The role of inheritance in structuring hyperextended rift systems: some considerations based on observations and numerical modelling

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Abstract

A long-standing question in Earth Sciences is related to the importance of inheritance in controlling tectonic processes. In contrast to physical processes that are generally applicable, assessing the role of inheritance suffers from two major problems: firstly, it is difficult to appraise without having insights into the history of a geological system; and secondly all inherited features are not reactivated during subsequent deformation phases. Therefore, the aim of this paper is to give some conceptual framework about how inheritance may control the architecture and evolution of hyperextended rift systems.

In this paper, we use the term inheritance to refer to the difference between an "ideal" layer-cake type lithosphere and a "real" lithosphere containing heterogeneities. The underlying philosophy of this work is that the evolution of hyperextended rift systems reflects the interplay between their inheritance (innate/"genetic code") and the physical processes at play (acquired/external factors). Thus, by observing the architecture and evolution of hyperextended rift systems and integrating the physical processes, one my get hints on what may have been the original inheritance of a system.

We first define 3 types of inheritance, namely structural, compositional and thermal inheritance and develop a simple and robust terminology able to describe and link observations made at different scales using geological, geophysical and modelling approaches. To this, we add definition of rift-induced processes, i.e. processes leading to compositional changes during rifting (e.g. seprentinization or decompression melting). Using this approach, we focus on 3 well-studied rift systems that are the Alpine Tethys, Pyrenean-Bay of Biscay and Iberia-Newfoundland rift systems. However, as all these examples are magma-poor, hyperextended rift systems that evolved over Variscan lithosphere the concepts developed in this paper cannot be applied universally. For the studied examples we can show that:

1) the inherited structures did not significantly control the location of breakup and the structures in the southern North Atlantic example
2) the inherited thermal state may control the mode and architecture of rift systems, in particular the architecture of the necking zone
3) the architecture of the necking zone may be influenced by the distribution and importance of ductile layers during decoupled deformation and is consequently controlled by the thermal structure and/or the inherited composition of the curst
4) conversely, the deformation in hyperextended domains is strongly controlled by weak hydrated minerals (e.g. clay, serpentinite) that result form the breakdown of feldspar and olivine due to fluid and reaction assisted deformation
5) inherited structures, in particular weaknesses, are important in controlling strain localization on a local scale and during early stages of rifting
6) inherited mantle composition and rift-related mantle processes may control the rheology of the mantle, the magmatic budget, the thermal structure and the localization of final rifting

These key observations show that both inheritance and rift-induced processes played a

significant role in the development of the magma-poor southern North Atlantic and Alpine Tethys rift systems and that the role of inheritance may change as the physical conditions vary during the evolving rifting and as rift-induced processes (serpentinization; magma) become more important. Thus, it is not only important to determine the "genetic code" of a rift system, but also to understand how it interacts and evolves during rifting. Understand how far these new ideas and concepts derived from the well-studied hyperextended rift systems of the southern North Atlantic and Alpine Tethys can be translated to other less explored hyperextended rift systems will be one of the challenges of the future research in rifted margins.

1. Introduction

Although there is general agreement in the community that inheritance has an important influence on the architecture and tectonic evolution of rifted margins, it is not yet clear how inheritance controls the initiation and evolution of rifting and lithospheric breakup. Most studies advocating for a strong role of inheritance refer to the observed along strike changes of the rift architecture and segmentation of rifted margins that seem to correlate with changes in the onshore geology (Dunbar and Sawyer, 1989; Pique and Laville 1996; Ring 1994; Tommasi and Vauchez, 2001). While Taylor et al. (2009) and Gerya (2012) showed that major oceanic transform faults do not extend into the continents, Behn and Lin (2000) discussed the importance of the segmentation observed along the U.S. East coast passive margin for the incipient structure of the oceanic lithosphere. Thus, at present it is not clear what is the importance of crustal inheritance in controlling the segmentation in oceanic domains. Other studies discussed the reactivation of pre-existing faults, or localization of deformation along major lithological boundaries (Butler et al. 2006; (Bellahsen et al., 2012). These observations are most of the time local and are unable to explain why early stages of rifting are often diffuse and do in most cases not coincide with the location of final breakup (e.g. Dunbar and Sawyer 1989; Bertotti et al. 1993; Manatschal and Bernoulli 1998). In the North Atlantic (Lundin and Doré, 2011) and the NW Shelf of Australia (Direen et al. 2008), the final breakup cuts not only through ancient orogenic belts, but also through earlier rift systems, suggesting that inheritance does neither control the location of breakup, nor the onset of seafloor spreading. Unfortunately, thick sedimentary and/or magmatic sequences prevent detailed investigation of the pre-rift crustal and lithospheric structures and composition. Thus, although the common belief is that inheritance has an important control on the evolution of rift systems, at present it is difficult to assess under what conditions and how it controls rifting and lithospheric breakup.

The aim of this paper is to define types of inheritance and to show, using observations and models, how these different types of inheritance may control subsequent rifting events. We use examples from the ancient and present-day Alpine Tethys, Pyrenean-Bay of Biscay and Iberia-Newfoundland rifted margins to illustrate and discuss ideas and concepts on how to integrate the role of inheritance in the study of rifted margins. Since all rift systems discussed in this paper affect Variscan lithosphere and are magma-poor, the ideas developed in this paper may not be applied globally. Thus, the aim is not to propose models that can explain in a global way the influence of inheritance on rift systems, but to discuss how "inheritance" can be described and integrated in descriptions of hyperextended rift models leading to seafloor spreading.

Due to the very limited data from hyperextended rifted margins available in the open domain and the difficulty to link their yet little understood evolution to inherited

orogenic structures, we will focus our paper on the well explored Variscan system in Western Europe and its relation to the late rift evolution of the Alpine Tethys and southern North Atlantic. Due to the novelty of the subject, this paper does not review but does rather present some considerations on how inheritance may control hyperextended rift systems based on observations and simple numerical models.

2. Defining inheritance in rift systems and rifted margins

Although the terms "inheritance" and "hyperextended rifted margins" are often used, clear and universal definitions are necessary in order to discuss the impact of inheritance on hyperextended rift systems. The aim of this chapter is to develop a terminology that is simple, robust and that can describe and link observations made at different scales using geological, geophysical and modelling approaches. The underlying philosophy is that each rift system inherited a set of particular properties and boundary conditions similar to a "genetic code", and that the evolution of rift systems reflects the interplay between their inheritance (innate) and the physical processes at play (acquired) during rifting.

2.1 Inheritance

In contrast to oceanic domains that are newly formed and do, in theory, not bear inheritance, continental domains are the result of polyphase tectonic, magmatic and metamorphic events, often resulting from the superposition of several Wilson cycles. Continental domains preserve a long history with a complex lithological and structural record that cannot be described by few physical parameters only. Therefore, defining the original state of the lithosphere at the onset of a new tectonic phase, in this case rifting, is difficult. In this study, we use the term "inheritance" to refer to the difference between a "real" and an "idealized" lithosphere (Fig. 1). The "idealized" lithosphere (Fig. 1a) is defined as a thermally equilibrated horizontally homogeneous layer-cake lithosphere made of a quartzo-feldspatic upper and middle crust and a mafic lower crust and a peridotitic mantle (for definition of single layers see also (Sutra et al., 2013). In contrast to this "idealized lithosphere", a "real" lithosphere (Fig. 1b) may differ by its thermal state, compositional heterogeneities and structural complexity.

We distinguish between three types of inheritance, namely "thermal", "structural" and "compositional" inheritance. Although the three can, practically, not be disassociated, for simplicity we will discuss and exemplify the three independently. Figure 1b shows, in a conceptualized manner, the 3 types of inheritance in a section across a lithosphere at the end of an orogenic collapse but before onset of rifting. In this paper we only show examples of rift systems that developed over Variscan post-orogenic, lithosphere. Other types of lithospheres, not discussed in this paper, are cratonic or complex accreted lithosphere, the latter formed by different terrains and separated by multiple suture zones (e.g. Western North America).

2.1.1 Thermal inheritance

Two components have to be considered: 1) inherited heat and 2) potential of heat production. Inherited heat decreases with time and is consequently a function of the age of the lithosphere. This implies that the older the lithosphere, the colder the thermal structure and the stronger the lithosphere. Thus, the time lag between the last thermal perturbation of the lithosphere and the onset of rifting is of major importance in assessing the thermal state of the lithosphere at the start of a rifting event. In contrast, the potential of heat production depends on the presence of radiogenic elements, which is mainly controlled by the composition and age of the crustal rocks and magmatic additions.

2.1.2 Compositional inheritance

Compositional inheritance can be defined as the difference between the real composition of a lithosphere relative to that of an idealized lithosphere made of layers with an homogeneous rheology (layer cake). Compositional variations may occur in the mantle as well as in the crust as a result of sedimentary, tectonic, magmatic, metamorphic and metasomatic processes during previous orogenic and extensional cycles. These compositional variations may be difficult to determine using seismic imaging methods alone and are often only determined by direct sampling of the lithosphere through xenoliths studies or sampling of remnants of former rift systems in orogens. As shown in Fig. 1b, in a post-orogenic crust, major compositional inheritances are remnants of flexural foreland basins and older post-orogenic intercontinental rift basins; ophiolite belts commonly associated with deep marine, clay/fluid rich thick sedimentary sequences, arc systems rich in new juvenile magmatic crust, and later orogenic underplated gabbroic bodies. In the mantle lithosphere, depletion and fertilisation of the lithospheric mantle are mainly related to subduction processes, slab breakoff, mantle plumes and post-orogenic extension. For instance, the mantle within a supra-subduction setting may be locally depleted or hydrated, while mantle rocks belonging to the down-going plate may not be significantly modified. The occurrence of underplated mafic bodies, often emplaced during slab breakoff or orogenic collapse can also result in the depletion of the underlying mantle. For cratonic lithosphere, long-term processes related to metasomatism and fertilization associated with magma-percolation may modify the composition of the lithosphere through time.

2.1.3 Structural inheritance

The existence of structures inherited from previous orogenic events in the crust and the underlying lithospheric mantle result in mechanical weak zones. These "weak" structures may correspond to ancient thrust belts or remnants of subduction zones. When these structures formed in the brittle layers, they are often linked to hydration processes. When they occur in the ductile layers, mainly as mylonitic shear zones characterized by grain size reduction and crystallographic preferred orientations, they can result in a strong anisotropy (Tommasi and Vauchez, 2001; Ring 1994). Of particular importance are intra-continental strike-slip systems, because these structures can affect the whole lithosphere (Tommasi and Vauchez, 2001), conversely to extensional or compressional structures, which are typically limited to the brittle lithosphere and become more diffuse in the ductile layers. It is also important to note that not every inherited structure has the potential to be reactivated. Reactivation depends on its orientation relative to the stress field, on it's lateral continuity and on the rheological contrast of the encompassing rocks relative to the other lithospheric layers (e.g. Chenin and Beaumont, 2013).

2.2 Hyperextended, magma-poor and polyphase rift systems

The understanding of rift systems and rifted margins underwent a paradigm shift in the last two decades. This is mainly due to the development of new reflection and refraction imaging techniques at present-day deep water hyperextended rifted margins, deep-sea drilling along the Iberia/Newfoundland and East Greenland/Mid-Norwegian conjugate margins and studies of remnants of rift systems in collisional orogens (Peron-Pinvidic et al., 2013a). These studies resulted in new observations and interpretations of how rift systems evolve in time and space. This process enables us to describe the evolution of rifted margins and to classify them according to the following characteristics:

1) the volume and timing of magma formation during rifting (e.g. magma-poor vs. magma-rich); 2) single vs. polyphase evolution (i.e. superposition of several rifting events), and 3) the distribution and partitioning of deformation along strike (e.g. segmentation associated with shear margins vs. orthogonal rifted margins).

2.2.1 Magma poor vs. magma-rich rift systems

Drilling of rifted margins by the Ocean Drilling Project (ODP) offshore Greenland and Norway (ODP Legs 104 (Endholm et al. 1989), 152 (Larsen and Saunders 1998) and 163) Larsen et al. 1999), ODP Legs and Iberia and Newfoundland (ODP Leg 103 (Boillot and Winterer, 1988), ODP Leg 149 (Whitmarsh and Sawyer, 1996), ODP Leg 173 (Whitmarsh and Wallace, 2001) and ODP Leg 210 (Tucholke and Sibuet, 2007) enabled to distinguish between magma-rich and magma-poor rifted margins. While earlier models suggested that these margins are end-member margins that evolved in a completely different way, more recent studies suggest that they share many similarities and that both evolve in a more similar way during early stages of rifting (Péron-Pinvidic et al. 2013). Two processes seem to be common to all rifted margins: 1) tectonic thinning of the crust/lithosphere and 2) lithospheric breakup induced magmatically/thermally (Fig. 2). Many authors (Franke, 2013; Lundin and Doré, 2011; Peron-Pinvidic et al., 2013) (Keen et al. 2012) suggest that magma-rich margins go through a phase of hyperextension pre-dating the magmatic breakup (Fig. 2a). Thus, the major difference between magma-rich and magma-poor rifted margins relates to their capacity to form large volumes of magma over relatively short time frames at the time of breakup. Indeed, while at magma-poor margins the time between onset of magma and lithospheric breakup can last over several tens of millions of years (about 30Ma in the case of the Iberia margin), at magma rich margins the timing between onset of magma formation and breakup is very short (< 5myr); see Fig. 2). In contrast to magma-rich margins, magma-poor margins offer two major advantages in studying magmatic systems associated to rifting: 1) the final rift structures are not masked by massive magmatic flows, and 2) exhumation of mantle rocks offers the possibility to investigate the early stages of magma infiltration and its reaction with deeper lithospheric levels.

In this paper, we focus on the examples of the magma-poor, hyperextended rift systems of the southern North Atlantic and the Alpine Tethys since these are the only ones whose hyperextended part benefits from either drill hole and high quality reflection and refraction data or direct outcrop observation.

2.2.2 Polyphase rifting

Although the classical pure shear (McKenzie, 1978), simple shear (Wernicke, 1985) or the combination of the two (Lister et al., 1986) models are still largely used to explain the evolution of rifted margins, the increased access to data and the development of new dynamic models enable more accurate and observation-constrained scenarios for the evolution of rifted margins. For instance, the dynamic model proposed by Lavier and Manatschal (2006) builds on a large body of observations derived from the Alps and Iberia. This model describes the evolution of rifting as a sequence of events referred to as stretching, thinning, exhumation and seafloor spreading (see also (Péron-Pinvidic and Manatschal, 2009) for a definition of these terms). In a more recent paper, (Sutra et al., 2013) showed that mechanical coupling/decoupling of the crust and mantle during rifting is reflected in the transition between thinning and exhumation. In this paper we mainly use the terminology of these authors to describe the first-order evolution of rifting. All these models have in common that rifting is, at first order, a strain localization process, i.e. initiates over a wide area, localizes subsequently during necking and remains localized at

the mid-ocean ridge from breakup and onset of steady-state magmatic activity. However, more and more evidence show that many rifts initially fail, migrate or step outside the initial rift system before breakup occur resulting in so called “offset basins” (Chenin and Beaumont, 2013). Here we call this phenomenon "polyphase rifting". Well documented examples of polyphase rift include the NW Shelf offshore Australia (Direen et al. 2008), the North Atlantic (Lundin and Doré, 2011), the northern segment of the South Atlantic (Blaich et al., 2008) or the Southern-North Atlantic (Péron-Pinvidic and Manatschal, 2010). Why rift systems may fail or migrate and how far the localization/delocalization of rift systems may be controlled by inheritance will be part of the discussion of this paper.

2.2.3 Along strike variations, segmentation and 3D evolution of rift systems

Most studies and models of rifted margins are focused on a 2D approach looking at dip sections parallel to the kinematic transport direction. It is important to note that these sections are not necessarily perpendicular to the margin, in particular in oblique or shear margins. The along strike architecture of rifted margins characterized by segmentation and strong lateral changes of crustal structures and/or the volume of magmatic additions is not well understood. The hypothesis that inheritance and/or rift-induced processes may control the lateral evolution and segmentation of rift systems is discussed further below, based on the example of the southern North Atlantic.

3. The role of Variscan inheritance in structuring the Alpine Tethys and Iberian rift-systems

Here we use the term “Variscan inheritance” to refer not only to the orogenic processes but, also to the late to post-orogenic events, distinguishing between crustal and mantle inheritance. We focus on the Alpine Tethys and Iberian Atlantic rift systems that formed over a lithosphere with a strong Variscan inheritance. Although the fate of these two conjugate margins are different, i.e. one became part of a big ocean while the other probably never developed into a steady state ocean, we consider the two systems as belonging to the same Late Triassic to Early Cretaceous rift system that formed over Variscan orogenic lithosphere. Until now, most discussions of the importance of inheritance focused either on local or large-scale examples, without considering that the resolution of the observations is strongly scale-dependent. Here we try to integrate this idea by discussing the role of structural and compositional inheritance at different scales, ranging from lithospheric- to outcrop-scale.

3.1 Influence of inheritance on a large-scale: the example of “Variscan” inheritance

3.1.1 Variscan processes controlling crustal inheritance

3.1.1.1 Pre- and syn-Variscan structures

Akin to modern orogens, the Varsican orogen overprinted a continental crust/lithosphere inherited from older orogenic events. The limit between parts of the crust/lithosphere that were affected relative to those that remained unaffected by a severe Variscan overprint corresponds to the transition between the fold and thrust belt and the flexural foreland basin (Fig 3a and b). While this limit is well preserved across Northern Germany and Southeastern England and Ireland, its southern limit is strongly overprinted by the Mesozoic rifting and the Alpine orogeny (see Fig. 3). The main difference between the distinct orogenic domains shown in Fig. 3a are, apart from their compositional and structural inheritance, the age of the last tectonic/thermal event.

Capturing the details of the Variscan orogeny in Western Europe is at this stage difficult, since the overall evolution and architecture is still a matter of debate (e.g., Martinez Catalan 2011). Indeed, remnants of the Variscan orogeny are dissected and overprinted by later rifting and the Alpine orogenic cycle, which makes correlations between the different parts of the former Variscan orogen difficult. The better-preserved and continuous domain of the Variscan belt can be mapped through Eastern and North Western Europe corresponds to the northern Variscan foreland and the associated fold and thrust belt. More debated is the extent of the Rheic suture zone and that of arcs and back-arc systems in Western Europe. This is mainly due to the poor preservation of this system and the likely subduction of most material of the former internal parts, including the remnants of the Rheic ocean and of the former arc and back-arc systems. Debated is also the southern limit of the Variscan orogeny and the location of its associated foreland. The occurrence of un-deformed Ordovician and Devonian sediments in the Southern Alps suggests that the southern front coincides with the domain now occupied by the present-day Alps (Raumer et al. 2012). However, possible late orogenic large-scale transtensional or strike-slip systems may have dissected and displaced parts of the former orogen, resulting in present-day juxtaposition of initially side by side terranes in a late stage of the orogenic evolution (Raumer et al. 2012; Muttoni 2003). Another debate concerns the relationships between the Amorican and Iberian branches of the Variscan system. The most popular interpretation is that the Variscan orogen formed an orocline (Pastor-Galan et al. 2012), which is supported by the lack of Variscan overprint on the Newfoundland margin. The drilling of Culm-type sediments, characteristic for the Variscan foreland, at ODP Site 1069 along the deep Iberia margin supports this interpretation (for location see Fig. 3a; (Wilson et al., 2001). Although details of the correlations and extent of the Variscan range are not yet well constrained, at first order, the major paleogeographic domains, including foreland basins, fold and thrust belts and internal parts (suture zone) can be mapped on the scale of Western Europe. However, the lithospheric structure, including the upper mantle and lower crust, have been strongly overprinted and modified during the late to post-orogenic evolution of the orogeny as shown in the following sections.

3.1.1.2 Late- and post-Variscan structures

Within the Variscan domain, the lithospheric and crustal structures are the result of a complex evolution that ended with major late- to post-Variscan extensional and magmatic events. Although the exact timing and processes of this evolution are not yet well constrained on the scale of the whole orogen, it looks as if at about 310 to 300 Ma the magmatic system changed from an arc-dominated to an intracontinental magmatic setting. This change is associated with the emplacement of prominent underplated mafic bodies and extrusive magmatic systems simultaneous with crustal thinning (Fig. 3c). Evidence for magmatic underplating is widespread and can be observed in all exposed sections preserving pre-rift lower crust (e.g. Ivrea zone in the Southern Alps (Peressini et al. 2007), the Malenco-Campo-Grossina units in the Eastern Alps (Hansmann et al. 2001; Tribuzio et al. 1999), the Santa Lucia complex (Rossi et al. 2006) and Calabria (Caggianelli et al. 2012) as well as in xhenolites from the Variscan belt (e.g. xhenolites from the French massif (Féménias et al. 2003)).

This event, also referred to as the "Permian event", was concomitant with a change in the morpho-tectonic evolution of the former orogenic system. It is expressed as a shift from a classical mountain belt topography to one dominated by plateaus dissected by intra-continental sedimentary basins (Burg et al. 1994; Froitzheim et al. 2008). This evolution can be compared with the observed orogenic collapse in the Basin and Range

Province of the western US. The Permian Variscan collapse is not only recorded by the important magmatic additions to the crust (intrusions and underplating), but also by granulite facies metamorphism (Galli et al. 2011, see also Mohn et al. 2010 for a review), which indicates a very high thermal gradient (Müntener et al. 2000, Schuster and Stüwe 2008). These authors described the thermal evolution during the late- to post–orogenic event and determined the thermal state of the lithosphere at the end of Permian time, i.e. before onset of rifting. These studies showed that the magmatic and thermal evolution recorded in the continental crust is intimately related to processes acting in the underlying mantle lithosphere (Fig. 3c). As discussed in the following, these processes strongly controlled the final structure of the lithosphere at the end of the Variscan cycle and pre-determined the conditions at the onset of the Mesozoic rifting.

3.1.2 Variscan processes controlling mantle inheritance

In most rift systems, the composition and evolution of the sub-continental mantle involved in rifting is unknown and very difficult to either image or sample. Exceptions are the Iberia-Newfoundland and the ancient rift systems/margins exposed in the Pyrenees and the Alps. In these examples the mantle underneath the continent during rifting was exhumed at the seafloor at the end of rifting. This exhumed mantle was drilled in the Iberia margin and can directly be sampled in the Alps and Pyrenees. This offers the unique opportunity to investigate the nature of subcontinental mantle and its evolution during rifting and to explore the importance of inherited compositional variations.

3.1.2.1 Pre- to syn-Variscan inheritance

Mantle rocks drilled along the conjugate Iberia-Newfoundland margins (e.g. ODP Sites 637 (Galicia Margin), 987, 899, 1068, 1070 (Iberia Abyssal Plain) and 1277 (Newfoundland) (see Fig. 4b) show distinct compositional heterogeneities (Fig. 4a; (Müntener and Manatschal, 2006) and references therein). While the mantle rocks from the Newfoundland margin (ODP Site 1277) are made of depleted harzburgites (high rations of Cr# in clinopyroxene vs. spinel), those from the Iberia margin are less depleted and consist of infiltrated plagioclase peridotites (Müntener and Manatschal, 2006). The authors interpreted the strongly depleted harzburgites drilled at ODP Site 1277 at the Newfoundland margin as derived from the hanging wall of a Caledonian subduction (Fig. 4b). Thus, these mantle rocks sampled the strongly depleted supra-subduction mantle that may have produced the magma of the Bay of Island arc complex now exposed in Newfoundland. This interpretation assumes that mantle rocks from the Ocean Continent Transition (OCT) are derived from the pre-rift sub-continental mantle proving the existence of a strong mantle inheritance. The observation that the mantle rocks from the Iberia-Newfoundland margins show different degrees of depletion is a key observation, because mantle composition is a key parameter controlling the magmatic volume that can be produced during rifting.

3.1.2.2 Post-Variscan to early rift inheritance

In the Alps and Pyrenees, remnants of mantle rocks are common and associated with lower to middle crustal granulites and deep marine sediments (Jammes, 2009; Lagabrielle, 2009; Manatschal and Müntener, 2009) (Fig. 4d). Studies of these mantle rocks in the Pyrenees (Lherz, LeRoux et al. 2007), Liguria and Corsica (Piccardo, 2003; Rampone and Hofmann, 2012) (Montanini et al. 2012, and the Alps (Lanzo, Malenco, Totalp; (Müntener et al., 2010) Piccardo et al. 2009,) showed that the mantle rocks exhibited different degrees of modifications prior and during rifting. The uppermost subcontinental mantle that locally still preserves primary contacts to the overlying crust

(e.g. Manlenco, Hermann et al. 1997) was too cold to either melt or react with magma during the Permian orogenic collapse and subsequent rifting. This type of mantle, referred to as inherited mantle, is exposed in the most proximal parts of the OCT where it is often associated with remnants of the thinned continental crust (e.g. Malenco or Tasna; Müntener et al. 2004, 2010) (Fig. 4d). In contrast, the mantle rocks occurring in the more oceanward part of the OCT show a more complex evolution (Fig. 4d). Müntener et al. (2004) showed that these mantle rocks record two major events: 1) a strong depletion during the late- to post-orogenic event; and 2) an enrichment of the mantle during early stages of rifting. The first event is dated as Permian by Sm/Nd model ages, interpreted to date the melting of this mantle under asthenospheric conditions (Rampone and Hofmann, 2013; Müntener et al. 2004, 2010, and Piccardo et al. (2009). This event may be linked with the emplacement of the gabbros shown in Fig. 3c at the base of the crust during Permian time. The second event is associated with the infiltration and reaction of magma with the previously depleted mantle. Thus, this second event has to be post-Permian but pre-exhumation. These observations show that the subcontinental mantle records a complex, polyphase evolution including depletion and fertilisation cycles. Thus, the inherited subcontinental mantle underlying the domains that underwent rifting experienced a strong late- to post-Variscan imprint. How this may have controlled the subsequent rifting is discussed in the following section.

3.1.3 Control of Variscan inheritance on the large-scale evolution of the Tethyan and Iberian rift systems

The extent of late Variscan overprint in the lower crust and mantle compared to pre- and syn-orogenic compressional structures begs to the question of their relative influence on the large-scale evolution of the subsequent rift system. In the following, we discuss the role of the Variscan inheritance, in particular of lower crustal and mantle compositional inheritance in controlling the large-scale strain distribution and magmatic evolution during the subsequent rift event.

An interesting observation is that the location of hyperextension and later breakup of the Alpine Tethys and Iberian Atlantic localized along the former boundaries of the Variscan orogeny, leaving the central parts - thus all the previous suture zones - of the orogenic lithosphere preserved (e.g. Bohemian massif, Massif Central and Armorican belt and in the Iberian peninsula; Fig. 3a). This observation questions if the location of continental necking and lithospheric breakup is pre-determined by orogenic or post-orogenic processes.

How far gravitational processes, i.e slab breakoff and/or delamination of the orogenic roots of an over-thickened orogenic crust are responsible for the subsequent rift evolution compared to late- to post-Variscan inheritance is yet unclear. The observation that in Western Europe crustal necking and breakup affected the boundaries between internal and external parts of the former orogeny may suggest an “orogenic” control. One factor that may control the location of necking is the late- to post-Variscan mafic underplating, associated with the depletion of the underlying lithospheric mantle (Fig. 4c; further discussion in section 5). Indeed, this process seems to be widespread but localized underneath internal parts of the Variscan orogen (Fig. 3b). Thus, as shown by the observations reported above, the Permian magmatic event strongly modified the crustal and mantle composition underneath the realm affected by the Variscan orogen. This event is well documented in all the geological sections that expose remnants of the Late Variscan lower crust and mantle in the former distal portions of the Tethyan margin. It is also drilled along the most distal Iberia margin (ODP Site 1967 in Fig. 4b), found in

xenolithes all over Western Europe (Féménias et al. 2003) and in remnants exhuming lower crustal rocks in the Alps (e.g. Malenco; Herrmann et al. 1997). This suggests that the last thermal event that may have modified in a sever way the lithosphere and modified the composition of the lithospheric mantle and the lower crust is not related to the Variscan collision, but the late- to post-orogenic extensional collapse of this orogen. In the case of the Variscan orogeny, the post-orogenic collapse resulted in a stronger lithosphere underneath previously overthickened crust comparing to the more external and thinner crust in the former fold and thrust belt that was not modified (Fig. 3b).

A determinant observation for the magmatic evolution of the rifting event is that the lithospheric mantle underneath Western Europe exhibits different degrees of modifications prior and during rifting (Fig. 4a). The existence of depleted and enriched mantle domains underneath crustal domains undergoing extension is important, since it may explain differences in the magma production during rifting. Models assuming a homogeneous subcontinental mantle explain the lack of syn-rift magma as either related to very slow extension rates or a very cold mantle lithosphere or a combination of the two. The depleted mantle domains that formed during the late to post-orogenic Permian event may have lacked the necessary mineral components to produce melts during the subsequent extensional event. It may explain why the rift system remained magma-poor. Müntener et al. (2004) suggested, using the examples of the Alps that magmas first reacted with the previously depleted lithospheric mantle before they were extracted (Müntener et al.. 2009). Assuming that these observations are generally valid, it would mean that the subcontinental mantle was depleted during orogenic collapse and was unable to produce a lot of magma during the following rifting. The first magma that was formed by decompression reacted with the overlying depleted lithospheric mantle leading to its fertilisation without any surface expression. This suggests that magma-poor rifted margins may develop in the absence of plumes over a lithosphere that was depleted during orogenic collapse. In contrast, magma-rich margins may occur over fertile mantle lithosphere, either because it has never been depleted or because it was fertilized by a plume or during previous rift-events. How important and how general these processes are, is difficult to determine. However, it shows that linking ß-factors to volume of produced magma is an oversimplification and may not apply to domains where the subcontinental mantle is heterogeneous.

3. 2 Control of orogenic inheritance on a regional-scale: crustal necking and rifted margin segmentation

3.2.1 Control of inheritance vs. rift-induced processes on the morphology of necking profiles

The Western Iberia margin is often considered as the archetype magma-poor hyperextended rifted margin (Péron-Pinvidic et al. 2013). The numerous studies performed along this rifted margin (e.g. (Tucholke et al., 2007), Péron-Pinvidic and Manatschal, 2009) benefited from the dense covering of refraction and reflection seismic and drill hole data. In particular the discovery of hyperextended continental crust begged the questions of how the crust thins and what controls extreme crustal thinning? In order to answer to these questions, Sutra and Manatschal (2012) defined and described necking zones using reflection and refraction seismic lines from the Iberia rifted margin. These necking zones correspond to areas where the crust is thinned from its initial thickness to a crustal thickness of < 10km.

Observations

The architecture of the necking zone along the Western Iberia margin is shown in three sections in Fig. 5a). The northern section corresponds to the ISE1 line (Iberia Seismic Experiment; (Zelt et al., 2003), the central section to the composed LG12 - TGS line (Sutra and Manatschal, 2012) and the southern section to the IAM 9 line (Dean et al., 2000) (for location along the Iberia margin see Fig 3b). All sections are dip lines, i.e. parallel to the kinematic transport direction. The necking domain has been defined as the domain limiting original thick crust (30km) from crust thinner than 10km. The necking zone is defined by an oceanward increase in the total ß value, space accommodation and the coupling of the deformation between crust and mantle (for a more detailed definition of the "necking domain" see Sutra et al., 2013). In the northern section (ISE 1) the necking profile is wide and defined by a double neck. In the central section (LG12-TGS) the necking profile shows a single neck of intermediate width. In the southern section (IAM 9) the necking is sharp. In contrast to the change of the necking profile along strike of the margin, it is interesting to note that the width as well as the angle of aperture (angle between top and base of the basement, the later corresponding to a "S" type reflection; e.g. Sutra et al. 2013) of the coupled domain remains similar along the margin (Fig. 5a).

Implications

Thinning of the crust/lithosphere is one of the most fundamental processes related to rifting. Extreme crustal thinning is typically localized within the so-called necking domain, which can display strong lateral changes, as shown for the Iberian margin (Sutra and Manatschal, 2012). The necking profile at rifted margins can be controlled by the inherited bulk rheology of the crust, in particular by the relative distribution of ductile vs. brittle layers (Mohn et al., 2012; see also discussion of modelling results in Section 4). The observations show that the more ductile the crust, the more difficult it is to neck. Conversely, strong and more rigid crust tends to produce sharper necks. Other controlling parameters are the obliquity of extensional systems and the strain rate (Jammes et al., 2010). For the example, the Iberian margin formed by orthogonal extension more than 100 myr after the last thermal imprint, suggesting that the lithosphere was thermally equilibrated, and under similar strain rates along its whole length (see Sutra et al., 2013). As the necking profiles are quite different along the margin, one can assume that their architecture is primarily controlled by the composition of the crust and mantle. If we assume a homogeneous mantle composition, the change in the necking profile observed along the Iberia margin may provide some information about the inherited crustal composition and indirectly about the bulk rheology of the crust at the onset of rifting. As shown in Fig. 5b, the 3 sections cross different parts of the Variscan belt, which are likely to show different crustal compositions (i.e. different percentages of ductile material and/or inherited structures able to localize deformation). In the North (Fig. 5b), where the necking profile is a wide double neck, the crustal structure is likely to be complex, probably bearing several brittle-ductile transitions and/or strong lateral variations of the crustal structure. In the South, where the necking profile is narrower, the crustal structure is likely to be simpler, consisting of more brittle material and/or controlled by an important inherited crustal scale structure (Alves, 2011).

Another important observation is that all sections shown in Fig. 5a display similar width and angle of aperture of the coupled domain along strike (Fig. 5a). The map in Fig. 5c shows that the width of the coupled domain, which lies between the point of exhumation and the coupling point (see Sutra et al., 2013 for details), does not change along the margin, conversely to the width of the necking domain. Deformation in the coupled domain only occurs via brittle processes, while it is primarily controlled by ductile layers in the decoupled domain.

Based on the drilling results, the refraction velocities and in analogy with material that can be found in comparable positions in the Alps, we assume that the crustal rocks are strongly hydrated and that the bulk rheology of the hyperextended crust is mainly controlled by the newly created phylosilicates such as illite, chlorite and serpentinite. This would explain that, when the crust is thinned to less than 10 km and becomes entirely brittle, the initial compositional or structural differences do not impact deformation any more and all crusts, independent of the initial composition, start to deform in a similar way. Thus, while thinning and formation of the necking domain may be controlled, on a crustal scale, by inheritance, deformation in the hyperextended and exhumation domains is most likely not controlled by inheritance but by hydration processes, i.e. by a rift-induced process. This applies, however, only to the uppermost parts of the hyperextended crust and exhumed mantle (< 6km deep). On a lithospheric scale, the thinning of the lithosphere and onset of magma may eventually control the localization of deformation and magma-production controlling final breakup (see discussion section for more discussion).

3.2.2 Control of inheritance on the segmentation of rifted margins

To what extend oceanic structures in general and transform faults in particular are inherited from the previous rift system is highly debated (Bonatti et al., 1994). Taylor et al. (2009) showed that oceanic transform faults do not penetrate into the margin, and (Gerya, 2012, 2013) showed that prominent structures observed on the continent do not necessary continue into oceanic crust. However, at most present-day rifted margins the initial stages associated with the development of highly segmented rift-transform margins are often masked by thick sedimentary sequences and the relation between the rift structures, syn-tectonic sediments and magmatic additions remain poorly constrained. A good place to understand the role of inheritance on the segmentation of margins and hyperextended domains, as well as the location of breakup and structures within the oceanic crust is the Iberia Atlantic and Bay of Biscay – Pyrenean domain (Fig. 6). Welford et al. (2012) and Tugend et al. (2014) showed that in these domains deformation was distributed over tens of millions of years before it resulted in the opening of a spreading ocean that propagated into a hyperextended and strongly segmented rift-transform system during Late Jurassic to Mid Cretaceous time.

Observation

Alves (2011) showed that major lineaments observed and mapped onshore Iberia can be observed and followed on potential field maps into the Zone of Exhumed Continental Mantle (ZECM). This observation supports not only the idea that the mantle is of subcontinental origin in the ZECM (see previous chapter), but also that these structures are rooted into the lithospheric mantle. However, these lineaments disappear further oceanwards as indicated by the observation that the magnetic anomaly 34 is continuous and not affected by transfer faults along the Iberia-Newfoundland margins (Fig. 6a). Thus, in this case, the structures observed in the continent do not control the segmentation in the adjacent oceanic crust. Furthermore, none of the major transfer structures along the southern North Atlantic, with the exception of the Torre Madeira – Nova Scotia and Charlie Gibbs fracture zones seem to penetrate into the hyperextended domains. This also suggests that inherited continental structures are unlikely to control the structure of the adjacent oceanic crust.

Mapping of rift domains in the Bay of Biscay – Pyrenean system enabled the description of inherited pre- and syn-rift structures and their relationships to the first oceanic crust. Jammes, (2009) and Jammes et al. (2010b) showed that a network of

inherited structures such as the Toulouse and Pamplona faults localized and segmented hyperextended rift basins. According to potential field maps, both the Toulouse and Pamplona faults root deeply in the crust/lithosphere (Jammes et al. 2010b). A detailed map of the rift domains (Tugend et al. 14) (Fig. 6b) shows the complex relationships between the inherited Variscan and Late- to post-Variscan structures and the evolution of the rift domain in space and time. On the large-scale maps (Fig. 6a) as well as on a smaller-scale one inherited structures in the continent seem not to control the structures in the adjacent ocean. This is well shown by the observation that the limit between the hyperextended and oceanic domains remains unaffected by inherited structures mapped in the continent. However, Tugend et al. (2014) and Jammes et al. (2010b) showed that structuration of the hyperextended domain is strongly controlled by inheritance. Inherited structures can either transfer extension (e.g. Pamplona fault), or control the polarity of hyperextended margins, separating upper from lower plated margins (see Tugend et al. 2014).

Implications:

The observation that major structures observed in the continent can be prolonged into the zone of exhumed mantle (e.g. Iberia Abyssal Plain; Alves, 2011) supports the idea that the mantle underneath the OCT has a subcontinental origin. However, in the case of the Iberian margin, these structures do not control the segmentation in the adjacent oceanic crust. Furthermore, the major transfer structures in the oceanic crust along the southern North Atlantic do not penetrate into the continent, suggesting that inherited continental structures do not control the structure of the adjacent oceanic crust. The observation that the first oceanic crust truncates earlier rift structures in the southern North Atlantic (Fig. 6) suggests that breakup and seafloor spreading are not inflenced by inheritance but are the result of deep-seated asthenospheric processes. However, at present the process of breakup is not yet sufficiently constrained to understand the underlying controlling factors. The study of the Bay of Biscay – Pyrenean system shows that stretching, thinning and exhumation are strongly controlled by inherited structures, contrary to breakup and seafloor spreading. The latter may be controlled by heterogeneity within the lithospheric mantle underlying the hyperextended domains (e.g. section 3.1.2).

3.3 Control of inheritance on a *local-scale*

In this section we use two examples to discuss how inherited structures and compositional variations may control strain distribution on a local scale during hyperextension. The key learning that we want to convey is the importance of inherited weak layers controlling strain distribution and style of rifting. While the existence of pre-existing faults in the brittle layer may be a common feature on rifted margins, the pre-rift salt example can be used as an analogue for weaknesses such as weak layers in the crust or in the mantle able to localize deformation.

3.3.1 Relation between pre-rift structures and detachment faulting (Err detachment system, SE Switzerland)

General situation

The Err detachment system exposed in the Lower Austroalpine Err nappe in SE Switzerland is one of the world's few exposed and preserved rift-related detachment systems (e.g., Masini and Manatschal 2013). This structure can be mapped over about 200 km^2 exhibiting primary relationships with pre-rift Permian structures and Alpine thrust faults. The Err detachment system formed during the final stage of rifting by sequential

faulting and exhumation. The detailed mapping of this rift related detachment system (Froitzheim and Eberli, 1990; Manatschal and Nievergelt 1997; Masini et al. 2012) shows the importance of the pre-rift structures for the localization of this long offset low-angle extensional detachment structures.

Observations

The excellent exposure and preservation of the Err detachment system (Fig. 7a) permit the determination of the relationships between inherited late Variscan (Permian) structures and younger Alpine thrust faults (green and red structures in Fig. 7). Over wide areas the Err detachment system corresponds to two well-defined, top-to-the-west surfaces, referred to as the Err and the Jenatsch detachment faults (dark and light blue structures in Fig. 7). These detachment surfaces are either overlain by extensional allochthons, or, where the faults were directly exhumed at the seafloor, by syn-rift sediments (Fig. 7c; see also Masini et al. 2012 for details).

A first observation establishes the relation between the detachment system and a late Variscan batholith made mainly of granite. Throughout most of the observed area, the granites form the footwall of the detachment system, while they are nearly absent in the hanging wall. The primary intrusive contact between the host rocks and the granites, as well as the occurrence of host rocks within the granites, can be observed in many places in the footwall, suggesting that the detachment localized along, or near, the roof of the batholith. A second key observation is the relationship between the detachment system and a fault bounding a former Permian basin filled with a >500m thick volcano-sedimentary sequence of Permian age and sealed by lower Triassic shallow marine dolomites (Fig. 7b). As shown in Fig. 7c, the Jurassic detachment is overlain by the thick Permo-Triassic succession in the area of Piz Surgonda. Directly to the North, the lower Triassic dolomites are directly overlain, along a primary contact, by micaschists and orthogneisses that are in the hanging wall of the Jurassic detachment (Fig. 7c). This observation suggests that the Permian basin terminated abruptly north of the Surgonda area along an East-West striking steep, south-dipping fault forming the northern termination of the Permian basin (see Fig. 7d). Detailed mapping of the detachment surfaces shows that the northern termination of the Permian basin corresponds to the location where the Jurassic detachment fault system forms an East-West trending lateral ramp. Thus, the lateral ramp of the detachment fault reactivates an inherited surface corresponding to the northern termination of the Permian basin (see Fig. 7d). Mapping of the major top-to-the-west Alpine thrust fault shows that it reactivates the Jurassic detachment system in the south. In the north, where the Jurassic detachment fault forms a lateral ramp, the Alpine thrust truncates this structure. This observation suggests that the Alpine thrust was unable to reactivate the lateral ramp and incised into the footwall of the Jurassic detachment. This can explain the observation that the Jurassic detachment system was not reactivated during Alpine convergence and preserved its former relationship between footwall and hanging wall in the vicinity of the lateral ramp.

Implications

The mapping of the well preserved Permian, Jurassic and Alpine structures in the area of Piz Err – Piz Surgonda shows that the detachment faults localized either at existing lithological contacts such as the limit between intrusive granites and the surrounding host rocks, or that the detachment reactivated pre-existing structures. In the example of Err detachment system, the complex 3D structure of the detachment system is strongly controlled by the existence of a Permian basin as shown by the section that is perpendicular to transport direction in Fig. 7c. These observations show that reactivation

of inherited structures or lithological boundaries with a major rheological contrast are important and can control, on a local scale, the structures in hyperextended systems.

3.3.2 Relationships between pre-rift salt and detachment faulting (Mauléon and Parentis basins)

General situation

The Mauléon and Parentis basins in the Western Pyrenees – Bay of Biscay formed as hyperextended basins during Late Aptian to Albian time (Fig. 8a). Within these basins salt structures have been drilled and are seismically imaged (e.g. Ibis glacier in Parentis basin: Mathieu, (1986); Biteau et al. (2006)) and observed in the field (e.g. massif des Arbailles, Roquiague diapir in Arzacq-Mauléon basin; Canérot (1989)) (Fig. 8b). These basins offer therefore the opportunity to study the role of Triassic pre-rift evaporites during the subsequent Late Aptian to Albian hyperextension using seismic and drill hole data and field observations. Jammes et al. (2010a) and Lagabrielle et al. (2010) showed that the pre-rift salt played an important role controlling the final structure of the hyperextended basins. We consider the pre-rift salt as inherited compositional weaknesses, similar to weaknesses in the crust and/or mantle, with the only difference that evaporites are better exposed and their initial geometry better understood.

Observations

Within the Mauléon basin, the subsalt (basal Triassic Muschelkalk and Buntsandstein units) and the supra-salt units (Upper Triassic to Lower Cretaceous carbonates) are never juxtaposed within one continuous stratigraphic section. The sub-salt Lower Triassic sediments (Buntsandstein and Muschelkalk) are associated with the Paleozoic meta-sediments, while the supra-salt Upper Triassic to Lower Cretaceous carbonates form massive tilted blocks, interpreted as extensional allochthons that overlay Triassic evaporites and either continental or mantle rocks (e.g. Sarrance, Mail Arrouy blocks (Jammes et al., 2009). Thus, in the Mauléon basin, the Triassic evaporites served as a decoupling horizon separating sub-salt from supra-salt units. Although the present-day position of these blocks is the result of Pyrenean compression, the occurrence of growth structures of Early Aptian age (e.g. Urgonian carbonates in the Arbailles) indicates that these blocks tilted during Early Cretaceous rifting using the Triassic evaporites as decollement horizons. The observation of mantle rocks in salt diapirs within Upper Cretaceous sediments show that salt had to be in direct contact with the exhumed mantle already before Pyrenean convergence. An additional and independent argument is that a hydrothermal metamorphic event, associated with mantle exhumation, and dated at 107+/-3Ma, affected Triassic evaporites (Thiébaut et al., 1988, 1992). All these observations show that there is a close relationship between salt, detachment faults and mantle exhumation.

In the Parentis basin, diapiric structures occur mainly in the southern part (Mathieu, 1986; Biteau et al., 2006). Salt migration initiated in mid Cretaceous time and lasted until the end of Eocene time. The relation between the salt and the rift structures can be observed within a section running parallel to the coastline, referred to as the "Aquitaine coast section" (Fig.8c). In this section, Jammes et al. (2010a) described along the southern part of the section an extensional detachment structure that is related to salt migration. Apart from the unmistakable evidence for salt diapirs (supported by drilling), there are some hints for asymmetric blocks bounded by growth structures that are underlain by a

continuous, strong sub-salt reflector (Fig. 8c). The overall structure resembles that of extensional allochthons. These structures look also similar to those observed in the Arbailles, where tilted, fault-bounded blocks with Urgonian growth structures are associated with salt diapirs (Canérot, 1989; see previous paragraph). About 2 km eastward of the seismic section, two wells, SGM1 and CTS1, were drilled across the strong basal reflector and penetrated into basement. Well SGM1 penetrated beneath Tertiary and Upper Cretaceous sediments, 1737 m of Keuper salt series and 800 m of Triassic Muschelkalk dolomite and Buntsandstein sandstones, and tectonized quartzite dated as Carboniferous (Serrano et al., 2006). About 15 km to the north, located further basinward, well CTS1 penetrates Tertiary and Upper Cretaceous sediments, 1189 m of salt, 500 m of Lower Triassic sediments, and tectonized dolomites dated as Devonian (Serrano et al., 2006). The large thickness of the salt series penetrated by these two wells supports the presence of diapiric structures. The strong continuous reflector identified in the seismic data corresponds to the top of the Triassic Muschelkalk dolomite, which lies at the base of the Keuper salt. The continuity of this reflector under tilted blocks suggests that a decollement occurs in the overlying salt series, an interpretation similar to that proposed for the Mauléon area (see previous paragraph). The fact that the top of the basement is tectonized shows that this deformation is not only due to gravitational deformation. A major decollement located in the evaporitic Keuper is required to decouple deformation in the supra-salt units from deformation in the sub-salt units. While the supra-salt deformation was mainly accommodated by block tilting, the sub-salt deformation was accommodated along a top basement detachment fault drilled at wells SGM1 and CTS1 (Fig. 8c). We propose therefore that the tilting of the Upper Triassic to Lower Cretaceous carbonates, associated with the growth-structures (probably of Aptian age) occurred simultaneously with, but decoupled from major crustal thinning in the underlying continental crust (see model in Fig. 8d). The interaction of detachment faults and decollements, the latter within the Keuper evaporites, can explain, as shown by Jammes et al. (2010a), the unconformities observed and drilled in the southern Parentis and Mauléon basins as well as the extreme thinning of the crust and local mantle exhumation observed in these basins.

Implications:

The study of Triassic evaporite layers and their relationships to younger extensional detachment faults exhuming deeper crustal and mantle rocks in the Parentis and Mauléon basins enables us to suggest that the presence of a pre-rift salt layer in an area undergoing extreme crustal thinning can control the geometry and evolution of rift systems and obscure the rift related structures in the underlying basement. During an initial stage of rifting, pre-rift salt layers act as a decoupling horizon between sub- and supra-salt units suppressing the extension of the crustal detachment faults into layers above the salt and, by doing so, eliminating breakaways at the seafloor (Fig. 8d). As a consequence, the detachments sole out along the ductile salt layer and no sub-salt material can be exhumed to the seafloor. Thus, sub- and supra-salt layers deform by different deformation modes, which makes detachment faults difficult to identify on seismic images. In a later stage, when salt has migrated and thinned out, sub- and supra-salt layers can locally couple and detachment faults can daylight, resulting in the exhumation of basement surrounded by extensional allochthons formed by supra-salt sedimentary units. Thus, the pre-rift salt is a good illustration of the importance of inherited weak layers in decoupling and localizing the deformation on a local scale during rifting. It may be seen as an analogue for other potential weak layers in the curst or mantle, related to inherited compositional inheritances.

4. Control of thermal state on the evolution of rifting: a modelling approach

4.1 Modelling approach

In his pioneering work, (Buck, 1991) showed that the thermal structure of the lithosphere is controlling, together with the composition and crustal thickness, the evolution of rift systems. Depending on the thermal structure and crustal thickness, Buck (1991) distinguished between "narrow", "wide" or "core complex" extensional systems. Analogue models (Brun and Beslier, 1996) and numerical models (Huismans and Beaumont, 2007; Jammes et al., 2010) showed that, apart from the thermal structure and composition of the lithosphere, also the strain rate and obliquity may control the architecture of rift systems. The aim of this modeling study is to investigate the effect of the thermal structure on the dynamic behavior of the lithosphere during hyperextension. We consciously limit our models to very simple/basic to better investigate the very impact of different thermal structures. In contrast to compositional and structural inheritance, thermal inheritance is transient, therefore it would be difficult to investigate the effects of the thermal inheritance on hyperextension using fossil examples.

Following the work of Lavier and Manataschal (2006) we test the effects, on extensional styles and evolution, of variations in initial thermal structure in a rheologically layered lithosphere including a mafic (gabbroic) lower crust (Fig.9) assumed to be inherited from a phase of post orogenic collapse (see paragraph 3.1.1.2). The mantle is modelled as dry olivine and is assumed to be strong and depleted. We also include weakening of the crust by hydration where fault zones or ductile shear zone are active (for details see Fig. 9c) (Lavier and Manatschal, 2006). Finally, to limit the size of the parameter space we assume that the initial crustal thickness is dependent on the initial thermal structure. The numerical domain is initially 450 km wide and 100 km thick. (Fig. 9a). The crustal thickness is 35 km including a 7 km thick mafic layer at its base. Both numerical techniques and boundary conditions are identical to those used in Lavier and Manatschal (2006). To simulate rifting an extensional half rate of 2.5 mm/yr is applied on both side of the model (Fig. 9a). We vary the thermal structure assuming that the lithosphere cools rapidly after the phase of post-orogenic collapse. The lithospheric thickness is fixed to 100 km and we calculate an initial geotherm based on a fixed value of temperature at the Moho given by geological observation (i.e. Müntener et al. 2001). We define three cases: (1) the Moho temperature is set at 450°C for an old lithosphere (500 Myr) (Fig. 10), (2) 100 Myr later at 600°C (Fig. 11) and (3) immediately after post orogenic collapse at 750°C (Fig. 12). The first case could correspond to an ancient mobile belt and the intermediate case to the Tethyan rifting similar to what is observed in the Alps (for details see (Manatschal, 2004)). Even though these models are relatively simple when compared with the complex inheritances that could be ascribed (see previous sections), we obtain a wide range of tectonic behaviours and evolutions.

4.2 Modelling results

Figures 10, 11 and 12 show the results of the 3 different cases (1 to 3) with variable Moho temperatures. The figures show the evolution of the viscosity and strain rate over 30 Myr for each case (Fig. 10, 11 & 12). We distinguish 3 different styles of deformation: (see captions for details): (1) localized extension over a symmetric graben that roots in the mantle lithosphere that leads to rapid thinning and necking of the lithosphere until the mantle is exhumed at the surface (Fig. 10), (2) distributed extension followed by a phase of thinning during which necking of the lithosphere occurs for an intermediate lithosphere (Fig. 11), (3) highly distributed extension and near pure shear

deformation in case 1 with a hot/young lithosphere (Fig. 12). At first order the main control on the different styles of extension is whether brittle deformation in the upper crust is coupled to brittle deformation in the mafic lower crust and uppermost lithospheric mantle. In case 3 the hot ductile lower crust allows for decoupling to be maintained over very long time scales (30 myrs). In case 2 after being decoupled for ~ 13 Myr the middle to lower crust eventually are thin enough that brittle upper crustal deformation can couple to the deformation in the uppermost embrittled lithospheric mantle. At this moment in time the lithosphere thins over a major lithospheric scale fault zone extending in the mantle. In case 1 the lower crust and mantle lithosphere are mostly brittle and the deformation is immediately coupled, leading to very localized rifts rooted in the mantle lithosphere. The initial rapid subsidence of this keystone block, defined as block H in Lavier and Manatschal (2006), is followed by rapid necking of the lithosphere and extension and thinning of the internal parts of block H. Case 3 is an analogue of the stretching phase and case 1 of the thinning phase defined by Lavier and Manatschal (2006). Weakening of the crust by hydration (Lavier and Manatschal, 2006) develops listric detachment faults that distribute the deformation in the upper crust. It also allows for the formation of a weak middle crustal layer that decouples brittle deformation between the upper and lower gabbroic crust. Finally the strong gabbroic lower crust formed during the previous phase of extensional collapse act as a guide that distribute the stress elastically or viscoelastically across the whole extensional domain. It becomes a stress concentrator when brittle deformation can localize and propagate through it by plastic (Mohr Coulomb yield) weakening as a function of plastic strain. Figure 13 shows the state of the lithosphere and its topography after 150 km of extension for each case. At first order case 1 develops a narrower conjugate margins with a large zone of exhumed continental mantle. Case 3 develops upper-lower plate asymmetric conjugate margins in a wider rift but a narrow zone of final thinning. This model does not reach exhumation after 150km of extension. Case 2 is similar to that described in Lavier and Manatschal (2006). The extension has stretched the lithosphere down to 20 km thickness (stretching phase) before thinning it down to 10 km during the thinning phase preceding the exhumation phase that brings some serpentinized mantle to the surface along a rolling-hinge exhumation fault.

4.3 Implications

These models confirm that the thermal structure is a key parameter in controlling the rift evolution (Buck, 1991). The results of the models shown in Figures 10, 11 and 12 also emphasize the importance of the thermal structure controlling the necking architecture of the rifted margin. For higher temperature, a larger fraction of the upper lithospheric mantle and the continental crust are ductile. This results in the delocalization of the deformation and makes it hard to develop an area of necking where thinning of the crust and mantle are coupled and Moho-topography develops (Fig 12). In contrast, thinning of the crust and mantle are easily coupled and form sharper necks in colder settings (Fig 10). Thus, the evolution of a rift system depends on the thermal structure of the lithosphere in addition to its composition, rates and obliquity. For the same crustal thickness or tectonic province, young and hot lithosphere may be difficult to thin resulting in wide necking profiles, while old and cold lithosphere (Archean or Paleoproterozoic) may be strong, resulting in narrow necking profiles. The Variscan lithosphere underlying the Alpine Tethys and Iberia Atlantic margins follows a phase of Permian orogenic collapse that introduces a large amount of heat by magmatic processes and lithospheric thinning about 70 myr before onset of rifting. It may still be quite hot and may represent

an intermediate case between very young and very old lithosphere. Of course, these ideas do neither consider compositional variations, nor strain rate, which may strongly modify the bulk rheology of the extending lithosphere (see paragraph 3.2.1 and Fig. 5). At the moment, we do not see how we can overcome the problem of constraining complex structural and compositional inheritance, since we are not yet able to place realistic "structural" or "compositional" inheritances in model and allow them to compete with the thermal one. However, already taking into account variations in the thermal structure may explain why most rift systems, except for those that are related to subductions (e.g. back-arc situations; Tyrrhenian sea; Woodlark, Gulf of California), can not develop directly from an orogenic collapse into a rift system (e.g. Basin and Range). It is likely that rift systems cannot start within a crust/lithosphere that is too hot, probably because thinning of the crust and the mantle cannot be coupled. In contrast, it may also be difficult to initiate a rift system in a very cold lithosphere. This may explain why many rift systems try, in an early stage, to avoid these domains and may prefer weak orogenic lithosphere or other types of weaknesses.

5. Discussion

5.1 Relative role of inheritance vs. rift-induced processes

The integration of inheritance in rift models is a key to eventually be able to predict and understand the evolution of rift systems. However, as compositional, structural and thermal inheritances cannot be disassociated one from each other, it remains at present difficult to link particular rift structures or their evolution directly to one specific type of inheritance. Furthermore, existing structures or weaknesses are not always entirely reactivated as exemplified by the relation between Jurassic extensional detachment fault and Alpine thrust in the vicinity of the a lateral ramp in paragraph 3.3.1, (Fig. 7). Reasons may be that the geometry of the pre-existing structures/weaknesses are complex (e.g. listric faults, ramps etc.), limited in their lateral extend (e.g. confined evaporite layers), their orientation was not favourable to be reactivated, or the inherited weakness is located in a layer not stiff enough to concentrate the stress (e.g. Chenin and Beaumont, 2013). Another important point is that inherited weaknesses can either interact or not depending on their scale and location within the lithosphere. For instance, mantle weaknesses are unlikely to control the location of deformation at the surface as long as deformation in the upper crust is decoupled form the underlying mantle.

In order to introduce this complexity into the description of rift systems, it is important to define types of inheritance and to think about how, when and at what scale they may control the rift evolution or rift architecture. In the previous chapters we showed, based on natural examples that:

- inherited structures, in particular weaknesses, are important in controlling strain localization on a local scale (e.g. Err detachment and Mauléon-Parentis examples; Figs. 7 and 8)
- the necking zones may be influenced by the thickness and distribution of ductile layers during decoupled deformation (e.g. Iberia example; Fig.5; thermal results Figs. 10, 11 and 12)
- conversely, the deformation in the uppermost parts of the crust and exhumed mantle in hyperextended, coupled domains seems not to be controlled by inheritance, but by newly formed weak materials (clay, serpentinite, ...)
- mantle composition and mantle processes occurring before and during rifting may control the magmatic budget, the thermal structure and the localization of final rifting

(e.g. Alps, Pyrenees and Iberia examples; Fig. 4) in addition to the rheology of the mantle

- inherited structures within the crust do not significantly control the location of breakup and the structures in the oceanic crust (e.g. Bay of Biscay and Iberian Atlantic; Fig. 6)
- the inherited thermal structure controls the mode and architecture of rift systems, in particular the necking architecture (e.g. modelling approach; Figs. 10, 11, 12)

These key observations enable us to introduce inheritance in the rift models proposed to describe the evolution of rifted margins. However, it is important to distinguish between inheritance, and rift-induced processes, both of which can control rifting. Examples of rift-induced processes include formation of clay, serpentinization or decompression melting. It is also important to note that the role of inheritance may change as the physical conditions vary during the evolving rifting and as rift-induced processes become more important. Inherited mantle composition may become particularly important when the thermal structure of rifting evolves and magma production becomes an important factor.

5.2 The role of inheritance within an evolving rift system

Modes of rifting

Lavier and Manatschal (2006) proposed a dynamic model for magma-poor rift systems, for which they defined different modes of rifting depending on the deformation process(es) at play. These modes are referred to as stretching, thinning and exhumation (Fig. 14). As discussed by Bronner et al. (2011), breakup at magma-poor rift systems may only occur if an excess magmatic event is able to break the strength of the lithosphere and to trigger steady state seafloor spreading. In order to capture the breakup process, we define a “magmatic mode" pre-dating the "seafloor-spreading mode" (onset of steady-state seafloor spreading).

Independently of their inheritance, margins may go through a sequence of stretching, thinning, exhumation (not mandatory) and magmatic breakup. Although these modes have been defined for the Iberia-Newfoundland and Alpine Tethyan margins (Sutra et al. 2013, Mohn et al. 2013), they can be also recognized in most of the world rifted margins, their importance and expression may change depending on the inheritance. Therefore understanding which mode is controlled by what type of inheritance is a key to describe and predict the evolution of rift systems (Fig. 14).

Stretching mode

During the stretching mode the total horizontal extension and the related crustal and lithospheric thinning are minor, thus the thermal perturbation and the magmatic production due to decompression melting are negligible. Therefore, in the absence of magma (plume), the strength of the lithosphere is controlled by the strongest brittle layer, which plays the role of a stress guider (see Lister and Davis. 1989). Brittle deformation is decoupled along ductile crustal layers or ductile shear zones allowing brittle layers to deform independently from each other. Thus, the evolution of rift basins in the stretching mode is mainly controlled by inheritance/weaknesses in the uppermost brittle layer, which is usually the upper crust. Inherited compositional weaknesses (pre-existing basins or suture zones) or inherited structures (faults or shear zones) may control the localisation during these initial stages of rifting. If magmatic activity occurs at this stage (often related to plumes), magma can control the localization of deformation at this early stage of rifting through thermal weakening (see East African rift).

Thinning mode

Onset of the thinning mode corresponds to the failure of the brittle layers in the lithosphere and terminates with the attenuation of the last ductile layer (Pérez-Gussinyé and Reston, 2001). Thus, the location of necking may be controlled by weaknesses in the strongest brittle layer. For low to moderate thermal gradients this may correspond to the upper mantle (see deep necking depths of (Bassi et al., 1993). Thus, inheritance within the strong upper mantle may be important and control the location of the necking zone and ultimately the area of breakup. The necking process is mainly controlled by the capacity of the crust to decouple from the mantle lithosphere, i.e. the relative abundance of ductile crust over brittle crust, enabling to partition the deformation between the brittle and ductile layers. In contrast to the stretching mode where deformation is mainly limited to the crust, deformation is localized at a lithospheric scale during the thinning mode. How deformation is partitioned, how much pure-shear thinning occurs in the ductile layers and what processes really control the thinning of the crust are still a matter of debate. At this stage of rifting, it is very important to understand how brittle and ductile layers are distributed within the crust and lithosphere. Compositional and structural inheritance may control onset of thinning, while magma and fluids may become more important during the subsequent hyperextension. Therefore knowing the bulk rheological structure of the crust is essential. This may either be controlled by weak phases (hydrated phases, quartz rich rocks or metasediments in the crust) or strong phases (mafic bodies or granulites). The underthrusting or accretion of sediments during subduction or the existence of multiple sutures during accretion may create strong inherited weaknesses that may control the crustal thinning process. In systems with either high geothermal gradients or high percentages of ductile layers coupling may never be accomplished and Moho topography remains subdued despite of high amounts of extension (e.g. Basin and Range Province; Fig. 10). The overall thermal state is mainly controlling the brittle ductile transition. This becomes important because it ultimately controls the relative amount of ductile vs. brittle material in a crustal section. The colder the thermal state the more brittle the crust and lithospheric mantle and the faster coupling between crust and mantle will occur (Fig. 12). Conversely, the more ductile material the more extension is needed to reach the coupling point (see Figs. 10, 11 and 12 as well as Fig. 5).

Exhumation mode

During the exhumation mode, all the remaining continental crust is made of residual brittle layers, either upper curst, mafic lower crust, or former ductile crust that cooled and became brittle during thinning. At this stage, active brittle faults facilitate fluid transport and hydration of the crustal and mantle rocks, resulting in the serpentinization of the uppermost parts of the subcontinental mantle. Intense hydration modifies the original mineralogy controlling rifting (quartz, feldspar and olivine) to a new one controlled by clays and serpentine within the uppermost 6 kilometres of the basement. Thus, the rheology of the remaining crust is strongly controlled by weak, hydrated material. However, field observations show that, even during this late stage, local inheritance is important and the presence of inherited low-friction material (e.g. salt) or inherited structures may control the architecture of detachment systems on a local scale. However, on a larger scale it is not excluded that the newly formed weak material favours pure shear deformation to occur simultaneous to exhumation along localized detachment faults. Indeed, hydration of the crustal rocks (mainly feldspar to illite and chlorite) and mantle rocks (in particular olivine to serpentine) results in the formation of low friction mineral

that may change in a significant way the rheology of the extending uppermost crust and exhumed mantle. It is important to note that these processes are a consequence of rifting rather than controlled by inheritance. This control by the change in mineralogy may be illustrated by the width of the hyperextended crustal wedge along the Iberian margin, which remains the same along the whole margin while the necking profiles vary strongly (e.g. Fig. 5c and discussion in chapter 3.2.1). This suggests that the processes controlling hyperextension and exhumation are different from those controlling necking. We assume that, within the coupled domain (< 10km thick), the material is formed by hydrated residual brittle parts of the continental crust, while further outboard, where the crust is thinned to less than 6km, it may overlie a serpentinized mantle behaving like a Mohr-Coulomb material with a low friction coeofficient.

Thus, the architecture of the hyperextended domain may be primarily controlled by newly formed clay, serpentine or talc resulting from hydration of the crust and/or mantle after their embrittlement, rather than by inheritance.

Magmatic mode

Although magmatic activity may occur at low ß (< 1.5) in rift systems, mostly as a consequence of magma input from a deep mantle source (e.g. East African and Rio Grande rifts), rift-related magmatism is usually a consequence of decompression melting related to extreme lithospheric thinning. The production of magma in rift settings is controlled by a number of processes and parameters, of which the thermal state, the rate of thinning, the shape of the base of the lithosphere, the content of hydrous phases in the lithospheric mantle and its composition. Composition and amount of water in the lithosphere (e.g. inherited depleted or fertile lithospheric mantle) are inherited and the thermal state allowing for magma formation is usually reached after the lithosphere has been thinned by at least 50%.

In most rift models, mantle is considered to be homogenous and magma production controlled by strain (ß). However, the study of mantle rocks derived from the Alpine Tethys and Iberia-Newfoundland rift systems showed that the continental lithosphere involved in rifting is strongly heterogeneous. Although it may be difficult to determine the composition and history of the subcontinental mantle rocks, observations tend to show that two processes are important in controlling the composition of the lithosphere, which are: 1) magmatic processes associated with former subduction zones (supra-subduction vs. subducted mantle); and 2) the post-orogenic thinning and melting (cleaning) of the mantle. It is interesting to note that the Variscan orogenic collapse resulted in strongly depleted subcontinental mantle in the Alps and in Iberia before the onset of rifting. This subcontinental mantle was subsequently fertilized by the melt produced during lithospheric thinning pre-dating lithospheric breakup. This "buffer effect" may explain the delay in the major magmatic pulse during rifting at these two magma-poor rifted margins. In this context it is also interesting to note that some of the classical volcanic rifted margins (magma-rich margins) may have initiated as magma-poor rift systems that underwent hyperextension prior to magmatic breakup. Examples are the North Atlantic Vøre and Møring margins or the NW Shelf off NW Australia (Péron-Pinvidic et al., 2013, Direen et al., 2008). Indeed, at all these margins, the final magmatic breakup overprinted older hyperextended domains, suggesting that the breakup was not related to the previous extensional event, but triggered by the excess of magma. Similar observations can be made in other margins, where the final breakup crosscuts previous rift structures (see southern North Atlantic, Lundin and Doré, 2011). This begs the question of how much magmatic processes related to continental breakup are controlled by

inheritance? For example, for the Norwegian margin two possibilities can be envisaged: either the mantle was enriched as a consequence of earlier thinning (e.g. fertilisation of mantle during earlier rift events as discussed for the example of the Alps (Müntener et al. 2004 and 2009), or it was triggered by the arrival of the plume, in which case no "inheritance" is needed to explain the magmatic breakup.

This hypothesis suggests that the volume of magma generated during breakup is not necessarily a function of inheritance, but can be related to processes evolving in the underlying asthenosphere during hyperextension (e.g. Bronner et al. 2011). One of the reasons rift systems may fail, even after they went to mantle exhumation, is that they were not able to form enough magma to trigger the breakup (e.g. Porcupine, Rockall Trough, Orphan). This suggests that magmatic processes may represent the key control for continental breakup, being able to localize extension and force the lithosphere to break at any time. If this is true, lithospheric breakup is not controlled by extension or inheritance, but by asthenospheric (magmatic) processes.

Conclusion

The key question that guided this study was to what extent inheritance may control the architecture, evolution and magmatic budget of hyperextended rifted margins. We defined 3 types of inheritance, namely structural, compositional and thermal inheritance and developed a simple and robust terminology able to describe and link observations made at different scales using geological, geophysical and modelling tools. Based on this approach, we focused on 3 well-studied examples that are the Alpine Tethys, Pyrenean-Bay of Biscay and Iberia-Newfoundland rift systems. Our work tends to show that, for the examples discussed in this paper:

- inherited structures are important in controlling strain localization on a local scale and very early in the rift system.
- shape of necking zones may be controlled by the occurrence of ductile layers within the crust, while necking efficiency may depend on stiffness of the necked layers
- location of the necking zone may be controlled by the location where the strongest brittle layer may yield (in cold systems this may be the location where the upper lithospheric mantle starts to break).
- deformation in the hyperextended and exhumation domains is primarily controlled by newly formed weak materials (clay, serpentinite, ...) that result from fluid circulation and reaction assisted deformation.
- mantle composition and mantle processes occurring before and during rifting may control the rheology of the mantle, the magmatic budget, the thermal structure and the localization of deformation during late stages of continental crust attenuation and lithospheric breakup, while early stages are rather controlled by upper crustal inherited structures
- inherited structures in the continental crust do not significantly control the location of breakup and the structures in the oceanic crust.
- the inherited thermal structure controls the mode and architecture of rift systems, in particular the necking architecture

We conclude that both inheritance and rift-induced processes play a significant role in the development of rifted margins and need therefore to be introduced in the study of rift systems. However, as this study is exclusively based on the magma-poor Iberia-Newfoundland and Alpine Tethys hyperextended rift systems that developed over

Variscan lithosphere, the applicability of our conclusions to other, less explored systems remains to be tested.

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Figures

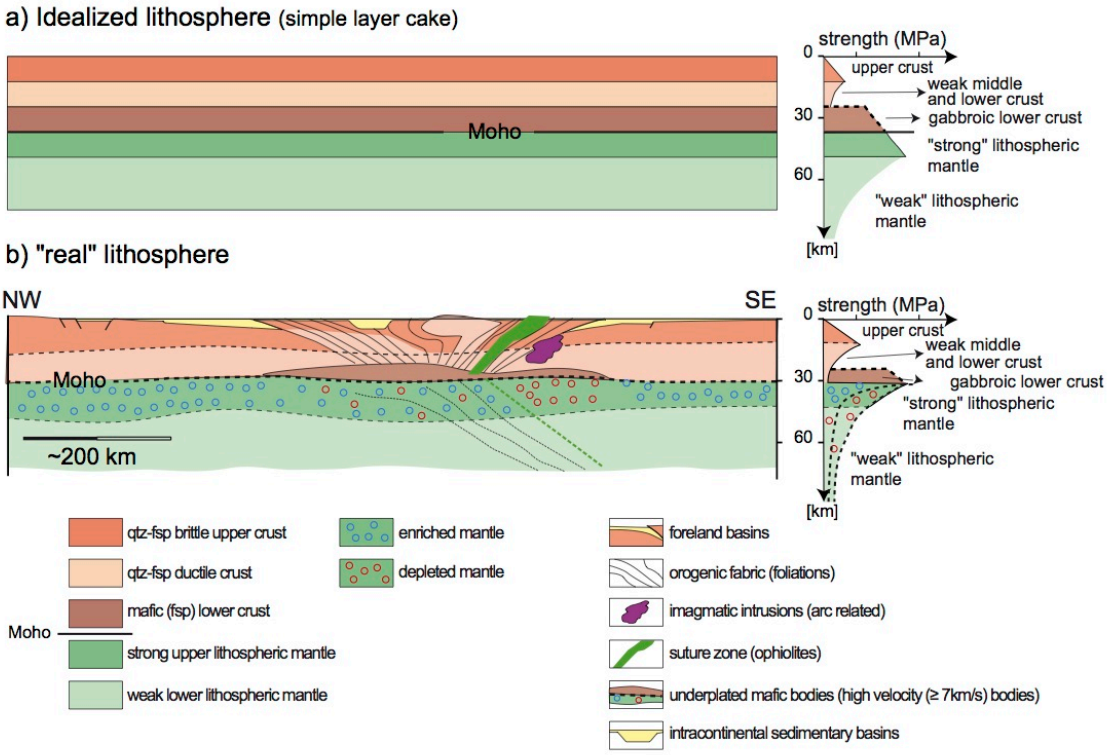


Fig. 1

Fig. 1: Lithospheric scale sections showing (a) an idealized lithosphere made of a layer cake and thermally equilibrated) and (b) a "real" post-orogenic lithosphere showing inherited structural and compositional complexity.

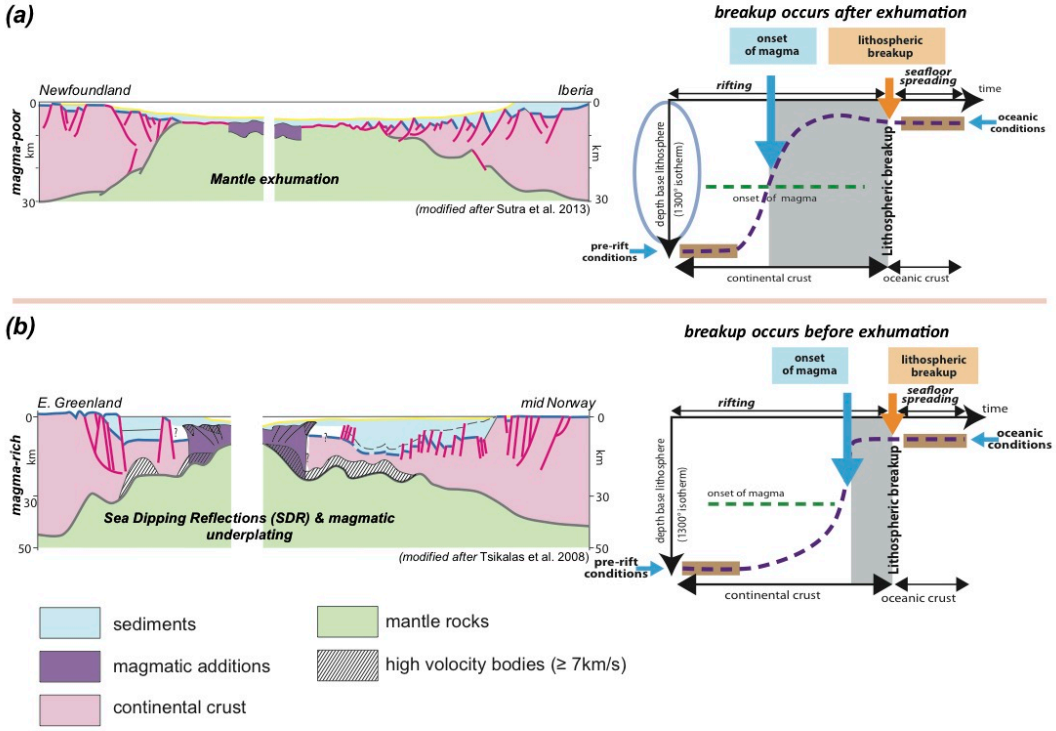


Fig. 2

Fig. 2: Two sections showing (a) a magma-poor rifted margin and (b) a magma-rich rifted margin. The T-t evolution for a magma-rich margin and a magma poor margin is shown in a diagram where the vertical axis corresponds to the depth of the lithosphere (defined as the 1300°C isotherm) and the horizontal axis to the time axes showing the evolution of a system from rifting to seafloor spreading. The violet line shows the depth of the lithosphere as a function of time. The green line corresponds to the onset of magma-production that occurs when the lithosphere is thinned to < 50%. The grey domain corresponds to the time between first production of magma and the lithospheric breakup. Note that at magma-poor rifted margins breakup occurs after mantle has been exhumed, while at magma-rich rifted margins, breakup may occur before separation of the two continents.

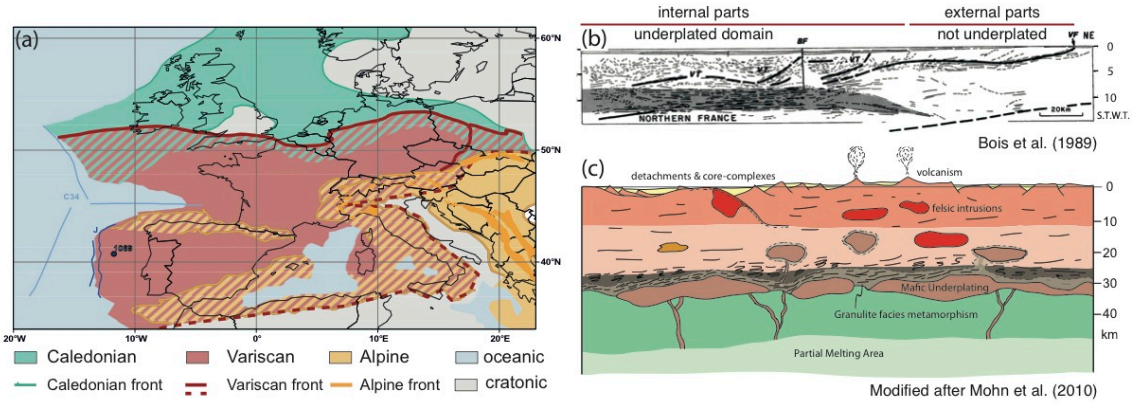


Fig. 3

Fig. 3: (a) Map showing the nature of the lithosphere in Western Europe. Between cratonic lithosphere in the northeast (Fenoscandinavia) and oceanic lithosphere in the West (North Atlantic) and in the Mediterranan (back arc basins) the reminder of the curst is either Caledonian, Variscan or Alpine (modified after Stojadinovic et al. 2013; Schmid et al. 2004; Robertson et al. 2013; Cowie & Kusznir 2012, Barrier et al. 2004). (b) Seismic section across the Northern Variscan domain in NE France showing the occurrence of underplated mafic crust (high velocity reflective crust) underneath the domain affected by the Variscan orogen. Note that the termination of the underplated body coincides with the northern border of the Variscan orogen (from Bois et al. 1989). (c) Schematic cross section showing the composition and major structures forming the crust and upper lithosphere at the end of late Permian extensional collapse (modified after Mohn et al. 2010). Note the occurrence of mafic (gabbroic) magmatic bodies at the base of the crust as well as in mid crustal levels while the upper crust is formed predominantly by felsic intrusions preferentially emplaced in upper crustal levels before 300Ma. In contrast most mafic intrusions range between 300 and 250Ma. Note also the occurrence of granulite facies contact metamorphism of Permian age that is associated with the emplacement of the gabbros. The shallow position of the asthenosphere during the Permian magmatic event results in high geothermal gradients and the depletion of the mantle rocks (see text for explanations and references).

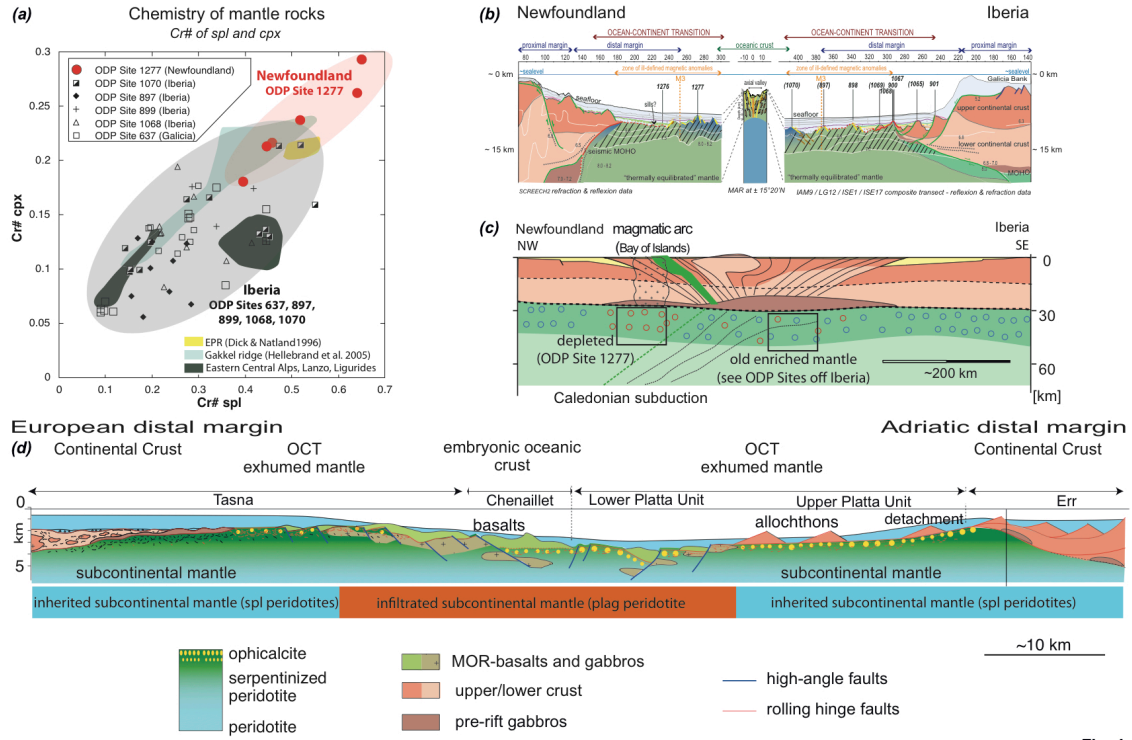


Fig. 4

Fig. 4: Nature and composition of mantle rocks drilled along the Iberia-Newfoundland rifted margins and exposed in the Alps. (a) Cr# of cpx versus Cr# of spinel in peridotites from the Iberia–Newfoundland rifted margins. Light blue shaded field are spinel peridotits from the Lena trough (Dick and Natland, 1996), dark grey shaded field are spinel and plagioclase peridotites from the Eastern Central Alps (Müntener et al., 2004 and 2009), and the yellow shaded field is from the East Pacific Rise (Dick and Natland, 1996). Note that all samples from Site 1277 (Newfoundland margin) are aligned along the Cr# covariation trend for spinel peridotites, while many samples from Iberia show substantial deviation towards higher Cr# in spinel. This offset has been interpreted by partial equilibration with plagioclase (for more details see Müntener and Manatschal, 2006). (b) Present-day section across the Iberia-Newfoundland rifted margin showing the location of the drill samples (modified after Péron-Pinvidic and Manatschal, 2009). Numbers refer to basement rocks drilled along the Iberia-Newfoundland margin. (c) Simplified section across Iberia-Newfoundland showing the situation prior to onset of rifting. Numbers refer to basement rocks drilled along the Iberia-Newfoundland margin; for present-day position see section shown in Fig. 3b. (d) Distribution of subcontinental inherited and infiltrated domains of mantle peridotite in the reconstructed OCT in the Alps (modified after Manatschal and Müntener, 2009, and Müntener et al., 2009).

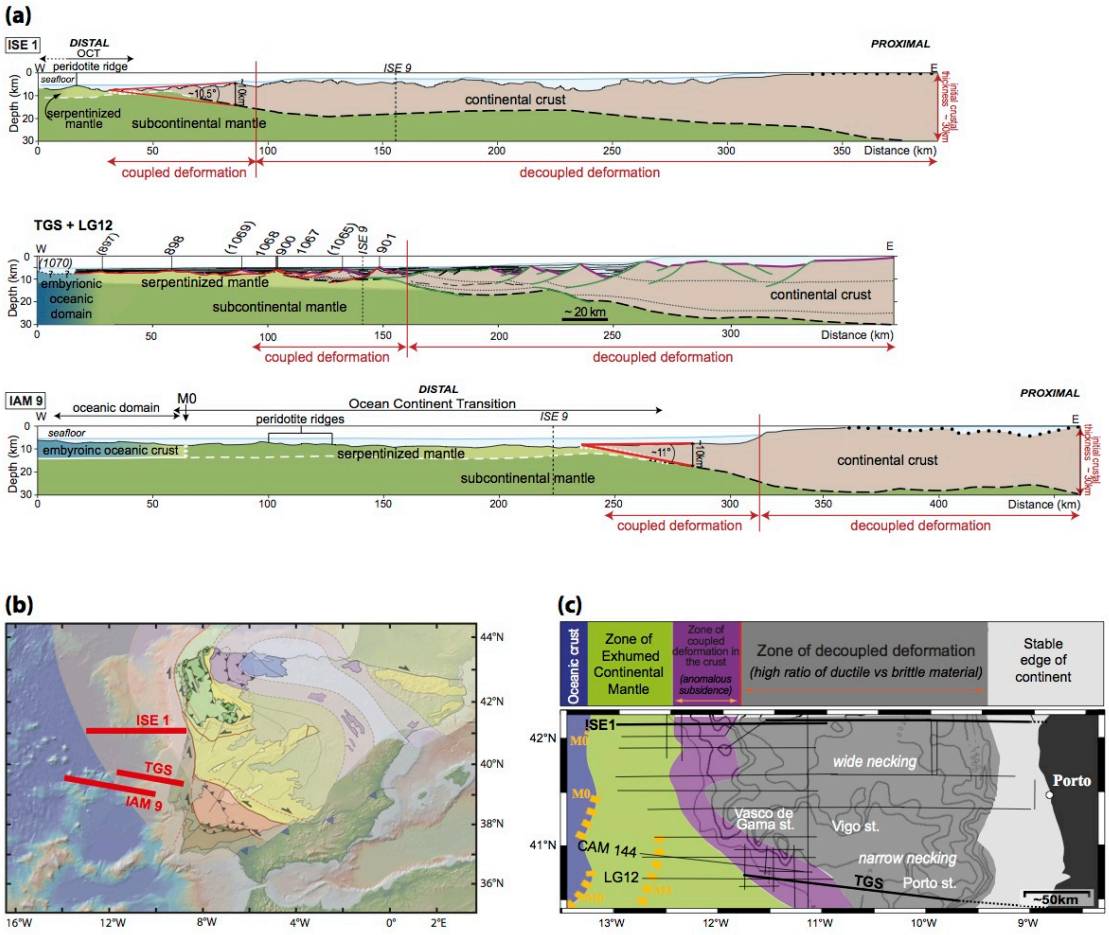


Fig. 5

Fig. 5: Crustal architecture and necking profiles along the Iberia margin (a) Sections across the Iberia margin (for details see text, for location see Fig. 5b). The lines show the first order architecture of the continental crust, i.e. the change in crustal thickness across the margin. In the sections domains of decoupled thinning (e.g. necking zone), can be distinguished from a domain of coupled deformation (hyperextended crust) that lies between the coupling point and the point of exhumation (for details see Sutra et al., 2013). Angle of crustal wedge within the coupled domain is indicated in red. (b) Map showing the position of the lines shown in Fig. 5a. The map shows also the distribution of the crustal domains in Iberia compiled from Martínez-Catalán (2012), Ryan et al. (2009) and Higgins and Leslie (2000). (c) Map showing the distribution of decoupled deformation (necking domain) from coupled deformation (hyperextended domain) along the northwestern Iberia margin (modified after Sutra et al., 2012).

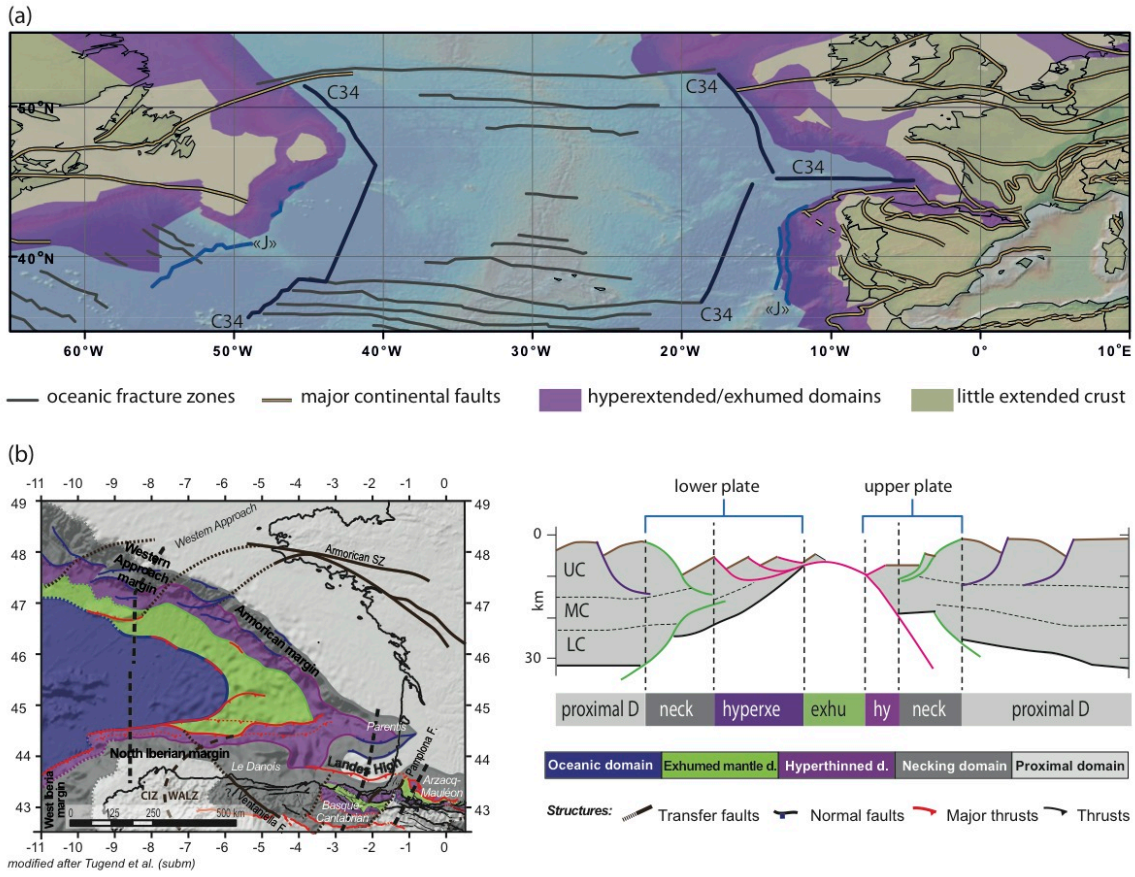


Fig. 6

Fig. 6: (a) Map of the southern N-Atlantic showing the major structures such as oceanic fracture zones, OCT, necking zones, continental fracture zones and the magnetic anomalies 34 and J that are supposed to correspond to the continent ward limit of unequivocal oceanic crust and embryonic oceanic crust respectively. Maps compiled after Arenas et al. (2007), Ziegler and Dézes (2006), Mosar (2003) and Ryan et al. (2009). (b) Map of rift domains preserved in the Bay of Biscay and Western Pyrenees ; CIZ: Central Iberian zone. WALZ: West Asturian-Leonese zone. NPF: North Pyrenean fault (modified after Tugend et al. subm).

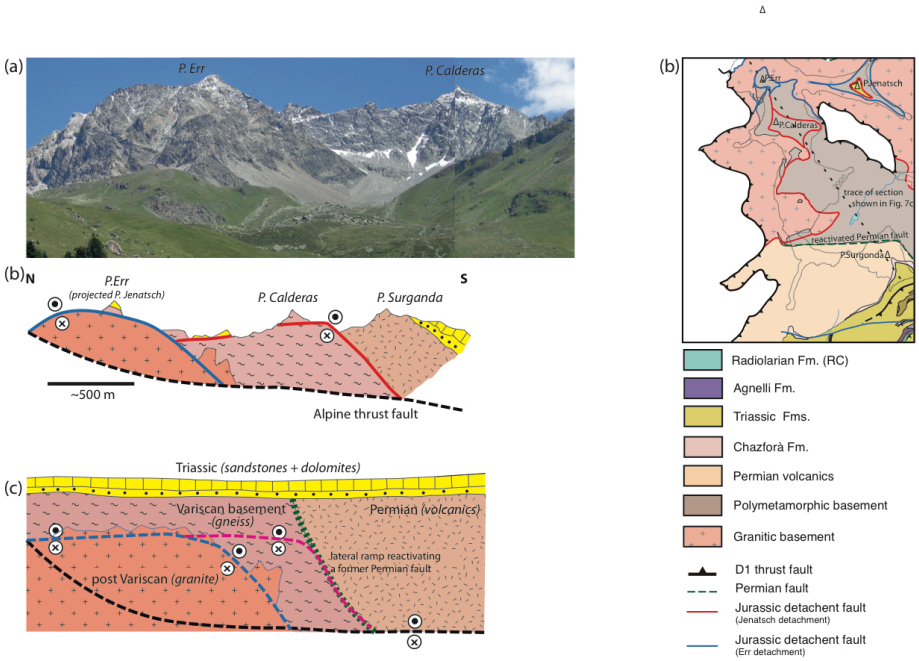


Fig. 7

Fig. 7: (a) Panoramic view of the Jurassic extensional Err and Jenatsch detachment faults exposed in the Err nappe in SE Switzerland. View from west towards the East. (b) Map of the Err nappe in the area of Piz Err – Piz Surgonda, showing relations between the Jurassic extensional detachment system (in blue and red colours), a pre-existing Permian normal fault (in green colour) and an Alpine thrust fault (in black). (c) NW-SE cross section across the Err nappe showing the present-day relation between the Jurassic extensional detachment faults, the Permian fault and the Alpine D1 thrust fault. (d) A restored section showing the relationships before onset of Jurassic rifting and the traces of the future faults.

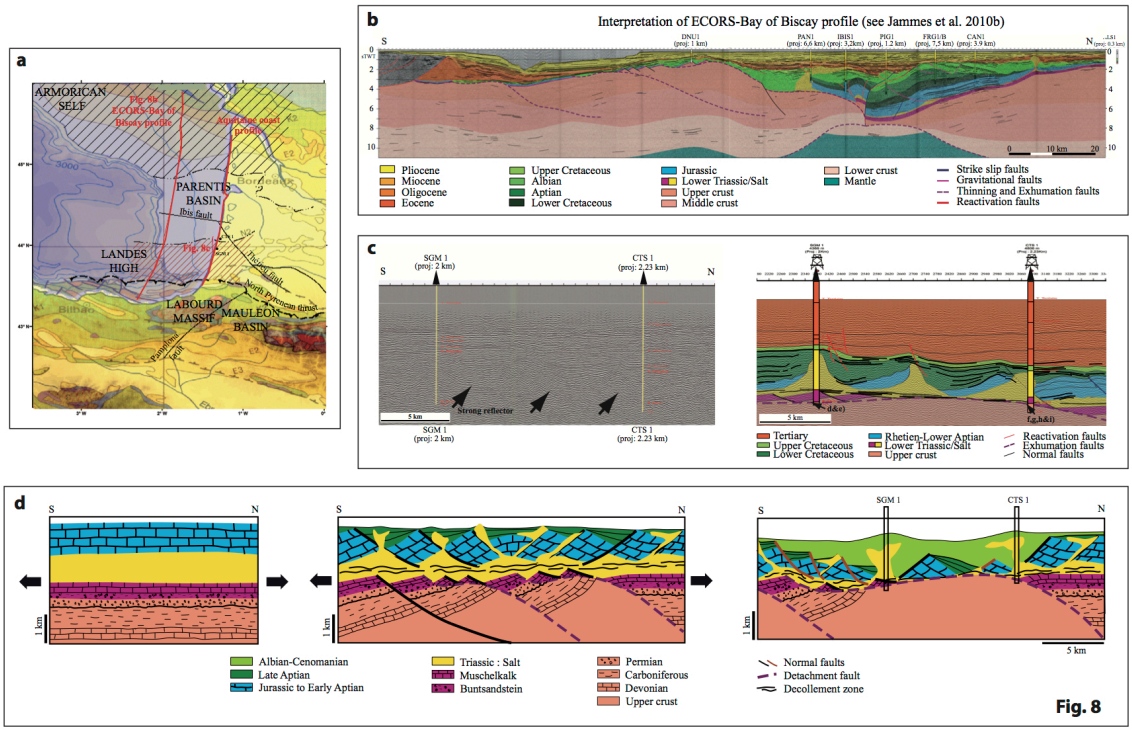


Fig. 8: Relationships between pre-rift salt and extensional detachment faults in the Parentis and Mauléon hyperextended basins. (a) Map showing location of the Parentis and Mauleon basins and location of seismic profiles shown in this figure. (b) ECORS-Bay of Biscay profile with location of drill-holes. Interpretation taken from Jammes et al. (2010b). CAN1 = Cormoran; FRG1/B = Frégate; PIG1 = Pingouin; PAN1 = Pélican; IBIS2 = Ibis 2; DNU1 = Danu; ALS1 = Albatros; TWT = two-way traveltime. (c) Zoom on the Aquitaine coast profile (modified after Jammes et al. 2010b). Note the presence of the asymmetric blocks bounded by growth structures and soled by a strong reflection indicated by black arrows. To the right, interpretation of the profile proposed by Jammes et al. (2010b). Note also that drill holes penetrate deformed basement underneath the Upper Triassic evaporate horizon. For more discussion see text. SGM1 and CTS1 = Saint Girons en Marensin and Contis wells; line is in TWT (= two-way travel time). (d) Conceptual model showing the function of the Triassic pre-salt during extreme crustal thinning from the pre-rift stage to the stretching stage to the thinning stage. Note the decoupling between the detachment faults responsible for crustal thinning and the pure shear deformation within the Upper Triassic to Jurassic limestones. For location of wells SGM1 and CTS1 see Fig. 8c and for details Jammes et al. (2010b).

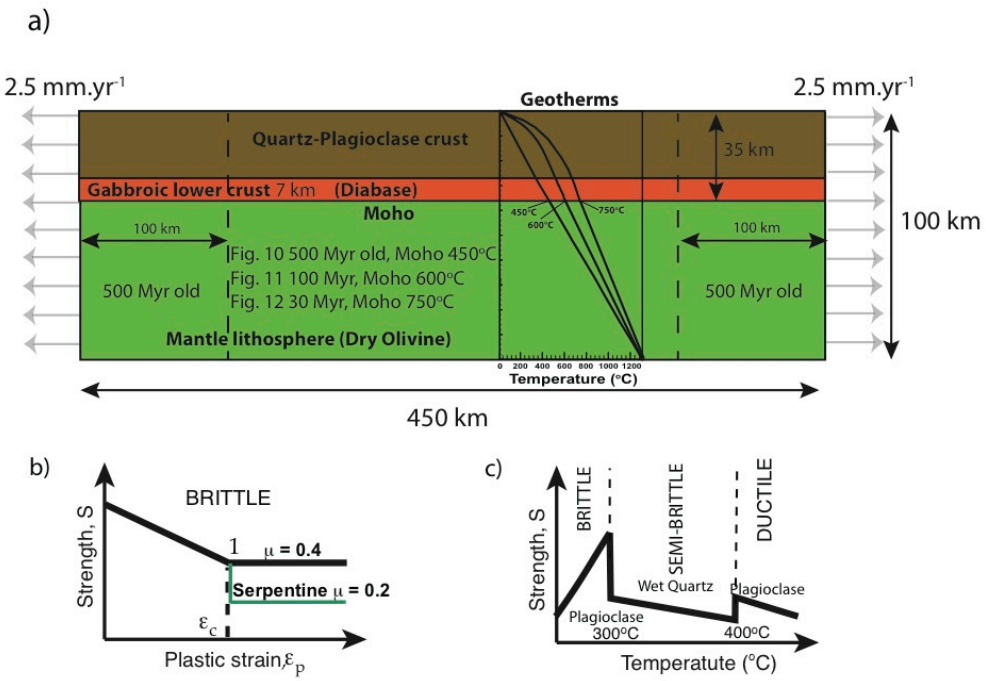
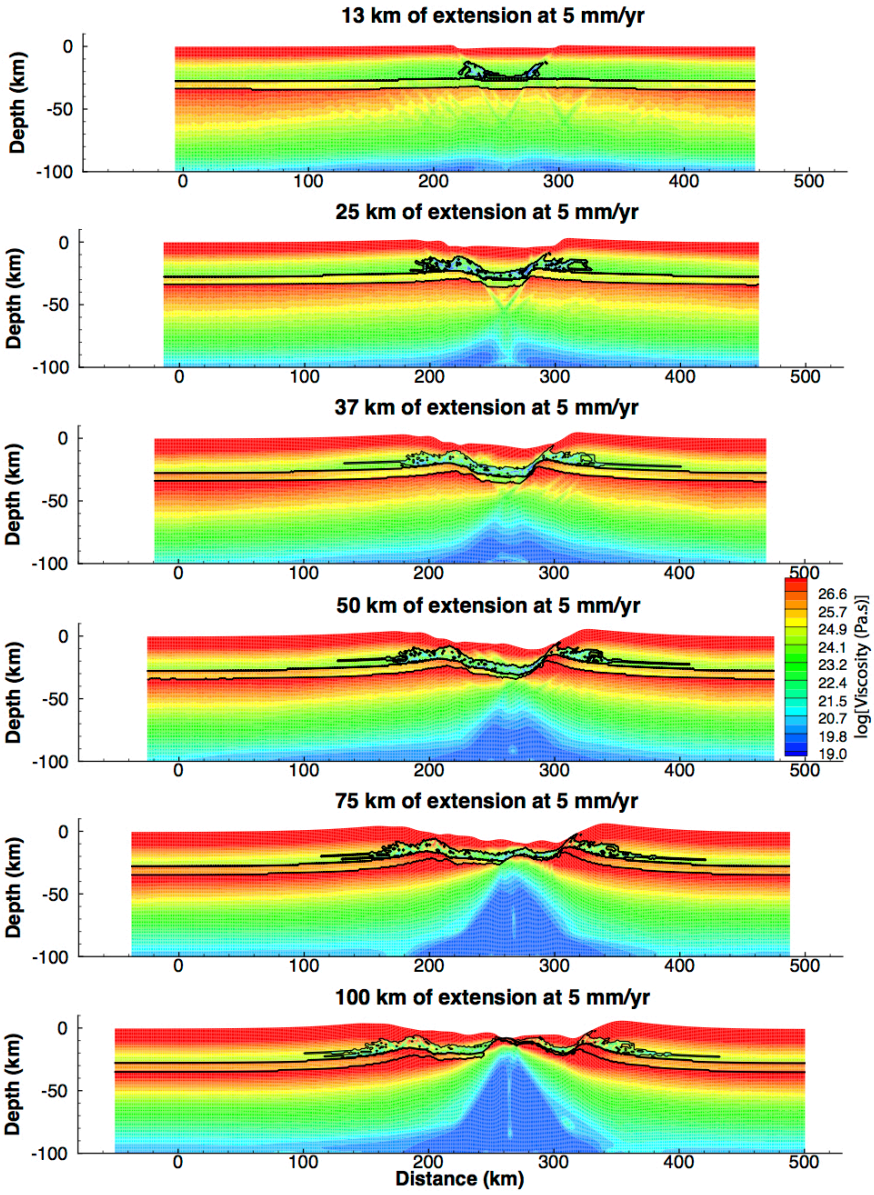


Fig. 9: (a) The sides of the model are pulled apart at a constant half-rate of 2.5 mmyr^{-1}. Isostatic equilibrium is simulated at the bottom while the top surface is stress free to model the topographic evolution. The model is 450 km wide and 100 km deep. Two zones of 100 km of each side of the models are set with a geotherm corresponding to a 500 Myr old lithosphere. The center is set at 500 Myr, 100 Myr and 30 Myr depending on the cases analyzed (see figure 10, 11 & 12). The crust is 35 km thick with a 7 km thick gabbroic layer in the lower crust. In the brittle areas, the lithosphere is elastoplastic with a Mohr-Coulomb yield criterion. Crustal and mantle density are 2800 kg.m^{-3} and 3300 kg.m^{-3} respectively. Power law creep parameters are: crust, quartz (exponent, n = 3, Activation energy, Q = 2 10 J.mol^{-1}, pre-exponent, A = 5. 10^{2} MPa^{-n}.s^{-1}) and plagioclase (n = 3.2, Q = 2.38 105 J.mol^{-1}, A = 3.3 10-4 MPa^{-n}.s^{-1}), gabbroic lower crust (n = 3.05, Q = 3.5 105 J.mol^{-1}, A = 1.25 10^{-1} MPa^{-n}.s^{-1}), mantle, dry olivine (n = 3, Q = 5.2 10^{5} J.mol^{-1}, A = 7. 10^{4} MPa^{-n}.s^{-1}) (see Lavier and Manatschal (2006) and references therein for additional details). b) The material is both frictional and cohesional (initially the friction coefficient, μ is 0.6 and the cohesion, C is 44 MPa). Both cohesion and friction decrease locally as a function of plastic strain (10%) to trigger the formation of shear bands (to μ = 0.3 and C = 4 MPa). Serpentinization occurs in the mantle for temperature < 600ºC and depth < 10 km. There we further decrease the friction coefficient to 0.4. c) Schematic decrease in viscosity associated with the formation of a ductile shear zone. The viscosity decreases from that of plagioclase to quartz when the total work is equal to 4.10^{6}J and the temperature is between 300ºC and 400ºC. This parameterization leads to the formation of a mixed zone of weak ductile material mixed with strong brittle material (for additional details see Lavier and Manatschal (2006)).

A- Evolution of the viscosity through time - Low temperature Moho (450°C).



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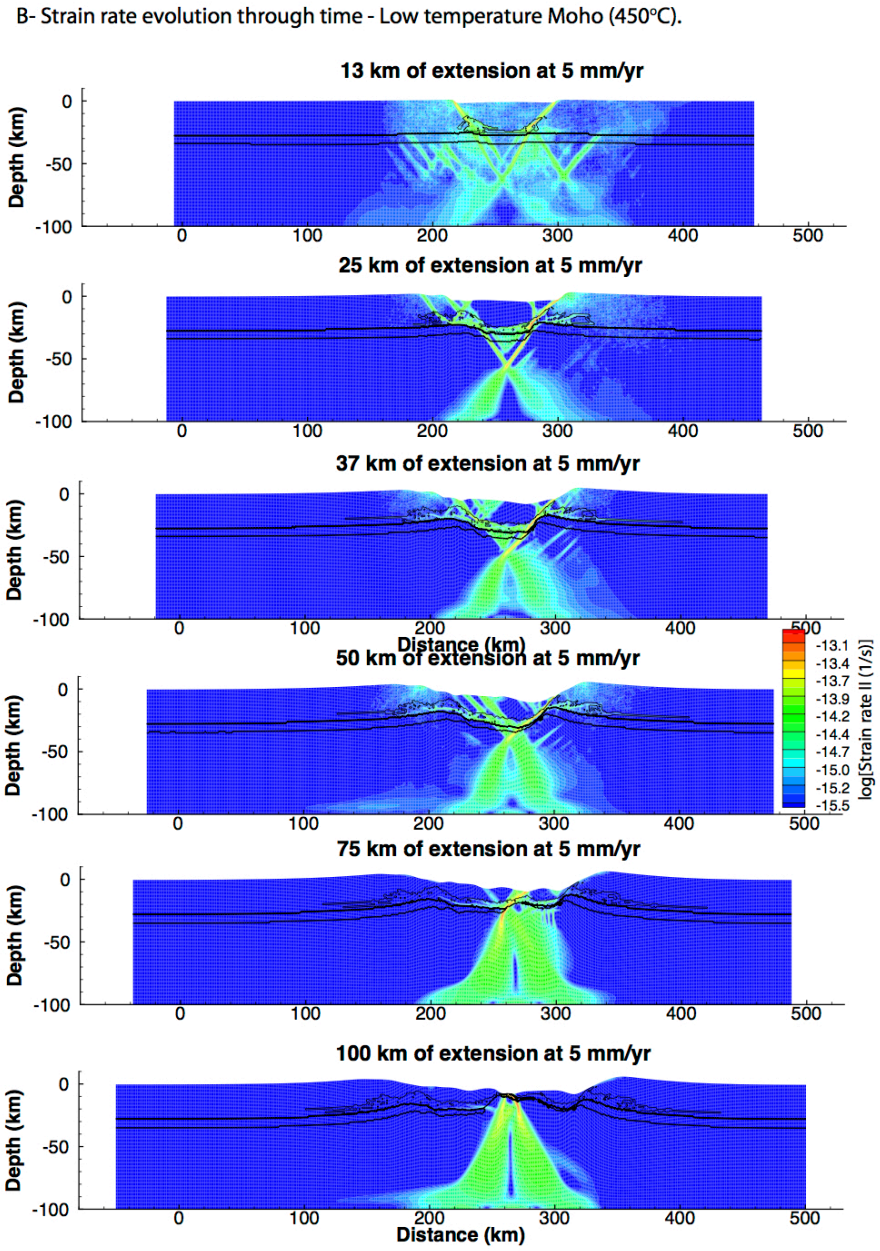
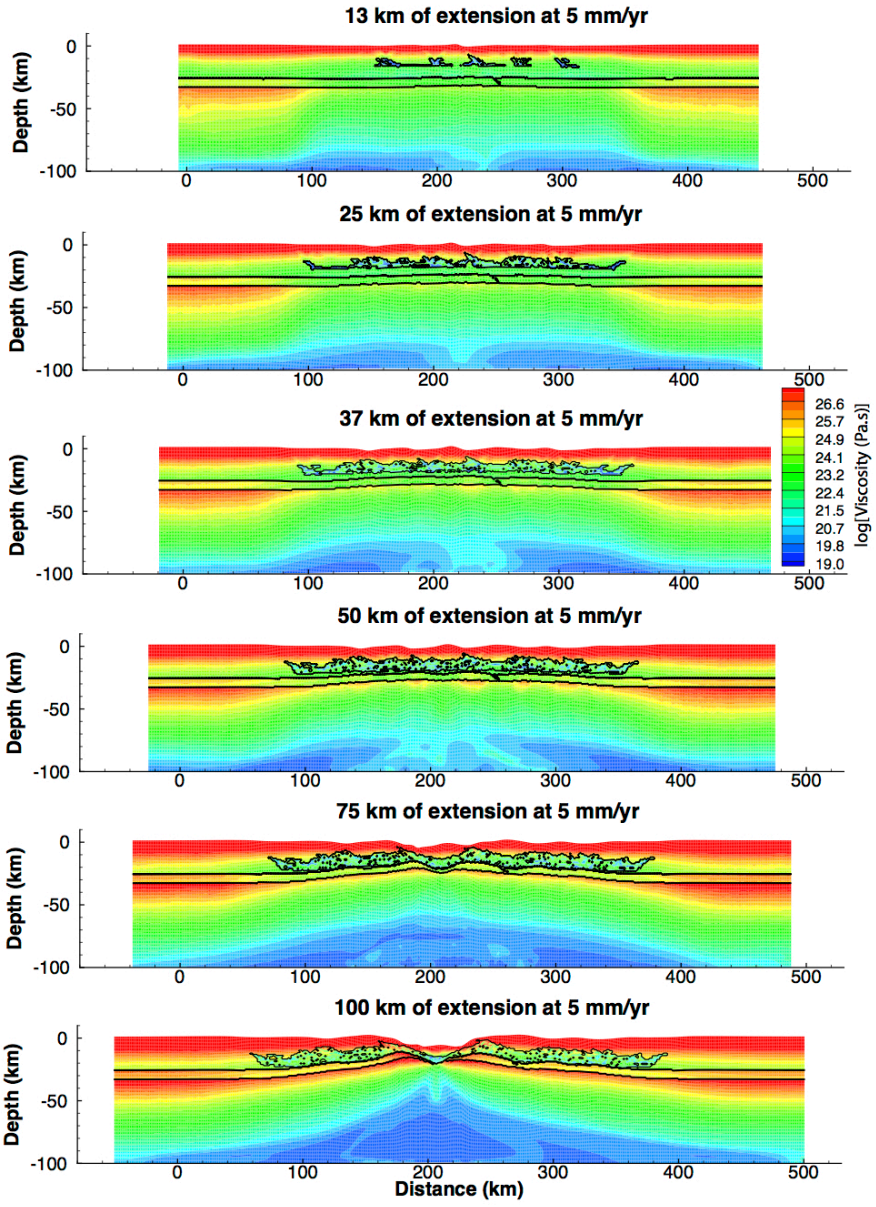


Fig. 10: The thermal structure is set such that the Moho temperature is 750ºC and the bottom temperature is 1330ºC. This situation is the case of a hot lithosphere corresponding to conditions similar to that past orogenic collapse. The lower and middle crusts are very weak and the mantle is mostly ductile. A) Evolution of the viscosity over 100 km or 20 Myr of extension. The lithosphere has a low average viscosity (~10^{23} Pa.s). The deformation is mostly viscous and crust and mantle deformation are decoupled for most of the evolution. Three small basins form and accommodated part of the extension in a brittle manner. As more extension is accommodated a shallow layer of hydrated middle crust forms and link the different basin. The crust and mantle thin uniformly while the

middle crust is slowly exhumed close to the surface. After 75 km of extension the crust is uniformly thinned down to ~20 km. After 100 km, extension localizes in a very narrow basin or keystone block (near the distance 190 km), the deformation is still very distributed and the weak middle crust is exhumed on the flanks of the keystone. B) Until 50 km of extension, the deformation is distributed in the middle to lower crust and mantle lithosphere with some normal faulting in the crust. After this period of ductile and distributed thinning, a large number of normal faults form across the zone of extension. This marks a period of embrittlement of the lithosphere that leads to the localization of the extension in one narrow keystone block (100 km) of extension.

A- Evolution of the viscosity through time - Medium temperature Moho (600°C).



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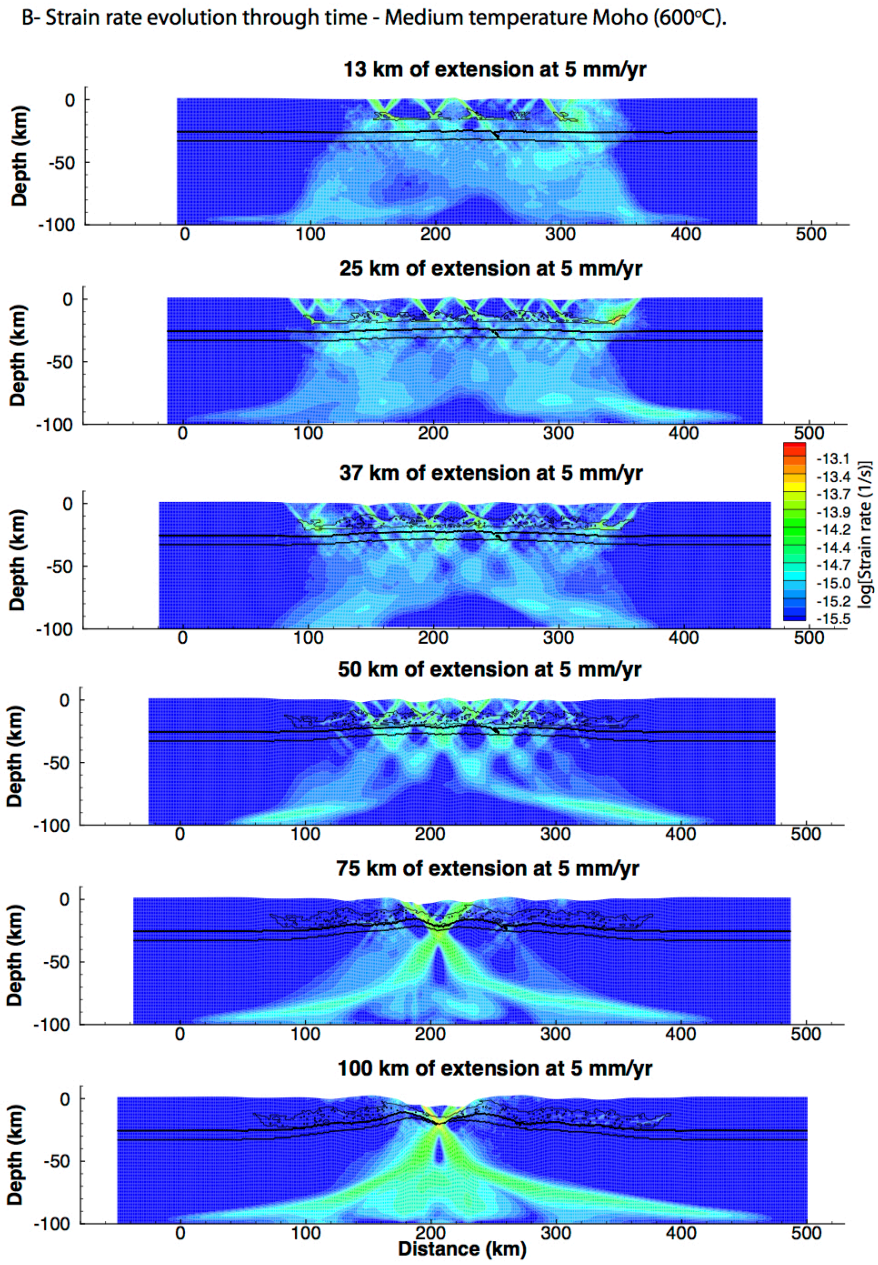
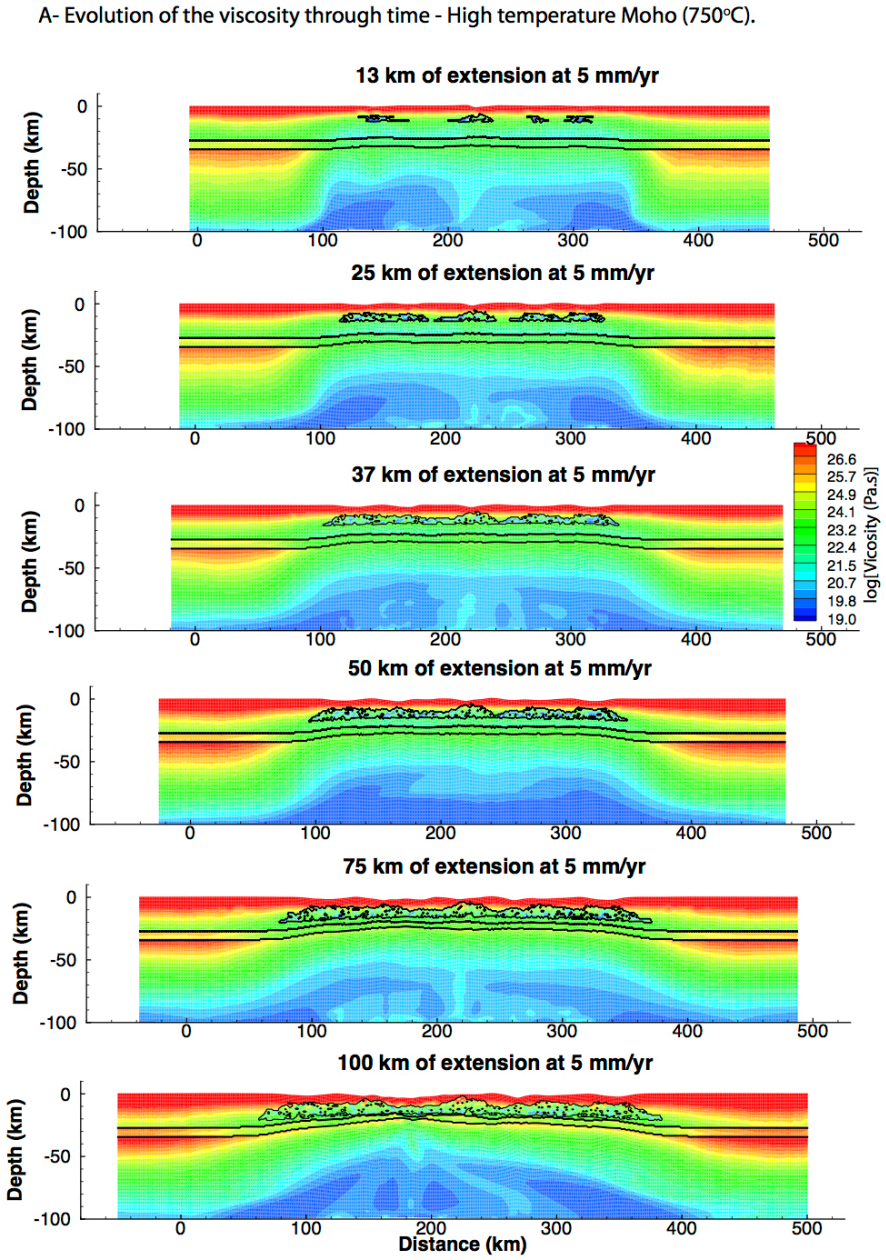


Fig. 11: The thermal structure is set such that the Moho temperature is 600ºC and the bottom temperature is 1330ºC. This situation is the case of a hot lithosphere corresponding to conditions of a young basin (100 Myr past orogenic collapse). A) Evolution of the viscosity over 100 km or 20 Myr of extension at 5 mm.yr^{-1}. In that case a keystone block H forms between 50 and 75 km of extension. This follows a period of pure shear distributed extension in the upper crust accompanied by the formation of mid-crustal layer of hydrated crust. The deformation in the crust and the mantle is decoupled until the keystone forms. The keystone is narrower (~40 km wide) than in the previous case. After 100 km of extension the crust is thinned to < 100 km and the mantle

lithosphere upwelling in a narrow channel below the crust. B) Deformation is distributed in the upper crust from the onset of deformation suggesting that a large proportion of the lithosphere can behave in a brittle manner. After 25 km of extension multiple half grabens have formed in the upper and the mantle lithosphere starts to deform in a brittle manner along multiple normal shear zone. After 50 km of extension the deformation zone narrows until it localizes in a keystone block (75 km of extension). Between 75 and 100 km of extension the crust and lithosphere thin in the keystone block (see previous case).



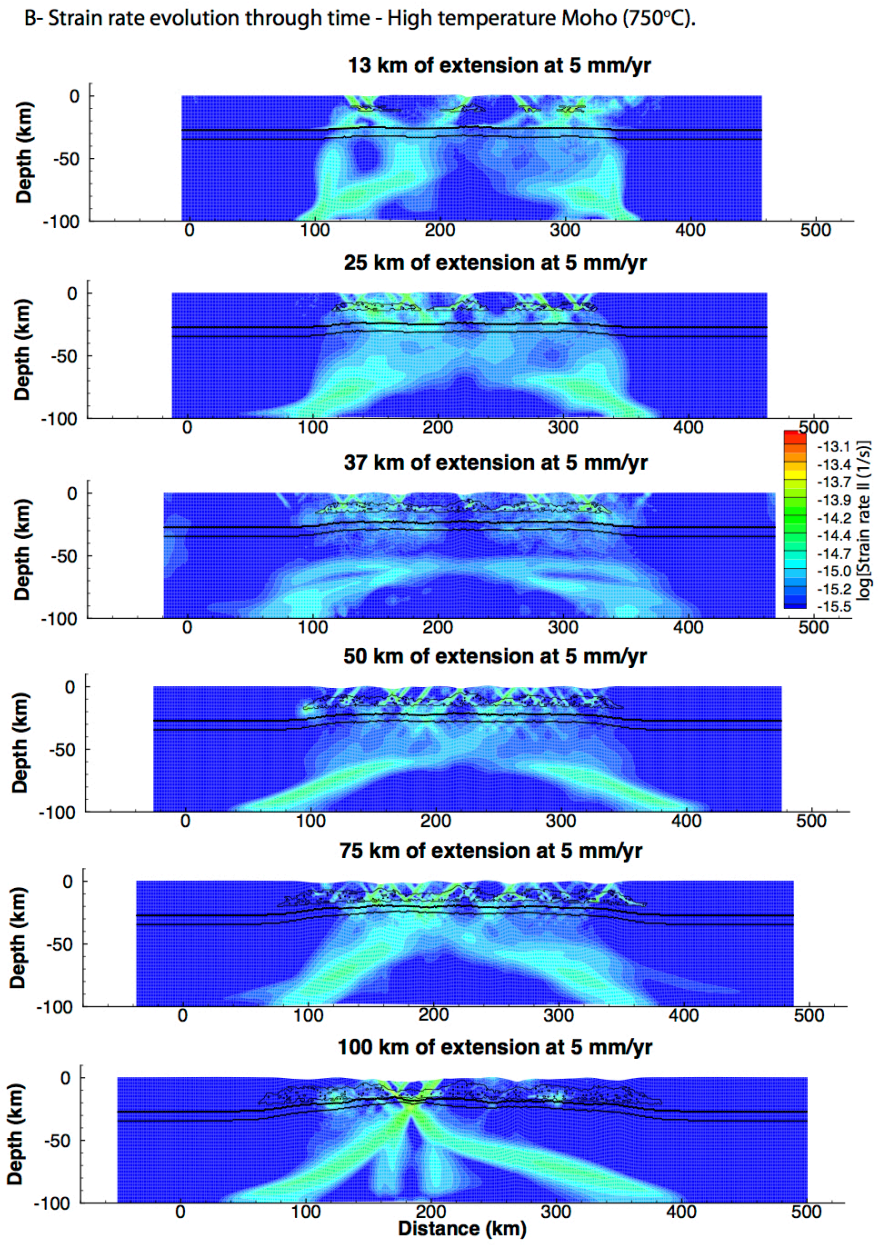


Fig. 12: The thermal structure is set such that the Moho temperature is 450ºC and the bottom temperature is 1330ºC. This situation is the case of a cold lithosphere. These are conditions in a very old orogeny (500 Myr) and a very strong lithosphere. The lower and middle crusts are very strong and brittle and the mantle lithosphere is brittle down to 50 km. A) Evolution of the viscosity over 100 km or 20 Myr of extension at 5 mm.yr^{-1}. The models show localization of the deformation in a 90 km wide keystone (initial block H) that evolves in a very localized and narrow set of conjugate margins (100 km of extension). Zones of weak hydrated crust (outlined in with black lines in the middle crust) form at the base of the brittle shear zone in the crust. The rift flanks are very developed

(~5 km high) as well as the depth of the basin (~ 5 km). This situation is simulating an extreme case of thinning phase of the lithosphere. The crust thins from 35 km to almost 0 km in 75 km of extension. This is unlikely to happen since the force needed to develop such a rift is > 10^{13} $N.m^{-1}$. Additional weakening from magmatic processes is needed (Buck, 2005). B) Evolution of the second invariant of the strain rate that represents the effective shear strain rate and instantaneous deformation. From the beginning the deformation is localized on a few brittle fault zones and is coupled between the crust and mantle lithosphere. After 37 km of extension the deformation migrates inside the initial keystone block. After 75 km the deformation localizes into a major rolling-hinge that exhumes the mantle lithosphere at the bottom of a basin. This may correspond to the exhumation phase (Lavier and Manatschal, 2006).

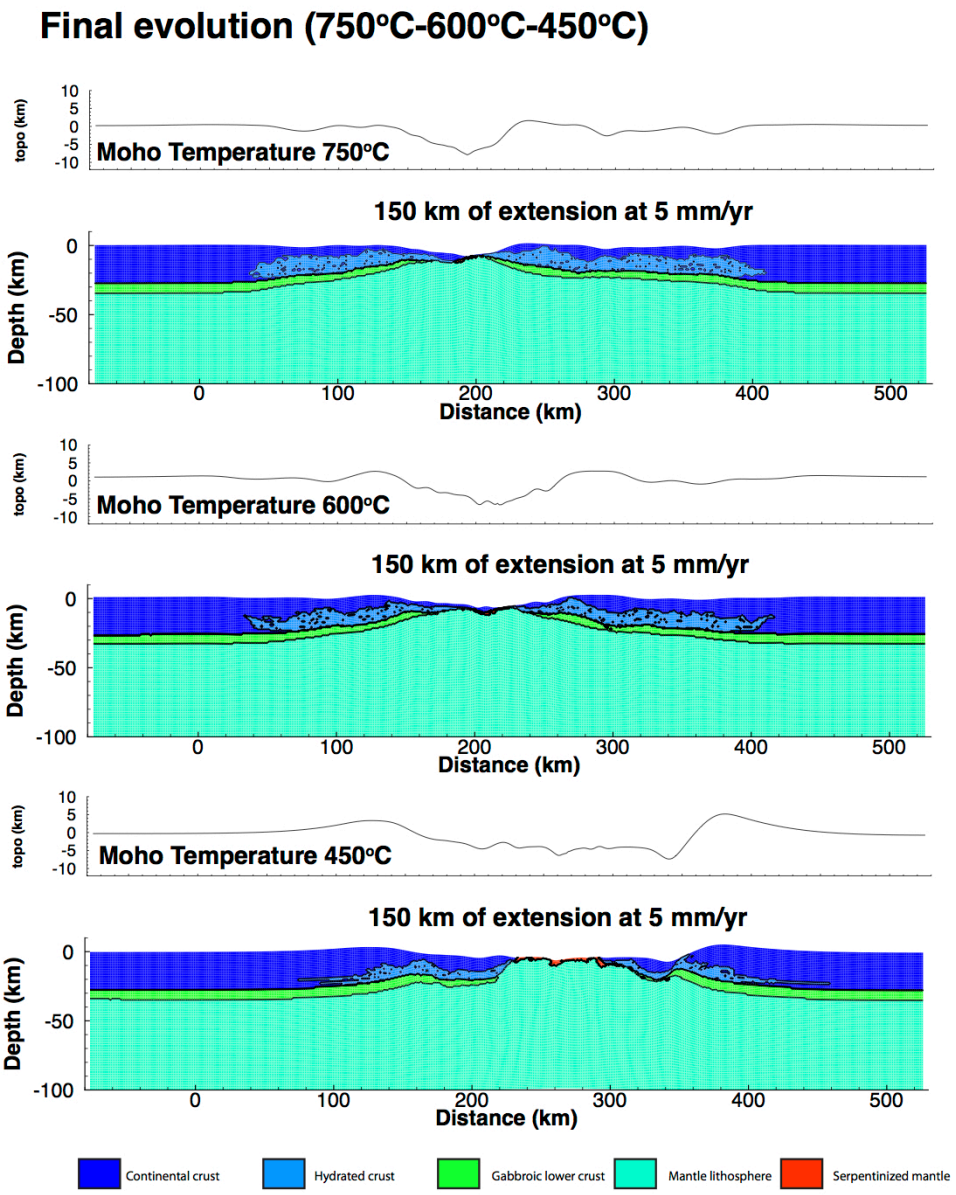


Fig. 13: Final structure and topography of the lithosphere for each case after 150 km of extension. In all cases a localized thinned zone of crust forms. In the coldest case (Moho at 450ºC) the final basin is wide (~230 km) with a large zone of exhumed serpentinized mantle and a very pronounced topography (max ~10 km). For the hottest case (Moho at 750ºC) the basin is narrower (<100 km) and the topography is very asymmetric. The margin has developed a very clear upper-lower plate structure. Moreover the hydrated middle crust is exhumed close to the surface. Finally the medium case (Moho at 600ºC) is akin to that of the Lavier and Manatschal (2006) case. The basin is asymmetric and some exhumation has occurred and the basin width is intermediate after the thinning phase (~160 km).

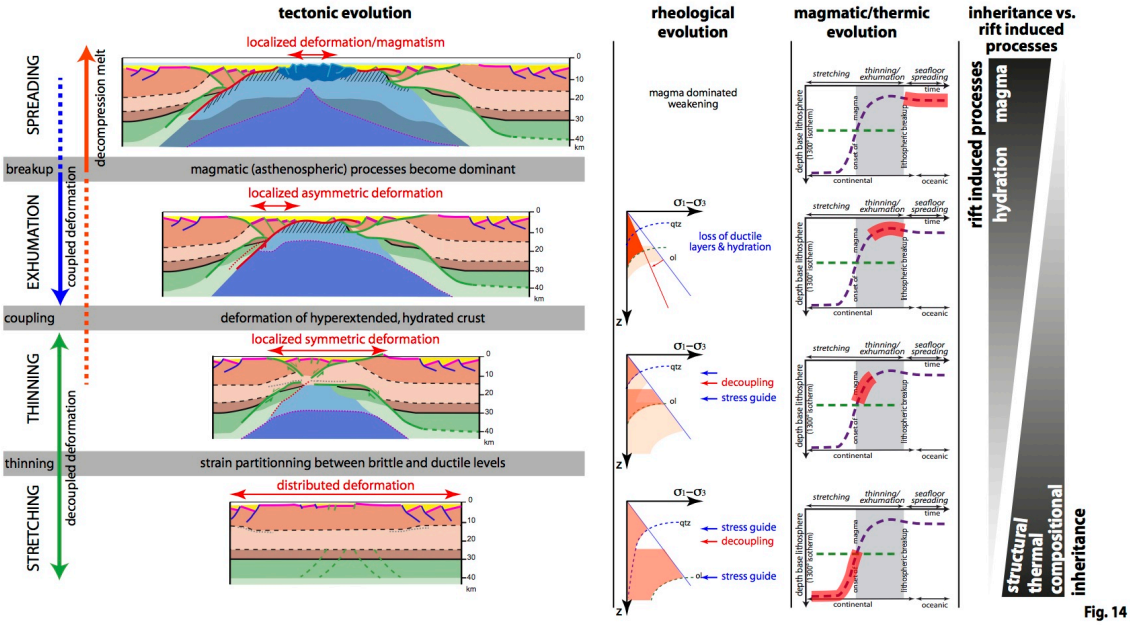


Fig. 14

Fig. 14: Conceptual model showing the tectonic, rheological and magmatic and thermal evolution of a magma-poor rifted margins and its relation to inheritance and rift induced processes.

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