oman, Earth Planet. Sci. Lett. 143 (1996) 173–182.
[77] K. Lambeck, Structure and evolution of the intra-cratonic basins of central Australia, Geophys. J. R. Astron. Soc. 74 (1983) 843–886.
[78] S. Cloetingh, E. Burov, A. Poliakov, Lithosphere folding: primary response to compression? (from central Asia to Paris Basin), Tectonics 18 (1999) 1064–1083.
[79] P. Wessel, W.H.F. Smith, Free software helps map and display data, EOS Trans. AGU 72 (1991) 441–446.
[80] A.B. Watts, Isostasy and Flexure of the Lithosphere, Cambridge University Press, Cambridge, 2001, 458 pp.
[81] S.H. Kirby, A.K. Kronenberg, Rheology of the lithosphere: Selected topics, Rev. Geophys. 25 (1987) 1219–1244.
[82] B.E. Parsons, J.G. Sclater, An analysis of the variation of ocean floor bathymetry and heat flow with age, J. Geophys. Res. 82 (1977) 803–827.
[83] T. Seno, Y. Yamanaka, Double seismic zones, compressional deep trench-outer rise events, and superplumes, in: G.E. Bebout, D.W. Scholl, S.H. Kirby, J.P. Platt (Eds.), Subduction Top to Bottom, Monograph 36, AGU, Washington, DC, 1996, pp.347–355.
[84] D.W. Burbank, Causes of recent Himalayan uplift deduced from deposited patterns in the Ganges basin, Nature 357 (1992) 680–683.
[85] P.J. Waschbusch, L.H. Royden, Spatial and temporal evolution of foredeep basins: lateral strength variations

and inelastic yielding in continental lithosphere, Basin Res. 4 (1992) 179–196.
[86] J.M. Lorenzo, G.W. O'Brien, J. Stewart, K. Tandon, Inelastic yielding and forebulge shape across a modern foreland basin: North West Shelf of Australia, Timor Sea, Geophys. Res. Lett. 25 (1998) 1455–1458.
[87] N.L. Carter, M.C. Tsenn, Flow properties of continental lithosphere, Tectonophysics 136 (1987) 27–63.