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ORIGINAL PAPER

Control of rheological stratification on rifting geometry: a symmetric model resolving the upper plate paradox

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Abstract Numerical experiments reproduce the fundamental architecture of magma-poor rifted margins such as the Iberian or Alpine margins if the lithosphere has a weak mid-crustal channel on top of strong lower crust and a horizontal thermal weakness in the rift center. During model extension, the upper crust undergoes distributed collapse into the rift center where the thermally weakened portion of the model tears. Among the features reproduced by the modeling, we observe: (1) an array of tilted upper-crustal blocks resting directly on exhumed mantle at the distal margin, (2) consistently oceanward-dipping normal faults, (3) a mid-crustal high strain zone at the base of the crustal blocks (S-reflector), (4) new ocean floor up against a low angle normal fault at the tip of the continent, (5) shear zones consistent with continentward-dipping reflectors in the mantle lithosphere, (6) the mismatch frequently observed between stretching values inferred from surface extension and bulk crustal thinning at distal margins (upper plate paradox). Rifting in the experiment is symmetric at a lithospheric scale and the above features develop on both sides of the rift center. We discuss three controversial points in more detail: (1) weak versus strong lower crust, (2) the deformation pattern in the mantle, and (3) the significance of detachment faults during continental breakup. We argue that the transition from wide rifting towards narrow rifting with a pronounced polarity towards the rift center is associated with the advective growth of a thermal perturbation in the mantle lithosphere.

Keywords Continental rifting · S-reflector · Detachment fault · Parallel-dipping faults · Numerical modelling

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Introduction

Over the past decades, deep sea drilling, seismic campaigns, diving cruises, and the investigation of Jurassic and Cretaceous margins in the Alps have greatly increased our knowledge about the architecture and subsidence history of passive continental margins. It has become clear that several large-scale observations cannot be explained satisfactorily by the classic symmetric rifting model of McKenzie (1978). A discussion has emerged on whether McKenzie's model needs to be refined or whether rifting geometry can be fundamentally different. Many studies have explained the geometry of continental margins through a rifting process that is asymmetric at the lithosphere scale (e.g. Boillot et al. 1988; Froitzheim and Manatschal 1996; Reston et al. 1996; Driscoll and Karner 1998; Manatschal and Bernoulli 1999; Whitmarsh et al. 2001). Asymmetric models usually propose the existence of enormous solitary normal faults, so-called detachments, cutting through the entire lithosphere (Wernicke 1985). Such detachments would accommodate tens of kilometers of extension. Ongoing displacement along a detachment would eventually lead to the breakoff of the continental crust and the exhumation of subcontinental mantle at the surface. According to McKenzie's model, two conjugate margins should mirror each other at a large scale. Asymmetric rifting models, on the other hand, predict fundamentally different conjugate margins with one margin being the footwall and the other being the hanging wall of the detachment, referred to as the lower- and upper-plate margins, respectively.

Figure 1 shows a cross-section through one of the best investigated margins, the Cretaceous Galicia Margin west of Portugal. The Galicia Margin and the adjacent Iberian Abyssal Plain Margin have been the subject of intense research including all of the above mentioned techniques, and often serve as reference margins in studies on magma-poor rifting (ODP legs 103, 149, and 173, Boillot et al. 1980, 1987, 1988, 1995; Mauffret and Mondatert 1987; Hoffmann and Reston 1992; Reston et al. 1996; Pickup et al. 1996; Whitmarsh et al. 2000; Pérez-Gussinyé et al. 2006). Researchers working on ancient margins now preserved in the Alps often compare their results with findings from the Galicia Margin (Manatschal and Bernoulli 1999; Whitmarsh et al. 2001; Manatschal 2004). The Cretaceous margin at the northern border of the Briançonnais domain, exposed in the Tasna nappe in eastern Switzerland, is currently interpreted by most authors as belonging to the same margin as the Galicia Margin (Florineth and Froitzheim 1994). Several features displayed in Fig. 1 appear at least to be typical of magma-poor margins including those found in the Alps (e.g. Manatschal and Bernoulli 1999; Whitmarsh et al. 2001):

1. At the distal margin, the crust tapers from approximately 25 km thick down to zero over a distance of 50–100 km. This taper consists of tilted blocks separated by consistently oceanward-dipping normal faults. On the continentward side of this crustal taper, there is a wide area of moderate crustal extension without any pronounced fault polarity towards the rift-center (e.g. Boillot et al. 1988).
2. The tip of the continent is made up of rocks with typical upper-crustal composition, often with pre-rift sediments preserved on top of the fault-bounded blocks. Lower crust is only scarcely preserved or completely absent (Manatschal and Bernoulli 1999; Whitmarsh et al. 2000). At magma-poor margins like the Galicia margin, the upper crust rests directly on exhumed subcontinental mantle. Among the most generally observed features of rift systems is that the entire distal margin is made up of rocks derived from the uppermost crust, and this finding has also been reported from several magmatic margins (Kusznir et al. 2004, and references therein).
3. The base of the crust coincides with a bright seismic reflector, the so-called S-reflector, which was first described in the Bay of Biscay (De Charpal et al. 1978). This reflector reaches the ocean floor at the tip of the continent, where it has yielded samples of greenschist-facies mylonites with a strong cataclastic overprint (Beslier et al. 1990; Boillot et al. 1995). The S-reflector is now viewed as a mid-crustal shear-zone with a top-to-the-ocean sense of shear (Manatschal et al. 2001). While a few of the oceanward-dipping faults cut the S-reflector, most of them appear to be rooted in this major shear-zone (Pickup et al. 1996; Reston et al. 1996).
4. At the tip of the continent, the ocean floor corresponds to the surface of a normal fault that is rotated into a subhorizontal orientation (Boillot et al. 1995). Hence, the distal continental margin represents the footwall of a normal fault.
5. At the distal margin and the adjacent oceanic lithosphere, continentward-dipping reflectors are found within the upper mantle. These reflectors are particularly well sampled at the western margins off the Iberian peninsula (Pickup et al 1996).

With the exception of the continentward-dipping reflectors, all of these observations have been used in the past to infer asymmetric rifting at the Iberian and Alpine margins, but not always with the same conclusion (Fig. 2). The structural architecture (i.e. the tilted blocks, the S-reflector, and the normal fault surface at the distal margin) lead several authors to assume that these margins were actually lower plate margins (e.g. Froitzheim and Manatschal 1996; Reston et al. 1996; Manatschal and Bernoulli 1999). The S-reflector was interpreted as the detachment and the parallel-dipping fault array as subsidiary faults in its hanging wall of it (Reston et al. 1996; Manatschal and Bernoulli 1999). On the other hand, the absence of lower crust at the distal margin was interpreted as resulting from a continentward-dipping detachment (Boillot et al. 1988). The enigmatic lack of deeper crust has far reaching implications for the syn-rift subsidence of the margin, and this subject has interested the hydrocarbon industry for decades. This absence of deep crust is in agreement with

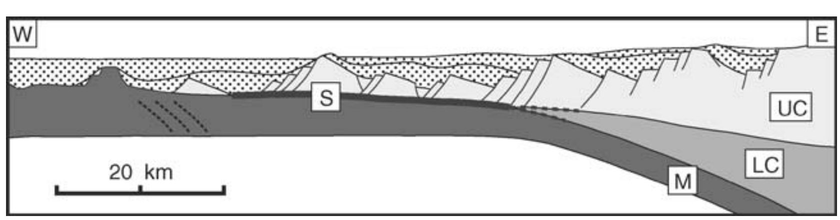

Fig. 1 Schematic cross-section through the Cretaceous Galicia Margin west of Portugal (Boillot et al. 1995). Abbreviations: *UC* upper crust; *LC* lower crust; *M* mantle; *S* S-reflector

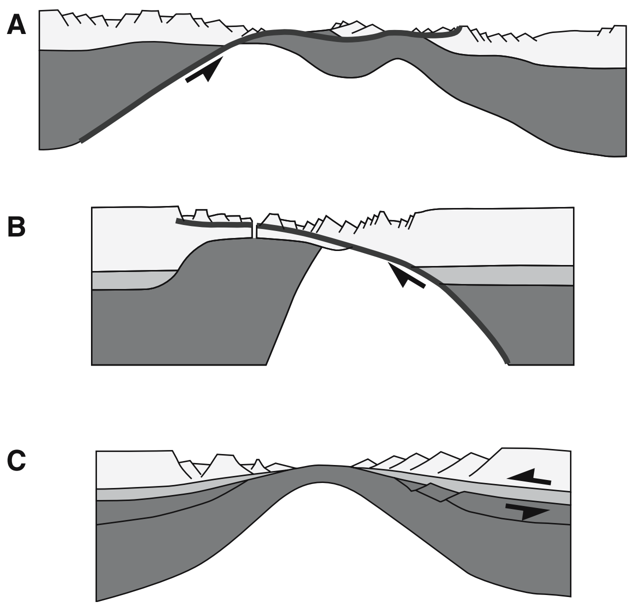

Fig. 2 Conceptual interpretations of the typical margin architecture shown in Fig. 1. **a** Based on the tilted block array, the oceanward-dipping faults, and the top-to-the-ocean-directed high-strain zone, Froitzheim and Manatschal (1996) inferred asymmetric rifting along an oceanward-dipping detachment at the Adriatic Margin in the Alps. For the same reasons, the Galicia Margin was also interpreted as a lower plate margin (e.g. Manatschal and Bernoulli 1999). **b** Based on the lack of lower crust at the distal margin, Boillot et al. (1988) proposed asymmetric rifting along a continentward-dipping detachment at the Iberian margins. **c** Brun and Beslier (1996) explained the observations in terms of an overall symmetric rifting model with weak lower crust acting as a ductile shear zone with a top–to-the-ocean sense of shear

the mismatch between bulk crustal thinning and stretching factors derived from normal faulting observed at the surface. At the distal margin, crustal thinning and associated subsidence is often greater than would be predicted from the extension accommodated by the observed normal faults. This feature is so common that some authors have proposed a general depth dependency of stretching during rifting, with stretching factors increasing with depth in the crust (Davis and Kusznir 2004; Kusznir et al. 2004). Authors studying crustal thinning and subsidence history usually identify upper plate margins when attempting to explain their data in terms of asymmetric rifting (Driscoll and Karner 1998). This apparent contradiction has become known as the upper plate paradox (Driscoll and Karner 1998). It seems that studies focusing on the structural record commonly identify lower plate margins, whereas studies on subsidence and bulk architecture lead to interpretations in terms of an upper plate margin. Unfortunately, even though the Newfoundland margin is conjugate to the Iberian margins, it is less well known and observations from this margin are inconclusive with respect to the problem of asymmetric rifting. In the Alps, continental margins are well preserved only in the hanging walls of later subduction zones. All conjugate margins in this area have been subducted, and are either completely lost or strongly deformed and/or metamorphosed. It appears that the well-studied Mesozoic margins west of Iberia and in the Alps show the structural arrangement expected of lower plate margins. However, these same margins have been interpreted as upper plate margins based on the distribution of crustal thinning and associated subsidence.

In a recent short paper (Nagel and Buck 2004), we proposed that the above described architecture of rifted margins could be explained by an overall symmetric model in which the lithosphere has a particular vertical and horizontal stratification. Here, we summarize this model and then present a series of numerical modeling experiments with varying lower-crustal strengths. Our aim is to discuss some controversial aspects of continental rifting and breakup that have previously received little or no attention. We cover the following points: (1) the strength of the lower crust, (2) the deformation pattern in the mantle, including the significance of the continentward-dipping reflectors, and (3) the role of detachment faults in the rifting process.

A symmetric model

With a weak layer in the mid crust and a pronounced horizontal weakness at the rift center, numerical models reproduce the above mentioned key elements of rifted continental margins on both sides of the rift center (Nagel and Buck 2004). Figure 3 shows the setup of such an experiment. The continental crust consists of three layers: (1) a brittle upper crust, (2) a weak linear-viscous middle crust and (3) a relatively strong lower crust with the brittle and ductile properties of diabase (Ranalli 1995). The mantle follows a flow law for dry olivine (Ranalli 1995) and is therefore also strong. The horizontal weakness is imposed through the addition of a pronounced thermal perturbation in the rift center (Fig. 3). Virtually all the strength beneath the brittle crust is taken away in the center of the thermal swell. On the rift flanks, the strong lower crust is firmly coupled to the upper mantle (Fig. 3).

During extension, mantle and lower crust show localized necking at the site of the thermal perturbation (Fig. 4). Extension in the brittle upper crust is accommodated through a series of dominantly inward-dipping faults on both sides of the rift center (Fig. 4a). Within the weak mid-crustal layer, the distributed deformation in the upper crust is transferred into the necking site of the deeper lithosphere. In other words, a shear zone connects the brittle faults with the center of the rift. Hence, the upper crust

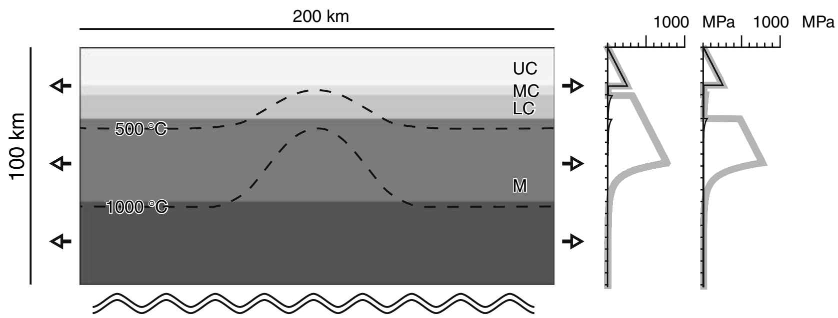

Fig. 3 Setup for the numerical experiment shown in Fig. 4. The model is 200 km wide, 100 km deep and has four layers. *UC* upper crust (16 km, elasto-plastic); *MC* middle crust (4 km, visco-elastic); *LC* lower crust (10 km, visco-elasto-plastic); *M* mantle (70 km, visco-elasto-plastic). The boundary at 65 km is added only to visualize deformation in the mantle, and corresponds approximately to the transition from spinel peridotite to garnet peridotite. The layers UC, LC and M follow a Mohr-Coulomb yield criterion. Brittle strain localization is achieved through a cohesion and friction loss with brittle strain. Between 0.5 and 50 % brittle strain, the cohesion and the friction angle are reduced from 10 to 5 MPa and from 30° to 28°, respectively. The middle crust has a linear viscous behavior with a viscosity of 10^{20} Ps. Lower crust and mantle follow temperature-dependent and non-linear flow laws for diabase and olivine, respectively (Ranalli 1995). Viscosities are cut off at 10^{20} Ps, i.e. no viscosities lower than 10^{20} Ps are allowed. The background thermal gradient is 15°C/km, and the thermal gradient in the center of the perturbation is 30°C/km. The left and right sides move laterally with a velocity of 2.5×10^{-10} m/s (1.58 cm/a). The lower boundary is floating and deformable (Winkler foundation). Mantle material is added to replace material that is pulled upward. The upper surface is unconstrained. The grid is made up of 100 × 200 elements. Two strength profiles illustrate the effect of a thermal perturbation for a lithosphere with strong (*left*) and weak (*right*) lower crust (using diabase and wet granite, respectively, from Ranalli 1995). Each *section* shows a profile for the center of the thermal swell (30°/km—*thin black line*) and a section for the thermal background gradient (15°/km—*thick gray line*). Viscous strength is calculated for a strain rate of 5×10^{-14} s^{-1}. An explicit time-marching, Lagrangian finite difference technique (*FLAC*) is used for calculation (Cundall and Board 1988; Cundall 1989; Poliakov et al. 1994). It is well described in the cited references

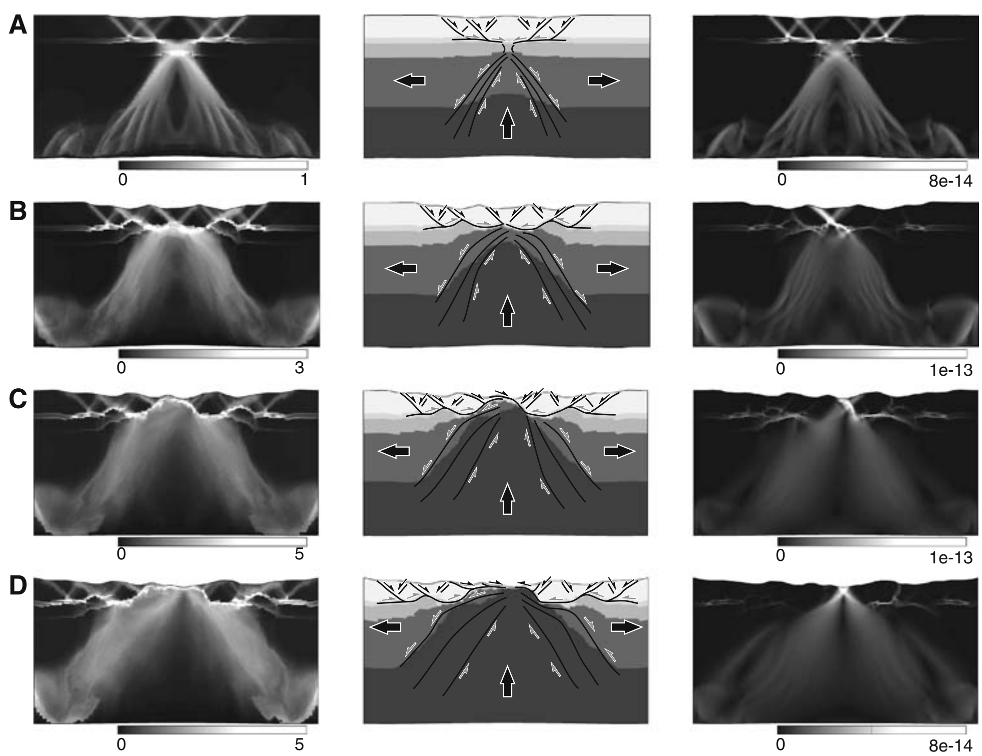

Fig. 4 Numerical model using the setup shown in Fig. 3, after about 10 (**a**), 45 (**b**), 75 (**c**), and 100 (**d**) km of extension. *Left column* total strain (brittle + viscous); *middle column* distribution of upper crust, middle crust, lower crust and mantle. Color code is the same as in Fig. 3. *Black lines* indicate shear zones, *small arrows* shear senses (*black* brittle; *gray* viscous). *Thick black arrows* indicate the general motion of lithospheric blocks; *right column* snapshots of total strain rate (brittle and viscous) at the corresponding stage of deformation

collapses into the opening rift center on top of a weak mid-crustal high-strain zone, which exhibits a top-to-the-rift center sense of motion. The lower crust tears after about 25 km of extension and is further passively moved to the side. After about 35 km of extension, the fault array in the upper crust is largely abandoned and faulting localizes in the rift center (Fig. 4b). The upper crust breaks up after about 60 km of extension as subcontinental mantle is exposed at the surface (Fig. 4c). Further deformation is localized in the newly formed oceanic lithosphere. The resulting margins show striking similarities to the cross-section in Fig. 1 (Fig. 4d): the margins consist of arrays of tilted blocks with normal faults dipping towards the rift center. At the tip of the continent, upper crust rests on subcontinental mantle and the base of the tilted blocks corresponds to a mid-crustal high-strain zone with top-to-the-ocean sense of shear. This shear zone reaches the ocean floor at the tip of the continent. Finally, the newly formed surface-area at the distal margin is represented by a rotated normal fault.

The numerical modeling also predicts the apparently stronger thinning of the deeper parts of the crust at both distal margins, and thus provides a solution for the upper plate paradox. Thinning does not occur due to more intense stretching in the lower crust, but rather because stretching is more localized than in the upper crust, i.e. the lower crust is necked away from the site of future breakup, whereas the upper crust collapses into the rift center on top of the un-thinned lower crust on the rift flanks. Integrating the amount of stretching over the entire rift system gives the same result for all levels of the lithosphere. This might provide a way to test our model against models of true depth-dependent stretching. According to the presented scenario, strongly thinned crust not balanced by normal faulting at the surface should be restricted to the distal margin, whereas, at more proximal positions, upper-crustal extension should be stronger than the bulk crustal thinning.

Our model is mechanically very similar to previous attempts to explain the architecture of magma-poor continental margins through an overall symmetric rifting model (Brun and Beslier 1996; Boillot and Coulon 1998). These authors already proposed a top-to-the-ocean-directed ductile shear zone in the lower crust on both sides of the rift, with very similar deformation patterns in the mantle lithosphere (see section on mantle deformation below). There are two main differences between our model and existing models. First, we assume that the lower crust needs to be strong and coupled to the upper mantle in order to explain the architecture of rifted margins. Second, we emphasize that the weak layer beneath the brittle upper crust has to be thin to ensure distributed deformation and the formation of parallel dipping faults (Nagel and Buck 2004; Nagel and Buck 2006). In the following sections, we discuss three controversial aspects of rifting in more detail: (1) the problem of weak versus strong lower crust, (2) the geometry of deformation in the mantle lithosphere, and (3) the significance of detachment faults during continental breakup.

Three controversial subjects

Weak versus strong lower crust

In recent years, there has been a discussion about the rheological properties of the lower crust particularly during rifting. Some studies have inferred weak lower crust during rifting (Brun and Beslier 1996; Whitmarsh et al. 2000). We consider that the observed rift geometry can only be explained by a relatively strong lower crust that undergoes boudinage at the lithospheric scale (Nagel and Buck 2004; Lavier and Manatschal 2006). A number of authors have proposed a lower crust considerably stronger than the middle crust, based on experimental work (e.g. Ord and Hobbs 1989) as well as regional studies of extensional (Handy and Zingg 1991; Müntener et al. 2000) and collisional (Schmid et al. 1996) tectonics. The petrological argument is that lower crust would be compositionally different from upper crust and/or might be water depleted. The lower crust of the Adriatic plate exposed in the Eastern and Central Alps is indeed rich in mafics and granulites (Müntener et al. 2000). Petrological studies have shown that this lower crust was not particularly hot at the onset of rifting and was therefore probably strong (Müntener et al. 2000). For the calculation presented in our Fig. 4, we use a strong flow law for a mafic composition, i.e. diabase (Ranalli 1995).

Figure 5 shows a numerical experiment calculated with exactly the same setup as Fig. 4, the only difference being that a wet granite flow law (Ranalli 1995) is used to define the properties of the lower crust. Under lower-crustal conditions, the viscosity of deforming granite is as low as the weak mid-crustal layer even on the rift flanks. Thus, the entire upper crust now rests on top of a 14-km-thick weak channel. The resulting deformation pattern in the crust is very different from the one calculated for a mafic lower crust (Fig. 4). The resistance of a weak viscous channel to localized faulting in the overlying brittle layer depends on the viscosity (Huismans et al. 2005), and particularly on the thickness of the channel. A thicker channel promotes localization of the faulting (Nagel and Buck 2006). Figure 5 shows deformation in the upper crust becoming rapidly concentrated at a single fault (a crustal detachment). This is due to the thick channel, which allows the weak lower crust to flow easily into the developing depression. This flow is illustrated by high-strain zones with opposite sense of shear developing at the upper and lower surfaces of the weak channel, especially in the footwall block of the detachment (Fig. 5b–d). As a result, highly deformed lower crust accumulates beneath the crustal detachment and is subsequently exhumed. The efficient spreading of lower crust also implies a larger amount of extension until crustal breakup (breakup is not

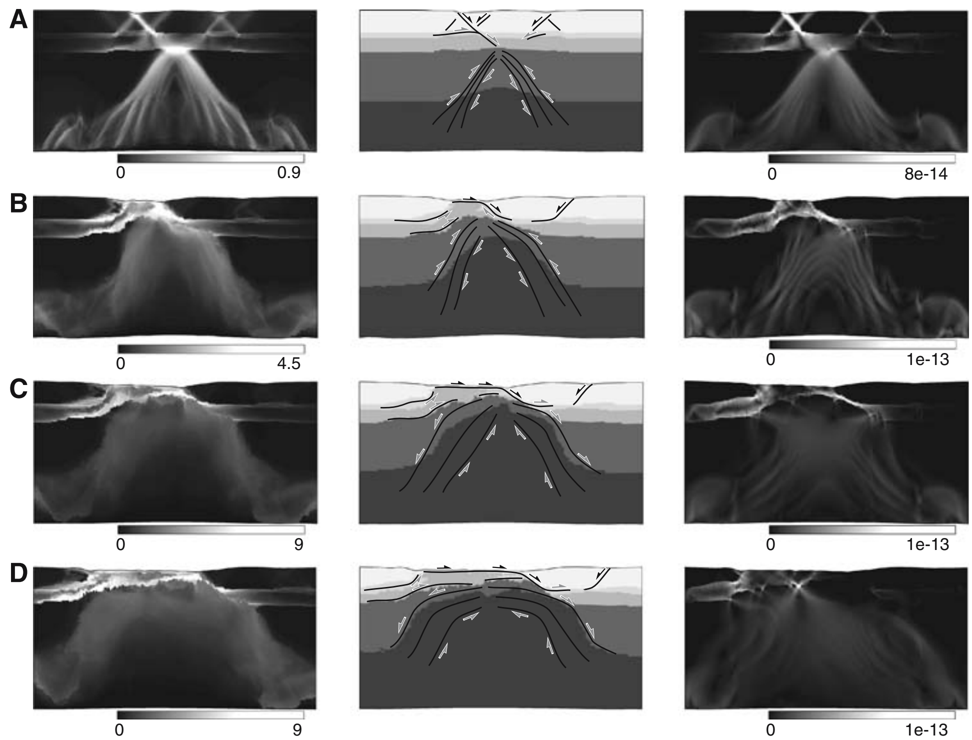

Fig. 5 Same model as in Fig. 4, except that a wet granite flow law (Ranalli and Murphy 1987) is assumed for the lower crust

even completed after 100 km of extension, as shown in Fig. 5d). The resulting margins are asymmetric at a crustal scale. The footwall of the crustal detachment shows highly deformed lower-crustal rocks across more than 50 km at the distal margin, whereas the upper plate is made up of undeformed upper-crustal rocks. On the side of the developing upper plate margin, crustal thickness increases more rapidly than on the side of the lower plate margin.

Figure 6 gives a more complete picture of the effects of lower-crustal rheology on the fault pattern in the upper crust. It shows five experiments with successively decreasing lower-crustal strength after 75 km of extension (a: dry diabase, Mackwell et al. 1998; b: diabase, Ranalli 1995; c: diorite, Ranalli and Murphy 1987; d: dry granite, Ranalli 1995; e: wet granite, Ranalli 1995). Figure 6b corresponds to Fig. 4c, and Fig. 6e corresponds to Fig. 5c. There is an evident increase in the amount of strain localization from stronger to weaker flow laws. Only experiments 6a and 6b reproduce the passive margin architecture shown in Fig. 1. In experiments 6c and 6d, the central block does not fail and two more or less well developed core complexes are located on both sides of the rift. As seen before, there is only a single core complex developing in Fig. 6e. In Fig. 6d and e, considerable amounts of lower crust are accumulated in the footwall of the detachments.

The crustal detachment in Fig. 5 does not form exactly above the thermal perturbation. Instead, it is offset from the necking site in the upper mantle lithosphere where asthenospheric mantle is being pulled up. As a result, the two margins show the topographic evolution proposed for asymmetric rifting (Fig. 7, Wernicke 1985), although extension in the upper mantle lithosphere is not accommodated along a solitary detachment. The upper plate margin with almost completely preserved crust, which is located above the thermal swell in the asthenosphere, is topographically higher than the lower plate margin with thinned crust and colder mantle lithosphere. Hence, the experiment shows similarities with rifting models assuming lithospheric detachments because failure in the crust and mantle are decoupled by a thick weak zone. Such a result was also obtained in physical experiments presented by Michon and Merle (2003). However, this contrasts strongly with the numerical modeling of Huismans and Beaumont (2002, 2003), who predicted that asymmetric rifting only occurs if the entire lithosphere is completely coupled and brittle. In summary, while numerical experiments simulating rifted margins with a thick weak lower crust are relevant for general discussion about symmetric and asymmetric rifting, the situation appears completely different from the Iberian and Alpine margins. The lower plate margin of the crustal detachment should show tens of kilometer of exposed lower crust at the distal margin. With a weak lower crust, the extension of the upper crust is more localized than for the lower crust. This is contrary to the observations. Moreover, the upper plate margin of Fig. 5

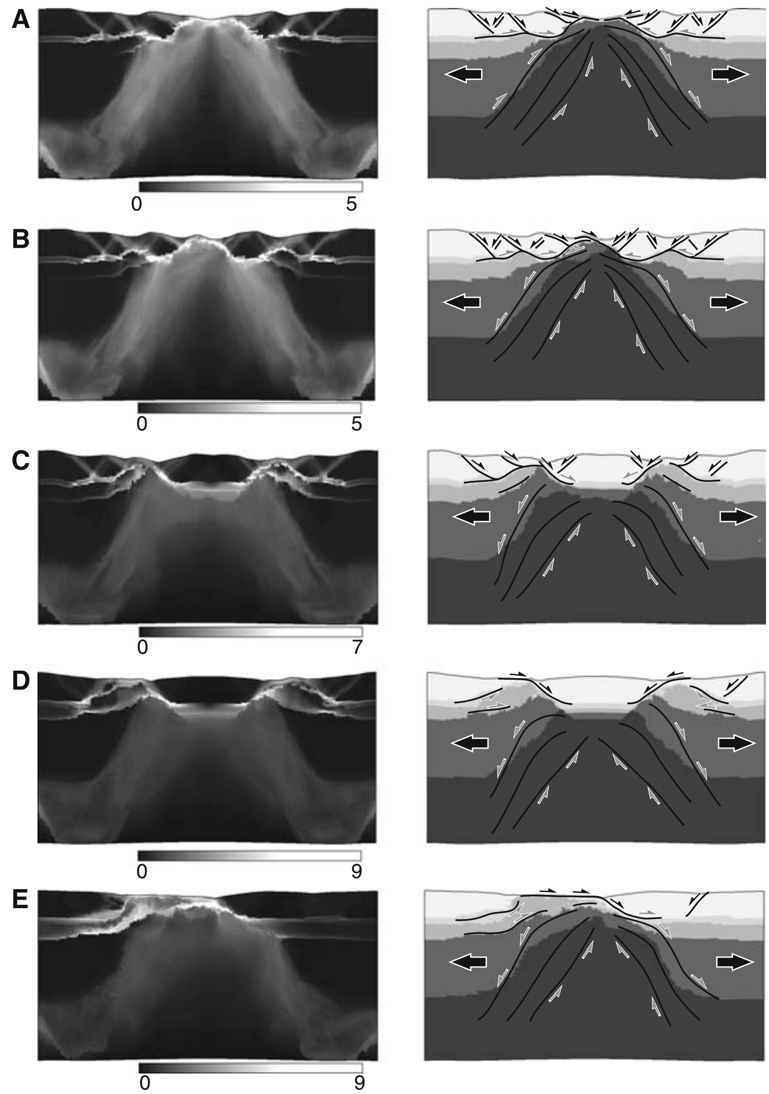

Fig. 6 Numerical models using the setup shown in Fig. 3 for five different lower-crustal rheologies (**a** dry diabase, Mackwell et al. 1998; **b** diabase, Ranalli 1995; **c**: diorite, Ranalli and Murphy 1987; **d**: dry granite, Ranalli 1995; **e**: wet granite, Ranalli 1995). *Left column* total strain after about 75 km of extension; *right column* distribution of upper crust, middle crust, lower crust, and mantle. Same key as middle column in Figs. 4 and 5

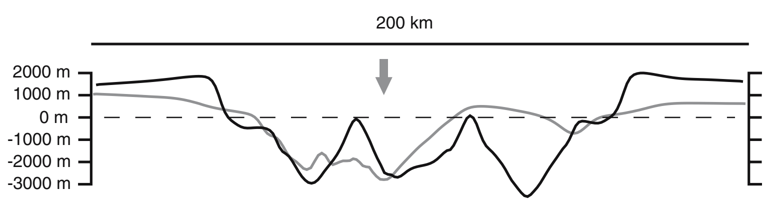

Fig. 7 Rift topography of the two experiments shown in Figs. 4 (*black*) and 5 (*gray*) after 45 km of extension. *Arrow* indicates emergence of the crustal detachment at the seafloor in Fig. 5

represents a poorer fit to the Iberian margin than the two margins in Fig. 4. There is no formation of an array of oceanward-dipping faults, and even the crustal structure at a large scale is different since the crust thins over a much shorter distance. This occurs because the crustal detachment in the experiment has a steep dip, and thus only rotates into a horizontal orientation along its exhumed portion in the lower plate margin. We conclude that the lower crust must have been strong during the formation of the Iberian and Alpine margins.

Deformation and exhumation of subcontinental mantle

Symmetric rifting models should show very different deformation fields in the mantle when compared to asymmetric models involving a lithospheric detachment. Accordingly, different structures and metamorphic histories should be preserved in the exhumed mantle at the distal continental margin. Continentward-dipping reflectors in the upper mantle (Pickup et al. 1996) have been interpreted as a succession of shear zones that become younger towards the rift center (Brun and Beslier 1996; Boillot and Coulon 1998). These shear zones would represent a system that is conjugate to the oceanward-dipping faults in the upper crust, and should be present on both sides of the rift. This corresponds to the pattern observed in our models. Plots of finite strain (Fig. 4, left column) show bell-shaped highly strained domains with a general dip towards the continent in the mantle beneath the evolving margins. Strain rate plots (Fig. 4, right column) illustrate how these margins evolve. Only the inner portion of the strained area is active as a shear zone. The mantle beneath these zones is pulled upwards, while the material above them moves sideways with the entire lithosphere (Fig. 4, middle column). Hence, the active shear zones have a normal sense of shear. Material flows through the shear zones and is further transported passively to the side. The portions of the bell-shaped domain on the continentward side represent material that has been deformed in the same shear zone at an earlier stage of rifting. Our model thus confirms existing ideas about the origin of continentward-dipping reflectors in an overall symmetric rift setting (Brun and Beslier 1996; Boillot and Coulon 1998). The ductile shear zones in the upper mantle and the conjugate top-to-the-ocean-directed shear zones at the base of the brittle upper crust (the S-reflector) merge at the tip of the strong lithospheric wedge that is extracted to the side (Fig. 4). As a result, the two shear zones with opposite sense of shear come into direct contact. Froitzheim et al. (2006) have coined the term ''extraction fault'' for such a surface. At the Galicia margin, Beslier et al. (1990) have reported opposite-directed shear senses in peridotites at the tip of the continent. Greenschist-facies to brittle structures show top-to-the-ocean directed motion, whereas opposite-directed shearing occurs at temperatures of 850–1,000°C.

In a scenario of lithospheric detachments, the mantle material at the tip of the continent should be located immediately beneath the crust at the onset of rifting. The symmetric model in Fig. 4 predicts mantle material pulled up from greater depth at this location. Although some material from shallower depth is sheared towards the rift center, most of the exhumed mantle is at depths greater than 60 km at the start of rifting (stability field of garnet peridotite; see color code in Fig. 4, middle column). Studying the pressure–temperature–time evolution of samples from continental margins might provide us with a way of testing the rifting models. Foliated ultramafic rocks from the Platta nappe in Eastern Switzerland equilibrated in the stability field of spinel peridotite (Desmurs et al. 2001), therefore at depths shallower than 65 km (Bucher and Frey 1994). In a symmetric rift setting, mantle rocks not far from the tip of the continent should have been exhumed from depths within the stability field of garnet peridotite. However, it is very likely that garnet peridotites would re-equilibrate to spinel peridotites during deformation associated with exhumation. Away from the margin, ultramafics in the Platta nappe are even present as feldspar peridotites (Desmurs et al. 2001) and have evidently re-equilibrated close to the surface. In the absence of time constraints, the petrological record should be regarded as inconclusive.

Detachment-faults during continental breakup

In the past few years, several authors have proposed that the formation of detachment faults would be restricted to the final stage of rifting, i.e. when the crust is already thinned down to a thickness of 7–10 km (Whitmarsh et al. 2001; Manatschal 2004; Lavier and Manatschal 2006). Recent structural and petrological studies on the Tasna detachment in the Alps have revealed that this fault at the continent–ocean transition does not have a high-grade metamorphic portion and was active only at relatively shallow levels (Manatschal et al. 2006). A transition from symmetric graben-type faulting to core-complex formation can be triggered by thermal advection and associated thinning of the strong upper portion of the lithosphere. Thinning of the brittle layer promotes the formation of single faults with large offsets, as the footwall of the normal fault can be rotated without the occurrence of huge bending stresses (Lavier et al. 2000). To a certain extent, the experiment illustrated in Fig. 4 shows this type of behavior. The right-dipping fault in the rift center is more active than the conjugate left-dipping fault during late rifting (Fig. 4b, c). The gradual transition from a graben-like to a single-fault mode starts when the crust is thinned to 10–12 km. The right-dipping fault accumulates some twenty km of offset and accounts for the ultimate breakup of the continental crust (Fig. 4c). The ductile continuation of this fault in the mantle shows a steeper dip than the fault plane itself. Hence, the entire shear zone can be viewed as a downward steepening detachment fault (Manatschal 2004; Lavier and Manatschal 2006). The strong activity of this fault during late rifting causes a slight asymmetry of the two evolving margins. Mantle exposed at the tip of the continent is derived from shallower depths in the footwall of this fault than in the hanging wall (see reference horizon in the mantle in Fig. 4d). Hence, our model accords with the view that the ultimate continental breakup can be associated with a detachment fault (Manatschal 2004; Lavier and Manatschal 2006). However, this asymmetry is not expressed in the fundamental characteristics of the two conjugate margins.

Discussion

The experiment shown in Fig. 4 reproduces all the major characteristics of magma-poor margins in an overall symmetric rifting geometry. At a late stage of rifting, a degree of transition takes place towards asymmetric rifting along a single detachment fault in the crust. However, none of the key observations that were viewed in the past as supporting asymmetric rifting are related to this stage. These features include the formation of the S-reflector, the formation of the tilted block array, and the lack of deeper crust at the distal margin, as well as the associated mismatch of stretching factors. We propose that all these features develop during symmetric rifting. Our model requires a specific rheological stratification of the lithosphere, i.e. the presence of a thin weak layer in the mid crust and a pronounced horizontal weakness in the rift center. Both of these weaknesses are strong and predefined in our experiments. A number of studies have inferred that the viscous middle crust is weaker than the lower crust. The choice of a linear viscous middle crust of 10^{20} Ps in our experiments is clearly somewhat arbitrary. Evidently it would be more satisfactory to use a model with three visco-elasto-plastic layers bahaving like the experiment in Fig. 4. However, using a granitic upper crust, and assuming a non-linear visco-elasto-plastic flow law (with a thermally defined brittle-ductile transition), we did not observe the development of a mid-crustal high-strain zone and deformation remained localized in the thermally weak rift center. While we achieved some success using viscous strain weakening in a non-linear granitic upper crust (see Lavier and Manatschal 2006), the results obtained are not so clear as with a linear viscous middle crust. Also, the predefined thermal perturbation has an arbitrary magnitude—it acts almost as a gap in the lithosphere (Fig. 3). In nature, unweakened lithosphere is probably so strong that it cannot be effectively rifted (Buck 2004). Numerical modeling experiments with a constant extension velocity and without any thermal anomaly lead to distributed faulting in the upper crust and unrealistically high stresses (>500 MPa) in the upper mantle. Deformation can be localized by a small perturbation in the mantle lithosphere, which may be caused, for example, by a magmatic event. If the thermal advection at this site escapes thermal diffusion, the perturbation starts to grow and a runaway process is initiated. We obtained similar results with a smaller thermal perturbation (e.g. 23°C/km instead of 30°C/km in the center of the rift, 15°C/km on the rift flanks). This is because the extension velocity is relatively high (1.6 cm/a) and even the smaller perturbation represents a significant relative weakness and thus efficiently localizes the deformation. We propose that the rift geometry can be explained by the formation of a ‘‘runaway weakness’’ in an otherwise strong upper mantle and lower crust. This weakness does not need to be produced entirely by a thermal anomaly. Runaway effects can also be initiated by viscous strain weakening.

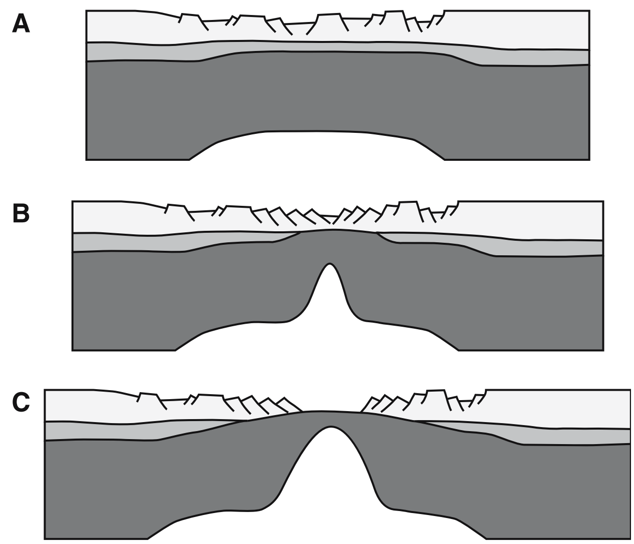

Fig. 8 Conceptual sketch showing different rifting stages of magma-poor margins. **a** Early rifting without a horizontal weakness—extension is homogeneous at a lithospheric scale. **b** Rifting stage after development of a central weakness localizing the deformation in the mantle lithosphere and lower crust. **c** Spreading stage—deformation is entirely localized, margins are tectonically quiet. The architecture of distal margins is mainly formed during stage B

The early stage of rifting without any localized weakness might be recorded in the wide area of moderate extension typically found at the continentward side of the tapering crust at the distal margin. During this phase, rifting is probably slow and uniform at the lithospheric scale, and normal faults in the upper crust do not show the consistent dip towards the site of future breakup (Fig. 8a). With the development of a thermal and/or mechanical weakness, deformation becomes localized in the deeper lithosphere and the adjacent upper crust collapses into the rift center (Fig 8b). Once the crust tears, the site of strongly localized deformation in the mantle reaches the surface (Fig. 8c). The presence of a crustal wedge, tilted block array and S-reflector characterize the rifting stage, during which the mantle lithosphere and the lower crust are already subject to spreading beneath the crust.

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References

Beslier M-O, Girardeau J, Boillot G (1990) Kinematics of peridotite emplacement during North Atlantic continental rifting, Galicia, northwestern Spain. Tectonophysics 184:321–343

Boillot G, Coulon C (1998) La déchirure continentale et l’ouverture océanique: Géologie des marges passives. Gordon and Breach, Paris, pp 1–208

Boillot G, Grimaud S, Mauffret A, Mougenot D, Kornprobst J, Mergoildaniel J, Torrent G (1980) Ocean-continent boundary of the Iberian margin: a serpentinite diapir west of the Galicia Bank. Earth Planet Sci Lett 48:23–34

Boillot G, Recq M, Winterer EL et al (1987) Tectonic denudation of the upper mantle along a passive margin: a mode based on drilling results (ocean drilling program Leg 103, Western Galicia Margin, Spain). Tectonophysics 132:335–342

Boillot G, Winterer EL, Meyer A et al (1988) Proceedings of the ocean drilling project, scientific results, 103. Ocean Drilling Program, College Station, TX, pp 1–848

Boillot G, Agrinier P, Beslier MO, Cornen G, Froitzheim N, Gardien V, Girardeau J, Gil-Ibarguchi J-I, Kornprobst J, Moullade M, Schaerer U, Vanney JR (1995) A lithospheric syn-rift shear zone at the ocean-continent transition: Preliminary results of the GALINAUTE II cruise (Nautile dives on the Galicia Bank, Spain). C R Acad Sci (Paris) Sér IIa 322:1171–1178

Brun JP, Beslier MO (1996) Mantle exhumation at passive margins: Earth planet. Sci Lett 142:162–173

Bucher K, Frey M (1994) Petrogenesis of metamorphic rocks. Springer, Berlin, pp 1–318

Buck R (2004) Consequences of asthenospheric variability in continental rifting. In: Karner GD, Taylor B, Driscoll NW, Kohlsted LD (eds) Rheology and deformation of the lithosphere at continental margins. Columbia University Press, NewYork, pp 1–30

Cundall PA (1989) Numerical experiments on localization in frictional materials. Ing Arch 58:148–159

Cundall PA, Board M (1988) A microcomputer program for modelling large-strain plasticity problems. In: Swoboda G (ed) Numerical methods in geomechanics. Balkema, Rotterdam, pp 2101–2108

Davis M, Kusznir NJ (2004) Depth-dependent lithospheric stretching at rifted continental margins. In: Karner GD (ed) Proceedings of NSF rifted margins Theoretical Institute. Columbia University Press, New York, pp 92–136

De Charpal O, Guennoc P, Mondatert L, Roberts DG (1978) Rifting, crustal attenuation and subsidence in the Bay of Biscay. Nature 275:706–711

Desmurs L, Manatschal G, Bernoulli D (2001) The Steinmann trinity revisited: mantle exhumation and magmatism along an ocean-continent transition: the Platta nappe, eastern Switzerland. In: Wilson RCL, Whitmarsh RB, Taylor B, Froitzheim N (ed) Non-volcanic rifting of continental margins: evidence from land and sea. Geol Soc (London) Spec Publ 187:235–266

Driscoll NW, Karner GD (1998) Lower crustal extension across the Northern Carnarvon Basin, Australia: evidence for an eastward dipping detachment. J Geophys Res 103:4975–4992

Florineth D, Froitzheim N (1994) Transition from continental to oceanic basement in the Tasna nappe (Engadine window, Graubünden, Switzerland): evidence for Early Cretaceous opening of the Valais ocean. Schweiz Mineral Petrogr Mitt 74:437–448

Froitzheim N, Manatschal G (1996) Kinematics of Jurassic rifting, mantle exhumation, and passive margin formation in the Austroalpine and Penninic nappes (eastern Switzerland). Geol Soc Am Bull 108:1120–1133

Froitzheim N, Pleuger J, Nagel TJ (2006) Extraction faults. J Struct Geol 28:1388–1395

Handy MR, Zingg A (1991) The tectonic and rheological evolution of an attenuated cross section of the continental crust: Ivrea crustal section, southern Alps, northwestern Italy and southern Switzerland. Geol Soc Am Bull 103:236–253

Hoffmann HJ, Reston TJ (1992) Nature of the S reflector beneath the Galicia Banks rifted margin: preliminary results from pre-stack depth migration. Geology 20:1091–1094

Huismans RS, Beaumont C (2002) Asymmetric lithospheric extension: the role of frictional plastic strain softening inferred from numerical experiments. Geology 30:211–214

Huismans RS, Beaumont C (2003) Symmetric and asymmetric lithospheric extension: relative effects of frictional-plastic and viscous strain softening. J Geophys Res 108NOB10:2496. doi:10.1029/2002JB002026

Huismans RS, Buiter SJH, Beaumont C (2005) Effect of plastic-viscous layering and strain softening on mode selection during lithospheric extension. J Geophys Res 110:B02406. doi:10.1029/2004JB003114

Kusznir NJ, Hunsdalew R, Roberts AM (2004) Timing of depth-dependent lithosphere on the S. Lofoten rifted margin offshore mid-Norway: pre-breakup or post-breakup?. Basin Res 16:279–296

Lavier LL, Manatschal G (2006) A mechanism to thin the continental lithosphere at magma-poor margins. Nature 440:324–328

Lavier LL, Buck WR, Poliakov ANB (2000) Factors controlling normal fault offset in an ideal brittle layer. J Geophys Res 105:23431–23442

Mackwell SJ, Zimmerman ME, Kohlstedt DL (1998) High-temperature deformation of dry diabase with application to tectonics on Venus. J Geophys Res 103:975–984

Manatschal G (2004) New models for evolution of magma-poor rifted margins based on a review of data and concepts from West Iberia and the Alps. Int J Earth Sci 93:432–466

Manatschal G, Bernoulli D (1999) Architecture and tectonic evolution of nonvolcanic margins: present-day Galicia and ancient Adria. Tectonics 18:1099–1199

Manatschal G, Froitzheim N, Rubenach M, Turrin BD (2001) The role of detachment faulting in the formation of an ocean-continent transition: Insights from the Iberia Abyssal Plain. In: Wilson RCL, Whitmarsh RB, Taylor B, Froitzheim N (eds) Non-volcanic rifting of continental margins: evidence from land and sea. Spec Publ Geol Soc (London) 187:405–428

Manatschal G, Engström A, Desmurs L, Schaltegger U, Cosca M, Müntener O, Bernoulli D (2006) What is the tectono-metamorphic evolution of continental break-up: the example of the Tasna ocean continent transition. J Struct Geol 28:1849–1869

Mauffret A, Mondatert L (1987) Rift tectonics on the passive continental margin off Galicia (Spain). Mar Petrol Geol 4:49–70

McKenzie DP (1978) Some remarks on the development of sedimentary basins. Earth Planet Sci Lett 40:25–32

Michon L, Merle O (2003) Mode of lithospheric extension: conceptual models from analogue modeling. Tectonics 22:1028. doi:10.1029/2002TC001435

Müntener O, Hermann J, Trommsdorff V (2000) Cooling history and exhumation of lower-crustal granulite and upper mantle (Malenco, eastern central Alps). J Petrol 41:175–200

Nagel TJ, Buck WR (2004) Symmetric alternative to asymmetric rifting models. Geology 32:937–940

Nagel TJ, Buck WR (2006) Channel flow and the development of parallel-dipping normal faults. J Geophys Res 111:B08407. doi:10.1029/2005JB004000

Ord A, Hobbs BE (1989) The strength of the continental crust, detachment zones and the development of plastic instabilities. Tectonophysics 158:269–289

Pérez-Gussinyé M, Morgan JP, Reston TJ, Ranero CR (2006) The rift to drift transition at non-volcanic margins: insights from numerical modelling. Earth Planet Sci Lett 244:458–473

Pickup SLB, Whitmarsh RB, Fowler CMR, Reston TJ (1996) Insight into the nature of the ocean-continent transition of west Iberia from a deep multichannel seismic reflection profile. Geology 24:1079–1082

Poliakov ANB, Herrman HJ, Podladchikov YY, Roux S (1994) Fractal plastic shear bands. Fractals 2:567–581

Ranalli G (1995) Rheology of the Earth, 2nd edn. Chapman & Hall, London, pp 1–413

Ranalli G, Murphy DC (1987) Rheological stratification of the lithosphere. Tectonophysics 132:281–295

Reston TJ, Krawczyk CM, Kläschen D (1996) The S reflector west of Galicia (Spain): evidence for detachment faulting during continental breakup from prestack depth migration. J Geophys Res 101:8075–8091

Schmid SM, Pfiffner OA, Froitzheim SN, Schönborn G, Kissling E (1996) Geophysical-geological transect and tectonic evolution of the Swiss-Italian Alps. Tectonics 15:1036–1064

Wernicke B (1985) Uniform-sense normal simple shear of the continental lithosphere. Can J Earth Sci 22:108–125

Whitmarsh RB, Dean SM, Minshull TA, Tompkins M (2000) Tectonic implications of exposure of lower continental crust beneath the Iberia Abyssal Plain, northeast Atlantic Ocean: Geophysical evidence. Tectonics 19:919–942

Whitmarsh RB, Manatschal G, Minshull TA (2001) Evolution of magma-poor continental margins from rifting to seafloor spreading. Nature 413:150–154