Mantle detachment faults and the breakup of cold continental lithosphere

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ABSTRACT

We use a novel numerical approach that fully couples the energy, momentum, and continuum equations to investigate the physics of extension and breakup of cold continental lithosphere to form new ocean basins. Unlike hot continental systems, where flat-lying detachment faults are nucleated in the strong part of the upper crust, cold continental systems have flat-lying detachment faults nucleating in the strong upper mantle at a relatively early stage. These detachment faults subsequently control the development of a mantle core complex and associated crustal structures. The observed structures are analogous to those developed in mid-crustal core complexes during extension of relatively thick and hot continental crust. In the cold environment, however, a strong elastic layer is developed within the mantle, shifting the stress-bearing part of the system to below the Moho. Our modeling results reproduce key tectonic elements of a natural system (the Iberia margin, offshore the Iberian Peninsula) by stretching a randomly perturbed, unpatterned lithosphere. Results also explain the "upper plate paradox" by doming of continental lithospheric mantle separated from the crust by two diffuse detachment zones dipping toward the two future continental margins. Doming is facilitated by channel flow of the lower crust.

Keywords: extension, detachment faults, passive margin, numerical modeling, mantle exhumation, core complex.

INTRODUCTION

Detachment faults are low-angle normal faults that play a major role in continental extension (Lister and Davis, 1989), particularly in association with mid-crustal metamorphic core complexes (Fig. 1A). Detachment faults are also recognized in passive margin environments (e.g., Froitzheim and Manatschal, 1996; Reston, 1996; Driscoll and Karner, 1998; Manatschal and Bernoulli, 1999), where, following the model of Wernicke (1981, 1985), they have been thought to represent asymmetric translithospheric structures rooted in the mantle (Whitmarsh et al., 2001) (Fig. 1B). In the magma-poor Iberian continental margins offshore the Iberian Peninsula in southwestern Europe, serpentinized peridotites, interpreted to represent old subcontinental mantle, are exhumed as a mantle core complex in the footwall of an extensional detachment dipping toward the continent (Boillot et al., 1988).

Increased knowledge of the architecture of ocean-continent transitions in magma-poor rifted margins and the realization of the importance of detachment faulting and mantle exhumation (Manatschal, 2004) have led to a number of conceptual models for the evolution of continental breakup (Froitzheim and Manatschal, 1996; Whitmarsh et al., 2001; Davis and Kusznir, 2004). Numerical and analogue models have investigated the physics of these systems (Brun and Beslier, 1996; Huismans and Beaumont, 2002; Nagel and Buck, 2004; Lavier and Manatschal, 2006), but our understanding of the physics remains incomplete, particularly regarding the origin and development of detachment faults.

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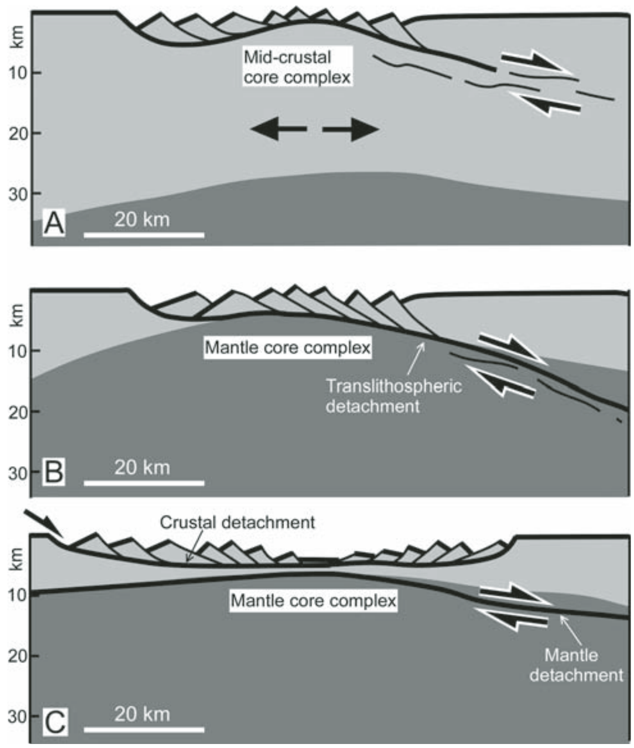

Figure 1. Different interpretations for the geometry of detachment faults. A: Detachment fault separating mid-crustal metamorphic core complex in footwall from tilted blocks in hanging wall. Detachment terminates at mid-crustal levels, governing partitioning between noncoaxial strain in upper-middle crust and pure shear in lower crust (after Lister and Davis, 1989). B: Translithospheric detachment fault (Wernicke, 1985) leading to exhumation of mantle core complexes (e.g., Reston, 1996; Whitmarsh et al., 2001). C: Mantle detachment faults, originating in the brittle-ductile transition in upper mantle, resulting in mantle core complex (this study). Final geometry resembles previous model, but mantle detachment is initially not connected to upper and middle crustal structures. In addition, mantle detachment originated within mantle and not along Moho.

In this paper, we reproduce key features of magma-poor continental rifted margins by using a novel formulation of the rheological behavior of the lithosphere, and without imposing major thermal or rheological heterogeneities or ad hoc weakening rules. This approach is fundamentally different from that of previous models, in which rifting was influenced by the imposition of weak layers or weakening rules that control strain localization and development of detachment faults (Nagel and Buck, 2004; Lavier and Manatschal, 2006). Our results indicate that symmetric rifting and mantle exhumation are controlled by development of intramantle and intracrustal detachment faults associated with variable strain rates with depth. Intramantle detachments represent neither the root of mid-crustal detachments, nor reactivation of a weak layer, but originate spontaneously within the strong upper mantle.

NUMERICAL APPROACH

The approach used here is based on energy feedback effects within and between different lithospheric layers, which are responsible for a dynamic evolution of the lithospheric strength profile (Regenauer-Lieb et al., 2006). This approach relies on brittle and ductile localization feedback processes, and their communication across the strong brittle-ductile transition. The physical processes and the numerical approach have been described previously (Benallal and Bigoni, 2004; Regenauer-Lieb and Yuen, 2004; Kaus and Podladchikov, 2006; Regenauer-Lieb et al., 2006). The brittle localization feedback is pressure sensitive, and the ductile localization feedback is characterized by creep instabilities and is temperature sensitive (Regenauer-Lieb et al., 2006). These two feedback loops operate simultaneously, and the dominance of one feedback mechanism over the other depends on pressure, temperature, and rheology.

In earlier contributions, we have shown that during the extension of relatively hot and thick (40–60 km) continental crust, an elastic layer develops within the quartz-dominated crust (Rosenbaum et al., 2005; Regenauer-Lieb et al., 2006). The term elastic layer, also known as elastic core, is used here to describe a layer of maximum elastic energy that may also undergo either viscous flow or brittle rupturing. This is the stress-bearing part of the system that controls the communication between ductile localization in the lower crust with brittle localization in the upper crust. In this system, two detachment zones develop: (1) an upper- to mid-crustal detachment immediately above the elastic layer, where steep upper crustal normal faults sole out; and (2) a detachment zone immediately below the Moho, detaching the mantle from the lower crust. The intracrustal detachment zone develops into a weak zone within what is otherwise the strongest part of the crust, and controls development of crustal core complexes. The Moho in our earlier models (Regenauer-Lieb et al., 2006) has remained essentially flat. In this new contribution, a cold lithosphere is extended, leading to the development of within-mantle detachment zones that control continental breakup.

The numerical model has been inspired by extensional structures in the Iberia-Newfoundland conjugate margins. The initial crustal thickness is 30 km and the surface heat flux is 50 mW/m^2. We assume a typical Variscan crust rheological structure (Strehlau and Meissner, 1987), with a 20-km-thick quartz layer, underlain by a 10-km-thick feldspar layer on an olivine-dominated mantle. In the models, the maximum deviatoric stress is not allowed to exceed the pressure (see details of the model setup in the GSA Data Repository[1]). We stop each numerical model after development of a subcontinental mantle core complex, but before final crustal rupture. This is because the rheology of the weaker part of the brittle crust (top 2 km) is not explicitly modeled. At shallow crustal levels, microcracks are increasingly important, but our numerical code does not calculate surface energies related to their formation. We therefore simplify shallow crust deformation by considering its behavior to be similar to soil, with vanishing cohesion. This simplification does not significantly influence the results because of the low strength of shallow crust.

RESULTS

A symmetric extension velocity of 1.2 cm/yr was applied (0.6 cm/yr in each direction) to a lithosphere cross section (200 km wide and 80 km deep) in plane strain. This corresponds to an effective pure shear strain rate of

$$\dot{\varepsilon}_{\mathrm{eff}}^{\mathrm{pure}} \equiv \sqrt{\frac{1}{2}\dot{\varepsilon}_{ij}\dot{\varepsilon}_{ij}} = 1.9\times10^{-15}\,\mathrm{s}^{-1} \quad (1)$$

for a box deforming homogeneously in plane strain. The lithosphere contains randomly distributed local perturbations with a wavelength of ~100 m. If the energy feedback mechanism is not considered, the model deforms by pure shear at constant strain rate. Energy feedback leads to significant departure from pure shear through brittle fracture, ductile shear localization, and dynamic creep instabilities. All of these lead to strain localization into zones of high strain rate variation with depth (Fig. 2).

In our models, three elastic layers develop, one in each rheological layer (the one in the feldspar layer is only weakly developed). These three elastic layers are defined by the highest initial regional stresses. Further extension and shear localization due to energy feedback effects turn these layers into zones of highest regional energy dissipation. Because energy dissipation is the paramount weakening process in flow localization, the strongest layer forms weak detachment faults (Regenauer-Lieb et al., 2006). In order to avoid ad hoc weakening effects, we maintain the rheological parameters constant throughout the models. Any weakening effects such as mantle serpentinization would further focus strain localization.

Temperature profiles in the models show nearly adiabatic decompression, with only a slight diffusion of advected geotherms. Unlike the hot model (Regenauer-Lieb et al., 2006), the olivine layer here is the strongest

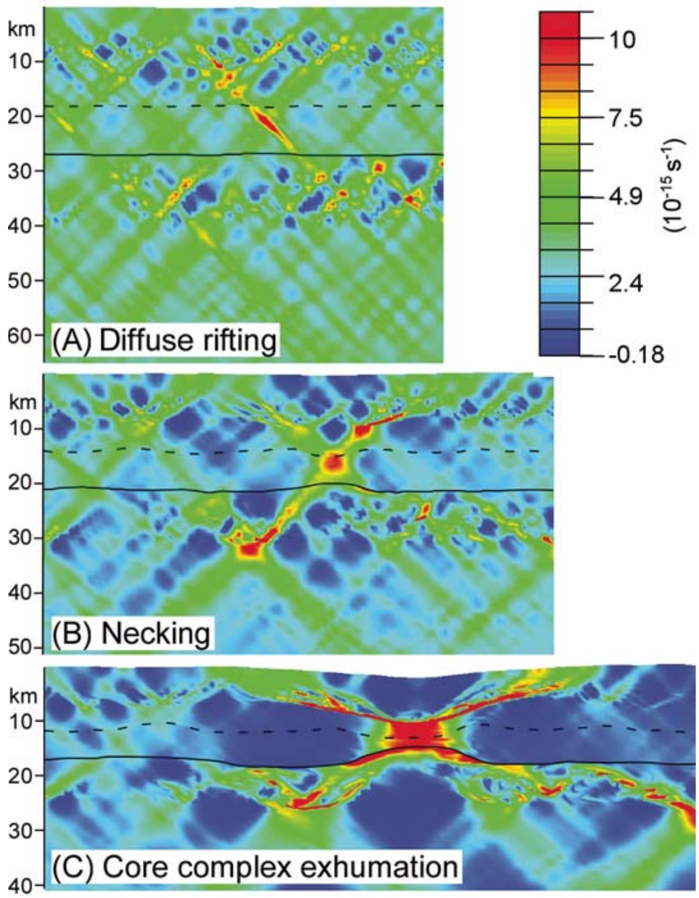

Figure 2. Modeled evolution of mantle core complex during the breakup of nonvolcanic continental margins. Crust-mantle boundary (Moho) is indicated by continuous black line. Dashed line separates quartz-dominated upper-middle crust and feldspar-dominated lower crust. Initial crustal thickness is 30 km and surface heat flow is 50 mW/m^2. Diagrams show strain rate for (A) diffuse rifting (β = 1.1), (B) crustal necking (β = 1.4), and (C) core complex exhumation (β = 1.8) stages. Low-angle shear zones are initiated during necking stage both in crust and upper mantle. No vertical exaggeration.

[1]GSA Data Repository item 2007253, a brief description of the model set up and the list of parameters used, is available online at www.geosociety.org/pubs/ft2007.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

of the three elastic layers, characterized by the highest values of energy dissipation. This elastic layer controls the development of a weak zone in the mantle, which becomes a detachment shear zone, characterized by high strain rates and stresses.

The evolution of rifting in our models is subdivided into diffuse rifting, crustal necking, and mantle core complex exhumation stages (Whitmarsh et al., 2001). The diffuse rifting stage (Fig. 2A) is characterized by distributed normal high-angle shear zones in the upper crust, above a mid-crustal elastic layer, and in the upper mantle, above a stronger mantle elastic layer. Strain rate and strain distributions are characterized by conjugate high-angle shear bands, which intersect immediately above the crustal and mantle elastic layers and continue into these layers at much lower values.

The crustal necking stage (Fig. 2B) marks the beginning of mantle core complex formation, when, in addition to the high-angle shear zones, there is a geometric weakening through the ensuing of Moho doming accommodated by newly formed low-angle detachment zones. This stage is achieved after 1.3 m.y. and $\beta = 1.4$, coinciding approximately with the amount of extension at which detachment faults were activated in the Iberia abyssal plain (Manatschal et al., 2001).

The core complex exhumation stage is the last evolutionary stage. It is characterized by the growth of a mantle core complex with an initial wavelength of 260 km (Figs. 2C and 3A) associated with second-order domes with wavelengths of ~32 km and amplitudes of 2 km. High strain and strain rates in the mantle are focused in a thick band below the Moho, characterized by a number of relatively narrow shear zones. This high strain zone is characterized by opposite dips and shear senses on either side of the dome, which together accommodate roughly symmetrical mantle doming.

During mantle doming, strain in the upper crust is focused into low-angle and flat detachment zones that dip toward the future oceanic crust (the mantle dome apex) and define major crustal necks or grabens. The location of these necks is controlled by the second-order underlying mantle domes. The horsts between are characterized by rotated blocks, 5–6 km wide, weakly deformed and separated by low-angle normal faults (Fig. 3). Block rotation is accommodated by pervasive flow in the feldspar-dominated lower crust and the lower part of the quartz-dominated layer.

Deformation of the originally vertical grid column (Fig. 3A) indicates a significant channel flow of the lower crust. This region controls the crustal necking stage (Figs. 2B, 2C) and undergoes considerable thinning, decoupling mantle from upper crust, and is bounded by the mid-crustal and the mantle detachments. The mantle detachment is characterized by significant simple shearing within a zone below the Moho that accommodates 40 km of lateral displacement away from the apex of the doming mantle (25 km to the left, 15 km to the right).

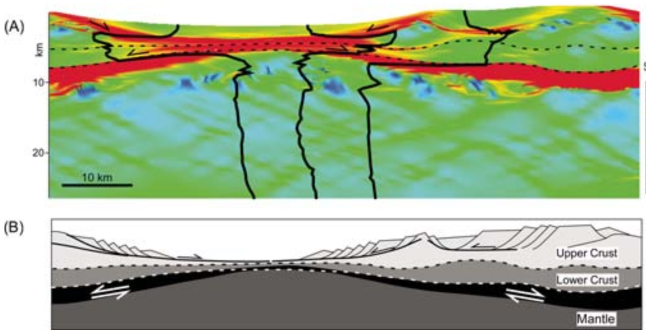

Figure 3. A: Strain distribution during the breakup of nonvolcanic continental margins after 15 m.y. (same parameters as Fig. 2). Dashed lines mark boundary between quartz, feldspar, and olivine layers. Three black curved steep lines indicate deformation of originally vertical lines, defining strong channel flow within lower crust. Normal faults at tip of continent dip toward basin (oceanward) and involve extreme thinning of feldspar lower crust (Nagel and Buck, 2004). Mantle dome has amplitude of 18 km and is 3.8 km from model surface. Mantle is separated from crust by wide detachment. This developed initially as narrow band within mantle, but as stretching increased, detachment widened to reach the base of the crust. Feldspar-dominated lower crust above dome is reduced to ~1 km thick. Maximum surface topography from peak to trough is ~2 km. B: Schematic cross section showing effect of crustal and mantle detachment faulting in nonvolcanic rifted margins based on interpretation of numerical results. Exhumation of subcontinental mantle dome is controlled by two detachment faults with opposite sense of movement that operate simultaneously (similar to model by Davis and Kusznir [2004, their Fig. 4.18]). Thick lines in crust and black band in upper mantle correspond to detachment faults.

DISCUSSION

The results indicate that decoupling between the doming subcontinental mantle and the stretched continental crust is facilitated by strain localization into diffuse zones of high strain rates, forming mantle detachments with opposite dips. Lithospheric breakup in the absence of magmas starts with high-angle normal faults (diffuse rifting stage) followed by the development of detachments nucleated within the elastic layers. Mantle detachment initiates in the crustal necking stage when zones of localized strain in each of the three elastic layers start to link ($\beta = 1.4$). Thereafter, the elastic layer in the olivine layer controls the exhumation of a subcontinental mantle core complex, facilitated by channel flow and necking of the feldspar layer. Channel flow in this case is driven by shear related to mantle upwelling.

The mantle core complex lacks the typical asymmetry expected for core complexes, characterized by the dominance of a single major detachment dipping in one main direction (Figs. 1A, 1B). In detail, however, local asymmetries arise regardless of the perfectly symmetric boundary conditions. The models considered only a randomly perturbed system without major primary anisotropies or asymmetric boundary conditions. The presence of such anisotropies in real geological examples might modify the results.

Our model supports the conceptual model for continental breakup in nonvolcanic passive margins (Boillot and Froitzheim, 2001), but differs because of the spontaneous formation of detachment faults within the mantle, which ultimately control the denudation of subcontinental mantle and continental breakup. Unlike previous models (e.g., Nagel and Buck, 2004), a mid-crustal detachment horizon develops self-consistently out of our dynamic rheology formulation, and although it plays an important role in the evolution of upper crustal structures, it has only a secondary significance in the wider process of continental breakup. The thinning of the feldspar layer at the crest of the mantle core complex could explain the absence or only local preservation of lower crustal rocks in the distal Iberia margin (Manatschal and Bernoulli, 1999).

The intramantle detachment and mantle core complex developed in this cold, magma-poor system are all related to the development of a strong elastic layer within the mantle. This evolution is directly analogous to that developed within the crust for hot continental lithospheres (Regenauer-Lieb et al., 2006). The main difference is that in hot environments, the mantle is too hot, and a strong elastic layer develops only in the quartz-dominated crust. This crustal elastic layer becomes the stress-bearing part of the system (Rosenbaum et al., 2005; Regenauer-Lieb et al., 2006), and a mid-crustal detachment and crustal core complexes develop without significant involvement of Moho topography. Thus, the difference in the mode of extension between hot and cold systems is primarily controlled by the vertical migration of the strongest elastic layer from the upper crust to the upper mantle.

The model results also help explain the "upper plate paradox" (Driscoll and Karner, 1998). In the context of continental rifting by asymmetric core complex development, the upper plate (hanging wall of detachments) should be exposed on one continental margin, and the lower plate (footwall) should be exposed on the other. However, commonly both margins have characteristics interpreted to correspond to the upper plates of detachments (Driscoll and Karner, 1998).

This paradox is related to a well-documented difference in the finite stretching of the upper crust, accommodated by and measured from brittle fault displacements, compared to the total lithospheric stretching estimates. The smaller values of upper crustal stretching are likely related to an increase in strain rate with depth, and exhumation of subcontinental mantle (Davis and Kusznir, 2004), as documented in nonvolcanic margins (e.g., Whitmarsh et al., 2001). Davis and Kusznir (2004) pointed out that existing models of formation of rifted continental margin fail to explain depth-dependent stretching, subcontinental mantle exhumation, and the upper plate paradox. Our results support their conceptual model to explain these features, where a broadly symmetrical subcontinental mantle dome develops, decoupled from continental crust by diffuse detachment zones dipping toward the continents, associated with strong channel flow and stretching of the ductile lower crust.

CONCLUSION

We have implemented a thermo-mechanical approach to investigate cold lithospheric extension of rheologically stratified crust. This has been done without prescribing additional weakness zones or large-scale thermal perturbation. Our model results have produced a number of fundamental features recognized in nonvolcanic passive margin, including listric upper crustal faults, extreme thinning of the lower crust, and within-mantle detachments that control exhumation of a symmetrical subcontinental mantle dome. Activity on these structures can eventually lead to continental breakup, replicating the inferred evolution of magma-poor extensional lithosphere reported in the literature and explaining the upper plate paradox. We note that the essence of our results, particularly the nature of the dynamic lithospheric strength profile and the origin of detachments above an elastic layer, might also explain the origin of oceanic core complexes formed in the vicinity of spreading ridges (Tucholke et al., 1998).

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