Symmetric alternative to asymmetric rifting models

Thorsten J. Nagel
W. Roger Buck
Lamont-Doherty Earth Observatory of Columbia University, Palisades, New York 10640, USA

ABSTRACT

We describe the first numerical simulation of continental rifting that reproduces the three major structures observed at magma-poor margins such as the Galicia Margin west of the Iberian Peninsula or the Apulia Margin in the Alps: (1) Distal continental margins consist of fault-bounded blocks separated by oceanward-dipping normal faults. (2) At the tip of the continent, lower crust is scarcely preserved or absent, and upper crust directly overlies exhumed mantle. (3) The base of the rotated crustal blocks is a prominent seismic reflector and represents a high strain zone with a top-to-the-ocean sense of shear. In our model, these structures do not reflect asymmetric rift geometry at a lithospheric scale. Instead, they derive from upper-crustal collapse over a mid-crustal shear zone into the rift center and are present on both sides of the rift axis. The model has a horizontal weak zone in the middle crust on top of strong lower crust and a localized vertical zone of thermal weakness in the rift center. We hypothesize that the development of a thermal perturbation and associated strain localization in the deeper lithosphere may cause the transition from widely distributed faulting and crustal thinning to constricted faulting directed toward a well-defined rift center.

Keywords: rifting, continental margin, detachment, numerical model, lower crust.

INTRODUCTION

One of the most discussed questions about continental margins is whether the geometry of rifting was asymmetric or symmetric at a lithospheric scale. Symmetric models show similar structural evolution on both margins with the rift axis as the center of symmetry (McKenzie, 1978; Brun and Beslier, 1996). Asymmetric models assume significant deformation along solitary normal faults (detachments) that cut through crust and lithospheric mantle (Wernicke, 1985). These faults would accommodate tens of kilometers of offset and eventually lead to the exposure of mantle material at the surface. Two conjugate margins would represent the upper and lower plate of a detachment fault and would show fundamentally different geometries. Apart from the economic interest in the architecture of continental margins, the existence of lithospheric detachments would have far-reaching implications for the mechanical properties of the lithosphere.

Figure 1 displays the geometry of two classical magma-poor margins, which have been the subject of comparative studies (Whitmarsh et al., 2001; Wilson et al., 2001; Manatschal and Bernoulli, 1999)—the Early Cretaceous Galicia Margin west of Spain and the Jurassic Apulia Margin now incorporated in the Alps. The margins west of the Iberian Peninsula (the Galicia Margin and the adjoining Iberian Abyssal Plain Margin) have been intensely studied by deep-sea drilling cruises (Ocean Drilling Program Legs 103, 149, and 173), submersible cruises, and seismic experiments (Boillot et al., 1980, 1987, 1995; Mauffret and Mondatert, 1987; Hoffmann and Reston, 1992; Pickup et al., 1996; Reston et al., 1996; Whitmarsh et al., 2000). The margin of the Apulian microcontinent that was situated toward the Alpine Tethys (Piemont Ligurian Ocean) is exposed in eastern Switzerland and has been used for onshore investigation of rifting processes (Froitzheim and Manatschal, 1996; Manatschal and Bernoulli, 1999). At both locations, the oceanic lithosphere close to the continental margin shows serpentinized ultramafic rocks of subcontinental composition covered with deep-sea sedimentary rocks. Basalts are only sporadically present but become more abundant oceanward and finally merge into oceanic crust a few tens of kilometers farther outboard (Boillot et al., 1987; Manatschal and Bernoulli, 1999).

Continental crust at the distal margin is made up of rocks with typical upper-crustal composition. Lower-crustal rocks are absent or only locally preserved (Manatschal and Bernoulli, 1999; Whitmarsh et al., 2000). The continental crust tapers from 20 or 30 km thickness to zero over ~50 km. This wedge consists of an array of tilted blocks separated by consistently oceanward-dipping normal faults. At the Galicia Margin, some of these faults close to the edge of the continent penetrate the crust-mantle boundary. However, most faults seem to root in a subhorizontal zone of high seismic reflectivity, which was first described from the Bay of Biscay (De Charpal et al., 1978; Pickup et al., 1996; Reston et al., 1996). This so-called S-reflector is present at the base of the crustal wedge for tens of kilometers and reaches the ocean floor at the tip of the continental crust. There, it is represented

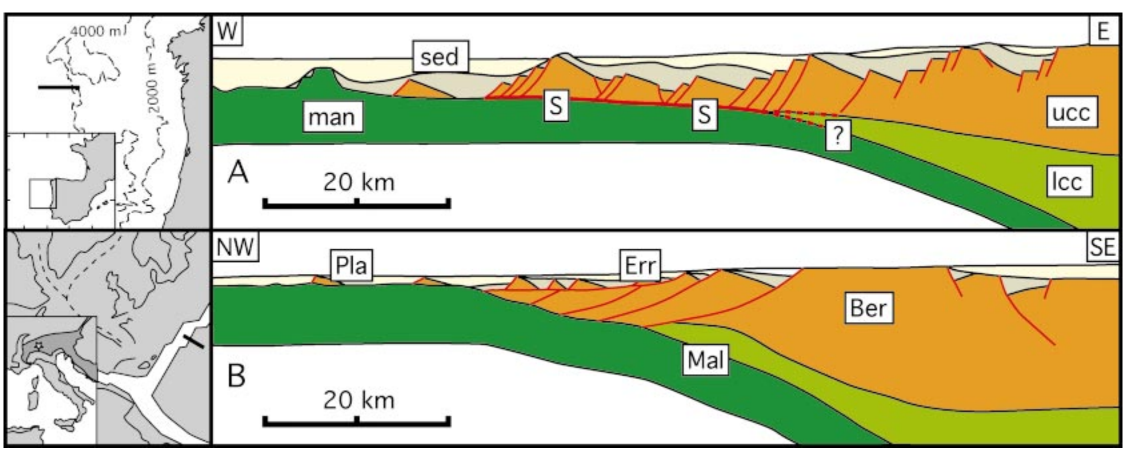

Figure 1. Schematic cross sections through (A) Galicia Margin west of Iberian Peninsula (Boillot et al., 1995) and (B) Jurassic margin of Apulian continent toward Alpine Tethys (Manatschal and Bernoulli, 1999). Inset in B displays present location of margin in Swiss Alps and paleoposition of section in Jurassic time. Abbreviations in A: lcc—lower continental crust; man—mantle; S—S-reflector; sed—sedimentary rocks; ucc—upper continental crust. Tectonic units in B that incorporate corresponding positions (color coded) at rifted margin: Ber—Bernina nappe system (large horst developed during early stage of rifting); Err—Err nappe (distal continental margin); Mal—Malenco unit (lower crust and adjacent upper mantle); Pla—Platta nappe (oceanic lithosphere).

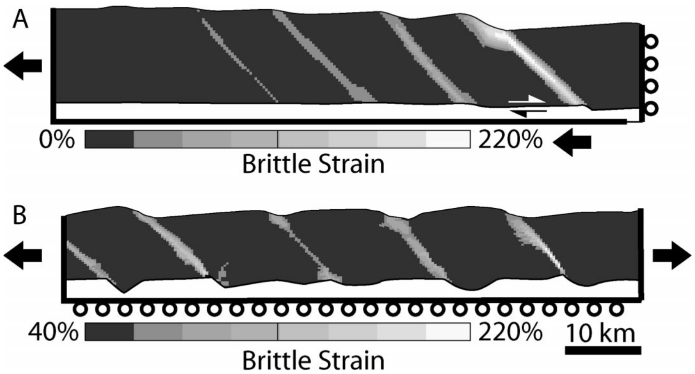

Figure 2. Three-layer extension experiments using vertically fixed lower boundaries. We used numerical code that has been applied in many studies of extension (e.g., Buck and Poliakov, 1998). Models are 75 km wide and 14 km thick (225 × 42 grid elements). Upper layer (11 km) follows Mohr-Coulomb yield criterion; lower layer (3 km) is linear-viscous with viscosity of 1 × 10^{20} Pa·s. Strain localization in brittle shear zones is induced by parameterization of strain weakening. Between 1% and 100% brittle strain, cohesion and friction angle are reduced from 15 MPa to 5 MPa and from 30° to 25°, respectively. Figures show brittle-strain pattern resulting from application of fixed extensional velocity at sides of model domain. A: Lower and left boundaries move to left, and right boundary is horizontally fixed. This setup leads to right-lateral shear flow in lower layer. Extension velocity is 3 × 10^{-10} m/s, and total extension is ~8 km. B: Lower boundary is vertically fixed but shear-stress free. Extension velocity is 5 × 10^{-10} m/s, and total extension is ~15 km. Strain below 40% has been cut away to highlight parallel orientation of significant faults. At displayed stage, only visible faults are active.

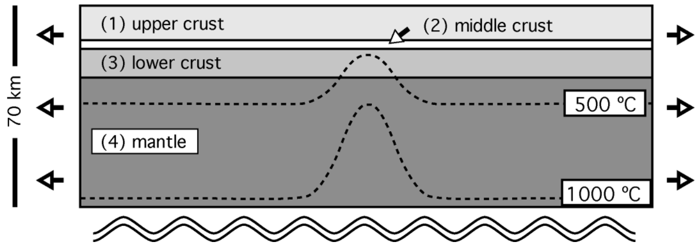

Figure 3. Setup for rifting model. Model is 200 km wide and consists of four layers (top to bottom): (1) brittle upper crust (12 km), (2) ductile middle crust (3 km), (3) lower crust (10 km), and (4) upper mantle (45 km). Viscous behavior of mantle and lower crust is determined by temperature-dependent creep parameters for olivine and feldspar, respectively (Ranalli, 1995). Layers 1, 3, and 4 follow Mohr-Coulomb yield criterion. Between 0.5% and 100% brittle strain, cohesion and friction angle are reduced from 20 MPa to 5 MPa and from 30° to 15°, respectively. Ductile middle crust is linear-viscous and has viscosity of 1 × 10^{20} Pa·s. Background thermal gradient is 15 °C/km. In addition, narrow bell-shaped thermal perturbation with maximum thermal gradient of 30 °C/km is present at center of model, which produces pronounced weakness in lower layers. Left and right boundary of model move with horizontal velocity of 3 × 10^{-10} m/s outward. Lower boundary is supported by lithostatic normal stress and deformable. Mantle material with temperature of lower boundary (1050 °C) is added where surface is pulled upward. Upper boundary is unconstrained.

by mylonites formed under mid-crustal, i.e., greenschist facies, conditions with a cataclastic overprint (Boillot et al., 1995). It is now accepted that the S-reflector represents a shear zone with top-to-the-ocean sense of motion (Manatschal et al., 2001). The continentward termination of the S-reflector is ambiguous. At both margins, the crustal wedge translates into a broad zone of moderately thinned crust in which normal faults do not consistently dip oceanward.

In the past decade, these observations were widely considered to support asymmetric rifting models. The oceanward-dipping normal faults at the distal margin have been interpreted as steep subsidiary faults in the hanging wall of a detachment, and the S-reflector at the Galicia Margin is construed as the detachment (Manatschal and Bernoulli, 1999; Hoffmann and Reston, 1992; Reston et al., 1996; Froitzheim and Manatschal, 1996). Accordingly, the Galicia and Apulia Margins would be lower-plate margins. Such a rifting geometry cannot account for the absence of lower crust at the edge of these margins, as abundant lower crust should be expected at lower-plate margins. The lack of lower crust has therefore been attributed to a predetachment stage of rifting, during which relatively strong lower crust would have been necked (Manatschal and Bernoulli, 1999; Whitmarsh et al., 2001).

Other studies have explained the polarity of normal faults at continental margins in an overall symmetric deformation pattern (McKenzie, 1978). Brun and Beslier (1996) proposed the formation of oceanward-dipping normal faults on top of weak lower crust that is sheared with a top-to-the-ocean sense of motion on both sides of the rift. In their interpretation, the S-reflector would represent this highly strained, ductile lower crust.

PARALLEL-DIPPING NORMAL FAULTS

We were led to consider the evolution of nonvolcanic rifted margins as an outgrowth of work on the general problem of parallel normal faulting. Parallel normal faults, like the oceanward-dipping arrays of the Galicia and Apulia Margins, are a common feature of many regions of continental extension, including the Gulf of Suez and the Basin and Range. During a series of simple numerical experiments we were able to simulate parallel faults only when we used a three-layer model with a strong brittle layer separated from a rigid lower boundary by a relatively weak ductile zone (Fig. 2). Figure 2A shows an experiment with boundary conditions typical for physical experiments. The lower boundary and left side of the model form a rigid frame that moves with a constant velocity to the left, whereas the right boundary is fixed in the horizontal direction. Offset accommodated along normal faults in the brittle material is transmitted to the right side of the model through the viscous lower layer. Hence, this setup forces a right-lateral shear flow in the viscous layer. Depending on precise model parameters such as thickness and viscosity of the weak layer, an array of right-dipping normal faults may form in the brittle layer. Similar results have been obtained in physical experiments using sand as a brittle material (Brun et al., 1994). The explanation of these experiments using sand was that, apart from extensional stress, a horizontal shear stress would be imposed on the brittle layer by the shear flow in the viscous layer. This shear stress would deflect stress orientations and make one direction of normal faults kinematically favorable (Mandl, 1987). It was concluded that horizontal shear stress is the crucial factor in the formation of parallel normal faults (Mandl, 1987; McClay and Ellis, 1987).

Shear stress associated with viscous flow and topography clearly determines the polarity of the fault array in Figure 2A toward the center of deformation on the right side. However, we find that the critical condition for the formation of parallel faults is that the lower boundary of the viscous layer resists vertical motions. By using a vertically fixed lower boundary, parallel-dipping faults develop in our experiments even if the lower boundary is shear-stress free (Fig. 2B). Faults in experiment B are initiated with both left and right dips. However, with increasing strain, faults with one or the other orientation are abandoned. Experiments using floating lower boundaries show a strong tendency to localize deformation at one place as the weak substratum can be easily pulled up below faults in the brittle layer.

On the basis of these results we reasoned that a model with a weak layer in the middle crust on top of strong lower crust might lead to the kind of upper-crustal blocks overlying mantle observed at non-

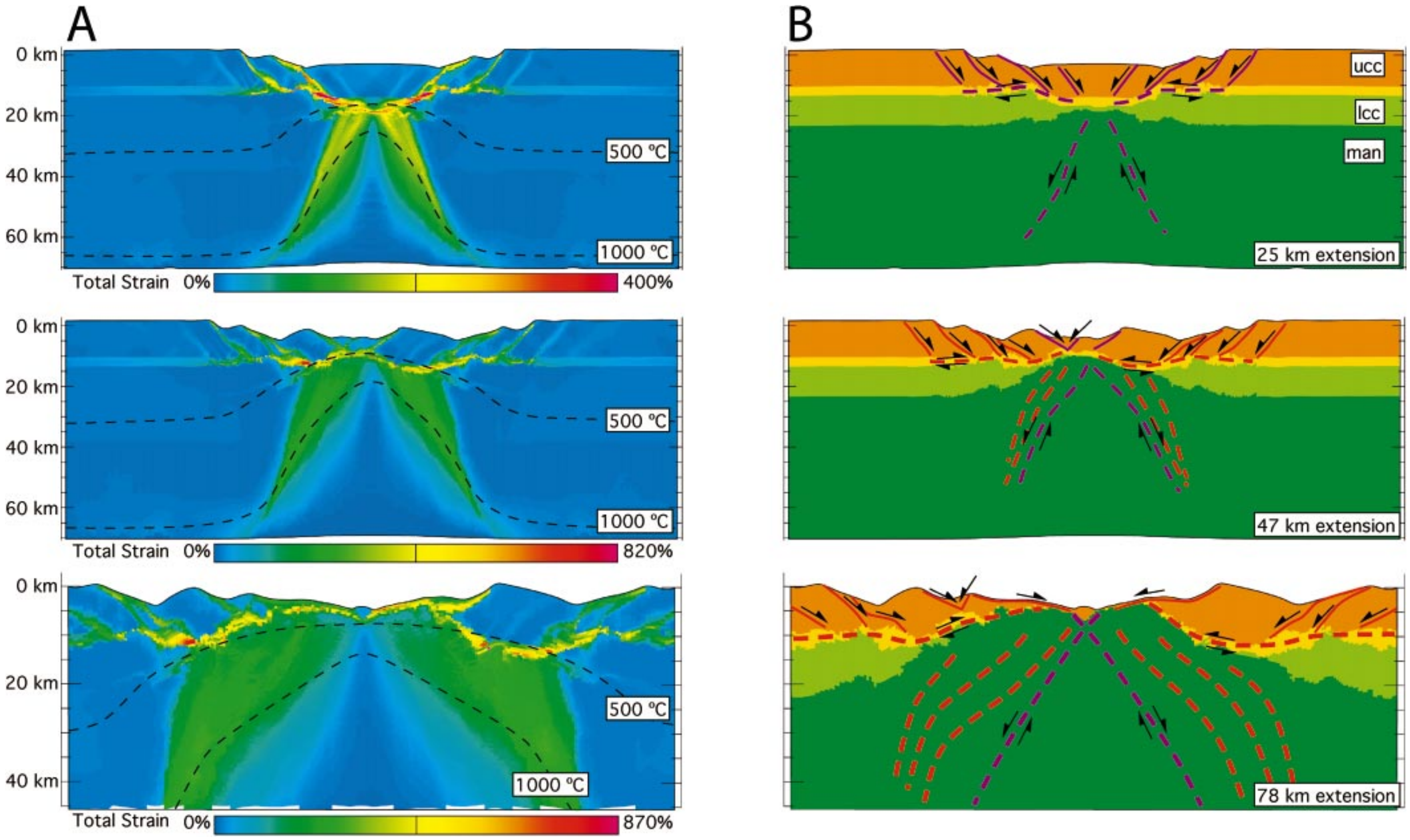

Figure 4. Rift model after ~25, 47, and 78 km of extension. A: Total strain (viscous and brittle). Dashed lines indicate 500 °C and 1000 °C isotherms, respectively. B: Distribution of upper crust, middle crust (unlabeled), lower crust, and mantle at same deformation steps as in A. Abbreviations: lcc—lower continental crust; man—mantle; ucc—upper continental crust. Purple lines—active high strain zones; red lines—abandoned ones. Solid lines—brittle deformation; dashed lines—ductile deformation. Thick dashed line highlights high-strain zone in weak mid-crustal layer (S-reflector).

volcanic margins. The middle crust, i.e., the ductile part of the upper crust, is often considered to be weaker than the lower crust owing to compositional differences or water depletion of the lower crust (Kohlstedt et al., 1995). Figure 3 shows the setup of a numerical model with a three-layer crust. Besides the vertical rheologic layering, a lateral perturbation was introduced through a thermal anomaly in the center of the model.

NUMERICAL MODELING OF RIFTING

During model extension, the uppermost mantle and the lower crust undergo localized necking with viscous flow in the hot weak site in the center of the model (Fig. 3). In the mantle below, two conjugate viscous high-strain zones with steep orientations develop, which intersect in the necking area. Material passes through these zones with flow directions upward in the lower region and outward on the left and right sides. Deformation in the brittle upper crust starts as a single graben above the necking area in the lower crust and mantle. Subsequently, both sides of the graben evolve into an array of parallel normal faults. These faults root in the weak mid-crustal layer, where the distributed deformation in the brittle upper crust is transferred into the necking area in the strong lower parts of the model. After ~25 km of extension, the lower crust tears; however, small remnants are locally preserved at the base of the upper crust. Uprising mantle fills the opening gap, and the collapsing upper crust comes directly on top of mantle rocks on both sides of the rift. After 40 km of extension the array of normal faults is abandoned and deformation in the upper crust is localized in the center of the model. After 75 km of extension the continental crust also breaks up and mantle is exposed at the surface. Ongoing extension is localized in the center of the new oceanic lithosphere, leaving behind a tectonically quiet passive continental margin. The top of the basement at the ocean-continent transition represents an unroofed normal fault, which is rotated into a gently dipping orientation.

DISCUSSION

The model reproduces the three major characteristics of magma-poor margins: (1) the oceanward-dipping normal fault array, (2) the upper-crustal blocks on top of subcontinental mantle, and (3) the mid-crustal shear zone with top-to-the-ocean sense of shear. The geometry of rifting is symmetric at a lithospheric scale. The model has a predefined vertical and horizontal stratification that is essential for the mechanical behavior—a relatively strong lower crust beneath weaker middle crust and a lateral thermal weakness in the rift center.

Some authors have inferred weak lower crust during continental rifting (Brun and Beslier, 1996; Manatschal and Bernoulli, 1999; Whitmarsh et al., 2000). However, we find it easier to explain rift geometries by assuming relatively strong lower crust. Weak lower crust would flow into the developing rift depression and would be abundant at the tip of the continent. Furthermore, experiments with a thick weak layer result in the formation of horsts and grabens rather than parallel-dipping faults, as the brittle upper crust is not affected by the resistance of the strong lower boundary. At the Apulia Margin, pre-Mesozoic gabbroic lower crust at the crust-mantle boundary was relatively cool (~550 °C) and therefore probably strong at the start of rifting (Münterer et al., 2000).

Our model has a predefined thermal weakness in the lithosphere. Such a perturbation may not be present at the onset of extension. However, owing to the temperature-dependent flow properties, extension in the mantle has a strong tendency to localize. A thermal weakness, e.g., as a result of small intrusions, can efficiently concentrate deformation. Such a weakness would further increase through advection. This is also the case in our model (Fig. 4A), where the long-term thermal structure is governed by the extension rate rather than the precise shape of the initial perturbation. Offshore Galicia, small amounts of gabbro dating shortly before continental breakup are found in the ultramafic basement (Schärer et al., 2000).

Normal faulting and moderate crustal thinning during early rifting affect regions hundreds of kilometers wide perpendicular to the continent-ocean transition at the Apulia Margin and the Galicia Margin. This geometry may reflect distributed stretching of thermally homogeneous upper mantle and lower crust. A transition to less distributed extension and to a geometry in which normal faults dip systematically toward a well-defined rift center may reflect strain localization in the lower lithosphere but not necessarily a switch to an asymmetric rifting geometry.

Rather than exploring conditions for symmetric and asymmetric rifting geometry (Huismans and Beaumont, 2002), this study shows that many observations used as evidence for asymmetric rifting are consistent with symmetric models. Similar to Brun and Beslier (1996), we interpret the structures observed at distal continental margins as the result of upper-crustal collapse into the rift center and not as an expression of asymmetric deformation at a lithospheric scale. In our model, the S-reflector represents a mid-crustal ductile shear zone that transferred distributed collapse in the brittle upper crust into a spreading center in the deeper lithosphere. Major deformation along the S-reflector was ductile and predated ultimate continental breakup. In contrast to existing models, we propose that the removal of lower crust from the site of future breakup is associated with slip along the S-reflector and the formation of the oceanward-dipping fault array.

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