LETTERS

The effect of energy feedbacks on continental strength

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The classical strength profile of continents[1,2] is derived from a quasi-static view of their rheological response to stress—one that does not consider dynamic interactions between brittle and ductile layers. Such interactions result in complexities of failure in the brittle–ductile transition and the need to couple energy to understand strain localization. Here we investigate continental deformation by solving the fully coupled energy, momentum and continuum equations. We show that this approach produces unexpected feedback processes, leading to a significantly weaker dynamic strength evolution. In our model, stress localization focused on the brittle–ductile transition leads to the spontaneous development of mid-crustal detachment faults immediately above the strongest crustal layer. We also find that an additional decoupling layer forms between the lower crust and mantle. Our results explain the development of decoupling layers that are observed to accommodate hundreds of kilometres of horizontal motions during continental deformation.

The strength profile of the continental lithosphere is understood to be a combination of the pressure-dependent brittle strength of the upper crust (Byerlee's law), and the viscous, temperature-dependent and strain-rate-dependent strength of the lower continental crust, dominated by either quartz or plagioclase rheology[3]. The brittle–ductile transition is typically defined for a given characteristic geological strain rate (often $10^{-15}\,s^{-1}$) as the depth where the two strength curves intersect. As temperature increases with depth, the lower crust becomes increasingly weaker, until reaching the mantle, which is dominated by the generally stronger olivine rheology. This is known as the Brace-Goetze strength profile[1–3], or informally as the 'Christmas tree' (Fig. 1a). The weak lower crust sandwiched between the strong brittle–ductile transition area and the strong mantle defines what is known as the 'jelly sandwich'. This view is simple and powerful, and has led us a long way towards understanding the deformation of continents in the context of plate tectonics.

The 'Christmas tree' strength profile, however, implies that the strongest part of the crust is located at the layers immediately above the brittle–ductile transition (Fig. 1a). This raises a major problem in continental tectonics, because field evidence shows strain localization and the development of major detachment faults at typical depths of the brittle–ductile transition[4–8]. The reason for this apparent paradox may be that existing models have not considered the dynamic evolution of the strength profile due to loading and the history of deformation and strain localization.

In order to evaluate the rheological response to energy feedback effects, we designed numerical models of extension with an initial crustal thickness of 42 km. We assume a free top surface and zero tangential stress (free slip) on other boundaries. Extension is driven by velocity boundary conditions of $1\,cm\,yr^{-1}$ applied on either side of the model. Implementing such boundary conditions using the classical Mohr-Coulomb approach[9] produces a pure shear style of extension, which is accommodated by steeply dipping normal faults and fails to predict the development of low-angle normal faults. However, in our fully coupled simulation (see Methods and Supplementary Appendix 1), localization feedbacks develop in and between the brittle and ductile layers (Fig. 2). These are expressed by the development of listric faulting and detachment faults in the brittle upper crust and ductile shearing in an elasto-viscoplastic lower crust. In this process, brittle faulting may rupture at seismogenic rates (for example, 10^2–$10^3\,m\,s^{-1}$), whereas shear zones propagate at much slower rates (up to $3\times10^{-9}\,m\,s^{-1}$)[10]. This contrast in strain rates leads to complex interactions at the brittle–ductile transition[11].

Complex structural evolution, related to dynamic changes in strength, is recognized in the extensional models in Fig. 3. During the early stages of loading (Fig. 3a), two distinct depth levels develop sub-horizontal high strain decoupling segments—a deeper and a shallower one. The deep decoupling zone is rooted in the lowermost

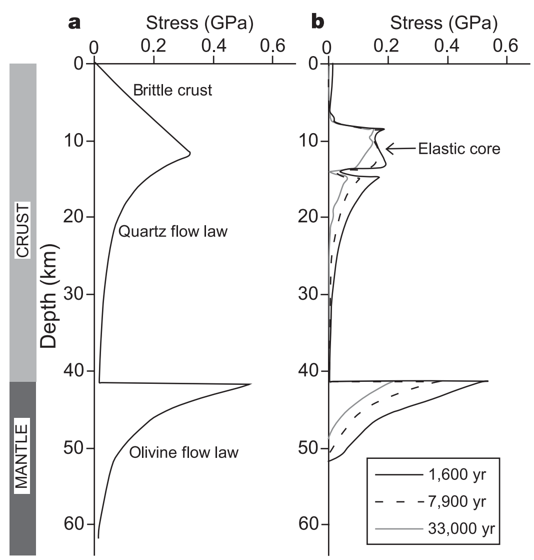

Figure 1 | Strength profiles of the lithosphere. a, Simplified Brace-Goetze strength profile for parameters in Table 1 in Supplementary Appendix 1. **b**, An example of early stress evolution for the dynamic feedback calculation, showing the development of an elastic core that is gradually eroded by the dynamic weakening in the upper mantle and the lower crust (for a more evolved stress profile, see Supplementary Appendix 3). Note that the brittle crust has failed while the elastic core retains its strength. The pronounced weak zone at ~14 km is developed owing to ductile localization feedback effects.

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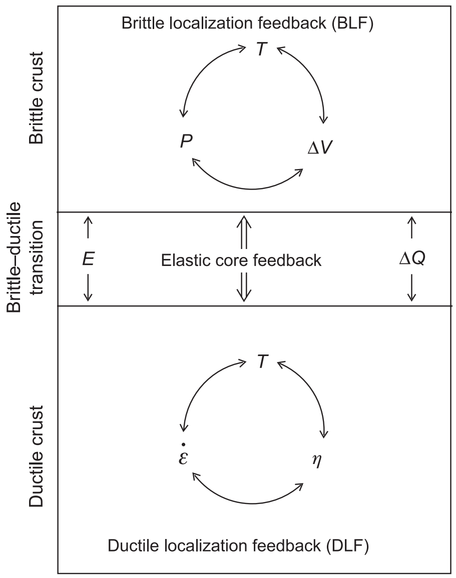

Figure 2 | Strain localization feedbacks in upper and lower crust. The brittle localization feedback (BLF) involves pressure (P), temperature (T) and volume change (ΔV) and gives rise to brittle faults in upper crust. The ductile localization feedback (DLF) gives rise to ductile shear zones in lower crust, and also includes components of viscosity (η), strain rate ($\dot{\varepsilon}$) and temperature (T), dampened by heat diffusion. The two layers communicate by changes of stored elastic energy (E) and heat (ΔQ) in the strong mid-crustal elastic core, which corresponds to a brittle–ductile transition zone. The depth of this transition is a function of P, T, t, strain rate, shear modulus, Poisson ratio and activation energy.

crust on top of the mantle. It can be attributed to rheological contrasts between quartz and olivine. This contrast is amplified by a thermo-mechanical ductile localization feedback (DLF in Fig. 2), resulting in an effective viscosity drop just above the Moho, which increases with deformation (Supplementary Appendix 3). In the mantle, small-scale conjugate pairs of shear instabilities nucleate on a short wavelength (of the order of 100 m) around thermal perturbations. These are ephemeral structures that gradually cascade to larger wavelengths. This leads to shear focusing onto a number of master shear zones (Fig. 3b). The net effect of these shear instabilities is that the mantle evolves dynamically to a weak rheology with shear stress less than 0.3 GPa (Fig. 1b). Quite opposite to the prediction of Fig. 1a, the mantle is not the strongest layer but evolves to be weaker than the upper crust (Fig. 1b and Supplementary Appendix 3).

The efficient weakening of the mantle layer relies on the development of strong shear zones reaching from the deep crust decoupling layer into the mantle. Conversely, shear zones with an upward continuation from the deep decoupling layer into shallower crust are diffuse and die out into an elastic core. This core is defined as the region of maximum elastic energy storage in the crust. It appears initially without shear zones and deforms pervasively at a slower rate than the localized shear zones in the upper and lower crusts (Supplementary Appendix 3). Above the elastic core, deformation is governed by the brittle localization feedback (BLF; Fig. 2), which forms conjugate faults (Fig. 3a). Like in the mantle, ductile shear zones in the lower crust collocate into the largest possible structures and wavelength (Fig. 3b). In the upper crust, faults gradually become listric, forming a detachment layer of maximum shear heating (that is, maximum dissipation) on top of the elastic core (Fig. 3b).

The complex deformation of the elastic core, dominated simultaneously by brittle and ductile strain localization, is the key to understanding the deformation of the lithosphere in general and the communication between the upper and lower crust in particular. This core is characterized not only by the maximum elastic energy in the crust, but also by having the lowest ratio between Young's modulus and yield strength. This leads to a competition between storing elastic energy and dissipating heat, which is translated into the development of contemporaneous short-term, rapid, brittle cracks localized on pressure anomalies, and short-term, slow, ductile shear zones on temperature anomalies. The high stored energy in this area allows many dynamic shear zones to develop. However, they die out because their size is below the critical length scale for survival, defined by the equilibration between shear heating and heat diffusion (of the order of kilometres, equivalent to the square root of the ratio between heat diffusivity and strain rate in the shear zone[12]).

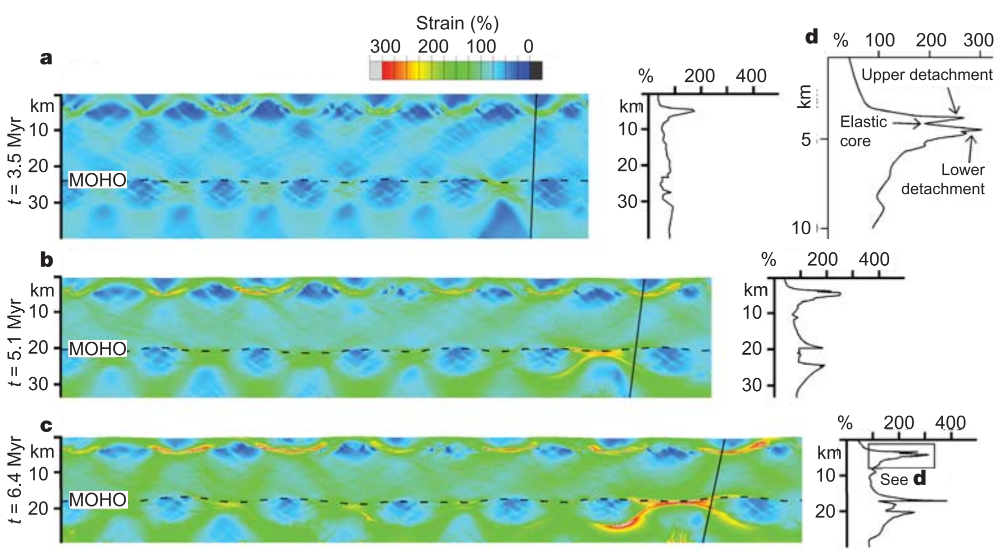

Figure 3 | Model results. a–c, Results of the 70 mW m^{-2} heat flow extension model using parameters from Table 1 in Supplementary Appendix 1. Additional results for 60 mW m^{-2} are shown in Supplementary Appendix 2. The coloured panels show vertical cross-sections through the lithosphere plotted on a 1:1 x:y spatial scale. Note the development of mid-crustal detachments and an additional decoupling zone on top of the Moho. The strain profile on the right of each panel is plotted for a straight line, which is initially vertical (at 0 Myr) and tilts significantly owing to differential deformation of crust and mantle, causing simple shear out of pure shear boundary conditions. **d**, Enlargement of the strain profile at 6.4 Myr, showing the development of two upper crustal detachment levels.

The propagation of the lower tip of a brittle fault is driven by the combination of loading and brittle localization feedback effects. Fault propagation is hampered as it reaches the top of the elastic core, where brittle and ductile processes interact, because it encounters a viscous response that blunts the fault tip and eliminates the stress singularity. Further arrest of the brittle fault is driven by ductile thermal-mechanical feedback effects, which act to decrease the differential stress at the fault's tip by lowering the effective viscosity[11]. The softening occurs at a subhorizontal layer, transforming the fault into a heat source and a detachment defined by a mylonitic shear zone.

Simultaneously, ductile shear zones propagate upwards towards the elastic core. However, these shear zones die out because of increasing crustal strength and because they reach the area in which brittle localization feedback dominates. Their energy is translated into a local, positive amplitude disturbance of the elastic stress field. This extra stress, loaded to the core from below, eventually starts to interact with similar features loaded from above. When they coalesce, the core is disrupted leading to a second, deeper detachment rooted at the base of the mid-crustal core. In this way, the upper mid-crustal detachment zone develops into a 1.5-km-wide low viscosity zone defined by two detachments separated by an elastic core (Fig. 3c, d). This detachment zone is located immediately at or near the strongest part of the crust, and coincides with a zone of low effective viscosity (Supplementary Appendix 3). Significant mantle uplift develops below the detachment, while the surface topography remains relatively moderate (570 m maximum amplitude, see Supplementary Appendix 4). The dynamic evolution of the brittle–ductile transition is controlled by three significant aspects of evolution: (1) the apparent weakness of the uppermost crust reflects the fact that stress in this layer is released on faults; (2) similarly, the gradual weakening of the uppermost mantle controls the development of collocated, wide ductile shear zones (Fig. 3b, c); and (3) development and interaction of the detachment zones above and below the narrowing elastic core, which ultimately breaks up, leading to a lithospheric scale continental break-up. Quite contrary to the quasi-static prediction of Fig. 1, it is not the mantle but the mid-crustal brittle–ductile transition which dynamically evolves into the stress-bearing part of the lithosphere for most of its deformation history.

Thus, in these models the lithosphere evolves through several steps as stress is loaded and strain is localized: first, crust and mantle decouple, followed by the weakening of the mantle through the development of shear zones. The mid-crustal detachment then develops at the top of the elastic core. The gradual weakening of the elastic core through loading then leads to the second ductile detachment at the base of the core, followed by crust-wide shear zones which lead ultimately to lithospheric break-up. In models with a lower thermal gradient, as in the 60 $\mathrm{mW\,m^{-2}}$ run, the strong elastic core is initially wider, and consequently the mid-crustal detachment has a similar but yet much richer evolution (Supplementary Appendix 2).

Crustal-scale detachments or décollements are characteristic features of thin-skinned terranes, accommodating large relative displacements between allochthonous, brittle upper slabs, and ductilely deformed lower slabs. Detachments have been documented in extensional environments (for example, Basin and Range[4,5,8] and the Aegean Sea[13]), and in shortening environments (for example, Helvetic nappes in the Alps[14] and Canadian Cordillera[15]). The extensional detachments were originally thought to penetrate the whole crust to the Moho[16]. However, further investigation showed that they tend to root at the brittle–ductile transition, at crustal depths around 10 ± 5 km (refs 4–8, 17). Similarly, convergent terranes have well-established décollements at mid-crustal levels[15] and at the base of the crust[18].

In contrast with observations, the Brace-Goetze strength profile predicts that the brittle–ductile transition is the strongest part of the crust and therefore the least likely part to develop a detachment. This paradox has commonly been resolved by postulating that detachments develop on zones of crustal weakness, such as evaporites[19], shale and marl[14], or low-viscosity zones related to rheological stratification[20]. Although weak layers could plausibly initiate detachment faults, as demonstrated by the Moho decoupling in Fig. 3b and c, our calculations demonstrate that they are unnecessary. Mid-crustal detachments arise spontaneously from thermo-mechanical feedback effects between the elasto-viscoplastic and brittle layers, even in a homogeneous crustal medium. Furthermore, the results suggest that crustal detachment may be followed by steeply dipping lithosphere-wide shear zones, which ultimately enables lithospheric break-up[21].

Decoupling of crust and mantle across the Moho is a much discussed issue[3,20,22,23]. However, despite the usage of strong rheological contrasts across the Moho, the development of this decoupling layer has not been recognized in numerical models. Our results illustrate the importance of thermo-mechanical feedback effects for the spontaneous development of this decoupling layer.

The evolution of the lithospheric strength profile, as shown by our model, also resolves a paradox mentioned in ref. 24 with regard to the aseismic behaviour of the upper mantle. According to the 'jelly sandwich' strength model of the continental lithosphere, the upper mantle is expected to be strong and produce earthquakes upon loading. The strong upper mantle should also be found in flexural rigidity analyses of continents. Nonetheless, data show that earthquakes do not occur in the continental upper mantle, and that the latter is commonly not reflected in flexural rigidity analyses. This apparent paradox is resolved in our models, which demonstrate that the mantle evolves from strong to weak as ductile instabilities develop and stabilize in the olivine layer. Although our models do not progress into active seismic events, there is a strong propensity of the quartz layer to develop fast ductile slip reminiscent of slow earthquakes[25], whereas the olivine layer has the tendency to deform by stable ductile flow.

The energy approach to modelling the evolution of the continental lithosphere shows that its strength profile differs significantly from that derived from the classical quasi-static 'Christmas tree' constructs. This difference hinges on thermo-mechanical feedback effects. Out of a most simple quartz-olivine composite lithospheric composition, the models develop natural features, such as the thickness of the crustal seismogenic zone, mid-crustal detachments at the otherwise strongest crustal layer, lower crustal channel flow above the crust–mantle decoupling zone, and lack of flexural strength or seismicity in the continental upper mantle[10].

METHODS

Model overview. The approach used here is to solve the fully coupled continuum, momentum and energy equations (Supplementary Appendix 1). This differs from traditional approaches, where the energy equation is not fully coupled or is neglected. We solve the equations for equilibrium and seek the minimum value of free energy[26], using an implicit adaptive time step technique[26,27].

Energy equation. In order to understand the temporal evolution before and after the initiation of failure, the dissipation of the system must be solved. The master equation that provides the evolution of the dissipation function is the energy equation, with its time derivative. By coupling the energy equation we are able to study the self-consistent history of strain localization through either the brittle or the ductile localization feedback. The strength of our method is that it avoids prescribing shear localization. By doing so, the entropy of the system and the second law of thermodynamics are considered. This contrasts with prescriptive methods, such as the Mohr-Coulomb approach, where, for instance, in order to stabilize shear zones in numerical models, shear weakening after failure is imposed, with no account of the second law of thermodynamics, which it could be violating.

Modelling brittle and ductile behaviour. Brittle behaviour in our model differs from classical approaches, which use non-coaxiality in stress and strain rate tensors (non-associative behaviour) by prescribing the dilatancy angle. This implies that localization in itself is prescribed rather than arises spontaneously, and also implies that failure will take place in preferred planes with prescribed angles. This approach, which was set in the mid-1970s[28,29], only considers the phenomenon of localization and is not tailored for describing further evolution of deforming systems after failure. For this, energy, and consequently entropy, needs to be considered. In our approach, the brittle regime is solved using standard pressure-dependent yield strength, without referring to the prescriptive Mohr-Coulomb rheology. Instead we use a natural localization phenomenon that results from the full feedback between thermal expansion and (recoverable) temperature changes related to pressure variations (isentropic work; see ref. 30, p. 8).

In our calculations, noise-level strength heterogeneities lead to local thermal expansion due to focusing of work. This leads to a pressure pulse, which further creates thermal expansion and leads to a local departure from coaxiality, bringing the material locally above the yield stress and spontaneous localization of shearing. Our system is thus associative, where pressure controls yielding and thermal expansion controls localization. This approach is particularly faithful to semi-brittle conditions at the brittle–ductile transition, where the dilatancy angle is negligible due to the overburden pressure, and the fracture angle is 45° in plane strain. However, this approach would require specific coupling to the energy evolution of dilatant fractures to appropriately model near surface processes. Hence, our models produce a smoother topography than expected in a natural system.

Ductile deformation also tends to localize into ductile shear zones as a result of feedback between shear heating and thermal softening (Fig. 2). In order to calculate the composite brittle and ductile rheology everywhere, we use the additive strain rate decomposition where temperature- and pressure-dependent strain rates are added. The viscous constitutive law used is power-law. For simplicity, we assume that any strain rates below $10^{-16}\,s^{-1}$ are elastic.

Received 10 November 2005; accepted 5 May 2006.

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Acknowledgements We thank V. Lyakhovsky for comments on the manuscript. K.R.-L. acknowledges support from the Johannes Gutenberg-University Mainz and the Predictive Mineral Discovery Cooperative Research Centre.

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