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Modes of continental rifting as a function of ductile strain localization in the lithospheric mantle

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ABSTRACT

Analogue and numerical models have shown that the strength of the lithospheric mantle controls the mode of lithosphere deformation. In extension, the presence or absence of a high strength brittle mantle respectively leads to localized or distributed rifting. However, first order geophysical data question the existence of such a brittle mantle. Here we use 2-D finite-element large strain modelling to quantify the impact of a ductile localizing mantle – instead of brittle – in triggering continental rifting. As a novelty, the mantle rheology considers the effect of grain boundary sliding during strain-induced grain size reduction, which may promote a significant strength drop and subsequent strain localization at low mantle temperature (<700–800 °C). Our results reveal that such ductile localizing mantle implies varying modes of continental rifting that mainly depend on both the amount of weakening in the ductile mantle and the strength of the lower ductile crust. A medium to strong lower crust implies coupling between the upper crust and ductile localizing mantle, yielding to narrow continental rifting. In contrast, a weak lower crust implies decoupling between the upper crust and ductile localizing mantle, giving rise to a switch from distributed faulting at incipient strain to localized faulting at large strain. Ductile strain localization in the lithospheric mantle is therefore sufficient to trigger continental rifting, although a critical amount of weakening is required. Such ductile localizing mantle provides a relevant geological and mechanical alternative to the brittle mantle. It moreover provides a wider variety of modes of upper crustal faulting that are commonly observed in nature.

1. Introduction

Thermal and rheological layering of the continental lithosphere exerts a direct control on the mode of lithosphere deformation. While the crustal strength dominates the lithosphere strength for a Moho temperature (T_M) higher than 700 °C, the high-strength sub-Moho mantle controls the lithosphere strength for T_M lower than 700 °C (Gueydan et al., 2008). The presence of a high strength uppermost mantle together with a weak deep crust that permits horizontal ductile shearing along the Moho is essential to promote strain localization at lithosphere scale (Allemand and Brun, 1991; Brun and Beslier, 1996; Gueydan et al., 2008). In analogue and numerical experiments, the high strength uppermost mantle is modelled as a brittle mantle, such as predicted by the classical strength profile of the lithosphere (Brace and Kohlstedt, 1980; Carter and Tsenn, 1987; Sawyer, 1985). Some of these numerical models promote the development of mantle fault zone through an ad-hoc drop of friction coefficient with increasing strain (Gueydan et al., 2008; Huismans and Beaumont, 2003; Huismans et al., 2005). Bos and Spiers (2002) predict indeed a decrease of analogue fault strength with increasing strain. This weakening process can be applied to upper crustal faulting and could well explain the weakening of major lithospheric faults (Holdsworth, 2004). However, the applicability of a brittle weakening process to the uppermost mantle remains to be demonstrated. Furthermore, the existence of a brittle mantle is now questioned. The rare occurrence of earthquake within the continental lithospheric mantle (Maggi et al., 2000), together with the low elastic thickness inferred in deformed region (Audet and Bürgmann, 2011; Jackson, 2002), suggest in opposite that the crustal strength controls the lithosphere strength. These features contradict therefore the existence of a high strength brittle uppermost mantle, but they also conflict with pre-requisites of the lithosphere dynamics that involve a high strength and localizing mantle to explain both lithosphere-scale strain localization and the large values of elastic thickness beneath cratons (Burov and Watts, 2006).

As an alternative, an entirely ductile mantle with an initial high strength that decreases during deformation can reconcile these opposite views of lithosphere strength. Indeed, Frederiksen and Braun (2001) have shown that mantle shear zones can be generated in response to strain-induced weakening of the uppermost lithospheric mantle, providing a simple explanation for lithosphere-scale strain localization. Based on structural and microstructural observations in the Ronda peridotite and using available olivine flow laws (Hirth and Kohlstedt, 2003), we have shown that viscous deformation dominated

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by dislocation-accommodated grain boundary sliding (disGBS) can induce a significant weakening during dynamic recrystallization at low mantle temperature (Précigout et al., 2007). 1-D aggregate-scale numerical modelling of the disGBS-induced weakening moreover shows that this process is efficient to strongly reduce the mantle strength for low mantle temperature (Précigout and Gueydan, 2009). These results exemplify the role of ductile strain localization within the ductile mantle and allow us to define a ductile localizing mantle. Nonetheless, the impact of such a ductile localizing mantle on lithosphere-scale deformation remains to be constrained. This is the main purpose of the present study through 2-D numerical analyses focusing on 1/the role of the amount of weakening in the ductile mantle and 2/the impact of ductile crust rheology for the mode of continental rifting.

2. Continental lithosphere rheology

2.1. Mantle rheology

We account here for a composite olivine rheology that combines four deformation mechanisms with their respective flow laws: dislocation creep (r), diffusion creep (d), disGBS (g) and low-temperature plasticity (e) (Drury, 2005; Goetze, 1978; Hirth and Kohlstedt, 2003; Précigout and Gueydan, 2009; Précigout et al., 2007). Each mechanism contributes to the overall strain rate ($\dot{\varepsilon}$) of an olivine aggregate as follows:

$$\dot{\varepsilon} = \dot{\varepsilon}_r + \dot{\varepsilon}_d + \dot{\varepsilon}_g + \dot{\varepsilon}_e \tag{1}$$

With dislocation creep

$$\dot{\varepsilon}_r = \mathrm{A_r} \cdot \exp\left(\frac{-\mathrm{Q_r}}{\mathrm{RT}}\right) \cdot \tau^{\mathrm{n_r}} \tag{2}$$

diffusion creep

$$\dot{\varepsilon}_d = \mathrm{A_d} \cdot exp\left(\frac{-\mathrm{Q_d}}{\mathrm{RT}}\right) \cdot \tau^{\mathrm{n_d}} \cdot \mathrm{d}^{-\mathrm{m_d}} \tag{3}$$

disGBS

$$\dot{\varepsilon}_g = \mathrm{A_g} \cdot \exp\left(\frac{-\mathrm{Q_g}}{\mathrm{RT}}\right) \cdot \tau^{\mathrm{n_g}} \cdot \mathrm{d}^{-\mathrm{m_g}} \tag{4}$$

and low-temperature plasticity (or exponential creep)

$$\dot{\varepsilon}_e = A_e \cdot \exp\left[-(Q_e/RT) \cdot \left(1-\tau/\tau_p\right)^{n_e}\right] \tag{5}$$

where, $\dot{\varepsilon}$, τ, d, m, T, R, A, Q, and n are, respectively, the strain rate (in s^{-1}), the equivalent shear stress (in MPa), the grain size (in µm), the grain size exponent, the temperature (in K), the gas constant (J/°), the pre-exponential constant (MPa^n/s), the activation energy/enthalpy (in J, the effect of pressure is disregarded here) and the stress exponent. The subscripts r, d, g and e stand for respectively dislocation creep, diffusion creep, disGBS and exponential creep. τ_p is the Peierls stress defined for low-temperature plasticity (Goetze, 1978).

The composite olivine rheology thus involves several mechanisms that compete each other; the mechanism with the highest strain rate/lowest stress dominates the deforming aggregate depending on stress, grain size, temperature and overall strain rate. The conditions for which each mechanism dominates are displayed through a so-called deformation map, i.e., a stress-grain size log–log plot for a range of temperature at constant overall strain rate. Based on the iso-temperature curves, this graph highlights four stress-grain size fields, respectively for dislocation creep at large grain size and low stress, for exponential creep at high stress, for diffusion creep at low grain size and low stress, and for disGBS at intermediate conditions (Fig. 1). While dislocation creep and low-temperature plasticity are grain-size-insensitive (GSI) creep, as shown by horizontal curves on the deformation map, both diffusion creep and disGBS are grain-size-sensitive (GSS) creep at a different degree, as indicated by different slopes for the iso-temperature curves (Fig. 1). This degree directly depends on the grain size exponent (m) of the olivine flow law (Eqs. (3) and (4)).

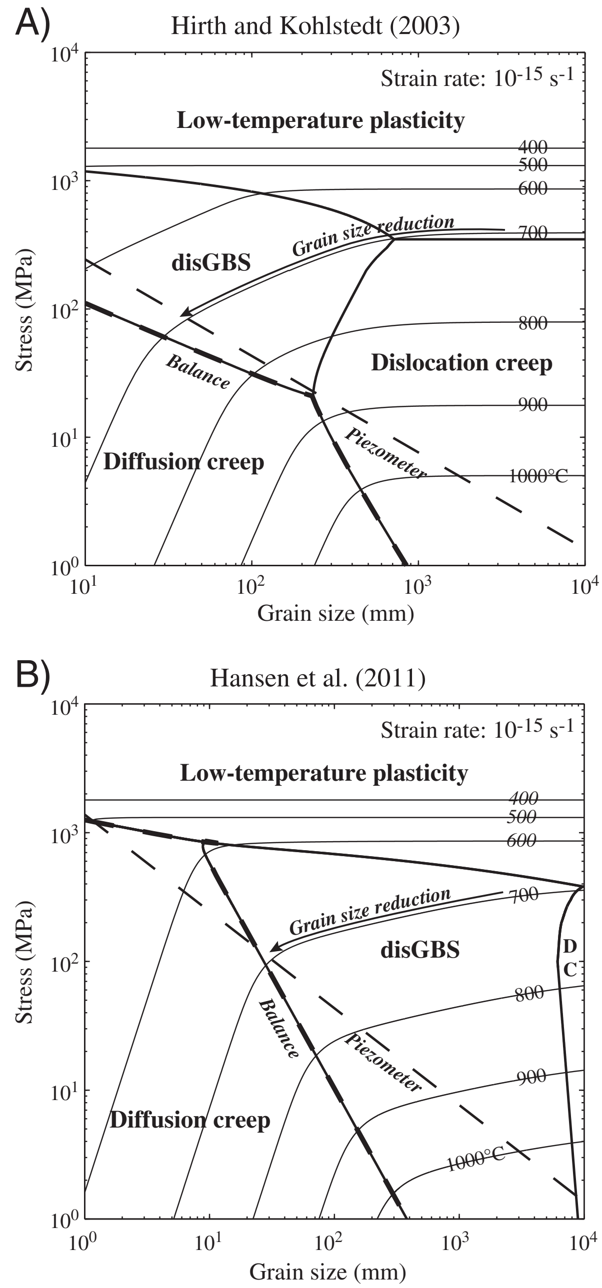

Fig. 1. Olivine deformation maps (shear stress vs. mean grain size) at a constant strain rate ($10^{-15} \cdot s^{-1}$) showing the four deformation mechanisms that compete to control the mantle rheology: low-temperature plasticity (exponential creep), dislocation creep, disGBS and diffusion creep. Iso-temperature curves are provided to show stress/grain sizes for temperatures ranging from 1000 to 400 °C. A/deformation map including the disGBS flow law from Hirth and Kohlstedt (2003); B/deformation map including the disGBS flow law from Hansen et al. (2011). Hypotheses for the recrystallized grain size are either equilibrium (at the boundary between grain size sensitive and grain size insensitive creeps, thick dashed line; de Bresser et al., 1998) or experimentally constrained paleopiezometer (thin dashed line; Van der Wal et al., 1993). Values of the ductile flow laws are given in Table 1.

During strain-induced grain size reduction of olivine, the contribution of GSS mechanisms to the deformation of mantle rocks is enhanced at the expense of GSI mechanisms. As a consequence, olivine aggregate suffers a stress drop/weakening that may trigger strain localization