Asymmetric lithospheric extension: The role of frictional plastic strain softening inferred from numerical experiments

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ABSTRACT

Plane-strain thermomechanical finite element model experiments of lithospheric extension are used to investigate the effects of strain softening in the frictional plastic regime on the asymmetry of extension. Strain softening is considered in cases where the crust is either strongly or weakly coupled to the mantle, and as the extension velocity varies from 0.3 to 30 cm/yr. In the absence of strain softening, extension is symmetric (SS mode). When strain softening takes the form of a reduction in the internal angle of friction with increasing strain, lithospheric extension may be asymmetric at a lithospheric scale (AA mode), or exhibit crustal asymmetry concomitant with mantle symmetry (AS mode). The different styles depend on the relative control of the system by the frictional plastic and ductile layers, which promote asymmetry and symmetry, respectively. High extension velocities and weak ductile crust-mantle coupling tend to suppress the fundamental asymmetry induced by frictional strain softening. This is because they, respectively, increase the effective strength of the lower lithosphere and decrease the control by frictional plasticity.

Keywords: lithosphere, extension, rifting, strain softening, numerical modeling.

INTRODUCTION

Are rifts and conjugate passive margins fundamentally asymmetric, and if so, what controls the asymmetry? The seismically defined deep structure of continental rifts and passive margins has been interpreted both in terms of symmetric and asymmetric rifting models (Wernicke and Burchfiel, 1982; Lister et al., 1986; Keen et al., 1989; Mutter et al., 1989; White, 1990; Boillot et al., 1992; Sibuet, 1992; Louden and Chian, 1999). Whether the inferred asymmetries stem from whole lithosphere faults and shear zones remains a key question. Although a range of kinematic models has been proposed, including pure shear (McKenzie, 1978), simple shear (Wernicke, 1985), and combinations of these models (Lister et al., 1986), such models provide little insight into extensional processes. Most dynamical models (Braun and Beaumont, 1987; Chéry et al., 1992; Bassi et al., 1993) are either intrinsically symmetric or predict symmetric extension, and therefore the origin of possible asymmetry remains enigmatic, although inherited inhomogeneity may always be invoked.

Here we focus on asymmetry that stems from localized deformation owing to strain-dependent softening of the brittle crust and upper mantle. Forward numerical thermomechanical modeling is used to investigate the effect of a reduction in the effective internal angle of friction with increasing strain on the mode of lithospheric extension. Previous work (Buck, 1993; Lavier et al., 1999) shows that loss of cohesion on faulting can lead to large offsets on brittle normal faults during extension of an elastic-plastic layer. However, if the layer is of crustal thickness (>20 km), faults are abandoned and replaced after only a few kilometers of offset. Viscous shear localization can occur during the transition from dislocation creep to diffusion creep when there is ambient cooling (Braun et al., 1999). Frederiksen and Braun (2001) used a simple empirical viscous strain-softening model to demonstrate that ductile strain localization in the mantle lithosphere can dominate deformation during extension. Our study complements theirs by investigating brittle, as opposed to ductile, strain localization.

MODEL DESCRIPTION

A fully thermomechanically coupled, plane strain, incompressible viscoplastic model is used to investigate extension of a layered lithosphere with frictional plastic (brittle) and thermally activated power-law viscous (ductile) rheologies (Fig. 1). In the frictional plastic regime all materials have a pressure-dependent Drucker-Prager failure criterion. Initially, the effective angle of friction is 7°, but during strain softening this angle decreases with the second strain invariant (Fig. 1A). The physical rationale is that weakening of faults and shear zones may be caused by high, transient or static fluid pressures, or strain that produces gouge or mineral transformations (Sibson, 1990; Streit, 1997; Bos and Spiers, 2002). The deformation mechanism in the model depends on the ambient conditions; at yield flow it is plastic, below yield it is viscous.

The model has a free top surface and the other boundaries have zero tangential stress (free slip). Extension, driven by velocity boundary conditions and seeded by a small plastic weak region (Fig. 1), is computed using an arbitrary Lagrangian-Eulerian method for the finite element solution of incompressible viscoplastic creeping flows (Fullsack, 1995).

RESULTS

Reference Model, No Strain Softening—Model 1

In the reference model the crust and upper mantle are coupled and both are frictional plastic. This is achieved by increasing the effective viscosity for the wet quartz by a factor of 100 (Fig. 1A). There is no strain softening. Deformation is symmetric and is initially controlled by focused cross-conjugate frictional-plastic shears that extend from the surface and root in the viscous mantle, where they become diffuse (Fig. 1B). During this first phase there is little thinning of the ductile mantle. Later, during phase two (Fig. 1C) the conjugate frictional-plastic shears are abandoned, strain increases in the ductile lower lithosphere, and the model extends in a distributed near pure-shear symmetric necking mode.

Frictional Strain Softening—Model 2

Model 2 is the same as model 1 except that strain softening (Fig. 1A) is now included. During phase one strain softening preferentially focuses strain into one of the conjugate frictional shears, and deformation is asymmetric (Fig. 2A). Asymmetry is a consequence of the positive feedback between strength reduction (decreased internal angle of friction) and increasing strain. Any perturbation of the symmetric extension will lead to one shear gaining an advantage through the positive feedback mechanism. This shear will develop at the expense of the other. As in model 1, phase one is dominated by deformation of the frictional upper lithosphere, and the lower lithosphere, which is ductile and has no asymmetric tendency, deforms passively. The asymmetry offsets the loci of upper and lower lithosphere thinning by ~50 km (Fig. 2A).

During phase two (Fig. 2B) the frictional shears become less important and the style changes to pure shear, which thins both the crust and mantle lithosphere by necking of the region between the near surface frictional shears. The embedded frictional shear zone is
rotated and upwarped and is subhorizontal above the uplifted lower lithosphere, and its end roots diffusely in the ductile region below.

The two protomargins have pronounced differences in width and geometry; a wide lower plate margin with mantle lithosphere structurally emplaced beneath strongly extended upper crust, and a narrower upper plate margin dominated by the early conjugate shears and later stage rotation and shearing of the upper crust (Fig. 2B).

Frictional Strain Softening and Crustal Decoupling—Model 3

Model 3 has the same frictional strain softening as model 2, but the crust and mantle are decoupled by scaling the effective viscosity of the wet quartz by a factor of one, thereby allowing ductile flow in the lower crust (Fig. 1A). During phase one (Fig. 3A) some of the asymmetry of model 2 is retained, but crustal asymmetry is diminished because the conjugate frictional shears sole out in the decoupled ductile lower crust. Lower lithosphere asymmetry is still apparent in the diffuse shear zone and in the geometry.

During phase two (Fig. 3B) the crust develops secondary frictional shears within the rift zone, thereby creating boudin-like necks bounded by the frictional shears which sole out in the lower crust and propagate beneath the rift flanks. Corresponding ductile necking of the lower lithosphere during phase two is nearly symmetric.

Figure 1. A: Model geometry including weak seed (von Mises yield strength, \(\sigma_y\)) and velocity boundary conditions (\(V_{\text{ext}}\), \(V_b\)), with \(V_b\) chosen to achieve mass balance. Initial, laterally uniform temperature (\(T\)) increases from surface (0°C) in accord with uniform crustal heat production (0.8 \(\mu W/m^2\)) and basal heat flux (20 \(mW/m^2\)). Mantle lithosphere geothermal gradient is uniform and sublithospheric mantle is isothermal at 1330°C. Thermal conductivity is 2.25 W/m°C and thermal diffusivity is \(1 \times 10^{-6} m^2/s\). Densities (\(\rho\)) are shown at 0°C and thermal expansivity is \(3.1 \times 10^{-5}/C\). At right internal friction angle (\(\phi\)): initial \(\phi = 7^\circ\) (solid lines) and strain (\(\epsilon\)) softened, \(\Phi = 1^\circ\) (dashed lines) representative strength envelopes of coupled and decoupled models when \(V_{\text{ext}} = 0.3 cm/yr\). Frictional plastic strain softening with \(\epsilon\) is shown at top. Effective viscosity \(\eta_{\text{eff}} = A^{-n} e^{1 - m_{\epsilon}} \exp[(Q/PV)/nR T]\), where for wet quartz (Gleason and Tullis, 1995), initial constant \(A = 1.1 \times 10^{-28} Pa^{-n}\), power law exponent \(n = 4.0\), activation energy \(Q = 223 kJ/mole\), activation volume \(V = 0\), and for dry olivine (Karato and Wu, 1993), \(A = 2.4 \times 10^{-16} Pa^{-n}\), \(n = 3.5\), \(Q = 540 kJ/mole\), \(V = 25 \times 10^{-6} m^3/mole\), \(e\) is second strain rate invariant, \(P\) is pressure, and \(R\) is gas constant; \(x\) and \(z\) are horizontal and vertical coordinates, \(t\) is time, \(\Delta x\) is total amount of extension. B, C: Reference coupled model 1 [\(\eta_{\text{eff}}\) (wet quartz × 100)], no strain softening, showing deformed Lagrangian mesh, velocity vectors, and sample isotherms after extension of 120 and 211 km, respectively, for dashed area in A. Model layers denote upper and lower crust, strong frictional upper mantle lithosphere, ductile lower lithosphere, and ductile sublithospheric mantle. Scaling of quartz viscosity makes upper three layers frictional plastic. Note symmetric extension. Color versions at: http://is.dal.ca/~huismans/publications.htm.

**DISCUSSION**

Our results, together with others, indicate three fundamental rift modes for this type of model (Fig. 5). The modes are (1) SS, fully symmetric rifting, implying symmetry in both upper and lower lithosphere extension (models 1 and 4); (2) AA, fully asymmetric rifting, implying asymmetry of both layers (model 2, phase one); and (3) AS, asymmetric upper lithosphere rifting concomitant with symmetric lower lithosphere rifting (models 2 and 3, second phase).

Strain softening is the fundamental cause of asymmetry and all models are symmetric when there is no strain softening. However, the degree of asymmetry accompanying strain softening depends on other factors, and it is important to understand how softening interacts with other model properties and processes.

**Frictional Plastic Strain Softening Versus Evolving Geometry**

The primary factor governing asymmetry in a purely frictional plastic model is the feedback loop between strain softening and localized shear, but this may be opposed or reinforced by the evolving geometry. Geometrical hardening occurs when the model evolves so that the active shears are no longer geometrically viable. This type of hardening opposes continued increase in asymmetry and, depend-
ing on the amount and range over which strain softening occurs, weak shear zones will lock and be abandoned. Symmetry is favored when shears are abandoned early and the feedback loop is broken. In contrast, the evolution may lead to geometrical softening, e.g., the necking instability when a plastic layer thins. Such geometrical softening reinforces the intrinsic material softening.

Frictional Plastic Versus Ductile Material Distributions

A second factor is the extent to which the distribution of frictional plasticity geometrically dominates the model lithosphere. In the coupled models the entire crust and uppermost mantle are frictional plastic and strain softening efficiently creates asymmetry. In contrast, in the decoupled models the thick upper frictional plastic layer is replaced by a laminate containing a lower crustal ductile core, which shows no preference for asymmetry. Strain-softened upper crustal shears sole out in the lower crust and no longer force asymmetry as effectively. A continuum of behavior exists between models dominated by frictional plasticity, which exhibit the most asymmetry, and those where geometrically larger ductile regions promote symmetry.

Frictional Plastic Versus Ductile Strength Distributions

A third factor is the relative strengths, as opposed to geometries, of the plastic and ductile regions. Ductile strength increases with strain rate, i.e., higher stresses at higher rifting velocities, whereas the frictional strength is independent of strain rate. This strain-rate dependence of strength explains the difference between models 2 and 4. During phase one of model 2 the asymmetric shear originating in the frictional plastic layers projects into the lower lithosphere, forcing distributed ductile shear (Fig. 2A); plastic asymmetry dominates over ductile symmetry. In contrast, in model 4 the corresponding forced shearing is largely absent (Fig. 4A) because the strain rate, effective viscosity, and strength of the lower lithosphere are all higher. The strong ductile region actually suppresses the asymmetric deformation preferred by the overlying frictional plastic layer; ductile symmetry dominates over plastic asymmetry.

During phase two in both models the lower lithosphere necks, the preferred symmetric mode for a ductile layer. This result indicates that even in model 2 ductile necking prevails over forced shearing in phase two, which accounts for the change of mode from AA to SS.

Thermal Evolution

A fourth factor is the thermal state of the model, which also contributes to the asymmetry by determining the relative distribution of frictional plastic and ductile regions as the model evolves. When extension is slow, the thermal Peclet number is small and the lithosphere cools as it extends, thereby continuously renewing the frictional plastic layers as the model evolves. This promotes asymmetric behavior by comparison with models that have a large thermal Peclet number.
CONCLUSIONS

The factors controlling extension and the corresponding modes are summarized in Figure 5. All models described here have a two-phase development; phase one is controlled by the frictional behavior and phase two is controlled by the ductile rheology. The asymmetric AA mode is favored by strain softening of a coupled, cold frictional upper lithosphere at low rifting velocities, whereas the AS mode occurs in the equivalent decoupled model. The symmetric SS mode is favored by the absence of strain softening or by high rifting velocities. Asymmetry of conjugate passive margins at the crustal scale has been interpreted to indicate the AA mode (Wernicke and Burchfiel, 1982; Lister et al., 1986; Mutter et al., 1989; Boillot et al., 1992; Sibuet, 1992; Louden and Chian, 1999). The occurrence of the AS mode in the dynamical models is an alternative (Lister et al., 1991).

The SS and AA modes are, respectively, similar to the pure-shear (McKenzie, 1978) and simple-shear (Wernicke, 1985) kinematic models of lithospheric extension. In dynamical models frictional plastic strain softening can control which of these two styles is selected. This same mechanism may operate in natural systems. Our work differs from the Buck (1993) and Lavier et al. (1999) models in that the amount of strain softening is greater than that owing to loss of cohesion. Fault zones may also be weakened by static or transient high fluid pressures and mineral transformations (Sibson, 1990; Streit, 1997), and by the preponderance of phyllosilicates in deformed polymineralic zones (Bos and Spiers, 2002). The results shown here are, however, model dependent and depend on the model mesh resolution. They demonstrate the feasibility of the mechanism, but more accurate, higher resolution model experiments are needed.

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