

The long-term strength of continental lithosphere: “jelly sandwich” or “crème brûlée”?

E.B. Burov, *Laboratoire de Tectonique, University of Paris 6, 4 Place Jussieu, 75252 Paris Cedex 05, France, evgenii.burov@lgs.jussieu.fr*, and **A.B. Watts**, *Department of Earth Sciences, University of Oxford, Parks Road, Oxford OX1 3PR, UK, tony.watts@earth.ox.ac.uk*

ABSTRACT

There has been much debate recently concerning the long-term (i.e., >1 m.y.) strength of continental lithosphere. In one model, dubbed jelly sandwich, the strength resides in the crust and mantle, while in another, dubbed crème brûlée, the mantle is weak and the strength is limited to the crust. The different models have arisen because of conflicting results from elastic thickness and earthquake data. We address the problem here by first reviewing elastic thickness estimates and their relationship to the seismogenic layer thickness. We then explore, by numerical thermomechanical modeling, the implications of a weak and strong mantle for structural styles. We argue that, irrespective of the actual crustal strength, the crème-brûlée model is unable to explain either the persistence of mountain ranges or the integrity of the downgoing slab in collisional systems. We conclude that while the crème-brûlée model may apply in some tectonic settings, a more widely applicable model is the jelly sandwich.

INTRODUCTION

The strength of Earth's outermost layers has been a topic of debate ever since the turn of the last century when Joseph Barrell first introduced the concept of a strong lithosphere that overlies a weak fluid asthenosphere (Barrell, 1914). The concept played a major role in the development of plate tectonics (e.g., Le Pichon et al., 1973), and the question of how the strength of the plates varies spatially and temporally is a fundamental one of wide interest in geology.

One proxy for strength is the effective elastic thickness of the lithosphere, T_e (see Watts, 2001 and references therein). By comparing observations of flexure in the region of long-term loads such as ice, sediment, and volcanoes to the predictions of simple elastic plate (flexure) models, it has been possible to estimate T_e in a wide range of geological settings. Oceanic flexure studies suggest that T_e is in the range of 2–40 km and depends on load and plate age. In the continents, however, T_e ranges from 0 to 100 km and shows no clear relationship with age.

The results of flexure studies are qualitatively consistent with the results of experimental rock mechanics. The Brace-Goetze failure envelope curves (Goetze and Evans, 1979;

Brace and Kohlstedt, 1980), for example, predict that strength increases with depth and then decreases in accordance with the brittle (e.g., Byerlee) and ductile deformation laws. In oceanic regions, the envelopes are approximately symmetric about the depth of the brittle-ductile transition (BDT), where the brittle-elastic and elastic-ductile layers contribute equally to the strength. Since both T_e and the BDT generally exceed the mean thickness of the oceanic crust (~7 km), the largest contribution to the strength of oceanic lithosphere must come from the mantle, not the crust.

In the continents, the strength envelopes are more complex, and there may be more than one brittle and ductile layer. Despite this, Burov and Diament (1995) have been able to show that a model in which a weak lower crust is sandwiched between a strong brittle-elastic upper crust and an elastic-ductile mantle accounts for the wide range of T_e values observed due to the wide variation in composition, geothermal gradient, and crustal thickness possible in continental lithosphere.

Recently, Jackson (2002) challenged this so-called jelly sandwich model for the rheology of continental lithosphere. He stated the model was incorrect, proposing instead a model in which the upper crust is strong, but the mantle is weak. We dub this here the “crème-brûlée” model (Fig. 1). Jackson (2002) bases his model on the observations of Maggi et al. (2000), which suggest that earthquakes in the continents are restricted to a single layer (identified as the seismogenic thickness, T_s) in the upper brittle part of the crust and are either rare or absent in the underlying mantle.

It is well known that experimental rock mechanics data are based on relatively low temperatures and pressures and strain rates that are orders of magnitude greater than those that apply to the lithosphere (~ 10^{-6} – 10^{-4} s $^{-1}$ compared to ~ 10^{-17} – 10^{-13} s $^{-1}$). Hence, it is not really possible to use these data to distinguish between different rheological models (e.g., Rutter and Brodie, 1991). We therefore take a different approach. First, we review the T_e estimates because they, we believe, best reflect the integrated strength of the lithosphere. Then, we use numerical models to test the stability and structural styles associated with rheological models.

In order to focus the debate, we limit our study to the two rheological models considered by Jackson (2002): crème brûlée and jelly sandwich. This is not intended to exclude other models. We regard crème brûlée as including all models with a weak mantle and jelly sandwich as all models with a strong mantle, not just those with a weak lower crust. The effect of strong, rather than weak, lower crust depends on its strength with respect to the mantle. If the mantle is weaker than the lower crust, and the crust is not strong at the Moho depth, then the system is mechanically decoupled (e.g., upper panel, Fig. 1C) and we obtain a plate with a strength that is not significantly different from crème brûlée. In contrast, if the mantle is stronger than the lower crust, and the crust is strong at the Moho depth, then the system is coupled (e.g., lower panel, Fig. 1C), and this yields a plate that is capable of considerable strength.

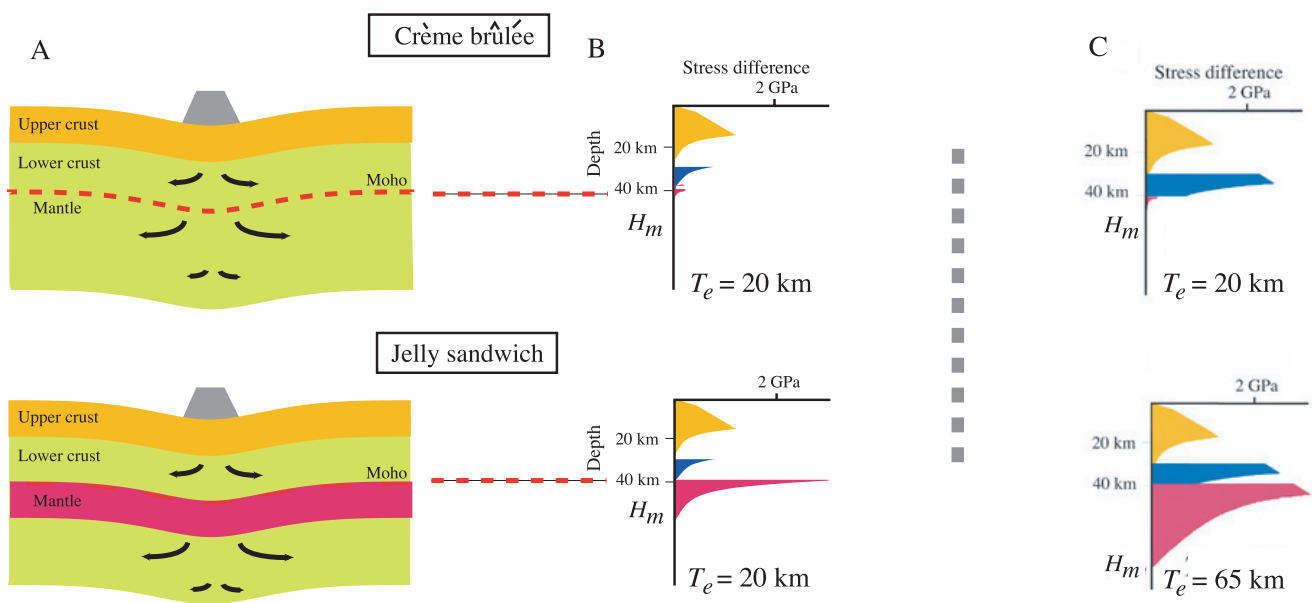


Figure 1. Schematic diagram illustrating different models for the long-term strength of continental lithosphere. In the crème-brûlée model, the strength is confined to the uppermost brittle layer of the crust, and compensation is achieved mainly by flow in the weak upper mantle. In the jelly sandwich model, the mantle is strong and the compensation for surface loads occurs mainly in the underlying asthenosphere. (A) Models of deformation. Arrows schematically show the velocity field of the flow. (B) Brace-Goetze failure envelopes for a thermotectonic age of 150 Ma, a weak, undried granulite lower crust, a uniform strain rate of 10^{-15} s^{-1} , and either a dry (jelly sandwich) or wet (crème brûlée) olivine mantle. H_m is the short-term mechanical thickness of the lithosphere; T_e is the long-term elastic thickness. Other parameters are as given in Tables 1 and 2. The two envelopes match those in Figures 5B and 5D of Jackson (2002). They yield a T_e of 20 km (e.g., Burov and Diament, 1995), which is similar to the thickness of the most competent layer. This is because the competent layers are mechanically decoupled by weak ductile layers and so the inclusion of a weak lower crust or strong mantle contributes little to T_e . (C) Brace-Goetze failure envelopes for a thermotectonic age of 500 Ma. Other parameters are as in (B) except that a strong, dry, Maryland diabase has been assumed for the lower crust. The two envelopes show other possible rheological models: in one, the upper and lower crusts are strong and the mantle is weak (upper panel); in the other, the upper and lower crusts and the mantle are strong (lower panel). The assumption of a strong lower crust in the weak mantle model again contributes little to T_e because of decoupling, although T_e would increase from 20 to 40 km if the upper crust was strong at its interface with the lower crust. In contrast, a strong lower crust contributes significantly to the T_e of the strong mantle model. This is because the lower crust is strong at its interface with the mantle and so the crust and mantle are mechanically coupled.

TABLE 1. SUMMARY OF THERMAL AND MECHANICAL PARAMETERS USED IN MODEL CALCULATIONS

Thermal	Surface temperature (0 km depth)	0 °C
	Temperature at base of thermal lithosphere	1330 °C
	Thermal conductivity of crust	$2.5 \text{ Wm}^{-1} \text{ °C}^{-1}$
	Thermal conductivity of mantle	$3.5 \text{ Wm}^{-1} \text{ °C}^{-1}$
	Thermal diffusivity of mantle	$10^{-6} \text{ m}^2 \text{ s}^{-1}$
	Radiogenic heat production at surface	$9.5 \times 10^{-10} \text{ W kg}^{-1}$
	Radiogenic heat production decay depth constant	10 km
	Thermo-tectonic age of the lithosphere	150 Ma (Fig. 1B); 500 Ma (Fig. 1C)
Mechanical	Density of upper crust	2700 kg m^{-3}
	Density of lower crust	2900 kg m^{-3}
	Density of mantle	3330 kg m^{-3}
	Density of asthenosphere	3310 kg m^{-3}
	Lamé elastic constants λ , G (here, $\lambda = G$)	30 GPa
	Byerlee's law—Friction angle	30°
	Byerlee's law—Cohesion	20 MPa

TABLE 2. SUMMARY OF DUCTILE PARAMETERS ASSUMED IN MODEL CALCULATIONS*

	Composition	Pre-exponential stress constant, A $\text{MPa}^{-n} \text{ s}^{-1}$	Power law exponent, n	Activation energy, Q KJ mol^{-1}	Figure 1
Upper crust	Wet quartzite	1.1×10^{-4}	4	223	B, C
Lower crust	Dry Maryland diabase	8 ± 4	4.7 ± 0.6	485 ± 30	C
	Undried Pikwitonei granulite	1.4×10^4	4.2	445	B
Mantle	Dry olivine	4.85×10^4	3.5	535	B, C (jelly sandwich)
	Wet olivine	417	4.48	498	B, C (crème brûlée)

*The failure envelopes in this paper match those in Jackson (2002); the Jackson (2002) envelopes are based on Figure 4 in Mackwell et al. (1998), who did not list all the parameters, referring instead to primary references. We therefore list here the parameters we have used.

GRAVITY ANOMALIES AND T_e

Gravity anomalies, especially their departures from local isostatic models (e.g., Airy, Pratt), have long played a key role in the debate concerning the strength of the lithosphere. Modern isostatic studies follow either a forward or inverse modeling approach. In forward modeling, the gravity anomaly due, for example, to a surface (i.e., topographic) load and its flexural compensation is calculated for different values of T_e and compared to the observed gravity anomaly. The best fit T_e is then determined as the one that minimizes the difference between observed and calculated gravity anomalies. In inverse (e.g., spectral) models, gravity and topography data are used to estimate T_e directly by computing the transfer function between them as a function of wavelength and comparing it to model predictions. The different approaches should yield the same results.

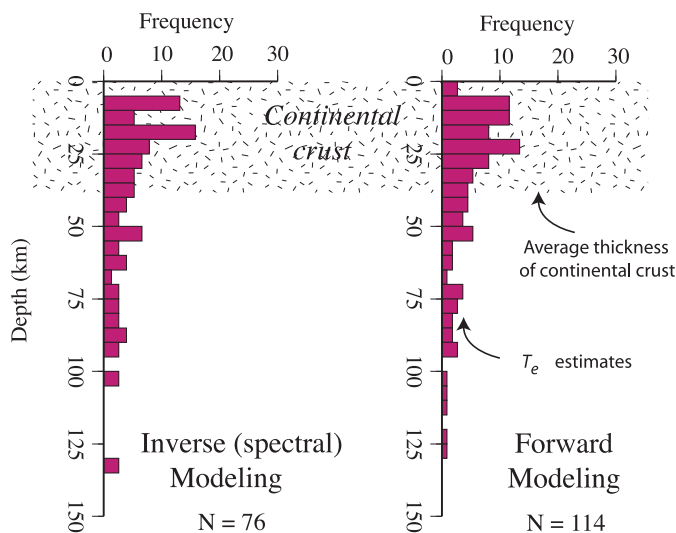


Figure 2. Histograms showing continental T_e estimates based on forward and inverse (i.e., spectral) gravity anomaly modeling methods. The histograms are based on data in Tables 5.2 and 6.2b of Watts (2001) and references therein. The data reflect a mix of tectonic settings. The spectral estimates mainly reflect old cratons, but include orogenic belts and rifts. The forward estimates are mainly from foreland basins. They reflect mainly rifts, since their mechanical properties are usually inherited during foreland basin formation, but include old cratons. The two modeling methods yield similar results and show that continental lithosphere is characterized by both low and high T_e values. In general, low values correlate with rifts, intermediate values with orogenic belts, and high values with old cratons. N—number of estimates.

In oceanic regions, forward and inverse models do yield similar T_e values. This is exemplified along the Hawaiian-Emperor seamount chain in the Pacific Ocean. Forward modeling reveals a mean T_e of 25 ± 9 km, while inverse (spectral) modeling using a free-air admittance method obtains 20–30 km (Watts, 1978). When the T_e estimates are plotted as a function of load and plate age, they yield the same result: T_e increases with age of the lithosphere at the time of loading, small (2–6 km) over young lithosphere and large over old lithosphere (>30 km).

In continental regions, the two modeling approaches have yielded different results. The earliest spectral studies, for example, recovered T_e values that were significantly smaller than those derived from forward modeling (see Cochran, 1980, and references therein). The subsequent development of more robust methods of determining T_e using a Bouguer coherence method (e.g., Forsyth, 1985; Lowry and Smith, 1994), however, has yielded values more compatible with forward modeling (Fig. 2).

Recently, McKenzie and Fairhead (1997) argued that most previous continental T_e estimates based on the Bouguer coherence (spectral) method are overestimates rather than true values. They used a free-air admittance method to argue that continental T_e was low, <25 km. Since their estimates of the elastic thickness are comparable to the seismogenic layer thickness, T_s , they proposed that the strength of the continental lithosphere resides in the crust, not the mantle.

Tests with synthetic and observed gravity anomaly and topography data show, however, that when the Bouguer coherence

and free-air admittance methods are similarly formulated, they yield the same results; namely, that continental T_e ranges from a few km to >70 km and that it varies spatially over relatively short (~100 km) horizontal scales (e.g., Pérez-Gussinyé and Watts, 2005).

While T_e values that exceed the local thickness of the crust do not indicate which layer, crust or mantle, is strong (Burov and Diament, 1995), they imply high mantle strength. For example, the Bouguer anomaly associated with the flexure of the Indian shield beneath the Ganges foreland basin by the load of the Himalaya requires a T_e of ~70 km (Fig. 3), which is significantly higher than the local crustal thickness of ~40 km. It is difficult using the Brace-Goetze failure envelopes to explain such high values without invoking a significant contribution to the strength from the subcrustal mantle.

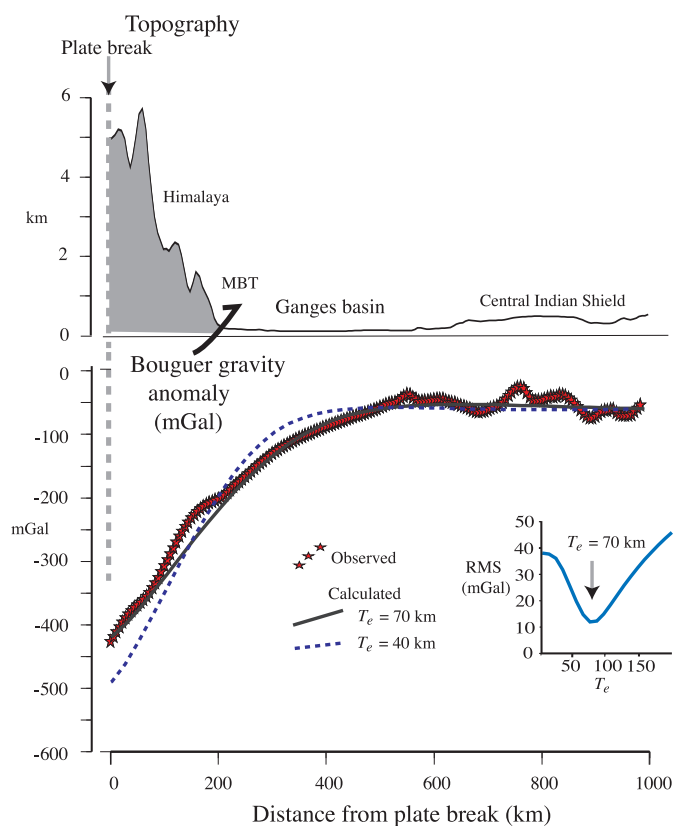


Figure 3. Comparison of observed and calculated Bouguer gravity anomalies along a profile of the Himalayan foreland at about longitude 80° E. The observed profile is based on GETECH (UK geophysical consultancy) Southeast Asia Gravity Project (SEAGP) gravity data. The calculated profile is based on a discontinuous (i.e., broken) elastic plate model, a load comprising the topography (density = 2650 kg m^{-3}) above sea level between the Main Boundary Thrust (MBT) and plate break (gray shaded region) and the material (density = 2650 kg m^{-3}) that infills the flexure (density = 2650 kg m^{-3}); a mantle density of 3330 kg m^{-3} ; and elastic thickness, T_e , of 70 km. Jordan and Watts (2005) have shown that this combination of load and elastic thickness parameters yields the best fit to the observed Bouguer anomaly. The dashed blue line shows the calculated Bouguer anomaly for the same load, but with $T_e = 40$ km, which is the value preferred by McKenzie and Fairhead (1997). The difference between the observed and calculated Bouguer anomaly for this T_e is, however, up to -70 mGal beneath the load and up to +30 mGal in flanking regions. The inset shows the root mean square (RMS) difference between the observed and calculated Bouguer anomaly for $0 < T_e < 180$ km.

Irrespective of the different methodologies, it is clear from Figure 2 that T_e can be high and may well exceed not only the seismogenic layer thickness, T_s (typically 10–20 km), but also the local crustal thickness. We are not surprised by this result. T_e reflects, we believe, the long-term integrated strength of the lithosphere, while T_s is representative of the strength of the uppermost of the crust on short time scales. It is difficult to use the failure envelopes to predict the actual depth to which seismicity should occur. However, if we assume that Byerlee's law is applicable to great depths, then the BDT increases from 15–25 km for the relatively low strain rates of flexure to 50–70 km for relatively high seismic strain rates. We attribute the general absence of mantle earthquakes at these latter depths to the lack of sufficiently large tectonic stresses (>2 GPa) to generate sliding.

SIMPLE PHYSICAL CONSIDERATIONS

The crème-brûlée and jelly sandwich models imply fundamental differences in the mechanical properties of mantle lithosphere. In the crème-brûlée model, for example, the mantle lithosphere is mechanically indistinguishable from asthenosphere, which suggests a very low viscosity. Here, we explore the stability of mantle lithosphere by posing the question “What do the different rheological models imply about the persistence of topography for long periods of geological time?”

The mean heat flow in Archean cratons is $\sim 40 \text{ mW m}^{-2}$, which increases to $\sim 60 \text{ mW m}^{-2}$ in flanking Phanerozoic orogenic belts (Jaupart and Mareschal, 1999). As Pinet et al. (1991) have shown, a significant part of this heat flow is derived from radiogenic sources in the crust. Therefore, temperatures at the Moho are relatively low ($\sim 400\text{--}600 \text{ }^\circ\text{C}$). The mantle must maintain a fixed, relatively high, viscosity that prevents convective heat advection to the Moho. Otherwise, surface heat flow would increase to $>150 \text{ mW m}^{-2}$, which would be the case in an actively extending rift (e.g., Sclater et al., 1980). Since heat flow this high is not observed in cratons and orogens, a thick, cool, stable mantle layer should remain that prevents direct contact between the crustal part of the lithosphere and the convective upper mantle.

The negative buoyancy of the mantle lithosphere at subduction zones is widely considered as a major driving force in plate tectonics. The evidence that the continental mantle is $\sim 20 \text{ kg m}^{-3}$ denser than the underlying asthenosphere and is gravitationally unstable has been reviewed by Stacey (1992), among others. Although this instability is commonly accepted for Phanerozoic lithosphere, there is still a debate about whether it applies to the presumably Mg-rich and depleted cratonic lithospheres. Irrespective, volumetric seismic velocities, which are generally considered a proxy for density, are systematically higher in the lithosphere mantle than in the asthenosphere. Depending on its viscosity, the mantle lithosphere therefore has the potential to sink as the result of a Rayleigh-Taylor (RT) instability (e.g., Houseman et al., 1981).

We can estimate the instability growth time (i.e., the time it takes for a mantle root to be amplified by e times its initial value) using Chandrasekhar's (1961) formulation. In this formulation, a mantle Newtonian fluid layer of viscosity, η , density, ρ_m , and thickness, d , is placed on top of a less dense fluid asthenospheric layer of density, ρ_a , and the same thickness.

(We note that this formulation differs from that of Conrad and Molnar [1997] who used a fluid layer that is placed on top of a viscous half-space. However, both formulations are valid for instability amplitudes $< d$). The most rapidly growing instability wavelength, λ , is Ad , where $2.5 < A < 3.0$ and the corresponding growth time, t_{min} , is $B\eta[(\rho_m - \rho_a)gd]^{-1}$, where $6.2 < B < 13.0$ and g is average gravity. We can evaluate t_{min} for a particular η by assuming $(\rho_m - \rho_a) = 20 \text{ kg m}^{-3}$ and $80 < d < 100 \text{ km}$. If the continental mantle can support large stresses (>2 GPa) and has a high viscosity ($10^{22}\text{--}10^{24} \text{ Pa s}$), as the jelly sandwich model implies, then t_{min} will be long ($>0.05\text{--}2 \text{ b.y.}$). If, on the other hand, the stresses are small (0–10 MPa) and the viscosity is low ($10^{19}\text{--}10^{20} \text{ Pa s}$), as the crème-brûlée model suggests, then it will be short (0.2–2.0 m.y.).

The consequence of these growth times for the persistence of surface topographic features and their compensating roots or anti-roots are profound. The long growth times in the jelly sandwich model imply that orogenic belts, for example, could persist for up to several tens of m.y. and longer, while the crème-brûlée model suggests collapse within a few m.y.

We have so far considered a Newtonian viscosity and a large viscosity contrast between the lithosphere and asthenosphere. However, a temperature-dependent viscosity and power law

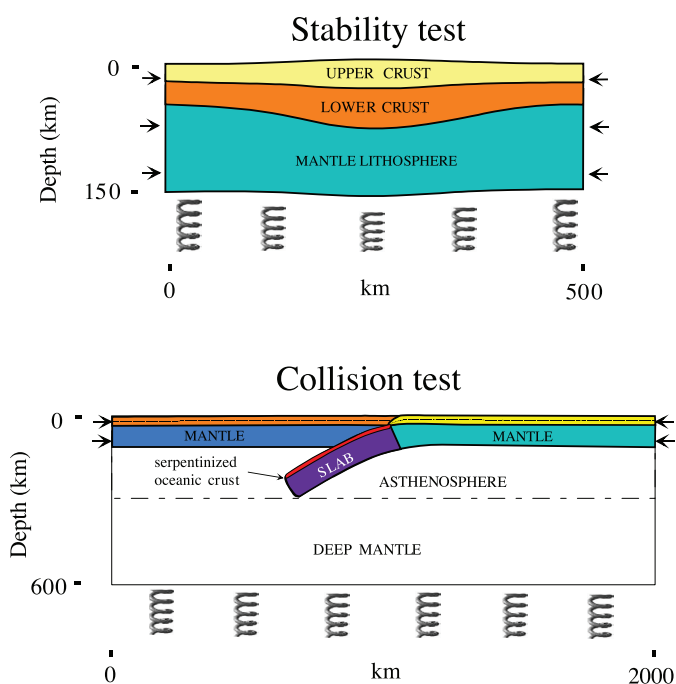


Figure 4. Numerical model set-up. The models assume a free upper surface and a hydrostatic boundary condition at the lower surface (depicted by springs in the figure). (A) The stability test is based on a mountain range height of 3 km and width of 200 km that is initially in isostatic equilibrium with a zero elevation 36-km-thick crust. We disturbed the isostatic balance by applying a horizontal compression to the edges of the lithosphere at a rate of 5 mm yr^{-1} . The displacements of both the surface topography and Moho were then tracked through time. (B) The collision test was based on a continent-continent collision initiated by subduction of a dense, downgoing, oceanic plate. We assumed a normal thickness oceanic crust of 7 km, a total convergence rate of 60 mm yr^{-1} , and a serpentinized subducted oceanic crust (Rupke et al., 2002). Rheological properties and other parameters are as given in Tables 1 and 2.

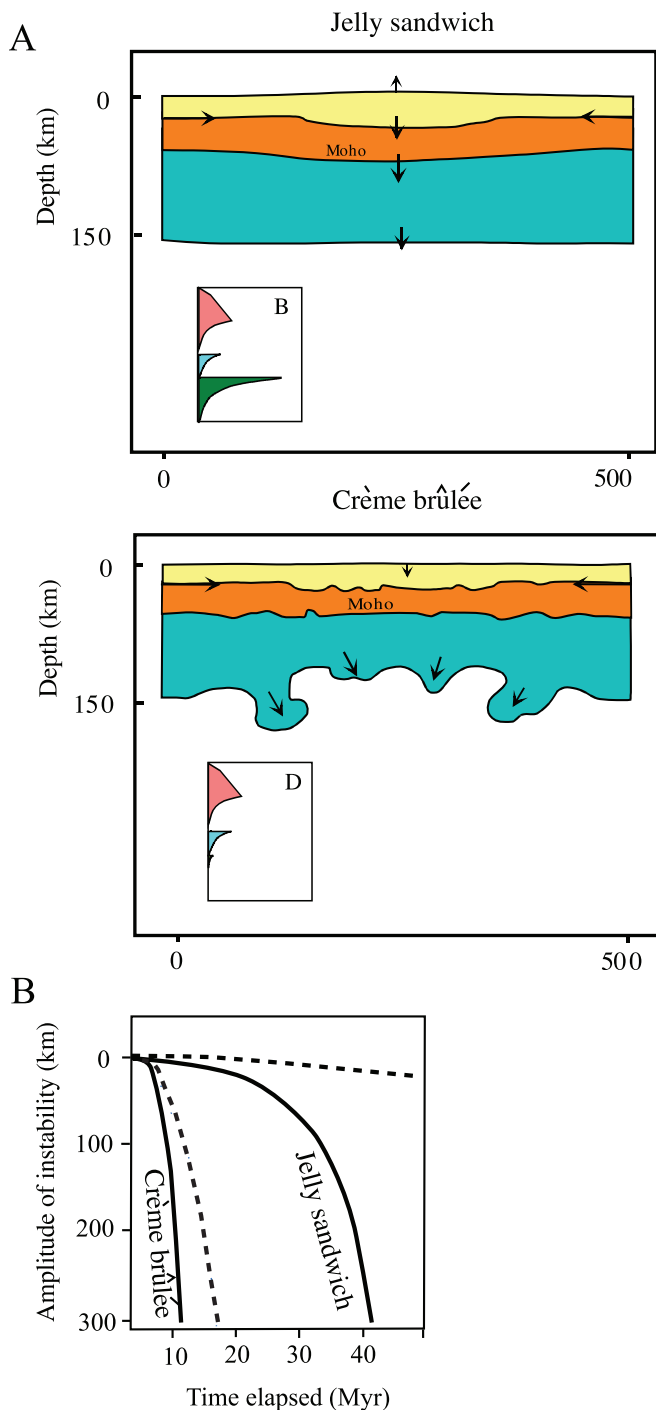


Figure 5. Tests of the stability of a mountain range using the failure envelopes associated with the jelly sandwich (Figure 5B of Jackson, 2002) and crème-brûlée (Figure 5D of Jackson, 2002) rheological models. The thermal structure is equivalent to that of a 150 Ma plate. (A) Crustal and mantle structure after 10 m.y. has elapsed. (B) The amplitude of the mantle root instability as a function of time. The figure shows the evolution of a marker that was initially positioned at the base of the mechanical lithosphere (i.e., the depth where the strength = 10 MPa). This initial position is assumed to be at 0 km on the vertical plot axis. The solid and dashed lines show the instability for a weak, young (thermotectonic age = 150 Ma) and strong, old (thermotectonic age = 500 Ma) plate, respectively.

rheology result in even shorter growth times than the ones derived here for constant viscosity (Conrad and Molnar, 1997; Molnar and Houseman, 2004). Moreover, if either the viscosity contrast is small or a mantle root starts to detach, then Equation (1) in Weinberg and Podladchikov (1995) suggests that the entire system will begin to collapse at a vertical Stokes flow velocity of $\sim 1 \text{ mm yr}^{-1}$ for the jelly sandwich model and $\sim 100\text{--}1000 \text{ mm yr}^{-1}$ for the crème-brûlée model. (We note that these flow velocities depend strongly on the sphere diameter, which we assume here to be λ). Therefore, our assumptions imply that a surface topographic feature such as an orogenic belt would disappear in $<0.02\text{--}2 \text{ m.y.}$ for the crème-brûlée model, whereas it could be supported for as long as 100 m.y.–2 b.y. for a jelly sandwich model.

DYNAMICAL MODELS

In order to substantiate the growth times of convective instabilities derived from simple viscous models, we carried out sensitivity tests using a numerical model that allows the equations of mechanical equilibrium for a viscoelasto-plastic plate to be solved for any prescribed rheological strength profile (Poliakov et al., 1993). Similar models have been used by Toussaint et al. (2004), for example, to determine the role that the geotherm, lower crustal composition, and metamorphic changes in the subducting crust may play on the evolution

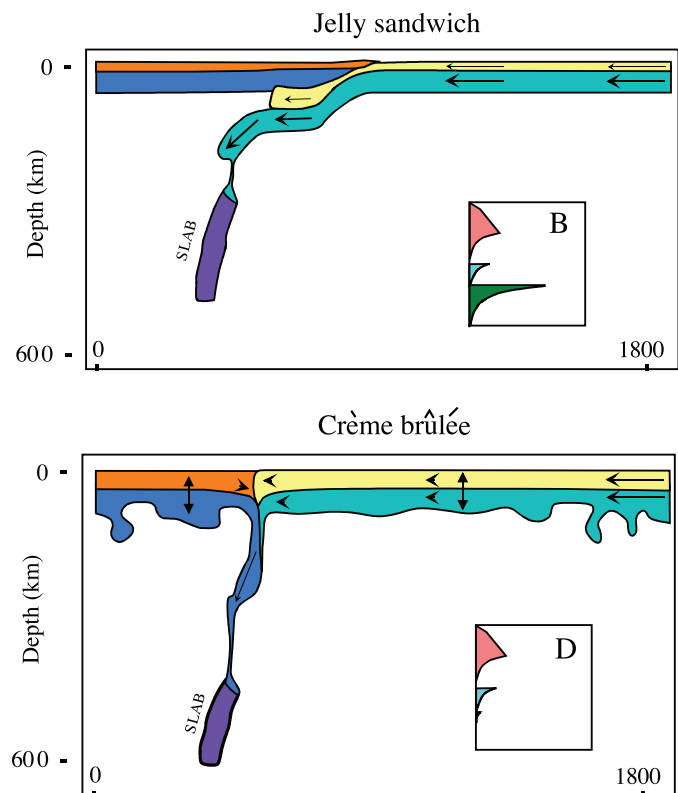


Figure 6. Tests of the stability of a continental collisional system using the failure envelopes associated with the jelly sandwich (B—Figure 5B of Jackson, 2002) and crème-brûlée (D—Figure 5D of Jackson, 2002) models. The elastic thickness, T_e , and Moho temperature are $\sim 20 \text{ km}$ and $600 \text{ }^\circ\text{C}$, respectively, for both models. The figure shows a snapshot at 5 Ma of the structural styles that develop after 300 km of shortening.

of continental collision zones. We ran two separate tests (Fig. 4) using rheological properties that matched the envelopes in Figures 5B and 5D of Jackson (2002). Our aim in using these envelopes, which otherwise yield a similar T_e (Fig. 1), was to determine what the crème-brûlée and jelly sandwich models imply about the stability of mountain ranges and the structural styles that develop.

Figure 5 shows the results of the stability tests. The figure shows a snapshot of the deformation after 10 m.y. We found that in the crème-brûlée model the crust and mantle already become unstable after 1.5–2.0 m.y. By 10 m.y., the lithosphere disintegrates due to delamination of the mantle followed by its convective removal and replacement with hot asthenosphere. This leads eventually to a flattening of the Moho and tectonic erosion of the crustal root that initially supported the topography. The jelly sandwich model, on the other hand, is more stable, and we found few signs of crust and mantle instability for the duration of the model run (10 m.y.).

Figure 6 shows the results of a collision test. The figure shows a snapshot of the deformation after 300 km of shortening, which at 60 mm yr⁻¹ takes 5 m.y. The jelly sandwich model is stable and subduction occurs by the underthrusting of a continental slab that, with or without the crust, maintains its overall shape. The crème-brûlée model, on the other hand, is unstable. There is no subduction, and convergence is taken up in the suture zone that separates the two plates. The crème-brûlée model is therefore unable to explain those features of collisional systems that require subduction such as kyanite- and sillimanite-grade metamorphism. The jelly sandwich model can explain not only the metamorphism, but also some of the gross structural styles of collisional systems such as those associated with slab flattening (e.g., Western North America—Humphreys et al., 2003), crustal doubling (e.g., Alps—Giese et al., 1982), and arc subduction (e.g., southern Tibet—Boutelier et al., 2003).

CONCLUSIONS

We have shown here that observations of flexure and the results of thermal and mechanical modeling are compatible with the view that the mantle part of the lithosphere is strong and is capable of supporting stresses (and surface and subsurface loads) for long periods of geological time.

Oceanic flexure studies show that T_e is high and that large loads such as oceanic islands and seamounts are supported, at least in part, by the subcrustal mantle. While the role of the mantle lithosphere in the continents is more difficult to quantify, there is evidence from cratonic regions and both forward and inverse (i.e., spectral) gravity modeling that T_e is high and can locally significantly exceed the crustal thickness.

We find no difficulty in reconciling the results of seismogenic layer, T_s , and elastic thickness, T_e , studies. While both parameters are proxies for the strength of the lithosphere, they are not the same. T_s reflects the thickness of the uppermost weak brittle layer that responds on historical time scales to stresses by faulting and earthquakes. T_e , in contrast, reflects the integrated strength of the entire lithosphere that responds to long-term (>10⁵ yr) geological loads by flexure.

There is almost certainly no one type of strength profile that characterizes all continental lithosphere. We have only tested two possible models in this paper. Nevertheless, they are rep-

resentative and useful. They are based on the same failure envelopes that Jackson (2002) used to argue that the mantle is weak, not strong. Moreover, they allow us to speculate on the stability of other models. The crème-brûlée model considered by Jackson (2002) yields a T_e of 20 km that is at the *high* end of the seismogenic layer thickness (typically 10–20 km). We have already shown that the crème-brûlée model is mechanically unstable. Therefore, weaker crème-brûlée models (e.g., ones with a weaker upper crust and $T_e < 20$ km) will be even more unstable. The jelly sandwich model of Jackson (2002) yields a T_e of 20 km that is at the *low* end of continental T_e estimates (which, as Fig. 2 shows, may exceed 70 km). We have already shown that this model is stable. Stronger jelly sandwich models (e.g., ones with a strong, coupled, lower crust and $T_e > 20$ km) will be even more stable. The wide range of continental T_e estimates suggest that while the crème-brûlée model may apply to some specific settings (e.g., young, hot, rifts such as parts of the Basin and Range, western USA; the Salton Sea, southern California; and Taupo volcanic zone, north island New Zealand), the jelly sandwich model, and its stronger variants, is more widely applicable (e.g., rifts, orogenic belts, cratons).

Thermomechanical modeling of lithospheric deformation suggests that the persistence of surface topographic features and their compensating roots require that the subcrustal mantle is strong and able to act as both a stress guide and a support for surface loads. It might be thought that it would not matter which competent layer in the lithosphere is the strong one. However, our tests show that the density contrast between the crust and mantle is sufficient to ensure that it is the mantle, rather than the crust, that provides both the stress guide and support. In our view, subduction and orogenesis *require* a strong mantle layer. We have found this to be true irrespective of the actual strength of the crust. Weak mantle is mechanically unstable and tends to delaminate from the overlying crust because it is unable to resist forces of tectonic origin. Once it does delaminate, hotter and lighter mantle asthenosphere can flow upward to the Moho. The resulting increase in Moho temperature would lead to extensive partial melting and magmatic activity as well as further weakening such that subduction is inhibited and surface topography collapses in a relatively short interval of time.

We conclude that rheological models such as crème brûlée that invoke a weak mantle are generally incompatible with observations. The jelly sandwich is in better agreement and provides a useful first-order explanation for the long-term support of Earth's main surface features.

ACKNOWLEDGMENTS

The kernel of the numerical code used in Figs. 5 and 6 is based on Parovoz (Poliakov et al., 1993). We thank P. Molnar, D. McKenzie, and S. Lamb for discussions, E. Humphreys, W. Royden, B. Kaus, W. Thatcher, and K. Howard for their constructive reviews, and the Dyeti and the University of Paris 6 visiting professor program for support.

REFERENCES CITED

- Barrell, J., 1914, The strength of the Earth's crust. I. Geologic tests of the limits of strength: *Journal of Geology*, v. 22, p. 28–48.
- Boutelier, D., Chemenda, A., and Burg, J.-P., 2003, Subduction versus accretion of intra-oceanic volcanic arcs: insight from thermo-mechanical analogue experiments: *Earth and Planetary Science Letters*, v. 212, p. 31–45, doi: 10.1016/S0012-821X(03)00239-5.
- Brace, W.F., and Kohlstedt, D.L., 1980, Limits on lithospheric stress imposed by laboratory experiments: *Journal of Geophysical Research*, v. 85, p. 6248–6252.

- Burov, E.B., and Diament, M., 1995, The effective elastic thickness (T_e) of continental lithosphere: What does it really mean?: *Journal of Geophysical Research*, v. 100, no. B3, p. 3905–3928, doi: 10.1029/94JB02770.
- Chandrasekhar, S., 1961, *Hydrodynamic and hydromagnetic stability*: Oxford, Oxford University Press, 704 p.
- Cochran, J.R., 1980, Some remarks on isostasy and the long-term behavior of the continental lithosphere: *Earth and Planetary Science Letters*, v. 46, p. 266–274, doi: 10.1016/0012-821X(80)90012-6
- Conrad, C.P., and Molnar, P., 1997, The growth of Rayleigh-Taylor-type instabilities in the lithosphere for various rheological and density structures: *Geophysical Journal International*, v. 129, p. 95–112.
- Forsyth, D.W., 1985, Subsurface loading and estimates of the flexural rigidity of continental lithosphere: *Journal of Geophysical Research*, v. 90, p. 12,623–12,632.
- Giese, P., Nicolich, R., and Reutter, K.-J., 1982, Explosion crustal seismic studies in the Alpine-Mediterranean region and their implications to tectonic processes, in Berckhemer, H., and Hsu, K.J., eds., *Alpine-Mediterranean geodynamics*: American Geophysical Union, p. 347–376.
- Goetze, C., and Evans, B., 1979, Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics: *Geophysical Journal of the Royal Astronomical Society*, v. 59, p. 463–478.
- Houseman, G.A., McKenzie, D.P., and Molnar, P., 1981, Convective instability of a thickened boundary layer and its relevance for the thermal evolution of continental convergent belts: *Journal of Geophysical Research*, v. 86, p. 6135–6155.
- Humphreys, E.D., Hessler, E., Dueker, K., Farmer, G.L., Erslov, E., and Atwater, T., 2003, How Laramide-age hydration of North American lithosphere by the Farallon slab controlled subsequent activity: *International Geology Review*, v. 45, p. 575–595.
- Jackson, J., 2002, Strength of the continental lithosphere: Time to abandon the jelly sandwich?: *GSA Today*, v. 12, no. 9, p. 4–10, doi: 10.1130/1052-5173(2002)012<0004:SOTCLT>2.0.CO;2.
- Jaupart, C., and Mareschal, J.C., 1999, The thermal structure and thickness of continental roots: *Lithos*, v. 48, p. 93–114, doi: 10.1016/S0024-4937(99)00023-7.
- Jordan, T.A., and Watts, A.B., 2005, Gravity anomalies, flexure and the elastic thickness structure of the India-Eurasia collisional system: *Earth and Planetary Science Letters*, v. 236, p. 732–750.
- Le Pichon, X., Francheteau, J., and Bonnin, J., 1973, *Plate tectonics*: Amsterdam, Elsevier Scientific, 300 p.
- Lowry, A.R., and Smith, R.B., 1994, Flexural rigidity of the Basin and Range–Colorado Plateau–Rocky Mountain transition from coherence analysis of gravity and topography: *Journal of Geophysical Research*, v. 99, p. 20,123–20,140, doi: 10.1029/94JB00960.
- Mackwell, S.J., Zimmerman, M.E., and Kohlstedt, D.L., 1998, High-temperature deformation of dry diabase with applications to tectonics on Venus: *Journal of Geophysical Research*, v. 103, p. 975–984, doi: 10.1029/97JB02671.
- Maggi, A., Jackson, J.A., Priestley, K., and Baker, C., 2000, A re-assessment of focal depth distributions in southern Iran, the Tien Shan and northern India: do earthquakes occur in the continental mantle?: *Geophysical Journal International*, v. 143, p. 629–661, doi: 10.1046/j.1365-246X.2000.00254.x.
- McKenzie, D.P., and Fairhead, D., 1997, Estimates of the effective elastic thickness of the continental lithosphere from Bouguer and free-air gravity anomalies: *Journal of Geophysical Research*, v. 102, p. 27,523–27,552, doi: 10.1029/97JB02481.
- Molnar, P., and Houseman, G.A., 2004, The effects of buoyant crust on the gravitational instability of thickened mantle lithosphere at zones of intracontinental convergence: *Geophysical Journal International*, v. 158, p. 1134–1150, doi: 10.1111/j.1365-246X.2004.02312.x.
- Pérez-Gussinyé, M., and Watts, A.B., 2005, The long-term strength of Europe and its implications for plate-forming processes: *Nature*, v. 436, p. 381–384, doi: 10.1038/nature03854.
- Pinet, C., Jaupart, C., Mareschal, J.-C., Gariépy, C., Bienfait, G., and Lapointe, R., 1991, Heat flow and lithospheric structure of the eastern Canadian shield: *Journal of Geophysical Research*, v. 96, p. 19,923–19,941.
- Poliakov, A.N.B., Cundall, P., Podladchikov, Y., and Laykhovskiy, V., 1993, 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 Stone, D.B., Runcorn, S.K., eds., *Flow and creep in the solar system: Observations, modeling and theory*: Kluwer, Holland, Dynamic Modeling and Flow in the Earth and Planets Series, p. 175–195.
- Rupke, L.H., Morgan, J.P., Hort, M., and Connolly, J.A.D., 2002, Are the regional variations in Central American arc lavas due to differing basaltic versus peridotitic slab sources of fluids?: *Geology*, v. 30, p. 1035–1038, doi: 10.1130/0091-7613(2002)030<1035:ATRVIC>2.0.CO;2.
- Rutter, E.H., and Brodie, K.H., 1991, Lithosphere rheology—a note of caution: *Journal of Structural Geology*, v. 13, p. 363–367, doi: 10.1016/0191-8141(91)90136-7.
- Sclater, J.G., Jaupart, C., and Galson, D., 1980, The heat flow through oceanic and continental crust and the heat loss of the Earth: *Journal of Geophysical Research*, v. 85, p. 269–311.
- Stacey, F.D., 1992, *Physics of the Earth*: Brisbane, Brookfield Press, 513 p.
- Toussaint, G., Burov, E., and Jolivet, L., 2004, Continental plate collision: Unstable vs. stable slab dynamics: *Geology*, v. 32, p. 33–36, doi: 10.1130/G19883.1.
- Watts, A.B., 1978, An analysis of isostasy in the world's oceans—1. Hawaiian-Emperor Seamount Chain: *Journal of Geophysical Research*, v. 83, p. 5989–6004.
- Watts, A.B., 2001, *Isostasy and flexure of the lithosphere*: Cambridge, Cambridge University Press, 458 p.
- Weinberg, R.F., and Podladchikov, Y., 1995, The rise of solid state diapirs: *Journal of Structural Geology*, v. 17, p. 1183–1195, doi: 10.1016/0191-8141(95)00004-W.

Manuscript received 16 June 2005; accepted 28 Oct. 2005.

TWO NEW POSITION STATEMENTS ADOPTED BY GSA COUNCIL

Geoscience Data Preservation

Contributor:

Jamie Robertson, Chair

Panel Members:

Odin Christensen, Linda Gunderson,
Christopher Keane, Emi Ito, Robert Schafer,
Marvin Carlson, Kerstin Lehnert, and Warren Allmon

Position Statement:

The Geological Society of America supports the preservation of geoscience samples and data sets for the public good and urges public and private sector organizations and individuals to routinely catalog and preserve their collections and make them more widely accessible.

Adopted by GSA Council October 2005.

Background information and recommendations for implementation of this policy are detailed at www.geosociety.org/aboutus/position9.htm.

Geoscience and Natural Hazards Policy

Contributors:

Lou Gilpin, George Linkletter, Co-Chairs

Panel Members:

David Applegate, Vic Baker, Susan Cannon,
John Costa, Orrin Pilkey

Position Statement:

The Geological Society of America (GSA) urges the public, scientists, and policy makers to work together to reduce our vulnerability to natural hazards.

Adopted by GSA Council October 2005.

Background information and recommendations for implementation of this policy are detailed at www.geosociety.org/aboutus/position6.htm.



SCIENCE ■ STEWARDSHIP ■ SERVICE

To read GSA's position statements (2001 to present) and to find out more about how to participate in supporting and implementing these position statements, check out the geology and public policy section of GSA's Web site, www.geosociety.org/science/govpolicy.htm.

The most effective way to influence the success of GSA's geology and public policy efforts is to **communicate** among your colleagues and scientists in related professions, to students and members of the public and media, and with your congressional representatives.