Effects of Initial Weakness on Rift Architecture

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Abstract

Much work has been conducted investigating the primary factors controlling rift architecture, using both computational and laboratory methods. Here, we examine the effects and relative importance that different types of initial weaknesses have on extensional lithospheric deformation style. We find that the type of initial weakness included in the model plays a primary role in determining the subsequent rift mode. Models using single localised weak seeds produce symmetric narrow rifts, irrespective of whether or not strain-softening is included, while models that include regions of diffuse weaknesses produce a wide rift mode. Models that include an initial weak fault tend to produce more asymmetric rifting leading to a core complex mode. By distributing the strain laterally over large areas, the ductile lower crust tends to oppose the narrow rift mode forced by the initial weakness. These results suggest that initial weaknesses may play a major role in determining the mode of rifting. Our numerical experiments confirm that low angle faults can form as the result of rotation of initially high angle faults. While previous studies have suggested that the rheological and thermal profiles of the lithosphere play the most important role in rift mode determination, our results illustrate that initial weaknesses could play a major role in rift mode determination, highlighting the need to make initial weaknesses a primary consideration when modelling the extensional deformation of the lithosphere.
Introduction

Rift modes and rift architectures have been the subject of many studies in the past using both laboratory (Brun and Beslier, 1996; Brun, et al., 1994; Michon and Merle, 2003) and computational methods (Buck, 1991; Huismans and Beaumont, 2003; Lavier, et al., 1999; Nagel and Buck, 2004; Wijns, et al., submitted 2004). While most of these studies have focused on the importance of continental rheology and thermal regime, little consideration has been given to the role initial weakness may have in inducing stress and strain localisation (Dunbar and Sawyer, 1989). We present a quantitative analysis of the effects initial weaknesses have on the mode of continental extension and the geometry of rift architecture. Having tested various forms of localized and randomly distributed weakness, we find that the type of initial weakness embedded in the model, represents a primary controlling factor of continental rift architecture.

Continental Extension Modes and Rift Architectures

Regardless of the origin of the forces driving extension, continental rifts have been classified into three categories (Buck, 1991; Corti, et al., 2003; Ruppel, 1995). In the narrow rift mode, exemplified by the East African System, extensional deformation localizes along typically 100 to 150 km wide, continuous or segmented, rifts. In the wide rift mode, such as the Basin and Range Province, extensional deformation is more diffuse spreading over an horizontal length-scale several times larger than the thickness of the lithosphere. In contrast to narrow rifts, wide rifts display relatively small gradient in crustal thickness. The Aegean domain exemplifies the core complex mode of continental extension. As with the wide rifts, core complexes display little
variation in crustal thicknesses, however, they differ from the wide-rift mode in that deeper crustal levels are exhumed to the surface forming a domical metamorphic core wrapped by a low-angle normal fault connected laterally to a crustal-scale detachment fault. The décollement/detachment system is overlaid by low-grade rocks and detrital sediments (Coney, 1980).

The consensus is not yet reached on the significance of the three different modes of continental extension. Some authors see in the core complex mode a mere variant of wide rift mode (e.g. Brun, 1999) whereas for others, core complex and wide rift modes represent the result of fixed-boundary collapse and free boundary collapse respectively (Rey, et al., 2001). For some, the three modes may represent various stages of long-lasting continental extension (Olsen and Morgan, 1995). Rheology is commonly accepted as a key parameter. Wide rift mode is thought to be promoted by power-law rheology that dominates at higher temperature and low strain rate, whereas plastic deformation that dominates under lower temperature and high strain rate, is thought to favor the development of narrow rifts (Bassi, 1991; 1995). Continental extension modes are also debated in terms of rheological stratification of the continental lithosphere. In particular the depth of the brittle-ductile interface in the crust, the coupling between brittle and ductile layers, as well as the number of mechanically contrasting layers in the lithosphere, are of primary importance in controlling the style of continental extension (Allemand and Brun, 1991; Allemand, et al., 1989; Benes and Davy, 1996; Brun and Beslier, 1996; Brun and Tron, 1993; Huismans, et al., 2005). Rifting modes are also discussed in terms of competition between gravitational forces and driving forces (i.e. integrated lithospheric strength) somehow related to the relative thicknesses of the brittle and ductile layers (Benes and
Davy, 1996; Brune and Ellis, 1997). There is a lot of overlap between the various interpretations of the three modes of continental extension and overall they all point toward the conclusion that, continental geotherm, rheology, and strain rate conspire to dictate the rift mode.

Many studies have also emphasized the role of localized initial weakness (as thermal anomalies, compositional heterogeneities or both) in localizing the development of narrow rifts (Bassi, 1991; 1995; Braun and Beaumont, 1987; Dunbar and Sawyer, 1989; Fletcher and Hallet, 1983), or promoting the development of core complexes (Brun, 1999; Brun, et al., 1994). We examine here a range rheological models in which localized and diffuse rheological heterogeneities are superimposed onto coupled and decoupled rheological profiles.

Model Description

We use the Ellipsis code, a Lagrangian integration point finite element code capable of tracking time dependant variables embedded in an Eulerian mesh, allowing for accurate tracking of surfaces and boundaries through time (Moresi, et al., 2002; Moresi, et al., 2001). While the parameters in the model are scaled to achieve maximum computational efficiency (see Appendix), “real-world” values parameters are listed in Table 1. Our continental lithosphere is 450 km long and 140km thick and includes a 35 km thick continental crust. Extension is driven by a constant velocity condition applied to the right boundary of the model (Fig 1), giving an initial strain rate of $1 \times 10^{-15} \text{ s}^{-1}$ over the horizontal length of the model. A free-slip boundary condition is applied at the top and bottom surfaces. The basal heat flux of 20 mWm$^{-2}$ and a crustal radiogenic heat production ($1.2 \times 10^{-6}$ Wm$^{-3}$) was adjusted to achieve a
Moho temperature of 550 °C while the 1300 °C isotherm is reached at 140km. We use the standard rheological profile of Brace and Kohlstedt (1980) for the continental lithosphere in which a Mohr-Coulomb failure law, augmented by a strain weakening function, is the dominant failure mechanism at low temperature and high strain rate, whereas the Frank-Kaminetskii approximation (Frank-Kamenetskii, 1955) of the temperature dependent Arrhenius (η) viscosity is used at high temperature and low strain rate:

\[ \eta(T) = \eta_0 e^{-\frac{T}{T_0}} \]  

Equation (1)

where \( T \) is the temperature, \( \eta_0 \) is the viscosity at a temperature of 0 °C and \( T_0 \) is a constant. Coupling of the crust to the mantle is implemented by altering the values of \( \eta_0 \) and \( T_0 \). Coupled models have values of \( \eta_0 \) and \( T_0 \) chosen such that the entire crust is initially in the brittle regime while decoupled models have values of \( \eta_0 \) and \( T_0 \) chosen such that the lower crust is initially in the ductile regime (Fig. 1).

Strain weakening in the Mohr-Coulomb failure law is implemented via a power law function \( f(\varepsilon) \)

\[ f(\varepsilon) = \frac{1 - (1 - a) \left( \frac{\varepsilon}{\varepsilon_0} \right)^{\varepsilon_n}}{a} \]

Equation (2)

where \( \varepsilon \) is the accumulated plastic strain, \( \varepsilon_0 \) is the saturation strain beyond which no further weakening takes place, \( \varepsilon_n \) is an exponent which controls the shape of the function and \( a \) is a maximum value of strain weakening beyond which no further
weakening occurs. In our models strain weakening is linear ($\varepsilon_n = 1$) with a maximum of 80% weakening (equivalent to a $\varepsilon_n$ of 0.2) taking place at a saturation strain ($\varepsilon_0$) of 0.5 (Figure 1), similar to the values used in previous work (Huismans and Beaumont, 2003; Wijns, et al., submitted 2004).

Initial weaknesses are implemented in the following methods:

- **Single weak seeds:** square regions of low viscosity material are included in the lithosphere possibly representing regions of higher radiogenic heat production (if anomaly is located in the crust), regions of locally reduced viscosity due to thermal/pressure perturbations or regions of weakness caused by a material inhomogeneity.

- **Regions of diffuse weakness:** random uniformly distributed regions of pre-strained, pre-weakened material are dispersed through the crust corresponding possibly to accretionary continental crust with crustal inhomogeneities or localised regions of higher radiogenic heat production decreasing the strength of the continental crust at discrete points over a large region.

- **Faults:** discrete regions of low viscosity material cutting the upper and lower crust at an angle of 45° are included representing a fault or a planar weakness such as a suture zone.

**Results**

We test the differing types of weakness on models where the crust is coupled to the mantle (brittle lower crust) representative of older continental lithosphere as well as on models where the crust is decoupled from the mantle (ductile lower crust), representative of younger continental lithosphere (Brun, 1999).
Single Localised Weak Seeds

In these models, a single square of low viscosity material represents a localised weak seed. Different models were tested by embedding the weak seed at various depths within the lithosphere. Upon extension, strain focuses around the weak seed resulting in the development of a narrow rift (Figs. 2-8). Previous studies (Buck, 1991) have demonstrated that narrow rift mode preferentially evolves in lithosphere with high integrated strength, common of cool and thick lithosphere. Our modelling indicates that a narrow rift mode can also develop as a result of the inclusion of a single weak seed in the lithosphere. We tested the sensitivity of the model to the magnitude of the weakness and found that a seed 10 times weaker than that of the surrounding material is sufficient to localise strain. In coupled models, conjugate normal faults originating from the weak seed lead to one single rift zone whose width corresponds to the depth of the seed (Fig. 2-5). This narrow rift accommodates ongoing extension. Allemand and Brun (1991) found that conjugate rift faults meet at the base of the brittle layer and therefore the thickness of the brittle layer controls the width of the rift basin. By placing the weak seed at increasing depth we are locally increasing the depth to the effective base of the brittle layer (the top of the weak seed) and subsequently increasing the width of the resulting rift basin.

Decoupling the crust from the mantle (Figs. 6-8) results in different rift architecture than the coupled models, although the narrow rift mode still accommodates extension. Regardless of where the seed is located, periodic but out of phase zones of necking develop in the upper crust and upper mantle. A weak seed placed in the upper crust results in a small rift basin, the width being controlled by the depth to the crustal brittle-ductile transition. Lower crustal flow towards the rift centre feeds back into
the upper mantle causing strain localisation and necking. The displacement from upper mantle necking is then transmitted by lower flow inducing further faulting of the upper crust at distances corresponding to the depth of the mantle brittle-ductile transition. When a seed is placed in the ductile lower crust or in the upper mantle, two small narrow basins form in association with two conjugate normal faults. At the surface, the distance between the conjugate faults is about the depth of the crustal brittle-ductile transition. The distance between the two smaller basins is equal to the depth of the brittle-ductile transition in the mantle. This may have implications for using basin structure to obtain information about the rheology of the rift, with the rift basin widths and distances between rift basins having a direct correlation to the depths of the crustal and mantle brittle-ductile transitions or to crustal heterogeneities.

Huismans and Beaumont (2003) suggested that rift asymmetries can be caused by preferential localisation of extension over a single weak seed being taken up along one conjugate fault of a rift as a result of the inclusion of strain weakening. The model setup used here is very similar to that of Huismans and Beaumont (2003), however, our single seed models produce symmetric rift basins both with and without the inclusion of strain softening (Figs. 4 and 5). This may be that our models retain computational symmetry while those of Huismans and Beaumont (2003) do not. This serves to illustrate the need for further benchmarking of available dynamic modelling codes. Rift basins formed including strain weakening produce more localized rift basins with more pronounced topography due to increased focusing of the strain along the conjugate faults of the rift, but the general rift architecture is the same between the two models. This was observed in all models, indicating that while strain weakening plays a role in modulating the topography of the rift basin, it is not of first order
importance in rift mode selection. The topography generated in the models, even models without strain weakening included, does seem to be exaggerated. In some cases rift flank topography exceeds 10km. Inclusion of erosional and sedimentary processes (Burov and Poliakov, 2001) in the model in the future may serve to help account for the over-estimated topography.

Region of Diffuse Weakness

In these models we implemented the initial weaknesses as small, random (uniformly distributed) regions of weaker material confined to the central 350km of the models. Upon extension, uniformly distributed weak seeds lead to a wider zone of extension (~500 km after 50% extension) made of numerous, discrete and synchronous basins (Figs. 9 and 10). In both coupled and decoupled models, each seeds initially produce a set of conjugate normal faults. Upon ongoing extension, only a smaller number of normal faults remain active. The natural selection process is such that in the final configuration the basins are bounded by normal faults merging at the brittle-ductile transition. The final configuration in the decoupled model is very similar to that of our decoupled/single seed models. However, the strain distribution is a significant difference. In the diffuse weakness model, the basins are synchronous and therefore they accommodate similar amount of extension. Overall, the lateral distribution of the stretching factor is rather homogeneous. This clearly contrasts with the wide extensional zone that results from a single weak seed in decoupled models. In the latter, progressively younger basins develop for the initial weak seed. Consequently, older basins accommodate a larger amount of extension and the stretching factor is heterogeneous over the length of the model. This model may represent the diffuse rift mode of Ruppel (1995), where topographically high-standing features representing the footwalls of normal faults are interspersed with intervening valleys.
The distance between rift bounding faults and between adjacent rifts in the coupled model approximately corresponds to the distance to the depth of the brittle-ductile transition. In decoupled models the distance between rift bounding faults corresponds roughly to the depth of the crustal brittle-ductile transition while the distance between adjacent basins corresponds roughly to the depth of the brittle-ductile transition in the mantle.

Although the parameters of our models place them within the narrow rift mode field of Buck (1991), the presence of diffuse weakness produces a wide rift mode via a series of adjacent narrow rifts, while the geometry of the rift fits the broad description of a wide or diffuse rift (Ruppel, 1995). It could be argued that the temporal evolution of our wide rift model does not fit the description of a wide rift (Buck, 1991), since strain is initially distributed in many regions with a wide rift mode active from the onset of rifting, instead of strain migrating through. Clearly the definition of the wide rift mode needs to be revisited.

**Faults**

Faults were implemented in the model as a thin strip of weak material inclined at 45° representing a planar feature such as a fault or a suture zone. The coupled fault model (Fig. 11) produces a strongly asymmetric rift and clearly demonstrates that an initially high angle fault can rotate to become low angle detachment. This result is consistent with the rolling hinge model of Lavier et al. (1999). As the lithosphere extends, rotation of the west dipping high angle fault results in a low angle fault being exposed at the surface which has accommodated substantial movement. After about ~10% extension a secondary, synthetic high angle fault appears dipping east. Both faults
continue to rotate with the final rift geometry (50% extension) showing two low angle faults dipping in opposite directions, generated from one initially westward dipping high angle fault.

While single seed weaknesses resulted in narrow rift mode, imposing a fault as initial weakness favour a core complex mode. In the coupled model (Fig. 11) a metamorphic core complex initiates with lower crust exposed at the surface over the eastern rift arm after around 35% extension. Following further extension, an oceanic core complex forms with the upper mantle lithosphere being exposed at the surface after ~50% extension. Decoupling the crust and mantle (Fig. 12) results in lateral diffusion of extension. In a decoupled system the high angle fault almost acts as a single weakness with a final rift architecture similar to that seen in the decoupled single upper crustal weak seed model (Fig. 6). Previous studies have required weak rheological profiles, weak lower crust being a particular condition (Brun, et al., 1994; Buck, 1991; Wijns, et al., submitted 2004) as a requirement for the formation of core complexes. Here we have shown the core complex mode can occur in coupled crust as a result of the inclusion of an initial high angle fault.

**Conclusion**

While most previous work has concentrated on regional strength discontinuities (such as rheological layering etc), here we demonstrate that localised strength anomalies can result in localisation of rifting and can also force selection of one of the three modes of rifting (narrow, wide and core complex). Our results suggest that the magnitude of these weaknesses does not have to be large, localised strength reductions to 1/10 that of surrounding materials are enough to localise strain and onset the rifting process. An initial weakness implemented as a single seed will result in a
narrow rift mode, a region of diffuse, random noise will result in a wide rift mode while an initial weakness implemented as a fault with result in a core complex mode. Decoupling the upper crust and mantle by including a ductile lower crust acts to suppress the rifting mode favoured by the initial weakness, acting to diffuse the rifting over larger lateral areas. The inclusion/omission of strain weakening does not affect the symmetrical evolution of our single seed models, as has previously been suggested (Huismans and Beaumont, 2003); it only serves to enhance the topography and create deeper rift basins due to strain being preferentially focussed more along the bounding faults of the rift.

In single seed coupled models, the spacing of rift bounding faults is a function of the effective depth of the weak seed. In addition, in single seed decoupled models where multiple rifts develop, the spacing of adjacent rifts is controlled by the depth of the brittle-ductile transition in the mantle. In a coupled configuration of the diffuse weakness models, the distance between rift bounding faults and between adjacent rifts roughly corresponds to the depth of the brittle-ductile transition. In the decoupled configuration the distance between rift bounding faults correspond to the depth of the crustal brittle-ductile transition while the distance between adjacent basins corresponds to the depth of the brittle-ductile in the mantle. The decoupled models show out of phase stretching in the upper crust and upper mantle. We suggest that the core complex mode of rifting can be achieved with the inclusion of a weak fault. This conclusion is an alternative to previous studies which found the rheological profile dictates the formation of core complexes (Buck, 1991; Wijns, et al., submitted 2004). It should be noted that both our and previous studies into factors controlling core
complex formation are not mutually exclusive. The results presented here highlight that rift mode selection can be forced by more than one dominant model parameter.

References


Michon, L., and Merle, O. (2003), Mode of lithospheric extension: Conceptual models from analogue modeling, Tectonics, 22, -.


Figure Captions

Figure 1
(top) Model geometry showing material layer thicknesses. Upper and lower crust are a wet quartz composition with density 2700 kg/m$^3$ and 2800 kg/m$^3$ respectively while Mantle is a dry olivine composition with density 3300 kg/m$^3$. Extension is driven by a boundary velocity $V_{ext}$ chosen to achieve a strain rate of $1 \times 10^{-15}$ over the model. The initial temperature field, controlled by a basal heat flux of 20mW/m$^2$ and radiogenic heat production of $1.2 \times 10^{-6}$ W m$^{-3}$ in the crust, is initially laterally uniform and increases with depth from 0°C at the surface, 550°C at the base of the crust and to 1300°C at the base of the model. Erosion and sedimentation are not included in the model. (bottom left) Representative yield strength envelopes of a coupled and decoupled model for a strain rate of 1e-15 with initial strength denoted by solid lines and strain weakened denoted by dashed lines. (bottom right) Strain softening behaviour showing strength weakening from 100% to 20% after an accumulated strain of 0.5, after which no further weakening occurs. Dashed lines show the effect of the exponential parameter (En) on the curve (see Equation 2 in the text).

Figure 2
Coupled upper crustal localised single seed weakness model including strain weakening after extension of 10%, 20%, 35% and 50%. Coupling is controlled by variation of the T0 parameter in the viscosity approximation (see text). Upper crust is orange, lower crust is brown, mantle is red while the imposed weak seed is purple. A shows strain (blue colouring) while B shows only material colouring at the same time deformation step.
Figure 3
Coupled lower crustal localised single seed weakness model including strain weakening after extension of 10%, 20%, 35% and 50%.

Figure 4
Coupled model including strain weakening with single seed weakness imposed in the upper mantle. Notice rifting is symmetric in this model.

Figure 5
Coupled model with single seed weakness imposed in the mantle and no strain weakening. Rifting is still symmetric in this model, however, a shallower rift basin forms compared to models which include strain softening (Fig. 4).

Figure 6
Decoupled model results after extension of 10%, 20%, 35% and 50% including strain weakening with a localised single seed weakness in the upper crust. Decoupling is controlled by variation of the T0 parameter in the viscosity approximation (see text). Imposing a single seed weakness in the upper crust results in a W type architecture of the narrow rift mode (see text for discussion).

Figure 7
Decoupled model results after extension of 10%, 20%, 35% and 50% including strain weakening with a localised single seed weakness in the lower crust.

Figure 8
Decoupled model results after extension of 10%, 20%, 35% and 50% including strain weakening with a localised single seed weakness in the upper mantle.

Figure 9
Coupled random weakness model results. Weak seeds (blue triangles) are uniformly randomly distributed through the crust between 50km and 400km. Strain (blue colouring in A) is localised in many basins resulting a wide rift mode.

Figure 10
Decoupled model with weak seeds (blue triangles) uniformly randomly distributed through the crust between 50km and 400km.

Figure 11
Coupled model including strain weakening with a 45 fault imposed in the crust. Rifting is very asymmetric and a core complex forms after ~70% extension with mantle exposure at the surface. A characteristic feature of the Iberian Margin is also demonstrated here with the upper continental crust directly overlaying the mantle after ~35% extension (see text for further discussion).

Figure 12
Decoupled model including strain weakening with a 45 fault imposed in the crust.
<table>
<thead>
<tr>
<th>Parameter</th>
<th>Natural Value</th>
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<tr>
<td>density upper crust</td>
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</tr>
<tr>
<td>density lower crust</td>
<td>2800 kg m$^{-3}$</td>
</tr>
<tr>
<td>density mantle</td>
<td>3300 kg m$^{-3}$</td>
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<td>basal heat flux</td>
<td>$20 \cdot 10^3$ W m$^{-2}$</td>
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</table>

Table 1

List of natural (non-scaled) model parameter values.
Appendix – Scaling

The ellipsis model input parameters are scaled from real world values into dimensionless values in order to minimise computation time and increase the accuracy of the solution. We use the non-dimensional scaling approach given by:

\[ E = N/S \]  

Equation (1)

where \( E \) is a dimensionless Ellipsis variable, \( N \) is the dimensional real world parameter and \( S \) is a dimensional scaling factor.

Scaling from real model length dimensions of 450km wide by 150km deep is done using a length scaling factor \( S_L \) of \( 1.5 \times 10^5 \) m resulting in non-dimensional model geometry of 3 units wide by 1 unit deep.

Using a Thermal diffusivity scaling factor \( S_\kappa \) of \( 1 \times 10^{-6} \) a time scaling factor \( S_t \) can be found using:

\[ S_t = S_L^2 / S_\kappa \]  

Equation (2)

With a viscosity scaling factor \( S_\eta \) of \( 1 \times 10^{21} \) Pa and a gravity scaling factor \( S_g \) of 1 m/s\(^2\), a density scaling factor \( S_p \) can be found:

\[ S_p = \frac{S_g \times S_L \times S_t}{S_\eta} \]  

Equation (3)
Using a velocity scale \((S_u)\) defined by:

\[ S_u = \frac{S_L}{S_t} \quad \text{Equation (4)} \]

a non-dimensional velocity of 67.5 is calculated to achieve a real world strain rate of \(1 \times 10^{-15} \text{s}^{-1}\) over the length of the model.

Temperature is scaled between non-dimensional values of 0.17 and 1 corresponding to temperatures of 273K and 1603K respectively using a temperature scale \((S_T)\) of 1603K.
Figure 1
Figure 4

A

B

10% extension

20% extension

35% extension

50% extension

100km
Figure 5

A

B

10% extension

20% extension

35% extension

50% extension

100km
Figure 6
Figure 7
Figure 8

A

B

10% extension

20% extension

35% extension

50% extension

100km
Figure 9
Figure 10

A

B

10% extension

20% extension

35% extension

50% extension
Figure 11

50% extension
35% extension
20% extension
10% extension

100km
Figure 12