Extension of stable continental lithosphere and the initiation of lithospheric scale faults

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Abstract. We address the physical conditions which control the style of continental extension. Geological evidence suggests that once lithospheric scale zones of localized deformation have been formed, they strongly affect continental deformation. It is the purpose of this paper to investigate mechanisms which may cause lithospheric scale faults to initiate in stable continental lithosphere which is laterally fairly homogeneous. Faults and shear zones cutting strong layers in the lithosphere will have a very significant influence on the evolution during extension. Based upon experimental flow laws, a strength maximum can be expected in the mantle directly beneath the Moho. Strain localization in the shallow upper mantle is therefore expected to have a very pronounced effect on the evolution of the extending lithosphere. Low viscosities in the lower crust decouple the crust mechanically from the upper mantle. Therefore, causes for strain localization in the sub-Moho mantle must be found in the mantle itself. Two potential causes satisfying this requirement are boudinage and strain weakening. We use thermal-mechanical finite element models which incorporate the elastic, visco-plastic, and viscous response of lithospheric rocks. The results of our model experiments suggest that boudins do not evolve during extension of continental lithosphere in most situations; only when the extension rate is fast relative to thermal reequilibration may homogeneous boudinage result. By its very nature, however, this type of boudinage cannot produce lithospheric scale faults. Our model results suggest that shear zones may evolve after strain weakening. However, the style of extension on the scale of the lithosphere is pure shear like because the shear deformation in localized zones is balanced; in most cases the shear zones occur in conjugate pairs. Initiation of lithospheric scale faults is concluded to be unlikely in stable and homogeneous lithosphere in interior parts of continental plates.

Introduction
The lithospheric stretching model which was originally proposed by Artemjev and Artyushkov [1971] and analyzed by McKenzie [1978] and the simple shear model of Wernicke [1981, 1985] represent end members of a range of models which describe the kinematics of continental extension. With these models, and intermediate derivatives from them [Royden and Keen, 1980; Kasznir and Egan, 1989; Lister and Davis, 1989], it appeared possible to sketch the first-order kinematics of extending continental lithosphere. However, interpretations are not always unique; in the North Sea basin, for instance, both the simple shear and the pure shear stretching model are advocated [Gibbs, 1984; Beach, 1986; Gibbs, 1987; Klemperer and White, 1989; White, 1989; Latin and White, 1990]. In some basins the pure shear model seems a reasonable description of the extensional strain of the underlying lithosphere, whereas the simple shear model is a reasonable description of the kinematics of other basins. Thus the problem of lithospheric extension is to understand what conditions determine the mode of extension.

Once lithospheric scale zones of localized deformation have been formed, they tend to stay operative in subsequent tectonic phases [White et al., 1986]. Lister and Davis [1989] argue that weak zones cutting rheologically strong layers control the evolution of extending lithosphere. Therefore one of the central issues related to the problem of lithospheric extension is to understand the conditions under which lithosphere-cutting faults are initiated. One possibility is that these weak zones grow along lateral material discontinuities, which may for instance be the result of accretion processes. We do not exclude the possibility that strain often tends to get localized along material boundaries, but it is imperative to note that there is no compelling evidence that it always does so. For instance, localized strain in mantle peridotites, which might be related to lithospheric scale faults, is not associated with lateral inhomogeneities (R. L. Vissers, personal communication, 1993). Therefore in this paper we investigate the conditions for strain localization in the upper mantle of continental lithosphere which is laterally homogeneous. We focus on stable continental lithosphere, i.e., lithosphere which thermally and mechanically is close to equilibrium, implying a steady state geothermal gradient and strain rates lower than 3 \times 10^{-6} \text{s}^{-1} (1% strain in 100 million years (m.y.)). It is clear that lithosphere which has been subject to a recent thermal and/or tectonic event is not stable, since stresses have not yet equilibrated to the new boundary conditions and temperatures are transient. The problem of continental extension in the context of nonstable lithosphere is investigated by Govers and Wortel [1993].

We put special emphasis on the region just beneath the Moho because the upper mantle has the largest load-bearing capacity [Kirby and Kronenberg, 1987, and references therein]; localization of strain in upper mantle rocks will undoubtedly have very prominent effects on the evolution of the extending lithosphere. Causes for breaking the strong sub-Moho layers must be sought in the mantle itself, since the upper mantle and lower crust are expected to be mechanically decoupled by low viscosities in normal lower crust at an average geothermal gradient. We investigate two mechanisms which have been proposed for strain localization in continental lithosphere: boudinage and strain softening.

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Several studies predict that (parts of) the lithosphere will boudinage during extension [Fletcher and Hallet, 1983; Zuber et al., 1986; Ricard and Froidevaux, 1986; Martinod and Davy, 1992; Fleitout et al., 1995]. Boudinage has been put forward to explain observations of structures which exhibit spatial periodicity, both in oceanic [Morceau and Fleitout, 1989] and continental lithosphere [Fletcher and Hallet, 1983; Froidevaux, 1986; Ricard and Froidevaux, 1986]. The nonhomogeneous deformation associated with pinch and swell structures could be relevant in the context of initiating lithospheric scale faults. We will therefore investigate large-deformation models of continental extension under conditions which are predicted to be favorable for the development of boudinage.

Large-deformation models of continental extension [Houseman and England, 1986; Braun and Beaumont, 1987, 1989a, b; Dunbar and Sawyer, 1987; Bari, 1991; Christensen, 1992] are based upon steady state flow laws which relate composition, strain rate, temperature, and stress at a constant microstructure. The assumption of constant microstructure, i.e., that material parameters like grain size, preferred orientation, and metamorphic phase are constant, is very important. In large-strain models, the effect of variations in microstructure become prominent because a strain-induced change of deformation mechanism is likely to occur [White, 1973, 1976, Kurato et al., 1986, Brodie and Rutter, 1987, Rutter and Brodie, 1988, Handy, 1989; Drury et al., 1991]. One especially important consequence of a nonconstant microstructure could be that deformation may localize in pervasive zones instead of being controlled by the rheology of a significant volume of lithospheric rocks, the overall strength of the lithosphere would be determined by the rheology of a minor part of the volume in this case. As the consequences for strain localization for the problem of lithospheric extension are clear, we investigate whether strain weakening may result in initiation of lithospheric scale faults.

Boutillier and Keen [1994] recently published results from finite element models of extension of lithospheric extension which a priori contain large crustal faults. The models are extended by imposing a horizontal velocity to lateral model boundaries which is constant with depth. Interestingly, their results indicate that shear zones evolve in the mantle, extending from the initial fault, in some cases leading to an asymmetric style of extension on the scale of the whole lithosphere. These results are, however, a direct consequence of their initially assumed geometry, in which the initial fault was allowed to generate once but no subsequent initiation of new faults is possible. In fact, conspicuous conjugate shear systems evolve in the crust of their models, which would have been accommodated by a conjugate fault would they have allowed new faults to grow. From a numerical point of view, modeling initiation of new faults is problematic, however. Therefore we opt to not include preexisting faults and to look for regions where strain becomes localized in response to extension.

**Thermal-Mechanical Model and Finite Element Approach**

To model large-deformation continental extension, we employ a finite element method to obtain solutions to the coupled differential equations which control the mechanical and thermal evolution of the lithosphere. The selected mechanical finite element code, TECTON, developed by Melosh and Raufsky [1980] (see also Melosh and Raufo [1981, 1983] and Melosh and Williams [1989]), solves the mechanical equilibrium equations for nodal displacements by linear shape functions using an implicit time stepping scheme. We have supplemented the code with a thermal finite element program which solves for nodal temperatures by linear shape functions on the same grid with an implicit scheme to forward the solutions with time.

We assume plane strain, i.e., that the horizontal extent of our models perpendicular to the cross-sectional plane is large compared to the scale of the model in the plane of the figures (e.g. Figure 1). The models are symmetric around a vertical plane on the left-hand boundary of the model; for reasons of economy, calculations are performed only in the right-hand side of the model (Figure 1). Mechanical and thermal boundary conditions on the left-hand boundary are formulated to reflect the symmetry: hori-

![Figure 1](image.png)

**Figure 1.** Generalized model of continental lithosphere. For reasons of economy only the right half is modeled. Mechanical and thermal boundary conditions on the left side reflect symmetry. The model is assumed to continue to the right infinitely. This is reflected in the mechanical and thermal boundary conditions of no tilt and no horizontal heat flow. The top of the lithosphere is at 30°C, and a constant basal heat flow \( q_b \) is applied to the lower boundary. Isostatic rebound forces act on density interfaces, and a uniform force is applied to the right boundary. See text for explanation.
ztional displacements and horizontal heat flux are zero on this side. Only results in the right-hand side of the model will be displayed.

We use the "density-stripping" method [Braun, 1988; Williams and Richardson, 1991] to calculate the isostatic restoring pressures at density interfaces and to obtain the uplift of the surface. This method is a good approximation to solving the full mechanical equations, including gravity, in the absence of strong horizontal density gradients [Braun, 1988]. A small density inversion of 50 kg/m² is modeled on the lithosphere/asthenosphere boundary. Thermal boundary conditions on horizontal interfaces are 0°C temperature at the surface and a constant basal heat flux on the lower boundary of the model. The boundary conditions on the right-hand side of the model reflect the assumption that the plate continues to the right; the horizontal heat flow is zero and, mechanically, this is a no-tilt boundary.

The continental lithosphere is modeled as a three-layer system, upper crust, lower crust, and mantle. Mechanically, the model lithosphere is a continuum which can achieve permanent strain by thermally activated viscous flow or by visco-plastic failure, which is selected to approximate Byerlee's [1978] brittle strength using a von Mises yield criterion. The rheology is elastic, viscous, or visco-plastic depending on the composition, pressure, stress, and temperature. In the upper crust we assume a "wet" quartzite viscous rheology [Paterson and Luan, 1990], which yields a viscosity which is intermediate among current experimental rheological laws. For the lower crust we use the "wet" Adirondack granulite rheology [Wills and Carter, 1990], which gives a low viscosity relative to other literature values, and for the mantle we assume a low-viscosity end-member, wet olivine [Kutter and Brodie, 1988] (Table 1). The steady state flow laws we adopt are the result of extrapolation of empirical results and should therefore be considered as first-order estimates of the stresses at specific strain rates. For calculation of (initially steady state) geotherms we use the thermal parameters suggested by Chapman [1986]. Pressure and temperature dependence of the upper crustal conductivity is included by selecting average conductivities which are parameterized in terms of the initial surface heat flow.

Boudinage

The topography in one of the world's best studied extended continental regions, the Basin and Range Province of the western United States, exhibits a distinct periodicity with wavelengths ranging between 30 and 50 km [Zoback et al., 1981]. This regular pattern was explained by Fletcher and Halley [1983] in terms of boudinage of the upper crust. On the basis of observed Rouquier anomaly patterns, Froidevaux [1986] suggested that the mantle beneath the Basin and Range Province exhibits pinch-and-swell structures with a 200-km wavelength and ascribed this to mantle boudinage during extension. The conceptual model suggested by these studies therefore predicts simultaneous crustal and mantle boudinage during extension, with different wavelengths. Figure 2 indicates how out-of-phase boudinage can potentially be important for the problem of lithospheric extension; pinched crustal and mantle regions could link up into a lithospheric scale zone of localized deformation.

Physically, boudinage is caused by instabilities in the displacement field in the direction perpendicular to extension. As a result, a small perturbation in the initial displacement field grows exponentially during extension. Potential causes for stretching instabilities in continental lithosphere are (1) thickness variations in a lithosphere subject to a tensile load, (2) for non-Newtonian materials, viscosity variations in combination with strain rate variations, and (3) the density inversion at the lithosphere-asthenosphere boundary. The reason why these three mechanisms can potentially cause unstable vertical displacements during extension is that they all have a positive feedback character. The first cause can be understood if one thinks of a beam of variable thickness to which a tensile force is applied. Stresses in portions of the beam with a smaller cross section will be higher than stresses in thicker parts of the beam. As a result, thin portions of the beam will experience more strain (i.e., extension) than thick parts and therefore will thin more relative to thicker parts of the beam. This, in turn, will further focus stresses. Note that stress-induced thickness variations may lead to boudinage in elastic and Newtonian viscous materials. The second cause for stretching instabilities can be understood by examining the nature of nonlinear viscous flow laws, typically, strain rate is proportional to stress to some power greater than 1. As a result, a local increase in stress leads to an increase in strain rate and to increased thinning. Localized thinning leads to stress focusing and subsequently causes an increase in strain rate. The third cause, finally, is the classic example of a mechanical instability; a small uplift of the density interface generates buoyancy forces which further amplify the uplift, and so on.

The inverse dependence of viscosity on the temperature is an important factor for the potential of the lithosphere to boudinage. Lithospheric extension typically leads to steepening of the geotherm due to advection followed by a phase of cooling. A consequence of nonuniform extension is that more extended portions of the lithosphere cool while (transiently) heating adjacent less extended regions. Thermal conduction has the effect of delocalizing extension, i.e., smearing out the deformation over larger regions. For instance, extension of the mantle and no extension of the overlying crust is followed by cooling of the mantle and a transient increase of temperatures in the crust. This causes a viscosity increase in the mantle and a viscosity drop in the crust which may lead to a shift in the locus of extension from mantle to crust during ongoing extension. It is therefore to be expected that thermal conduction counteracts boudinage.

Various model studies have investigated the conditions leading to stretching instabilities [Ricard and Froidevaux, 1986; Zuber et al., 1986; Bassi and Bonnin, 1988; Martinod and Davy, 1992; Fleitout et al., 1995]. Typically, these studies investigate the effect of one or two of the above three causes for boudinage. An often used method is to use analytical solutions derived from perturbation theory for various rheologies. Disadvantages of this approach are that the results hold only for small deformations and that, in most cases, elasticity is neglected. Our aim is to circumvent these limitations by using a large-deformation finite element approach in order to investigate the significance of boudinage for the problem of continental extension. We will present the results of three numerical simulations which are intended to highlight various pertinent aspects of lithospheric boudinage. This approach enables us to make a direct comparison with previous studies and to gain insight into the relative importance of the aforementioned physical mechanisms for boudinage.
Table 1. List of Parameters

<table>
<thead>
<tr>
<th>Quantity</th>
<th>Symbol</th>
<th>Value</th>
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<tbody>
<tr>
<td>Young’s modulus</td>
<td>$E$</td>
<td>$1 \times 10^{11}$ Pa</td>
</tr>
<tr>
<td>Poisson ratio</td>
<td>$\nu$</td>
<td>0.25</td>
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<tr>
<td>Specific heat at constant pressure</td>
<td>$c_p$</td>
<td>$1300$ J kg$^{-1}$ K$^{-1}$</td>
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<td>Volumetric thermal expansivity</td>
<td>$\alpha$</td>
<td>$3.2 \times 10^{-3}$ K$^{-1}$</td>
</tr>
<tr>
<td>Coefficient of brittle friction</td>
<td>$\mu$</td>
<td>0.8</td>
</tr>
<tr>
<td>Pore fluid pressure</td>
<td>$P_f$</td>
<td>hydrostatic</td>
</tr>
<tr>
<td>Gravitational acceleration</td>
<td>$g$</td>
<td>$9.8$ m s$^{-2}$</td>
</tr>
<tr>
<td>Power law</td>
<td>$\dot{\varepsilon}_y = A_p \sigma_y^{n_p} \exp \left( \frac{Q_p}{RT} \right) \sigma_y'$</td>
<td></td>
</tr>
<tr>
<td>Strain rate tensor component</td>
<td>$\dot{\varepsilon}_{ij}$</td>
<td>$\sigma_{ij}$</td>
</tr>
<tr>
<td>Deviatoric stress tensor component</td>
<td>$\sigma_y'$</td>
<td>$(\frac{1}{2} \sum_{ij} \sigma_{ij}' \sigma_{ij}')^{\frac{1}{2}}$</td>
</tr>
<tr>
<td>Effective stress</td>
<td>$\sigma_y$</td>
<td>$(\frac{1}{2} \sum_{ij} \sigma_{ij}' \sigma_{ij}')^{\frac{1}{2}}$</td>
</tr>
<tr>
<td>Effective strain</td>
<td>$\varepsilon_y$</td>
<td>$(\frac{1}{2} \sum_{ij} \varepsilon_{ij}' \varepsilon_{ij}')^{\frac{1}{2}}$</td>
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<tr>
<td>Universal gas constant</td>
<td>$R$</td>
<td>$8.3144$ J mol$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>Temperature</td>
<td>$T$</td>
<td></td>
</tr>
<tr>
<td>Diffusion creep law</td>
<td>$\dot{\varepsilon}<em>y = A</em>{dc} \sigma_y^{n_d} \exp \left( \frac{Q_{dc}}{RT} \right) \sigma_y'$</td>
<td></td>
</tr>
<tr>
<td>Grain size</td>
<td>$d$</td>
<td></td>
</tr>
<tr>
<td>Upper crust: wet quartzite$^*$</td>
<td>$n_{pl}$</td>
<td>3.1</td>
</tr>
<tr>
<td>power law sensitivity</td>
<td>$Q_{pl}$</td>
<td>135 kJ mol$^{-1}$</td>
</tr>
<tr>
<td>power law activation enthalpy</td>
<td>$A_{pl}$</td>
<td>7.7621 $\times 10^{-7}$ Pa$^{-1}$ s$^{-1}$</td>
</tr>
<tr>
<td>mass density</td>
<td>$\rho_y$</td>
<td>2876 kg m$^{-3}$</td>
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<tr>
<td>volumetric heat production$^*$</td>
<td>$H_{110}$</td>
<td>1.37 $\mu$W m$^{-1}$</td>
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<tr>
<td>thermal conductivity$^*$</td>
<td>$k_{110}$</td>
<td>2.56 W m$^{-1}$ K$^{-1}$</td>
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<td>Lower crust: granulite$^4$</td>
<td>$n_{pl}$</td>
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<td>$Q_{pl}$</td>
<td>243 kJ mol$^{-1}$</td>
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<tr>
<td>power law activation enthalpy</td>
<td>$A_{pl}$</td>
<td>9.5534 $\times 10^{-7}$ Pa$^{-1}$ s$^{-1}$</td>
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<tr>
<td>mass density</td>
<td>$\rho_y$</td>
<td>2876 kg m$^{-3}$</td>
</tr>
<tr>
<td>volumetric heat production</td>
<td>$H_{1c}$</td>
<td>0.45 $\mu$W m$^{-3}$</td>
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<tr>
<td>thermal conductivity$^5$</td>
<td>$k_{1c}$</td>
<td>2.60 W m$^{-1}$ K$^{-1}$</td>
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<tr>
<td>Mantle: wet olivine$^3$</td>
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<td>3.0</td>
</tr>
<tr>
<td>power law sensitivity</td>
<td>$Q_{pl}$</td>
<td>420 kJ mol$^{-1}$</td>
</tr>
<tr>
<td>power law activation enthalpy</td>
<td>$A_{pl}$</td>
<td>8.5746 $\times 10^{-15}$ Pa$^{-3}$ s$^{-1}$</td>
</tr>
<tr>
<td>diffusion creep activation enthalpy</td>
<td>$Q_{dc}$</td>
<td>240 kJ mol$^{-1}$</td>
</tr>
<tr>
<td>diffusion creep pre-exponent</td>
<td>$A_{dc}$</td>
<td>0.6253 Pa$^{-1}$ s$^{-1}$</td>
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<tr>
<td>mass density</td>
<td>$\rho_y$</td>
<td>3300 kg m$^{-3}$</td>
</tr>
<tr>
<td>volumetric heat production</td>
<td>$H_m$</td>
<td>0.02 $\mu$W m$^{-3}$</td>
</tr>
<tr>
<td>thermal conductivity$^6$</td>
<td>$k_m$</td>
<td>3.20 W m$^{-1}$ K$^{-1}$</td>
</tr>
<tr>
<td>Asthenosphere: mass density</td>
<td>$\rho_a$</td>
<td>3250 kg m$^{-3}$</td>
</tr>
</tbody>
</table>

$^*$ Paterson and Luan [1990].

$^4$ For initial surface heat flow of 60 mW m$^{-2}$.

$^5$ Wilks and Carter [1990].

$^6$ Rutler and Brodie [1988].

### Stretching Instabilities in the Absence of Thermal Conduction

Most of the previous studies on stretching instabilities neglect the effects of both viscosity variations resulting from deformation-induced thermal advection and thermal conduction. An exception is the study by Fiebig et al. [1995], in which thermal advection is accounted for. In order to mimic the results of these studies, as far as this is possible in the large-deformation approach, we investigate the evolution of continental lithosphere with unrealistically low thermal conductivities. Any material point has a temperature which stays constant with time and, consequently, viscosity changes induced by temperature changes do not occur in this model. Displacement of material points during extension causes thermal advection. This model should be seen as one which maximizes temperature gradients and, accordingly, lateral viscosity gradients. Consequently, our model should exhibit a stronger tendency to mechanical instability than the models from previous studies which did not include thermal advection.
Initially, the model lithosphere consists of a 15-km-thick upper crust, a 15-km-thick lower crust, and a 45-km-thick lithospheric mantle (Figure 3). Thermally, the initial model is in equilibrium with a surface heat flow of 60 mW/m² (Table 1). The initial Moho temperature is 473.3°C. The selected rheologies promote the development of out-of-phase boudinage. A modest density inversion of 50 kg/m³ is modeled at the lithosphere-asthenosphere boundary. Random perturbations with maximum peak-to-trough amplitudes of 500 m are added to the initial vertical position of density interfaces, i.e., to the free upper surface, to the Moho, and to the lithosphere-asthenosphere boundary. Figure 4a shows the initial spectrum of the Moho. In order to facilitate comparison between linear stability models and our modeling results, we adopt a constant strain rate boundary condition on the right-hand side of our model. The model is extended at a constant horizontal strain rate of $3 \times 10^{-15}$ s⁻¹ (i.e. a relative velocity of 6 cm/yr initially).

Figure 5 shows the deformed finite element grid 6 m.y. after stretching started. It is clear that a harmonic deformation pattern has evolved with a dominant wavelength of approximately 200 km (compare Figure 4b). This style of deformation is called homogeneous boudinage by Marinov and Davy [1992]. They predict that homogeneous boudinage will not occur during extension of continental lithosphere. Homogeneous boudinage is, however, consistent with the results of Ricard and Froidevaux [1986] and Zuber et al. [1986]. The cause for boudinage must be thickness variations resulting from stress concentrations, viscosity variations caused by strain rate variations for non-Newtonian materials, or the density inversion at the lithosphere-asthenosphere boundary. To investigate the effect of the density inversion, we reran the finite element model without the density inversion; we found that the results are virtually identical. However, in the density-stripping approximation which is adopted in this paper, horizontal components of buoyancy forces associated with the dis-
placement of density interfaces are neglected. Therefore our models tend to underestimate the effects of density-driven instabilities. It is, in fact, very likely that the effects of a density inversion work on longer timescales. However, since we find no indication of any effects related to the density inversion on timescales investigated here, we are inclined to conclude that potential cause 3 is not relevant to boudinage on lithospheric scale. More probably, the occurrence of boudinage in this model is the combined effect of stress-induced thickness variations and strain-rate-induced viscosity variations.

**Stretching Instabilities and Temperature Variations**

The next model to consider is identical to the first one except that we do allow thermal conduction and the associated variations in viscosities to occur. The results (Figure 6) are strikingly different, a single necked region rather than a periodic deformation pattern develops. Clearly, thermal conduction has an opposing effect on boudinage. Probably, the reason why deformation localizes in a single region instead of being distributed over a number of pinched zones is that strain rates are higher in a single necking region (at the same extension rate). At higher strain rates, thermal conduction is less relevant, so that the change in deformation style from boudinage to a single necking region can be seen as a response of the lithosphere to counterbalance the (stabilizing) effect of thermal conduction.

The relative importance of thermal conduction decreases with increasing extension rate. This is illustrated in Figure 7, which shows the deformed grid of a finite element model extended at twice the strain rate of the previous model (12 cm/yr initial velocity). After 5 m.y., homogeneous boudinage has again occurred. Qualitatively, the effects of decreasing the brittle strength or increasing the initial geothermal gradient are expected to be similar; both will tend to decrease the relative importance of viscous deformation and therefore of the counteracting effects of cooling. In general, neglecting the (strong) temperature-induced viscosity changes during lithospheric extension [Ricard and Froidevaux, 1986; Zuber et al., 1986; Bassi and Bonnin, 1988; Martinod and Davy, 1992] yields results which are hardly indicative for the style of finite deformation.

A source of discrepancies between the response to lithospheric extension (our study) and the results of models based upon linear

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**Figure 4.** (a) Initial amplitude spectrum of the Moho. (b) Amplitude spectrum of the Moho after 6 m.y. of extension.

**Figure 5.** Deformed finite element grid of the boudinage model without thermal reequilibration, after 6 m.y. and 229 km of horizontal extension. Effective stresses \( \sigma_E \) are shown as patterned curves, displaying regions of plastic and viscous flow.
stability analysis, which thus far has received little attention, is the neglect of elasticity in the latter. Elastic fiber stresses in strong parts of the lithosphere strongly inhibit vertical motions during extension (in compression, vertical motions will be amplified more). Given the fact that the mantle immediately beneath the Moho is probably the strongest part of continental lithosphere, the common observation in extended regions of a relatively flat Moho is not surprising. To illustrate this point, we show in Figure 8 the results of a model which is identical to the model in Figure 5 (no thermal conduction) except that we decreased to elastic strength by selecting $E = 10^9$ N and $\mu = 0.05$. Clearly, the boudinage-like response which followed extension of the model in Figure 5 is absent in Figure 8, and we note a tendency to boudinage with increasing elastic strength. In the limit to infinite elastic strength, boudinage can therefore be expected. Visco-plastic models can be viewed as elasto-visco-plastic models with infinite elastic strength, meaning they show no elastic deformation as a result of an imposed stress. Consequently, visco-plastic models can be expected to exhibit boudinage where elasto-visco-plastic models do not.

Returning to the problem of continental extension, we find that in few cases homogeneous boudinage occurs and that a single necked region is more common. The important point within the context of this study is that out-of-phase boudinage, as predicted by various linear stability analyses, does not occur. Since we did not perform a full study of lithospheric boudinage for all possible parameter values, we conclude that on the basis of insights gained from our few model experiments, boudinage probably is not a viable mechanism for initiation of lithospheric scale faults in stable continental lithosphere.

**Strain Weakening**

Hobbs et al. [1990, p.143] note that "it is a prominent theme in both the materials and the geological literature that strain softening... is a necessary and sufficient condition for localization of deformation." This statement is not undisputed since it has been noted by others that strain softening is not a sufficient condition for shear localization (see discussion and references of Drury et al. [1991]). In this section we investigate whether strain softening in the mantle portion of stable continental lithosphere leads to localized deformation and initiation of lithospheric scale faults.

**Potentially Important Strain Localization Mechanisms**

Drury et al. [1991] give an overview of mechanisms which may lead to localized deformation. In the context of this paper, we will not consider shear localization induced by the imposed

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**Figure 6.** Deformed finite element grid of the boudinage model with thermal reequilibration, after 8 m.y. and 340 km of horizontal extension. The effect of allowing temperature variations to occur is that no boudinage develops; rather, a single necked region develops.

**Figure 7.** Deformed finite element grid of the boudinage model with thermal reequilibration, extended at double the strain rate of the previous two models. The figure shows the mesh after 5 m.y. and 473 km of horizontal extension. The effect of the higher extension rate is that thermal reequilibration is suppressed, resulting in homogeneous boudinage.
physical conditions, i.e., boundary conditions or material properties, during experimental deformation of rocks. Localization resulting from imposed mechanical boundary conditions may potentially be important, as was shown by the noncoaxial deformation experiments of Fransson and Spiers [1990]. The amount of experimental evidence is limited, however. We review only the localization mechanisms known as "intrinsic", i.e., mechanisms which lead to localized deformation even in rigidly confined, materially and thermally homogeneous and isotropic rocks.

The best known localization mechanism is brittle failure. The importance of brittle failure in the lithosphere as a strength limiting mechanism is generally acknowledged. The large number of brittle faults, especially in the continental crust, allows brittle failure to be approximated by continuum deformation on a lithospheric scale in most cases.

Deeper in the lithosphere, at elevated temperatures, materials deform by crystal plastic or diffusion processes. Here, the strain can become localized as a result of decreasing viscosities with increasing strain ("strain softening"). Strain softening mechanisms can be divided into five main classes: (1) thermal softening as a result of shear heating, (2) reaction softening resulting from a (metamorphic) phase transformation, (3) geometric softening due to changes in the crystallographic fabric (the average orientation and shape of crystals that constitute a rock), (4) brittle softening induced by a combination of brittle and ductile processes, and (5) structural softening induced by changes in microstructure [see Drury et al., 1991, and references therein].

A distinction can be made between mechanisms which lead to localization of deformation and mechanisms which are required to keep ongoing deformation localized. In the absence of phase transformations, thermal softening is typically a mechanism that keeps deformation localized; in the absence of localized deformation, shear heating is negligible in typical tectonic processes.

Syntectonic reactions, either as pure phase transformations or as chemical reactions, probably are a very efficient means for localizing deformation [e.g., Tullis and Yund, 1985; Handy, 1989; Drury et al., 1991]. We are not aware of any experiments in which quantitative estimates are made of the amount of weakening due to syntectonic reactions, so that reaction weakening cannot be modeled.

Wenk et al. [1991] estimate the effect of geometric softening on viscosities to be less than a factor of 10. Therefore geometric softening is considered irrelevant for large-scale modeling.

Hobbs et al. [1990] and Tullis et al. [1990] discuss localization due to brittle processes in rocks in which the bulk deformation occurs by ductile mechanisms. It is conceivable that brittle softening does occur in the brittle-ductile transition zone, but, again, quantitative estimates are not available.

**Diffusion Creep**

Various workers [e.g., Rutter and Brodie, 1988; Handy, 1989] address the issue of strain localization due to a change of deformation mechanism. It is especially the experimental work of Karato et al. [1986], Brodie and Rutter [1987] and Rutter and Brodie [1988] which is important, in that some quantitative relations are established for the weakening behavior of olivine with a controlled grain size. Recently, diffusion creep has begun receiving attention in the geophysical literature [Voroney and Furlong, 1992; Hopper and Buck, 1993], but the impact of a transition from dislocation creep to diffusion creep has not yet been explored.

In Figure 9a the effects of grain size and strain (or time) on the steady state flow stress are shown. If it is assumed that the starting material initially has a grain size of 1 cm and a temperature of 600°C, it is clear from Figure 9a that recovery-controlled dislocation creep is the controlling mode of deformation at geologically significant strain rates (10^{-12} to 10^{-16} s^{-1}). The flow stress does not depend on the grain size in this field, but creep polygonization, rotation recrystallization, and possibly migration recrystallization [Drury et al., 1991] lead to a reduction of grain size with increasing strain. After "some strain" the grain size has been reduced sufficiently, and the deformation mechanism changes to diffusion creep.

Within the grain size sensitive field, the stress drop can be estimated from the rheological path proposed by Rutter and Brodie [1988]. (Handy [1989] discusses alternative rheological paths, but no quantitative estimates for the stress drop can be made from these.) It is assumed that, initially, mantle rocks deform by dislocation creep. Figure 9b shows the viscosity as function of natural
strain at various temperatures and stresses. No strain weakening occurs at low stresses or strain rates. Above a critical stress, which depends on temperature, the transition to diffusion creep will occur and induce a strong decrease in mantle viscosity. Note that the stress to initiate strain weakening is an increasing function of temperature.

The strain necessary to hit the diffusion creep field depends on the initial grain size distribution within the mantle. Neither the initial grain size distribution in the lithosphere nor the amount of strain required for the onset of diffusion creep, starting from a known grain size distribution, are known. These quantities are parameterized in our modeling by $\varepsilon_0$, the natural strain at which weakening starts. We assume that our modeling starts right before the onset of weakening, and we set $\varepsilon_0$ to 10%. In their experiments on fabric softening, White et al. [1985] observed that weakening was complete after 40% natural strain. Following Rutter and Brodie [1988], we therefore assume that the amount of strain involved in the complete transition from dislocation creep to diffusion creep is 40%, too. Therefore $\varepsilon_1$, the natural strain at which the stress drop is complete, is 50%.

In our analysis we neglect the influence of grain growth. We appreciate that grain growth is a relevant process in the context of diffusion creep. In fact, piezometric grain size relations probably reflect a balance between competing processes: grain growth and grain size reduction. Including the effect of grain growth, however, would reduce the viscosity drop upon weakening. We opted not to incorporate grain growth in order to arrive at the maximum weakening effect. Our model result should therefore be viewed as a case of maximum strain weakening.

**Strain Weakening Model**

To investigate the effect of strain weakening in the lithospheric mantle, we designed a numerical (finite element) experiment at...
conditions which enable a dramatic and rapid decrease in viscosity due to straining. The reason why we focus on strain weakening in the mantle only was discussed in the introduction. Initially, the model is close to mechanical equilibrium, i.e., stresses are so close to hydrostatic that strain rates are very low (of the order of $10^{-16}$ s$^{-1}$ or less). A modest Moho thickness variation from 40 km near the model symmetry center to 35 km is modeled in order to focus the thinning. As a result, slight deviation from mechanical equilibrium occurs in the transition zone between 50 and 75 km from the vertical symmetry center (Figure 10), where the Moho thickness decreases from left to right. Recent papers [Krause et al., 1991; Bird, 1991] emphasize that the lower crust, because of its low viscosities, is very efficient at attenuating strong lateral density gradients. A relatively mild Moho depth gradient, like in our model, can be expected in continental lithosphere that is close to mechanical equilibrium. The starting model is in local isostatic equilibrium.

Initially, the model is in thermal equilibrium. The surface heat flow varies from 63 mW/m$^2$ near the symmetry center to 38 mW/m$^2$ at the right. The Moho temperature ranges from 626°C at the base of the thicker crust to 530°C at the base of the thinner crust. We assume a uniform value of $\omega$ and $\xi$ in the lithospheric mantle. The right-hand boundary is loaded between 0 and 0.1 m.y. with a horizontal tensile force of $8 \times 10^{12}$ N/m. This force is kept constant from 0.1 m.y. outward.

Figure 11 shows the axial effective stress $\sigma_z$ and natural strain $\varepsilon$ after 0.1 m.y., 0.5 m.y. and 2 m.y. Before the onset of strain weakening, a classical Christmas tree stress profile develops. After 0.5 m.y., strain weakening has started to occur in the lithospheric mantle immediately beneath the Moho. As a result, a low viscosity channel develops between depths of 35 and 44 km. This response reflects the temperature and stress dependence of the viscosity drop associated with strain weakening (Figure 9); compared to the deeper lithospheric mantle, temperatures in the sub-Moho mantle are lower and stresses are higher. Consequently, the effect of strain weakening is most pronounced in the shallowest mantle. The stress peak at 46 km depth marks the vertical transition between mantle regions where the viscosity drop associated with strain weakening is more and less significant. The temperature at this depth is 749°C after 0.5 m.y. Stresses which were previously supported by the uppermost mantle are transferred to the crust and deeper mantle after strain weakening has started occurring. With time, the increase of stress in the deeper mantle results in the onset of strain weakening in this part of the lithosphere and in an associated widening of the low viscosity channel after 2 m.y.

Figure 12 displays the deformed finite element grid and contours of effective strain. The low-viscosity mantle channel shows up as a more strongly deformed region. The most conspicuous feature is the dextral shear zone developing in the mantle beneath the (initially) thinner crust. Contours of effective stress (Figure 13) show that the strength of this shear zone is relatively low. Closer to the model symmetry axis, conjugate sets of shear zones develop in the strain-weakened mantle. Strain weakening is apparent not only from the display of stresses but also from the rugged Moho topography; the total loss of the sub-Moho mantle strength has resulted in a drop in the elastic fiber stresses which, normally, strongly inhibit vertical motions of the Moho.

The resulting deformation patterns can most easily be understood in terms of a broken elastic sheet which is immersed in a viscous fluid. The strength of the sub-Moho mantle has not been affected by strain weakening on the right-hand side of the model lithosphere; in fact, this region is hardly strained and can therefore be thought of as a horizontal elastic sheet (three-dimensional). The in-plane tensile force acting on the right-hand model side predominantly loads the elastic sheet. As a result, the elastic sheet is pulled to the right after strain weakening near the symmetry center has occurred. Drag between the elastic sheet and the viscous medium surrounding it generates dextral shear in the mantle and sinistral shear in the lower crust. Viscous drag in the wake of the elastic sheet causes conjugate shear zones to develop in the mantle near the symmetry axis.

The important point in the context of this study is that although shear localization results from strain weakening, the deformation

Figure 10. Initial geometry of the strain weakening model. The Moho thickness increases from 35 to 40 km between $x = 50$ and $x = 75$ km. The model is, initially, in thermal and local isostatic equilibrium. A slight deviation from mechanical equilibrium occurs in the transition zone.
Figure 11. Strain weakening model, showing effective stress $\sigma_E$ and effective natural strain $\log(1 + \varepsilon_E)$ at the symmetry axis. After 0.5 m.y., strain weakening in the sub-Moho mantle begins, producing a low-viscosity channel in this region.

Figure 12. Strain weakening model, showing (part of) deformed finite element mesh and contours of effective strain $\varepsilon_E$ after (top) 2 m.y. and (bottom) 5 m.y. Strain localization does occur in the sub-Moho mantle, but it does not result in lithospheric scale simple shear extension.
Figure 13. Strain weakening model, showing contours of effective stress $\sigma_E$ after (top) 2 m.y. and (bottom) 5 m.y. A low-stress region develops in the sub-Moho mantle as a result of strain weakening near the symmetry axis.

Pattern on a lithospheric scale is best described by pure shear extension. Dextral shear in the shear zone dipping to the right is balanced by sinistral shear in the deep lower crust, and, closer to the symmetry center, shear zones occur in conjugate pairs. None of the shear zones grow into the shallow lower crust or into the upper crust. The uniform style of extension in this model reflects the fact that none of the (mechanical) boundary conditions induce nonuniformities. It was shown before [Govers, 1993] that the selected boundary conditions are appropriate representations (models) of plate boundary processes in interior parts of continental plates.

Experiments with the same model at different spatial discretizations show that the results are sensitive to the mesh density in the sense that the width of the shear zones and the strain taken up along them varies with the grid size. The overall style of extension, however, is completely insensitive to the spatial discretization. From our few model experiments we conclude therefore that we find no indication of lithospheric scale asymmetries evolving as a result of strain weakening for boundary conditions which are appropriate for interior parts of lithospheric plates.

Discussion and Conclusions

Based upon empirical ductile flow laws and realistic geotherms, a strength maximum can be expected in the sub-Moho
mantle. Strain localization in this region will undoubtedly have prominent effects on the evolution of extending lithosphere as a whole. Low viscosities in the lower crust decouple the mantle and crust mechanically so that the causes for strain localization must be found within the mantle itself. In this paper we therefore focus on mechanisms which may lead to strain localization in the upper mantle. We consider two potential causes for localization: nonhomogeneous deformation as a result of (1) boudinage and (2) strain localization due to a deformation mechanism change from dislocation creep to diffusion creep.

Our modeling of extensional boudinage was motivated by the results of analytic model studies, which predict boudinage to occur during continental extension. Potential causes for boudinage resulting from stretching instabilities during continental extension are (1) thickness variations associated with stress variations, (2) stress rate induced viscosity variations, and (3) the density inversion at the lithosphere-asthenosphere boundary. In the analytic model studies, typically, only one or two of the above causes are taken into account. Other limitations are that the results apply to small strains only and that the adopted rheologies are plastic or viscous. Our model is more general in that it can accommodate larger strains and more realistic rheologies: elastic, visco-plastic, or viscous, depending on temperature, stress, and pressure. Also, all of the three potential causes for boudinage can be taken into account. Our aim was to test the predictions of the linear stability analyses at finite deformation. On the basis of their modeling, Martinod and Davy [1992] conclude that "boudinage of opposite phase" would result from continental extension. Parameters in our finite element model are selected to promote development of this deformation mode as this might lead to initiation of lithospheric scale fault zones.

From our model results we conclude the following. Homogeneous boudinage may result in cases when continental extension occurs at stretching rates that are high enough to inhibit reequilibration. At slower stretching rates, temperature-induced viscosity changes result in a single zone of thinning rather than boudinage at a specific wavelength. The critical stretching rate above which boudinage may occur is therefore not a simple function of thermal diffusivity but also of the thermal activation energy controlling the viscosity. The results show that neglecting the temperature-induced viscosity changes, as has been done in nearly all of the analytic studies, yields predictions for the style of continental extension which are not accurate in most cases.

In a model which was designed to promote the conditions for "out-of-phase" boudinage, i.e., a model without thermal conduction, we did not observe this mode of boudinage. Our conclusions are therefore at odds with those of Martinod and Davy [1992]. In our point of view, it is also due to the fact that these authors do not incorporate elasticity in their analysis that their results are different. Elastic fiber stresses have a very strong inhibiting effect on vertical motions during extension. Therefore neglecting elasticity will yield growth rate factors which are profoundly different from a lithosphere which does have an elastic strength. On the basis of the results of our modeling, we conclude that out-of-phase boudinage on the scale of the lithosphere is probably unlikely.

Rheological instabilities resulting from strain localization can be triggered by strain softening processes. Quantitative estimates of strain weakening are available for few of the dozen weakening processes. From recent experiments [Karato et al., 1986; Brodie and Rutter, 1987], the weakening effect resulting from a transition of deformation mechanism from dislocation creep to diffusion creep can be estimated. In our modeling we have incorporated this specific strain weakening law which aims at estimating the maximum weakening effect due to this deformation mechanism transition. Strain weakening in our model occurs only in the lithospheric mantle not in the crust. The model results show that mantle shear zones do evolve due to strain weakening but that the resulting style of deformation is pure shear on the scale of the lithosphere. A conclusion we draw from this model is that for the strain softening mechanism we studied, simple shear deformation along a lithospheric scale fault is probably unlikely in stable lithosphere.

For continental lithosphere which has been subject to recent orogenic events, Govers and Wortel [1993] showed that the numerical methods employed in the present study are capable of producing lithospheric scale shear zones/faults. The fact that no large-scale faults are generated in the models presented in this paper highlights the strong dependence of the style of intraplate extension on the mechanical boundary conditions. The selected boundary conditions reflect conditions which are appropriate for interior parts of continental plates, i.e., away from plate boundaries [Kuszniir, 1982; Govers, 1993]. We conclude from our work that initiation of whole lithosphere-cutting faults/shear zones, due to either boudinage or strain weakening, is unlikely in this setting, unless it occurs as a result of causes not incorporated in the present models.

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