Observations from the Alpine Tethys and Iberia–Newfoundland margins pertinent to the interpretation of continental breakup

G. MANATSCHAL[1], O. MÜNTENER[2], L. L. LAVIER[3], T. A. MINSHULL[4] &
G. PÉRON-PINVIDIC[1]

[1]*CGS-EOST, 1 rue Blessig, F-67084 Strasbourg, France (e-mail: manat@illite.u-strasng.fr)*

[2]*Institute of Geological Sciences, University of Bern, Baltzerstrasse 1, CH-3012 Bern, Switzerland*

[3]*University of Texas Institute for Geophysics, Jackson School of Geosciences, Austin, TX, 78759, USA*

[4]*Southampton Oceanography Centre, European Way, Southampton SO14 3ZH, UK*

Abstract: Although the Iberia–Newfoundland and Alpine Tethys margins are of different age and ultimately had a different fate, they share remarkable similarities. Both pairs of margins show a change from initially distributed and decoupled extension to later localized, coupled and asymmetric extension that results in thinning of the crust and exhumation of subcontinental mantle. The change in the mode of extension together with the localization of deformation reflects an evolution of the bulk rheology of the extending lithosphere. In this paper we summarize the pertinent geological observations for the Iberia–Newfoundland and Alpine Tethys margins. We describe the stratigraphic evolution, the fault geometry, basin architecture, and magmatic and metamophic evolution of the two pairs of margins from initial rifting to final continental breakup. This description forms a basis for understanding the evolution of the bulk rheology and how the various processes interact during progressive lithospheric extension. For the Iberia–Newfoundland and Alpine Tethys margins initial rifting appears to be controlled by inherited heterogeneities and mechanical localization processes, whereas final rifting and lithospheric rupture is controlled by serpentinization, magmatic and thermal weakening. At other margins, these modes may interact in a different way depending on the prerift conditions and the evolution of the rheology during rifting.

In the last decade, studies of rifted margins have benefited from an increasing quantity of high-quality data and significant developments in numerical modelling. The ability to use more realistic boundary conditions in numerical modelling not only enables us to investigate the complex interaction between processes controlling rifting but also leads to a demand for more and better datasets. Although in the last years a lot of high-quality geophysical data have become available from distal passive continental margins, direct observations on the subseismic scale from this domain are still the exception. One way forward is to take a comparative approach using direct observations and unlimited sampling on analogous structures preserved on land. Based on this approach, Manatschal & Bernoulli (1998, 1999), Boillot & Froitzheim (2001), Whitmarsh *et al.* (2001*a*) and Wilson *et al.* (2001) were able to demonstrate that marine-derived hypotheses of margin development proposed from drilling and seismic surveys can be supported and further developed using observations from the ancient margins exposed in the Alps.

In this paper we present pertinent observations from the Alpine Tethys (Fig. 1) and Iberia–Newfoundland margins (Fig. 2). The aim of this paper is to demonstrate how observations can help to constrain the tectonic and rheological evolution of these magma-poor rifted margins. More generally, this study attempts to discuss how extensional processes evolve during continental rifting and how they interact with sedimentary, hydrothermal and magmatic processes. More detailed observations and explanations, in particular concerning the Alpine examples, are given in Manatschal (2004). A more general description of the physical boundary conditions of the modes of extension discussed can be found in Lavier & Manatschal (2006).

In the first part of this paper, the tectonic and palaeogeographic evolution is discussed together with some ideas and models that evolved from the study of the Alpine Tethys and Iberia–Newfoundland margins. Then, a set of fundamental observations of lithospheric extension is presented, which includes geological mapping, structural/sedimentological and petrological observations, marine multichannel

From: KARNER, G. D., MANATSCHAL, G. & PINHEIRO, L. M. (eds) *Imaging, Mapping and Modelling Continental Lithosphere Extension and Breakup*. Geological Society, London, Special Publications, **282**, 291–324.
DOI: 10.1144/SP282.14 0305-8719/07/$15.00

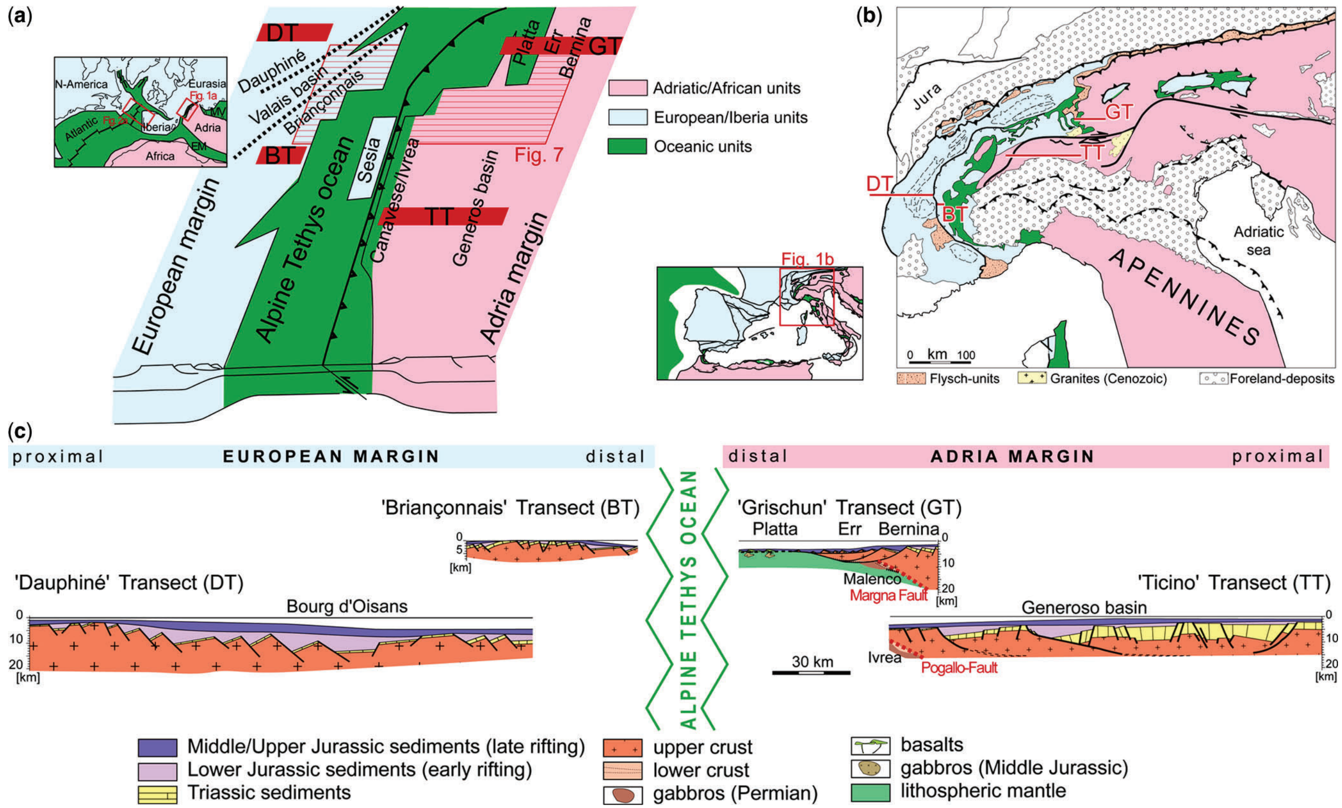

Fig. 1. (**a**) Palaeogeographic situation of the Alpine Tethys ocean and adjacent margins during Late Cretaceous time (modified after Dal Piaz 1995). Inset map shows the palaeogeographic relation between the Tethys and Atlantic oceanic basins during Late Cretaceous time. Abbreviations: EM, Eastern Mediterranean/Neo-Tethys; MV, Meliata–Vardar Ocean. (**b**) Tectonic map of the Alps (modified after Polino *et al.* 1990) showing the distribution of the major palaeogeographic units. Inset map shows the distribution of the palaeogeographic units in Western Europe. (**c**) Reconstructed palaeogeographic sections across the Alpine Tethys margins; the Grischun Transect modified after Manatschal & Bernoulli (1999); Briançonnais and Dauphine transects after Lemoine *et al.* (1987); and the Ticino Transect after Bertotti *et al.* (1993). For the present-day and Late Cretaceous position of the transects see (a) and (b).

(a)

(b)

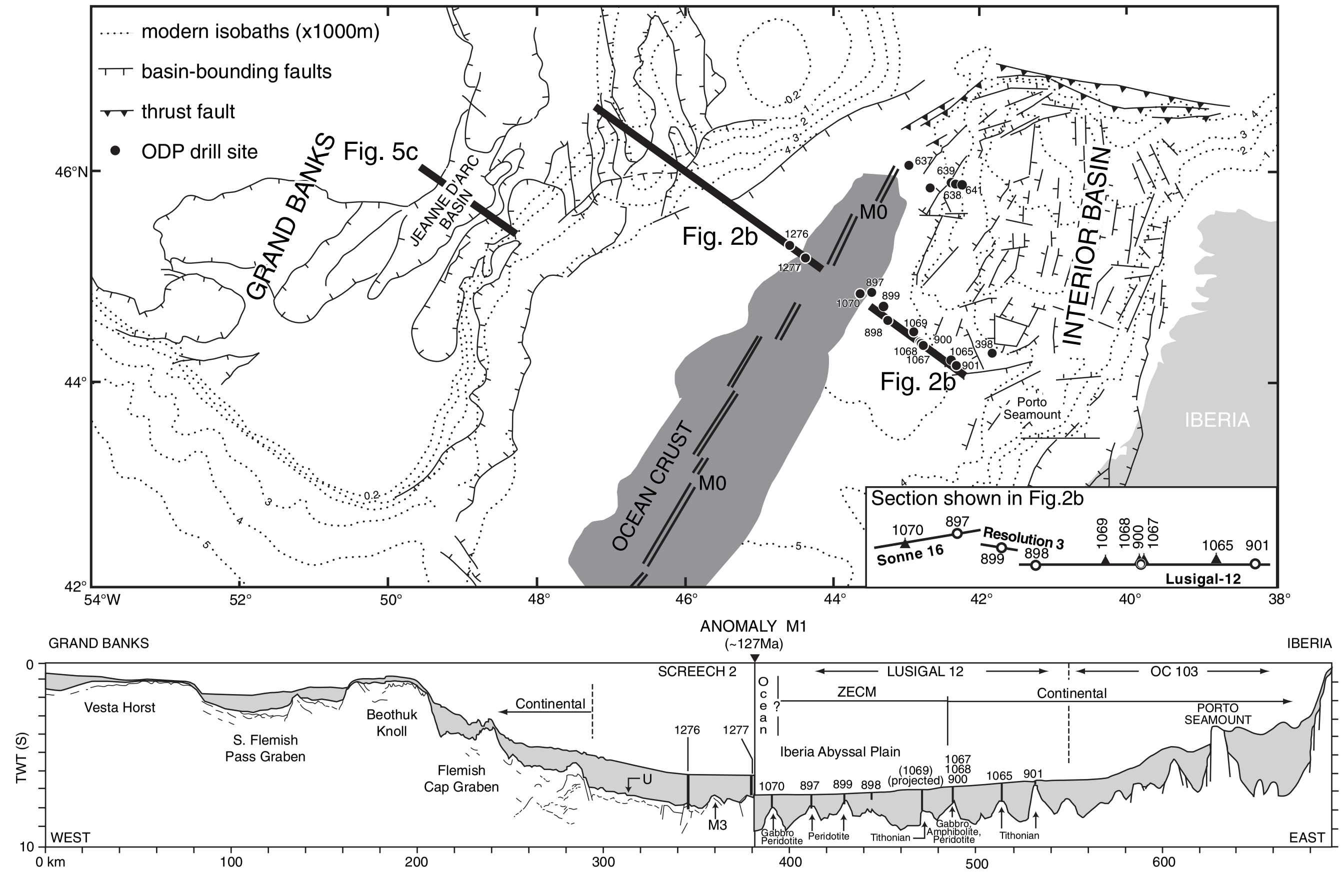

Fig. 2. (**a**) Map of the Cretaceous Atlantic between Iberia and Newfoundland reconstructed to anomaly M0 time (*c*. 125 Ma) based on the reconstruction pole of Srivastava *et al.* (2000) with the North American Plate fixed (modified after Shipboard Scientific Party 2004). (**b**) Conjugate interpreted seismic reflection profiles from the Newfoundland and Iberia margins, juxtaposed at approximately anomaly M1. Sediments are shaded grey above basement (Iberia) or basement and/or the U reflection (Newfoundland). Newfoundland margin: simplified interpretation of SCREECH 2 line with location of ODP sites 1276 and 1277; Iberia margin: composite seismic section (Sonne 16, JOIDES Resolution 3 and Lusigal 12) along the conjugate Iberia drilling transect (modified after Shipboard Scientific Party 2004).

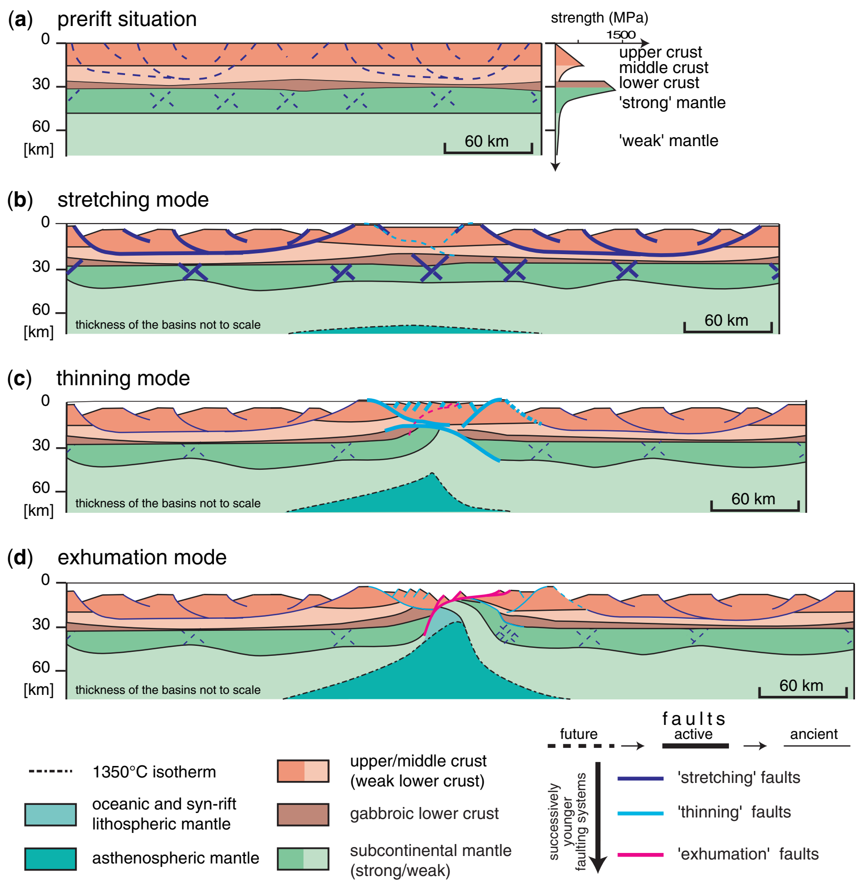

Fig. 3. Conceptual model showing the temporal and spatial evolution relative to a fixed left-hand edge from (**a**) the prerift situation, to (**b**) stretching mode to (**c**) thinning mode to (**d**) exhumation mode (modified after Lavier & Manatschal 2006). The single sections have been drawn so that the continental crust is balanced approximately. In reality, erosion has to be taken into account, but, given the scale of the model, it can be neglected and in any case during the later stages the continental crust was covered by deep water so no erosion occurred. The contact between the strong and the weak subcontinental mantle corresponds, at least during the initial stage of rifting, to a stable boundary. Therefore, the surface of strong subcontinental mantle is balanced in the four stages, although thermal erosion and serpentinization will affect and change the thickness of the 'strong' subcontinental mantle in a final stage of rifting. The top of the asthenosphere, as drawn in the model, corresponds to the 1350 °C isotherm. Consequently, asthenospheric mantle can, by cooling, change into lithospheric mantle. The major uncertainty in the model is related to the spatial and temporal evolution of the thermal structure within the distal margin during final breakup, which is at the moment not resolved by any observational data.

seismic reflection and refraction data, and drilling results. The final sections discuss extensional processes at magma-poor rifted margins, the associated exhumation and magmatic processes, and the rheological evolution of the extending lithosphere deduced from the observations presented in this paper. In order to simplify the description and discussion of the various rift structures representing different domains and stages within the two pairs of margins, all observed sites are located in the conceptual model presented in Figure 3.

From observations to a conceptual model

The Alpine Tethys and Iberia–Newfoundland margins: an overview

The Alpine Tethys and Iberia–Newfoundland margins result from rifting of a lithosphere previously affected by the Variscan orogeny (Fig. 4b). The Alpine Tethys margins evolved as part of an equatorial Late Triassic–Jurassic rifting system, which extended from the Caribbean to the eastern

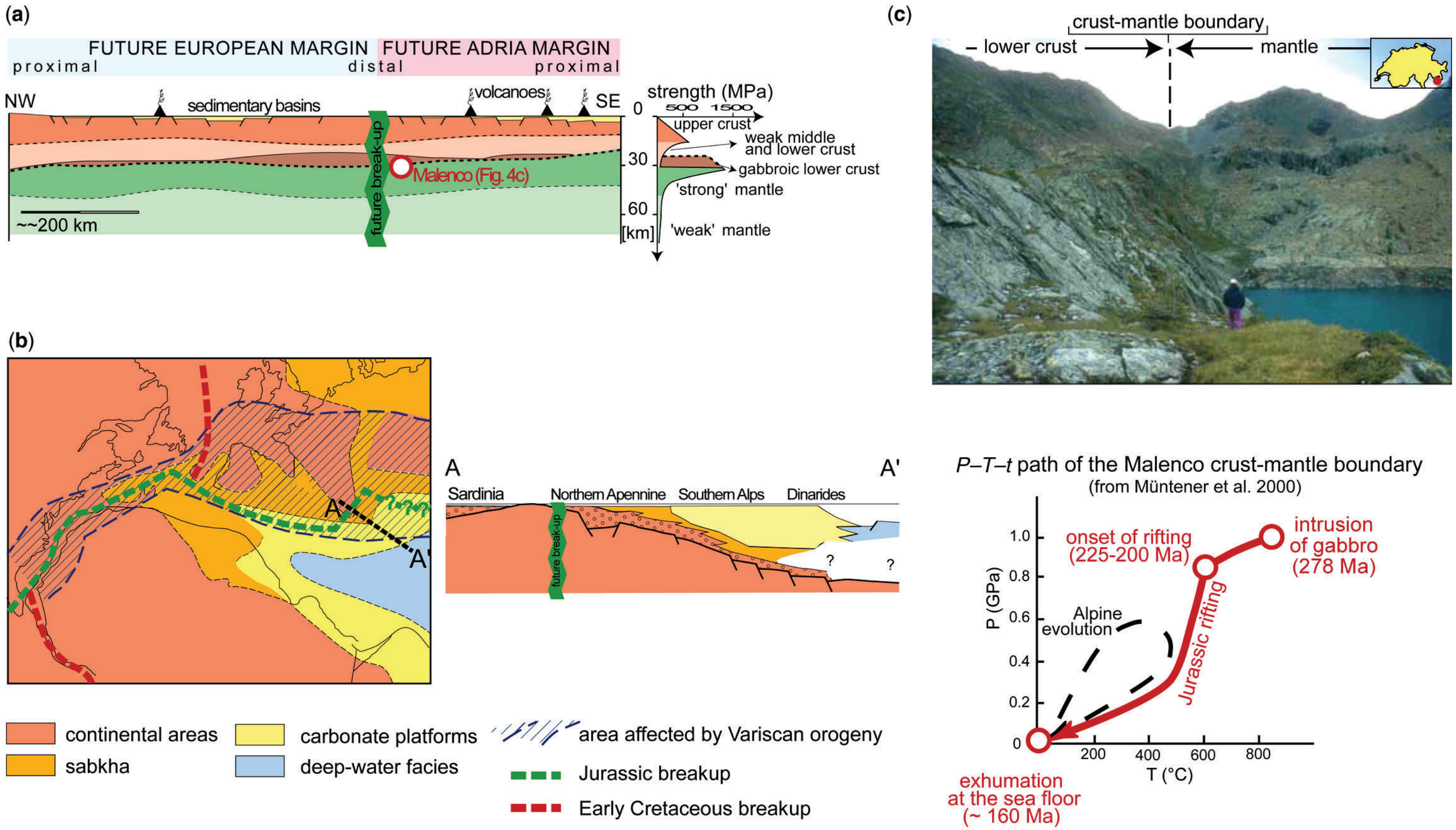

Fig. 4. The prerift conditions as constrained from the Alps. (**a**) Schematic cross-section across the area of future breakup showing the position of the Malenco crust–mantle boundary, the distribution of lithologies in the prerift lithosphere, the position of prerift basins and volcanic systems, and the general rheology at the onset of rifting. (**b**) Late Triassic restoration of the Mediterranean and central Atlantic realm, and a section through the area of future Jurassic breakup showing the prerift depositional architecture relative to the future Alpine Tethys Ocean (modified from Bosellini 1973 and Laubscher & Bernoulli 1977). (**c**) The Malenco crust–mantle boundary as exposed in the field and pressure–temperature–time (*P–T–t*) diagram for the Malenco crust–mantle boundary. The red arrow shows the exhumation path of Permian gabbros welding the crust–mantle boundary and the dashed black line the successive Alpine history (modified after Müntener & Hermann 2001).

Mediterranean area (Fig. 1a). The Iberia–Newfoundland margins resulted from Jurassic–Early Cretaceous rifting and opening of the Northern Atlantic (Fig. 2a). Thus, the different segments of the Atlantic–Tethys Ocean system opened at different times and the closure of the Alpine Tethys segment was contemporaneous with seafloor spreading in the Central and North Atlantic.

A wealth of data have been collected from the ancient Alpine Tethys and present-day Iberia–Newfoundland margins. These data have enabled a precise description to be made of the architecture and the temporal and spatial evolution of these two pairs of magma-poor rifted margins. At present, these two pairs of margins can be considered as the best-studied examples of so called magma-poor rifted margins. An overview of the Alpine margins including their reconstruction is given by Manatschal (2004), whereas a review of the work carried out over the past three decades at the Iberia margin can be found in Whitmarsh & Wallace (2001). The major results obtained from these margins are: the existence and mapping of exhumed mantle in the ocean–continent transition (OCT) (Decandia & Elter 1972; Boillot *et al.* 1987; Pickup *et al.* 1996; Chian *et al.* 1999; Dean *et al.* 2000); the discovery and description of detachment systems (Froitzheim & Eberli 1990; Reston *et al.* 1995; Krawczyck *et al.* 1996; Manatschal & Nievergelt 1997; Manatschal *et al.* 2001) and the changing basin architecture and sedimentary evolution across the margin (Lemoine *et al.* 1987; Froitzheim & Eberli 1990; Wilson *et al.* 2001).

Rift concepts derived from the Iberia–Newfoundland and Alpine Tethys margins

Observations or surveys can resolve either finite geometries at ancient margins or the spatial distribution of tectonic activity in present rifts. In order to understand the temporal and spatial evolution of the whole rifting process, data sampled across different scales have to be compared and examples of margins preserving different stages of the process have to be studied. Therefore, comparative studies played an important role in the understanding of rifted margins.

In the 1970s a comparative approach allowed sedimentary facies such as the upper Jurassic Radiolarian Cherts and Majolica Limestones to be traced from the Central Atlantic into the Alpine realm (Bernoulli & Jenkyns 1974). Subsequent studies then compared exposed rift basins or fault systems in the Alps with seismically imaged structures in proximal margins offshore (Graciansky *et al.* 1979; Bally *et al.* 1981). These studies resulted in a precise description of the structures in the proximal parts of the margins (e.g. Lemoine *et al.* 1986) that supported the pure shear model proposed by McKenzie (1978).

In the late 1980s the first drilling of the deep Iberia margin resulted in the unexpected discovery of exhumed mantle. Several modifications of the simple shear model initially proposed by Wernicke (1981) were proposed to explain mantle exhumation (see fig. 3 in Reston *et al.* 1996). Since then, ODP legs 149, 173 and 210 and geophysical surveys have enabled a much more complete description of the geophysical characteristics of the deep Iberia margin (Whitmarsh & Wallace 2001). Simultaneous with the study of the deep Iberia margin, mapping of detachment faults in the Alps (Froitzheim & Eberli 1990; Florineth & Froitzheim 1994; Manatschal & Nievergelt 1997) and detailed structural (Manatschal 1999); petrological (Müntener *et al.* 2000, 2004; Desmurs *et al.* 2002) and isotopic (Schaltegger *et al.* 2002) investigations have allowed the geological characteristics of the distal margins to be described. These studies have demonstrated the limitations of the previously proposed pure-shear and simple shear models (McKenzie 1978; Wernicke 1981) and lead to new and more evolved models. A comparison of the Alpine Tethys and Iberia–Newfoundland margins shows that, although they are of different age and ultimately had a different fate, they share many similarities. These similarities permit a comparison of direct observations from the ancient margins with the borehole and geophysical data from the present-day margins. Based on these data, a conceptual model to explain the tectonic evolution of the Iberia–Newfoundland and Alpine Tethys margins was proposed by Whitmarsh *et al.* (2001*a*). Modified versions of this conceptual model are presented by Manatschal (2004) and Lavier & Manatschal (2006). Although these models capture only the most general features of the rifting process, they are designed such that they respect all data derived from ODP legs 103, 149, 173 and 210 and all major Alpine observations. Based on time constraints from the stratigraphic record, isotope ages and magnetic anomalies, four stages of development can be described and will be referred to as the 'prerift situation', the 'stretching mode stage', the 'thinning mode stage' and the 'exhumation mode stage' (Fig. 3).

Fundamental observations of lithospheric extension

Observations constraining the conditions before onset of rifting (Fig. 3a)

Crustal thickness estimated from the prerift stratigraphic record. Observations constraining the prerift conditions can be obtained from the prerift stratigraphic record, which is poorly known for the Iberia–Newfoundland margins, but excellently preserved and well described for the Alpine Tethys system (Fig. 4). The prerift stratigraphic record in the Alpine Tethys domain is characterized, apart from local exceptions, by a general evolution from continental siliciclastic–sabkha and shallow-marine carbonate dominated depositional systems (Fig. 4b). Carboniferous–Permian clastic sediments and volcanic rocks form the base to the succession. They are commonly interpreted to be deposited in fault-bounded basins forming during post-orogenic extensional collapse of the Variscan orogen (Handy & Zingg 1991; Handy *et al.* 1999). Triassic depositional environments changed across the future Alpine realm from deeper marine environments to the SE across platform carbonates and sabkha environments into continental siliciclastics systems to the NW (Fig. 4b). The sedimentary sequence is generally thicker in the east and south (Central Austroalpine and Southern Alps; 1–5 km), thins towards the NW (Lower Austroalpine and Briançonnais; less than 500 m) and is locally only a few tens of metres in the future proximal European margin. The thicker Triassic sequences in the eastern and SE parts of the Alpine realm can be explained by its proximity to the Triassic Meliata ocean and the thermal subsidence following the breakup of this ocean (Channell & Kozur 1997; Dercourt *et al.* 1986; Robertson 2004). For the more western parts, forming the future distal margins of the Alpine Tethys, the stratigraphic record suggests that the prerift continental crust was in isostatic equilibrium after Permian time (Müntener *et al.* 2000). Therefore, at the onset of rifting the crustal thickness in the future distal margin is estimated to be in the order of about 30 km (Fig. 4a).

Prerift Moho temperature and composition of the lower crust and upper mantle. Constraints on the Moho temperature and composition of the lower crust and upper mantle before onset of rifting can be obtained from an exhumed prerift fossil crust–mantle boundary exposed in the Malenco unit in northern Italy (Müntener & Hermann 1996; Hermann *et al.* 1997) (Fig. 4c). This crust–mantle boundary is welded by an Early Permian gabbro, which intruded into lower crustal rocks formed by metapelites and metacarbonates. Granulite facies metamorphism lasted for about 20 Ma (Hermann & Rubatto 2003). Müntener *et al.* (2000) demonstrated that after the intrusion of the gabbro in Permian time, the mantle and lower crustal rocks cooled more or less isobarically over 50 Ma. They calculated the temperature and pressure conditions at the crust–mantle boundary for the onset of rifting to be at about 550 °C at 0.9–1 GPa, corresponding to a 33–36 km-thick crust (Fig. 4a). Based on these data, a rheological model with a brittle upper crust, a ductile middle crust, a gabbroic strong lower crust, a brittle upper mantle and a plastic lower mantle can be proposed for the prerift lithosphere (Fig. 4a). Evidence for a widespread lateral continuity of the underplated gabbros beneath the future proximal margins is given by the common occurrence of Permian gabbro xenoliths in Tertiary volcanic rocks in Western Europe (Féménias *et al.* 2003). Pre-existing Variscan and Permian structures, local prerift volcanic activity and underplated gabbros along the crust–mantle boundary resulted in rheological heterogeneities (Handy 1989), which, as discussed below, may have strongly controlled the distribution of rift structures at the onset of lithospheric extension.

Observations documenting early rifting (Fig. 3b)

Stratigraphic record of early rifting. The age of onset of rifting is diffuse and poorly constrained at both the Iberia–Newfoundland and Alpine Tethys margins. In the Iberia–Newfoundland margins interpretations of the onset of rifting are based mainly on geometrical arguments, i.e. determination of geometry related to synsedimentary normal faulting such as onlap onto tilting hanging walls and thickening into footwalls (Driscoll *et al.* 1995; Wilson *et al.* 2001). In the Alps, the age of onset of rifting is based on sedimentological, stratigraphic and structural arguments (Eberli 1988; Bertotti *et al.* 1993; Wilson *et al.* 2001). The onset of rifting is marked by differential subsidence and associated faulting during Late Triassic time. However, because sedimentation kept pace with differential subsidence, the prerift platform conditions persisted during initial rifting. Also, the most proximal parts of the Adriatic margin may have been affected by post-rift subsidence related to the opening of the Vardar–Meliata Ocean in Triassic time (Fig. 4a). This renders identification of the onset of rifting at the Alpine Tethys margins difficult. Larger rift basins, which are related unambiguously to the rifting leading to the Alpine Tethys Ocean, formed only later, during Early Jurassic time (Hettangian–Sinemurian). Basins such as the Il Motto basin (Fig. 5b) and the Generoso Basin (Fig. 6b) formed simultaneously with large basins on the conjugate proximal European margin (e.g. Bourg d'Oisans Basin: Chevalier *et al.* 2003) (Fig. 5a). Rifting in the proximal parts of the future margin ceased in the middle–late Early Jurassic, but some of the faults became inactive even earlier and were sealed by Upper Sinemurian sediments (e.g. Il Motto: Eberli 1988; Conti *et al.* 1994) (Fig. 5b). Later rifting was focused, as discussed below, on the future distal margin.

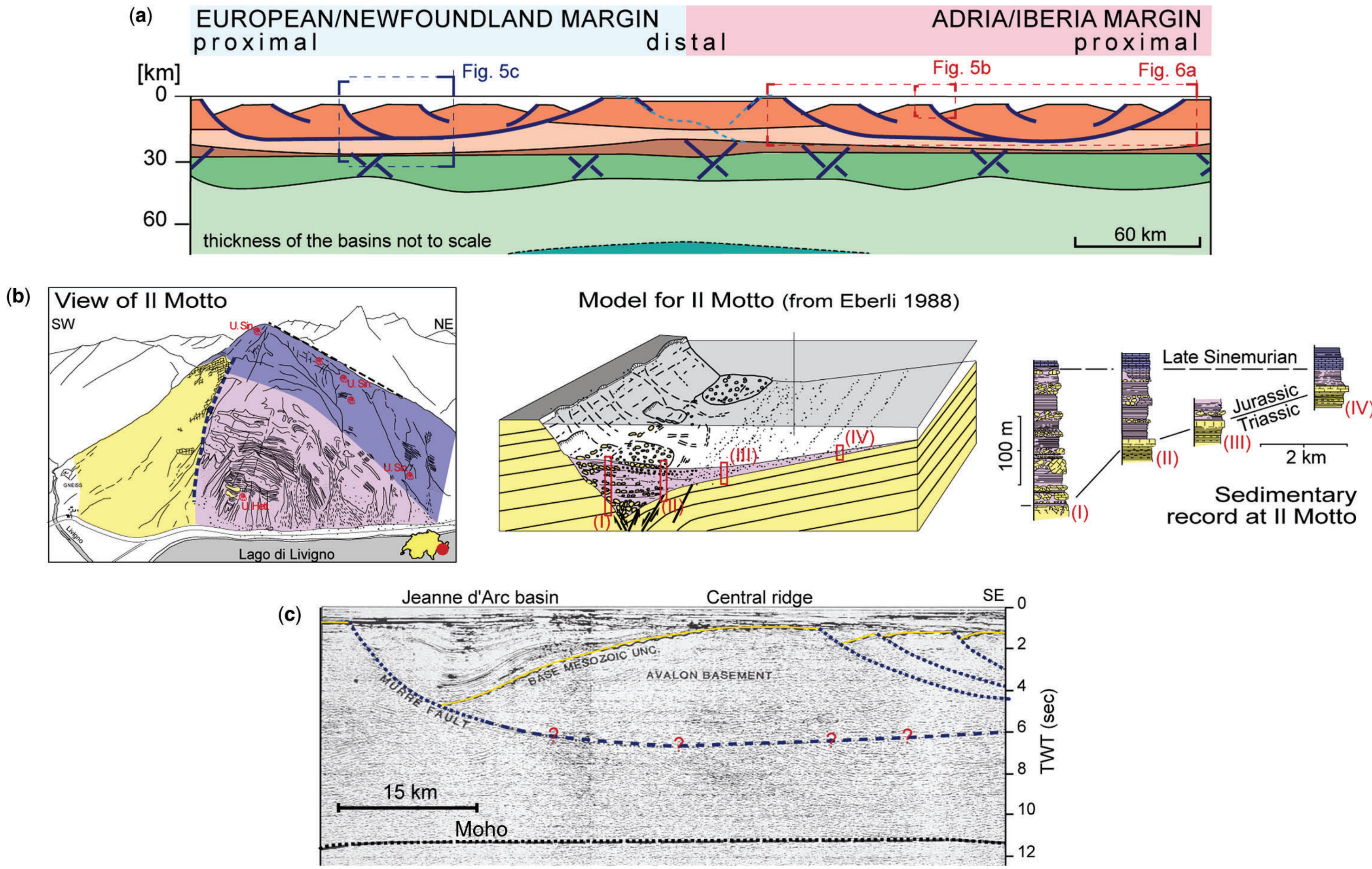

Fig. 5. (**a**) Reconstructed section across the future margins during initial rifting and locations of examples of early rift structures in the proximal parts of rifted margins. (**b**) View of Il Motto showing the break-away of a high-angle normal fault that is sealed by Upper Sinemurian sediments (to the left) and a model for the Il Motto Basin (middle part) based on sedimentary sections across the basin (to the right) (modified after Eberli 1988). For location of the Il Motto Basin see (a). (**c**) Reflection seismic section across the Jeanne d'Arc Basin (from Tankard *et al.* 1989). For location see Figure 2a.

Early rift structures preserved in the future proximal margins. The most prominent rift structures in the proximal margins are fault-bounded rift basins. Such basins have been imaged in seismic reflection profiles from the proximal Iberia margin (Montenat *et al.* 1988; Murillas *et al.* 1990) and drilled on the Newfoundland margin (Keen *et al.* 1987; Driscoll *et al.* 1995). A typical example is the Jeanne d'Arc Basin (Fig. 5c), which is bounded by a major fault that is well imaged in the upper crust but disappears at middle–lower crustal levels and does not affect the Moho, which is regionally up-warped beneath the extended basin (Keen *et al.* 1987).

Examples of fault-bounded basins are spectacularly exposed in the area of Bourg-d'Oisans near Grenoble in the external French Alps (Lemoine *et al.* 1986; Chevalier *et al.* 2003) and at Il Motto in the Ortler nappe along the Swiss–Italian border in the eastern Alps (Conti *et al.* 1994) (Fig. 5b). The most complete section preserving rift structures of the former proximal Adriatic margin is exposed in the Lombardian and southern Swiss Alps (Bertotti *et al.* 1993) (Fig. 6a). In this section, the basin architecture is exposed over more than 200 km from the Trento plateau to the east to the Canavese to the west. Bernoulli (1964) demonstrated, based on mapping of abrupt changes in facies and thickness of the Early Jurassic formations across the Lugano–Val Grande fault, the existence of an Early Jurassic rift basin, about 30 km wide and filled with several thousand metres of synrift sediments (e.g. Generoso Basin) (Fig. 6b). Bertotti (1991) mapped the Lugano–Val Grande fault over a horizontal distance of more than 30 km. He demonstrated that this fault cuts across prerift sediments into basement, changing from a localized brittle fault into a ductile mylonitic shear zone (Fig. 6b). The reconstructed geometrical relationship between fault zone and bedding in the overlying synrift sediments led Bertotti (1991) to conclude that the fault had a listric geometry. Based on microstructural investigations, he demonstrated that the fault soled out in greenschist facies conditions, i.e. at about 300–350 °C and about 10–12 km depth, assuming a normal temperature gradient of 30 °C km^{-1}.

The synrift sediments in the rift basins in the proximal margin are characterized by abrupt changes in facies and show significant thickness variations. Basin infill is dominated by turbidites interbedded with hemipelagic sediments. Mass-flow deposits, such as debris flows or rock falls derived from exposed normal faults, are observed only in a few basins. A particularly instructive section preserving the relationships between the sediment infill and the master fault bounding the basin is exposed at Il Motto (Fig. 5b). The synrift sedimentary sequence is characterized by fining- and thinning-upwards in the basin and, away from the fault zone, an inversion of the clast stratigraphy in the basin owing to erosion and redeposition of the footwall block (Eberli 1988). Erosion of the upper Triassic sediments in the footwall block and their redeposition in the basin suggests subaerial exposure and erosion of the tip of the tilted footwall block. In the Generoso Basin and the Bourg-d'Oisans the highs remained submarine. In both places a change from shallow to deeper marine sedimentation has been documented from the footwall near the break-away of the fault. This observation indicates that both hanging wall and footwall were subsiding, but at different rates relative to sea level, and that rift-shoulder uplift was less than the general subsidence of the extending margin.

Thermal and magmatic evolution during early rifting. Constraints on the thermal evolution of the margin during initial rifting can be obtained from cooling ages of basement rocks. Published cooling ages obtained from the footwall and hanging wall of the Lugano–Val Grande fault range from 220 to 180 Ma (Sanders *et al.* 1996) (Fig. 6c). However, cooling ages cannot be related to exhumation and uplift along the Lugano–Val Grande fault because there was marine sedimentation onto the footwall. Bertotti & Ter Voorde (1994) and Bertotti *et al.* (1999) interpreted these cooling ages to be related to thermal equilibration of a Ladinian magmatic event described from the eastern southern Alps and the Austroalpine realm (Lu *et al.* 1997). Their interpretation ignores the fact that Late Triassic–Early Jurassic cooling ages are widespread in the Alpine realm, suggesting that rifting in the proximal margins was associated with cooling of the extending lithosphere.

Volcanic activity was widespread during the Triassic across the future European and Adriatic margins. This is well established by tuff horizons and local dolerites within the Triassic sedimentary sequences. However, evidence for volcanic activity during Late Triassic–late Middle Jurassic rifting is not recorded. This observation suggests that rifting related to the opening of the Alpine Tethys ocean can be considered as non-volcanic.

Rates of extension during early rifting. The problem of objectively defining the age of onset of rifting generates big error bars on the determination of the duration of rifting and consequently the rates of extension. For the West Iberia margin, rift duration is poorly constrained and commonly overestimated (Wilson *et al.* 2001). Observations show that rifting is polyphase and that strain rates are difficult to obtain. In the best case, extension rates can be determined for parts of the margins

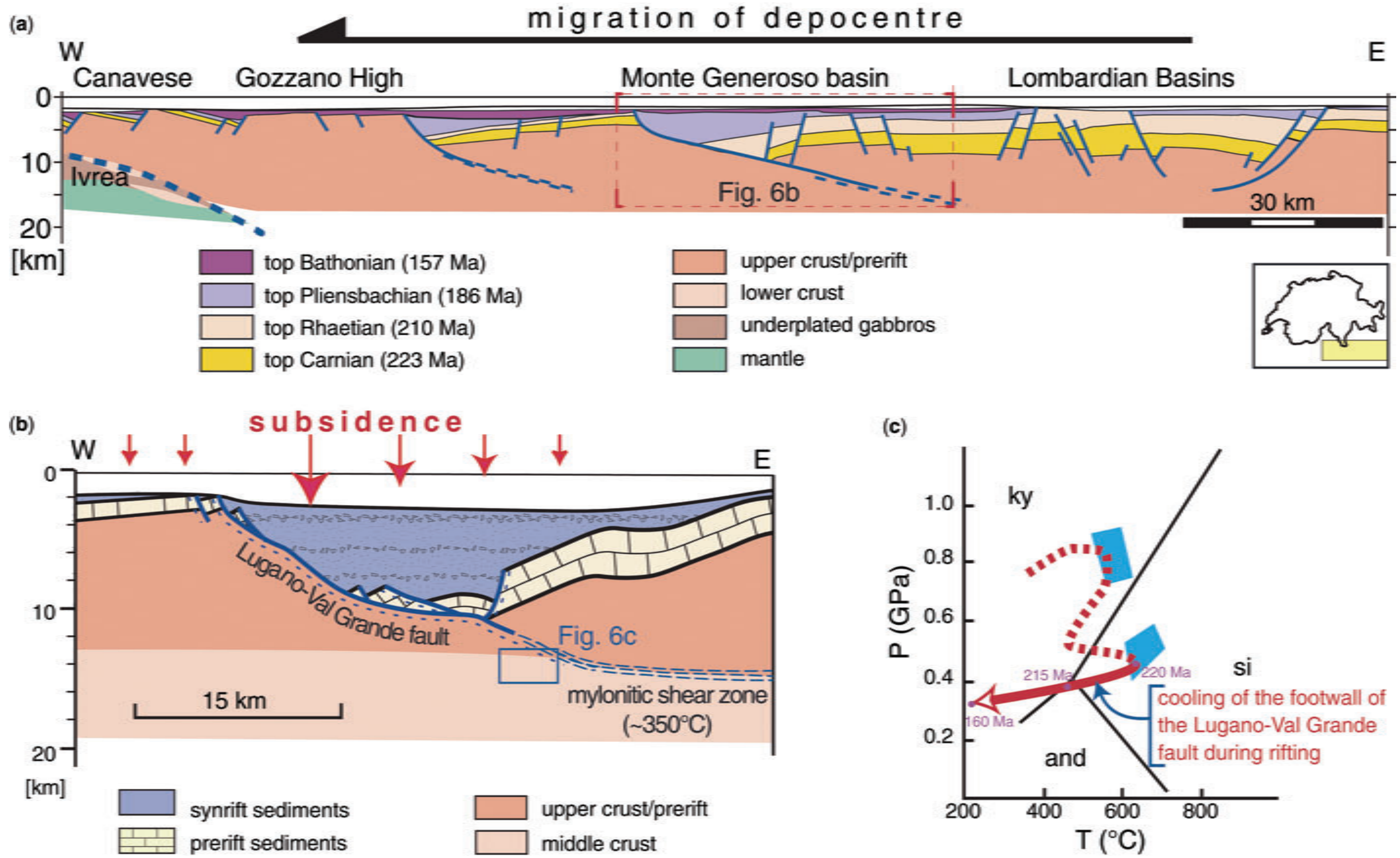

Fig. 6. (**a**) Basin architecture and depocentre migration in the Ticino Transect (for location see Fig. 1) (modified from Bertotti *et al.* 1993). (**b**) Palinspastic reconstruction of the Monte Generoso Basin and its relation to the Lugano–Val Grande fault bounding the basin (modified after Bertotti 1991). (**c**) P–T–t evolution from the rocks sampled in the vicinity of the Lugano–Val Grande fault (Sanders *et al.* 1996).

or single events. Bertotti *et al.* (1993) defined for the Ticino Transect average velocities of rifting ranging between 0.46 and 1.28 mm year^{-1}. Strain rates were also determined for two faults bounding rift basins in the proximal European margin (i.e. Ornon fault bounding the Bourg d'Oisans Basin: Chevalier *et al.* 2003; and the Uzer fault: Dromart *et al.* 1998). Chevalier *et al.* (2003) showed that the motion along a single fault was discontinuous. Low rates of displacement (0.2–0.4 mm year^{-1}) were accommodated by diffuse extensional deformation, whereas high rates (up to 1.8 mm year^{-1}) were accommodated along the localized master fault bounding the basin. These authors showed that such high rates were observed to occur over very short time spans (one ammonite zone; less than 0.5 Ma). These studies show that most of the extension has been accommodated during short pulses and that activity is not simultaneous along different fault segments in the margin. This behaviour means that strain rates for the whole margin are difficult to obtain. Available data suggest that overall the extension associated with rifting appears to be slow ($<$10 mm year^{-1}), discontinuous and distributed over tens of millions of years (Bertotti *et al.* 1993; Chevalier *et al.* 2003). In contrast, slip rates along single faults were fast, but occurred only over a short time span. These rather slow rates may explain also the cooling ages obtained from the basement underlying the basins in the Ticino transect (Fig. 6c).

Observations documenting advanced rifting (Fig. 3c)

Stratigraphic record and early rift structures preserved in the future distal margins. In contrast to proximal margins, distal margins are more difficult to access. Present-day distal margins are at abyssal depth and covered by thick sedimentary units, and fossil examples preserved in collisional orogens are either subducted or strongly overprinted during collision. Therefore the early rift evolution of distal margins is so far only constrained from a few places. The available observations from the Alps show the migration of depocentres towards the area of future breakup, indicating a diachronous evolution of the proximal and distal margins as well as a successive localization of the deformation in the distal margin during early rifting (Lemoine *et al.* 1986; Eberli 1988; Bertotti *et al.* 1993; Manatschal & Bernoulli 1998). A similar evolution is also observed in the West Iberia margin, where rifting in the Interior Basin predates rifting in the Iberia Abyssal Plain (Wilson *et al.* 2001). A reconstruction of the distal Alpine Tethys margins during Toarcian time (*c*. 175 Ma), about 10 Ma before final breakup, is shown in Figure 7. In this reconstruction the distal parts of the future European margin are represented by the Briançonnais domain (see Briançonnais Transect in Fig. 1c). Parts of this domain escaped subduction and were obducted onto more proximal parts of the European margin (Fig. 1b). Distal parts of the Adriatic margin are represented by the Err–Bernina domain in the north and the Canavese–Gozzano domain further to the south (see the Grischun and Ticino transects in Fig. 1c). These units were in the hanging wall of the Tertiary subduction and therefore were overprinted only weakly during Alpine collision. The observed rift structures in the future distal margin are in many respects different from that described previously from the proximal margins. The diachronous evolution of the proximal and distal margins is best recorded by the observation that the early Middle Jurassic fault-bounded basins in the future distal margins formed contemporaneously with post-rift hemipelagic sequences in the proximal margins (see the Grischun Transect in Fig. 1c). Subsidence is indicated by the transition from platform to hemipelagic cherty limestones, which are contemporaneous with the synrift sediments in the proximal basins. In contrast to the proximal margin these sediments show a constant thickness of less than 100 m across the distal margin, clearly indicating that high-angle faulting and related block tilting were subdued in this part of the margin. These lower Jurassic, hemipelagic carbonates – the Agnelli Limestone of Furrer *et al.* (1985) – are overlain by unsorted, polymictic breccias with variable amounts of a reddish matrix and clasts predominantly derived from the basement. These breccias show a transition into siliciclastic turbidites, interbedded with hemipelagic marls, 200 and 450 m thick (e.g. Saluver Formation of Finger *et al.* 1982). These sediments are the first direct evidence for significant tectonic activity within the distal margin. The common occurrence of clasts of alkali granite, characteristic of the Bernina nappe (Spillmann & Büchi 1993), together with the arrival of sandstones, indicates uplift and subaerial exposure of parts of the Bernina domain contemporaneous with down-faulting of the Err domain. The age of this event is weakly constrained; it is younger than the hemipelagic sediments dated as Pliensbachian and older than the overlying upper Middle Jurassic Radiolarite Formation (Baumgartner 1987). Uplift of the Bernina domain is, however, indirectly dated as Toarcian (about 180 Ma) by the occurrence of gravity flows within ammonite-bearing hemipelagic limestones in the distal part of the proximal margin exposed in the Ela–Bernina nappe (Eberli 1988) (Fig. 7).

In the Briançonnais domain uplift and subaerial exposure is recorded by a prominent erosional

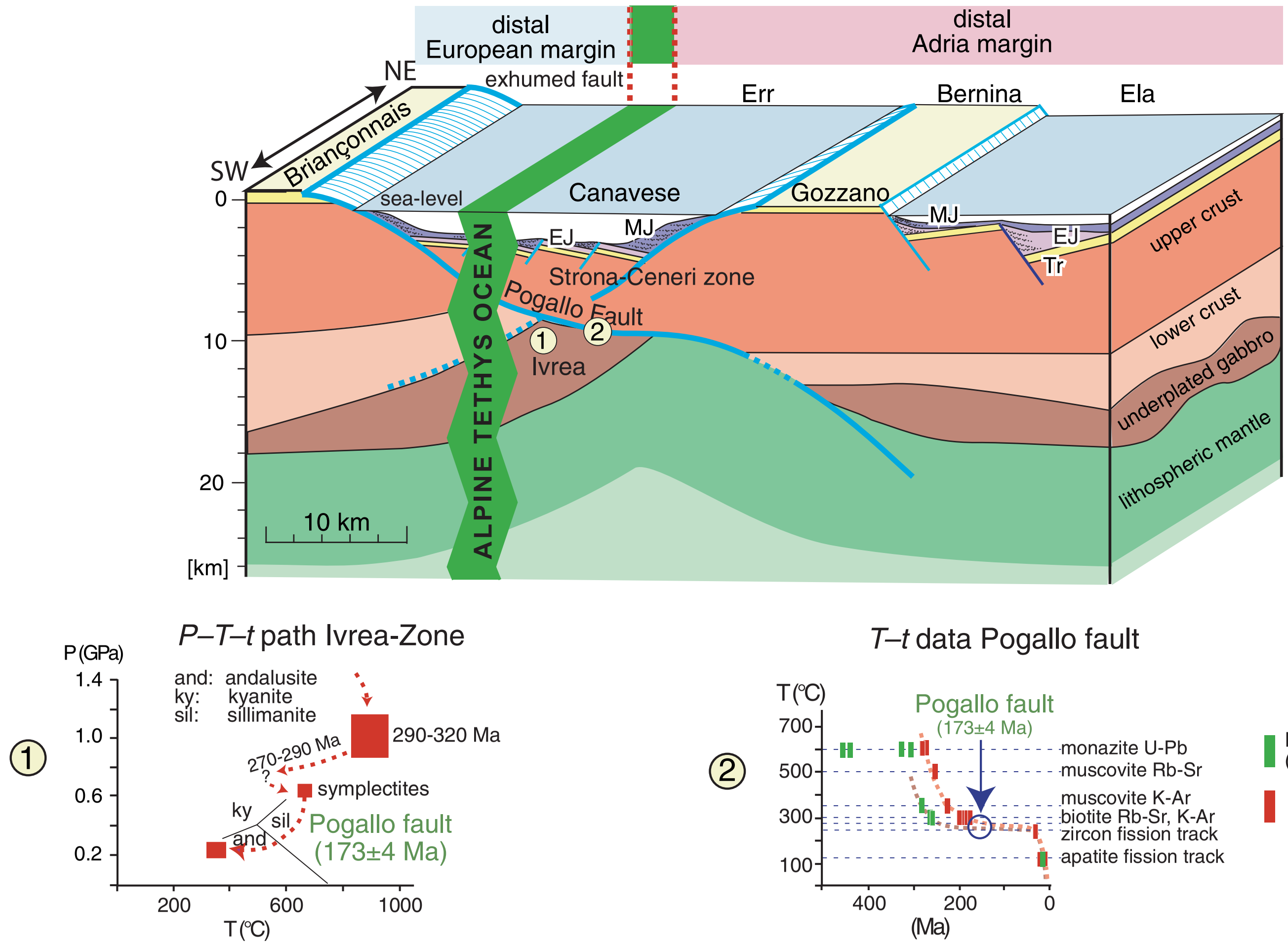

Fig. 7. Reconstructed section across the future distal margin during an early stage of rifting (Toarcian; 180 Ma) juxtaposing the Briançonnais domain and the Ivrea–Strona-Ceneri (Ticino Transect; Fig. 1) and Platta–Err–Bernina domains (Grischun Transect; Fig. 1). The section combines domains that belong to conjugate margins and had a different Alpine tectonic evolution. The P–T–t evolution of the Ivrea Zone and the T–t history of the Pogallo fault are presented below (see Handy *et al.* 1999). For more discussion see text.

surface and a karst system within the Triassic and Lower Jurassic shallow-water carbonates that was dated by Lemoine *et al.* (1986) and Claudel & Dumont (1999) as late Early to Middle Jurassic in age.

Constraints for crustal thinning during early rifting in the distal margin are obtained from the lower crustal rocks exposed in the Ivrea zone and the Malenco area in northern Italy (Figs 1c & 7). These lower crustal rocks are separated from upper crustal rocks along the Pogallo fault in the Ivrea zone (Handy 1987; Handy & Zingg 1991) (see Fig. 7) and the Margna fault in the Malenco area (Müntener & Hermann 2001) (see Fig. 1c). In reconstructed sections, these faults dip beneath the distal Adriatic margin (Handy 1987; Müntener & Hermann 2001). Cooling ages from the footwall of the Pogallo fault (Handy & Zingg 1991) and the Margna fault (Villa *et al.* 2000), together with thermo/barometric data (Müntener & Hermann 2001 and references therein), clearly show exhumation and major thinning of the crust to about 0.4–0.5 GPa (i.e. about 15 km) during initial rifting in Early Jurassic time. U–Pb dating on titanite from synmylonitic dykes from the Pogallo fault gave an age of 173 ± 4 Ma (Toarcian–Allenian), interpreted to result from a discontinuous retrograde metamorphic reaction associated with the crustal thinning that was localized along the Pogallo fault (Mulch *et al.* 2002).

These observations show that thinning of the crust along major mylonitic fault zones in the distal margin (e.g. Pogallo fault; Fig. 7) post-dates Sinemurian–Pliensbachian (200–184 Ma) sealing of fault-bounded basins in the proximal margins (e.g. Il Motto; Fig. 5b), and coincides with migration of the depocentre into the future distal margin (e.g. Ticino Transect; Fig. 6a), and uplift of the Briançonnais and Bernina domains (e.g. Fig. 7). We infer, based on these data, that there is a direct relationship between deep lithospheric thinning and migration of the depocentre into the distal margin. Lemoine *et al.* (1987) proposed that the contrasting behavior of the two future distal margins was directly related to breakup. In order to explain the uplift of the Briançonnais margin, they proposed a simple shear model for continental breakup with a detachment fault dipping beneath the European margin. Our observations suggest that uplift of the Briançonnais and Bernina domains is more probably related to the thinning of the continental crust along Pogallo type faults (see below for further discussion), which predate the final detachment faults exhuming the mantle (see later discussion) by 10–15 Ma. Based on these observations and as discussed below, we believe that the Pogallo and Margna faults are very important structures that explain thinning of the crust in the future distal margin predating final mantle exhumation.

Observations documenting final stage of rifting, breakup and transition to sea-floor spreading

The sedimentary record of final rifting in the Alpine Tethys margins shows that: (1) synrift sediments become younger towards the place of future breakup (Froitzheim & Eberli 1990; Bertotti *et al.* 1993) (Fig. 6a); (2) final rift basins are associated with downward concave faults leading to a very different architecture of the synrift sediments (Wilson *et al.* 2001); and (3) the isostatic evolution of the most distal margins is asymmetric (Lemoine *et al.* 1987). The most prominent structures related to final rifting are detachment faults, which are described and mapped in the Alps (Froitzheim & Eberli 1990; Florineth & Froitzheim 1994; Manatschal & Nievergelt 1997; Manatschal 2004) and seismically imaged as low-angle reflections off Iberia (de Charpal *et al.* 1978; Boillot *et al.* 1987; Reston *et al.* 1995; Krawczyk *et al.* 1996) (Fig. 8). These structures are interpreted to play an important role in the exhumation of subcontinental mantle during final rifting, leading to the formation of regions that show the characteristics neither of 'true' oceanic crust nor of thinned continental crust. Whitmarsh *et al.* (2001*a*) introduced the term 'zone of exhumed continental mantle' (ZECM) for such regions. In this paper the term 'ZECM' is used only for the part of the margin that is floored by subcontinental mantle lithosphere. In contrast, the term 'OCT' is used in a more general way for the transition from the distal continental margin to the oldest oceanic crust and implies neither a particular genetic evolution nor a specific composition of the underlying lithosphere.

We summarize the major observations related to final rifting and breakup of the lithosphere using two examples. The first is the Err–Platta OCT exposed in the Alps (Fig. 8b), and the second is the southern Iberia Abyssal Plain (Fig. 8c). More detailed descriptions of these two sites are given in Manatschal (2004 and references therein).

Err-Platta OCT

- *Detachment system in the Err–Platta OCT.* Remnants of a detachment system preserving primary relations between detachment faults, exhumed mantle rocks, magmatic rocks, and syn- and post-rift sediments are exposed in the Err–Platta nappes in SE Switzerland (Manatschal & Nievergelt 1997) (Fig. 9). For details of the reconstruction of the overall detachment system see Manatschal & Nievergelt

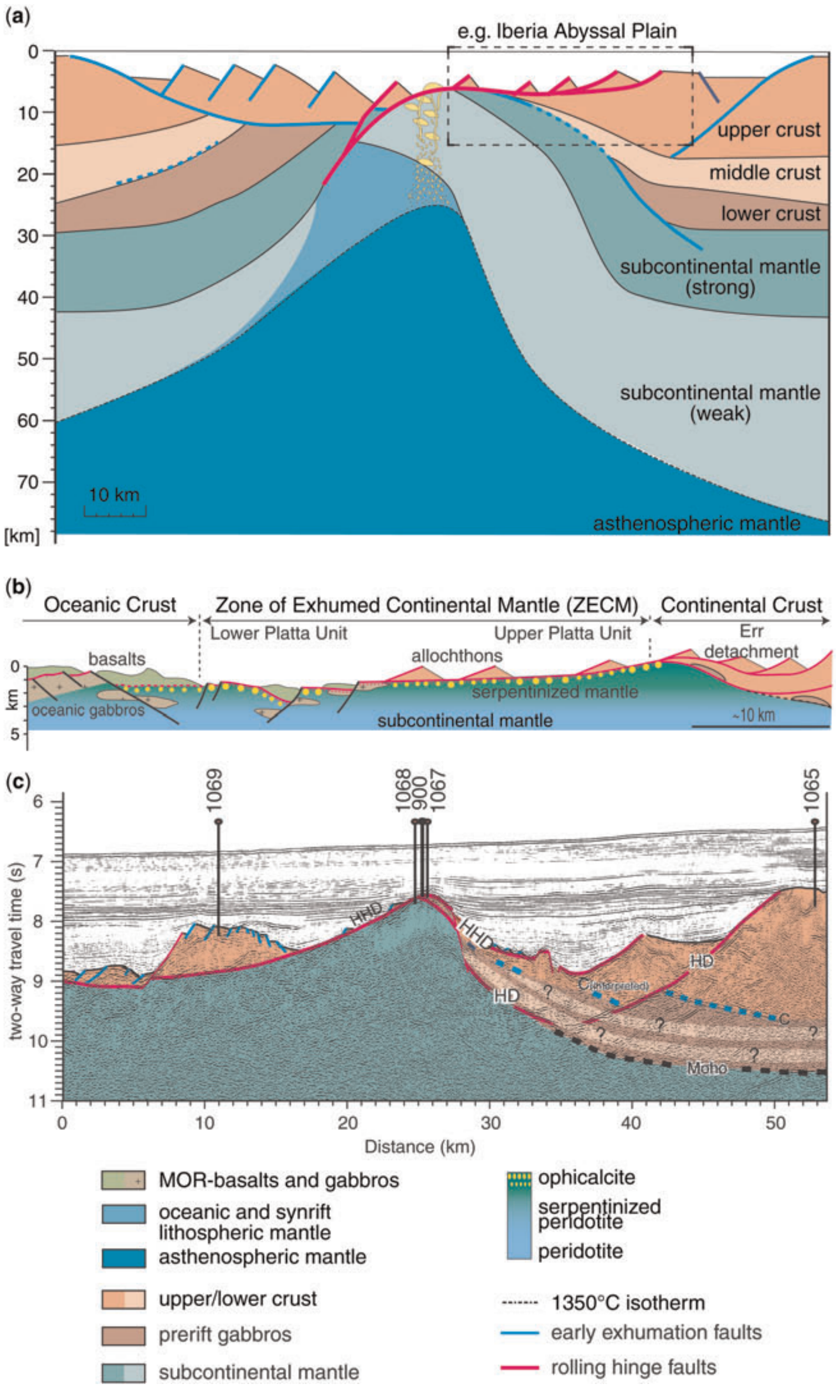

Fig. 8. (**a**) Generalized section showing the situation during final rifting based on observations from the Alpine Tethys and Iberia–Newfoundland margins. (**b**) Tentative palinspastic section showing the observed relationships between the detachment system and associated sediments, exhumed mantle rocks, and intrusive and extrusive magmatic rocks in the Err–Platta ocean–continent transition. (**c**) Geological interpretation of the depth-migrated Lusigal 12 profile (Krawczyk *et al.* 1996) showing the distribution of upper and lower crustal rocks, exhumed subcontinental mantle rocks and reflections interpreted as detachment faults. Numbers on top of the section refer to ODP sites.

(1997) and Desmurs *et al.* (2001). Although the actual transition from continental rocks to exhumed mantle rocks is not preserved in one continuous outcrop, the Err–Platta OCT represents at present the best-documented and most complete section across an OCT of a magma-poor rifted margin.

The detachment system reconstructed in the Err–Platta OCT forms one or several breakaways in the continental crust and cuts oceanwards from upper crust directly into subcontinental mantle. The relations between the detachment system and its hanging wall are spectacularly exposed over 6 km in the area of Piz d'Err–Piz Bial in the Err nappe (for more details see Froitzheim & Eberli 1990; Manatschal & Nievergelt 1997). The hanging wall is formed by fault blocks composed of continental basement and prerift sediments, which are tilted, toward the east (continent) and the west (ocean). The fault blocks, a few kilometres to some hundreds of metres across, are truncated systematically at their base by a detachment fault (the Err detachment; Manatschal & Nievergelt 1997) (Fig. 9). Similar blocks, interpreted as extensional allochthons, also overlie mantle rocks in the Parsettens area in the Platta nappe (Fig. 9). Locally, the detachment faults are directly overlain by tectono-sedimentary breccias, pillow basalts (Fig. 9) or post-rift sediments. Mapping of the detachment faults in the Err nappe shows that these structures are corrugated and composed of several fault zones. The geometrical relationships between the different fault zones are described in Manatschal & Nievergelt (1997).

- *Deformation structures associated with the detachment system.* In the continental crust, profiles across the footwall of the detachment faults exhibit a transition from undeformed granite into green cataclasites that are capped along a sharp and well-defined horizon by a characteristic black fault gouge (Manatschal 1999) (Fig. 9). The lack of any crystal plastic deformation in the quartz-rich basement rocks indicates that these faults were active under low-temperature conditions ($<300\,^{\circ}$C) within the uppermost 10 km in the crust. Based on geochemical, mineralogical and structural investigations, Manatschal (1999) demonstrated the importance of fluid- and reaction-assisted deformation processes for the localization of extension in the brittle upper crust. Of particular importance is the observation that mantle-derived elements such as chromium (Cr) and nickel (Ni) are enriched in the fault zone in the continental crust. Manatschal *et al.* (2000) explained this enrichment by the percolation of fluids with high concentrations in Ni and Cr that were derived from a serpentinizing mantle (Frueh-Green *et al.* 2002). This interpretation leads to the suggestion that serpentinization of the mantle started beneath an extending continental crust, and that serpentinization is intimately related to extensional deformation. This is compatible with the results obtained from the stable isotope investigations of Skelton & Valley (2000) and the modelling results of the rheological evolution during continental breakup by Pérez-Gussinyé & Reston (2001).

In the ZECM, a fault zone with a constant top-to-the-ocean sense of shear caps the exhumed mantle. This fault zone is formed by foliated serpentinite, serpentinite cataclasite and serpentinite gouges, recording a complex deformation and hydration history under lower amphibolite facies to sea-floor conditions. High-temperature shear zones, formed by peridotite mylonites–ultramylonites, are commonly oblique to the fault zone capping the exhumed mantle, and show a top-to-the-continent sense of shear. The relationship between the 'cold' fault zone and the 'high-temperature' mylonite shear zones is exposed in an outcrop at Sur al Cant in the Platta nappe (Fig. 9). In this outcrop, the high-temperature top-to-the-continent peridotite mylonites are cut by a top-to-the-ocean fault zone, which also cuts rodingitized basaltic dykes. This is a clear evidence that magmatic activity predates or was coeval with final movement along the top-to-the-ocean fault zone. This observation is fundamental, because it demonstrates that late, 'cold' mantle exhumation was associated with magmatic activity. The different transport directions between the 'high-temperature' ($>700\,^{\circ}$C) and later hydrous 'cold' deformation structures show a complex relationship between 'brittle' and 'ductile' deformation processes during continental breakup that is not yet fully understood.

Age constraints on the deformation history can be obtained from deformation structures within gabbros that have been dated by U–Pb on zircons at 161 ± 1 Ma (Bathonian–Callovian) (Schaltegger *et al.* 2002). Clasts of gabbros and albitites, also dated at 161 ± 1Ma by the same method, are found in pillow and sedimentary breccias overlain by middle–Upper Jurassic radiolarian cherts. This observation suggests that the gabbros were exhumed, probably along detachment faults, to the sea floor directly after their crystallization. Microstructures in these gabbros reveal a deformation history ranging from synmagmatic to brittle conditions. This deformation was acquired during their intrusion into partially serpentinized mantle rocks and their subsequent exhumation at the sea floor.

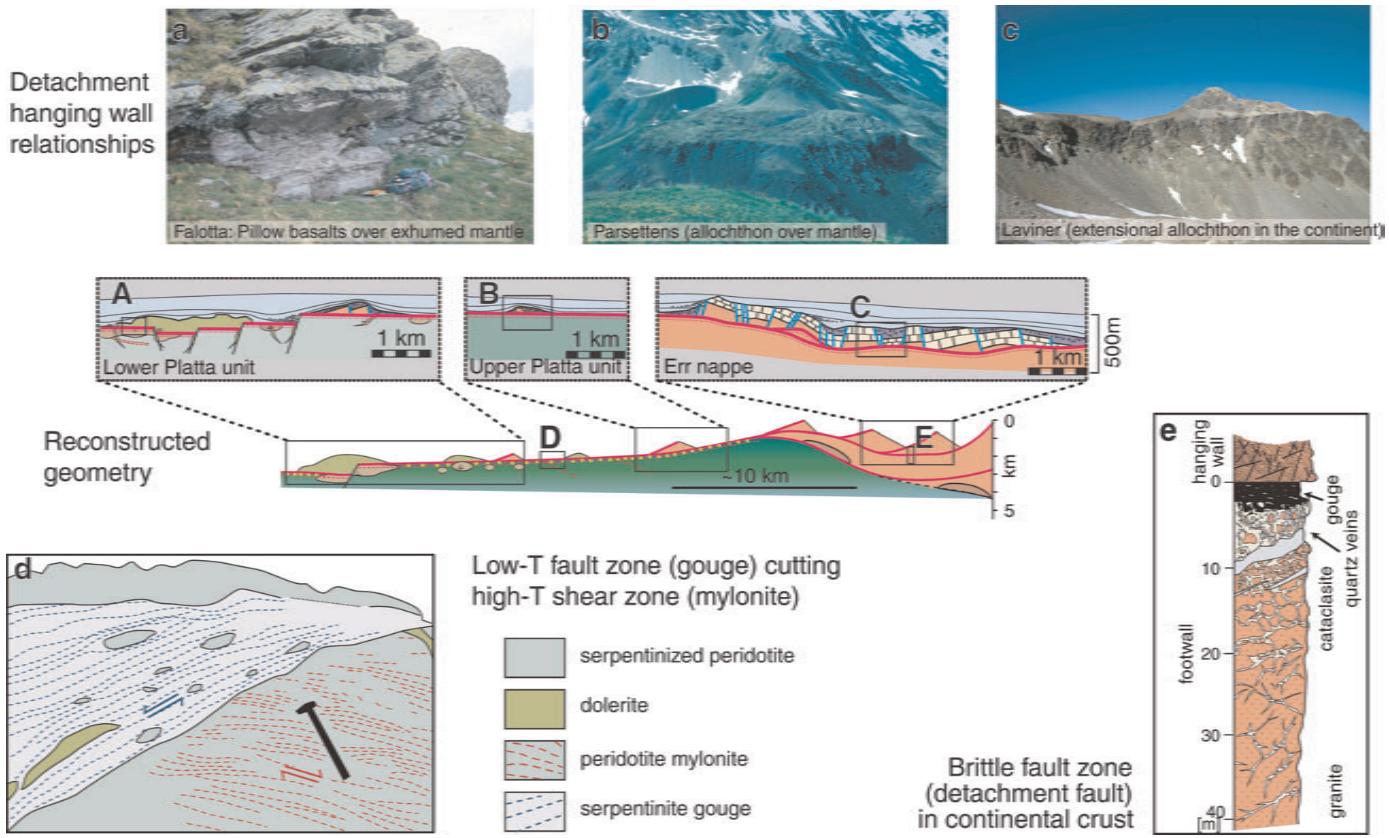

Fig. 9. Tentative palinspastic section across the Err–Platta OCT showing the observed relationships between the detachment system and its hanging-wall and footwall structures. In the three grey boxes overlying the section, reconstructed geometry of mapped rift-related structures are shown (for legend see Fig. 8). A–E: observed structures in the Err-Platta OCT: (**a**) pillow basalts overlying exhumed mantle in the Falotta area (Platta nappe); (**b**) continent-derived breccias separated from exhumed mantle along a sub-horizontal detachment fault in the Parsettens area (Platta nappe); (**c**) an allochthon truncated by the sub-horizontal Err detachment in the area of Piz Laviner (Err nappe); (**d**) preserved cross-cutting relationships between a top to the ocean serpentinite gouge zone, a top-to the continent high-temperature mantle shear zone and a dolerite dyke in the area of Al Cant (Platta nappe); and (**e**) brittle fault zone in the Err nappe related to the Err detachment. The fault zone architecture shows a transition from granite to cataclasites containing syntectonic quartz veins. The fault core is formed by a black gouge. For locations and more explications of the outcrops see Manatschal *et al.* (2003).

- *Relation between detachment faults and its sedimentary cover*. In several places in the Err–Platta OCT synrift sediments overlie the exhumed detachment surface at a low angle ($<20^\circ$) of discordance (Manatschal & Nievergelt 1997). Exposure of the detachment at the sea floor is also supported by the occurrence of clasts of green cataclasites and black gouge derived from the detachment in the synrift sediments (Froitzheim & Eberli 1990). These observations and the lack of any evidence for a major fault-related morphology produced by this fault, despite its large displacement (>10 km), can be explained only if the detachment fault was locally exhumed at the sea floor at a low angle. The observation that sediment sequences formed by turbidites and debris flows, classically interpreted as synrift sediments, are cut by detachment faults but also deposited onto already exhumed detachment faults shows that there are complicated polyphase relationships between synrift sedimentation and detachment faulting.

  Different types of tectono-sedimentary breccias, referred to as ophicalcite in the Alpine literature, include mainly serpentinite clasts embedded in a fine-grained, microsparitic, calcite matrix. The fabric of these breccias varies considerably between two end-member types. One end-member consists of serpentinite host rock with fractures filled by red limestone and/or white sparry calcite. An intermediate type consists of clast-supported breccias with *in situ* fragmented serpentinite clasts (Bernoulli & Weissert 1985; ophicalcites I of Lemoine *et al.* 1987). The other end-member consists of coarse, unsorted, matrix-supported breccias with fragments of serpentinite, gabbro, and continent-derived basement rocks and prerift sediments (ophicalcites II of Lemoine *et al.* 1987). Geopetal infill of sediment into crevasses and pockets in the mantle rocks indicate that these rocks were exposed at the sea floor. Indeed, the mantle rocks are also stratigraphically overlain by pillow-basalts or radiolarian cherts. Further evidence for the occurrence of mantle rocks at the seafloor is the presence of mass-flow breccias (ophicalcite type II of Lemoine *et al.* 1987) containing clasts of serpentinite and Triassic dolomites encased in a serpentine arenite or calcite matrix, and the occurrence of serpentinite arenites within the post-rift sediments (Decandia & Elter 1972).
- *Mantle and magmatic rocks in the ZECM*. The mantle rocks in the ZECM are invariably serpentinized peridotites derived from spinel lherzolites and harzburgites into which gabbros and basaltic dykes were intruded. A major and trace element study of mantle clinopyroxene across the Platta ZECM (Müntener *et al.* 2002) reveals that mantle rocks close to the continent may represent spinel peridotite mixed with (garnet)-pyroxenite layers while those at some distance from the continent are pyroxenite-poor peridotite that equilibrated in the plagioclase stability field. Textural relationships indicate that some plagioclase peridotites in the Platta nappe were formed by melt infiltration and melt-rock reaction (Müntener *et al.* 2004). Two-pyroxene and single-orthopyroxene thermobarometric data from the Platta nappe reveal an increase in the equilibration temperature from 850 ± 50 °C at 0.8–1.2 GPa near to the continent to more than 1000 °C further oceanwards (Desmurs 2001). The age of equilibration near the continent is likely to be Permian, as constrained from intrusive relationships with Permian gabbros in the Malenco peridotite (Müntener *et al.* 2000). In contrast, the oceanward increase in the equilibration temperature in the ZECM may be related to final continental breakup and onset of sea-floor spreading. The age for the infiltration event is, as suggested by clinopyroxene Nd isotopic data, very likely to be of Jurassic age (Müntener *et al.* 2004). However, because infiltration must have occurred under high-temperature conditions, this process must have initiated prior to final exhumation of the infiltrated mantle along cold brittle faults at the sea-floor.

  Gabbro bodies occupy less than 5% of the total observed serpentinite volume. Desmurs *et al.* (2002) described smaller spherical bodies, less than 100 m in diameter, and larger sill-like bodies. The smaller bodies are more homogeneous, consist of Mg-gabbro and show a decrease in grain size from the core towards the rim. The larger bodies show a great diversity in composition from primitive olivine-gabbros to highly differentiated Fe–Ti–P-gabbros and diorite, but up to 90% of the body is formed by Mg-gabbro.

  Massive basalts, pillow lavas, pillow breccias and hyaloclastites crop out in patches of variable thickness and size in the Platta nappe, and their abundance increases oceanward across the reconstructed section (Dietrich 1969). Away from the edge of continental crust, pillow lavas form isolated bodies less than 100 m in diameter and a few tens of metres thick. Oceanwards, the bodies are aligned and appear to be controlled by late, synmagmatic high-angle faults. The basalts stratigraphically overlie serpentinites, ophicalcites, gabbros and associated breccias, indicating that their emplacement post-dates exhumation of the mantle rocks and gabbros at the sea floor. However, basaltic dykes are also truncated by detachment faults showing that detachment faulting occurred within a magmatically active system. Based on Mg numbers and Ni contents of equilibrium olivine calculated from primitive basalts and gabbros, Desmurs *et al.* (2002) demonstrated that few mafic rocks found in the Err–Platta OCT are primary melts. Most represent fractionated compositions ranging from transitional mid-ocean ridge basalts (T-MORB) to normal MORB (N-MORB). The mafic rocks may be explained by low–moderate degrees of melting of a N-MORB type mantle, as indicated by initial Hf isotope data of zircons and Nd data of whole-rock samples (Schaltegger *et al.* 2002). The source of some basalt, however, is enriched in incompatible elements. This compositional variation seems to correlate, as indicated by a study of Desmurs *et al.* (2002), with the spatial distribution of the mafic rocks within the OCT in that mafic rocks with T-MORB signatures occur close to the continental margin whereas N-MORB signatures are more frequently found oceanwards.

- *Age constraints for continental breakup and emplacement of first magmatic rocks*. In the Alpine Tethys margins crustal breakup is constrained by the oldest biostratigraphic ages of the sediments sealing both oceanic and continental units. The biostratigraphic ages obtained for these first sediments, which are radiolarian cherts, give Bathonian–Callovian ages (Bill *et al.* 2001). These ages are similar to radiometric ages dating recrystallization and exhumation of the oldest magmatic rocks in the ZECM. U/Pb zircon ages on gabbros from the ZECM range between 165 and 157 Ma (Lombardo *et al.* 2002; Schaltegger *et al.* 2002), and Ar/Ar cooling ages on phlogopite in a pyroxenite of the Totalp serpentinite yielded a cooling age of 160 ± 8 Ma (Peters & Stettler 1987). Thus, all data obtained from the ZECM in the Alps converge towards a very short time period, implying that exhumation and cooling of the subcontinental mantle was associated with the onset of magma emplacement and the deposition of radiolarian cherts. This interpretation is compatible with the observation that detachment faulting leading to the exhumation of the mantle was linked directly to emplacement and exhumation of gabbros at the sea floor.

The Southern Iberia Abyssal Plain

- *Evidence for a detachment system in the Southern Iberia Abyssal Plain.* In the Southern Iberia Abyssal Plain, several strong intrabasement reflections have been observed in the region of thinned continental crust and interpreted as fault structures (Beslier *et al.* 1995; Krawczyk *et al.* 1996; Whitmarsh *et al.* 2000) (Fig. 8c). Mapping of these reflections show complex lateral variations from classical high-angle to low-angle top-basement structures (Péron-Pinvidic *et al.* 2007). In an east–west-directed section perpendicular to the margin, the L and H reflections form break-aways towards the continent, bound well-developed basins and cut into basement (Figs 8c & 10). Thus, the H and L reflections show clear characteristics of fault structures. In this paper, whenever we refer to these structures as reflectors, we use L and H, and when we refer to them as detachment faults, we use H-Detachment (HD) and L-Detachment (LD). Whereas L has been mapped to cut into the mantle defined by velocities higher than 8 km s^{-1}, H flattens within a crustal layer showing velocities of between 7 and 8 km s^{-1}, is up-warped beneath Hobby High and reaches the top of the basement at Hobby High, where it separates mantle rocks in the footwall (Site 1068) from lower crustal rocks in the hanging wall (sites 900 and 1067) (Fig. 8c).

A continentward-dipping reflection, the C reflection, is cut and displaced by HD and LD (Whitmarsh *et al.* 2000) (Fig. 8c) and must therefore be older than these structures. Only a few kilometres north of Lusigal 12, on seismic line CAM 144, Chian *et al.* (1999) described a 6.5–7.3 km s^{-1} interface at a depth similar to that of reflection C on Lusigal 12 and interpreted it as the crust–mantle boundary. Based on these data, Whitmarsh *et al.* (2000) also interpreted C as a contact between continental crust above and serpentinized peridotite below. A simple geometric argument, however, leads to an alternative interpretation of the C reflection. The C reflection is truncated by the HD and has therefore to change across HD from its footwall into its hanging wall going oceanwards. Assuming that the C reflection is a continuous structure, it must reach the top of the basement continentwards of where the HD reaches the top of the basement (Fig. 8c). At ODP sites 900 and 1067 drilling into the hanging wall of HD penetrated lower crustal rocks. Based on this line of evidence, Manatschal *et al.* (2001) interpreted the C reflection in the area of Site 1065 as an intracrustal reflection (e.g. Fig. 8c). The lower crustal rocks drilled at ODP sites 900 and 1067 show a pervasive brittle overprint at the top, which decreases in intensity down-hole (Manatschal *et al.* 2001) (Fig. 10).

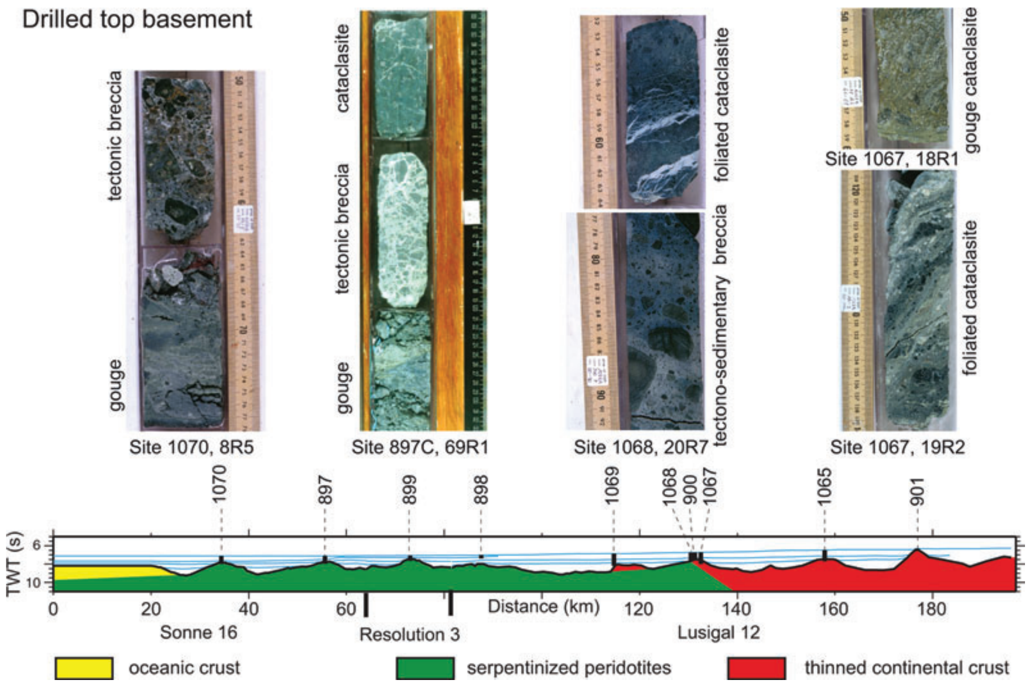

Fig. 10. Photographs of the rocks derived from the top of the basement drilled in the OCT in the Iberia Abyssal Plain. Numbers on top of the section refer to ODP sites.

$^{40}Ar/^{39}Ar$ plagioclase ages were obtained from these rocks: 136 Ma at Site 900 (Féraud *et al.* 1996) and 137 Ma at Site 1067 (Manatschal *et al.* 2001). These ages, interpreted as cooling ages, show that, at 136 Ma, these basement rocks were not yet at the sea floor, although they are at present overlain by Upper Cretaceous sediments (Fig. 8c). These observations suggest that Hobby High must be capped by a detachment fault in order to explain the cooling ages and the penetrative deformation at the top of the drilled basement sealed by sediments. At Site 1068, on the oceanward side of Hobby High and within the footwall of H, drilling penetrated mantle rocks. These rocks are capped by a serpentinite gouge and overlain by tectono-sedimentary breccias, providing further support for the existence of an exhumed detachment fault at Hobby High. This detachment, the Hobby High Detachment (HHD) (Manatschal *et al.* 2001), is interpreted to cut from the continental crust at sites 1067 and 900 into mantle at Site 1068 (Fig. 8c).

In seismic profile Lusigal 12, HHD is interpreted to break out about 15 km east of Hobby High (Fig. 8c). From there, it can be followed towards Hobby High, where it has been drilled. Further oceanwards it plunges beneath a high drilled at Site 1069. This high, with a topographic relief of about 1.5 km and 9 km long, is capped by pre-, syn- and post-rift sediments, and is interpreted as a continent-derived extensional allochthon emplaced onto exhumed subcontinental mantle (Fig. 8c). This interpretation requires a large offset along the HHD, which is also compatible with the observation that this fault forms the top of the basement over a distance of more than 20 km.

- *Zone of exhumed continental mantle (ZECM).* In the Southern Iberia Abyssal Plain, the thinned continental crust is bounded oceanwards by an up to 130 km-wide zone of exhumed continental mantle (ZECM). This zone shows distinct geophysical characteristics. Its seismic velocity structure differs from that of the adjacent stretched continental and oceanic crust. A 2–4 km-thick upper basement layer with a P-wave velocity of 4.5–7.0 km s^{-1} and a high velocity gradient (*c.* 1 s^{-1}) merges into a lower layer $\leq$4 km-thick with velocities of approximately 7.6 km s^{-1} and a low velocity gradient ($<$0.2 s^{-1}) (Chian *et al.* 1999). Moho reflections are weak or absent. The top-basement velocity is lower than that of the adjacent continental crust, while the velocity in the lower layer is too high to represent magmatically intruded or underplated continental crust or even oceanic layer 3. The velocity structure can best be explained by mantle serpentinization decreasing with depth. The seismic velocity structure gradually changes oceanwards to typically oceanic approximately 20 km west of the peridotite ridge (Whitmarsh *et al.* 2001*b*).

North–south-trending, low-amplitude magnetic anomalies in the ZECM indicate that basement magnetization is typically much lower than that of the oceanic basement further to the west (Srivastava *et al.* 2000; Russell & Whitmarsh 2003). These weak magnetic anomalies have been explained to result either from small magmatic bodies within the ZECM (Russell & Whitmarsh 2003) or from serpentinization within the hinge of downward-concave faults during exhumation of mantle rocks in the ZECM (Sibuet *et al.* 2007). Morphologically, the acoustic basement may be divided into two regions of north–south-trending basement ridges and deep (*c*. 9 km) relatively low-relief basement; both narrow northward.

ODP cored exhumed mantle rocks at four sites (sites 897, 899, 1068 and 1070) and each time penetrated serpentinite gouge and/or cataclasite in several places overlain by tectono-sedimentary breccias (Fig. 10). Because drilling targeted basement highs, sampling is probably biased, but drilling results are compatible with the observed velocity structure, suggesting widespread serpentinization of the mantle in the Iberia Abyssal Plain. Primary-phase chemistry and clinopyroxene trace element compositions obtained from the Iberia margin indicate a heterogeneous mantle depleted by less than 10% partial melting and percolated by mafic melts (Beslier *et al.* 1996; Abe 2001; Hébert *et al.* 2001). Serpentinization began, at least locally, before sea-floor exhumation of the peridotite (Skelton & Valley 2000). Despite strong serpentinization, the exhumation of mantle rocks from deep lithospheric levels to the ocean floor has been documented by a sequence of deformation structures under decreasing temperatures. This is shown by peridotite mylonite shear zones, serpentinization, the formation and subsequent cataclastic reworking of serpentinite mylonites, and low-temperature replacement by calcite (Beslier *et al.* 1996). The serpentinized mantle is capped by tectono-sedimentary breccias reworking adjacent exhumed basement (Manatschal *et al.* 2001). At Site 1070 (Fig. 10), which is situated above a margin-parallel basement ridge, upper Aptian (112.2–116.9 Ma) sediments overlie subcontinental mantle, which is intruded by gabbro veins. These gabbro veins are the only rift-related intrusive magmatic rocks drilled in the Iberia margin. Extrusive magmatic rocks were recovered only as clasts within sedimentary breccias in the ZECM (Sawyer *et al.* 1994).

- *Dating continental breakup*. The age of continental breakup, referred to in this paper as the process that leads to the final mechanical separation of two lithospheric plates that is followed by sea-floor spreading, has been classically determined by the first sea-floor spreading magnetic anomaly and/or a breakup unconformity. Such dating was based on the assumptions that: (1) breakup of continental crust is immediately followed by sea-floor spreading; (2) rifting is of the same age across and along the whole margin; and (3) the sediment architecture in the OCT is simple. The results obtained from the Iberia–Newfoundland margin show that all these assumptions have to be questioned, and that determining the age of continental breakup is difficult (Péron-Pinvidic *et al.* 2007).

  The oldest magnetic anomaly in the southern Iberia Abyssal Plain has been interpreted by Whitmarsh & Miles (1995) as M3 and by Srivrastava *et al.* (2000) as M17. These contrasting interpretations suggest that, at magma-poor margins, magnetic anomalies are not necessarily formed by classical sea-floor spreading processes and consequently are not reliable for dating onset of sea-floor spreading. The same is the case for the sediments. Because only drill sites over highs were able to reach the basement in the OCT, the oldest prerift sediments overlying oceanic as well as continental crust, suitable to date the age of breakup, have not yet been drilled.

  Dating magmatic rocks emplaced during continental breakup in the OCT was possible only for one albitite clast from a breccia drilled at Site 1070. U/Pb dating on a single zircon yielded an age of 127 ± 4 Ma (Beard *et al.* 2002), which is the age of the overlying magnetic anomaly J. U/Pb ages on zircons from a meta-gabbro dredged at the Deep Galicia margin gave an age of 121.7 ± 0.4 Ma, which is the age of the overlying magnetic anomaly M0 (Schärer *et al.* 1995). In the Alps, the emplacement of gabbros in the ZECM is contemporaneous with the deposition of the first sediments overlying oceanic crust. Therefore, it seems that dating gabbros from the OCT is a reliable way to constrain the age of accretion of the surrounding crust.
- *Rates of extension during continental breakup*. Rates of extension during final rifting can be determined from a kinematic inversion of profile Lusigal 12, presented in Figure 11. In this reconstruction, the thickness of the predetachment

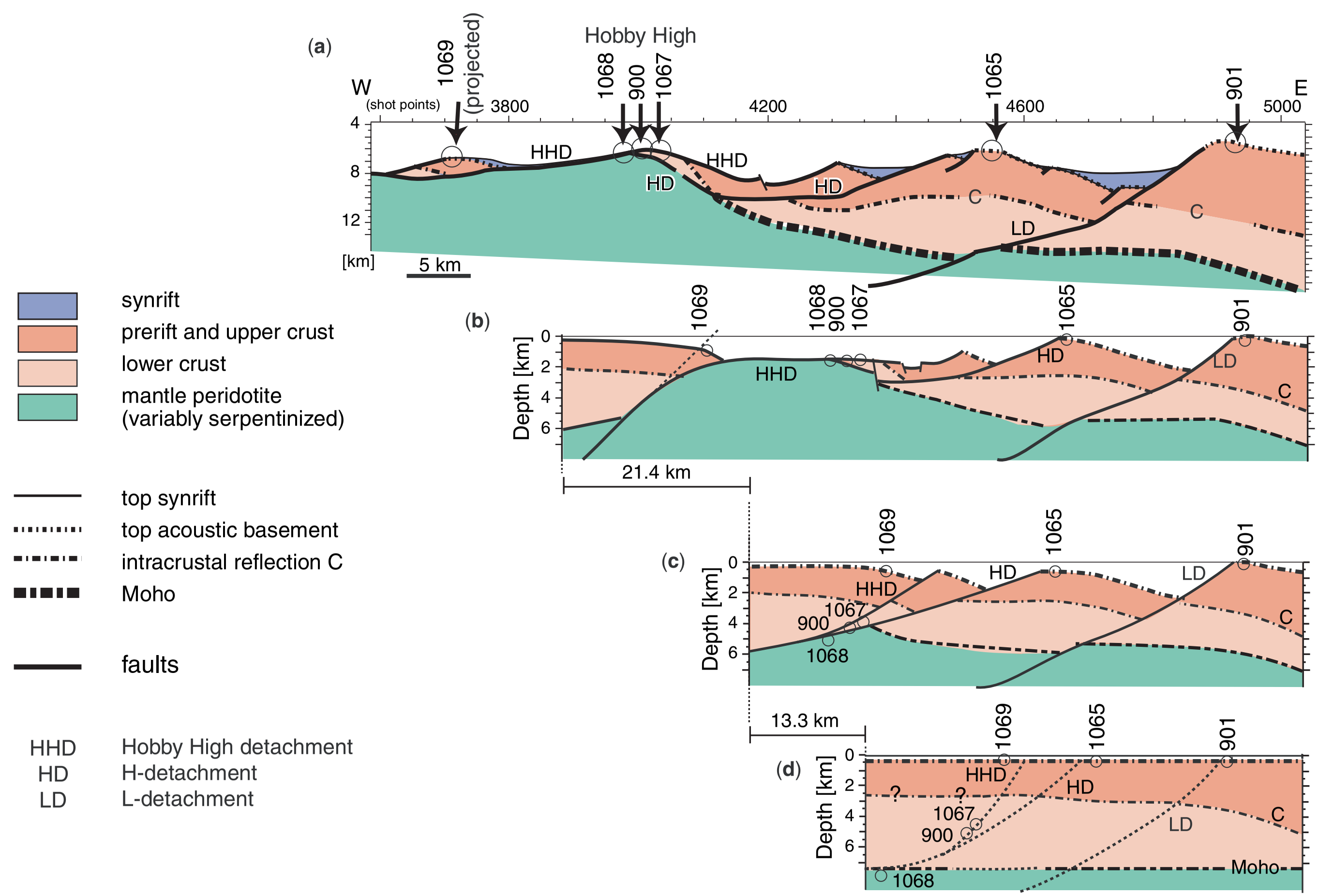

Fig. 11. (**a**) Geological interpretation of the depth-migrated Lusigal 12 profile showing the distribution of upper and lower crustal rocks exhumed subcontinental mantle rocks and reflections interpreted as detachment faults. Numbers on top of the section refer to ODP sites. (**b**)–(**d**) Evolution of faulting as determined from kinematic inversion of the interpreted Lusigal 12 profile shown in (b): (**b**) mantle exhumation along downwards-concave fault HHD, the bathymetry is constraint for this stage for Site 1069 (less than 1500 m) (Kuhnt Urquhart 2001; Urquhart 2001); (**c**) crustal extension accommodated by upwards-concave faults LD, HD and HHD; and (**d**) situation before onset of extension along LD, HD and HHD (it is important to note that the crust was already thinned to less than 10 km at this stage). For further discussion see the text.

crust and the former crustal location of the ODP sites (circles labelled with ODP site numbers), as well as amount of thinning, are obtained from the step-by-step kinematic inversion of the profile. For details of the kinematic inversion and a more extended discussion see Manatschal *et al.* (2001).

Within this inverted profile three time markers exist: (1) the Tithonian sediments drilled and dated at sites 901, 1065 and 1069, which are tilted and consequently predate the detachment faults; (2) the $^{40}Ar/^{39}Ar$ plagioclase ages of 136 Ma at Site 900 (Féraud *et al.* 1996) and 137 Ma at Site 1067 (Manatschal *et al.* 2001), dating cooling of the Hobby High basement rocks across the 150 °C isotherm and therefore representing a maximum age for first mantle exhumation; and (3) 128 Ma, the age of the M3 magnetic anomaly which is well identified in the Iberia margin (Whitmarsh & Miles 1995, Russell & Whitmarsh 2003). In order to determine the rate of extension one can calculate from the step-by-step kinematic inversion of the profile the total amount of extension that was accommodated along the LD, HD and the HHD in order to exhume the first mantle at the sea floor. The reconstruction shows that the total extension accommodated by LD, HD, and HHD is of the order of 34.7 km (initial, predetachment length is 34.6 km, final post-detachment length is 69.3 km). This estimate of extension is similar to the approximately 40 km of continental extension estimated from crustal thickness variations by Minshull *et al.* (2001) for the IAM9 profile 50 km to the south. This total amount of extension had to be accommodated between the Tithonian (145 Ma) and 136 Ma, when the first mantle rocks were within the reach of the sea floor. Using these numbers, an average velocity of 3.8 mm year^{-1} is obtained. The velocity of mantle accretion in the ZECM between the first exhumation at Hobby High at about 136 Ma and the first magnetic anomaly M3 (128 Ma) 80 km further oceanwards is approximately 10 mm year^{-1}, similar to the half-spreading rate immediately to the west, which is approximately 10–14 mm year^{-1} (Whitmarsh & Miles 1995). These values are of the same order as those observed at ultra slow-spreading ridges (Dick *et al.* 2003).

Extensional processes in magma-poor rifted margins

The very similar strain evolution, stratigraphic record and isostatic response to rifting described from the Alpine Tethys and Iberia–Newfoundland margins suggest that the underlying processes controlling lithospheric extension in these two pairs of margins are the same or at least very similar. This observation enables us to combine the complementary datasets from the two pairs of margins to describe different types of fault systems and deformation modes, both of which depend on the initial conditions and rheological evolution of the extending lithosphere.

Fault systems in rifted margins

The previous descriptions show striking similarities between the observed fault systems in the Alps and seismically imaged structures in the Iberia–Newfoundland margins. Hölker *et al.* (2002*a*) modelled synthetic seismic sections using density, velocity and thickness data from the well-exposed detachment structures in the Alps (Hölker *et al.* 2002*b*, 2003), and compared the synthetic seismic profiles they obtained with seismic data and drilling results from Iberia. These studies provide some interesting insights into the seismic characteristics of detachments observed in the OCT, but also show the limitations of seismic imaging methods. Hölker *et al.* (2003) demonstrated that intrabasement detachment structures such as the Err detachment fault (Fig. 9) would be transparent and therefore not imaged in a seismic section. Nevertheless, we believe that some of the reflections observed in the seismic sections in the Iberia–Newfoundland margins have their equivalents in observed fault structures in the Alps. Based on a combination of field observations with seismic reflection data we define three types of fault systems, which are described in the section below (Fig. 3).

Upwards-concave faults. Reflections interpreted as upwards-concave faults are typically found in the proximal margins but occur also in the distal margins (e.g. H and L reflection; Fig. 8c). Because most of these reflections are observed on time sections, the downwards increase in seismic velocity tends to accentuate the apparent concave-upwards curvature. However, structures comparable to these reflections are also observed at land. In the Alps, the Ornon fault bounding the Bourg d'Oisans Basin (Chevalier *et al.* 2003) and the Lugano-Val Grande fault bounding the Generoso Basin (Bertotti 2001; Fig. 6b) were described as upwards-concave (listric) faults. On a map scale, the Lugano–Val Grande fault cuts across the prerift sedimentary cover into the basement over a horizontal distance of more than 30 km soling out in a greenschist facies mylonitic shear zone within a quartz-rich middle crust, i.e. at about 10–15 km depth (Bertotti 1991). Equivalent structures to the H and L reflections are likely to exist in more internal parts of the Alps, but were not identified in the Err–Platta OCT.

Downwards-concave faults. Downwards-concave reflections are not observed at the Iberia margin, but drilling results from the Iberia Abyssal Plain led to an interpretation of the top-basement reflection as a detachment fault. This structure forms the exhumed part of a downwards-concave fault dipping beneath the Newfoundland margin (Manatschal *et al.* 2001). The best-studied structure is the HHD, which was drilled during ODP legs 149 and 173. Analogue structures to the HHD are exposed in the Alps. More continentward parts of this fault system, not drilled off Iberia, are preserved in the Err nappe (e.g. Err detachment; Fig. 9). The transition from the continent to the exhumed mantle (e.g. Hobby High) is exposed in the Tasna OCT (Florineth & Froitzheim 1994) and situations corresponding to more oceanward portions (not drilled in Iberia) are found in the Platta nappe. These fault structures are characterized by: (1) a top-to-the-ocean transport direction; (2) a fault zone, some tens to some hundreds of metres across, affected by fluid- and reaction-assisted brittle deformation processes and capped by a gouge horizon; (3) depositional contacts with sediments overlying the fault dips at a low angle; and (4) the local occurrence of hanging-wall blocks interpreted as extensional allochthons stranded on both exhumed continental and mantle rocks (Fig. 9). These faults, referred to as 'top-basement detachment faults' (Hölker *et al.* 2003), are therefore interpreted as exhumed segments of no longer active downwards-concave detachment faults that accommodated tens of kilometres of offset and exhumed continental and mantle rocks at the sea floor.

Continentwards-dipping shear zones. In the Adriatic distal margin, mylonitic shear zones separating lower from upper crustal rocks have been mapped in the Ivrea zone (e.g. the Pogallo fault: Hodges & Fountain 1984; Handy 1987; Handy & Zingg 1991) (Figs 1c & 7) and the Malenco area (e.g. Margna fault: Hermann & Müntener 1996; Bissig & Hermann 1999; Müntener & Hermann 2001) (Fig. 1c). P–T–t (pressure–temperature–time) data from the hanging-wall and footwall rocks from these shear zones show that they were active during an early stage of rifting and led to significant thinning of the continental crust by at least 15 km (0.4–0.5 GPa) (Handy & Zingg 1991; Müntener & Hermann 2001). Although the break-away zones of these faults are not preserved, we suggest in our reconstruction in Figure 7 that they formed the oceanward limit of the Briançonnais domain in the most distal European margin. The reconstruction of this fault system, now preserved in the two conjugate margins, suggests strong rift-shoulder uplift (Briançonnais domain) contemporaneous with down-faulting and little upper crustal extension of its hanging wall (distal Adriatic margin) (Fig. 7). Thus, subdued subsidence and upper crustal normal faulting are the surface expression of the fault system that thinned the crust by more than 15 km.

Structures comparable to the continentwards-dipping shear zones mapped in the Alps are not observed clearly in the Iberia margin. The only possible candidate is the C reflection. This reflection is cut by HD and the LD (Fig. 8c) and therefore predates final rifting. In contrast to the well-defined intrabasement reflections H and L, the C reflection is less sharp and coherent and has a more complex multicyclic nature (Hölker 2001). The seismic response of ductile shear zones in continental crust has been investigated by Fountain *et al.* (1984), Hurich *et al.* (1985) and more particularly for the Ivrea zone by Holliger & Levander (1994). These authors showed that mylonitic shear zones in the continental crust (e.g. Pogallo fault) are able to produce reflections comparable to the C reflection (see also Hölker 2001). If a shear zone is responsible for the C reflection, the reflection may additionally be amplified by a likely juxtaposition of mafic lower crustal rocks (e.g. rocks drilled at sites 1067 and 900) against upper crustal rocks. The continentwards-dipping mylonitic shear zones such as the Pogallo fault are key structures to explain the thinning of the continental crust during an early stage of rifting.

Modes of lithospheric extension observed in magma-poor rifted margins

The geological–geophysical reconstruction of the Alpine Tethys and Iberia–Newfoundland margins (Fig. 3) resulted in the identification of three different modes of extension, referred to as 'stretching mode', 'thinning mode' and 'exhumation mode' (Lavier & Manatschal 2006). The description of the different modes relies on the geometry and the strain evolution of extending fault systems and the related basin architecture and isostatic response. The terms 'stretching', 'thinning' and 'exhumation' are not intended to be a quantitative description of the mode of deformation but rather to describe qualitatively how stain is accommodated during different phases of lithospheric extension. In contrast to the more general terms 'wide rift mode', 'narrow rift mode' and 'core complex mode' proposed by Buck (1991); the modes introduced here are based on observations and identify 'common' styles of extension observed in magma-poor rifted

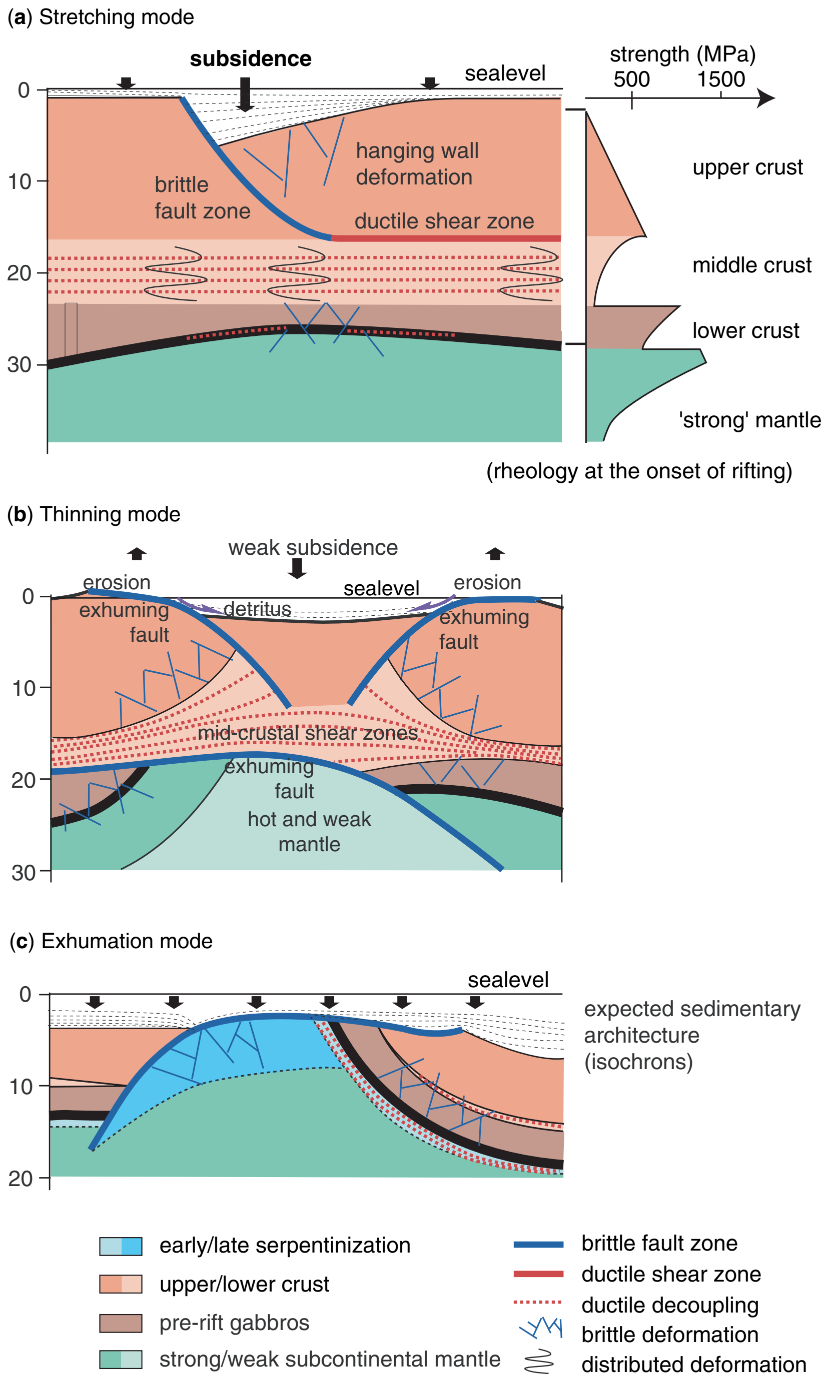

Fig. 12. Cartoons summarizing the sediment architecture, fault geometry and the strain distribution for: (**a**) stretching mode; (**b**) thinning mode; and (**c**) exhumation mode. For further explanations see the text.

margins. The aim of this paper is to describe the geological characteristics of the three modes. Lavier & Manatschal (2006) modelled the single modes. Their results show that the various modes depend on a set of physical conditions such as bulk rheology and Moho temperature at the onset of rifting. Therefore, the modes described below enable us to relate observations to rift processes and, by modelling, to investigate the underlying physics controlling lithospheric extension (Lavier & Manatschal 2006).

Stretching mode. The stretching mode is typically associated with fault systems cutting across the brittle upper crust and soling out at middle–lower crustal levels that behave ductilely and decouple deformation in the upper crust and upper mantle (Fig. 12a). Strain distribution in the upper crust is localized along high-angle faults with fault offsets of less than 10 km and producing a total extension, β, of less than 2. Strain distribution in the mantle is for this mode unconstrained by observations.

This mode results in classical fault-bounded asymmetric rift basins, up to 6 km deep and 30 km wide, filled commonly by marine deposits (e.g. Bosence 1998). Rift-shoulder uplift is suppressed, as indicated by the subsidence of both hanging wall and footwall. In this mode uplift and erosion of the highs are not observed, exhumation processes are insignificant and thinning of the crust is moderate to minor. Accommodation space increases by vertical movement, the sedimentary bedding lies at a high angle to the fault, and the overall architecture of the basin is determined by sedimentary sequences thickening towards the footwall and onlapping onto the hanging wall (Fig. 12a). Examples of basins formed by this mode of deformation are the Jeanne d'Arc Basin (Fig. 5c), the Generoso Basin (Fig. 6b) and the Bourg d'Oisans Basin (Chevalier *et al.* 2003). All these basins formed at an early stage of rifting and in the proximal parts of the future margin (Fig. 3b).

Thinning mode. The thinning mode is the most enigmatic phase of extension. This mode explains the thinning of the crust to less than 10 km (Lavier & Manatschal 2006). However, in contrast to the other two modes, it is less well constrained by direct geological observations mainly because the fault system and the sedimentary basins related to this mode are overprinted during later breakup and distributed across the two conjugate margins. Therefore, the overall system can not be observed in one piece. Portions of this system are observed in the Alpine margin. Examples of fault systems belonging to this deformation mode are the Pogallo and Margna faults in the Alps (Figs 1c & 7). P–T–t data indicate that these faults are very efficient in thinning the crust, and can explain local thinning to less than 10 km (Müntener *et al.* 2000). At the Iberia–Newfoundland margin the C 'reflection' (Fig. 8c) is in a similar position to the Pogallo and Margna faults in the Alps, but more observations are necessary to interpret this structure. Better examples for the thinning mode can be observed in parts of the margin where thinning was very pronounced but did not lead to continental breakup. Examples are the Galicia Interior Basin (Pérez-Gussinyé *et al.* 2003), the Porcupine Basin (Reston *et al.* 2001) and the Rockall Trough (England & Hobbs 1997). In all these examples the continental crust has been thinned to less than 10 km without leading to significant normal faulting and major fault-related morphology and/or tilted blocks in the upper crust.

Exhumation mode. The exhumation mode is characterized by downwards-concave faults that pull deeper crustal and mantle rocks from underneath a hanging wall and exhume them in the ZECM. This mode explains why the top of the basement in the ZECM must be considered as an exhumed fault plane. The fact that this fault exhumed deep-seated rocks to the surface and accommodated tens of kilometres of displacement without creating major fault-related topography is best explained by a downwards-concave fault geometry. Such downwards-concave faults have been proposed and referred to as rolling-hinge faults by Buck (1988) and Wernicke & Axen (1988) for the metamorphic core complexes in the Basin and Range in the SW USA. This model explains how large offsets are accommodated along active high-angles faults that rotated near to the surface to inactive lower-angle faults. A similar model has been proposed by Tucholke *et al.* (1998) to explain mantle exhumation at slow-spreading ridges, leading to so-called megamullions or 'oceanic core complexes'. Although the overall geometrical concept of downwards-concave faulting is very similar in all three settings, geologically these three settings are characterized by different crustal and lithospheric thickness and a different thermal structure.

The basin architecture and the stratal relationships associated with the exhumation mode are completely different from those of classical rift basins (Wilson *et al.* 2001). The basins form by pulling the footwall from underneath a hanging wall, i.e. the depositional area increases and is floored by an exhumed fault zone that also represents the main detritic sediment source (Fig. 12c). Thus, the relationships between the basal detachment and sediments is analogous to that of a conveyor belt, whereby the exhumed footwall rocks were fractured, exposed and redeposited along the same active fault system. This process explains the common occurrence of tectono-sedimentary breccias overlying detachment faults at low angles observed in the Alps and drilled off Iberia. These observations have also major implications for the interpretation of the stratigraphic record within OCT. In contrast to classical rift basins where synrift strata form wedges thickening into the footwall, in these basins the synrift sediments are dispersed and overlie the footwall at a low angle (Fig. 12c).

Mantle exhumation and magmatic processes during continental breakup

Mantle exhumation processes in the ZECM. Concave-downwards faults can explain in a relatively simple way the exhumation of subcontinental mantle rocks over a wide area in the ZECM without producing a high-amplitude, fault-bounded topography. Explicit in this interpretation is that the top of the mantle represents a top-basement detachment fault. This is the case in the Alps (Fig. 9) and off Iberia (Fig. 10) where the mantle is capped by a brittle fault zone and overlain by tectono-sedimentary breccias. As illustrated in Figure 13, the type of mantle rocks and their distribution in the ZECM strongly depends on the geometry of the exhuming fault system, the depth of the brittle–ductile transition, the number of faults and their overprinting relationships. In Figure 13a two end-member cases are illustrated.

In one case (example to the left in Fig. 13a) the fault soles out at the crust–mantle boundary. As a consequence, the exhumed mantle in the ZECM is formed only by the subcontinental mantle that was directly underlying the crust before detachment faulting started. In this example, the overall deformation between crust and mantle is decoupled along the crust–mantle boundary, and the asthenosphere rises passively and thermally overprints the tectonically exhumed mantle. Decoupling between deformation in the crust and the mantle may be explained by serpentinization.

At the other extreme (example to the right in Fig. 13a), the fault cuts into the mantle and soles out in a hot and ductile infiltrated lithospheric mantle overlying the rising asthenosphere. In this case the deformation between the crust and the mantle is coupled and, as a consequence, deep mantle levels can be exhumed in the ZECM. The asthenosphere is pulled out underneath the upper plate margin and the ZECM exposes a section across the lithospheric mantle. The results of Desmurs *et al.* (2001), Müntener & Piccardo (2003) and Müntener *et al.* (2004) demonstrate that the mantle rocks in the Alpine ZECM change from spinel peridotite mixed with (garnet)-pyroxenite layers next to the continent to pyroxenite-poor peridotite that equilibrated in the plagioclase stability field further oceanwards. These changes in the nature of the mantle rocks in the ZECM are compatible with a detachment fault cutting into the mantle. In contrast, the structural observations clearly indicate that the detachment fault leading to the exhumation of the mantle to the sea floor are late, cold structures that were only active in the serpentine stability field. High-temperature mylonite shear zones that may have been responsible for the exhumation from deeper and hotter mantle portions are cut by colder detachment faults showing an opposite sense of shear (Fig. 9; photo D). Thus, simple monophase detachment models are unable to explain the overall observations related to mantle exhumation. Lavier & Manatschal (2006) proposed a polyphase evolution of detachment faulting in which hot and deep mantle was first exhumed and emplaced underneath a thinned crust during the thinning mode before it was pulled out along a late and cold detachment to the sea floor during the exhumation phase. In order to test this model more data are needed regarding the precise distribution of different mantle rocks in the ZECM and the timing and kinematics of mantle shear zones.

In Figure 13b we illustrate how the interaction of detachment faults can explain characteristic features in the ZECM such as the formation of extensional allochthons or peridotite ridges. Propagating faults, i.e. faults that cut in front of the active fault, can explain the formation of allochthons that are soled by detachment faults (e.g. block drilled at Site 1069 imaged in the Lusigal 12 seismic section, Fig. 8c). This block is soled by a strong reflection interpreted as the continuation of the HHD, that separates a block formed by upper continental crust drilled at Site 1069 from the underlying mantle that was drilled at Site 1068 (Fig. 8c).

The formation of peridotite ridges can be explained by back-stepping faults, which would explain the formation of the highs as well as the observation that these ridges are, in contrast to allochthons, not soled by strong reflections, i.e. detachment faults. However, such a mechanism would imply that the peridotite ridges formed during mantle exhumation, which is, at least for some examples (sites 897 and 899 and 1277), shown not to be the case (Péron-Pinvidic *et al.* 2007). Therefore, back-stepping detachment faults may form peridotite ridges, but they cannot explain the present day basement topography of the ZECM.

Relation between extensional and magmatic processes. Although the Iberia–Newfoundland and the Alpine Tethys margins are commonly referred to as 'non-volcanic' rifted margins, several

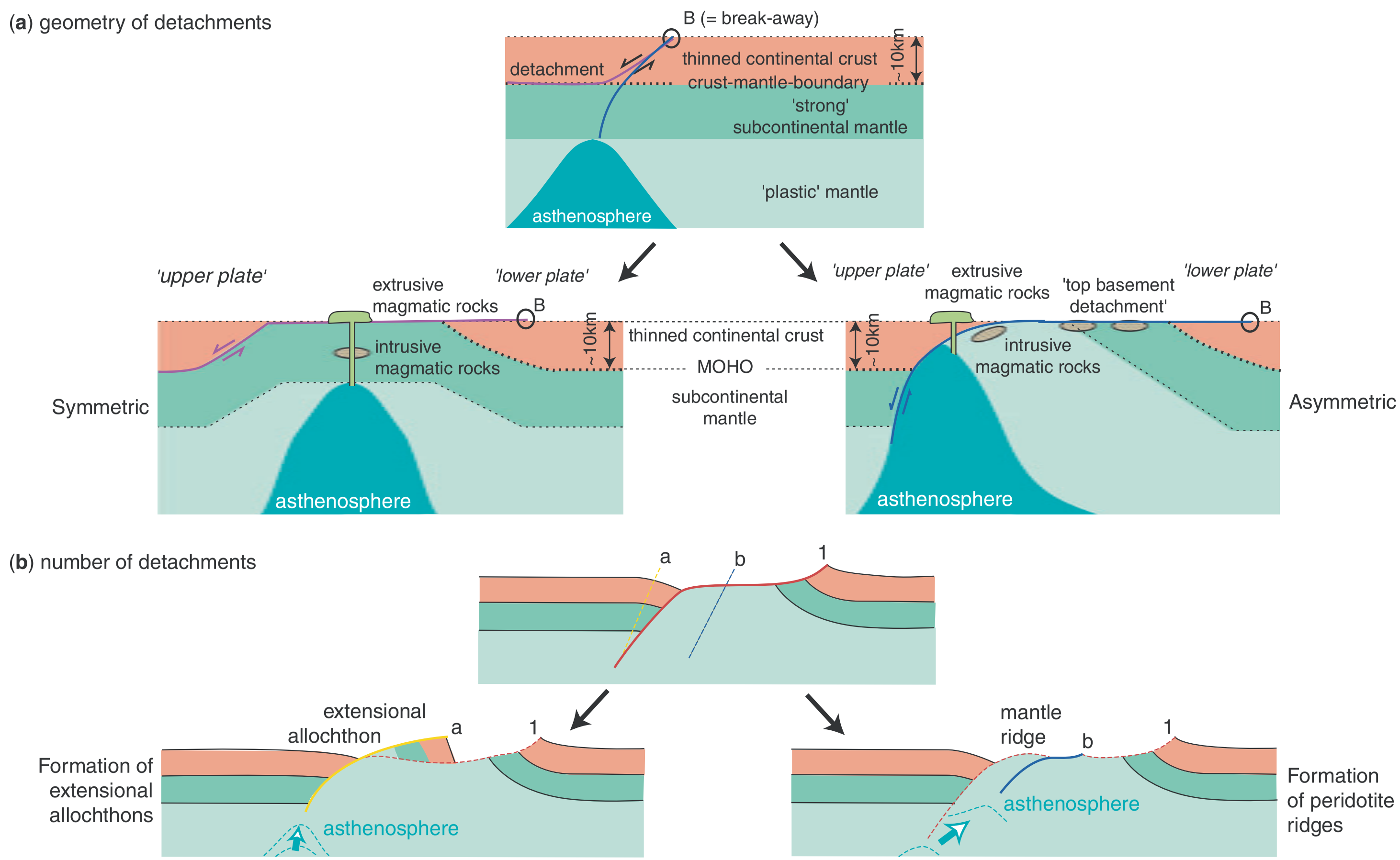

Fig. 13. (**a**) Geometry of detachment faults and its importance for the distribution of magmatic and mantle rocks in the ZECM: the example of a detachment fault soling out at the crust–mantle boundary (to the left) and a detachment cutting into the mantle lithosphere (to the right). (**b**) Interplay between detachment faults: sequence of propagating detachment faults cutting into the hanging wall leading to the formation of extensional allochthons (to the left) and detachment faults cutting in the footwall leading to the formation of peridotite ridges.

observations demonstrate that continental breakup within these two pairs of margins was preceded by magmatic activity. Detachment faults cut dolerites (Fig. 9; photo D), and infiltrated mantle peridotites are exhumed together with synrift gabbros along detachment faults at the sea floor. These observations from the Alps and Iberia clearly indicate that detachment faulting interacted with an active magmatic system before continental breakup. Therefore, we believe that the term 'magma poor' is more appropriate and should be used instead of 'non-volcanic' or 'non-magmatic' for the Iberia–Newfoundland and Alpine Tethys margins.

At present, it is not yet understood how extensional faulting and emplacement of igneous rocks interact during final rifting, and how magmatic rocks are distributed in the ZECM. Modelling results by Lavier *et al.* (1999) and Huismans & Beaumont (2002) show that strong thermal/magmatic weakening suppresses asymmetric fault systems and results in localized symmetric systems classically described at mid-ocean ridges. The existence of magma clearly changes the rheological evolution of the margin, but what is the volume necessary to start sea-floor spreading and how do different distribution processes (dyking v. percolation) affect the bulk rheology of the extending lithosphere during continental breakup? The lack of data, in particular from the Newfoundland and European margins, makes it impossible to propose a single model. The evolution of a magmatic system, its relation to detachment faults and the distribution of the magmatic rocks in the ZECM are illustrated in Figure 13a. The major difference between the two models is the degree of asymmetry, the position of the asthenosphere and the distribution of magma in the extending system. In the extreme case where a single downwards-concave fault cuts across the whole lithosphere, all extrusive and shallow intrusive magmatic rocks are emplaced in the 'upper plate' margin, while the intrusive magmatic rocks and percolated mantle rocks that are genetically linked to the extrusive rocks are exposed in the 'lower plate' margin.

Although it is generally accepted that the Iberia–Newfoundland and Alpine Tethys margins are asymmetric, surprisingly little is known about the conjugate margins: in Newfoundland drilling at ODP Site 1276 failed to sample basement rock, and in the Alps most of the European margin has been subducted. Therefore, more studies, in particular on the Newfoundland and Alpine European margins, are needed to constrain the generation and distribution of magma in magma-poor rifted margins, and to understand how magmatism interacts with serpentinization and deformation processes during final breakup.

Conclusions and outlook

In this paper we have used observations from the Iberia–Newfoundland and Alpine Tethys margins to describe and discuss the tectonic evolution of an extending lithosphere during magma-poor rifting. The most pertinent observations are:

- the polyphase nature of rifting;
- the evolution from initially distributed to final localized extension;
- the different stratigraphic, isostatic and tectono-metamorphic evolution of the distal and proximal future margins during early rifting;
- the observed close relationship between fault geometry, basins architecture and isostatic response, enabling us to distinguish between a stretching mode, a thinning mode and an exhumation mode;
- the importance of exhumation processes during a final stage of rifting;
- the complex relationship between mantle exhumation, detachment faulting and magmatic processes.

Models of the tectonic evolution of the Iberia–Newfoundland and Alpine Tethys margins will need to capture these observations. Future studies need to link observations with processes and to discuss the underlying physics controlling lithospheric extension. It will be important to determine the physical conditions for each of the described modes and to understand what controls their transitions into other modes and how these transitions are documented in nature. Of particular importance will be the understanding of the complex interactions of mechanical and magmatic as well as chemical and thermal processes during rifting. It seems likely that each of the processes can change the bulk rheology and consequently control the strain evolution of the lithosphere during rifting.

The increasing ability to use more realistic boundary conditions in modelling will also result in a demand for more complete and better datasets. In particular observations at a subseismic scale, constraints on age and palaeobathymetry, and data on tectono-metamorphic, hydrothermal and magmatic processes during final rifting are needed. Such observations are, however, difficult to obtain from rifted margins *in situ*. Future studies on land not only need to bridge the different observation scales, but also will need to quantify better the observations.

The paper benefited from the careful reviews of N. Christie-Blick, A. Robertson and N. Direen that enabled us to improve the manuscript. We thank also A. Bouzeghaia for improving the quality of the figures. Support for this study came for the GDR marges

(France) (G. Manatschal and G. Péron-Pinvidic), the Royal Society of London through a Joint Project grant (T. A. Minshull, G. Manatschal and G. Péron-Pinvidic) and the Swiss National Science foundation (grant no. 21-66923 (O. Müntener and G. Manatschal) and PP002-102809) (O. Müntener).

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