Initiation of asymmetric extension in continental lithosphere

R. Govers and M.J.R. Wortel

Department of Geophysics, Institute of Earth Sciences, Utrecht University, P.O. Box 80.021, 3508 TA Utrecht, The Netherlands

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ABSTRACT


The physical conditions are investigated under which the lithospheric-scale style of extension is pure shear or simple shear. We focus on the initial stages of continental extension to monitor how symmetric or asymmetric modes of extension evolve from specific tectonic conditions. Continental collision, magma intrusions and interaction between the lithosphere and the underlying mantle are investigated as sources for extension. We use a finite element method to model the thermo-mechanical evolution of continental lithosphere. Experimental flow laws are used to model the elastic, brittle, power law creep or diffusion creep rheology of lithospheric rocks. Our results indicate that if in-plane forces change from compressive to tensile immediately after a rapid mountain building phase, initiation of a lithosphere-scale detachment fault is possible. We find a strong dependence of the extensional style on the distribution with depth of residual stresses from the collision phase. This result is consistent with observations of gravitational collapse in regions, like the Aegean and the Basin and Range Province, where detachment faults have exhumed lower crustal rocks. The predicted dip direction of the fault also agrees with observations in these areas. Intrusion of magma into continental lithosphere, which is subjected to in-plane tensile forces, will cause localization of pure shear deformation. The style of deformation resulting from mantle plumes impinging to the base of the lithosphere is symmetric. Delamination of lithospheric mantle may initiate detachment faults if delamination occurs at the end of a collision phase, when in-plane forces change sign from compressive to tensile. This result also strongly depends on the assumed residual stress distribution. If delamination occurs during the mountain building phase, the style of deformation will be pure shear. Another interesting outcome from our modeling is that dramatic strain weakening as a result of a deformation mechanism change from dislocation creep to diffusion creep, reduces the tendency to strain localization.

Introduction

A large volume of the published work on continental extension has been concentrated on quantifying the stretching factor \( \beta \) of the pure shear model of McKenzie (1978a). In some basins it was found that the kinematics of deformation could be more accurately described by depth-dependent stretching factors (Royden and Keen, 1980), which introduced asymmetry near basin edges. In the simple shear model of Wernicke (1981, 1985), the asymmetry is complete as continental extension occurs along a lithosphere cutting detachment fault ("fault" is used here in a loose way to describe a high degree of localized deformation accommodated by an unspecified deformation mechanism). In terms of asymmetry, the pure shear and simple shear models represent end-members of a spectrum of models for continental extension that allow non-homogeneous deformation in some depth range (Royden and Keen, 1980; Hellinger and Sclater, 1983; Rowley and Sahagian, 1986; Gans, 1987; Kuszniir and Egan, 1989; Lister and Davis, 1989). These papers motivated work to investigate what geo-
logical or geophysical data allow the models to be discriminated (e.g., Furlong and Londe, 1986; Ruppel et al., 1988; Buck et al., 1988; Latin and White, 1990). Additionally, work has been done to determine the overall style of extension in sedimentary basins and passive margins (e.g., Sclater and Christie, 1980; Le Pichon and Sibuet, 1981; Barton and Wood, 1984; Beach, 1986; Lister et al., 1986; White, 1989; Latin and Waters, 1991).

Surprisingly, a realistic assessment of the physical conditions which determine the style of continental extension is made in a few papers only. In interior parts of the continental lithosphere, total strain rates tend to be uniform with depth, independent of the exact nature of the plate boundary forces (e.g., slab pull) that drive extension (Kusznir and Bott, 1977; Govers, 1993). Equivalently, the normal mode of intraplate deformation as a result of in-plane forces alone is pure shear. Govers and Wortel (in prep.) show that initiation of lithosphere-scale detachment faults is unlikely in continental lithosphere which, mechanically and thermally, is in equilibrium. They conclude that in-plane forces alone do not suffice to generate asymmetric deformation in lithosphere not containing pre-existing weak zones. Initiation of shallow-dipping lithosphere-scale faults therefore has to occur under different circumstances. Braun and Beaumont (1989a) and Dunbar and Sawyer (1989) investigate the style of extension in continental lithosphere containing offset pre-existing weaknesses in the crust and mantle. They conclude that the weak lower crust can act as a level of decollement between crust and mantle, along which the horizontally offset mantle and crustal weaknesses may link up.

Pre-existing weaknesses might control the initiation of detachment faults, but there is no evidence that this is always so. Also, an explanation of detachment faults in terms of pre-existing weaknesses disregards the possible relation between a pre-extension thickening period and subsequent asymmetric extension, as was suggested by Coney (1987) for the Basin and Range Province. Yin (1987, 1989) and Spencer and Chase (1989) show that stress trajectories in elastic plate-like continua which are subject to combinations of boundary conditions, agree with the low-angle nature of detachment faults. Their choice for an elastic rheology is motivated by the large deviatoric stresses which can be supported near the upper crustal brittle–ductile transition. For a more realistic rheology it can be expected that, at specific geothermal gradients, the upper crust, lower crust and mantle are decoupled to some degree by weak zones. In these cases, the results of Yin (1987, 1989) and Spencer and Chase (1989) can be interpreted to apply to these layers individually. The relevance of their models in the context of whole-lithosphere deformation is, however, more difficult to determine.

In a model with a simplified rheology, Melosh (1990) shows that principal stress trajectories flatten out in a low-viscosity layer beneath a strong layer containing a (steeply dipping) normal fault. It is argued that Yin (1989) and Spencer and Chase (1989) need to invoke boundary conditions which derive from hidden sources, but in his model, Melosh (1990) invokes an ad hoc end load; basically, he assumes that the upper and lower crust ride along on an extending upper mantle. It is, therefore, clear from the onset that shear between crust and mantle is required. For intraplate regions, the results of Kusznir and Bott (1977) and Govers (1993) show that a uniform in-plane force boundary condition is a more realistic approximation of the end load resulting from plate boundary processes. In Melosh's (1990) model it is assumed that the dip of the fault plane in strong layers is rather steep. It has been proposed (Jackson, 1987; Wernicke and Axen, 1988) that fault planes form in steep orientations, and rotate to shallow dips during extension. However, the large areal extent of denuded footwalls and relatively constant metamorphic grade of exposed footwall rocks in the Basin and Range Province of western North America (Lister and Davis, 1989), is an argument against the theory that low-angle normal faults originally formed in steeper orientations. Melosh's model is, therefore, considered not relevant in the context of initiation of an asymmetric mode of extension.

Typically, the overall style of deformation is symmetric, except when specific causes for asymmetric deformation exist. Our approach is to
identify the physical conditions which lead to asymmetric deformation and to focus on the whole-lithosphere simple shear end-member. Requirements for initiation of detachment faults can be formulated as that the overall style of deformation is asymmetric and that strain is localized in a zone of finite width. It is important to realize that localization of strain is not a sufficient requirement for asymmetric extension since, in most cases, strain will localize along conjugate fault systems and induce an overall symmetric style of extension. We, therefore, consider the boundary conditions which induce asymmetric extension as primary causes for localization. In the present paper we investigate processes which lead to conditions which are favorable for initiation of asymmetrical extension other than pre-existing weaknesses (in the following, processes which lead to conditions which are favorable for initiation of asymmetrical extension will be loosely referred to as “causes for initiation of asymmetrical extension”). We consider continental lithosphere which is not in mechanical and/or thermal equilibrium, i.e. the lithosphere recently has been subject to a tectonic and/or thermal event. Three causes for mechanical and thermal instability of continental lithosphere will be examined. The first two causes result in mechanical and thermal initial and boundary conditions within the lithosphere itself: lithospheric thickening and intrusions. The third cause results from mechanical and/or thermal interaction between the lithosphere and the underlying mantle.

Asymmetric extension: possible causes

Şengör and Burke (1978) recognize two end-members of rifting: “active” rifting, in which rifting is a result of mantle convection/lithosphere interaction, and “passive” rifting, in which rifting is a response to processes occurring in the lithosphere itself. A feature common to most rifts is volcanism, and Şengör and Burke (1978) identify the expected sequence of events for active rifting as doming—volcanism—extension, and as rifting—volcanism for passive rifting. They conclude that, at present, passive rifting is “by far more widespread” than active rifting. Continental thickening as a cause for subsequent extension is a purely passive mechanism. Thermo-mechanical interaction between mantle plumes and the overlying lithosphere is a purely active mechanism. All other causes of continental extension we consider are intermediate between passive and active mechanisms.

The potential energy contrast of regions of isostatically compensated thickened crust relative to normal continental lithosphere may generate tensile stresses (Love, 1911; Artyushkov, 1973). Continental thickening as a cause for subsequent extension has been proposed for the Altiplano in the Andes (Dewey, 1988), for the Tibetan Plateau in the Himalayas (England and McKenzie, 1982; Mercier et al., 1987), the Aegean Sea in the Mediterranean (McKenzie, 1972; Berckhemer, 1977; Le Pichon, 1983) and for the Basin and Range Province (Coney, 1987; Wernicke et al., 1987; Sonder et al., 1987). Given the evidence for major low-angle normal faults in the Aegean Sea (e.g., Lister et al., 1984; Lee and Lister, 1992) and in the Basin and Range Province (e.g., Wernicke, 1981; Davis, 1983; Wernicke, 1985) this mechanism must be considered a serious candidate for initiating lithosphere-scale faults.

Continental extension is commonly associated with intrusions and sometimes with volcanism. It is mostly unclear whether magmatism is caused by passive upwelling of thinning lithosphere or whether magma intrusions play an initiating role in continental extension. Active magmatism and intrusions may play an important role in localizing extensional deformation in pre-stressed lithosphere. We will investigate to what extent intrusions may lead to asymmetrical deformation.

The category of sub-lithospheric causes for detachment faults generates mechanical and/or thermal boundary conditions on the lower lithosphere boundary. Examples are lithospheric mantle delamination (detachment), rising or sinking plumes and buoyant subducted slabs. McKenzie (1978b) proposed that (parts of) lithospheric mantle may delaminate during thickening and sink into the mantle. Supportive evidence for this mechanism was found from numerical experiments by Houseman et al. (1981). Relative to the continental mantle, the asthenosphere is buoyant...
as a consequence of higher asthenospheric temperatures. In the delamination model, this density inversion causes (parts of) the mantle to detach from the overlying lithosphere during or after thickening. The delaminated mantle sinks into the asthenosphere and is replaced by material of asthenospheric temperatures, consequently generating isostatic uplift pressures and a drastic change in the lower thermal boundary condition.

Quantitative estimates of the effects of mantle convection on the overlying plates necessarily derive from indirect observations. Topographic swells in oceanic basins with up to 1000 meter amplitude are commonly attributed to mantle-lithosphere interaction. Crough (1978) found that the surface heat flow over these swells is 20–25% higher than in “normal” oceanic lithosphere and subsidence after the plate passed over the “hot spot”, and can be well explained by conductive cooling of a rejuvenated lithosphere. These observations strongly indicate that oceanic topographic swells are at least in part thermally supported. More recent studies do, however, not support these findings. Another effect of rising mantle plumes could be that a normal stress is applied to the base of the lithosphere (McKenzie, 1977). Dynamical pressures have an immediate effect on the surface uplift from the time a plume starts to impinge onto the base of the lithosphere, whereas a time comparable to the thermal time constant of the lithosphere is required for the thermal mechanism to generate substantial uplifts. It is difficult to separate thermal and dynamical contributions to topographic swells in oceanic plate interiors (Nakiboglu and Lambeck, 1985).

It is not our purpose to present a full study of the model parameter space. Instead, we select a few models which could lead to asymmetrical extension for each of the mechanical or thermal instability causes discussed above. Our aim is to investigate whether fairly common tectonic processes may lead to lithosphere detachment faults and asymmetric deformation. Govers (1993) discusses mechanisms which look promising for initiating lithosphere-scale faults in more detail.

**Modeling approach**

**Continental rock rheology**

Various deformation mechanisms contribute to the strain of continental rocks which are subject to stresses. The rate and relative importance of these deformation mechanisms depends mainly on temperature, pressure and composition. The rheologies which are usually considered most important in the context of whole-lithosphere deformation are elasticity, brittle failure and steady-state ductile flow. Isotropic linear elastic behavior is a good approximation of recoverable strains at low stresses (usually below one third of the uniaxial yield stress). Permanent strain at low temperatures is mainly achieved by brittle deformation. From laboratory experiments, the differential stress magnitude which is required to overcome the frictional resistance along fault planes, was found by Byerlee (1978) to be rather insensitive to rock type. This relation is not undisputed and geological and geophysical observations indicate...
that Byerlee's law provides an upper limit — perhaps significantly overestimated — to differential stresses at depths greater than 5 km. It is, however, one of the few quantitative relations available and we adopt Byerlee's law to determine the yield strength in brittle rocks, which is modeled using a visco-plastic rheology.

At higher temperatures, rocks respond by creep to applied differential stresses. In our rheological model we neglect transient creep and, based upon laboratory experiments, assume that power law dislocation creep laws provide a first order description of the strain rate at a given temperature, stress and composition. These empirical flow laws are the result of extrapolation of strain rates over many orders of magnitude and should therefore be considered as first order estimates at best. A second deformation mechanism which may be important at high temperatures is diffusion creep. Rutter and Brodie (1988) propose a quantitative deformation path along which a material, that initially deforms by power law creep, moves towards the diffusion creep field with progressive strain and becomes weaker. Govers and Wortel (in prep.) give a more elaborate discussion on strain-weakening mechanisms.

We assume that the development of specific types of continental extension can be studied in structures which are essentially two-dimensional, in the sense that the geometry and boundary conditions hardly vary in the third out-of-plane horizontal dimension ("plane strain" assumption). In this case the constitutive equations can be summarized as:

\[
\begin{align*}
\dot{e}_{xx} &= \frac{(1 + \nu)}{E} \left[ (1 - \nu) \sigma_{xx} - \nu \sigma_{yy} \right] \\
&\quad + \frac{(\sigma_E/\eta_{\text{eff}})^{n-1}}{4 \eta_{\text{eff}}} \left[ \sigma_{xx} - \sigma_{yy} \right] \\
\dot{e}_{xy} &= \frac{(1 + \nu)}{E} \left[ (1 - \nu) \sigma_{xy} - \nu \sigma_{xx} \right] \\
&\quad - \frac{(\sigma_E/\eta_{\text{eff}})^{n-1}}{4 \eta_{\text{eff}}} \left[ \sigma_{xx} - \sigma_{yy} \right] \\
\dot{e}_{yy} &= \frac{(1 + \nu)}{E} \left[ (1 - \nu) \sigma_{yy} - \nu \sigma_{xx} \right] \\
&\quad + \frac{(\sigma_E/\eta_{\text{eff}})^{n-1}}{2 \eta_{\text{eff}}} \sigma_{xy}
\end{align*}
\]

(c.f. Melosh and Raefsky, 1980) where \( \dot{e}_{ij} \) denote deviatoric strain rate components, \( \nu \) and \( E \) are the elastic Poisson ratio and Young's modulus, and \( \sigma_{ij} \) are stress tensor components. A dot indicates differentiation with respect to time. \( \eta_{\text{eff}} \) is the effective viscosity selected to approximate the relation between effective strain rate and stress for a specific deformation mechanism, material, temperature and pressure. \( \sigma_E \), the effective stress, is the second invariant of the stress tensor for plane strain:

\[
\sigma_E = \left( \frac{(\sigma_{xx} - \sigma_{yy})^2}{2} + \sigma_{xy}^2 \right)^{1/2}
\]

In this formulation incompressibility is automatically maintained during viscous flow; isotropic stresses only affect the elastic volumetric strain, viscous volumetric strains are always zero. Other useful parameters are the effective strain and effective total deviatoric strain rate. The definitions for effective strain and strain rate are similar to the definition of the effective stress. The second invariants of the stress tensor, strain tensor and strain rate tensor are useful measures of the magnitude of the deviatoric tensor. In what follows, total strain rate is referred to as "strain rate".

We model the continental lithosphere as a three-layer system; upper crust, lower crust and lithospheric mantle. In the upper crust we assume a wet quartzite power-law rheology (Patterson and Luan, 1990), the lower crust is modeled using the Adirondack granulite rheology of Wilks and Carter (1990) and in the mantle we employ the wet olivine power-law and diffusion creep rheologies from Rutter and Brodie (1988). From Figure 1 it is clear that the selected rheologies are at the low-viscosity end of the spectrum of upper crustal, lower crustal and mantle rheologies.

**Numerical model**

The above constitutive equations are incorporated in a finite-element code, TECTON, developed by Melosh and Raefsky (1980, 1981, 1983) and Melosh and Williams (1989). A displace-
ments based finite-element method is used to solve the mechanical equilibrium equation:
\[ \nabla \cdot \sigma + X = 0 \]
for body force \( X \). We have adapted TECTON to solve the transient heat conduction equation as a function of time \( t \):
\[ \rho C_p \frac{dT}{dt} = \nabla \cdot (k \nabla T) + H \]
by a finite-element method for temperatures \( T \), density \( \rho \), specific heat \( C_p \), thermal conductivity \( k \) and heat production \( H \). Overlapping thermal and mechanical finite-element grids are used to solve the heat and mechanical equilibrium equations sequentially. The differential equations are coupled via the effective viscosity and via thermal stresses and buoyancy forces. Table 1 gives parameter values which are adopted in this paper.

**Boundary conditions**

To calculate vertical surface uplifts we employ the “density stripping method” (Braun, 1988; Williams and Richardson, 1991). In this method it is assumed that gravity pre-stresses are hydrostatic and that the slope of density interfaces is small. Initial gravity body forces do not enter the system of partial equations and buoyancy forces are replaced by restoring pressures on density interfaces. Changes in density resulting from thermal expansion are included as body forces. We assume that the asthenosphere has a slightly (50 kg/m³) lower density than the overlying lithospheric mantle, as a result of higher asthenospheric temperatures.

Relative to normal continental lithosphere, continental lithosphere that has been thickened under isostatic conditions is very often in an average state of tension. The magnitude of the average differential stress can be calculated by comparing the density moment of two lithospheric columns (Artyushkov, 1973). In these calculations, it is implicitly assumed that force con-

**Table 1a**

| Powerlaw creep parameters adopted for the lithosphere and diffusion creep parameters for the lithospheric mantle |
|---|---|---|
| \( n_{pl} \) | \( Q_{pl} \) | \( A_{pl} \) |
| (kJ/mole) | (kJ/mole) | (Pa·s\(^{-1}\)) |
| Upper crust | Wet quartzite ¹ | 3.1 | 135 | 7.7621 × 10\(^{-26}\) |
| Lower crust | Granulite ² | 3.1 | 243 | 9.5534 × 10\(^{-21}\) |
| Mantle | Wet olivine ³ | 3 | 420 | 8.5746 × 10\(^{-15}\) |
| Diffusion creep grain size = 10 \( \mu \) m |

¹ Paterson and Luan (1990).
³ Rutter and Brodie (1988)
TABLE 1b
Thermal data. $H_{\infty}$ and $k_{\infty}$ are heat production and conductivity at a steady-state geotherm with a surface heat flow of 60 mW/m². $C_p$ and $\alpha$ are the specific heat and volumetric expansion coefficient, respectively.

<table>
<thead>
<tr>
<th></th>
<th>$H_{\infty}$ ((\mu W/m^3))</th>
<th>$k_{\infty}$ (W m⁻¹ K⁻¹)</th>
<th>$C_p$ (J kg⁻¹ K⁻¹)</th>
<th>$\alpha$ (K⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper crust</td>
<td>1.37</td>
<td>2.56</td>
<td>1300</td>
<td>3.2 x 10⁻⁵</td>
</tr>
<tr>
<td>Lower crust</td>
<td>0.45</td>
<td>2.60</td>
<td>1300</td>
<td>3.2 x 10⁻⁵</td>
</tr>
<tr>
<td>Mantle</td>
<td>0.02</td>
<td>3.20</td>
<td>1300</td>
<td>3.2 x 10⁻⁵</td>
</tr>
</tbody>
</table>

Contributions from sub-lithospheric mantle processes to stresses in thickened and normal lithospheric columns have equal magnitudes. This assumption is justified if the normal and thickened columns are proximate, since most mantle processes occur on large length scales and horizontal gradients of sub-lithospheric forces are, therefore, small. We calculate the average differential stress of thickened continental lithosphere relative to normal continental lithosphere. The resulting average differential stresses are included as uniformly distributed pre-stresses in thickened lithosphere. Smaller-wavelength buoyancy contributions, for instance resulting from delamination or secondary convection, can be included by separate basal pressures.

The finite-element models are symmetric about a vertical axis (Fig. 2), and thus the calculations are performed only in the right hand side of the model. Boundary conditions at the left-hand side of the actual model are formulated to reflect this symmetry; horizontal displacements and heat flux are zero on this side. We will display the results in the right-hand half of the model only. In the following, we will discuss the style of extension of the finite-element models in terms of "symmetric" and "asymmetric". We recognize that this classification might be misleading, since the models are completely symmetric around $x = 0$, as just explained. With "asymmetric extension" we intend to describe deformation on one symmetry side of the model, which is predominantly localized along a fault or shear zone with a uniform sense of shear.

Boundary conditions on the right-hand side of the models reflect the assumption that the model continues to the right; the horizontal heat flow is zero and mechanically this is a no-tilt boundary. In some of the models a uniform in-plane force is applied on the right-hand side. These forces are assumed to derive from plate boundary forces (ridge push, slab pull, etc.) which are transmitted as stresses to plate interiors. Based upon observations of depths of oceanic intraplate earthquakes, Govers et al. (1992) show that in-plane stress
magnitudes of the order of hundreds of MPa's are possible. Their study confirms model calculation results of the intraplate stress field in the Indo-Australian plate (Cloetingh and Wortel, 1985). In-plane stress magnitudes in our models are less than or equal to $1 \times 10^{15}$ N/m, which is equivalent to 100 MPa averaged over a 100-km-thick lithosphere. The upper surface is held at 0°C. We assume that a continental keel continues below the lithosphere (Anderson, 1989) in most models. The sub-lithospheric mantle is, therefore, part of the thermal boundary layer and a lower boundary condition for the heat flux is considered more appropriate than a constant-temperature lower boundary condition.

**Extension of thickened lithosphere**

As it is our aim to investigate common geological processes which may lead to lithosphere-scale asymmetric deformation, we select model parameters which promote this type of extension. In this section we study a few models of thickened continental lithosphere (TCL; nomenclature taken from Braun and Beaumont, 1989b).

The initial steady-state geotherm in the reference continental lithosphere (RCL) has been calculated for the average continental surface heat flow of 60 mW/m². The initial crustal thickness is 35 km, the Moho temperature is 548°C. Elevations are relative to the initial RCL surface.

The TCL results from instantaneous thickening of the RCL to produce a maximum laterally inhomogeneous structure. We adopt a uniform crustal thickening factor $\beta^{-1} = 2$, which is the upper limit of thickening factors found in the Himalayas. To maximize the potential energy contrast of the TCL relative to the RCL without invoking additional tectonic mechanisms such as mantle delamination, we assume that the lithospheric mantle has not been thickened and has passively been depressed. After thickening, isotherms in the thickened lithosphere are unstable for two reasons. First, heat-producing elements are concentrated in the upper crust and thickening of the crust eventually leads to a higher geothermal gradient. A second reason for thermal instability is that advective heat transport has occurred during the thickening phase. The surface elevation of the TCL relative to the RCL is calculated from local isostatic balance. Although we do not actually model the mantle beneath the TCL at depths greater than 100 km, the continental mantle is assumed to continue to greater depths.

The TCL is bounded by a transition region across which the crustal thickness returns to the reference value of 35 km. To promote lateral inhomogeneity, we select a 25 km width for the

![Fig. 3. Initial grid and geotherms of thickened lithosphere model.](image-url)
transition zone, which we consider a minimum, given that the topography of the TCL is more than 4 km (c.f. Davis et al., 1983). Figure 3 displays the initial grid and geotherms.

The strain in the models we consider in this paper will be shown to be laterally inhomogeneous. In the following discussion, we classify the strain in extended models as either symmetric or asymmetric on a rather arbitrary basis from inspection of the deformed finite element grid, and distributions of strain and strain rate. Thus far, an objective criterion which adequately describes the overall style of extension has not been found. Our approach to classifying the style of extension.

Fig. 4. Model 1 (Gravitational collapse model). Evolution of lithosphere model and temperatures.
is to look for lithosphere-scale asymmetries and, if the evidence for asymmetric behavior is vague or absent, to classify the style of extension as "pure-shear-like" or symmetric. Through figures of deformed model grids, we allow the reader to judge the symmetry or asymmetry of extension and weigh our classification. This approach obviates the problem that, although we might classify the style of extension on the scale of the lithosphere as symmetric, readers interested in the upper crust may decide for themselves that a specific mechanism leads to asymmetries in their region of interest. Anyway, would we have designed some objective criterion, it would not have been sufficient to present our results by merely classifying the style of extension on the basis of that criterion. The arbitrariness of preferring one criterion above another would be reflected in the fact that, in the grey area between pure symmetry and pure asymmetry, one criterion would classify a model as symmetric, while the other would classify it as asymmetric.

**Model 1. Gravitational collapse model**

In this model, the compressive in-plane force which thickened the crust is removed at \( t = 0 \). Figure 4 shows the evolution of the model lithosphere. The potential energy stored in the TCL does not suffice to cause significant extension in the first 10 Ma. Heating of the TCL has the combined effect of creating buoyancy forces and decreasing viscosities. This increases the rate of thinning which, in turn, increases the rate of mantle heating. It is therefore conceivable that, after approximately 20 Ma, extension accelerates to lead finally to complete rifting. Figure 5 displays surface topography, Moho temperature, strain and strain rate in the symmetry center \((x = 0)\) as a function of time. The top panel of Figure 6 shows contours of effective strain after 22 Ma. Contours of the logarithm of effective strain rates after 22 Ma are displayed in the bottom panel. As a result of lateral heat conduction, heating is most pronounced near the transition zone. Strain and strain rate maxima occur in the weak upper and lower crust near the transition zone. It is clear that the strain and strain rates are not homogeneous, a result which is hardly surprising in a laterally strongly varying structure. However, no clear asymmetry has developed and we consider the style of deformation,
therefore, as essentially symmetric. Gravitational collapse can well be described by depth dependent uniform stretching beneath the previously thickened region.

Assuming different rheologies or a different initial geotherm would alter the time constants of our modeling but not the overall style of deformation. Compared to vertical heat conduction, lateral heat conduction is relatively unimportant in heating the TCL and, consequently, heating is rather uniform. Both the driving load for thinning and changes in the viscosity structure derive from heating of the TCL and, therefore, thinning occurs mainly by pure shear. Thickening of the lithospheric mantle by the same amount as the crust would focus the heating and deformation more within the crust, but it would again be fairly uniform. Although a complete investigation of the parameter space would be required to make conclusive statements, gravitational collapse is probably not relevant for initiation of a lithosphere-scale asymmetrical mode of extension. This finding is in accordance with the results of Braun and Beaumont (1989b).

Model 2. Immediate extension of thickened continental lithosphere

In the previous model we removed the in-plane compressive force immediately after the thickening phase. In Model 2 the in-plane force changes sign from compressive to $5 \times 10^{12}$ N/m tension at $t = 0$. This force is equivalent to an in-plane stress of 50 MPa and is held constant from $t = 0$ onward.

Figure 7 shows the evolution of the extending lithosphere. The time scale of deformation is considerably shorter than in the previous model and after 4 Ma the lower crustal deformation is so large that the results are becoming inaccurate. The deep lower crust acts as a detachment level and the cold TCL mantle is pulled from underneath the thickened crust. Some 3 Ma after in-plane forces switched from compressive to tensile (Fig. 8), a dipping zone of elevated strains and strain rates cuts the whole continental lithosphere. This asymmetric structure develops in the mantle and upper crust in the downdip direction.
of the lower crustal shear zone. The conjugate zone (Fig. 8a), which develops in the TCL as a consequence of moment balance, displays more distributed strain rates (Fig. 8b) and more uniform deformation. Localized deformation along a fault dipping beneath the TCL is clearly preferred.

Localized thinning of the crust generates a foredeep (Fig. 9) which is partly flexurally supported. Braun and Beaumont (1989c) recognized that thinning may lead to buoyancy forces working on lithosphere with a finite flexural strength. They invoke this mechanism to give an explanation for flank uplifts in rift zones. Likewise, crustal thinning in our model generates downward-directed buoyancy forces which cause an upwarp in the RCL (shown by the change in slope of the surface topography at 4 Ma between 120 and 150 km) and a more significant flexural uplift of the TCL.

In this model we incorporate a strain-induced transition in the rheology of the mantle from dislocation creep to diffusion creep. Depending upon ambient temperature and strain rate, the effect of a switch to diffusion creep can induce a drop in viscosity of several orders of magnitude. As such, a deformation mechanism change from dislocation creep to diffusion creep is viewed as a reason for shear localization in the mantle (Poirier, 1980; see also references in Drury et al., 1991). Let us assume that strain weakening in the mantle starts soon after thickening, at 5% natural strain. Following Rutter and Brodie (1988), we assume that the transition to complete diffusion creep occurs over a strain interval of 40%, i.e. strain weakening is complete at a natural strain of 45%. Model 3 only differs in this aspect from the previous model. An in-plane tensile force, equivalent to a stress of 50 MPa which is uniformly distributed over the RCL thickness, starts extending the TCL immediately after thickening.

Figure 10 shows a few snapshots of the deforming lithosphere model. Mantle weakening starts after approximately 2 Ma. Comparison of Figures 7 and 10 shows that at 3 Ma, the Model 3 mantle has been stretched more in the TCL near the transition zone. Prior to weakening, mantle
strains are highest in this region and, consequently, this is the region where strain weakening starts. After 4 Ma the transition to diffusion creep is nearly complete in the mantle of the TCL. Weakening accelerates the overall rate of extension, and beyond 4 Ma the elements is arc strongly deformed and our results are no longer accurate.

The differences in style of deformation between the models with (Fig. 11) and without mantle weakening (Fig. 8) are significant. In Figure 11a, the weakening front (5%) and weakening tail (45%) contours are indicated. Before weakening starts (at 2 Ma), the style of deformation is identical to that of Model 2, i.e. a lithosphere-scale fault grows from the shearing lower crust.

After the weakening front starts sweeping the mantle, the style of deformation is better described by pure shear. The dipping zone in Figure 11a is a relict of the pre-weakening strain. The strain rates give an impression of the current style of deformation, which is pure shear at 3 Ma. At first sight this result is surprising, as strain weakening is called upon as a mechanism for localization. In the absence of weakening, shear deformation in the lower crust controls the style of deformation in the mantle beneath. After weakening has occurred, the mechanical coupling between lower crust and mantle disappears and the mantle flows by pure shear in response to in-plane tensile stresses.

We recognize that the conclusions we draw from this model do not apply to all strain-weakening mechanisms and that, even for the mechanism change to diffusion creep we do model, various assumptions were required to arrive at a suitable flow law (Rutter and Brodie, 1988). It is, however, clear from our modeling that strain weakening and shear localization are not synonymous. This conclusion is in accordance with results from a strain-weakening model of continental lithosphere which has been not been
subject to orogenic events and which, therefore, is close to thermal and mechanical equilibrium (Govers and Wortel, in prep.).

**Model 4. Influence of thickening pre-stress**

It will be the subject of a forthcoming paper (Govers, 1993) to present a more detailed investigation of model parameters and assumptions that are involved in our analysis. In the context of this paper, we need to comment upon the assumption of uniform thickening pre-stress, since it affects the results of Model 2 significantly.

By comparing the density moment of the TCL to that of RCL, the average differential stress between thickened and reference lithosphere is calculated (Artyushkov, 1973; England and McKenzie, 1982; Braun and Beaumont, 1989b). In the previous models, we incorporated this average differential stress as pre-stresses which are uniformly distributed. Figure 12a shows the initial (uniform) pre-stress at the symmetry center. Figure 12a also displays the elastic solution to the equilibrium equation, i.e. the pre-stresses in balance with gravity body forces. The "plastic" stress, indicated in Figure 12a, represents the horizontal deviatoric stress after brittle deformation (chiefly in the upper crust) and viscous deformation in low-viscosity parts of the lithosphere has occurred.

An alternative way to incorporate thickening pre-stress is to directly compare the pressure at some depth in the thickened column to the pressure at the same depth in the reference column, and to infer a stress difference from this (Le Pichon, 1983). This yields the deviatoric pre-stress as a function of depth. Figure 12b shows the resulting pre-stress distribution. Averaged over the lithospheric column, this pre-stress versus depth is the same as in Figure 12a. It is, however, important that tensile stresses in the crust are higher, and that mantle pre-stresses are compressive in this case. It is clear from the elastic stresses in Figure 12b that the pre-stresses exert an up-bending moment on the lithosphere. From the 'plastic' stresses we infer that the upper 15 km (measured from the top of the RCL) are in brittle failure.

Model 4 is identical to Model 2, with the exception that we replaced the uniform pre-stress by a depth-dependent pre-stress. Figure 13 shows the evolution of the finite-element grid, and should be compared with Figure 7. Figure 14 shows the strain and strain rate after 3 Ma, for comparison with Figure 8. In Model 2, the mantle is pulled from beneath the TCL, with the lower crust in the transition zone acting as a detachment. The combination of higher tensile pre-stress in the crust, and small tensile and compressive pre-stress in the mantle, shifts the focus of extension in Model 4 to the crust. Crustal deformation leads to more rapid attenuation of the Moho in Model 4. The overall style of deformation is therefore much more pure-shear-like (Fig. 14).

**Intrusion in extending lithosphere**

**Model 5. Instantaneous intrusion**

Let us consider the RCL with a steady-state initial geotherm and a surface heat flow of 60
mW/m². An in-plane tensile force is applied to the right-hand side of the model which is equivalent in magnitude to a 50 MPa stress, distributed evenly over the 100-km lithosphere thickness. This force is kept constant from \( t = 0 \) onward, and stresses are allowed to redistribute with depth. At \( t = 1.2 \) Ma asthenosphere material instantaneously intrudes the lower 60 km of continental lithosphere. Effectively, the mantle material in this column is replaced by the asthenospheric material, which has the same material properties as the mantle and which has a temperature of 1350°C.

The top panel of Figure 15 displays the finite-element grid at 1.2 Ma and the 450°C temperature anomaly contour. The topography of the surface above the intrusion (Fig. 16) at 1.2 Ma is negligible; the asthenospheric material penetrating the mantle up to 5 km below the Moho generates buoyancy forces which are flexurally supported mainly by the upper mantle. Tensile fiber stresses (175 MPa) supported by the upper mantle, oppose vertical motions of the Moho and the surface. The average strain rate in the in-

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**Fig. 13.** Model 4 (Extending TCL model with depth-dependent pre-stress). Evolution of deforming grid.

**Fig. 14.** Model 4 (Extending TCL model with depth-dependent pre-stress) after 3 Ma. (a) Contours of effective strain. (b) Contours of logarithmic effective strain rate.
truded column increases by nearly an order of magnitude due to thermal stresses and buoyancy forces. The intruded material cools (and solidifies) very rapidly, however, and the average strain rate drops to a value which is higher than the pre-intrusion strain rate. The topography of the Moho at 4.7 Ma (Fig. 15) is a consequence of localized extension of the intruded lithospheric column and widening of the thermal anomaly in the surrounding mantle; thinning of the intruded column causes subsidence and downward isostatic restoring forces, heating of mantle surrounding the intrusion generates buoyancy forces. It is clear from Figures 15 and 16 that the intrusion serves to localize deformation and leads to complete lithospheric failure at the constant in-plane force boundary condition which is applied in this model.

From the contours of effective strain at 4.7 Ma (Fig. 17) it is clear that, to conserve torque balance, two conjugate systems of enhanced shear develop in the lithosphere. The system is perfectly symmetric and neither of the two conjugate systems will develop preferentially. The overall style of deformation can therefore best be described by a pure shear model. The results of Braun and Beaumont (1989a) and Dunbar and Sawyer (1989) show that if some offset material weakness is added to the system, preferential development of one of the branches is promoted. A model with the intrusion occurring at 25 km from the symmetry center (x = 0) gives slightly asymmetrical results; as a consequence of the mechanical boundary conditions (Fig. 2), the slope of horizontal interfaces in the symmetry center is zero. Consequently, the flexural behavior on either sides of the intrusion is different, but the
effects of this asymmetry on the style of deformation are minor.

In the absence of offset weak zones or other causes for asymmetrical behavior, any model of vertical intrusions will illustrate the tendency of the lithosphere to overall pure shear deformation. We consider it very likely that if, at some stage, a fault (in the sense of a zone of intensely localized deformation) would develop on one branch in Figure 17 along which shear deformation occurs, the tendency to moment conservation would generate stresses on the other system to also initiate a fault. Based upon these arguments and our model results we consider intrusions alone, i.e. not in combination with causes for preferential development of one of the deformation branches, not relevant for initiating lithosphere-scale asymmetrical structures.

Mantle–lithosphere interaction

Model 6. Mantle plume

In this model we investigate asymmetric deformation caused by mantle plumes impinging to the base of the continental lithosphere. As discussed in the introduction, we consider the mechanical effects of mantle plumes only. A basal pressure is used to model the (mechanical) effect of a rising plume (Fig. 18). The plume is applied between $t = 0.2 \text{ Ma}$ and $t = 0.7 \text{ Ma}$, and kept constant onward, to a 100-km-wide zone left and right from the symmetry center ($x = 0$). The basal pressure ends abruptly at 100 km to create the maximum possible lateral pressure gradient. The magnitude of the basal pressure is derived from the observation of 1000-m amplitude oceanic swells; assuming that the swells are in local isostatic equilibrium and that uplifts are completely dynamically supported, yields a pressure of 32 MPa. The model is subject to a constant in-plane tensile force equivalent to 50 MPa, which is left to redistribute with depth before the mantle plume is applied.

Figure 19 shows the strain rate, surface topography, Moho temperature and strain at the symmetry center as a function of time. The surface topography resulting from the mantle plume is close to 1000 m, immediately after the plume is applied to the base of the lithosphere. The width of the mantle plume is larger than the flexural wavelength of the continental lithosphere and the system is close to isostatic equilibrium near the symmetry center. Four Ma after the plume was applied, the surface subsides below sea level as a result of crustal thinning. From Figure 18 it is clear that the style of deformation can well be described by pure shear. We conclude that mantle plumes are irrelevant for initiating an asymmetric mode of deformation.
Model 7. Delamination during thickening of continental lithosphere

In this model we assume that delamination occurs during a compression phase which uniformly thickened the crust by $B^{-1} = 2$ and which passively pushed down the undeformed mantle (c.f. Model 1). The lower 35 km of the mantle are assumed to detach from the lithosphere beneath the thickened region and to be replaced by asthenosphere material. We assume that sinking of the detached mantle will maintain a back flow of asthenospheric material, inhibiting downward cooling and "mantle healing". This assumption is reflected in a constant 1350°C temperature boundary condition beneath the TCL in our modeling. Like in Model 1, we use pre-stresses to model the potential energy difference between TCL and RCL. In-plane compressive forces are applied to balance the pre-stresses and to maintain the topography. Isostatic forces resulting from delamination are included as pressures working on the base of the TCL.

The combined effect of mantle heating and isostatic pressures is clear in Figure 20, which displays the finite-element grid and strain, 2.1 Ma after delamination has occurred. After about 3 Ma the mantle has become so deformed that the finite-element results are inaccurate. Figure 20 shows the onset of a "asthenospheric diapir" that rapidly "eats" its way through the continental mantle. The crust hardly deforms due to the in-plane containment forces. Effectively, the style of deformation is symmetric as a consequence of a non-deforming lower crust.

Model 8. Delamination after thickening of continental lithosphere

The difference between this model and Model 7 is that delamination occurs at the end of the
thickening phase, which is characterized by a switch of the in-plane forces from compressive to a tensile force which is equivalent to an average stress of 50 MPa. Figure 21 shows the deformed grid, the strain and strain rates, 2.1 Ma after delamination occurred and stretching started. A zone of intense deformation has developed from the lower crust which dips into the mantle. The style of deformation near the base of the lithosphere is pure shear. As the lithosphere–asthenosphere boundary rises the strain maxima link up (Fig. 22) to a lithosphere-scale detachment fault.

The dependence of the results from Model 8 on the distribution of thickening pre-stresses is, like in the case of Models 2 and 4, significant. In a forthcoming paper (Govers, 1993) we present a detailed investigation of factors which control the symmetry or asymmetry of extension after continental thickening and delamination. We show that the style of extension is much more symmetric when residual stresses from thickening are distributed non-uniformly with depth.

Discussion

In intraplate regions, there is a strong tendency for total strain rates to develop which are uniform with depth (Kusznir and Bott, 1977; Kusznir, 1982; Govers, 1993; Govers and Wortel, in prep.), or equivalently, the normal mode of intraplate deformation is pure shear. Initiation of shallow-dipping lithosphere-scale faults, therefore, has to occur under special circumstances. Pre-existing weaknesses might control the initiation of detachment faults, but there is no evidence that strain always localizes along mechanical discontinuities. Mechanisms for initiation of translithospheric detachments based upon pre-existing weaknesses lithosphere, disregard the possible relation between continental thickening and subsequent asymmetric extension. We have investigated physical conditions which may lead to initiation of asymmetric extension. In this study we have considered three classes of causes for initiation of asymmetrical extension on lithosphere-scale faults: continental collision, magma intrusions, and causes resulting from lithosphere–mantle interaction.

We find that development of detachment faults is likely in the transition zone of continental lithosphere which has been thickened very rapidly compared to the thermal time constant of the lithosphere, i.e. on the order of a few Ma. A second requirement is that the in-plane compressive force causing continental thickening, switches to in-plane tension rapidly (again compared to the thermal time constant). The fault zone is predicted to dip under the thickened continental
lithosphere, consistent with observations in the Aegean (Lee and Lister, 1992) and in the Basin and Range (Wernicke, 1981). Our models predict a normal sense of shear along a dipping zone of localized deformation which evolves from the dipping lower crust. Any delay in either the mountain building or the sign switch of in-plane force will result in thermal relaxation, generating diffusely distributed buoyancy forces which tend to decrease the slope of the Moho. Since Moho depth gradients appear to be crucial for initiating dipping fault zones, thermal relaxation will thus, decrease the tendency of the lithosphere to asymmetric deformation. We show that our results are sensitive to the pre-stress distribution in our modeling.

A vertical magma intrusion into the lithosphere alone cannot initiate large-scale asymmetric deformation. Two conjugate systems of localized strain evolve, but none of them develop preferentially. It is clear that the orientation of the intrusion influences the symmetry of extension. We did not actually model intrusions in more gently dipping orientations, since the results would be very similar to the findings of Braun and Beaumont (1989a) and Dunbar and Sawyer (1989), who study the style of extension of offset pre-existing weak zones in the crust and mantle. These authors show that if a cause for preferential development of one of the two conjugate branches is added (in their case an offset crustal weak zone), the system will evolve asymmetrically.

Mantle plumes are probably not relevant for initiating detachment faults.

Our model study indicates that mantle delamination resulting from continental thickening is a likely mechanism for initiating detachment faults, if delamination occurs at the end of the compression phase and if in-plane tensile forces are applied to the thickened system immediately after compression. Model experiments not reported here indicate that this result is sensitive to the way thickening pre-stresses are applied in the modeling. A particular feature of the delamination model is that an asthenospheric diapir evolves, which will generate melt as a result of its rapid ascent and decompression. Latin and White (1990) argue that melt generation is unlikely in simple shear deformation. It, however, delamination precedes simple shear extension, this is not necessarily so.

Another interesting result from our modeling is that dramatic strain weakening in the mantle decreases the tendency for localized asymmetrical deformation. Lithosphere-scale detachment faulting is promoted in a layered rheological system with a weak lower crust overlying a strong mantle. Weakening of the mantle mechanically decouples the deformation behavior of the lower crust from the mantle, which subsequently deforms by pure shear.

It is important to realize that some mechanisms, which are concluded not to generate lithosphere-scale faults, do predict crustal-scale detachments. Based upon experimental flow laws and realistic geotherms, the lower crust should be very weak and, therefore, represent an excellent level for decoupling the crust from the mantle. In general, crustal detachments will lead to mixed-mode extension. However, more can be learned about the controlling physical mechanisms by studying end-members of the extension models spectrum. Given the tendency to pure shear extension in intraplate deformation, we have, therefore, focussed on the physical mechanisms which are relevant in the evolution of whole-lithosphere asymmetric features. This paper intends to give insight in the overall conditions which determine the initiation of lithosphere-scale faults. We did not conduct a full parameter study but selected a few models of fairly common geological processes. Combinations of the above causes for asymmetrical deformation may occur, but are considered less likely. In a forthcoming paper (Govers, 1993) we investigate how parameters of the continental thickening mechanism affect the tendency to initiation of lithospheric detachment faults.

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