Int J Earth Sci (Geol Rundsch) (2004) 93:432–466
DOI 10.1007/s00531-004-0394-7

ORIGINAL PAPER

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New models for evolution of magma-poor rifted margins based on a review of data and concepts from West Iberia and the Alps

Received: 26 June 2003 / Accepted: 15 March 2004 / Published online: 3 June 2004

Abstract Direct observation and extensive sampling in ancient margins exposed in the Alps, combined with drill-hole and geophysical data from the present-day Iberia margin, result in new concepts for the strain evolution and near-surface response to lithospheric rupturing at magma-poor rifted margins. This paper reviews data and tectonic concepts derived from these two margins and proposes that extension, leading to thinning and final rupturing of the continental lithosphere, is accommodated by three fault systems, each of them characterized by a specific temporal and spatial evolution during rifting of the margin, by its fault geometry, and its surface response. The data presented in this paper suggest that margin architecture and distribution of rift structures within the future margin are controlled first by inherited heterogeneities within the lithosphere leading to a contrasting behaviour of the future distal and proximal margins during an initial stage of rifting. The place of final break-up appears to be determined early in the evolution of the margin and occurs where the crust has been thinned during a first stage to less than 10 kilometres. During final break-up, the rheology of the extending lithosphere is controlled by the thermal structure related to the rise of the asthenosphere and by serpentinization and magmatic processes.

Keywords Magma-poor rifted margins · Iberia/Atlantic · Tethys/Alps · Detachment faulting · Sedimentary basins

Introduction

Geophysical surveys and, in some rare examples, analysis of samples collected by deep-sea drilling and submersibles have revealed that continental break-up can lead to contrasting types of margins: volcanic rifted margins characterized by thick igneous crust including underplated material contrasting with non-volcanic or magma-poor rifted margins which contain virtually no igneous crust formed during break-up. These results presented new problems to the margins community : how does the strength of the lithosphere evolve during continental rifting and how do extensional faulting and emplacement of igneous rocks, which are the result of temperature-dependant processes of solid-state deformation and decompression melting, control rift architecture? This paper discusses the tectonic evolution of magma-poor rifted margins by reviewing data and concepts derived from the Iberia and Alpine Tethyan margins.

In the past two decades, numerous geophysical surveys, combined with deep-sea drilling in the Iberia margin and field investigations on well-preserved fragments of the ancient Alpine Tethyan margins, exposed in the Alps, enabled a precise description of the architecture of the Ocean-Continent Transition (OCT) in magma-poor rifted margins. These investigations demonstrated the importance of detachment faulting and mantle exhumation in the OCT (Boillot et al. 1987; Reston et al. 1995; Manatschal et al. 2001) and the changing sediment architecture across the margin (Wilson et al. 2001a). Some of the results have been reviewed in recent papers (Manatschal and Bernoulli 1999; Whitmarsh and Wallace 2001), or have been published in a volume of the Geological Society of London entitled "Non-Volcanic Rifting of Continental Margins: A Comparison of Evidence from Land and Sea" (Wilson et al. 2001b). Recently, Whitmarsh et al. (2001) published a conceptual model aiming to explain the tectonic evolution of magma-poor margins from onset of rifting to sea-floor spreading along magma-poor rifted margins.

This paper discusses the tectonic evolution and surface response of lithospheric rupturing in magma-poor rifted margins and focuses particularly on the deformation processes related to rifting. It examines how these processes control the architecture of rifted margins and how they are

"Dedicated to Daniel Bernoulli who taught me to compare the geological record of oceans and orogens"

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related to sedimentary, magmatic, and hydrothermal processes. In an introductory part of the paper I review observations derived from magma-poor margins recognized in the Alps and seismically imaged and drilled in Iberia and show how this data can be used to constrain the temporal and spatial evolution of the deformation processes from the onset of rifting to the beginning of seafloor spreading. In the second part of the paper, well-preserved rift structures in the Alps and in Iberia are described and compared in order to investigate the processes controlling the tectonic evolution of the margins. In a final part, some general questions and unresolved problems related to the evolution of rifted margins are addressed, aiming to present a more general context on the evolution of rifted margins which goes beyond that of the present-day Iberia and the ancient Alpine margins.

The benefits of comparing data from land and sea

Studies comparing data from land and sea have a long tradition and played an important role in the development of a conceptual understanding of rifted margins (Fig. 1) (e.g. Boillot & Froitzheim 2001). An excellent example of this iterative approach is the discovery and study of rift basins in pre-orogenic Alpine settings. Rift basins were mapped at sea using seismic techniques and on land using sedimentological, stratigraphic and structural methods. Bernoulli (1964) demonstrated, based on mapping of abrupt changes in facies and thickness of the Liassic formations across a fault zone in the southern Alps in southern Switzerland, the existence of Liassic rift basins and compared them subsequently to seismically imaged structures in rifted margins (e.g. Bally et al. 1981). Bernoulli's work demonstrates how careful observations and descriptions on land can lead to a coherent interpretation of structures imaged by seismic methods at sea. This study is certainly one of the most important contributions to the understanding of rifted margins. The importance of Bernoulli's research goes beyond the discovery and description of rift basins in the Alps and the interpretation of rift structures in seismic sections. Equally, if not more important, was the iterative approach which he used, bridging observations across different scales, from the drill-hole, through outcrop, to the seismic scale in order to compare observations made on land and at sea (e.g. Bernoulli and Jenkyns 1974).

Further examples demonstrating the strength of the land-sea approach are the discovery of mantle exhumation and detachment faulting in the OCT, and the concept of ophiolites (Fig. 1). In all these examples, extensive sampling and direct observations on spatially limited outcrops on land close the gap between drill-hole and geophysical surveys of analogous structures at sea. Because the data sets obtained from rifted margins will always be very limited in their spatial and temporal resolution, the use of complementary data and a comparative approach will remain very important. However, the comparison of data sets collected across a range of spatial scales and obtained from different methods is only justified if the comparison is rigorous. Interpretations based on comparative studies have to ensure that they neither result from biased sampling nor go beyond the resolution of the data sets. Comparisons, therefore, have to be limited to general and well-identifiable features. Single observations cannot necessarily be extrapolated to a larger or smaller scale, and not every observation resulting from one margin will be applicable to another margin. In order to make a comparison meaningful, the available data sets have to have sufficient overlap, which is the case for the Iberia and Tethys margins discussed here. From no other margin has so much information been obtained on the temporal and spatial evolution of rifting. A wealth of data has been collected in the past three decades from Iberia and over more than a century from the Alps. Recent studies showed that data and concepts derived from the ancient margins exposed in the Alps are able to support marine-derived hypotheses of margin development proposed from drilling and seismic surveys off Iberia (e.g. Manatschal and Bernoulli 1998, 1999; Whitmarsh et al. 2001; Wilson et al. 2001a). Thus, the interpretations presented in this paper for the Iberia and Alpine Tethyan margins are relatively well constrained by observations and data.

The paleogeographic framework

In the Atlantic and western Mediterranean area, Late Triassic to Early Cretaceous rifting and opening of different oceanic basins spatially followed approximately the extent of the Variscan orogeny in Western Europe. The resulting ocean basins were part of an equatorial spreading system, which extended from the Caribbean to the eastern Mediterranean area and beyond, including the Central Atlantic, the eastern Mediterranean, the Liguria-Piemonte ocean, and the Vardar-Meliata ocean to the east (Fig. 2). The evolution of the different branches of the Alpine Tethys was determined by the movements of the North America, the Africa, the Eurasia, and the smaller Iberia, and Adriatic plates (Fig. 2b,c) (cf. Ricou 1994). The evolution of the Liguria-Piemonte ocean, from which most of the Alpine ophiolites are derived, was contemporaneous with and kinematically linked to the opening of the Central Atlantic in the Jurassic and separated Eurasia/Iberia from Adria/Africa (Fig. 2c). The early Cretaceous opening of the Northern Atlantic was associated with rotation of the Iberian plate relative to Eurasia (Fig. 2b). The different segments of the Atlantic-Tethys ocean system opened at different times and the closing of the Liguria-Piemonte segment was contemporaneous with ongoing spreading in the Central and Northern Atlantic (Fig. 2b). Thus, the margins compared here are of different ages but both formed on a lithosphere previously affected by the Variscan orogeny.

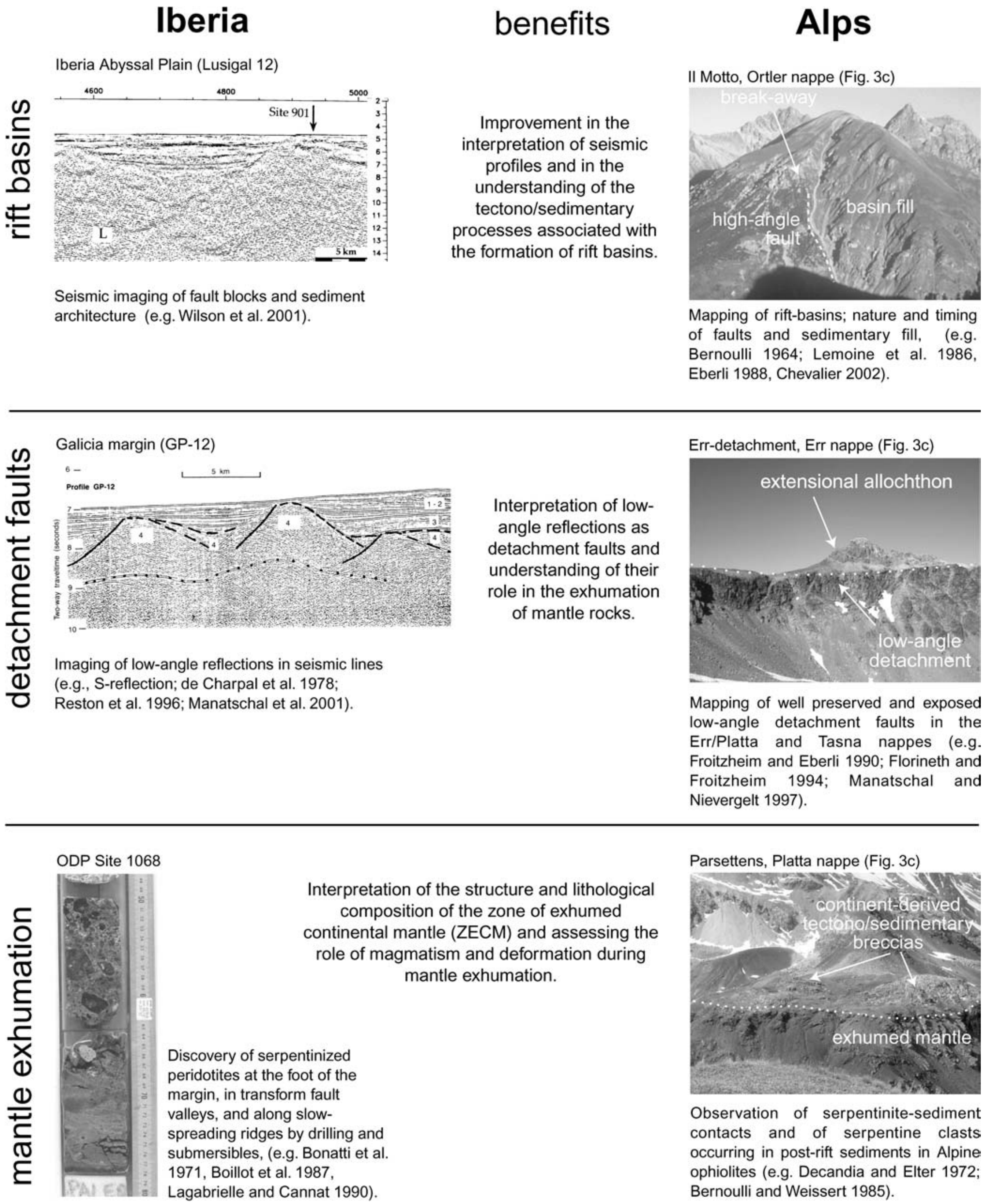


Fig. 1 Comparison of observations from land (*Alps*) and sea (*Iberia*) and the benefits of combining land/sea observations

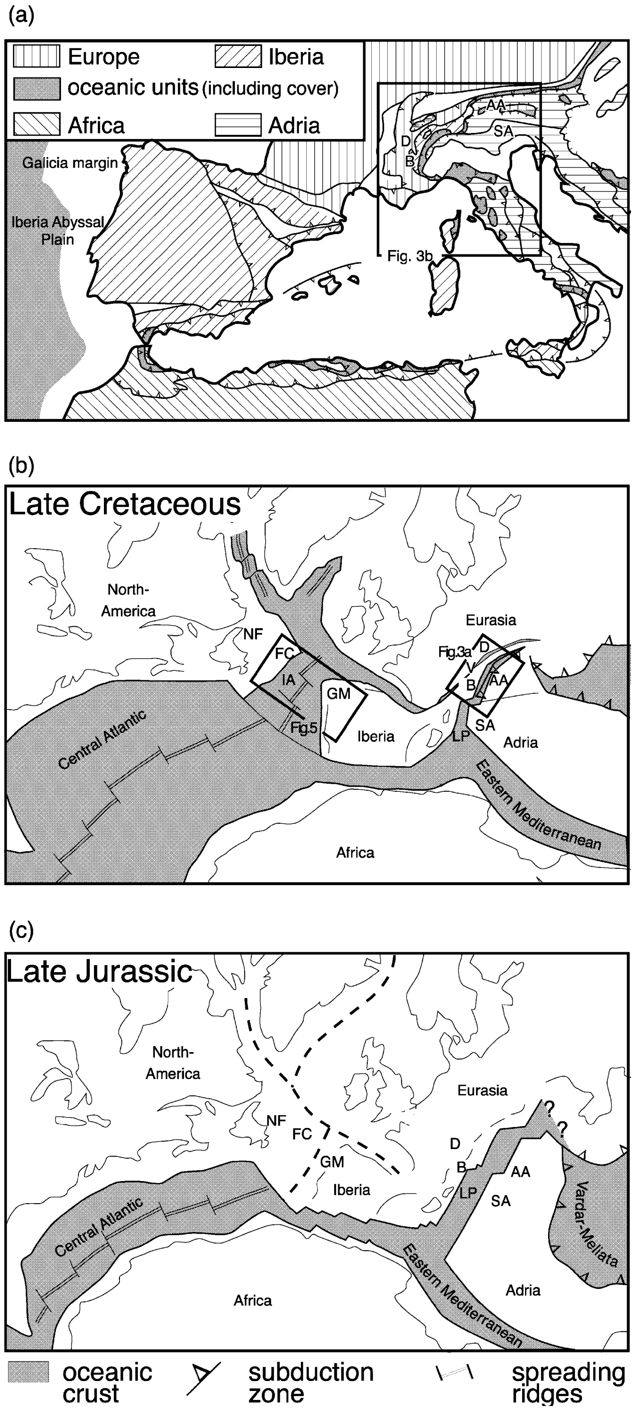


Fig. 2a–c Present-day map of the western Mediterranean and paleogeographic evolution: (**a**) Tectonic sketch map showing the distribution of the Eurasian, Iberia, Adria, and Africa continental areas and the distribution of the tectonic elements derived from them. Oceanic units are remnants of the Mesozoic Tethys. Large-scale paleogeography reconstructed for: (**b**) the Late Cretaceous and (**c**) the Late Jurassic. *AA*: Austroalpine; *B*: Briançonnais; *D*: Dauphiné; *FC*: Flemish Cap; *GM*: Galicia margin; *IA*: Iberian Atlantic; *LP*: Liguria-Piemonte ocean; *NF*: Newfoundland; *SA*: southern Alps (modified after Manatschal and Bernoulli 1999)

Remnants of ancient margins preserved in the Alps

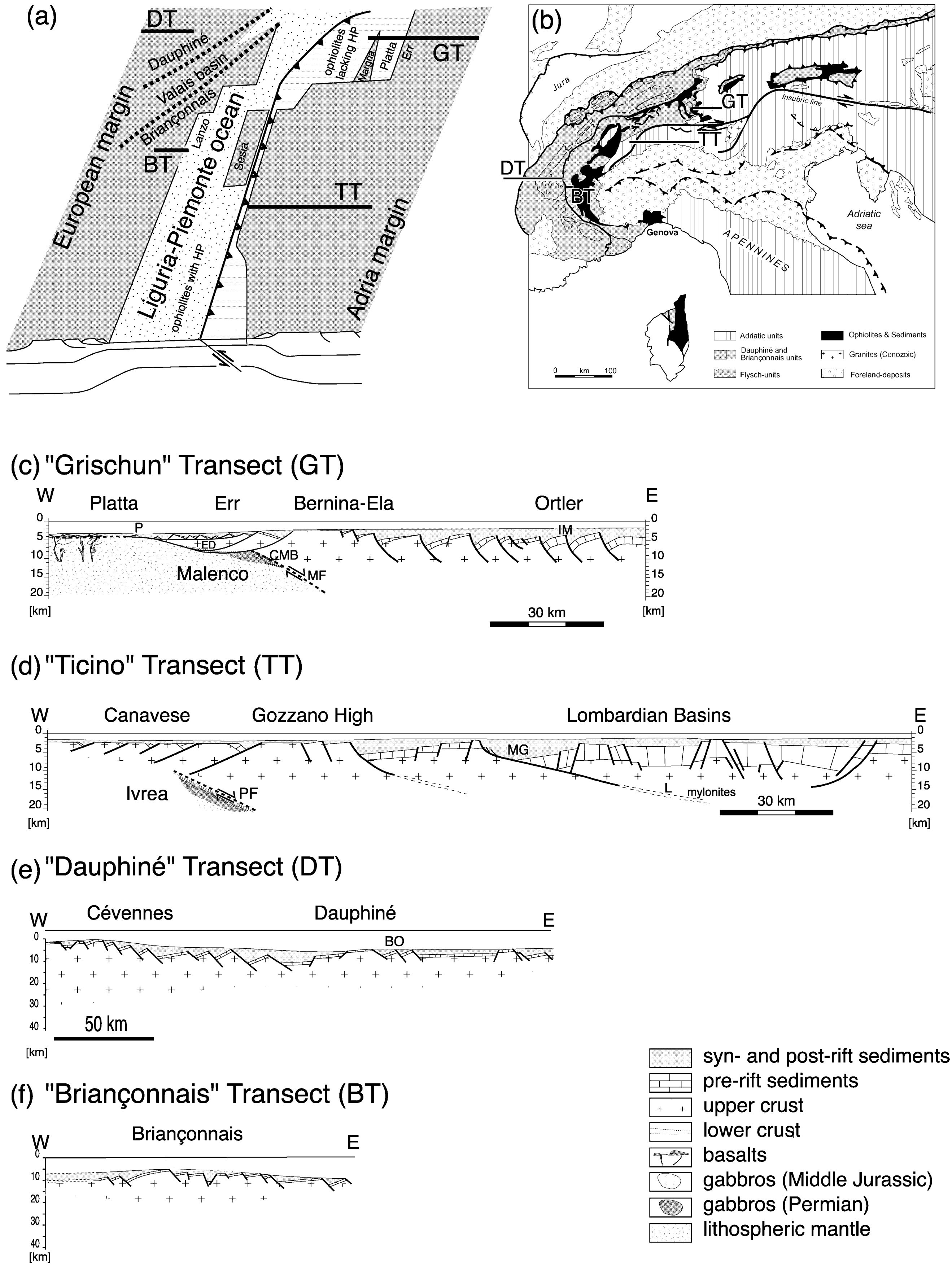
Reconstruction of ancient rift structures in the Alps

Understanding of the tectonic evolution of the Alps is a prerequisite to reconstruct the stratigraphic evolution and architecture of the ancient rifted margins preserved within them. However, explaining this evolution would go beyond the focus of this paper and, I refer, therefore, to the papers of Froitzheim et al. (1996) and Schmid et al. (1996) for modern interpretations of the Alps and to the papers of Lemoine et al. (1986), Bertotti et al. (1993), Froitzheim et al. (1994), Manatschal and Nievergelt (1997), and Manatschal and Bernoulli (1999) for details on the reconstruction of the ancient rift structures in the Alps.

In this paper, I will focus mainly on four transects in which rift structures are exceptionally well preserved (Fig. 3). The “Grischun” and the “Ticino” transects preserve rift structures of the Adriatic margin and are exposed today in the eastern and southern Alps. The “Dauphiné” and the “Briançonnais” transects preserve rift structures of the European/Briançonnais margin and are exposed today in the Dauphiné and Briançonnais domains in the western Alps (Fig. 3). In addition to these transects, a wealth of data is available from other scattered outcrops preserving the structural and stratigraphic record of rifting.

In the “Grischun” transect (Fig. 3c), remnants of the former Adriatic margin including the OCT have been first telescoped and then extended by E-W directed movements during the Late Cretaceous, leading to the formation of the south Pennine/Austroalpine nappe stack in Grisons. This nappe stack, only to a lesser extent affected internally by post-Cretaceous deformation, formed the “orogenic lid” during Tertiary N-S directed convergence and subduction of the oceanic and European lithosphere (Laubscher 1983). Thus, in Grisons a relatively straightforward coaxial deformation enables a simple kinematic inversion of the Cretaceous south Pennine/Austroalpine nappe stack to be made which permits the reconstruction of the former margin architecture (for details see Froitzheim et al. (1994) and Manatschal and Nievergelt (1997)). According to this kinematic reconstruction, the higher nappe units (the upper Austroalpine nappes) are derived from the proximal margin, whereas the lower nappes represent the distal margin (lower Austroalpine nappes including the Err nappe) and the ocean-continent transition (south Pennine Platta and Malenco nappes) (Fig. 3a,c). In the “Grischun” transect, higher crustal levels including sediments and shallow basement were detached from deeper structural levels. Thus, only shallow crustal rift structures have been sampled within the south Pennine/Austroalpine nappe stack (Conti et al. 1994).

The “Ticino” transect (Fig. 3d), exposed in southern Switzerland and northern Italy preserves a transect located south of the “Grischun” transect across the same margin (Fig. 3a,b). This domain of the former Adriatic margin escaped Late Cretaceous deformation to a large


(a)
DT
Dauphiné
Valais basin
Briançonnais
BT
Lanzo
European margin
Liguria-Piemonte ocean
ophiolites with HP
Sesia
ophiolites lacking HP
Margna
Platta
Err
GT
TT
Adria margin
(b)
Jura
GT
Insubric line
TT
DT
BT
Genova
APENNINES
Adriatic sea
0 km 100
Adriatic units
Dauphiné and Briançonnais units
Flysch-units
Ophiolites & Sediments
Granites (Cenozoic)
Foreland-deposits
(c) "Grischun" Transect (GT)
W
Platta
Err
Bernina-Ela
Ortler
E
P
IM
ED
CMB
MF
Malenco
[km]
30 km
(d) "Ticino" Transect (TT)
W
Canavese
Gozzano High
Lombardian Basins
E
MG
Ivrea
PF
L
mylonites
30 km
[km]
(e) "Dauphiné" Transect (DT)
W
Cévennes
Dauphiné
E
BO
50 km
[km]
(f) "Briançonnais" Transect (BT)
W
Briançonnais
E
[km]
syn- and post-rift sediments
pre-rift sediments
upper crust
lower crust
basalts
gabbros (Middle Jurassic)
gabbros (Permian)
lithospheric mantle

extent and was mainly affected by N-S directed shortening during Tertiary continent-continent collision (Schumacher et al. 1997). Thus, in the "Ticino" transect the N-S trending rift structures are sampled in a south-vergent thrust wedge and were partly reactivated as transcurrent faults and oblique ramps within the E-W trending fault-and-thrust belt (Schönborn 1992). The basin architecture in the proximal margin is well preserved and accessible in the central portion of this transect, but is poorly exposed and heavily deformed in the distal part, situated in the Canavese zone. Parts of the lower crust, belonging to the distal margin, are exposed in the Ivrea zone and are separated from the upper crust and its sedimentary cover along the Pogallo fault (Handy 1987). In the "Ticino" transect, rift structures can be traced into the basement (Bertotti 1991). Thus, in this transect the original architecture of the proximal margin is locally well preserved and large-scale structures observed in the basement and associated pressure – temperature – time (p-T-t) data can be related to events documented in the stratigraphic record within the overlying sediments (Bertotti et al. 1993).

The "Dauphiné" and "Briançonnais" transects (Fig. 3e,f) exposed today in the Dauphiné and Briançonnais domains in the western Alps, form remnants of the European/Briançonnais margin which were delaminated and accumulated in a number of crystalline-basement and sedimentary-decollement nappes forming the Helvetic/Subalpine and middle Pennine nappe pile. The original position of the two transects on the European/Briançonnais margin is difficult to determine because of the ill-constrained Alpine kinematics in the western Alps and the possible existence of an additional oceanic domain (Valais ocean of Stampfli 1993). However, it is at present generally accepted that the "Briançonnais" transect preserves distal portions and the "Dauphiné" transect proximal portions of the Jurassic European/Briançonnais margin. The nature and extent of the Valais domain, separating the two transects and their Cretaceous and Tertiary history, are still debated. Without entering into a discussion supporting or rejecting the oceanic nature of the Valais domain, it is undisputed that this zone formed a zone of crustal weakness during Cretaceous/Tertiary time and that the two transects did not form a continuous profile across the former European/Briançonnais margin as shown in Lemoine et al. (1986). Style and timing of rift structures observed in the "Dauphiné" transect are similar to those found elsewhere in the external domains of the Alpine chain and are therefore considered to be characteristic for the whole proximal European margin.

Fig. 3a–f The Liguria-Piemonte ocean and related margins: (**a**) Paleogeographic situation of the Liguria-Piemonte ocean and adjacent margins during Late Cretaceous time (modified after Dal Piaz 1995). (**b**) Tectonic map of the Alps (modified after Polino et al. 1990) showing the distribution of the remnants of the Liguria-Piemonte ocean. (**c**) The "Grischun" transect preserved in the South Pennine-Austroalpine nappes in Grisons (modified after Manatschal and Bernoulli 1999). *CMB*: crust-mantle boundary; *ED*: Err detachment; *IM*: Il Motto; *MF*: Margna fault; *P*: Parsettens. (**d**) The "Ticino" transect exposed in the southern Alps (modified after Bertotti et al. 1993). *MG*: Monte Generoso Basin; *L*: Lugano-Val Grande fault; *PF*: Pogallo fault. (**e**) The "Dauphiné" transect preserving the proximal European margin exposed in the Western Alps (modified after Lemoine et al. 1986). *BO*: Bourg d'Oisans. (**f**) The "Briançonnais" transect preserving the distal European margin exposed in the Briançonnais area, SE France (modified after Lemoine et al. 1986)

Reconstruction of sea-floor sequences preserved in the Alps

Palinspastic reconstructions of the ophiolite units in the Alps were obtained by kinematic inversion of the Alpine nappe stack. This approach, however, is only reasonable for ocean-floor sequences, which were telescoped into a simple nappe stack (e.g. Grisons) but not for sequences which were first subducted and then exhumed before being emplaced into the nappe stack (e.g. Franco/Italian Alps). The ophiolites exposed in Grisons lack a high-pressure metamorphic overprint, indicating that they have never been subducted. They show the same Alpine p-T path (Ferreiro-Mählmann 2001) and the same post-rift sedimentary evolution as the overlying continental Err nappe (Dietrich 1970), suggesting that the continental (Err nappe) and the ophiolitic units (Totalp-Platta-Malenco) exposed in Grisons came from paleogeographically adjacent areas, and that, together, they preserve a former ocean-continent transition (Manatschal and Nievergelt 1997) (Fig. 3a,c). This is further supported by the observation that mantle rocks in the Malenco unit locally preserve pre-Alpine contacts with lower crustal rocks (Hermann et al. 1997) and in the Platta nappe continent-derived allochthons are overlying exhumed mantle rocks (Manatschal and Nievergelt 1997).

The ophiolites exposed along the French/Italian Alps, except the uppermost unit in the Chenaillet ophiolite, have been affected by high-pressure metamorphism. Schwartz et al. (2000) demonstrated that different ophiolites show individual p-T paths, indicating that they were disrupted during subduction and/or later exhumation before being emplaced into the final nappe stack. Thus, a reliable kinematic inversion for such units is difficult, if not impossible, without a better understanding of the exhumation processes of high-pressure rocks in collisional orogens.

An alternative method to determine the original paleogeographic position of oceanic units may be possible if the age of the oceanic accretion can be determined and a simple accretion history can be assumed, i.e. the older the crust, the more marginal its position within the former oceanic basin. Interestingly, all available age data, in particular U/Pb zircon ages on gabbros obtained from the ophiolite sequences of the Liguria-Piemonte ocean, point towards a late Middle to early Late Jurassic age (about 160 Ma) (Lombardo et al. 2002; Schaltegger et al. 2002 and references therein). The narrow age range found in the Jurassic gabbros overlaps with the ages of the fist radiolarian cherts, sealing continental and oceanic sequences in the OCT (Bill et al. 2001), and the cooling ages obtained from gabbros and mantle rocks derived from Alpine ophiolites (Peters and Stettler 1987; Bill et al. 1997). These data demonstrate that the Alpine ophiolites represent the oldest and therefore also the most marginal parts of the former oceanic basin. This implies that most of the younger oceanic crust must have been subducted; indeed, the geochemistry and petrology of ophiolites show that unambiguous oceanic lithosphere is rare in the Alps and Apennines (Rampone and Piccardo 2000; Desmurs et al. 2002). These observations also have important implications for the sampling/accretion processes during Alpine convergence, clearly indicating that these processes are different from those observed in "complete" ophiolite units (e.g. Oman).

Accepting a simple accretion history within the Liguria-Piemonte basin and a subsequent simple geometry of the subduction leads to a straightforward palinspastic reconstruction in which all ophiolites with a high-pressure metamorphic overprint must be derived from the subducted plate, i.e. the European side. Those not affected by high-pressure metamorphism have to come either from the Adriatic side or may represent delaminated remnants of oceanic crust that escaped subduction and were thrusted onto the European margin (e.g. Chenaillet). Following this argument, the ophiolites exposed in Grisons are interpreted to come from the Adriatic side and those showing exposure to high-pressure conitions along the French/Italian Alps are interpreted to originate from the European/Briançonnais side (Fig. 3a). Because the ocean-floor sequences in Grisons show similarities with the rocks drilled in the OCT in Iberia one would expect, by analogy, that the ophiolites exposed in the Franco/Italian Alps could be equivalent to the sea-floor sequences off Newfoundland, an aspect which is going to be discussed later in this paper.

Architecture of the conjugate Adriatic and European/Briançonnais margins

Rifting in the Alpine realm began with differential subsidence and the formation of rift basins in more proximal parts of the future conjugate margins before it localized in more distal portions of the future margin. This resulted in a different basin architecture in the proximal and distal parts of the margins (Lemoine et al. 1986; Eberli 1988; Manatschal and Bernoulli 1998).

Overall, the rift structures in the proximal European/Briançonnais and Adriatic margins show a very uniform architecture, which is characterized by fault-bounded basins. Well-preserved examples of such basins are spectacularly exposed in the area of Bourg-d'Oisans in the "Dauphiné" transect (Fig. 3e), in the Monte Generoso area in the "Ticino" transect in southern Switzerland (Fig. 3d), and in the Ortler nappe in the "Grischun" transect in the eastern Alps (Fig. 3c). These basins range in width between 10 and 30 km and have accumulated a few hundred to some thousands of metres of syn-rift sediments. The basins were bounded by east- as well as west-directed high-angle faults. The sediment fill varies from basin to basin but is dominated by turbidites and other mass-flow deposits interbedded with hemipelagic sediments commonly showing fining and thinning upward cycles (Eberli 1988).

More distal parts of the Alpine-Tethyan margins, in particular of the European margin, are only preserved locally. In the Briançonnais domain, interpreted as a fragment of the distal European/Briançonnais margin, rift structures are only preserved locally and include high-angle faults (Claudel and Dumont 1999) (Fig. 3f). In this part of the margin, there is evidence for a change from initial subsidence and block tilting to uplift during a later stage of rifting. The uplift and erosion is well constrained by a prominent erosional surface showing karst features in Triassic to Lower Jurassic shallow-water carbonates filled with Middle Jurassic sediments (Lemoine et al. 1986). Because the conjugate Adriatic margin subsided beneath the calcite compensation depth (CCD) at the same time, as demonstrated by the deposition of radiolarian cherts, Lemoine et al. (1987) proposed a simple-shear model with a detachment fault dipping beneath the European/Briançonnais margin.

Remnants of an OCT belonging to the Briançonnais margin are preserved in the Tasna nappe (Florineth and Froitzheim 1994). Due to the uncertain paleogeographic position and age of this OCT, it is unclear how this structure can be related to the remnants of the distal margin exposed in the Briançonnais domain in south-eastern France. The OCT exposed in the Tasna nappe will be discussed in a later part of this paper.

Structures of the distal Adriatic margin are well exposed and preserved in the Err nappe in the "Grischun" transect (Fig. 3c). The most prominent structure is a system of low-angle detachment faults. The faults form break-aways in the continental crust, cut oceanwards into the mantle and are overlain by extensional allochthons (Fig. 3c). Pre-rift lower crustal rocks are not exposed along this detachment system but occur along the same margin in the Malenco area (Fig. 3c) and in the Ivrea zone (Fig. 3d). In both places, lower crustal rocks are separated from the upper crust by crustal-scale, continentward dipping faults; the Margna fault in the Malenco area (Müntener and Hermann 2001) and the Pogallo fault in the Ivrea zone (Handy and Zingg 1991) (Fig. 3c,d).

Architecture and composition of the oceanic (ophiolite) units

Neither the "ophiolites" in Grisons nor the sea-floor sequences drilled off Iberia show the characteristics of "true" oceanic crust. In order to characterize the genetic evolution and compositional and structural features of these sea-floor sequences, which are different to those of the adjacent oceanic and continental crusts, Whitmarsh et al. (2001) introduced the term "Zone of Exhumed Continental Mantle" (ZECM). In this paper the term "ZECM" is used only for the part of the margin which is floored by

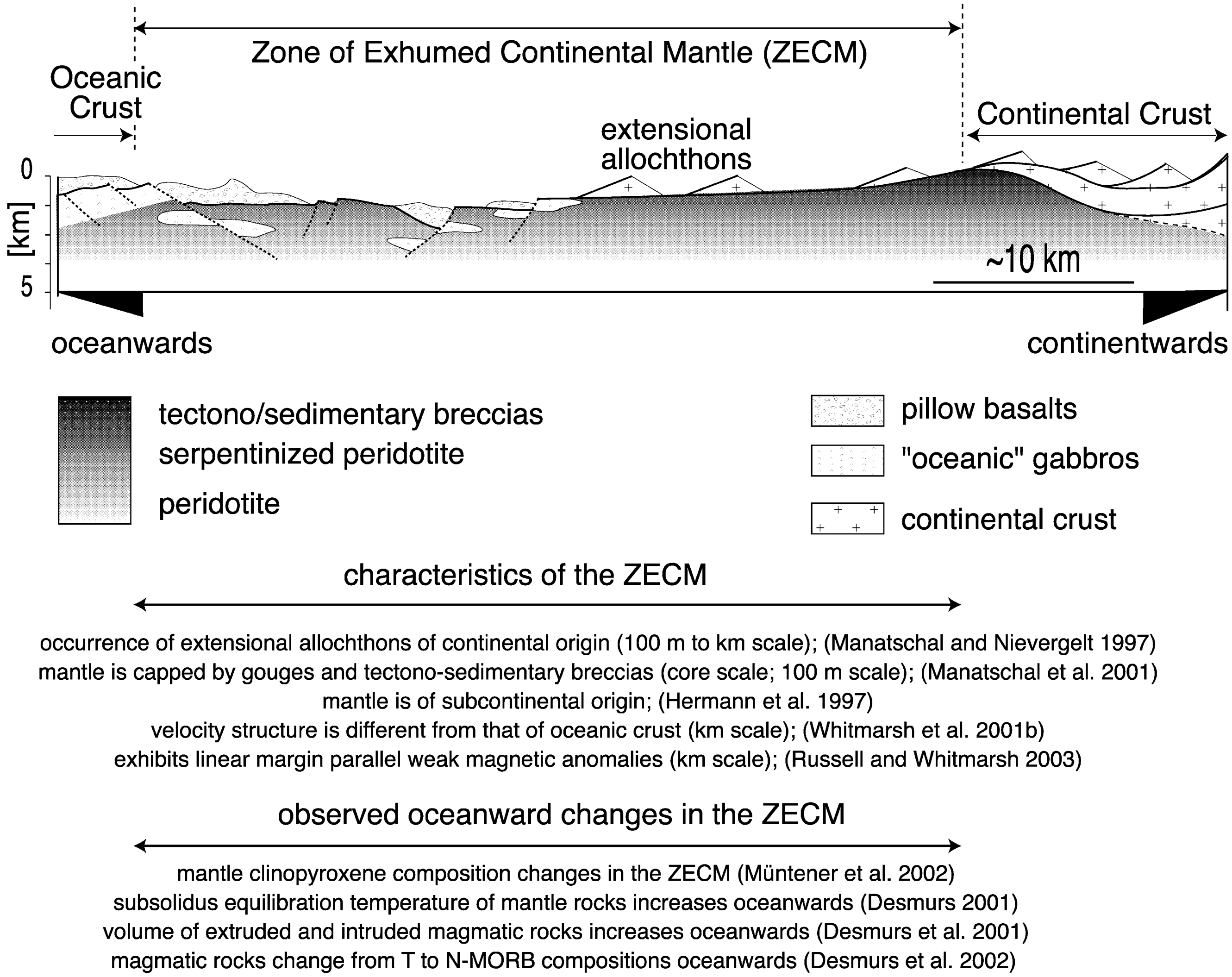


Fig. 4 Architecture and characteristics of the Zone of Exhumed Continental Mantle (ZECM) in magma-poor rifted margins

subcontinental mantle lithosphere (Fig. 4). In contrast, the term “OCT” is used in a more general way for the transition from the distal continental margin to the fist oceanic crust and implies neither a particular genetic evolution nor a specific composition of the underlying lithosphere.

Alpine ophiolites are in many ways different to the ophiolite sequence defined by the Penrose conference (Anonymous 1972). They lack sheeted-dyke complexes, have relatively small volumes of gabbros and basalts, and show a predominance of serpentinites. Based on the magmatic architecture and their relation to exhumed mantle rocks, Lagabrielle and Cannat (1990), Lagabrielle (1994), and Lagabrielle and Lemoine (1997) interpreted the ophiolites exposed along the Franco/Italian Alps to have formed within a slow-spreading ridge environment. The presence of extensional allochthons and tectono-sedimentary breccias with continent-derived material overlying mantle of subcontinental origin led Manatschal and Nievergelt (1997) and Desmurs et al. (2001) to propose that the ophiolites exposed in Grisons formed sea-floor sequences similar to those drilled in the ZECM off Iberia. This interpretation is supported by geochemical studies of the Alpine mantle rocks (Rampone and Piccardo 2000; Müntener et al. 2002). Although these interpretations appear to be incompatible with the fact that both ophiolite sequences formed at the same time in the same oceanic basin, in a later part of this paper a model is presented in which the first sea-floor sequences formed at opposite sides of an embryonic oceanic basin may be very different from one another.

The ophiolites in Grisons exposed in the Totalp, Platta, and Malenco units represent a typical example of a ZECM. The study of the lithologies and deformation processes within these ophiolites, combined with the geophysical data obtained from the Iberia margin, permits a better characterization of the ZECM (e.g. Whitmarsh et al. 2001). Some results of the most recent study of the lithologies from these ophiolites are summarized in this section (Fig. 4). For a more complete overview see Desmurs (2001), Desmurs et al. (2001, 2002), and Müntener and Piccardo (2003). The mantle rocks in these units are invariably serpentinized peridotites derived from spinel lherzolites and harzburgites into which gabbros and basaltic dykes were intruded. A 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 the ultramafic rocks at some distance from the continent are pyroxenite-poor peridotites that equilibrated in the plagioclase stability field. Bulk-rock analyses of the peridotites located further oceanwards show fertile to extremely depleted compositions and most clinopyroxenes equilibrated with plagioclase. Textural relationships indicate that some plagioclase peridotites in the Platta nappe were formed by melt infiltration and melt-rock reaction. Whether melt infiltration is related to the onset of sea-floor spreading or represents an older and independent event in the lithospheric mantle is, however, not yet clear. 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 to 1.2 GPa near to the continent to >1000°C further oceanwards (Desmurs 2001) (Fig. 4). The age at which equilibration occurred is not known.

In the Platta nappe, the gabbro bodies that intruded into partially serpentinized mantle occupy less than 5% of the total observed serpentinite volume. Desmurs et al. (2002) showed that smaller bodies, less then 100 metres in diameter, had a different magmatic evolution to 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. Troctolites were not found. Internal magmatic layering is not observed and primary contacts with the serpentinites, forming the host rocks, are commonly strongly tectonized. The main constituent in the larger bodies is Mg-gabbro forming up to 90% of the body. The relationships between Mg-gabbros and Fe-gabbros are diffuse but both are cut by the Fe-Ti-P-gabbros and still younger diorite dykes. U-Pb zircon ages from three different gabbros and one albitite yielded precise ages of 161 ± 1 Ma, indicating that all intrusive rocks were emplaced within a very short time range (Schaltegger et al. 2002).

Massive basalts, pillow lavas, pillow breccias, and hyaloclastites occur in patches of variable thickness and size in the Platta nappe and their abundance increases oceanward across the reconstructed section (Fig. 4). Away from the edge of continental crust, pillow lavas form isolated bodies less than 100 metres in diameter and a few tens of metres thick. Oceanwards, the bodies are aligned and appear to be controlled by late, syn-magmatic high-angle faults. The basalts stratigraphically overlie serpentinites, ophicalcites, gabbros, and associated breccias, indicating that their emplacement postdates exhumation of the mantle rocks at the sea floor. However, late fault zones, formed by serpentine gouge, truncate basaltic dykes clearly indicating that tectonic activity was still active when the basalts were extruded.

Based on mineral and bulk rock chemistry as well as simple modelling, Desmurs et al. (2002) demonstrated that the gabbro bodies record different magmatic processes, ranging from predominantly fractional crystallization to solidification without fractionation. Mg numbers and Ni contents of equilibrium olivine calculated from primitive basalts and gabbros indicate that few mafic rocks originate from primary melts but most represent fractionated compositions ranging from T- to N-MORB. Most mafic rocks may be explained by low to 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 basalts, 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.

In the ancient Alpine ZECM, magnetic anomalies and sediment architecture cannot be used to constrain the temporal evolution within the ZECM. In the Alpine examples, crustal break-up is constrained by the oldest biostratigraphic ages of the sediments sealing oceanic and continental rocks in the OCT, and by radiometric ages dating the emplacement and exhumation of magmatic rocks in the ZECM. The biostratigraphic ages in equivalent sediments in the Western Alps (Gets nappe) show Bathonian to Callovian ages (Bill et al. 2001), U/Pb zircon ages on gabbros from the ZECM show ages ranging 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 and Stettler 1987). Thus, all data obtained from the ZECM point to a very short time period, implying that exhumation and cooling of the subcontinental mantle was associated with the onset of magma emplacement. This interpretation is compatible with observed field relationships between faults, magmatic rocks, and sediments within the ZECM. These observations show that detachment faulting, leading to the exhumation of the mantle, was directly linked to emplacement and exhumation of gabbros at the sea floor. This contradicts the interpretation of Costa and Caby (2001) who proposed, based on Sm/Nd model-ages for gabbros, that magmatic activity within the margin initiated about 20 Ma before final break-up. Later in this paper, I will discuss the relationship between magmatism, serpentinization, and deformation based on observations made in the ZECM in the Alps and off Iberia.

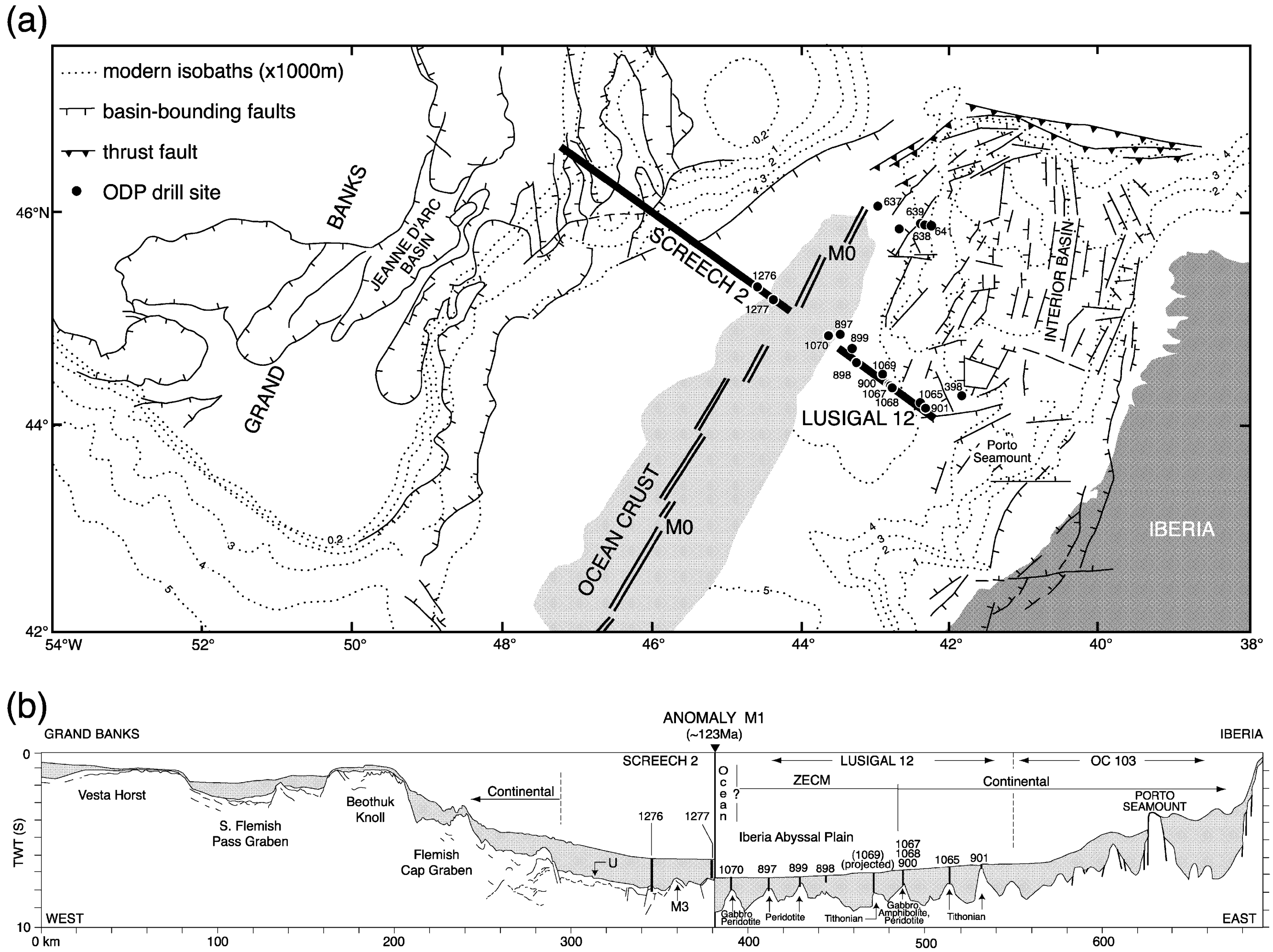


Fig. 5a – b The Iberia/Newfoundland conjugate margins: **(a)** Map of the Cretaceous Atlantic between Iberia and Newfoundland reconstructed to anomaly M0 time (~ 121 Ma) based on the reconstruction pole of Srivastava et al. (2000) with the Newfoundland plate fixed relative to present geographic coordinates (modified after Shipboard Scientific Party 2004). **(b)** Conjugate seismic reflection section from the Newfoundland and Iberia margins, juxtaposed at 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 Sites 1276 and 1277; Iberia margin: composite seismic section (Sonne 16, JOIDES Resolution, Lusigal 12, OC 103) along the conjugate Iberia drilling transect (modified after Shipboard Scientific Party 2004)

Architecture of the Iberia/Newfoundland conjugate margins

Tectonic overview

The Iberia margin off NW Spain and Portugal and its conjugate margin, the Newfoundland margin (Fig. 5), resulted from rifting and continental break-up between the North American and the Iberian plates during Early Cretaceous time (Fig. 2b). The Iberia margin has been extensively studied and drilled during ODP Legs 103, 149, and 173 and at present it is regarded by some as the typical example of a magma-poor rifted margin. In contrast to the Iberia margin, at the present time, no deep-sea drilling has been undertaken on the conjugate Newfoundland margin. However, reflection/refraction experiments suggest that this margin is different from the Iberia margin in many regards (Shillington et al. 2002). A more detailed analysis of the tectonic structure of the Iberia margin is presented in Pinheiro et al. (1996) and Whitmarsh and Wallace (2001), and of the Newfoundland margin by Jansa and Wade (1975) and Tucholke et al. (1989).

The proximal margin

Proximal parts of the margin are found in the N-S trending Interior and Porto Basins on the Iberian side (Montenat et al. 1988; Pérez-Gussinyé 2003) and the Jeanne d'Arc basin on the Newfoundland side (Driscoll et al. 1995). These basins are bounded by normal faults which dip toward the basin centres resulting in an assembly of moderately tilted blocks along the flanks and near-symmetrical grabens in the basin centres. As no wells penetrated deeper parts of these proximal basins, the reconstruction of their syn- and pre-rift sedimentary and stratigraphic evolution is little known except for the Jeanne d'Arc and Porto Basins. Wells in these two basins made the determination of several rift events (Murillas et al. 1990; Discoll et al. 1995) possible.

On the Iberia margin, the Vigo and Porto seamounts, together with Galicia Bank, form a NNW-SSE trending alignment of elevated highs which separate the Interior basin from the deep Galicia margin (Fig. 5). The geometry of the reflections, the nature of the seismic stratigraphic units, and the architecture of the basins are different east and west of these highs, suggesting that the high separates two different tectonic provinces of the margin.

The distal margin

Distal parts of the Iberia margin occur, west of Galicia Bank in the north, and at the Iberia Abyssal Plain further to the south (Fig. 5). The deep Galicia margin is characterized by N-S trending tilted fault blocks of continental basement and pre-rift sediments overlain by relatively thin syn- and post-rift sequences. These tilted blocks are underlain by a prominent reflection, the so-called *S* reflector (de Charpal et al. 1978, Reston et al. 1995). The *S* reflector can be followed oceanward towards a ridge consisting of serpentinized peridotites, the so-called peridotite ridge (Boillot et al. 1980). This ridge is supposed to separate "true" oceanic crust to the west from a "transitional" and thinned continental crust to the east (Sibuet et al. 1995).

To the south, in the southern Iberia Abyssal Plain, strong shallow-dipping reflections underlie tilted blocks of continental crust (H and L reflections) (Krawczyk et al. 1996). In contrast to the S reflection, the H and L reflections form distinct break-aways to the east, i.e. continentwards, and cut a weak, continentward dipping reflection, the C-reflection (Whitmarsh et al. 2000), interpreted in this paper as an older, continentward-dipping detachment fault. In a later part of this paper, I will propose that these reflections image detachment structures active during successive phases of continental thinning and final exhumation of mantle to the sea floor.

Little is known about the structure of the deep Newfoundland margin. Seismic reflection/refraction experiments show that the nature and structure of the crust in the distal margin is different from that of the conjugate margin in many aspects. The crust thins rapidly from 30 km to less than 5 km over a distance of only 75 km (Shillington et al. 2002) and further oceanwards the crust shows many seismic similarities to that of slow-spreading oceanic crust (Funck et al. 2003, Hopper et al. 2004). The lack of borehole data at the present time makes it, however, impossible to confirm this interpretation.

Zone of Exhumed Continental Mantle (ZECM) and oceanic crust

Off Iberia, the thinned continental crust is bounded oceanwards by a 40- to 130-km wide ZECM. This zone shows distinct geophysical characteristics. Its seismic velocity structure differs from those of the adjacent stretched continental and oceanic crusts. A 2–4 km thick upper basement layer with a P-wave velocity of 4.5–7.0 km/s and a high velocity gradient (~1 s^{-1}) merges into a lower layer $\leq$4-km thick with velocities of ~7.6 km/s 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 crust. The velocity structure probably reflects, therefore, mantle serpentinization decreasing with depth. The seismic velocity structure gradually changes oceanwards to typically oceanic 10–20 km west of the peridotite ridge.

N-S trending, low-amplitude magnetic anomalies indicate that basement magnetization is typically much lower than in the oceanic basement further to the west. This suggests that the upper seismic layer contains little magmatic material, which, however, increases in volume oceanwards. Morphologically, the acoustic basement may be divided into two regions of N-S-trending basement ridges and deep (~9 km), relatively low-relief basement; both narrow northward.

The ODP cored at nine sites in the Iberia Abyssal Plain and penetrated highly serpentinized peridotite four times. Because drilling only targeted basement highs, sampling is probably biased and the results of drilling need to be compared with geological and geochemical observations from the Alps in order to propose models for the lithological composition of the ZECM.

Primary-phase chemistry and clinopyroxene trace-element compositions obtained from the Iberia margin indicate a heterogeneous mantle depleted by <10% partial melting and percolated by mafic melts (Abe 2001; Hébert et al. 2001). Trace-element compositions are compatible with those expected from a subcontinental mantle. Serpentinization began, at least locally, before sea-floor exhumation of the peridotite (Skelton and Valley 2000). Despite strong serpentinization, the exhumation of mantle rocks from deep lithospheric levels to the ocean floor has been documented by their deformation under decreasing temperatures. This is shown by olivine 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 ODP Site 1070 (Fig. 5), which is situated above a margin-parallel basement ridge, 20 km (~2 Ma) oceanward of the peridotite ridge, upper Aptian (112.2–116.9 Ma) sediments overlying subcontinental mantle intruded by dykes were cored. Extrusive basaltic rocks were only found as small clasts and in small volumes in breccias in the ZECM (Sawyer et al. 1994). A continental allochthon, more than 10 km long and several hundreds of metres thick, was drilled at ODP Site 1069 in the southern Iberia Abyssal Plain.

The end of rifting is commonly defined at present-day rifted margins with the first sea-floor spreading magnetic anomaly, which gives a maximum age for the onset of sea-floor spreading. In the southern Iberia Abyssal Plain, mantle exhumation has not been dated directly, however, $^{40}Ar/^{39}Ar$ plagioclase ages of 136 Ma at Site 900 (Féraud et al. 1996) and 137 Ma at Site 1067 (Manatschal et al. 2001) date cooling of the gabbros and amphibolites recovered from these sites across the 150°C isotherm. Because the mantle rocks are within the same footwall and within less than 2 km from the dated rocks, exhumation of mantle across the 150 °C isotherm is interpreted to be of the same age. Thus, mantle exhumation had to occur near 137 Ma, i.e. the Berriasian/Valanginian boundary according to the time scale of Gradstein and Ogg (1996), but before 126 Ma, the age of the first undisputed sea-floor spreading magnetic anomaly (Whitmarsh and Miles 1995, Russell and Whitmarsh 2003). Thus, the delay between continental break-up (i.e. first mantle exhumation) and formation of the first unambiguous sea-floor spreading magnetic anomaly is less than 10 Ma. The velocity needed to explain the exhumation of 130 km of subcontinental mantle in the ZECM of the Iberia Abyssal Plain is in the order of 13 mm/yr, similar to that calculated for sea-floor spreading immediately to the west, which is in the order of 10–14 mm/yr (Whitmarsh and Miles 1995).

Tectonic evolution of magma-poor rifted margins

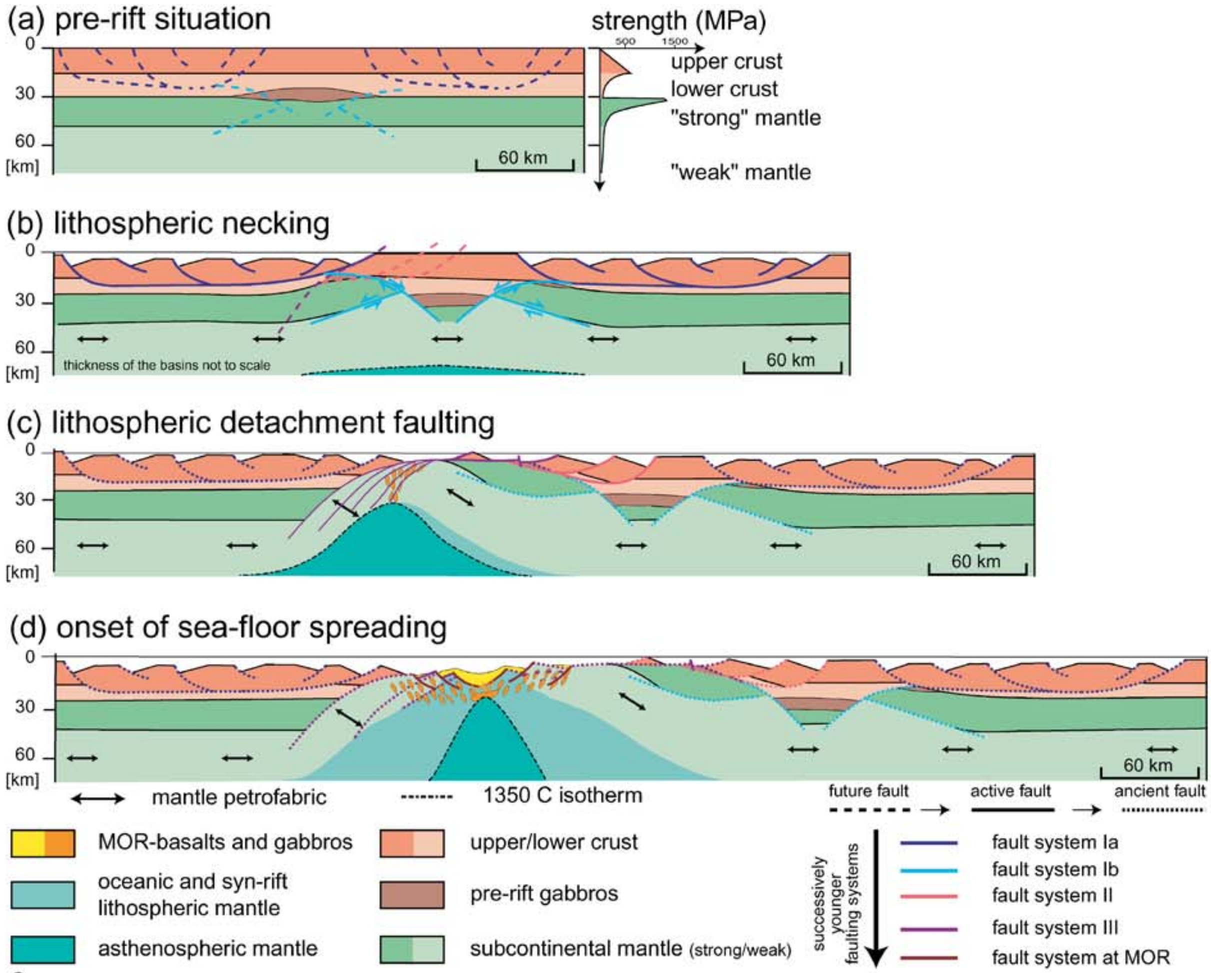
The following interpretation of the tectonic evolution of the Iberia – Newfoundland and European/Briançonnais – Adriatic conjugate margins is based on a conceptual lithospheric-scale model developed from one first published by Whitmarsh et al. (2001) (Fig. 6). The model is based principally on geophysical/geological observations from the Iberia margin and the Alps and it is constructed so that the final section corresponds to that imaged today across the Iberia and Newfoundland margin (Fig. 6d). Details of the architecture have been simplified in order to account for general features, nevertheless, all sites drilled during ODP Legs 103, 149, and 173 and all Alpine observations can be located in the section shown in Fig. 6d. Three stages of development are shown: the pre-rift stage, the necking stage, and the detachment stage. They are drawn so that the continental crust is approximately balanced . In reality, erosion has to be taken into account, but, given the scale of the model, it can be neglected and 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 break-up, which is not resolved by any observational data at the moment.

Conditions before onset of rifting

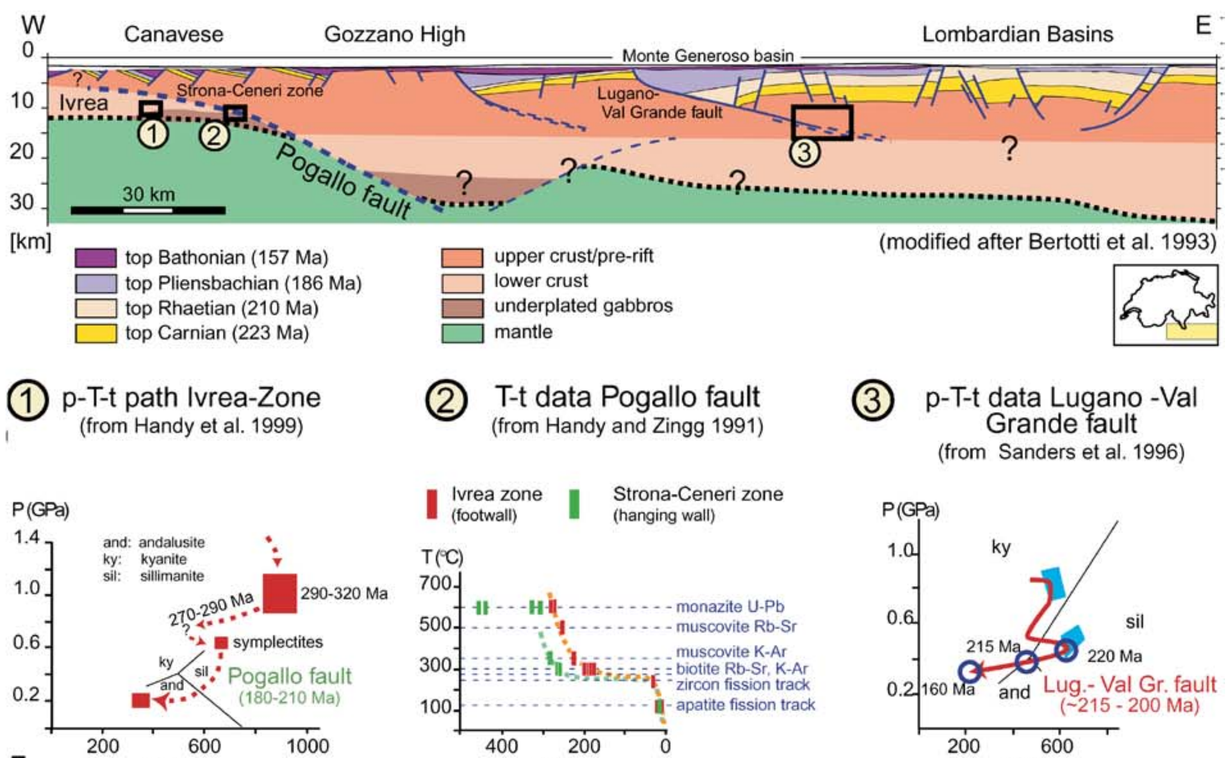
Little is known about the pre-rift conditions at the Iberia/Newfoundland margins, which means that they are poorly constrained for these margins. More data are available concerning the pre-rift stratigraphic evolution of the European/Briançonnais and Adriatic margins. The data show a very widespread late Carboniferous to Permian event recorded by clastic sediments and volcanic rocks, which were deposited in fault-bounded basins commonly interpreted as being formed in a transtensional/transpressional tectonic setting (Handy and Zingg 1991). Later Triassic depositional environments changed across the future Alpine realm from the southeast towards the northwest, from deeper marine to platform carbonates to sabkha environments, which graded into domains dominated by siliciclastics, marking the occurrence of emerged areas. Apart from local exceptions, the stratigraphic record in any one place shows a general evolution from subaerial deposits to shallow marine environments leading to dolomites and limestones, which are locally interbedded with sabkha deposits. The sedimentary sequence is generally thicker in the east and south (central Austroalpine and southern Alps, 1 to 5 km) and thinner in the northwest (lower Austroalpine and Briançonnais less than 500 m) and much thinner, locally only a few tens of metres thick, in the proximal European margin. These observations suggest that the pre-rift continental crust was isostatically equilibrated after Permian time, assuming a 30-km- thick continental crust at the onset of rifting (Fig. 6a).

Volcanic activity occurred during the Triassic, as indicated by tuff horizons and local dolerites within the Triassic sedimentary sequence. Within the Lower to lower Middle Jurassic syn-rift sedimentary sequences no evidence for volcanic activity has been found yet, indicating that rifting was, at least at the surface, not accompanied by magmatic activity.

Further constraints on the conditions before the onset of rifting can be obtained from an exhumed pre-rift fossil crust/mantle boundary, which is welded by a mid-Permian gabbro intrusion in Val Malenco (Hermann et al. 1997). 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 about 550ºC at 0.9 to 1 GPa, corresponding to


(a) pre-rift situation
strength (MPa)
500 1500
upper crust
lower crust
"strong" mantle
"weak" mantle
60 km
(b) lithospheric necking
thickness of the basins not to scale
60 km
(c) lithospheric detachment faulting
60 km
(d) onset of sea-floor spreading
60 km
[km]
mantle petrofabric
1350 C isotherm
future fault
active fault
ancient fault
MOR-basalts and gabbros
upper/lower crust
oceanic and syn-rift lithospheric mantle
pre-rift gabbros
asthenospheric mantle
subcontinental mantle (strong/weak)
successively younger faulting systems
fault system Ia
fault system Ib
fault system II
fault system III
fault system at MOR

6


W
E
Canavese
Gozzano High
Lombardian Basins
Monte Generoso basin
Ivrea
Strona-Ceneri zone
Lugano-Val Grande fault
Pogallo fault
30 km
[km]
(modified after Bertotti et al. 1993)
top Bathonian (157 Ma)
top Pliensbachian (186 Ma)
top Rhaetian (210 Ma)
top Carnian (223 Ma)
upper crust/pre-rift
lower crust
underplated gabbros
mantle
1 p-T-t path Ivrea-Zone
(from Handy et al. 1999)
2 T-t data Pogallo fault
(from Handy and Zingg 1991)
3 p-T-t data Lugano -Val Grande fault
(from Sanders et al. 1996)
P (GPa)
and: andalusite
ky: kyanite
sil: sillimanite
290-320 Ma
270-290 Ma
symplectites
Pogallo fault
(180-210 Ma)
T (°C)
Ivrea zone (footwall)
Strona-Ceneri zone (hanging wall)
monazite U-Pb
muscovite Rb-Sr
muscovite K-Ar
biotite Rb-Sr, K-Ar
zircon fission track
apatite fission track
(Ma)
ky
sil
and
215 Ma
220 Ma
160 Ma
Lug.- Val Gr. fault
(~215 - 200 Ma)

7

a 30 to 35 km thick crust. Applying these conditions to the overall pre-rift crust-mantle boundary, a classical four-layer rheological model with a brittle upper crust, a ductile lower crust, a brittle upper mantle, and a plastic lower mantle can be proposed for the onset of rifting (Fig. 6a).

Onset and early rifting

The age of the onset of rifting and its duration are poorly constrained and depend on the way rifting is defined (e.g. Wilson et al. 2001a). For the Iberia/Newfoundland margins interpretations of the onset of rifting are mainly based on geometrical arguments, i.e. determination of geometries related to syn-sedimentary normal faulting such as onlaps onto rotating hanging walls and thickening into footwalls. Because such features are largely absent in the example of the Iberia margin (Wilson et al. 2001a), a clear determination of the onset and age of rifting is difficult.

In the Alps, the onset of rifting is determined by abrupt changes in facies and/or thickness of sediments. Based on the stratigraphic record, rifting has been interpreted to start in either Permian (Winterer and Bosellini 1981), Late Triassic (Bertotti et al. 1993), or Early Jurassic (Bernoulli et al. 1990) times. Although the formation of Permian basins was related to extension/transtension and magmatic activity, this event apparently did not lead to a thinning of the crust to less than 30 km in the Alpine realm. Therefore, an extensional/transtensional regime, similar to that observed in the metamorphic core complexes in the southwestern United States, may be more appropriate to explain the Permian evolution. Müntener et al. (2000) showed that the Permian event was followed by isobaric cooling, lasting over more than 50 Ma, which is compatible with the little tectonic activity documented in the eastern Alpine realm during Triassic time. Thus, the Permian event is unlikely to represent the onset of rifting, but may have played an important role in the predetermination and localization of later rift structures.

The first evidence of strong, differential subsidence which is associated with faulting is recorded from the Late Triassic. While normal faulting affected the whole Alpine realm, subsidence was more pronounced in the southeastern part of the area, i.e. on the Adriatic margin (Bertotti et al. 1993; Berra 1995) and weaker on its northwestern side (Chevalier 2002). Sedimentation kept pace with subsidence so that faulting resulted only locally in a major morphological expression (Bertotti et al. 1993).

Fig. 6a–d Conceptual model showing the temporal and spatial evolution relative to a fixed left-hand edge from: (**a**) the pre-rift situation, to (**b**) initial rifting (lithospheric necking), to (**c**) final rifting (lithospheric detachment faulting), and (**d**) onset of sea-floor spreading (modified after Whitmarsh et al. 2001)

Fig. 7 Margin architecture and depocentre migration (modified from Bertotti et al. 1993) and p-T-t evolution (for references see figure) recorded in the "Ticino" transect. For further discussion see text

Only from the Early Liassic time onwards (Hettangian to Sinemurian), large rift basins formed all over the margin and led to a very uniform structural evolution in both the European/Briançonnais and Adriatic margins. In the Bourg-d'Oisans area, in the proximal European/Briançonnais margin, Chevalier (2002) was able to date and quantify the strain rate along single faults. The results show that rifting was diffuse and distributed before extension became localized along a few major faults. These faults operated independently of one another and most of the movement along them occurred over very short time periods. Rifting in the proximal parts of the future margin ceased in the Middle to Late Liassic, however, some of the faults became inactive even earlier and were sealed by Upper Sinemurian sediments (Eberli 1988; Conti et al. 1994). Later rifting became localized within the area of the future break-up.

Final stage of rifting, break-up and transition to sea-floor spreading

The temporal and spatial evolution of rifting is recorded well in the Alps by syn- and post-rift sedimentary sequences becoming younger oceanwards (Froitzheim and Eberli 1990). Basin architecture shows that earlier and more proximal basins are larger and bounded by listric faults soling out at mid-crustal levels, whereas the more distal basins are younger, smaller, and in some cases, truncated by low-angle detachment faults or reflections, respectively. In the Iberia margin, ODP failed, except at three sites, to penetrate syn- or pre-rift sequence (i.e. Tithonian sediments at Sites 901, 1065 and 1069). Similar sedimentary facies and thickness variations across the distal margins in Iberia and the distal Adriatic margins suggest a comparable sedimentological evolution and in turn a similar isostatic response during final rifting (Wilson et al. 2001a).

The stratigraphic record of latest rifting is well preserved in the Err nappe in the "Grischun" transect (Fig. 3c). The record shows Middle Liassic hemipelagic cherty limestones (Agnelli Formation; e.g. Manatschal and Nievergelt 1997), which terminate with a typical submarine hardground of Pliensbachian age, which is unconformably overlain by deep-water clastics, interbedded with hemipelagic marls (Saluver Formation, e.g. Finger et al. 1982). Unfortunately, the age of these sediments is poorly constrained. They must be younger than the Pliensbachian hardground and older than the overlying upper Middle Jurassic Radiolarite Formation, which gives a maximum duration of 30 Ma. The reconstruction of the basin geometry in the Err domain is based mainly on facies and thickness distributions and shows that the basins were less than 5 km wide and their sedimentary fill was in the order of some few hundred metres. In many places, syn-rift sediments directly overlie low-angle detachment faults exposed at the sea floor. The occurrence of clasts of characteristic black gouges derived from detachment surfaces in the syn-rift sediments, as well as the observation that syn-rift sediments locally seal the detachment faults, clearly demonstrates that at least the youngest syn-rift sediments were deposited during or even after detachment faulting and mantle exhumation at the sea floor. Thus, the syn-rift sediments in the distal margin may record a complex polyphase tectonic evolution.

Evidence of the stratigraphic evolution of the distal Adriatic margin contrasts with that found in the distal conjugate margin in the area of Briançonnais, where Lemoine et al. (1986) demonstrated that a change from initial subsidence associated with block tilting to later uplift occurred during a late stage of rifting. The uplift leading to the emergence, erosion, and karstification of the Triassic carbonate platform occurred simultaneously with subsidence beneath the CCD in the conjugate Adriatic margin. This observation is one of the strongest arguments for a strong asymmetry between the conjugate distal pairs of margins bounding the Liguria-Piemonte basin.

The age of continental break-up has been classically dated in present-day margins by the first sea-floor spreading magnetic anomaly and/or a break-up unconformity. In the Alps, the first sediments sealing oceanic and continental rocks in the OCT are radiolarian cherts that can be dated biostratigraphically and so enable the determination of the timing of the break-up. Radiometric ages dating the emplacement (crystallization ages) and exhumation (cooling ages) of magmatic rocks in the ZECM may also indicate the timing of this event. Such dating is based on the assumptions that: (i) break-up of continental crust is immediately followed by sea-floor spreading; (ii) rifting is of the same age across and along the whole margin; and (iii) the sediment architecture in the OCT is simple. All three assumptions have to be re-evaluated in magma-poor rifted margins and will be discussed in a later part of this paper.

Strain evolution during initial rifting

Rift structures in the proximal margin

Deformation structures related to the initial stage of rifting have been described from many locations in the Alps (e.g. Lemoine et al. 1986; Froitzheim 1988; Conti et al. 1994), but the best documented examples are observed in the "Ticino" transect in the southern Alps (Bertotti 1991) (Figs. 3d, 7). Along this transect, several fault-bounded basins exist, among which is the Monte Generoso basin. The Lugano-Val Grande fault bounding this basin cuts across the pre-rift sedimentary cover into the basement over a horizontal distance of more than 30 km. Along this fault, the mode of deformation changes from brittle to ductile as indicated by the transition from cataclasites to quartz mylonites along the fault. The geometrical relationship between the fault and the depositional geometry in the overlying sediments shows that this fault had a listric geometry (Bertotti 1991). Based on microstructural investigations, Bertotti (1991) demonstrated that the fault soled out in greenschist facies conditions, i.e., at about 10 to 15 km depth.

This situation is conspicuously similar to that observed in the Jeanne d'Arc basin in the Newfoundland margin, where fault geometry and architecture of the rift basins are well imaged in reflection seismic profiles (e.g. Keen et al. 1987). For this basin it can be shown that the bounding fault soles out at middle to lower crustal levels and that the Moho was not affected by faulting, but was only regionally up-warped beneath the extended zone. The similarity between these two examples leads to the suggestion that this mode of deformation, which includes a decoupling between deformation in the upper mantle and upper crust, may be also found below other proximal rift basins elsewhere in the European and Adriatic margins. Moreover, this mode of deformation is compatible with the results obtained by numerical and analogue modelling (Harry and Sawyer 1992; Michon and Merle 2003).

Rift structures in the distal margin

Observations in the Alps and off Iberia show that the rift structures observed in the distal margin cannot be explained in the same way as those in the proximal parts of the margin. Because the distal margin has been strongly overprinted by the final rifting event, structures related to previous stages of rifting are difficult to observe. In the model (Fig. 6), deformation in the distal margin is controlled by necking of the lithospheric mantle, which is associated with major detachment structures dipping towards the necking point and leading to a local thinning of the overlying crust. It is not understood why the position of the neck in the upper mantle is offset from the areas of extension at the surface, as observed in the Alps. Both numerical and analogue models show, however, that necking in the lithospheric mantle is commonly offset from maximal thinning in the continental crust (Harry and Sawyer 1992; Michon and Merle 2003).

Faults, analogous to those drawn in the model (Fig. 6b) in the distal margin are the Pogallo fault in the Ivrea zone (Handy and Zingg 1991) (Fig. 7) and the Margna fault in the Malenco area (Müntener and Hermann 2001) (Fig. 3c). Both faults are situated in a distal position within the Adriatic margin, dip eastwards, i.e. beneath the Adriatic margin, and separate lower crustal rocks in the footwall from upper crustal rock in the hanging wall. P-T-t paths of hanging wall and footwall rocks of the Pogallo fault show that this fault was active during Liassic time (simultaneously with the Lugano-Val Grande fault) and that the crust was thinned to about 10 km (0.3 to 0.4 GPa) along this fault (Handy and Zingg 1991; Fig. 7). Villa et al. (2000) and Müntener and Hermann (2001) showed, using p-T-t data, that the Margna fault thinned the crust to about 0.3 to 0.4 GPa (~ 10 km) during Early Jurassic time. Thus, there is independent evidence that within the distal margins, crustal-scale faults thinned the crust to about 10 km in an early stage of

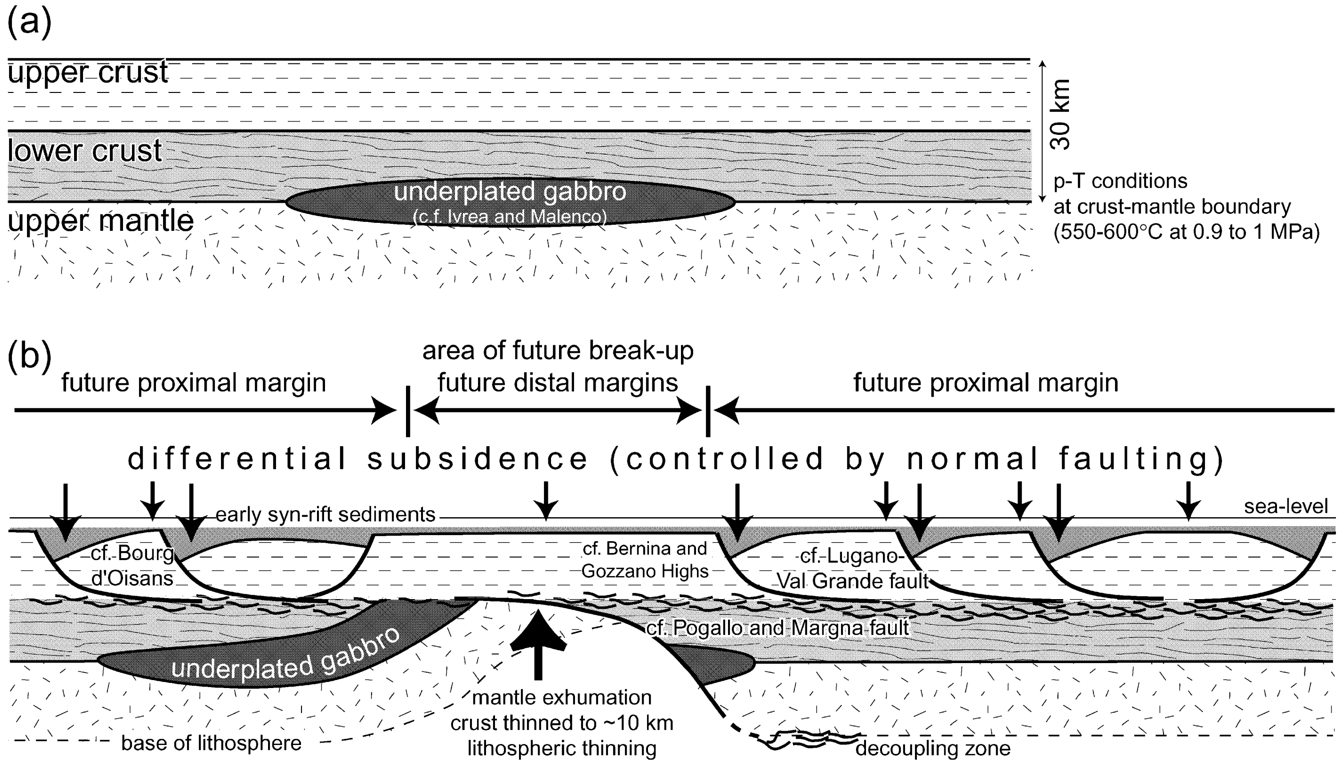


Fig. 8a–b Tectonic evolution during initial rifting as deduced from observations and data from the Alpine margins: **(a)** pre-rift situation; **(b)** situation during initial rifting

rifting. In a later part of this paper, I will discuss the importance of these structures for the evolution of the margin as well as their relationship with the oceanward dipping low-angle detachment faults observed in the OCT.

In both the "Ticino" and "Grischun" transects (Fig. 3c, d), depocentres were located during activity along the Pogallo and Margna faults within fault-bounded basins in the proximal parts of the margin (e.g. Monte Generoso basin in the Ticino; Fig. 7) while sedimentation in the distal margin was starved and only little evidence for the formation of rift basins within this part of the margin exists. Transition from platform to hemipelagic cherty limestones shows, however, that the distal margin was subsiding during this early stage of rifting. Thus, the surface response during an initial stage of rifting was different in the distal and proximal margins. This is particularly well recorded in the "Grischun" transect. Here, hemipelagic cherty limestones show a rather constant thickness across the entire distal margin (Agnelli Formation; Manatschal and Nievergelt 1997) demonstrating a lack of contemporaneous fault movement. In contrast, sediments of the same age fill fault-bounded basins in more proximal parts of the basin (Ortler basins in Fig. 3c; Eberli 1988). Thus, the Pogallo and the Margna faults did not produce much fault-related morphology in the distal margin, despite the fact that the crust has been thinned to about 10 km along these faults.

Comparable situations to those described above may exist in the Porcupine Basin off southwest Ireland and in the Galicia Interior basin. Reston et al. (2001) and Pérez-Gussinyé et al. (2003) demonstrated that these basins are floored by continental crust that was thinned to less than 12 km without producing a major fault-related morphology and/or tilted blocks. Later in this paper, I will propose that the C reflection in the southern Iberia Abyssal Plain may represent a structure analogous to the Margna and the Pogallo faults. These faults can explain local crustal thinning of the margin to about 10 km during an early stage of rifting.

Constraints on the thermal evolution during initial rifting can be obtained from p-T-t data of basement rocks in the southern Alps, where the Alpine metamorphic overprint was weak (<200°C). Cooling ages obtained from rocks derived from the footwall of the Pogallo fault (Fig. 7) and the Margna fault (Fig. 3c) together with thermo/barometric data clearly show that cooling was related to exhumation and major thinning of the crust to about 0.3 to 0. 4 GPa (i.e. about 10 km) (Fig. 8b). Cooling ages ranging from 220 to 180 Ma have been obtained from the footwall and hanging wall of the Lugano-Val Grande fault (Sanders et al. 1996) (Fig. 7). These cooling ages, however, cannot be related to exhumation along the Lugano-Val Grande fault, because marine sedimentation continued during Liassic time on both the hanging wall (about 4 km thick) and the footwall (about 100 m thick) and the facies of the sediments deposited on the footwall change from shallow to deeper marine conditions during the same time interval. Thus, in the case of the Lugano-Val Grande fault, both footwall and hanging wall were subsiding relative to sea level but at different rates (Fig. 8b). Therefore Bertotti and 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. This interpretation ignores, however, the fact that cooling ages of Late Triassic/Early Jurassic age are found over wide areas in the Alpine realm, suggesting that rifting was associated with the cooling of the extending lithosphere.

Together, these observations show that during the initial stage of rifting distal and proximal parts of the margins behaved differently. This leads to the following conclusions:firstly, the final place of break-up was determined early in the evolution of the margin and was probably controlled by inherited lithospheric heterogeneities (e.g. Permian underplated gabbros in Fig. 8a); secondly, final rifting was localized in places were the crust was thinned to less than 10 km during an initial stage of rifting (Fig. 8b).

Strain evolution during late-stage rifting

One of the major discoveries on rifted margins in the past two decades was the observation at sea of low-angle reflections (de Charpal et al. 1978; Boillot et al. 1987; Reston et al. 1995) and on land of low-angle detachment structures (Froitzheim and Eberli 1990; Florineth and Froitzheim 1994; Manatschal and Nievergelt 1997) in the OCT of magma-poor margins. Although these structures are thoroughly described and their importance for the exhumation of subcontinental mantle generally accepted, little is known about how they evolve, the mechanics associated with these faults, and the processes controlling their formation. Examples of exposed rift-related detachment faults have been described from the ancient margins in the Alps (Froitzheim and Eberli 1990; Florineth and Froitzheim 1994; Froitzheim and Manatschal 1996; Manatschal and Nievergelt 1997) and from the Red Sea margin (Talbot and Ghebreab 1997). An example of an active detachment fault related to rifting and opening of a small ocean basin exists in the Woodlark-D'Entrecasteaux rift system in the Western Pacific (Taylor et al. 1999b). The study of this active system yielded important information on the kinematic and geometric evolution of rift-related detachment systems. Abers et al. (1997) demonstrated that earthquakes associated with normal faulting occurred at depths of less than 10 km in the brittle field in an area with a seismically well-imaged reflection dipping at only 25°-30° (Abers 2001). This reflection was drilled recently during ODP Leg 180 where it crops out at the seafloor (Taylor et al. 1999a). The drilling showed that it represents a fault marked by a gouge zone. This shows that, in contrast to Andersonian mechanics (Anderson 1951), normal faults can be active at angles of 25°-30° within the brittle upper crust. GPS measurements between two islands located on the footwall and hanging wall, respectively, indicated very fast extension across the system (~30 mm/yr; Tregoning et al. 1998). Compared to many fossil detachment faults observed on land, which have dip angles of less than 10°, the normal fault in the Woodlark-D'Entrecasteaux rift is still relatively steep. Evidence of earthquakes along subhorizontal detachment faults (<10–15° dip) still has to be found (Abers 2001). In the following, I will describe two exposed subhorizontal (<15°) detachment systems preserved in the Alps and one drilled and seismically imaged in Iberia.

Detachment structures in the Err/Platta OCT (Grisons)

In the northwestern Adriatic margin, exhumation of subcontinental mantle at the sea floor and final emplacement of continent-derived extensional allochthons onto mantle and continental basement rocks occurred along low-angle, top-to-the-ocean detachment faults. The geometry of this detachment system within the thinned distal margin is spectacularly exposed and preserved over 6 km in the area of Piz d'Err-Piz Bial in the Err nappe (Froitzheim and Eberli 1990; Manatschal and Nievergelt 1997) (Figs. 9 and 10a,b). Excepting a few gaps due to later erosion, the footwall of one of the detachment faults, the Err detachment, can be traced over 18 km parallel to transport direction. Its hanging wall is formed by fault blocks of continental basement and pre-rift sediments, tilted toward the east and the west, i.e. along west- and east-dipping high-angle faults. The fault blocks, a few kilometres to some hundred metres across, are systematically truncated at their base by the Err detachment and are interpreted as extensional allochthons. An older detachment is preserved at Piz Jenatsch. This detachment fault is cut by the younger and lower Err detachment. Manatschal and Nievergelt (1997) determined a displacement of more than 10 km for the latter. Quartz mylonites are not observed along this detachment fault, indicating that, where observed, this structure formed under low-temperature conditions (<300°C) in the upper continental crust. Profiles across the footwall of the Err detachment exhibit a transition from undeformed granite into green cataclasites, which are capped along a sharp and well-defined horizon by a characteristic black-fault gouge. Manatschal (1999) demonstrated the importance of fluid- and reaction-assisted deformation processes as the dominant mechanism to produce such weak fault zones within a brittle upper crust.

Relics of a detachment system are also found in the Platta nappe (Manatschal and Nievergelt 1997). Here, the characteristic black gouge occurs at the base, or within blocks, of continental basement overlying serpentinized mantle peridotites. These blocks are covered by syn- to post-rift sediments and are interpreted as extensional allochthons overlying tectonically exhumed mantle (Figs. 9 and 10c).

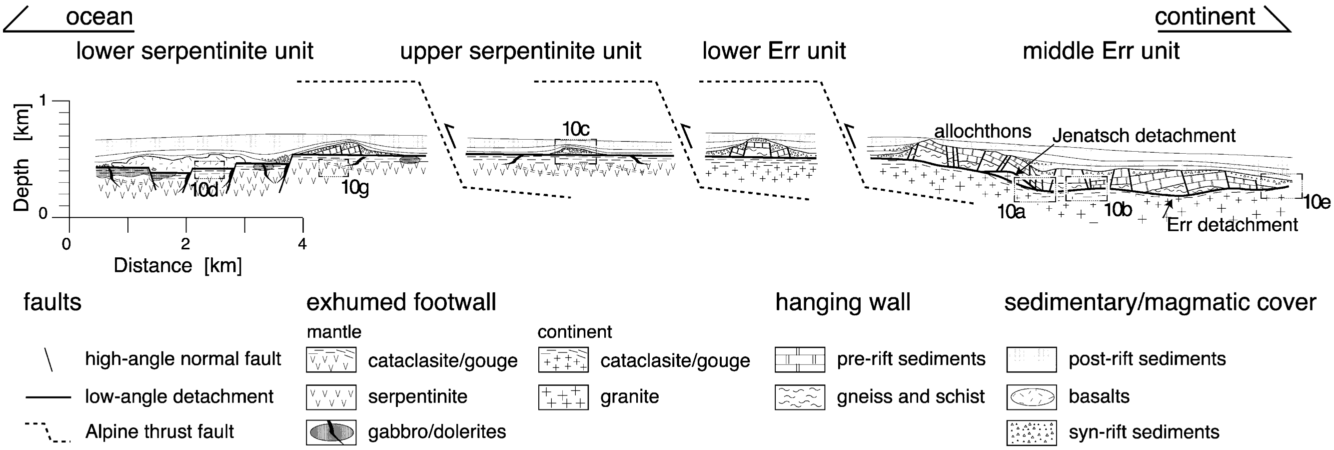
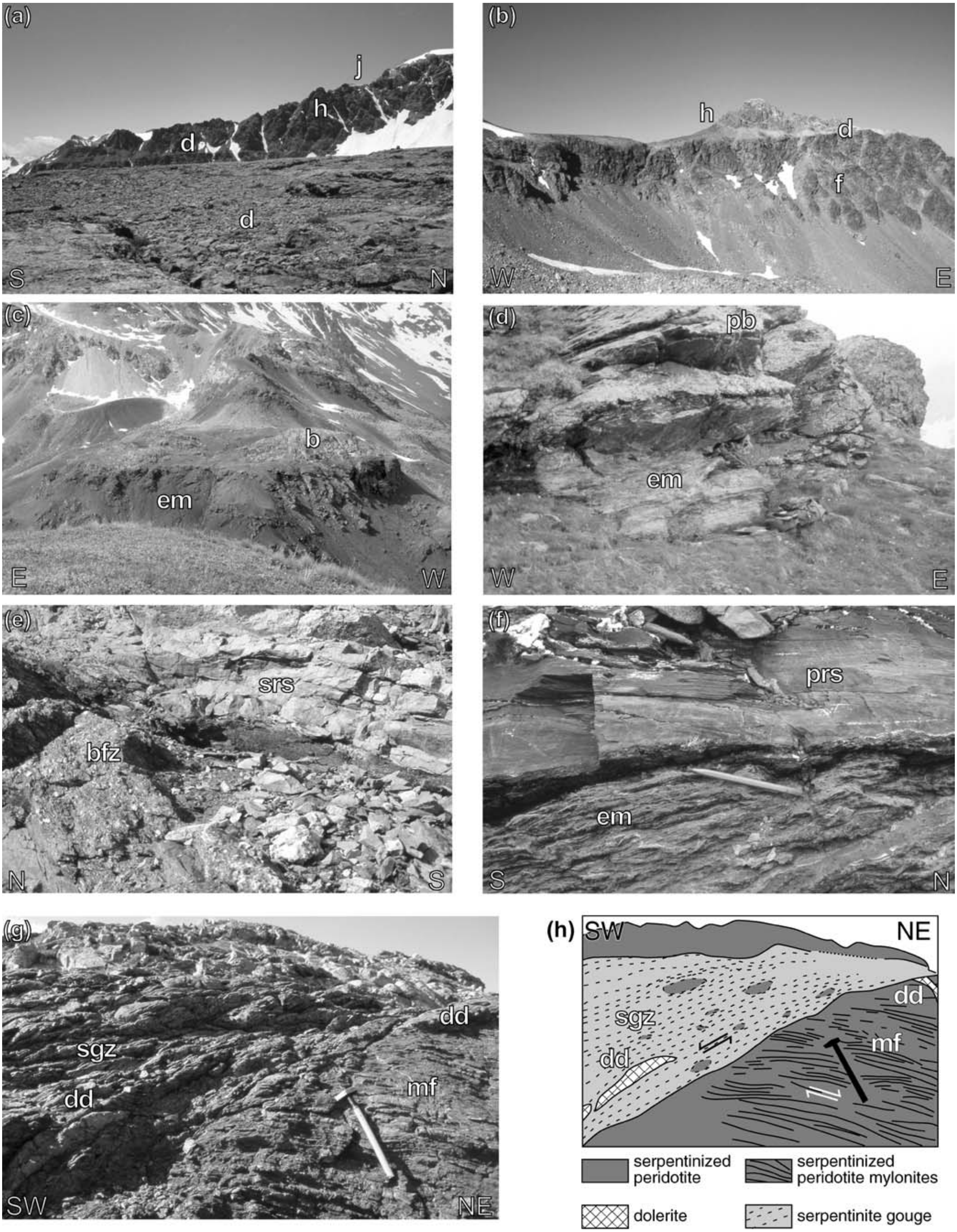


Fig. 9 Tentative palinspastic section showing the observed relationships between the Err detachment and associated sediments, exhumed mantle rocks, and intrusive and extrusive magmatic rocks in the ocean-continent transition preserved in the Err and Platta nappes in Grisons (SE Switzerland). Squares with numbers refer to photographs shown in Fig. 10 (modified after Manatschal et al. 2003)

Deformation in the mantle rocks indicates a polyphase evolution. Granulite-facies (i.e. anhydrous) deformation is characterized in the mantle units that were originally next to the continent by a penetrative spinel foliation and subparallel pyroxenites. Desmurs et al. (2001) demonstrated that, going oceanwards, the high-temperature, granulite-facies deformation changes: the rocks look less deformed, and pyroxenites are rare. This is valid, however, only for granulite-facies deformation. All other types of deformation appear to be similar throughout the ZECM. Localized high-temperature shear zones formed by peridotite mylonites to ultramylonites, are commonly inclined relative to younger serpentinite structures, and show a top-to-the-continent sense of shear. Tectonites reflecting deformation and hydration processes occurring under lower amphibolite facies and still lower metamorphic conditions are commonly preserved along the top of the mantle and include serpentinite mylonites, cataclasites, and gouges. These “colder” shear zones constantly show a top-to-the-ocean sense of shear. The relationships between the “cold” top-to-the-ocean and the “high-temperature” top-to-the-continent shear zones are spectacularly exposed in an outcrop at Sur al Cant south of Bivio in the Platta nappe (Fig. 10g,h). In this outcrop, the high-temperature top-to-the-continent mylonites are cut by colder and localized top-to-the-ocean serpentinite gouges. The gouge zone also cuts a rodingitized basaltic dyke, indicating that magmatic activity had already initiated when the top-to-the-ocean shear zone was active. This observation is fundamental, because it demonstrates that: (1) late, “cold” mantle exhumation was associated with magmatic activity; and (2) “high-temperature” (>700°C) and later hydrous “cold” deformation processes cannot be explained within the same kinematic framework.

Further information on the relationship between magmatism and detachment faulting can be obtained from the study of the deformation of the gabbros which were intruded at 161 Ma, i.e. during break-up, into partially serpentinized mantle rocks (Schaltegger et al. 2002). Microstructures in the gabbros reveal a deformation history ranging from syn-magmatic to brittle conditions. Because gabbros occur as clasts in pillow and sedimentary breccias which are overlain by middle-Upper-Jurassic radiolarian cherts, these gabbros had to be exhumed, probably along detachment faults, at the sea floor soon after their crystallization. In the gabbros, syn-magmatic to high-temperature deformation is localized in small shear zones, which are locally restricted and commonly not interconnected. The gabbros, therefore, were intruded in a tectonically active setting which was relatively cold, so that soon after recrystallization and cooling deformation became localized in the surrounding serpentinized mantle rocks. This may also explain why almost all contacts between mantle rocks, gabbros, and dolerites are strongly sheared, which is partly, but not exclusively, related to later Alpine reactivation.

Fluid flow, associated with hydration reactions and mass transfer, played an important role during final rifting. This is demonstrated by serpentinization of the mantle and the retrograde tectono/metamorphic evolution within shear zones ranging from amphibolite to sea-floor conditions. Manatschal et al. (2000) proposed thermally driven convection as a driving force for fluid flow during break-up in the OCT. Based on structural, mineralogical, and geochemical investigations of the fault rocks derived from the Err detachment, they demonstrated that marine fluids penetrated and interacted with mantle rocks before being captured and channelized along faults towards the sea floor. In the faults, the fluids triggered cataclastic deformation and mineral reactions leading to a complex interaction between deformation, fluid-flow, and mineral reactions and resulting in transient variations of the permeability structure of the fault. Fluid flow along the fault zone was associated with dissolution, transport, and precipitation processes resulting in a gain or a loss of dif-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dd
serpentinized peridotite
serpentinized peridotite mylonites
dolerite
serpentinite gouge

ferent elements. Mass-balance calculations based on major and trace element and stable isotope data permit the identification of a mantle source for the fluids percolating across the detachment, leading to an enrichment of mantle-derived elements such as Cr and Ni along the faults in the continental crust. This shows that large-scale detachment structures like the Err detachment were pathways for fluids derived from the mantle. Frueh-Green et al. (2002) demonstrated that the serpentinization of mantle rocks resulted in high concentrations of Ni and Cr within fluids. Thus, the enrichment of these elements along the Err detachment, active within the extending continental crust, is indirect evidence that this fault cut into the mantle and that serpentinization may have occurred at an early stage, already before the mantle was exposed at the sea floor.

The relationship of the detachment-system to the sediments is preserved in the continental units northwest of Piz d'Err and at several places elsewhere in the Err nappe (Manatschal and Nievergelt 1997) (Fig. 10e). The syn-rift sediments overlie the exhumed detachment surface at a low angle (<20°) of discordance. Exposure of the detachment surface at the sea floor is also supported by the occurrence of clasts of green cataclasites and black gouge derived from the Jurassic detachment in the syn-rift sediments (Froitzheim and 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.

The tectonized mantle in the ZECM is capped by tectono-sedimentary breccias and extensional allochthons of continent-derived material (Fig. 10c), as well as pillow basalts (Fig. 10d) and post-rift sediments. Different types of tectono-sedimentary breccias, referred to as ophicalcite in Alpine literature, occur in and include mainly serpentinite clasts embedded in a fine-grained, microsparitic, calcite matrix. The fabric of these breccias varies considerably from the serpentinite host rock with fractures filled by red limestone and/or white sparry calcite, to clast-supported breccias with in-situ fragmented serpentinite clasts (Bernoulli and Weissert 1985; ophicalcites I of Lemoine et al. 1987), to coarse, unsorted, matrix-supported breccias with fragments of serpentinite, gabbro, and continent-derived basement rocks and pre-rift sediments (ophicalcites II of Lemoine et al. 1987). Geopetal infill of sediment into crevasses and pockets of 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 sea floor 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 (Desmurs et al. 2001).

Fig. 10a – g Photographs from the OCT in the Alps (for location of photographs 10(**a**) to (**e**) and 10 (**g**) see Fig. 9, and for photograph 10(**f**) see Fig. 11a). (**a** and **b**) Err detachment at Piz Laviner (Err nappe). (**a**) View of the fault surface of the Err detachment (*d*) at Piz Laviner looking towards Piz Jenatsch, where the continuation of the Err detachment (*d*) can be observed. In the hanging wall (*h*) of the Err detachment a second fault can be observed which is the Jenatsch detachment (*j*). (**b**) View of the Err detachment (*d*) at Piz Laviner separating a cataclastically deformed granitic basement in the footwall (*f*) from an allochthon composed of Triassic dolomites (to the right) and schists (to the left) in the hanging wall (*h*). (**c**) Exhumed mantle (*em*) overlain by continent derived breccias (*b*) at Parsettens (Platta nappe). (**d**) Tectonized and exhumed mantle (*em*) overlain by pillow basalts (*pb*) at Falotta (Platta nappe). (**e**) Syn-rift sediments (*srs*) onlapping onto the exhumed detachment formed by a brittle fault zone (*bfz*) in the continental Err nappe at Fuorcla Cotschna. (**f**) Post-rift sediments (*prs*) stratigraphically overlying a tectonized, exhumed mantle (*em*) in the Tasna OCT. (**g** and **h**) Outcrop at Al Cant (Platta nappe) preserving the relationships between high- and low-temperature deformation in the mantle rocks and its relation to magmatic processes. A strongly localized serpentinite gouge zone (*sgz*) with a top-to-the-SW (top-to-the-ocean) sense of shear cut high-temperature peridotite mylonites showing a top-to-the-NE (top-to-the-continent) sense of shear. A dolerite dike (*dd*) cut the high-temperature mylonite (*mf*) foliation, but is truncated by the serpentinite gouge zone (*sgz*) and occurs as clasts in the latter

Detachment faults within an exposed OCT in the Tasna nappe (Grisons)

Within the Tasna nappe in the Alps of eastern Switzerland (Fig. 11a), a remnant of a former OCT is preserved (Florineth and Froitzheim 1994). This locality is at present the only place known where the lateral transition from continental crust to exhumed subcontinental mantle is exposed within one single outcrop and not disrupted by later thrusting. The outcrop extends over five kilometres along a SSW-NNE trending mountain ridge. Its general structure can be described as a wedge of heterogeneous and strongly tectonized lower continental crust, consisting of an assemblage of amphibolites, gabbros, tonalites, and migmatites thinning towards the north (Fig. 11a). The lower continental crust is bounded by two detachment faults, the Upper Tasna Detachment (UTD) and the Lower Tasna Detachment (LTD) (Froitzheim and Rubatto 1998) (Fig. 11a). The LTD separates the lower crustal rocks in the hanging wall from strongly serpentinized subcontinental mantle rocks in the footwall. The UTD is the trace of a low-angle detachment capping the continental basement to the SSW and the mantle rocks to the NNE. It is locally associated with polymictic tectono-sedimentary breccias and stratigraphically covered by shales which in turn are overlain by Lower Cretaceous calcareous siltstones and sandstones intercalated with breccia layers. The UTD has been interpreted by Florineth and Froitzheim (1994) to cut the LTD and to exhume the lower crustal and mantle rocks to the sea floor. The observation that the UTD is sealed by Lower Cretaceous sediments (Fig. 10f) confirms its pre-Alpine age. In the Tasna OCT, detachment faults first localized along the crust-mantle boundary (e.g. LTD) and only later did other detachment faults (UTD) cut into the mantle, thereby

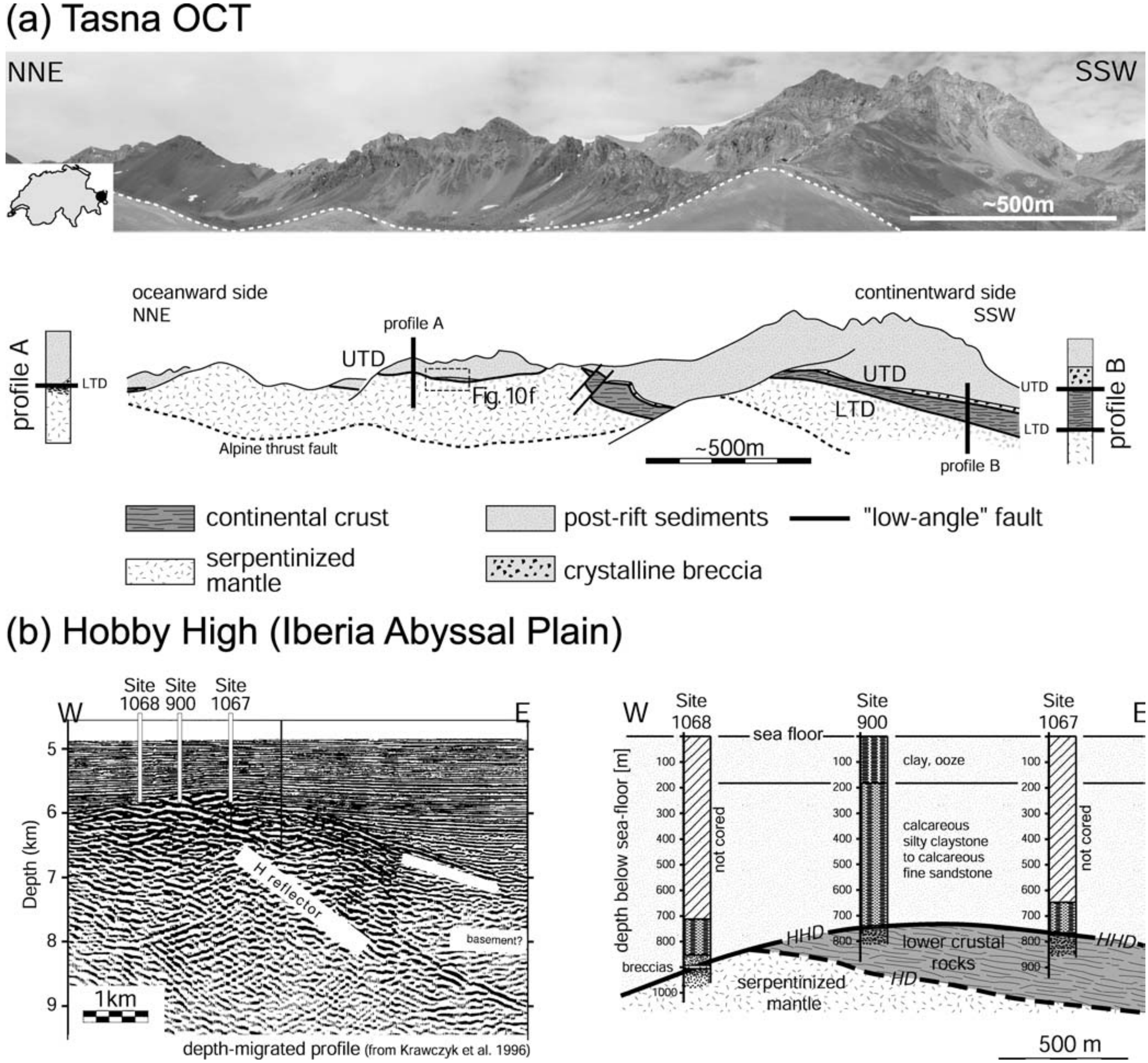


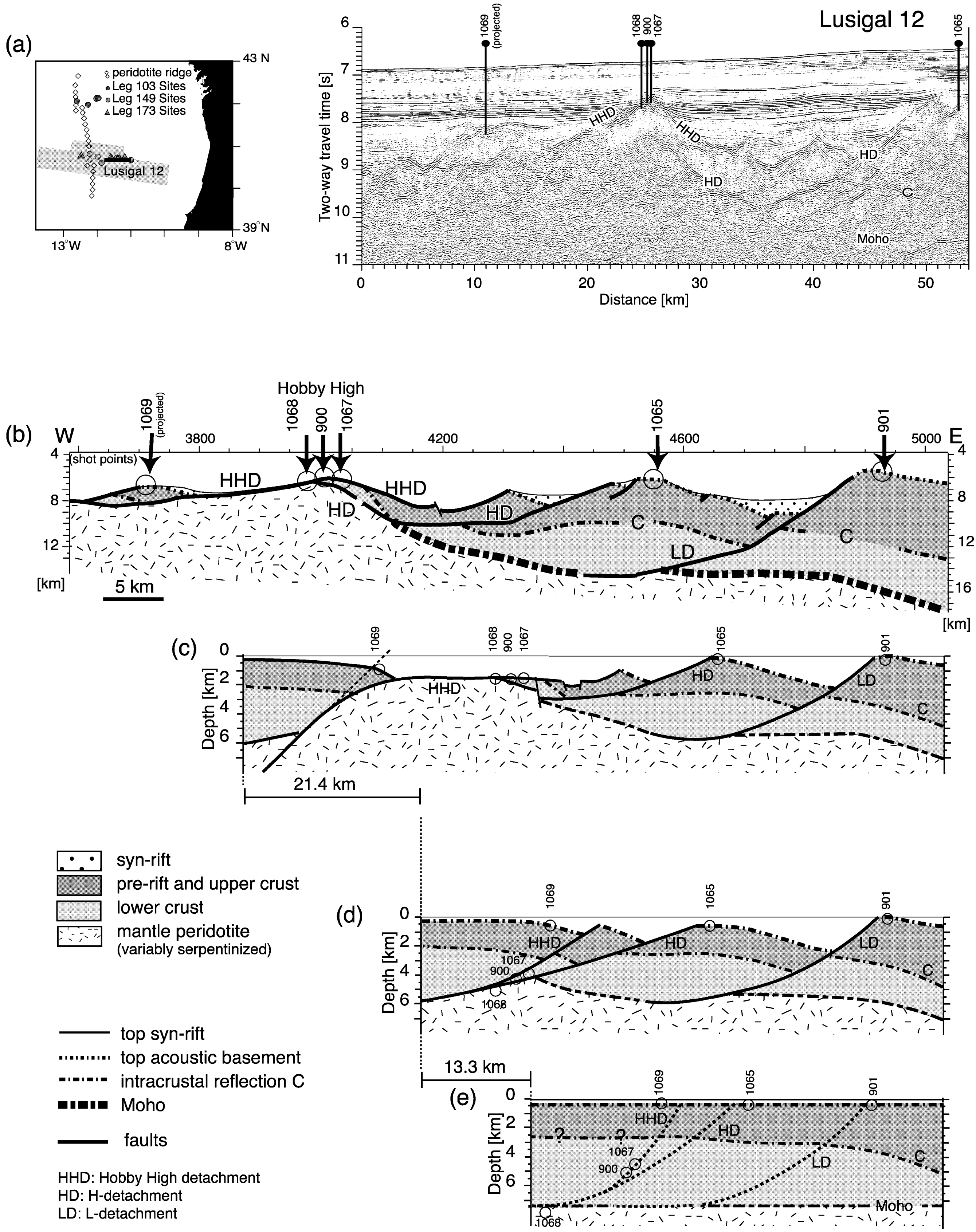
Fig. 11 (**a**) An ocean-continent transition exposed in the Tasna nappe in Grisons, SE Switzerland (modified after Florineth and Froitzheim 1994). *UTD*: Upper Tasna detachment; *LTD*: Lower Tasna detachment. (**b**) Depth-migrated seismic section (Lusigal 12, Krawczyk et al. 1996) and borehole profiles (ODP Sites 900, 1067, and 1068) drilled across Hobby High in the southern Iberia Abyssal Plain (modified from Hölker et al. 2002b)

coupling crustal and mantle deformation. This is compatible with the findings described earlier in this paper from the Err detachment but is also very similar to the situation observed through drilling and seismical imaging across Hobby High in the Iberia Abyssal Plain (Fig. 11b).

Seismically imaged and drilled detachment structures in the Iberia Abyssal Plain

In the Iberia Abyssal Plain, several strong reflections have been observed and described by Beslier et al. (1995),

Fig. 12 (**a**) Map of the Iberia margin showing location of Lusigal 12 seismic section relative to the peridotite ridge and ODP Sites and part of the Lusigal 12 seismic profile. (**b**) 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. (**c to e**) Evolution of faulting as determined from kinematic inversion of the interpreted Lusigal 12 profile shown in (**b**): (**c**) mantle exhumation along downward-concave fault *HHD*; (**d**) crustal extension accommodated by upward-concave faults *LD*, *HD*, and *HHD*; (**e**) 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 text



(a)
peridotite ridge
Leg 103 Sites
Leg 149 Sites
Leg 173 Sites
Lusigal 12
43 N
39 N
13°W
8°W
Lusigal 12
1069 (projected)
1068
900
1067
1065
Two-way travel time [s]
HHD
HHD
HD
HD
C
Moho
Distance [km]
(b)
Hobby High
W
E
(shot points)
3800
4200
4600
5000
HHD
HHD
HD
HD
C
C
LD
[km]
5 km
901
(c)
Depth [km]
21.4 km
(d)
13.3 km
(e)
syn-rift
pre-rift and upper crust
lower crust
mantle peridotite
(variably serpentinized)
top syn-rift
top acoustic basement
intracrustal reflection C
Moho
faults
HHD: Hobby High detachment
HD: H-detachment
LD: L-detachment

Krawczyk et al. (1996), and Whitmarsh et al. (2000) (Fig. 12a). The L and the H reflections form break-aways to the east and flatten at depth. The H reflection is upwarped beneath Hobby High and reaches the top of the basement at Hobby High where it is overlain by post-rift sediments (Fig. 11b). A continentward-dipping reflection, the C reflection, is cut and displaced along H and L (Whitmarsh et al. 2000) (Fig. 12b). Thus, the H and L reflections show clear characteristics of fault structures and are called henceforth H-Detachment (HD) and L-Detachment (LD) (Fig. 12).

The C reflection is cut and displaced by HD and LD and must consequently be older than these structures. Only few kilometres north of Lusigal 12, on seismic line CAM 144, Chian et al. (1999) described a 6.5 to 7.3 km/s interface at a depth similar to that of reflection C on Lusigal 12 and interpreted it as a crust/mantle boundary. Whitmarsh et al. (2000) interpreted, based on the velocity data, the C reflection as a contact between continental crust above and serpentinized peridotite below. A simple geometric interpretation, however, leads to a different interpretation of the C reflection. The C reflection is truncated by the HD and has therefore to change across HD from the footwall into the hanging wall of the HD going oceanwards. Thus, the C reflection is expected to cut out at the top of the basement continentwards of where the HD reaches the top of the basement (Fig. 12b). ODP drilling at Sites 900 and 1067 over Hobby High demonstrated that the rocks directly overlying the HD are formed of meta-gabbros, amphibolites, and meta-tonalite lenses which gave Paleozoic U-Pb zircon ages, indicating that these rocks represent pre-rift continental lower crust (Manatschal et al. 2001). Thus, based on the geological argumentation in this paper, the C reflection is interpreted as an intracrustal reflection separating lower crustal from upper crustal rocks.

Deformation of the lower-crustal rocks drilled at ODP Sites 900 and 1067 is pervasive and brittle at the top and decreases in intensity down-hole. $^{40}Ar/^{39}Ar$ 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) respectively. 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. These observations show that Hobby High has to be capped by a detachment fault, which exhumed the lower crustal rocks to the sea floor (Fig. 11b). This interpretation explains the cooling ages and the penetrative deformation at the top of the drilled basement at Hobby High. At Site 1068, on the oceanward side of Hobby High, drilling penetrated mantle rocks which are capped by a serpentinite gouge and tectono-sedimentary breccias, providing further support for the idea that Hobby High is capped by an exhumed detachment, the Hobby High Detachment (HHD) (Figs. 11b and 12b). This detachment is interpreted to cut from the continental crust at Sites 1067 and 900 into mantle at Site 1068, a situation which compares well with the situation described from the OCT within the Tasna nappe (Fig. 11a). Hölker et al. (2002b) discussed the similarity of the rocks drilled at Hobby High and exposed in the Tasna OCT and demonstrated, using seismic modelling, that the observed detachment structures in the Tasna OCT show the same seismic characteristics as the HD and the HHD at Hobby High.

In seismic section Lusigal 12, HHD is interpreted to break out to the east, near shot point 4300 between Sites 1065 and 1067 (Fig. 12b). From there, it can be followed towards Hobby High, where it has been drilled. Further oceanwards, following the F reflection of Krawczyk et al. (1996), it plunges beneath a high drilled at Site 1069. This high, about 1.5 km thick 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. 12b). 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. Hölker et al. (2003) proposed that the HHD is overlain by smaller allochthons similar to those observed along the Err detachment.

Kinematic inversion of the Lusigal 12 seismic profile

The previous description shows that HHD is in many respects different from the LD and the HD. A simple kinematic inversion of the interpreted seismic section Lusigal 12, presented in Fig. 12b to e, allows a description of the temporal evolution of detachment faulting in the Iberia Abyssal Plain. The kinematic inversion, which was kept as simple as possible, considers only rotational and translational deformation of seismically well-defined blocks, and conserves the cross-sectional area within the profile, assuming that the blocks moved parallel to the profile. For details of the kinematic inversion see Manatschal et al. (2001).

The thickness of the pre-detachment crust and the former crustal location of the ODP profiles (circles labelled with ODP site numbers), as well as the amount of thinning are obtained from the step-by-step kinematic inversion of the profile. The reconstruction shows that the total extension accommodated by LD, HD, and HHD is on the order of 34.7 km (the initial, pre-detachment length is 34.6 km, final post-detachment length is 69.3 km). The two most important results which were obtained from the reconstruction are that the crust was already thinned to about 10 km before LD, HD, and HHD became active (Fig. 12e), and that the geometry of the active faults changed from concave upwards (Fig. 12d) to concave downwards (Fig. 12c) during extension.

Initial faults were upwards concave, accommodating offsets of respectively 4.7 km (LD) and 8.6 km (HD) and soled out at about 10 km within a layer with a velocity higher than 7.3 km/s, interpreted as the crust/mantle boundary. This observation suggests the occurrence of a weak material which was either serpentinized mantle or a quartz-rich protolith. Marine Tithonian pre-rift sediments capping the footwall blocks near the break-aways at Sites 901 and 1065, as well as no evidence for truncation of the crests by erosion suggest that these blocks remained below sea level. Thus, during faulting along LD and HD, the footwall blocks were not significantly uplifted while the hanging wall subsided leading to rift basins up to 4 km deep and 15 km wide for the LD.

HHD differs from LD and HD because it is relatively flat (Fig. 12b). A displacement of several tens of kilometres is required to explain that for over more than 20 km this fault caps the top of the basement, only overlain locally by small allochthons (Hölker et al. 2003). Although HHD shows a complex architecture, which may result from post-kinematic isostatic equilibration, at present, nowhere does it dip more than 17°. This low inclination and the fact that the angle between the syntectonic breccias recovered at Site 1068 and the detachment fault is less than 10° indicates that the fault was at a low angle when it was exposed at the sea floor. However, if HHD was subhorizontal, how could this fault exhume mantle rocks and lower crustal rocks at the sea floor? Major tilting after its activity can be excluded because the overlying strata are subhorizontal. On the other hand, assuming a displacement of some tens of kilometres along a fault dipping 17° and having rigid hanging wall and footwall blocks would result in fault-related vertical relief of several kilometres (for 20 km displacement the topography would be in the order of 5.8 km). There is, however, neither evidence for subaerial exposure of the footwall, nor for subsidence of the hanging wall of the HHD. The lower Valanginian rocks recovered at Site 1069 in the hanging wall of the HHD (Fig. 12b) are chalks composed of a faunal assemblage of ostracods and benthic foraminifera thought to be indicative of an open-marine, outer shelf to upper slope environment not deeper than 1500 m (Urquhart 2001). Thus, the only geometrical solution to explain all observations is that the footwall was pulled out from underneath a relatively stable hanging wall along a concave downward fault (Fig. 12c). This is similar to the rolling-hinge model (Buck 1988; Wernicke and Axen 1988; Tucholke et al. 1998). Faults developing a similar geometry have also been modelled recently by Lavier et al. (1999). In their model, such faults form when the brittle layer is thin and the underlying material is weak.

Reflectors vs. detachment structures: a geological interpretation of rift-related structures

The previous descriptions show striking similarities between the observed detachments in the Alps and the seismically imaged and drilled reflections in Iberia. Hölker et al. (2002a) modelled synthetic seismic sections using density, velocity and thickness data from the well-exposed detachment structures in the Tasna OCT (UTD and LTD) (Hölker et al. 2002b) and the Err detachment (Hölker et al. 2003) and compared the synthetic seismic profiles he obtained with the seismic data and drilling results from Iberia. These studies provide some interesting constraints on seismic characteristics of detachment/reflections observed in the OCT. The study shows that the reflectivity of exhumed subcontinental mantle is characterized by variable amplitudes and numerous diffractions and contrasts with that of exhumed continental rocks which show a continuous and strong reflection. Moreover, shallow intrabasement detachments at the crust-mantle boundary are imaged as weak and discontinuous reflections with inverse polarity. However, their reflectivity depends strongly on the fracture healing behaviour of the surrounding rocks. If healing is fast and efficient, which has been demonstrated for the Err detachment (Manatschal 1999), such structures are transparent and will not be imaged (Hölker et al. 2002a).

Upward-concave faults (LD, HD, and LTD)

Structures comparable to the deeper and more proximal parts of HD and LD are unlikely to be preserved in the Alpine examples because the individual thrust sheets derived from the distal margin do not exceed 500 m in thickness and sampled only the uppermost crust including its cover. Further oceanwards, at Hobby High, the HD is in a shallow crustal position and truncated by the HHD. A comparable situation is exposed in the Tasna OCT where the LTD is cut by the UTD (Froitzheim and Rubatto 1998) (Fig. 11a). The LTD forms a tectonic contact between the lower crust and the upper mantle, exactly as proposed for the HD. Thus, the LTD may represent a structure analogous to the HD beneath Hobby High (compare Fig. 11a and Fig. 11b).

Downward-concave faults (HHD, Err detachment, UTD)

Structures corresponding to the HHD are exposed in the Alps: portions that were closer to the continent (not drilled in Iberia) are preserved in the Err nappe (e.g. Err detachment), the transition from the continent to the exhumed mantle (e.g. Hobby High) is exposed in the Tasna nappe, and situations corresponding to more oceanward portions (not drilled in Iberia) are found in the Platta nappe. These fault structures are characterized by: (1) 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 plane 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. These faults, henceforth called "top-basement detachment faults" (Hölker et al. 2003), are therefore interpreted as exhumed segments of no longer active downward-concave detachment faults which accommodated tens of kilometres of offset and exhumed continental and mantle rocks at the sea floor. The discovery of top-basement detachment faults dramatically

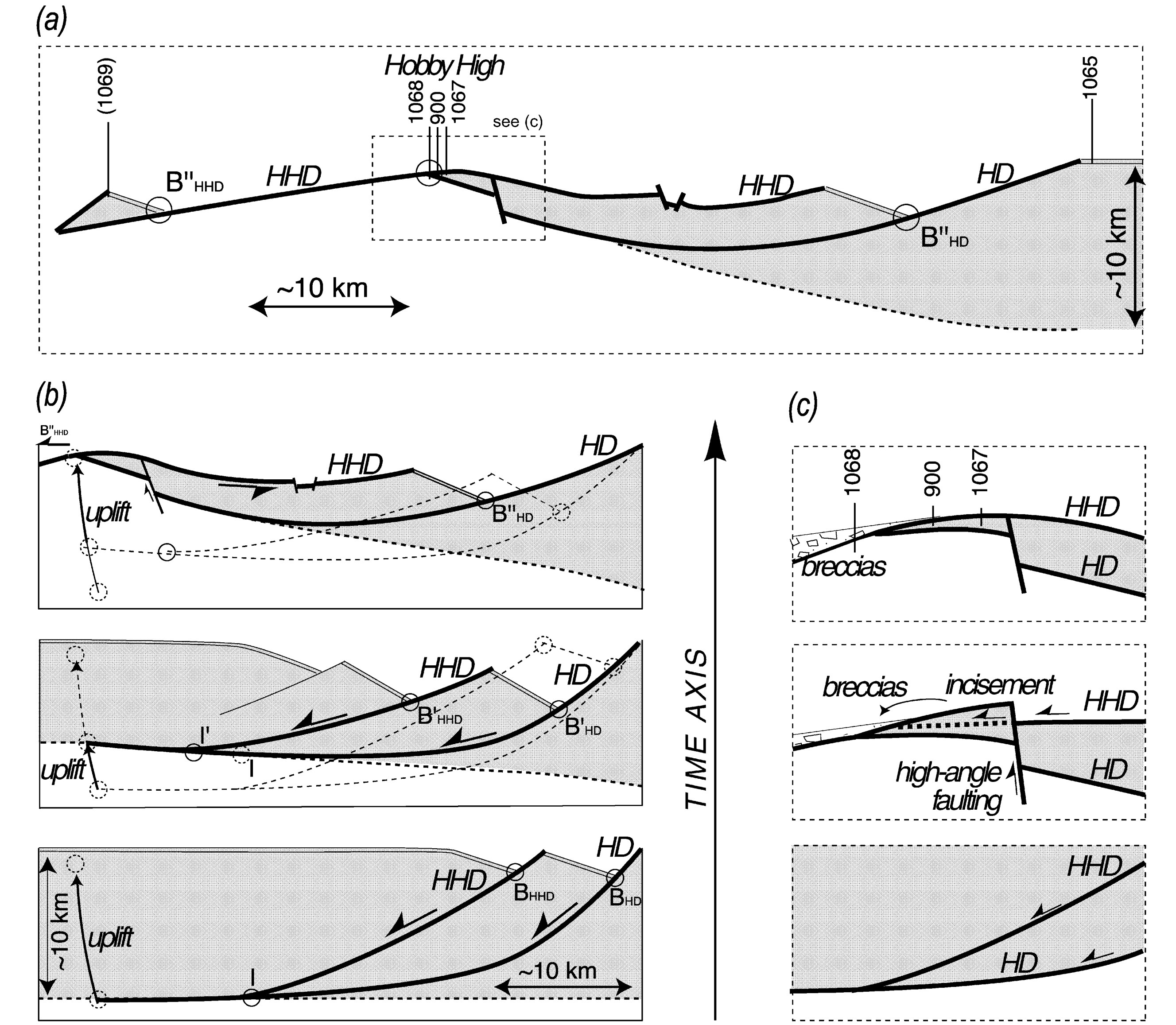


Fig. 13 (**a**) Present-day architecture of the OCT in the Iberia Abyssal Plain. (**b**) Evolution of the geometry of detachment faulting from upward-concave faults (*HD* and *HHD*) to downward-concave (*HHD*) during active uplift beneath Hobby High. (**c**) Relationship between low-angle and high-angle faulting as observed at Hobby High

changed the way the architecture and tectonic evolution of the OCT zone in magma-poor rifted margins is interpreted.

Late high-angle faults in the OCT

Several high-angle faults have been observed in the southern Iberia Abyssal Plain (Whitmarsh et al. 2000; Hölker et al. 2003). These faults are sealed by the sedimentary sequences overlying directly the HHD and consequently predate their deposition. Therefore, these high-angle faults are interpreted to form either during detachment faulting or directly afterwards. A direct relationship between detachment faulting and one well-defined high-angle fault is observed east of Hobby High (Fig. 13a). Whitmarsh et al. (2000) mapped and described this high-angle fault in several reflection lines and noted that it displaces the HD but not the top of the actual basement, interpreted as a top-basement detachment fault. Thus, the high-angle fault had to move after HD but before or simultaneously with HHD. The latter solution, shown in Fig. 13b (for details see also Fig. 13c) would imply that the up-warping of the HD, leading to the formation of Hobby High occurred while HHD was active. This solution forces one to assume that HHD was incising into the uplifting footwall of the active high-angle fault. Incisement into the uplifting hanging wall may also explain the formation of the breccias drilled on the oceanward side of Hobby High at ODP Site 1068 (Fig. 13c).

Hölker et al. (2003) described further high-angle faults truncating the HHD continentwards of Hobby High within the Lusigal 12 profile (Fig. 13a). Based on the observation that these high-angle faults affected the HHD but are sealed by the overlying sediments, it is very likely that most of the sediments observed in the Lusigal 12 profile between Hobby High and shot point 4300 (the break-away of HHD; Fig. 12b) have to be post-rift in age. Whether these high-angle faults accommodated some of the strain necessary to unbend the downward concave HHD or if the high-angle faults are unrelated to detachment faulting is not yet understood.

Early continentward dipping detachment faults (C, Margna and Pogallo faults)

Based on a geometrical interpretation and drilling results, the continentward dipping reflection C off Iberia is interpreted in this study as an intracrustal reflection separating upper crust from lower crust (for an alternative interpretation of the C reflection see Whitmarsh et al. 2000). Because C is truncated by the LD and the HD, it has to be older than these faults and it is interpreted as a detachment, active during an early stage of rifting leading to thinning of the crust to about 10 km. A similar interpretation can also be envisaged for the Pogallo fault (Handy and Zingg 1991) (Fig. 7) and for the Margna fault (Müntener et al. 2000) (Fig. 3c). In contrast to the C reflection, the Margna and Pogallo faults do not preserve direct relationships to oceanward dipping detachment faults, however, p-T-t data obtained from these faults demonstrate that they were active during an early stage of rifting and led to a significant thinning of the continental crust (Müntener and Hermann 2001). This is compatible with the results obtained from kinematic inversion of Lusigal 12 profile across the southern Iberia Abyssal Plain, which demonstrate that the crust had to be thinned to about 10 km before the HD and LD became active (Fig. 12). Thus, the Pogallo and Margna faults may represent structures equivalent to the C reflection.

In the deep Galicia margin, a structure equivalent to the C reflection is not observed. The most prominent reflection there is the S-reflection. Previous interpretations proposed an oceanward as well as a continentward dip of this structure (Reston et al. 1995). An alternative possibility is that the S reflection corresponds to a structure equivalent to C but which was reactivated or truncated by later structures comparable in style with the LD and the HD. Polyphase faulting and local reactivation of pre-existing structures may explain the different architecture observed in different transects.

Fault systems acting in rifted margins

The previous interpretation distinguishes between different fault structures which can be grouped into three different systems responsible for the successive thinning of the lithosphere from onset of rifting to final continental break-up (Fig. 6).

Fault system I: This fault system can be subdivided into two subsystems, fault system Ia and Ib. Fault system Ia is formed by faults cutting across the brittle upper crust and soling out at middle to lower crustal levels. These faults bound rift basins, up to 4 km deep and 30 km wide; fault offsets are <10 km and total extension is limited (β <2), and both hanging wall and footwall are subsiding which makes it impossible to exhume deeper crustal rocks along such faults. The type-example of fault system Ia is the Lugano-Val Grande fault (Bertotti 1991) (Figs. 7 and 8).

Fault system Ib is characterized by faults accommodating large amounts of extension in deeper crustal levels. They form simultaneously with fault system Ia and are therefore considered to be kinematically linked with it. Fault system Ib is supposed to be spatially related to the place where the necking of the lithospheric mantle occurs (Figs. 6b and 8). The architecture of this fault system is not yet well constrained, however, p-T-t data obtained from these faults as well as seismic interpretations from the Iberia margin suggest that these faults lead to a local thinning of the crust to less than 10 km. Good examples of fault system Ib are the Pogallo fault (Fig. 7) and the Margna fault (Fig. 3c) in the Alps and the C "reflection" in the southern Iberia Abyssal Plain (Fig. 12b).

Fault system II: This fault system shows many similarities with fault system Ia. The faults cut across a brittle layer and sole out within an underlying ductile layer (Fig. 12b). The fault-bounded basins are up to 4 km deep and 20 km wide and both hanging wall and footwall are subsiding. The overall extension and thinning by these faults are limited. The difference from fault system Ia is that these faults become active during a later stage of rifting and cut across the whole crust and sole out at the crust-mantle boundary (Fig. 12d). These faults can, therefore, only be active during a late stage of rifting, when the crust has already been extended and cooled. Thus, fault system II has to be preceded by initial thinning. Examples for fault system II are the LD and the HD in the Iberia Abyssal Plain (Fig. 12b).

Fault system III: This fault system is characterized by downward concave faults which are responsible for the exhumation of the mantle at the sea floor (Fig. 12c). These faults can accommodate large offsets (>10 km) without producing major fault related topography; they exhume deeper crustal and mantle rocks from underneath a relatively stable hanging wall. Examples of this fault system are the HHD in the Iberia Abyssal Plain (Fig. 12b) as well as the well-exposed Err detachment (Fig. 9) and the UTD (Fig. 11a) in the Alps.

These three fault systems formed and accommodated strain at different stages and positions within the margin. They can explain the thinning, localization, and final rupturing of the lithosphere as well as the local variability of the rift architecture. Thus, most of the extension appears to have been accommodated along shear zones. Homogeneous crustal flow may have occurred in quartz-rich lower-crustal levels; however, ductile flow cannot have been very important, accepting at a temperature of 550 to

600°C at the crust/mantle boundary at the onset of rifting (Müntener et al. 2000).

Fault systems Ia and Ib formed during an initial stage of rifting. These faults were probably kinematically coupled. The positions of deformation along these faults were guided by inherited structures and heterogeneities within a "cold" pre-rift lithosphere. The mode of extension along the two fault systems was, however, different in the distal and proximal margins. While extension in the distal margin affected mainly the lower crust, extension in the proximal margins affected the upper crust (Fig. 8). Late extension was localized in the previously strongly thinned distal margin and was first accommodated by the listric normal faults HD and LD, and, a later stage, also HHD (Fig. 13 b). These faults soled out at the crust-mantle boundary. Final rifting leading to mantle exhumation was controlled by downward-concave faults, the HHD in Iberia (Fig. 12b), and the Err detachment (Fig. 8) and the UTD (Fig. 11a) in the Alps. The similarity between single structures observed in the Alps and off Iberia, as well as their spatial and temporal evolution within the margin, suggest that the processes controlling their evolution are the same. Although these processes are not yet well understood in detail, it seems very likely that they reflect rheological changes resulting from the complex interaction of mechanical, chemical, and thermal processes during rifting. Thus, it is not the individual fault systems but rather their evolution in time and space that is characteristic of magma-poor margins. This may reflect rheological changes in the lithosphere due to serpentinization and heat advected by melts derived from the rising asthenosphere.

The transition from fault system type I to type II may reflect the extent of embrittlement (Whitmarsh et al. 2000) and the onset of serpentinization which starts when the crust reaches a thickness of about 10 km and the mantle arrives in the stability field of serpentine (Pérez-Gussinyé and Reston 2001). Serpentinization depends on fluid accessibility as well as on the stability field of serpentine. Therefore, serpentinization is expected to be limited during an initial stage to the uppermost part of the mantle, which is compatible with the observed localization of faulting along the crust-mantle boundary (e.g. HD; Fig. 12b).

The transition from fault system type II to type III may coincide with the onset of magmatic activity as indicated by the cross-cutting relationships between magmatic rocks and detachment faults in the Platta nappe (Fig. 10 g,h). Magmatic weakening is coupled with a rising asthenosphere and is, therefore, associated with a dynamic change in the thermal structure which is not well understood (Minshull et al. 2001). In contrast to the serpentinization-induced weakening, the thermo-magmatic weakening tends to have a rather vertical structure and is spatially and temporally related to the rising asthenosphere. Uplift and tilting in the footwall of HHD, observed underneath Hobby High, occurred while the hanging wall, drilled at ODP Site 1069, remained at relatively shallow marine conditions (less than 1500 m) (Urquhart 2001) compared to the 3 to 4 km water depth in the half graben basins to the east (Figs. 12c and 13b). These observations suggest a contribution of active rifting underneath the extended area. Melts derived from the rising asthenosphere may also have had a triggering effect on the changing mode of deformation from overall pure-shear deformation accommodated by the fault systems I and II to a transient phase of simple-shear deformation on a marginal scale, during which strain has been accommodated by fault system III (Fig. 6b,c).

Although the underlying mechanisms controlling the transition from one strain accommodation system to the other are not yet fully understood, they are likely to reflect changes in the bulk rheology associated with changes in the thermal structure of the extending lithosphere associated with a rising and narrowing "ridge" of asthenosphere. These changes include a general transition from mechanically (fault system I), to hydration/serpentinization (fault system II), to magmatically/thermally (fault system III) dominated localization processes. In detail, however, the interaction between the different processes is not yet well understood.

Concave-downward faults at the transition from rifting to sea-floor spreading

One of the major conclusions derived from the interpretation of detachment structures in the OCT is that detachment faults exhuming the mantle are concave downward. These faults coincide with the top of the acoustic basement over large distances. They are rooted in the mantle, and are overlain locally by continent-derived extensional allochthons and/or tectono-sedimentary breccias. Despite offsets of several tens of kilometres accommodated along these structures, their fault-related morphology is relatively smooth. The concept of concave downward faults is an integral part of the rolling-hinge model previously proposed by Buck (1988) and Wernicke and Axen (1988). This model proposed that the observed large offsets along low-angle detachment faults in the metamorphic core complexes in the southwestern United States were accommodated by faults that were active at high angles and rotated near to the surface to lower angles. A similar model has been proposed by Tucholke et al. (1998) to explain mantle exhumation at slow-spreading ridges, leading to so-called "mega mullions". Numerical modelling by Lavier et al. (1999) showed that such concave-downward faults accommodating large offsets form when the brittle layer is thin and underlain by a weak material. The concave-downward fault geometry proposed here for the HHD in the southern Iberia Abyssal Plain and for the Err detachment and the UTD is related to a different tectonic setting, namely the transition from rifting to sea-floor spreading within the OCT. Exhumation of mantle rocks in the OCT contrasts to continental extension by the amount of crustal thinning. In contrast to core complex formation along slow or ultraslow spread-

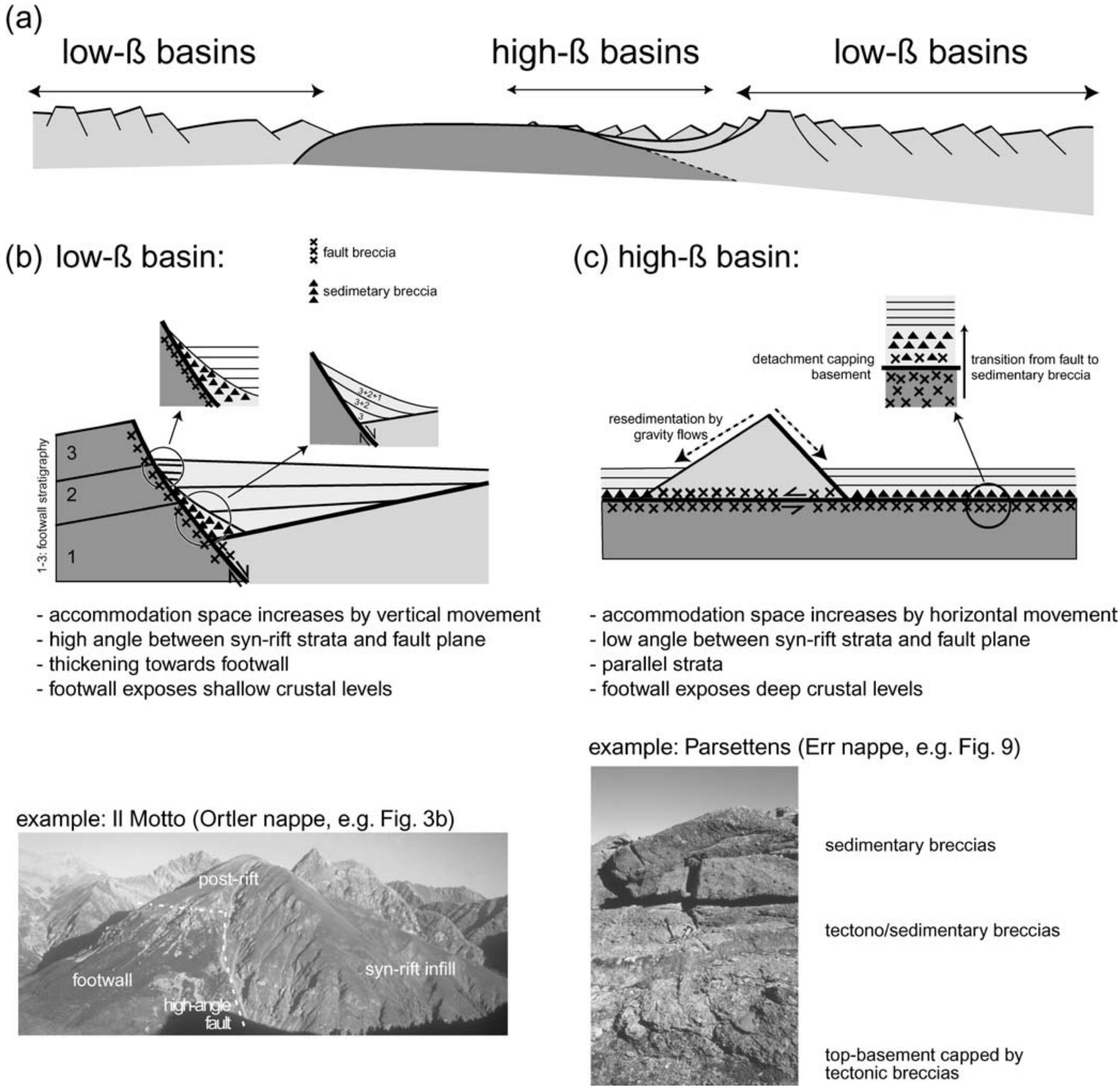


Fig. 14 (**a**) Distribution of low- and high-β basins in conjugate magma-poor rifted margins. (**b**) Sediment architecture and example of a low-β basin. (**c**) Sediment architecture and example of a high-β basin (modified from Wilson et al. 2001a)

ing ridges, extension in the OCT cannot be considered as a steady-state process.

The interpretation of concave-downward faults has major implications for the architecture of OCT zones, the early evolution of ocean basins, and the determination of strain (e.g. β-factors) and strain rates which are commonly used to predict subsidence and magma production along rifted margins (e.g. McKenzie 1978; White et al. 1987; McKenzie and Bickle 1988). Thus, the new observations require a re-examination of previous ideas and interpretations about the late tectonic evolution of magma-poor rifted margins. In the following part of this paper, I will explore the effects which concave-downward faults may have for the sedimentary architecture (Fig. 14) and for the distribution of magmatic rocks, and characteristics of mantle rocks in the OCT (Fig. 15).

Implications for the sediment-architecture in the OCT

Offset along individual faults within the southern Iberia Abyssal Plain increases oceanwards from some kilometres along LD and HD to some tens of kilometres along HHD. The increasing strain is accompanied by a change

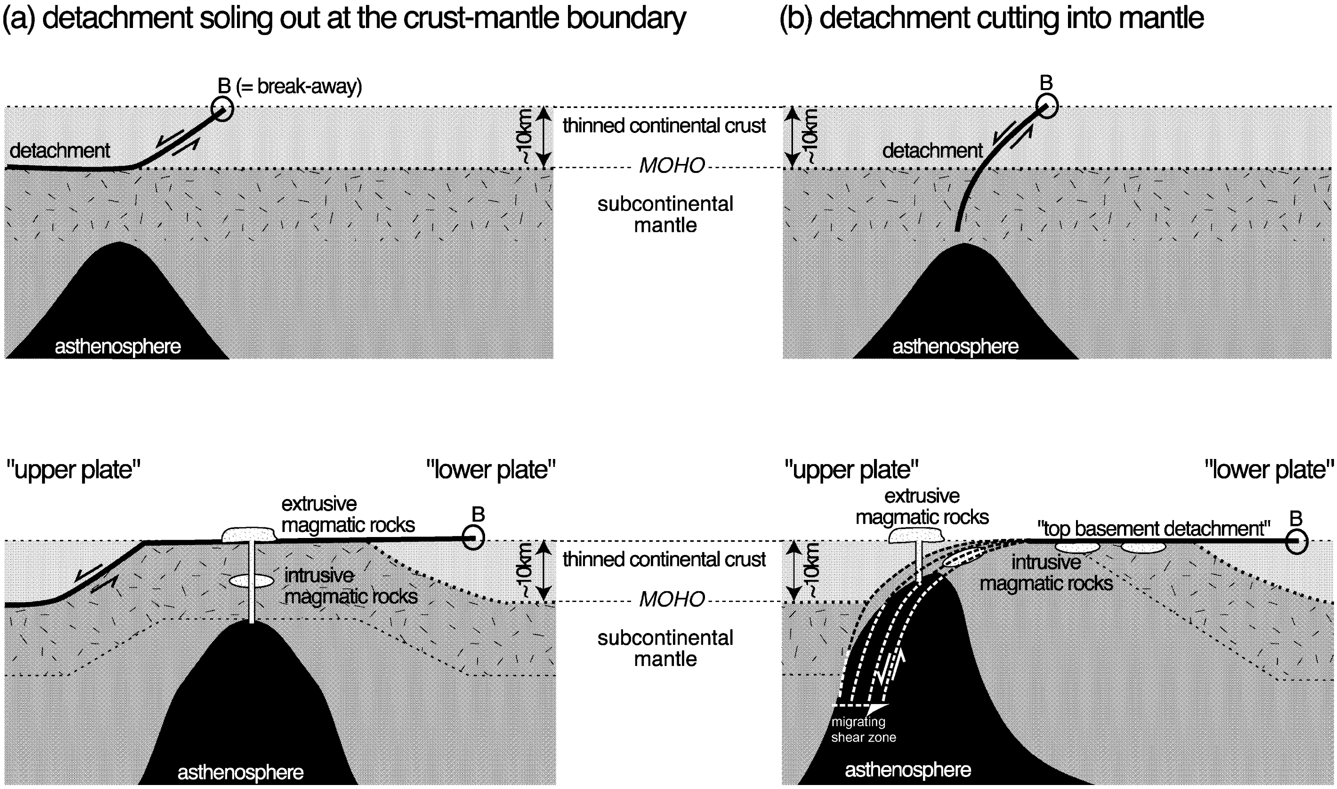


Fig. 15a–b Detachment geometry and its importance for the distribution of magmatic and mantle rocks in the ZECM: the example of: (**a**) a detachment fault soling out at the crust-mantle boundary; (**b**) detachment cutting into the mantle lithosphere. For discussion see text

in the geometry of the associated faults from concave-upward (e.g. LD) to concave-downward (e.g. HHD). The across-strike changes in geometry of the detachment faults are also accompanied by a change in the basin architecture leading to two end-member types of basins, henceforth called low-β and high-β basins (Wilson et al. 2001a) (Fig. 14a). Low-β basins are well studied and described in the literature (e.g. Bosence 1998). They are commonly filled by continental, shallow-marine, or submarine fan deposits associated with listric faults and hanging-wall subsidence which includes a rotation of the hanging-wall block leading to asymmetric rift basins. Within these basins, accommodation space increases by vertical movement, the footwall exposes only shallow-crustal levels, sediments in the basin onlap the faults at a high-angle, and the overall architecture of the basin is determined by sedimentary sequences thickening toward the footwall and onlapping onto the hanging wall (Fig. 14b).

High-β basins related to concave-downward faults are expected to show stratal relationships which are completely different from those of classical low-β rift basins (Wilson et al. 2001a). These basins form by pulling the footwall from underneath a stable hanging wall along concave-downward faults accommodating large fault offsets (Fig. 14c). Depositional area increases dominantly by horizontal movements, the basin is floored by detachment faults which exposes deep crustal and even mantle levels, and the syn- and post-rift strata are parallel or subparallel to the top-basement detachment surface. Within these basins the basal detachment might have acted as some kind of conveyor belt whereby the exhumed footwall rocks were fractured, exposed, and redeposited along the same active fault system. This interpretation may explain the common occurrence of tectono-sedimentary breccias overlying detachment faults observed in the Alps and drilled off Iberia and has some major implications for the interpretation of the stratigraphic record within the distal parts of these margins.

Implications for mantle characteristics and distribution of magmatic rocks in the ZECM

Concave-downward faults can explain the exhumation of subcontinental mantle rocks over a wide area in the ZECM without producing a high-amplitude, fault-bounded topography in a relatively simple way. How solid-state deformation and serpentinization processes are linked to the rise of the asthenosphere and how they interact with magma is not well understood in detail. A further complication, not discussed here, addresses the temporal and spatial evolution of the thermal structure and in particular how it is coupled in detail with detachment faulting. Here only the two most simple cases are considered and discussed: a detachment fault with a concave-upward geometry soling out at the crust-mantle boundary within a serpentinized upper mantle (Fig. 15a); and (2) a detachment fault rooting into a rising asthenosphere (Fig. 15b). These two end-member models demonstrate the expected relationships between mantle exhumation, deformation processes, and magmatism in the ZECM in a schematic way.

In both examples, the top of the mantle rocks exhumed at the sea floor represent an exhumed fault zone; also called a "top-basement detachment fault". This is compatible with the observation that wherever the top of the mantle has been sampled, in the Alps or off Iberia, the mantle is always capped by a brittle fault zone and associated with tectono-sedimentary breccias. Thus, most of the available mantle rocks sampled in the ZECM were formed in a fault zone. Assuming that fault zones represent zones of higher permeability, forming pathways for magma and fluids which result in strongly perturbed isotherms and complicated fluid-magma-rock interactions; the results obtained from these mantle rocks are not necessarily representative of mantle rocks forming deeper levels in the ZECM.

As shown in the two models in Fig. 15, the type and origin of the exhumed mantle rocks in the ZECM depend on the geometry of the detachment fault. Deeper mantle levels can only be exhumed with faults rooting in the asthenosphere (Fig. 15b), whereas faults soling out at the top of the mantle exhume only the uppermost subcontinental mantle (Fig. 15a). Desmurs et al. (2001) and Müntener and Piccardo (2003) demonstrated that the characteristics of the mantle rocks change in the ZECM. Next to the continent, the mantle rocks are formed by spinel peridotite mixed with (garnet)-pyroxenite layers, which equilibrated at lower temperatures. Oceanwards the mantle rocks are formed by pyroxenite-poor peridotite that equilibrated in the plagioclase stability field (Fig. 4). This change in the mantle characteristics across the ZECM observed in the Platta nappe supports the idea of a detachment rooting into the asthenospheric mantle (Fig. 15b). The existence of such a detachment fault necessarily results in a footwall rotation during exhumation. This may explain the up-warping of the HD beneath Hobby High in the footwall of the HHD (Fig. 13) and in a more general way, it shows that the observed orientations of structures and intrusive magmatic bodies in the ZECM do not necessarily correspond to their original orientation.

The geometry of detachment faults also controls the emplacement and distribution of magmatic rocks in the ZECM. In the case of a detachment soling out at the crust-mantle boundary, the magmatic system is limited to the footwall and can be considered to be symmetric (Fig. 15a). In the case of a detachment cutting into the mantle lithosphere, magmatic rocks can be emplaced in both hanging wall and footwall and can be separated later on by the detachment fault leading to a strongly asymmetric system (Fig. 15b). Intrusive and extrusive magmatic rocks emplaced in the hanging wall will finally form the "upper plate" margin, while intrusive magmatic rocks of the same age emplaced in the footwall will be exhumed on the "lower plate" margin (upper and lower plate refer only to the most distal parts of the margin and not to the whole margin). Thus, magmatic rocks forming simultaneously one above the other in a vertical rock column will finally be exposed side by side within the ZECM and will, after the onset of sea-floor spreading, form the conjugate borders of the oceanic basin (Fig. 15b). The model predictions are compatible with the observed trend in the ZECM in the Platta nappe, which is interpreted as a "lower plate" margin. There, basalts generally become more voluminous oceanwards and grade from T- to N-MORB in this direction (Desmurs et al. 2002). The model can also explain the apparent difference between the magma-rich ophiolites in the French/Italian Alps, interpreted as "upper plate" margin, and the magma-starved ophiolites in Grisons, interpreted as "lower plate" margin. By analogy to the Alpine ophiolites, the OCT off Newfoundland, interpreted as an "upper plate" should be more magmatic and similar to the ophiolites in the Franco/Italian Alps. If this hypothesis can be confirmed, then the term "non-volcanic margin" would become misleading and the concept of entirely non-magmatic mantle exhumation wrong. Although the total amount of magma appears to be small within magma-poor margins, observations in the Alps show that mantle exhumation by detachment faulting can be accompanied by magmatic activity. Therefore, I suggest that the "lack" of magmatism, as suggested from drilling on the Iberia margin is due to a drilling artefact and most of the magmatic rocks related to mantle exhumation in Iberia are expected to be emplaced in the conjugate Newfoundland margin.

Pending questions and unresolved problems related to the tectonic evolution of rifted margins

What is the stratigraphic response to lithospheric extension?

At present, the syn- and pre-rift stratigraphic record of the deeper and more distal parts of rifted margins is poorly known except for a few deep drill holes. Thus, the most limiting factor concerning the stratigraphic record of continental break-up is the lack of samples, i.e. of "real" data. This is the case for most distal parts of conjugate pairs of margins, including the Iberia/Newfoundland margins. How much of our knowledge about the pre-rift and syn-rift evolution of the Iberia/Newfoundland margins, and consequently of their isostatic evolution, is supported by "real" data? The sparse data obtained so far by scientific deep-sea drilling from the OCT show how little we know about this part of the margin and question previously proposed models for isostasy. In conclusion, more and deeper drill holes in the oceans are needed to understand the stratigraphic record of break-up. A lot of information is, however, already available and exposed in collisional orogens. The results presented in this paper show remarkable similarities between the two margins which permit us to make some predictions for what can be expected from the Iberia/Newfoundland conjugate margins. The stratigraphic record from the Alps shows that rifting initiated over a wide area and shifted during successive phases towards the future OCT. The complex temporal and spatial evolution of rifting results in syn- and post-rift sedimentary sequences becoming younger oceanwards, a result that contradicts the concept of the "break-up unconformity". A further observation, which is backed up by data from the Alps and seismic sections from the Atlantic, is that the rift architecture changes across the margin. Proximal basins are larger and bounded by listric faults, whereas in the more distal "high-β" basins the basin-fill is thinner and bounded by low-angle detachment faults (Fig. 14). Thus, the stratigraphic response to extension is different in proximal and distal margins. These differences have to be taken into account when margins are compared. Distal parts of the margins appear to be asymmetric, while proximal margins are, on the scale of the margin, symmetric.

β values, strain rates, and duration of rifting

In many rift models, thinning or extensional factors (β-values) and strain rates are used to determine subsidence or amount of expected magma in rifted margins (e.g. McKenzie 1978). Although pure-shear models can explain many observations in proximal margins, they are unable to explain the tectonic evolution and the basin architecture in distal margins. This is because distal margins are the result of a polyphase tectonic evolution, which is associated with a complex time-dependent thermal structure of the lithosphere (Fig. 6).

In order to determine strain rates, the duration of rifting has to be known. For the Alps, the age of break-up is well constrained, in contrast to the age of onset of rifting. Observations show that the onset of rifting is diffuse and widespread, associated with subsidence, and initiated 60 to 40 Ma before the onset of sea-floor spreading. Depending on the subsidence/accommodation rate, open rift basins may or may not have formed. Therefore a change in sedimentary facies showing a deepening of the basin does not necessarily indicate the onset of rifting. Although overall rifting (i.e. tectonic thinning of the continental lithosphere) may have lasted a long time, local areas in the margin may have been affected by rifting for only a short time. Strain rates measured for single faults bounding larger rift basins typically show that most of the strain has been accommodated during short pulses and that activity is not simultaneous along different faults across the margin. This means that strain rates for the whole margin are difficult to obtain. Available data suggest that the overall extension associated with rifting is slow (~10 mm/yr), discontinuous, and lasts tens of millions of years (Bertotti et al. 1993; Chevalier 2002) while movements along single faults and in particular final break-up are relatively fast, i.e. within less than ten millions of years. Tregoning et al. (1998) reported up to 30 mm/yr for the detachment fault in the Woodlark basin, which is related to continental break-up.

How are deep lithospheric/asthenospheric processes coupled with shallow crustal processes?

Changes in the architecture of rift basins across the margin, isostatic movements during rifting, and the along-strike segmentation of the margin are suggested to reflect indirectly deep lithospheric/asthenospheric processes and related rheological changes during rifting. Thus, on a margin scale, deep lithospheric/asthenospheric processes are coupled with surface processes, across rheology and density changes. The major problem in linking deep and shallow lithospheric processes is the lack of time constraints and the different temporal and spatial resolution, which can be obtained from the two domains.

If, as proposed in this paper, continental break-up is achieved by concave-downward detachment faults cutting into the mantle and progressively exhuming deeper mantle levels across the ZECM (Figs. 6d and 15b), the rocks finally exposed in the ZECM represent a section across the pre- and syn-rift lithospheric mantle. Thus, profiles across the present-day ZECM permit mantle processes to be studied occurring at different levels during rifting and to couple them to surface processes recorded in the sediments in the margins. Desmurs (2001) and Müntener and Piccardo (2003) showed that mantle rocks systematically change their characteristics across the ZECM (Fig. 4). One of the future challenges will be to unravel the complex, polyphase interaction between magmatic, hydrothermal, and deformation processes recorded in the mantle rocks and to combine these data with the temporal and spatial evolution inferred from the stratigraphic record. Combining these data sets will result in a better understanding of the coupling between deep-lithospheric and surface processes active during the formation of rifted margins.

Implications for other margins and possible predictions

Although the Iberia and Adriatic margins are of different ages and ultimately had a different fate, the remarkable similarities between them should not lead to the conclusion that all magma-poor rifted margins are similar. Comparative studies tend to highlight similarities rather than differences and it would be dangerous to suggest that every magma-poor margin evolves in a similar way. Therefore, one of the most important questions is: how far can observations derived from the Alps and Iberia be generalized to explain the evolution of rifted margins? An effective way to answer to this question is to test predictions derived from the study of these two margins. One major prediction is that entirely non-magmatic rifting does not exist, and that most of the magma is emplaced during final rifting in the upper plate. In the summer of

2003, ODP Leg 210 drilled the Newfoundland margin partly in order to test this idea. Let us see what the samples will tell us.

Conclusions

This paper reviews concepts for the formation of magma-poor rifted margins using data from the present-day Iberia margin and the ancient margins exposed in the Alps. Accepting the limits of such comparative studies, the following conclusion can be made:

1. Rifting is polyphase and extension is accommodated by three successive fault systems reflecting a changing thermal and rheological structure of the lithosphere driven by ascending asthenosphere during final rifting
2. Location of the position of continental break-up is pre-determined by pre-rift events showing the importance of inherited structures and of the pre-rift evolution of a margin.
3. Weakening processes change during rifting from mechanical weakening along faults to fault localization by serpentinization and magmatic processes, both of which are temperature dependent.
4. Entirely non-magmatic margins do not exist. Indeed some magmatism may even be a pre-requisite to exhume mantle rocks to the sea floor.
5. The observed changes in basin architecture across the margin reflect the temporal and spatial evolution of the bulk rheology of the lithosphere during extension.

Acknowledgments This paper is the result of a promise I gave long time ago to Wolfgang Franke to write a review paper about rifted margins exposed in the Alps. This paper results from a long, intense and successful collaboration with Daniel Bernoulli, Niko Froitzheim, Othmar Müntener, Laurent Desmurs, Andreas Hölker and many colleagues, too numerous to reference everybody here. I would like to thank the GDR marges (France), the Swiss National Science Foundation, and the Ocean Drilling Program for their support. I thank also Bob Whitmarsh and Yves Lagabrielle for helpful reviews and Chris Wilson for comments on an earlier version of this paper.

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