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Strain localisation and weakening of the lithosphere during extension

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ABSTRACT

We explore the sensitivity of extensional systems to the thermal structure of the lithosphere using numerical simulations that fully couple the energy, momentum and continuity equations. The rheology of the lithosphere is controlled by weakening processes, such as shear heating, that localises strain into shear zones and faults. Numerical models show that during extension of an initially unpatterned lithosphere, structures develop spontaneously out of basic thermodynamic energy fluxes, and without the imposition of *ad hoc* rules on strain localisation. This contrasts with the classical Mohr–Coulomb theory for brittle localisation, which prescribes the angles of faults by a mathematical rule. Our results show that the mode of extension is sensitive to subtle changes in rheology, heat flux and geometry of the system. This sensitivity lies at the core of the variety and complexity observed in extensional systems. Localisation processes make the lithosphere weaker than previously estimated from the Brace–Goetze quasi-static approach. Consequently, typical estimates for plate tectonic forces are capable of splitting the lithosphere under extension, even without the role of active magmatism.

1. Introduction

The strength of the lithosphere is one of the most fundamental parameters for understanding lithospheric extension. Yet our understanding of lithospheric strength is relatively rudimentary, and descriptions of the rheology of the lithosphere are a matter of debate. The current modelling paradigm is based on the concept of the Brace–Goetze strength profile (Goetze and Evans, 1979; Brace and Kohlstedt, 1980), informally known as the 'Christmas Tree'. It assumes a combination of frictional sliding and creep mechanisms, and is typically characterised by a weak lower crust sandwiched between the strong brittle–ductile transition area and the strong mantle (dashed curve in Fig. 1). The Brace–Goetze profile provides a quasi-static view of lithospheric strength and does not consider the role of weakening processes that localise strain. Consequently, Brace–Goetze profiles or other simplifications usually yield over-estimates of the strength of the lithosphere (Kusznir and Park, 1984; Ranalli and Murphy, 1987; Buck, 2006).

In this paper we describe an alternative approach for estimating lithospheric strength by considering energy fluxes in a system that fully couples the energy, momentum and continuity equations. Such an approach takes into account brittle and ductile energy feedback effects, which arise from temperature and pressure perturbations, and can lead to runaway processes of strain localisation (Regenauer-Lieb and Yuen, 2003; Regenauer-Lieb et al., 2006). We show that the strength of the lithosphere during extension is subjected to temporal and spatial changes. This allows the development of elastic zones bounded by weak zones, simulating typical core complex and detachment structures recognised in extensional environments (Rosenbaum et al., 2005; Regenauer-Lieb et al., 2006; Weinberg et al., 2007).

2. The quasi-static Brace–Goetze strength profile of the lithosphere

The classic approach to estimating the strength of the lithosphere, known as the Brace–Goetze profile, relies on a combination of pressure-dependent brittle layer overlying a temperature- and strain rate-dependent viscous layer (Goetze and Evans, 1979; Kohlstedt et al., 1995). The strength of the brittle layer is defined by Byerlee's (1978) law, which assumes a linear relationship between the shear strength of rocks and the pressure, with a constant friction coefficient. The latter is commonly assumed to be independent of temperature and material. The strength of the ductile layer is derived from steady-state creep experiments (Ranalli, 1995), and is most commonly described by a power law relationship between the stress and the strain rate. Commonly, a fixed strain rate is assumed (typically 10^{-15}–10^{-14} s^{-1}), allowing the definition of the strength of the whole lithosphere as an integral quantity.

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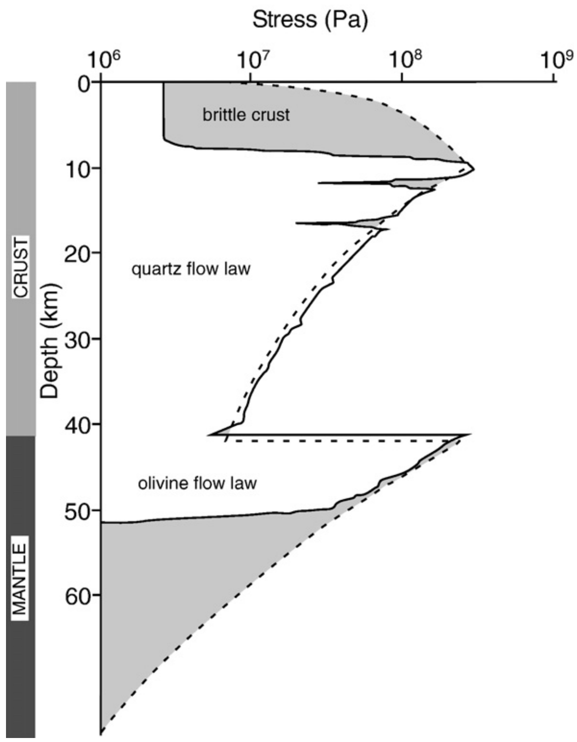

Fig. 1. A classical Brace–Goetze ("Christmas Tree") yield strength envelope (dashed line) shown against a dynamic elasto-visco-plastic strength profile (black line) (after Regenauer-Lieb et al., 2006). The dynamic profile evolves through time and is calculated here shortly after the onset of extension (130 kyr). The lithosphere is made up of a 42 km crust (quartz rheology) overlying an olivine mantle. The dynamic strength profile is not an average value for each depth, but is shown for a single 1D vertical line that intersects with two detachment faults in the 10–20 km depth range. Note that the energy feedbacks of the top purely brittle layer (down to ~2 km depths) are not considered, because at shallow depths, the surface energy of cracks, which is not incorporated in the current numerical model, becomes increasingly important. The shape of this profile changes from place to place and in time. Our particular choice of 1D profile represents one of the weakest profiles in the cross section and shows minimum strength values for the top brittle crust and for the area below 52 km, due to unloading of stresses by nearby shear zones. The maximum strength anywhere in the model is limited by the Brace–Goetze profile. Therefore, other dynamic strengths profiles in 2D and 3D will be restricted to the grey area.

In the context of the Brace–Goetze strength profile, the brittle–ductile transition is defined by the intersection of the brittle and the ductile strength curves. The material above the intersection is assumed to behave in a brittle manner, whereas the material below is assumed to flow viscously. However, this sharp brittle–ductile boundary overlooks any semi-brittle/semi-ductile behaviour such as observed in laboratory experiments and nature (Kohlstedt et al., 1995). This shortcoming remains unresolved even in more sophisticated numerical approaches that derive local strain rate from a momentum balance rather than assuming a constant strain rate everywhere (e.g. Albert et al., 2000).

The Brace–Goetze strength profile is an important concept that provides a first-order explanation for the geodynamics of plates. However, it does not adequately account for the dynamic evolution of deformation. In particular, it does not consider the complex communication of stresses and temperature (energy) between the plastically and viscously deforming parts of the system, or the feedback effects that lead to strain localisation and weakening of the system that is regulated by energy flux. The result of these simplifications is that it generates a region of maximum strength at the brittle–ductile transition in the crust, and also in the mantle, depending on the temperature profile (e.g. Ranalli and Murphy, 1987). However, geological evidence from extensional environments shows that strain is localised roughly at the crustal brittle–ductile transition (Davis and Coney, 1979; Miller et al., 1983; Wernicke, 1992; Jolivet et al., 1998; Chéry, 2001), an observation that is not reconciled with its supposed maximum crustal strength (Gueydan et al., 2004). In addition, based on seismological studies, there is no convincing evidence for the occurrence of seismic activity in the continental mantle, as would be expected from the Brace–Goetze strength profile (Maggi et al., 2000).

The concept of time dependent strength evolution of the lithosphere was already introduced in the 1980s (Kusznir, 1982; Kusznir and Park, 1982, 1984), but explicit dynamic coupling was not considered. These papers demonstrated that, by considering elasticity, the solution to lithospheric extension changes dramatically. In the simple case of constant applied load, creep in the lower crust amplifies the stresses in a strong, essentially elastic layer at the brittle–ductile transition to the extent that a low applied constant force can lead over time to whole scale lithosphere failure. Kusznir (1982) coined the term 'visco-elastic stress amplification' to describe this important phenomenon.

In numerical models that do not consider the energy flux through the system, strain localisation generally does not take place self-consistently, but requires the imposition of *ad hoc* rules, usually a Mohr–Coulomb rheology for the brittle layer and a strain softening rule for the viscous layer. The Mohr–Coulomb rheology formalises localisation as a mathematical concept through a particular shape of the yield envelope and a particular choice of flow rule. The most obvious effect of this imposition is that faults in the brittle layer have high angles constant with depth, so that listric and detachment faults do not arise naturally (Wijns et al., 2005). More subtly, *ad hoc* rules of localisation modify the geometric and thermal evolution of the system.

3. Energy feedback effects

In order to model the spontaneous process of strain localisation without imposing *ad hoc* rules, it is necessary to shift attention from critical stress states to critical energy states. This is because it is the process of energy dissipation that gives rise to feedback effects that localise strain. One way to do this is by investigating the dynamic evolution of the lithospheric strength profile using full coupling between continuity (e.g. conservation of mass), momentum (balance of forces) and energy equations (Regenauer-Lieb et al., 2006). The energy equation describes the dynamic evolution of the energy fluxes in the system, and it is the only time-dependent equation. Consequently, time-dependent instabilities only arise from feedback processes including the energy equation. This is a fundamental difference to the quasi-static Brace–Goetze approach, in which the brittle field is entirely time independent, and likewise, the ductile field has no time-dependent strength evolution. In order to overcome this deficiency, strain weakening laws have been proposed (e.g. Huismans et al., 2005; Lavier and Manatschal, 2006); however, these weakening laws are not derived self-consistently.

Our modelling approach incorporates energy feedback effects as driving forces for instabilities. These instabilities can be caused by self-localising pressure feedback effects that are dominant in the brittle parts of the system, and by self-localising temperature feedback effects that are dominant in the viscous parts of the system (Fig. 2). In these feedback loops, any pressure or temperature variations in the rock trigger feedback effects.

There are three possible sources for pressure variations in geological systems. One is by thermal expansion associated with thermal anomalies within the material. Mathematically, the isentropic work (W) of thermal expansion can be expressed as:

$$\frac{\mathrm{d}W}{\mathrm{d}t} = \alpha \, \Delta T \frac{\mathrm{d}p}{\mathrm{d}t} \tag{1}$$

where α is the thermal expansion coefficient and $\frac{\mathrm{d}p}{\mathrm{d}t}$ is the resulting pressure perturbation resulting from the thermal strain, expressed as the product of $\alpha \, \Delta T$.

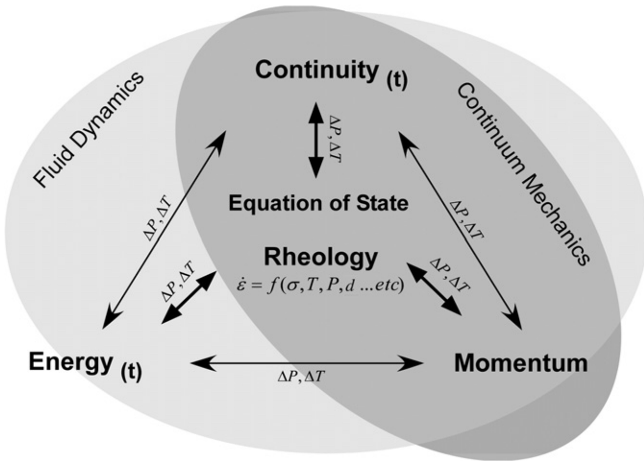

Fig. 2. Energy flux feedback loop linking the energy, continuity and momentum equations. Subtle changes in pressure and temperature can lead to a transition from elastic rheology to elasto-visco-plastic material behaviour, thereby having a large effect on the deformation of rocks. The key equations describing this transition are the constitutive rheological equations that define the strain rate ($\dot{\varepsilon}$) as a function of stress (σ), temperature (T), pressure (P), grain size (d) etc.

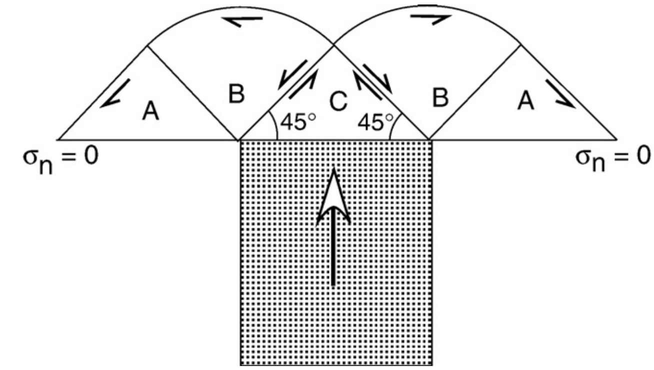

Fig. 4. Pressure variations caused by a rigid indenter pushing into softer material with normal stress free boundary conditions ($\sigma_n = 0$). The indenter causes the development of two symmetric fans of dextral and sinistral slip lines. For ideal plastic conditions with shear strength τ, the pressure variations can be calculated from the slip line theory (see Regenauer-Lieb and Petit 1997 for quantitative solutions). From area A to area C, the pressure increases systematically by a factor of 4.1 times the yield stress (Regenauer-Lieb and Petit, 1997).

Pressure perturbations can also result from the deformation of rocks with minerals of different strength contrasts. This is a fundamental mechanism in polycrystalline rocks under any pressure and temperature conditions. The magnitude of pressure variations is dependent on the geometry of rigid or soft inclusions, the deformation boundary conditions and rheology of the materials. For simple geometries, pressure variations can be calculated analytically, as done, for example, by Johnson et al. (1970) for rigid-plastic rheology, and Schmid and Podladchikov (2004) for linear viscous rheology. Pressure variations associated with an ellipse in simple shear and a rigid mineral indenting softer material are shown in Figs. 3 and 4, respectively.

The third possible source of pressure variations is associated with volume changes related to phase transitions. This is not considered in our models because it is tied to specific pressure–temperature conditions dependent on specific rock chemistry.

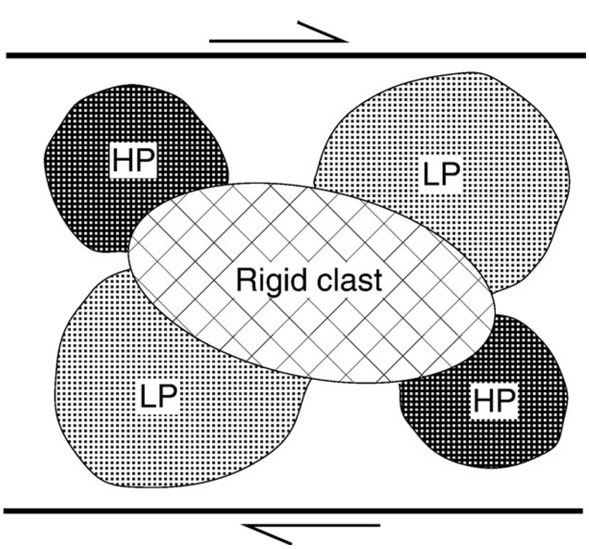

Fig. 3. Pressure variations associated with material strength contrasts in a simple shear setting. A rigid elliptic inclusion is embedded in a viscous matrix, causing low-pressure (LP) and high-pressure (HP) domains. Pressure variations are of the order of 3.5 times the background pressure (modified after Schmid and Podladchikov, 2004). Similar behaviour of pressure distribution can be obtained for elastic material in contact mechanics (Johnson, 1989).

Fig. 5 shows how local pressure variations can localise failure. Here, we consider a particle loaded to a value close to critical failure. Local pressure decreases around the particle can lead to failure, whereas increases lead to a move away from the failure envelope. Thus, failure is localised at low pressure sites, and large scale structures arise out of a catastrophic linkage of several of these regions across the rock mass.

Self-Localising temperature feedback processes rely chiefly on small temperature perturbation (caused, for example, by chemical reaction, heat of solution, radiogenic heat or phase changes) that weakens the rock, leading to increased strain rate and shear heating. The consequences of these feedback effects can be threefold: (1) stress can be relaxed under constant strain rate with no Localisation; (2) the deformation mode can switch over to a pressure-sensitive instability as described above; and (3) the elastic energy stored in the rock mass can be transformed into ductile shear heating in zones of localised strain. In this case, localisation results from a short-term runaway

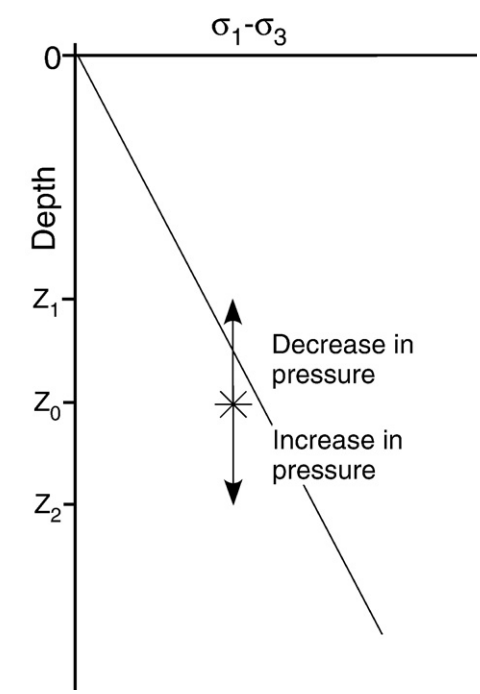

Fig. 5. Schematic diagram showing the effect of local pressure variations on brittle material in a sub-critical state. A positive (compressive) pressure perturbations leads to stabilisation, whereas negative (tensional) perturbations lead to failure.

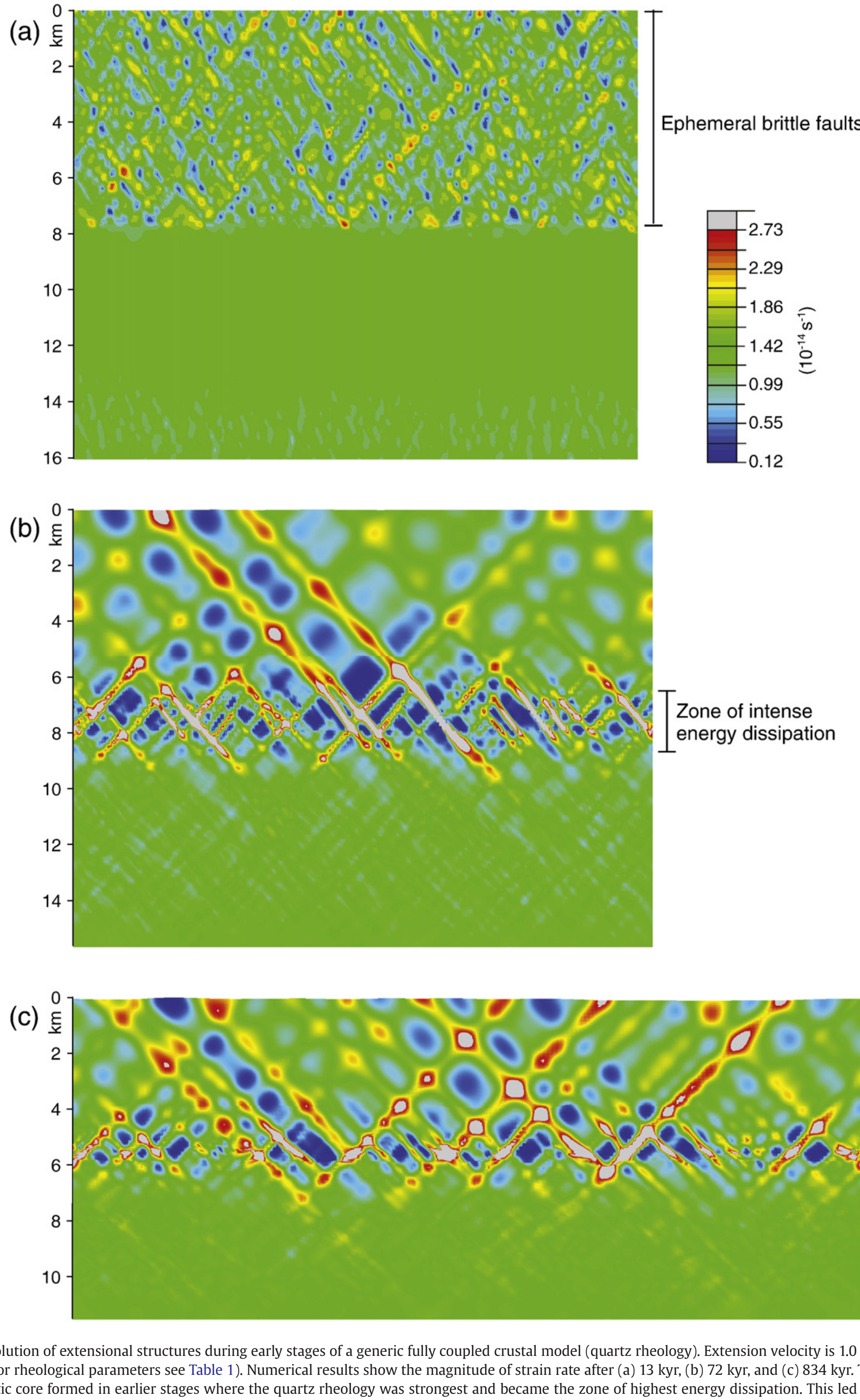

Fig. 6. Strain rate evolution of extensional structures during early stages of a generic fully coupled crustal model (quartz rheology). Extension velocity is 1.0 cm/yr and surface heat flow is 60 mW/m^2 (for rheological parameters see Table 1). Numerical results show the magnitude of strain rate after (a) 13 kyr, (b) 72 kyr, and (c) 834 kyr. The β factor for the last stage is 1.4. The elastic core formed in earlier stages where the quartz rheology was strongest and became the zone of highest energy dissipation. This led to most efficient strain localisation into zones of high strain rate surrounding lozenges of very low strain rate. Thus, the initially strongest layer became the weakest layer in the system.

process through an accelerating shear heating instability, whereby the temperature increase causes a decrease in viscosity, increased strain rate, and increased temperatures closing the loop.

The only negative feedback effects that prevent catastrophic failure are provided by (1) thermal conduction, dissipating the thermal pulse that develops through ductile or brittle instabilities, and (2) imposed constant velocity boundary conditions that stabilises velocities. The general framework of the brittle and ductile feedback loops, as shown in Fig. 2, shows that the equation of state is tied to the continuity equation, and the rheology acts as the most important feedback filter with its dependence on temperature, pressure, deviatoric stresses, grain strain, water content, activation energy and other material parameters. The temporal evolution of the process is thus controlled inherently by the energy equation, leading to weak structures that deform the lithosphere more efficiently by means of strain localisation.

This approach is an extension of classical fluid dynamics, in which pattern development (e.g., convection) is controlled by temperature evolution. The important difference is that classical fluid dynamics does not incorporate stored energy from elasticity. The incorporation of elastic rheology changes the fundamental energetics of the deforming bodies. The current limitation in the model is that the energy feedbacks of the top purely brittle layer (down to ~2 km depths) are not yet considered. This is because at shallow depths, the surface energy of cracks becomes increasingly important, but the current numerical scheme is unable to model this process. A more detailed description of the feedback loops, the equations employed and a comparison to the classical continuum mechanics approach is described elsewhere (Regenauer-Lieb and Yuen, 2003, 2004; Kaus and Podladchikov, 2006; Regenauer-Lieb et al., 2006; Regenauer-Lieb and Yuen, in press).

4. A dynamic approach to lithospheric strength

Static shear stress quantifies the quasi-static strength of material in continuum mechanics. The effective viscosity, defined as the shear stress divided by the associated strain rate, is the equivalent dynamic strength measured in fluid dynamics. A shear zone would normally have a high shear stress and would therefore be described as relatively strong in a continuum mechanics sense. However, its lower viscosity (resulting from associated high strain rate) accounts for a relatively low strength in a fluid dynamic sense. In order to compare the dynamic strength profile, which considers energy feedback mechanisms, with the quasi-static Brace–Goetze strength profile (Fig. 1), we refer only to the classical continuum mechanics shear stress definition of lithospheric strength.

The black curve in Fig. 1 shows a snapshot of a dynamic strength profile for calculations that fully couple the three equations in Fig. 2. The dynamic strength is calculated shortly (130 kyr) after the onset of extension. The dashed curve shows the equivalent static strength of the classical quasi-static Brace–Goetze profile for the same model.

At the early stage of deformation shown in Fig. 1, there is still some similarity between the dynamic strength profile and the quasi-static Brace–Goetze profile. At later stages, the shape of the dynamic strength curve would be profoundly different to that of the Brace–Goetze curve (see Fig. 7 in Regenauer-Lieb and Yuen, in press). Even at this early stage, it is obvious that the integrated strength of the lithosphere (the area to the left of each curve) is much weaker when localisation feedbacks are considered (Fig. 1). The top brittle crust and the area below 52 km already show drastic stress drops due to unloading of elastic stresses by active, high-stress shear zones. These areas have been subjected to considerable weakening through the positive feedback mechanism after the onset of visco-elasto-plastic flow. While the bottom layer is weakened through shear heating feedback and its effect on rheology, the top layer is weakened primarily through thermal expansion feedback. These weakening mechanisms amplify the visco-elastic stress amplification described earlier (Kusznir, 1982).

Another significant difference between the classical Brace–Goetze profile and the dynamic strength profile is the fact that the latter is not applicable in one-dimension. The effect of weakening in shear zones immediately leads to the focusing of stresses into the shear zones, thereby unloading the stress stored elastically in adjacent quasi-rigid blocks. This effect shows that elasticity plays a major role in a fully coupled energy approach. Stress variations in 2D or 3D in our models and in nature are considerably larger than those considered in a quasi-static approach (Fig. 1).

5. Self-organization: communication across brittle and ductile layers

The implementation of the energy feedback approach in numerical models, with its concept of dynamic strength profile, has profound implications for understanding geodynamic processes. Small perturbations in a particular area can have large effects for the whole model. Conversely, large variations can decay because they become energetically less favourable. Therefore, the incorporation of energy feedbacks into numerical simulations leads to rich solutions for the way faults nucleate and propagate through the lithosphere, resulting in a rich variety of extension styles.

Examples of 2D crustal extension models are presented in Figs. 6 and 7. Results show that the dynamics of extension is intimately linked with the development of strong elastic cores, which comprise regions with the highest regional stresses and where the dominant deformation mechanism is initially elastic. The strong elastic cores become zones of highest energy dissipation, which lead to strain localisation into zones of intense weakening (Fig. 6b) and ultimately become high strain rate and high stress detachment zones (Regenauer-Lieb et al., 2006). These zones are strong in the continuum mechanics sense and weak in the fluid dynamics sense. Multiple elastic cores can be developed depending on compositional and rheological stratification.

In the early stages of deformation, the model lithosphere is loaded and ephemeral brittle fracturing propagates downwards from the weak surface (Fig. 6a). These faults progressively propagate towards the brittle–ductile transition (Fig. 6b), coalescing into distinct zones of high strain rate (Fig. 6c). As loading continues, strain localisation also occurs in the ductile layer, propagating upwards towards the brittle–ductile transition (Fig. 7).

The existence of the strong elastic core is the key to understanding early stages of energy communication across brittle and ductile layers. In the process of upward shear zone or downward brittle fault propagation, energy in the form of heat and stored elastic energy is transferred from the termination of these localised bands to the elastic core, thus loading this region with stored energy. Communication between localisation in the upper and lower crusts takes place through the strong elastic band. At some point, loading of this strong layer leads to strain localisation into well-defined shear bands characterised by high strain rate (Fig. 6b) and shear heating, surrounding blocks of low strain rate. This process transforms the elastic layer from the strongest in the system, to the weakest. In other words, strain localisation is best developed in the strongest layer, leading to high energy dissipation and effective weakening.

Ductile creep is a much slower process than brittle faulting; this is the reason for the delay in localisation in the weak ductile parts of the lithosphere. In the absence of significant weakening related to microstructural modification, such as grain size reduction or metamorphic reactions in the fault zone, brittle faults are ephemeral. When ductile shear zones form, the system eventually organises itself through the communication between fast brittle faults and slow ductile shear zones (Fig. 7). This happens when brittle faults become rooted on the ductile shear zones and thus become permanent. Thus development of long term structures is controlled by these rooted brittle faults. At this stage, the elastic layer is preserved in well-defined elastic cores and detachment faults are developed in zones where brittle and ductile localisation phenomena interact (Fig. 7b). The different rates at which localised strain zones move implies that short timescale phenomena are controlled by brittle features whereas long timescale phenomena are likely to be controlled by ductile shear zones.

An important result is that the brittle–ductile transition is not defined by a single depth but appears in a broad transition zone in

Table 1
Parameters used in numerical models

Parameter	Name	Value	Units
X	Shear heating efficiency	1*	–
κ	Thermal diffusivity	Q=0.7×10^{-6} F=0.7×10^{-6} O=0.8×10^{-6}	m^2s^{-1}
α_{th}	Thermal expansion	3×10^{-5}	K^{-1}
c_P	Specific heat	Q=800 F=800 O=1000	J $kg^{-1}K^{-1}$
P	Density	Q=2800 F=2800 O=3300	kg m^{-3}
V	Poisson ratio	0.25	–
E	Young's modulus	4.5×10^{10}	Pa
A	Material constant—pre-exponential parameter	Q†=1.3×10^{-34} F§=7.9×10^{-26} O#=3.6×10^{-16}	Pa^{-n} s^{-1}
N	Power-law exponent	Q†=4 F§=3.1 O=3.5	–
H	Activation enthalpy	Q†=135 F§=163 O#=480	kJ mol^{-1}
B	Thickness of radiogenic layer	10	km
Qs	Surface heat flow	50/60	mWm^{-2}
Q_m	Mantle heat flow	20	mWm^{-2}

Note: Q=Quartz; F=Feldspar; O=Olivine. *Chrysochoos and Belmahjoub (1992), †Hirth et al. (2001), §Shelton and Tullis (1981), #Hirth and Kohlstedt (2004).

which brittle and ductile processes interact. Within this transition zone, detachment zones form in the area where ductile deformation communicates with brittle deformation. Such detachments can initially form at relatively shallow depth (~5 km, Fig. 7b) and are progressively exhumed during extension.

In summary, the strong elastic layer has a fundamental role in controlling the style of extensional deformation. In cases where the olivine-dominated upper mantle is hot, the strongest elastic core develops within the crust, dominating the development of crustal core complexes, while the mantle responds passively (Rosenbaum et al., 2005; Regenauer-Lieb et al., 2006). In cases where the mantle is cold, the dominant elastic core shifts to a brittle–ductile transition developed in the upper mantle, resulting in mantle core complexes (Weinberg et al., 2007).

6. Discussion

6.1. The role of shear heating

Shear heating is a fundamental physical process in our models and is considered to have a key thermodynamic effect in the processes of strain localisation in nature, such as was found for material sciences and metallurgy (e.g. Coffin and Rogers, 1967; Braeck and Podladchikov, 2007). However, shear heating has not been unambiguously documented in nature, and its role during deformation and metamorphism remains controversial (e.g. Camacho et al., 2001; Bjørnerud and Austrheim, 2002). At small spatial and temporal scales, evidence for co-seismic shear heating and rock melting is recognised in pseudotachylites (Magloughlin and Spray, 1992). A larger scale (~1000 km) effect of shear heating has also been postulated, for example, for the great 1994 Bolivia earthquake (e.g. Kanamori et al., 1998).

Temperature rises of the order of a centigrade are capable of localising strain (Regenauer-Lieb and Yuen, 1998). Using examples from material science, Braeck and Podladchikov (2007) propose that shear heating at a very small scale is a fundamental instability limiting

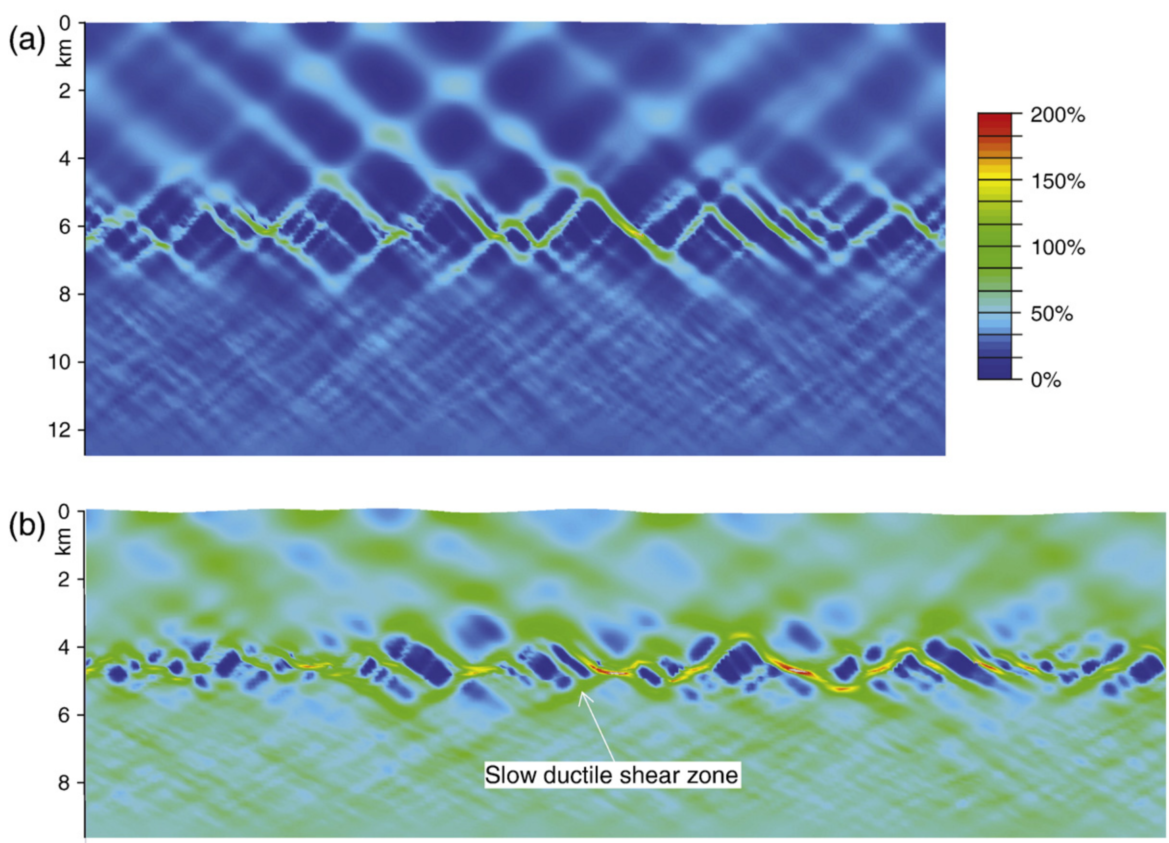

Fig. 7. Strain evolution of extensional structures in a generic fully coupled crustal model (quartz rheology). Extension velocity is 1.0 cm/y and surface heat flow is 60 mW/m^2 (for rheological parameters see Table 1). Numerical results show the magnitude of strain after (a) 595 kyr (β factor=1.3) and (b) 1.28 Myr (β factor=1.7).

the strength of all materials. Such small temperature increase is normally not detectable in the field. In some cases, however, strain localisation may lead to a thermal runway during seismic events (Kelemen and Hirth, 2007) to the point of forming melts. The difficulty in this case is that once this runway process occurs it is impossible to tell whether shear heating resulted from localisation or caused localisation. One possible way to detect the effect of shear heating is to look at larger scale shear zones that are equivalent to the characteristic thermal diffusion length scale of that shear zone. For example, assuming a shear zone with characteristic effective strain rate ($\dot{\varepsilon}_{\mathrm{eff}}$) of 10^{-12} s^{-1}, and rock diffusivity (κ_{t}) of 10^{-6} m^2/s, the characteristic thermal width (l) of the shear zone will be in the order of a kilometre (Eq. 2).

$$l \sim \sqrt{\frac{\kappa_{\mathrm{t}}}{\dot{\varepsilon}_{\mathrm{eff}}}} \tag{2}$$

The analyses of Braeck and Podladchikov (2007), as well as our own models (Regenauer-Lieb et al., 2006), suggest that self-localising instabilities without melting must be far more common than runaway melting instabilities. This self-localisation is very important for deformation on geological timescales, because they ensure longevity of faults. From a geodynamic perspective it is hence important to find evidence for shear heating on a kilometre length scale.

6.2. Sensitivity to heat flux

In our previous model results (Rosenbaum et al., 2005; Regenauer-Lieb et al., 2006; Weinberg et al., 2007), we have explored the main features of extensional systems, aiming at predicting rather than prescribing strain weakening phenomena. Here, we wish to demonstrate that the style of extensional features, and in turn, the extensional mode, is strongly sensitive to small variations in heat flux.

In a classical quasi-static approach to localisation, which excludes the energy feedback approaches, we would expect that raising the heat flow through the crust would result in a continuous trend of thinning of the brittle layer (or layers) related to the shallowing of the brittle–ductile transition (Ranalli and Murphy, 1987). A similar effect is also seen in our fully coupled dynamic approach, but this is complemented by additional, distinct, non-linear mode transitions. Fig. 8 shows the differences in results related to an increase in heat flow, equivalent to ΔT_{Moho} of 34 °C.

Results show that the slightly hotter model (b) is strongly dominated by doming of the mantle associated with well-developed mantle detachment and pronounced Moho topography. In contrast, the slightly colder model (a) has better developed crustal detachment and weakly developed mantle detachments, leading to an undulating Moho, lacking a well-defined long wavelength. Interestingly, the difference in Moho topography is opposite to the general trend obtained for higher temperature deviations, where a hot model resulted in a flat Moho (Regenauer-Lieb et al., 2006) whereas a cold model was characterised by distinct mantle doming (Weinberg et al., 2007). This apparent contradiction results from the sensitivity of the system to interference between crustal and mantle localisation wavelengths that can be constructive or destructive.

6.3. Is the lithosphere too strong?

A key question related to continental break-up is whether or not typical continental lithospheres are too strong to split apart unassisted by magmas (Buck, 2006). Underlying this question are rough estimates of lithospheric strengths based on Brace–Goetze strength

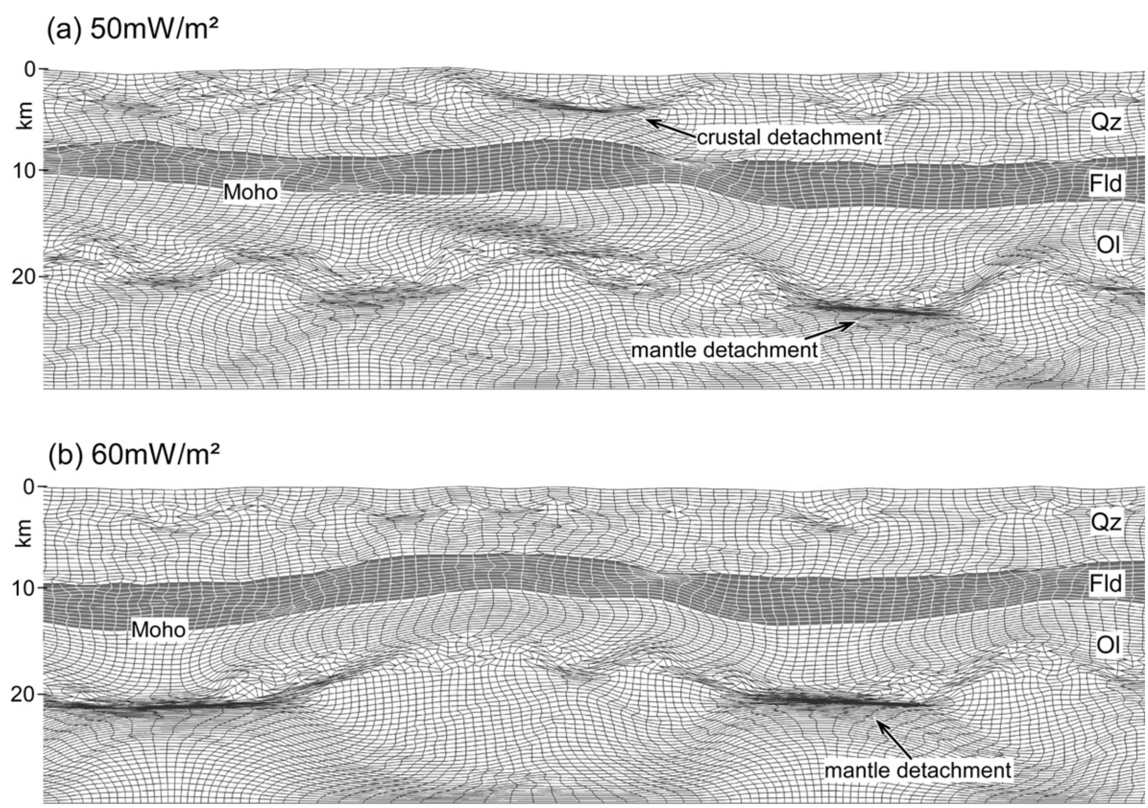

Fig. 8. Results of numerical models showing extension of a three layer (quartz (Qz), feldspar (Fld) and olivine (Ol)) continental lithosphere (see Table 1 for rheological parameters). Initial crustal thickness is 30 km. Results are calculated after 13.7 Myr extension with total velocity of 1.2 cm/yr. The feldspar layer is highlighted in dark grey. Initial thermal conditions follow a steady state isotherm from Equations 4–31 in Turcotte and Schubert (2002). The following initial conditions are used: thickness of the heat producing layer = 10 km; crustal conductivity = 2.8 $\mathrm{Wm^{-1}K^{-1}}$; mantle heat flow = 20 $\mathrm{mWm^{-2}}$; surface temperature = 280 K. The steady state isotherm is initially perturbed by 70 K amplitude white noise perturbations on 4% of the nodes. (a) Results for surface heat flow of 50 $\mathrm{mWm^{-2}}$; (b) Results for surface heat flow of 60 $\mathrm{mWm^{-2}}$.

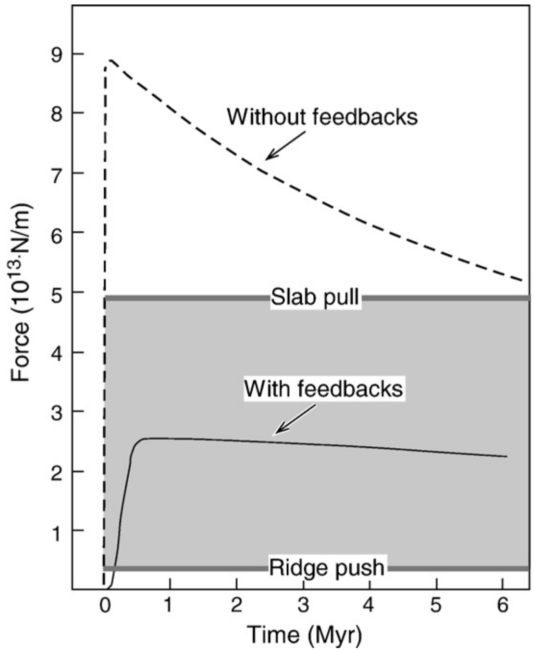

Fig. 9. Integrated forces calculated for lithosphere subjected to extension at 1.2 cm/yr shown in Fig. 8b. The original crustal thickness of 30 km is underlain by 50 km thick olivine mantle. The surface heat flow is 60 mWm^{-2}. Calculation is shown with and without feedback effects (black and dashed lines, respectively). The maximum force calculated for the coupled model at $2.55 \cdot 10^{13}$ N/m is equivalent to $3.54 \cdot 10^{8}$ Pa. This value is much lower than typical forces calculated for slab pull (4.9×10^{13} N/m) but higher than typical forces for ridge push ($3.41 \cdot 10^{12}$ N/m). Typical slab pull and ridge push values are from Turcotte and Schubert (2002). Note that these values are order of magnitude estimates due to uncertainties in extrapolation of laboratory experiments. Nevertheless, significant weakening of the coupled versus uncoupled solutions occurs independently of absolute strength.

profiles. Fig. 9 shows the calculated lithospheric strength measured as the integrated force applied in order to maintain a constant vertical boundary velocity. Results are shown for a fully coupled model and a traditional model that does not take into account feedback effects. For both scenarios the integrated force decreases with time because of the effect of lithospheric thinning. However, in the fully coupled model, the total strength of the lithosphere reaches a peak corresponding to a quarter of the peak reached when no feedback effects are considered. This implies that the lithosphere can rupture well below the maximum conceivable geodynamic forcing provided by slab pull. Therefore, plate tectonic processes other than subduction can have an effect on the mobility of the continental lithosphere. For example, it has been suggested that stresses induced by gravitational potential energy variations allow catastrophic weakening responsible for intraplate deformation (Coblentz and Sandiford, 1994; Sandiford et al., 2005). In a fully coupled framework, the role of weakening through magmatism does not have to be invoked.

7. Conclusions

Increasingly powerful numerical models in 2D and 3D are revealing the physical processes underlying extensional systems. Such models fall into two categories: (1) engineering-style models that use *ad hoc* rules for strain localisation; and (2) thermodynamic-style models that use an energy feedback approach to derive spontaneous localisation phenomena from the supposed underlying physical processes. The application of the second approach to geological problems is relatively new (Regenauer-Lieb et al., 2001; Regenauer-Lieb and Yuen, 2003; Kaus and Podladchikov, 2006; Regenauer-Lieb et al., 2006; Braeck and Podladchikov, 2007) and requires further field-based testing. Its application has produced encouraging results such as the spontaneous development of a number of structures that match well-documented features of extensional systems, including listric and detachment faults, as well as exhumation of crustal and mantle core complexes. It also indicates that the simplicity of the Brace–Goetze strength profile of the lithosphere might not appropriately define the strength of the lithosphere. When energy feedback effects are considered, the lithosphere weakens as strain accumulates. Most intensive strain localisation and energy dissipation through shear heating takes place in the initially strongest, elastic section of the lithosphere. This process transforms the strongest layer into the weakest. This process is what allows for continental break-up in the absence of mantle-derived dykes. The dynamic and structural evolution of such a system is strongly sensitive to initial conditions such as heat flux, and the nature of the initial rheological layering of the lithosphere. Therefore, in order to understand extensional systems it is necessary to take a multi-disciplinary approach that integrates the dynamic evolution of lithospheric strength and strain localisation with a much deeper understanding of the multiple roles of mafic and felsic magmatism and the ability of the mantle to produce melt.

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