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Architecture and tectonic evolution of nonvolcanic margins: Present-day Galicia and ancient Adria

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Abstract. A comparison of the reconstructed southeastern margin of the Tethys ocean with the present-day Galicia margin shows that although both margins are of different age and had a different fate, their architectures and tectonic evolutions are very similar. Along both non-volcanic margins the site of rifting shifted from a broad area in the future proximal margins to a localized area in its distal parts, accompanied by a change in the mode of extension. During the initial phase of rifting, extension was accommodated by symmetrically arranged listric faults which soled at midcrustal levels, indicating that deformation in the upper crust was decoupled from deformation in the upper mantle along a hot and weak lower crust. During advanced rifting, extension was dominated by simple shear along low-angle detachment faults with a top-to-the-ocean sense of movement. These shallow crustal structures formed a series of breakaways in the continental crust and cut into mantle rocks, indicating that now deformation in the upper crust and in the upper mantle was no longer decoupled. Cooling and strengthening of the lower crust during an initial stage of rifting apparently led to localization of deformation and a different style of deformation, documenting that the tectonic evolution of nonvolcanic margins is largely controlled by the thermal state of the lithosphere. Seafloor spreading initiated only after exhumation and exposure of the subcontinental mantle on the ocean floor and may have been accompanied by a loss of the yield strength of the upper mantle, due to a combination of simple shear extension, asthenospheric uplift, and increased melt production.

1. Introduction

Present-day divergent, nonvolcanic margins record one of the major plate tectonic processes, i.e., the rifting and breakup of continental lithosphere and the exhumation of subcontinental mantle preceding the onset of seafloor spreading. In present-day passive margins the reconstruction of their early histories relies mainly on the interpretation of reflection seismic profiles, magnetic and gravity data, and, in a few places along sediment-starved margins like the one of Galicia, on the analysis of samples collected by deep-sea drilling and from submersibles. By contrast, mountain belts provide extensive outcrops of oceanic and continental basement rocks, of overlying prerift, synrift, and postrift sediments, and of associated fault rocks. However, in these areas the continent-ocean transition is often tectonically decoupled, the elements of the former margin are dismembered, and only under favorable conditions can the history of the margin be reconstructed. This is the case in a few segments of the Alps, particularly in the south Pennine-Austroalpine boundary zone of the eastern Alps and in the southern Alps, where well-preserved rift-related faults allow for the palinspastic reconstruction of the passive margin and the ocean-continent transition zone.

One of the best investigated present-day continental margins is the nonvolcanic margin west of Iberia from which a wealth of data from deep-sea drilling, submersibles, and reflection seismic profiling is avalaible. Although the overall structure of the margin is well known and a large amount of excellent data exists, several contrasting models for the evolution of this margin have been proposed (compare Figure 3 of *Reston et al.* [1996]), and there appears to be little consensus on important questions such as the mode of extension (pure shear versus simple shear or a combination of both, symmetrical or asymmetrical margins) and the kinematics of the fault systems involved.

In this paper we shall compare the early history of the Cretaceous nonvolcanic margin of Galicia with that of a Jurassic margin of the Alpine Tethys which was involved in Alpine thrusting and nappe formation. Our comparison is based on stratigraphic, structural, petrological, and geochronological data from the eastern and southern Alps, both published and our own, on published geophysical and deep-sea drilling material and on personal observations of deep-sea drilling cores. We shall document the striking analogies in the architecture of these margins and reconstruct their tectonic and kinematic evolution from initiation of rifting to mantle exhumation and beginning seafloor spreading. Also, we shall discuss the mode and kinematics of detachment faulting and the role of the lower crust during rifting. We postulate that the similarities between the two margins and their common evolution from rifting distributed over a wide area to localized extension along low-angle detachment systems may be explained by a model that includes changes in the rheology of the continental lithosphere during progressive rifting. We argue for asymmetric rifting dominated by simple shear extension during the late phases of rifting, leading to the exhumation of subcontinental mantle, and for a lower plate position of both margins as opposed to the Newfoundland and Briançonnais margins. We further document in detail the analogies between the Galicia and Adriatic margins, whereas in an earlier short contribution [*Manatschal and Bernoulli*, 1999] we outlined the general situation of the Galicia-Newfoundland and Adria-Briançonnais margin pairs without a detailed reference to their structural, sedimentological, and metamorphic evolution.

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2. Paleogeographic and Tectonic Framework

2.1. General Situation

Late Triassic to Early Cretaceous rifting and opening of the different segments of the Mesozoic (Neo-)Tethys spatially followed, in the Atlantic and western Mediterranean area, approximately the Variscan orogeny. 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 (Figure 1c). The evolution of these different branches of Tethys was determined by the movements of the North American, the African, the Eurasian, and the smaller Iberia and Adria plates (Figure 1c) [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 from Adria. In the Early Cretaceous the Iberian plate separated from Laurasia and migrated eastward, leading to the opening of the Iberian Atlantic, the Bay of Biscay, and possibly also of another small ocean basin, the Valais ocean (Figure 1b) [cf. *Stampfli*, 1993]. Rotation of Iberia was kinematically linked with the opening of the Bay of Biscay and may have been associated with minor margin-parallel transfer movements along the west Iberian margin; however, most of the admittedly oblique extension was accommodated by the emplacement of oceanic crust and lithosphere. The different segments of the Atlantic-Tethyan ocean system opened at different times, and the closing of the Liguria-Piemonte segment was contemporaneous with ongoing spreading in the central and Iberian Atlantic and the opening of the North Atlantic (Figure 1b).

2.2. The Galicia Margin

The Galicia margin off NW Spain and Portugal (Figure 2) is a typical example of a non-volcanic, sediment-starved margin. It resulted from rifting and final breakup between the North American and the Iberian plates during Early Cretaceous time (Figure 1b). From this margin and its southern prolongation, a wealth of reflection seismic, magnetic, and gravity data were published, and rock samples are available from expeditions of the Deep Sea Drilling Project and the Ocean Drilling Program (Legs 47B, 103, 149, and 173), dredging, and submersible cruises. In this paper we shall concentrate on the northern portion of the Iberian margin, id est the Galicia margin west of the Galicia province in northwestern Spain. This portion of margin can be subdivided into three segments: the Interior and Porto Basins, Galicia Bank, and the Deep Galicia Margin (Figure 2). These segments can be distinguished by their different bathymetry and geological evolution during rifting.

2.2.1. The Interior and Porto Basins. The north-south trending Interior and Porto Basins form the proximal part of the margin and are considered to be the northward extension of the on-land Lusitanian Basin [*Montenat et al.*, 1988]. The two basins are bounded by normal faults which trend NNE-SSW and NW-SE and dip toward the basin centers, resulting in an assembly of moderately tilted blocks along the flanks and near-symmetrical grabens in the basin centers (Figure 2c). Transfer faults with a NE-SW to ENE-WSW orientation segment the basins. As no wells have been drilled in the Interior Basin, the reconstruction of its sedimentary and stratigraphic evolution is based mainly on correlation with

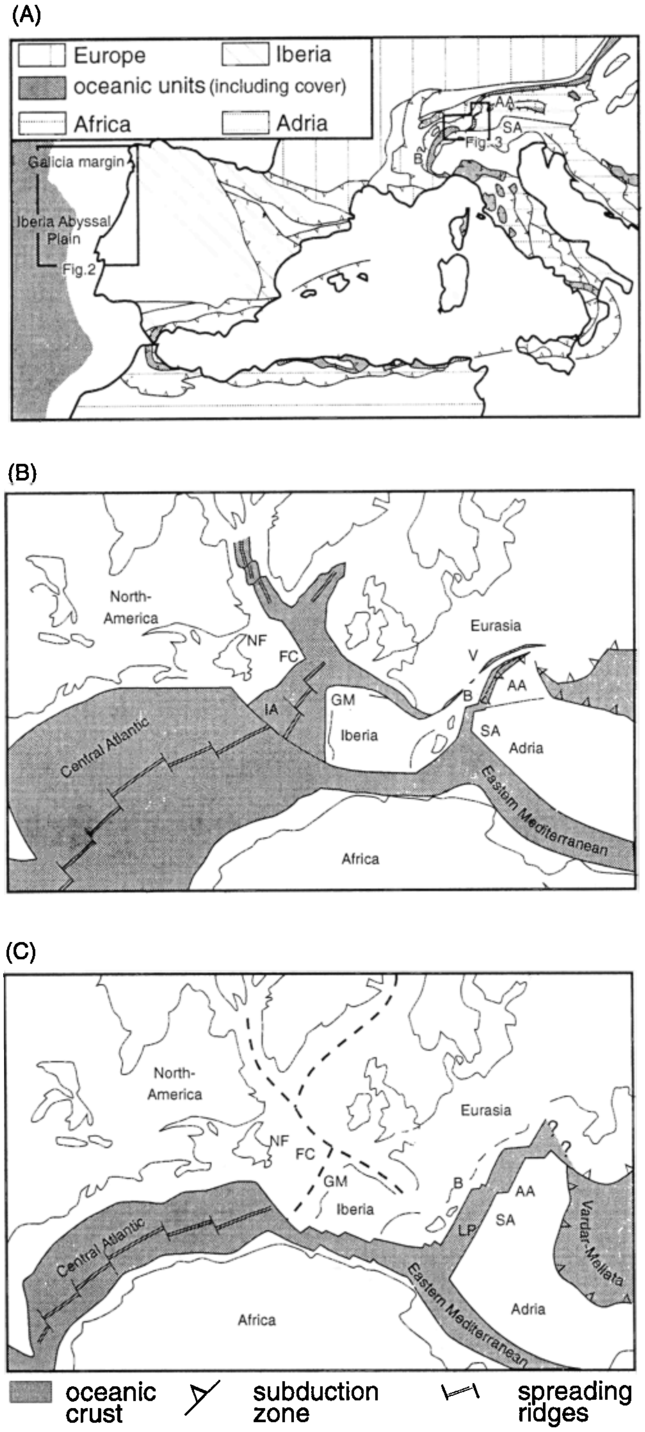


Figure 1. (a) Tectonic sketch map of the present-day west Mediterranean area showing the distribution of the European, Iberian, Adriatic, and African continental areas and the distribution of the tectonic elements derived from them. Oceanic units are remnants of the Mesozoic Tethys. (b,c) Large-scale paleogeography reconstructed for the Late Cretaceous (Figure 1b) and for the Late Jurassic (Figure 1c). AA, Austroalpine; B, Briançonnais; FC, Flemish Cap; GM, Galicia margin; IA, Iberian Atlantic; LP, Liguria-Piemonte ocean; NF, Newfoundland; SA, southern Alps, V, Valais ocean.

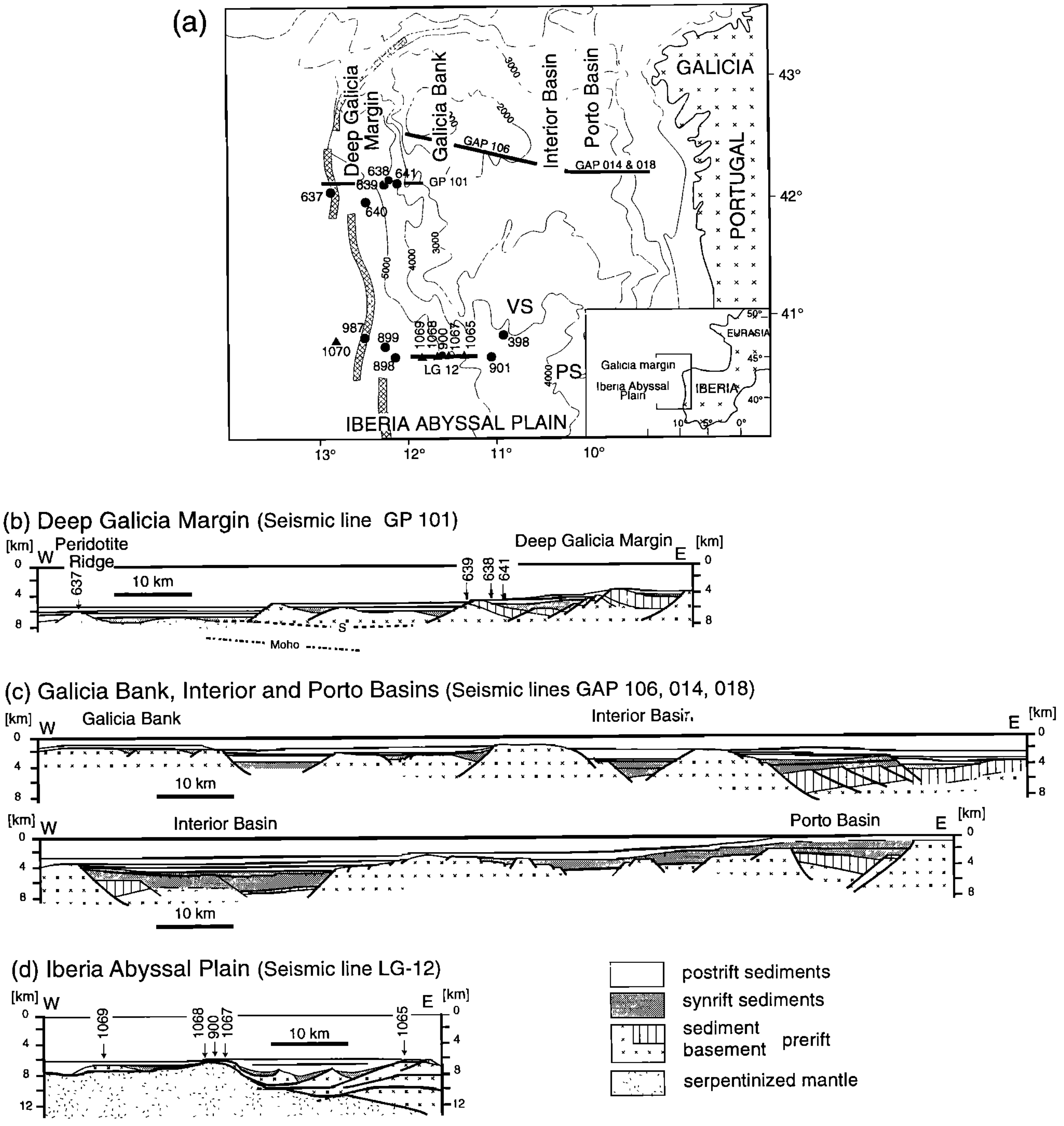


Figure 2. (a) Map of the Iberian margin west of Galicia and Portugal with the locations of Deep Sea Drilling sites and of seismic lines GP 101 and GAP 106, 014, and 018. Cross-hatched band marks the peridotite ridge. PS, Porto seamount; VS, Vigo seamount. (b) Seismic line GP 101 across the Deep Galicia Margin (after the work of *Mauffret and Montadert* [1987], discussed by *Pinheiro et al.* [1996]. Numbers refer to Ocean Drilling Program (ODP) Leg 103 Sites. S, *S* reflector. (c) Seismic lines GAP 106, 014, and 018 across the Galicia Interior Basin (after the work of *Murillas et al.* [1990], discussed by *Pinheiro et al.* [1996]). (d) Seismic line LG-12 across the Iberia Abyssal Plain after *Krawczyk et al.* [1996] with new interpretation.

wells in the Porto Basin and in the Deep Galicia Margin and on seismic stratigraphic correlation. In the Interior Basin, several events of extensional deformation can be defined, of which the most important is interpreted to be of Valanginian age [*Murillas et al.*, 1990].

2.2.2. Galicia Bank. Together with the Vigo and Porto seamounts, Galicia Bank forms a NNW-SSE trending alignment of elevated highs which separate the Interior Basin from the Deep Galicia Margin. Several authors [e.g., *Boillot et al.*, 1979] interpreted these highs as former horsts originating from Mesozoic rifting and uplifted during Tertiary compression. However, the geometry of the reflections, the nature of the seismic stratigraphic units, and the architecture of the basins are different east and west of Galicia Bank, suggesting that this ridge separates two different tectonic provinces of the margin.

2.2.3. Deep Galicia Margin. The Deep Galicia Margin includes the distal part of the margin and the continent-ocean transition, and it is characterized by N-S trending tilted fault blocks of continental basement and prerift sediments overlain by relatively thin synrift and postrift sequences. These tilted blocks are underlain by a prominent reflection, the so-called *S* reflector (Figure 2b) [*de Charpal et al.*, 1978; *Boillot et al.*, 1988]. The *S* reflector can be followed oceanward toward 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. Rifting in the distal margin and breakup between Newfoundland and Iberia are thought to have taken place from Valanginian to Aptian times [*Boillot et al.*, 1988].

2.3. The Adriatic Margin

2.3.1. Alpine tectonic evolution. By and large, the Alps are the product of continental collision between the Eurasian, the Iberian, and the Adriatic plates (Figure 1). This collision was preceded by the opening of three different oceans, the Vardar - Meliata (-Hallstatt) ocean to the east of Adria, the Liguria-Piemonte ocean between Adria and Briançonnais (a possible promontory of the Iberian plate), and the Valais ocean between Briançonnais and Europe. In the Alps, convergence started in the Early Cretaceous with the elimination of the Vardar - Meliata ocean and the collision with another microplate to the east [*Froitzheim et al.*, 1996]. As part of this orogeny, a thrust wedge propagated from the northeastern margin of Adria to the west, leading to a nappe edifice in the Austroalpine complex and its margin to the Liguria-Piemonte ocean. During this event the ocean-continent transition of the Piemonte-Adriatic boundary zone was telescoped into a number of thrust sheets.

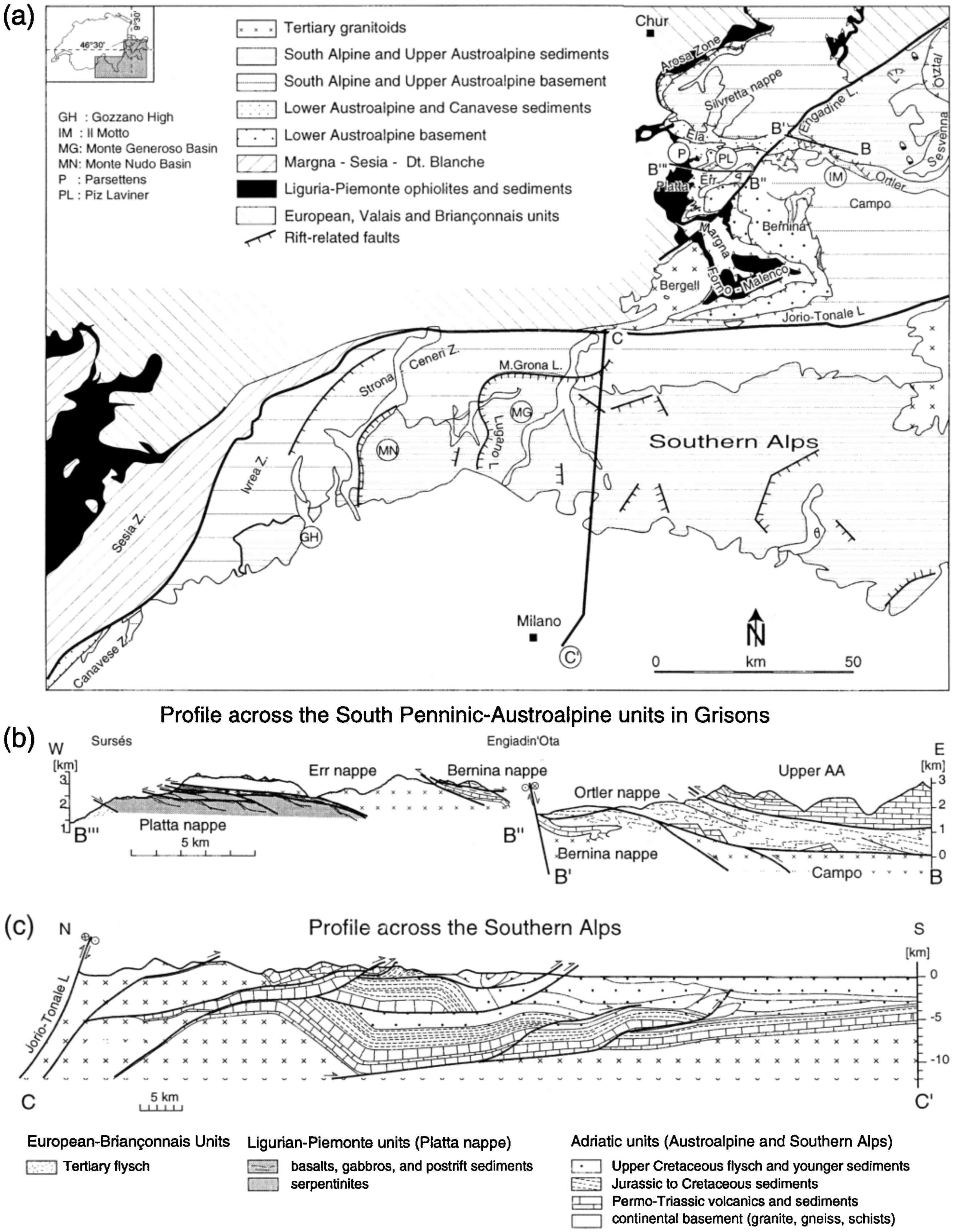
During the Late Cretaceous and early Tertiary, tectonic activity shifted to the west, and most of the Liguria-Piemonte ocean was subducted below the Adriatic plate (Figure 1b). During the Eocene, subduction and related metamorphism affected also parts of the Briançonnais, the Valais ocean, and, later on, the lower part of the European lithosphere, whereas the upper crust was delaminated and accumulated in a number of crystalline basement and sedimentary decollement nappes [*Pfiffner et al.*, 1997]. The thrust sheets derived from the Adriatic margin including the most continentward parts of the Liguria-Piemonte ocean, i.e., the Platta nappe, were always part of the upper plate and escaped subduction and high-pressure metamorphism.

The Tertiary continent-continent collision with the Briançonnais and with Europe affected the south Pennine-Austroalpine boundary zone of Grisons only weakly. The Cretaceous nappe edifice was thrust more or less "en bloc" to the north and over the middle and north Penninic units (Briançonnais and Valais). In contrast, a southward propagating thrust wedge affected the domains of the Adriatic margin located today in the southern Alps, which were less deformed by the previous Late Cretaceous orogeny [*Schumacher et al.*, 1997].

2.3.2. The south Pennine-Austroalpine transect. The evolution of the Adriatic margin can be reconstructed along two transects in the Alps (Figures 3 - 5). In Grisons the elements of the former margin have been telescoped by thrusting during the Late Cretaceous; however, the ocean-continent transition has never been subducted to great depth, and metamorphism did not exceed lower greenschist facies along our transect [*Ferreiro Mählmann*, 1995, and references therein]. West directed shortening was of the order of 200 - 300% (about 100 - 150 km). The resulting nappe edifice includes (from top to bottom) the upper Austroalpine (Ötztal, Silvretta-Sesvenna, and Campo-Ortler), the lower Austroalpine (Bernina-Ela and Err nappes), and the south Pennine nappes (Arosa zone, Platta nappe, and Forno-Malenco complex) (Figures 3a and 3b).

Late Cretaceous nappe stacking was followed by extension which occurred still during the Late Cretaceous [*Froitzheim et al.*, 1994, 1996]. The amount of extension, however, was small compared with the amount of previous shortening. Compression and extension were parallel to previous extension during Jurassic rifting, and part of the Jurassic basins was inverted along rift-related east dipping faults. During this process, higher crustal levels including sediments and shallow basement were detached from their original basement; however, the coaxial Cretaceous phases of deformation allowed us to visualize a relatively straightforward kinematic inversion of the margin (for details, see *Froitzheim et al.* [1994] and *Manatschal and Nievergelt* [1997]). According to this kinematic reconstruction, the higher nappe units are derived from the proximal margin (upper Austroalpine) whereas the lower nappes represent the distal margin (lower Austroalpine, Err and Bernina nappes) and the transitional crust (Platta nappe). Because of the detachment of shallow crust and sediments, the deep structures of the margin are not preserved in our transect; however, deeper crustal levels of the prerift stage are exposed to the south in the Austroalpine Margna nappe and the ultramafic Malenco complex [*Hermann et al.*, 1997; *Müntener et al.*, 1999]. In contrast, shallow crustal structures of the proximal margin (Ortler-Silvretta) [*Eberli*, 1988; *Froitzheim*, 1988; *Conti et al.*, 1994], the distal margin (Err) [*Handy et al.*, 1993; *Manatschal and Nievergelt*, 1997) and of the ocean

Figure 3. (a) Tectonic sketch map and distribution of early Mesozoic faults of the western southern Alps and the south Pennine-Austroalpine boundary zone, Italy, and southeastern Switzerland (modified after *Bernoulli et al.* [1990] and *Schönborn* [1992]). (b) Profile across the south Pennine-Austroalpine units in Grisons (modified after *Froitzheim et al.* [1994] and *Manatschal and Nievergelt* [1997]). (c) Profile across the southern Alps [after *Schönborn*, 1992]. For traces of the profiles, see Figure 3a.


(a)
46°30'
9°30'
GH : Gozzano High
IM : Il Motto
MG: Monte Generoso Basin
MN: Monte Nudo Basin
P : Parsettens
PL : Piz Laviner
Tertiary granitoids
South Alpine and Upper Austroalpine sediments
South Alpine and Upper Austroalpine basement
Lower Austroalpine and Canavese sediments
Lower Austroalpine basement
Margna - Sesia - Dt. Blanche
Liguria-Piemonte ophiolites and sediments
European, Valais and Briançonnais units
Rift-related faults
Chur
Arosa Zone
Silvretta nappe
Engadine L.
Ötztal
Sesvenna
Ela
B'
B
P
PL
B'''
Err
B''
IM
Ortler
Platta
Campo
Bernina
Margna
Forno - Malenco
Bergell
Jorio-Tonale L
C
Strona - Ceneri Z.
M.Grona L.
MG
MN
Lugano L.
Southern Alps
Ivrea Z.
Sesia Z.
GH
Milano
C'
N
0 km 50
Canavese Z.
Profile across the South Penninic-Austroalpine units in Grisons
(b)
W
Sursés
Engiadin'Ota
E
[km]
3
2
1
Err nappe
Bernina nappe
Upper AA
Ortler nappe
B'''
Platta nappe
5 km
B''
Bernina nappe
B'
Campo
B
0
(c)
N
Profile across the Southern Alps
S
[km]
0
-5
-10
Jorio-Tonale L
C
5 km
C'
European-Briançonnais Units
Tertiary flysch
Ligurian-Piemonte units (Platta nappe)
basalts, gabbros, and postrift sediments
serpentinites
Adriatic units (Austroalpine and Southern Alps)
Upper Cretaceous flysch and younger sediments
Jurassic to Cretaceous sediments
Permo-Triassic volcanics and sediments
continental basement (granite, gneiss, schists)

floor sequence (Platta) (*Dietrich* [1969, 1970] and this paper) are well preserved along our transect. Younger, N-S directed shortening and associated transcurrent faulting during Tertiary time are of subordinate importance within our transect [*Froitzheim et al.*, 1994] and are not considered in the kinematic inversion and palinspastic reconstruction.

2.3.3. The southern Alps transect. The southern Alps in southern Switzerland and northern Italy preserve another transect across the same margin to the south (Figures 1, 3a, and 3c) [*Bernoulli et al.*, 1979; *Winterer and Bosellini*, 1981]. Basement and sediments of the proximal margin segment are well exposed in the central portion of the southern Alps north of Milano, but the distal part, situated in the Canavese zone, is poorly outcropping and heavily deformed (Figure 3a). A precise reconstruction of the distal margin is therefore not possible along this transect. In the central Lombardian segment, north-south shortening during the Late Cretaceous and Tertiary was between 80 and 100 km and parallel to the strike of the Jurassic rift faults which were partly reactivated as transcurrent faults and oblique ramps within the east-west trending fault-and-thrust belt (Figure 3c) [*Schönborn*, 1992]. Nevertheless, a few of the Jurassic rift faults can be followed into the basement [*Bertotti*, 1991], and the anatomy of the proximal margin can be reconstructed from the depositional geometries of the well-exposed synrift sediments [*Bertotti et al.*, 1993].

Both sectors of the margin, the northern and the southern transect, are divided into segments with a characteristic structural and sedimentary evolution which is discussed in section 3. These segments are from west to east (1) a distal margin (Err in the north and Canavese in the south), (2) a boundary zone between the distal and the proximal margins (Bernina-Ela in the north and Gozzano in the south); and (3) a proximal part of the margin (Ortler-Silvretta in the north and Lombardian basin in the south) (Figures 4b and 4C).

3. Anatomy of the Margins

3.1. Continental Crust and Prerift Sediments

3.1.1. Galicia margin. The continental basement of the Galicia margin is known from very limited samples from drill holes and submersible dives, and its age and tectonic evolution are not well defined. Plutonic and metamorphic rocks [e.g., *Capdevila and Mougenot*, 1988] are locally overlain by thick, weakly metamorphosed sandstones, dolomites, and volcanoclastics of Late Devonian-early Carboniferous age [*Mamet et al.*, 1991]. A similar sequence is found in the Ossa-Morena zone in Portugal and southern Spain. This suggests that the Galicia margin is underlain by the northwestward prolongation of this external zone of the Variscan orogen. The occurrence of undeformed lower Paleozoic rocks at Flemish Cap in the conjugate Newfoundland margin [*King et al.*, 1986] indicates that this latter area was situated outside the Variscan belt and that the breakup between Newfoundland and Iberia followed more or less the Variscan front [*Capdevila and Mougenot*, 1988].

Lower crustal rocks have not been found so far along the Deep Galicia Margin, except for a few granulite samples. Fission track ages on zircons from two of them were Carboniferous to Early Permian [*Fuegenschuh et al.*, 1998], indicating that these rocks cooled to about 250°C at this time and were at upper crustal levels when rifting initiated.

Little is also known about the prerift sediments. In the Deep Galicia Margin, sandstones with volcanic detritus, interbedded with shaly dolomites and conglomerates, are overlain by about 400 m of Tithonian shallowwater carbonates. On Galicia Bank the basement is separated from these carbonates by a horizon which shows evidence of soil development during Mesozoic emergence [*Mamet et al.*, 1991]. Thus the area occupied by the Galicia margin was in shallow water or at sea level before rifting started in Cretaceous time.

3.1.2. Continental crust and prerift sediments of the Adriatic margin. In the southern Alps a pre-upper-Carboniferous (Variscan) metamorphic basement underlies the upper Paleozoic and Mesozoic formations. Some controversies exist about the importance and significance of earlier, Cadomian or early Paleozoic events which are documented by an increasing amount of radiometric data. Amphibolite-grade metamorphic mineral assemblages in the Variscan basement document extremely low-pressure (p), high-temperature (T) metamorphism as it is characteristic for extensional areas with a high heat flow [e.g., *Lardeaux and Spalla*, 1991]. Post-Variscan, Early Permian (290 Ma) extension is also documented by syntectonic intrusions of gabbros along the crust-mantle transition in the Ivrea zone [*Quick et al.*, 1994], accompanied by granulite formation and partial melting in the lower crust and the emplacement of granitic batholiths in the middle and upper crust, both contemporaneous with the formation of sedimentary basins and the extrusion of andesitic, dacitic, and rhyolitic volcanic rocks at the surface.

Like in the Ivrea zone, gabbros intruded during the Permian at the crust-mantle transition of the Malenco complex [Hermann *et al.*, 1997]. These gabbros crystallized at 10 - 12 Kbar, indicating that the Moho was at that time at about 30 km depth. Pressure-temperature-time data obtained from the gabbros and the surrounding granulites and ultramafic rocks document isobaric cooling which lasted for about 50 Myr, before these rocks were exhumed during Late Triassic and Early Jurassic rifting [*Müntener et al.*, 1999].

The sedimentary prerift sequence of the Adriatic margin starts with Upper Permian to Middle Triassic continental clastics of variable thickness. These subaerial deposits are overlain by Middle to Upper Triassic dolomites, limestones, and minor evaporites and shales. Varying thicknesses indicate differential subsidence and local tectonic activity, possibly in connection with ongoing transtensional/transpressional movements [*Handy and Zingg*, 1991] or events connected with the evolution of the Vardar-Meliata ocean to the east [*Bertotti et al.*, 1993]. The prerift sequence is generally thicker (1 - 5 km) in the east (upper Austroalpine) and south (southern Alps) and thinner (< 500 m) in the lower Austroalpine in the northwest.

3.1.3. Similarities in the prerift history of the two margins. The very different quality and density of data, as well as the poorly known crustal evolution of the Austroalpine and south Alpine areas during the Permian and Triassic, make a direct comparison of the continental basement and the prerift sedimentary sequence of the two margins difficult. There are, however, important similarities at least as far as the lower Austroalpine and the western south Alpine basement are concerned. For both margins we can assume (1) a Variscan overprint of the continental crust and (2) an isostatically equilibrated, about 30 km thick continental crust before initiation of rifting, as is indicated by limited subsidence and shallow water conditions prevailing over a long time before initiation of rifting and by the p-T-t data of the mantle-crust boundary in the Malenco area [*Müntener et al.*, 1999]. We therefore think that the continental crust underlying the two margins had a similar

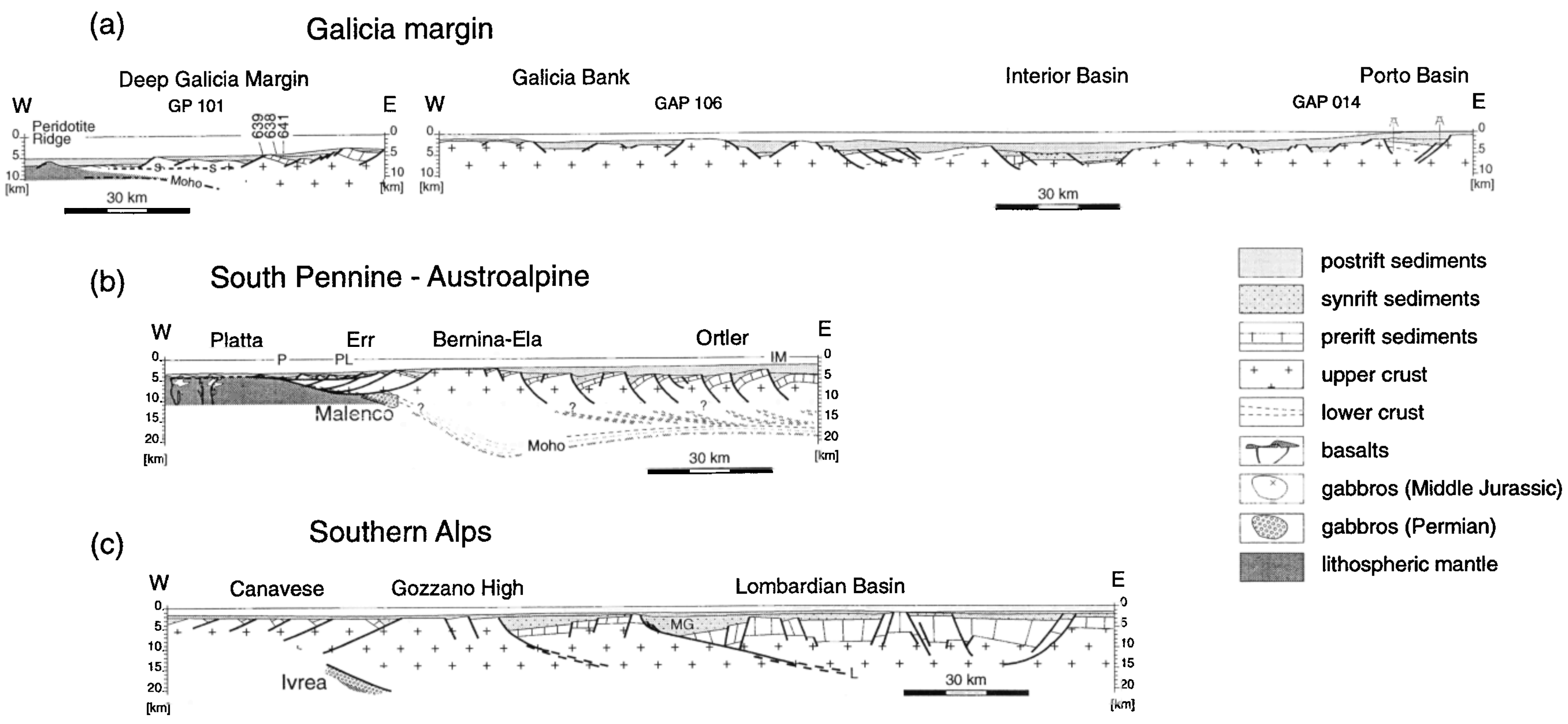


Figure 4. Reconstructed geological profiles across the Iberian (today) and Adriatic margins (Late Jurassic). (a) Galicia margin (modified after *Murillas et al.* [1990]). (b) South Pennine-Austroalpine boundary zone. IM, Il Motto; P, Parsettens; PL, Piz Laviner. (c) Southern Alps (modified after *Bertotti et al.* [1993]). MG, Monte Generoso Basin; L, Lugano-Monte Grona fault.

tectonic and thermal evolution before the onset of rifting and that the conditions at the onset of rifting were similar for the two margins.

3.2. Rift Basins and Synrift Sediments

Rift basins, bounded by normal faults and tilted blocks, are the most prominent structures defining the architecture of rifted margins (e.g., Figures 2b and 2c). Because the formation of basins and the deposition of synrift sediments are closely linked, they will be discussed together in this section. Rift-related structures, such as the intrabasement reflections and detachment faults, will be discussed in section 4.

The age of the onset of rifting and therefore also the duration of rifting depends on the way rifting is defined. In the Alpine Tethys, Permian and Middle Triassic extension have been interpreted by several authors as precursors of the opening of the Tethyan ocean [e.g., *Winterer and Bosellini*, 1981]. Likewise, Late Triassic to Early Jurassic crustal extension and basin formation in the Iberian and North Atlantic area have been related to the later opening of the corresponding ocean basins; however, these early extensional events are separated from break-up by a time interval of some 100 Myr during which tectonic activity and subsidence were discontinuous. Moreover, Permian and Triassic extension and related subsidence occur over large areas of northern Europe which never evolved into oceanic areas. In this paper we shall refer to rifting as the extensional processes continuously and progressively leading to break-up and continental separation. In the Alpine Tethys and along the Iberian margin, crustal extension leading to continental separation lasted about 30 - 50 Myr.

3.2.1. Rift basins and synrift sediments along the Galicia margin. Along the Galicia margin, rifting apparently started in the Interior Basin during the Valanginian, as postrift sedimentary sequences of assumed Hauterivian to Aptian age drape the rift basins [*Murillas et al.*, 1990]. Data from Deep Sea Drilling Project (DSDP) Leg 47B and Ocean Drilling Project (ODP) Leg 103 indicate that rifting along the Deep Galicia Margin was later, from Hauterivian to Aptian time [*Boillot et al.*, 1989a]. These scanty data suggest that rifting along the Galicia margin was diachronous, beginning in the Interior Basin and shifting only later to the Deep Galicia Margin. A similar spatial and temporal evolution of rifting is reported from the transect farther to the south between the Iberian Abyssal Plain and the Lusitanian Basin. In this transect, rifting started in late Oxfordian-early Kimmeridgian time in the proximal Lusitanian Basin [*Wilson et al.*, 1989] and shifted after the Tithonian, probably during the early Valanginian, to the distal Iberian Abyssal Plain [*Wilson et al.*, 1996]. Likewise, the formation of oceanic crust started earlier in the south (136 Ma) from where it propagated to the north [*Whitmarsh and Miles*, 1995].

In the Interior and Porto Basins most of the basins are bounded by antithetic high-angle and/or listric normal faults, and the overall geometry is that of a horst and graben structure (Figure 2c). The sedimentary fill is much thicker than that in the more distal basins of the Deep Galicia Margin (Figures 2b and 2c). Initial faulting led to drowning as well as to uplift and erosion of parts of the Tithonian-Lower Cretaceous carbonate platform and to the accumulation of a thick clastic sequence in the depocenters of the Porto Basin. Toward Galicia Bank the rift basins become shallower, and normal faulting appears to have been of lesser importance.

In the Deep Galicia Margin, rift basins are bounded by west dipping normal faults separating fault blocks tilted toward the east, i.e., toward the continent. These blocks are underlain by a strong reflection, the so-called *S* reflector. The size of the blocks decreases toward the ocean. The fault blocks are cut off and/or slightly offset by a discrete pattern of transfer faults which strike east-northeast or east-southeast and show a quite regular spacing of about 20 km. The blocks overlying the *S* reflector consist of continental basement rocks and prerift sediments and are unconformably overlain by a synrift sequence including marlstones, turbiditic sandstones, claystones, and hemipelagic limestones interbedded with debris flow deposits of Hauterivian to early Barremian age. In the half grabens these synrift sediments are up to 1 km thick but thin and locally pinch out toward the highs of the tilted blocks along onlap surfaces and internal unconformities [*Boillot et al.*, 1989a]. Quartzo-feldspatic sandy turbidites in the basins west of Galicia Bank are derived from the erosion of a subaerial, narrow high occupying the present area of Galicia Bank [*Winterer et al.*, 1988]. This implies tectonic uplift of Galicia Bank simultaneously with rifting and subsidence in the Deep Galicia Margin.

3.2.2. Rift basins and synrift sediments of the Adriatic margin. In the Alpine transect, rifting started in the areas which were to become the proximal parts of the future margin(s). In the Adriatic margin the geometry of the rift basins can be reconstructed from the relationships between prerift, synrift, and postrift sediments, their depositional geometries, and their relations to Jurassic fault scarps. At places, the Jurassic faults have not been reactivated, and their geometry can be observed directly [*Eberli*, 1988; *Froitzheim*, 1988; *Bertotti*, 1991; *Conti et al.*, 1994].

The first rift basins formed during the latest Triassic. They were depressions only a few kilometers wide in which coarse resediments, derived from active fault scarps, were deposited [*Bertotti*, 1991]. The shape and orientation of these basins could suggest a left-lateral component in their formation. During the Rhaetian, extension continued; however, because of high sedimentation rates, the faults had no major morphological expression [*Bertotti et al.*, 1993].

In the early Liassic the number of active faults decreased, and strain was gradually concentrated along a few major crustal faults. These faults controlled the sedimentation in the adjacent basins, which was characterized by olistoliths, coarse lithic breccias, slump complexes, and calcareous turbidites interbedded with hemipelagic spongolithic limestones [*Bernoulli*, 1964; *Eberli*, 1988]. Locally, these basinal deposits reach a thickness of a few kilometers. On the sediment-starved highs, submarine faulting produced a network of sediment-filled fractures (neptunic dikes) and complex tectono sedimentary breccias. 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 lower Liassic (upper Sinemurian) sediments (Figures 5 and 6f) [*Eberli*, 1988; *Conti et al.*, 1994].

In the southern Alps, facies and thickness of the synrift sediments abruptly change across these Jurassic faults, and their depositional geometries define the rift basins as half-grabens, up to 30 km wide (Figure 4c) [*Bernoulli*, 1964; *Bertotti*, 1991]. The Lugano-Monte Grona fault bounding the Late Triassic-early Liassic Generoso basin to the west can be traced from the surface into the Variscan basement over a horizontal distance of more than 30 km (Figure 7). This fault was active as a top-to-the-east normal fault and can be traced from the brittle into the ductile field where quartz mylonites formed. On the basis of the depositional geometry of the synrift sediments and on the temperature conditions during deformation, *Bertotti* [1991] could show that the fault had a listric geometry and soled out at a depth of about 15 km.

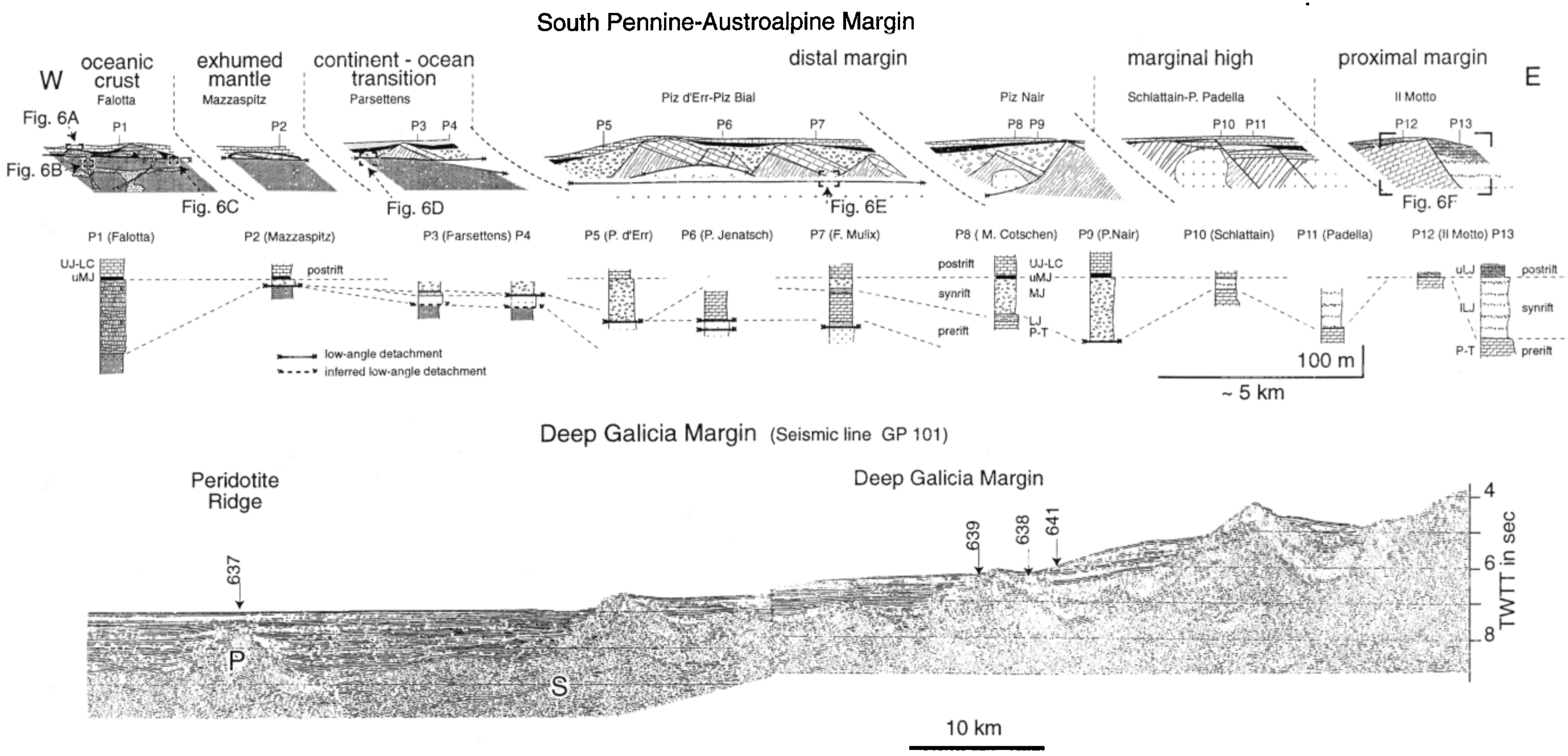


Figure 5. (top) Relationships between sediments, fault zones, and exhumed mantle rocks along the south Pennine-Austroalpine margin and (bottom) comparison with the Deep Galicia Margin. The top section represents portions of the margin, now in different tectonic imbricates, which preserve rift-related structures and depositional contacts with sediments. The corresponding lithostratigraphic columns document measured sections. For further reference to the stratigraphic columns, see *Eberli* [1988] and *Manatschal and Nievergelt* [1997]. Abbreviations are as follows: LC, Lower Cretaceous; UJ, Upper Jurassic; uMJ, upper Middle Jurassic; MJ, Middle Jurassic; LJ, Lower Jurassic; uLJ, upper Lower Jurassic; lLJ, lower Lower Jurassic; P-T, Permian-Triassic. The reflection seismic section below shows a portion of the Deep Galicia Margin including transitional crust in the peridotite ridge and continentward tilted blocks overlain by synrift and postrift sediments in the distal continental margin. P, peridotite ridge; S, *S* reflection, TWTT, two-way travel time. Numbers show locations of ODP sites. From *Boillot et al.* [1988].

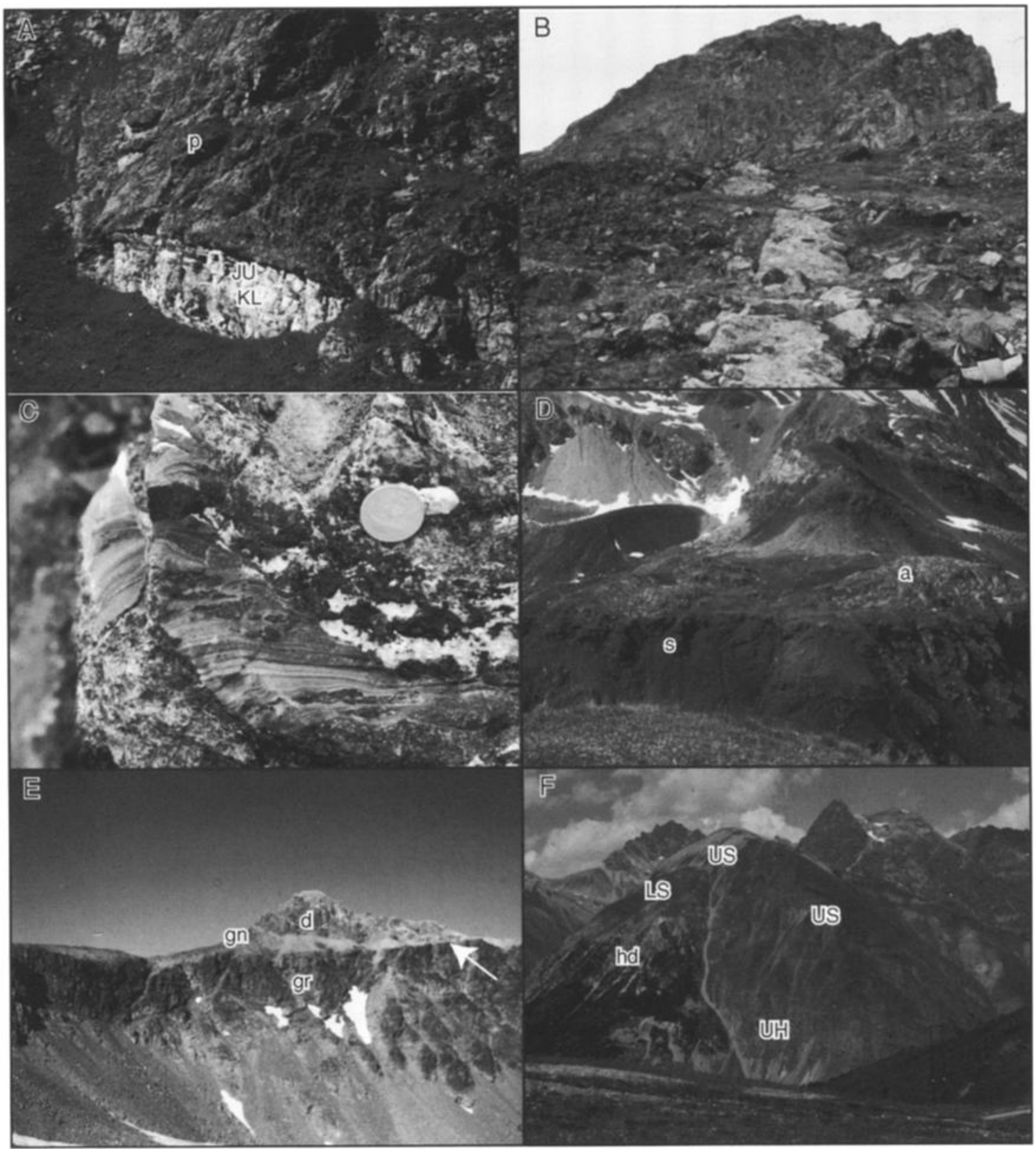


Figure 6. Geometrical relationships between postrift, synrift, and prerift sediments, continental, transitional and oceanic crust, and rift-related faults. (a) Pillow basalt (p), stratigraphically overlain by about 3 m of uppermost Middle to Upper Jurassic radiolarites (JU), and Lower Cretaceous white pelagic limestones (KL). The sequence is overturned. Platta nappe, Val Savriez, Grisons. (b) Undeformed, nonrodingitized basaltic dike cutting across serpentinized peridotites. Serpentinization predates emplacement of the dike (and extrusion of pillow lavas). Platta nappe, Starschagns, Grisons. (c) Pocket in serpentinized peridotite filled by internally deposited sediment. Layers and laminae of serpentinitic arenite with limestone matrix show distinct size-grading and lamination. Together with the radiolarites stratigraphically overlying the serpentinites, these internal sediments document exposure of the serpentinites at the seafloor. Coin is 2 cm in diameter. Arosa zone (lateral equivalent of Platta nappe), Totalp, Grisons. (d) Serpentinized peridotite (s), overlain by an extensional allochthon (a) composed of brittlely deformed granite and orthogneiss. The subhorizontal contact between the serpentinites and the allochthon is locally marked by a black fault gouge, clasts of which also occur in synrift debris flow breccias, documenting the Mesozoic age of emplacement of the allochthon. (e) Low-angle detachment fault (arrow) separating Variscan continental basement (Err granite, gr) below from tilted block composed of Variscan gneiss (gn) and prerift sediments (Lower Triassic sandstones and Middle Triassic dolomites (d), Err nappe, Piz Lavinèr, Grisons. (f) Jurassic high-angle fault separating Upper Triassic prerift sediments (Norian Hauptdolomit Formation (hd) to the left) from Liassic synrift sediments (Allgäu Formation, upper Hettangian (UH), lower Sinemurian (LS), upper Sinemurian (US)) to the right. Synrift sediments are hemipelagic spiculitic limestones with intercalations of thick debris flow breccias and proximal turbidites. The prerift sediments to the left are unconformably overlain by lower Sinemurian breccias, and the fault is sealed by upper Sinemurian basinal limestones. Il Motto, Sondrio Province, Italy.

Fault geometry and basin fill architecture of the south Alpine basins are conspicuously similar to that observed in reflection seismic profiles across the Jeanne d'Arc basin of the Newfoundland margin (Figure 7) [*Keen et al.*, 1987]. This basin is situated on the conjugate margin of the Iberian Atlantic; however, it evolved during early rifting when the evolution of the future margins was still symmetrical (see below).

In the Austroalpine proximal margin, exceptional exposures in the Ortler nappe allowed scientists to identify several high-angle normal faults [*Froitzheim*, 1988; *Conti et al.*, 1994]. Here the different subbasins appear to be smaller than those in the southern Alps; spacing between the faults is only of the order of 5 - 10 km, and the thickness of the basin fill is less than 1 km. At Il Motto (Figures 5 and 6f), one of these high-angle normal faults separates prerift Upper Triassic dolomites to the west from lower Liassic fault scarp-derived breccias interbedded with turbidites and hemipelagic sediments to the east. The synrift sediments show thinning and fining upward cycles and are overlain by pelagic and hemipelagic sediments of late Sinemurian age which also overlie the dolomites to the west sealing the fault. The reconstruction of the original fault geometry shows that near the surface the fault was dipping 60° - 70° toward the east.

In contrast to the basins in the proximal margin, the basins in the distal margin are less well preserved. The reconstruction of the basin geometry is mainly based on the depositional geometries of the synrift sediments which show that these basins were smaller [*Finger et al.*, 1982]. Their spacing was of the order of 3 - 5 km, and the sedimentary fill was only a few hundred meters thick. In many places, the synrift sediments overlie directly low-angle detachment faults exposed at the seafloor.

Rifting along the distal part of the margin (Err domain) initiated later than in the proximal margin, during the late Liassic (Toarcian) or earliest Middle Jurassic: middle Liassic hemipelagic cherty limestones (Agnelli Formation) show a constant thickness across the entire distal margin and no indication of synsedimentary faulting. The limestones terminate with a typical submarine hard ground and are unconformably overlain by deep-water clastics, interbedded with hemipelagic marls and between 200 and 450 m thick (Figure 5, Saluver Formation) [*Finger et al.*, 1982]. The formation includes siliciclastic turbidites, often with a reddish matrix and coarse, unsorted, polygenic breccias with variable amounts of matrix and clasts predominantly derived from the basement. The frequent occurrence of clasts of alkali granite, characteristic for the Bernina nappe [*Spillmann and Büchi*, 1993], indicates uplift and subaerial exposure of parts of the Bernina domain contemporaneous with downfaulting of the Err domain. The age of the Saluver Formation is weakly constrained; it is younger than the lower Pliensbachian hard ground along the top of the Agnelli Formation (H. Furrer, personal communication, 1997) and older than the overlying upper Middle Jurassic Radiolarite Formation [*Baumgartner*, 1987]; therefore a late Liassic (Toarcian) to early Middle Jurassic age may be assumed for these synrift sediments. Contemporaneous with initial faulting in the distal margin, a second pulse of gravity flow sedimentation is noted in the distal part of the proximal margin (eastern Bernina-Ela) [*Eberli*, 1988]. This late Liassic event may be associated with the uplift of the Bernina domain reactivating the fault(s) bounding the adjacent basin to the east. However, in contrast to the siliciclastic gravity flow deposits of the Saluver Formation, fault scarp derived breccias and turbidites of the proximal margins yield no basement clasts but only fragments of Triassic dolomites and limestones and penecontemporaneously displaced carbonate material.

3.2.3. Similarities in the synrift evolution of the two margins. Although the age of rifting is different in the two margins, there are remarkable analogies in their evolution. The duration of rifting along both margins is of the order of some 40 Myr. Along both margins, rifting shifted to the distal margin about 20 Myr before onset of seafloor spreading, reflected by the younging of the synrift and postrift sediments toward the ocean. A similar sedimentary evolution is suggested by comparable facies and thickness variations across the margins, in turn suggesting a similar isostatic response to extension. Another similarity is the change of the basin architecture from large-scale and overall symmetrical in the proximal margin to small-scale and asymmetrical in the distal margin.

3.3. Transitional and Oceanic Crust

3.3.1. Transitional and oceanic crust along the Galicia margin. A major result of ODP Leg 103 was the discovery of serpentinized mantle rocks exposed at the seafloor. The serpentinized mantle rocks occur along a 10 - 12 km wide segmented "ridge" [cf. *Boillot et al.*, 1987] which can be followed over 125 km parallel to the continent-ocean boundary. Basalts and gabbros occur only locally on the ridge itself [*Boillot et al.*, 1995b] but are more common along its western slope. Refraction seismic and magnetic data suggest the existence of a thin oceanic crust west of the peridotite ridge [*Whitmarsh et al.*, 1993]. On the eastern flank of the ridge the Galinaute II submersible cruise [*Boillot et al.*, 1995b] sampled a breccia with fragments of ultramafic, mafic, and continent-derived rocks. This breccia has been interpreted by *Boillot et al.* [1995b] as a tectonic breccia formed during Early Cretaceous rifting along a low-angle, brittle detachment fault.

The seismic velocity structure and magnetic anomalies of the peridotite ridge are distinctly different from those expected from either a true oceanic crust or a thinned continental crust [e.g., *Discovery 215 Working Group*, 1998]. Therefore this type of crust has been termed transitional [*Whitmarsh and Sawyer*, 1996]. The transitional crust is assumed to consist mainly of serpentinized peridotites separating thinned continental crust to the east from true ocean crust to the west. The serpentinization front is assumed to lie, on the basis of the seismic velocity structure, at 5 - 6 km below the seafloor [*Boillot et al.*, 1988] and to coincide with the Moho reflection observed in seismic profiles.

Samples collected by dredging, by drilling, and from submersibles show that the transitional crust consists of serpentinized plagioclase-bearing harzburgite and lherzolite [*Beslier et al.*, 1990], locally cut by rare plagioclase-rich veins and dioritic intrusions. The petrological and structural evolution of the serpentinized peridotite documents the following history of exhumation: (1) partial melting of the peridotite at temperatures of 1250° - 970°C, (2) formation of mylonites at temperatures decreasing from about 1000° to 850°C; (3) crystallization of pargasite (900° - 800°C) and other amphiboles under static conditions (750°C), (4) beginning hydrothermal alteration, (5) serpentinization (<300°C), and (6) brittle fracturing and filling of the fractures with serpentinite fragments and calcite cements (Figure 8), [cf. *Evans and Girardeau*, 1988; *Girardeau et al.*, 1988; *Beslier et al.*, 1990).

Foliated dikes of dioritic composition cutting across the serpentinites include brown amphibole overgrowing a high-

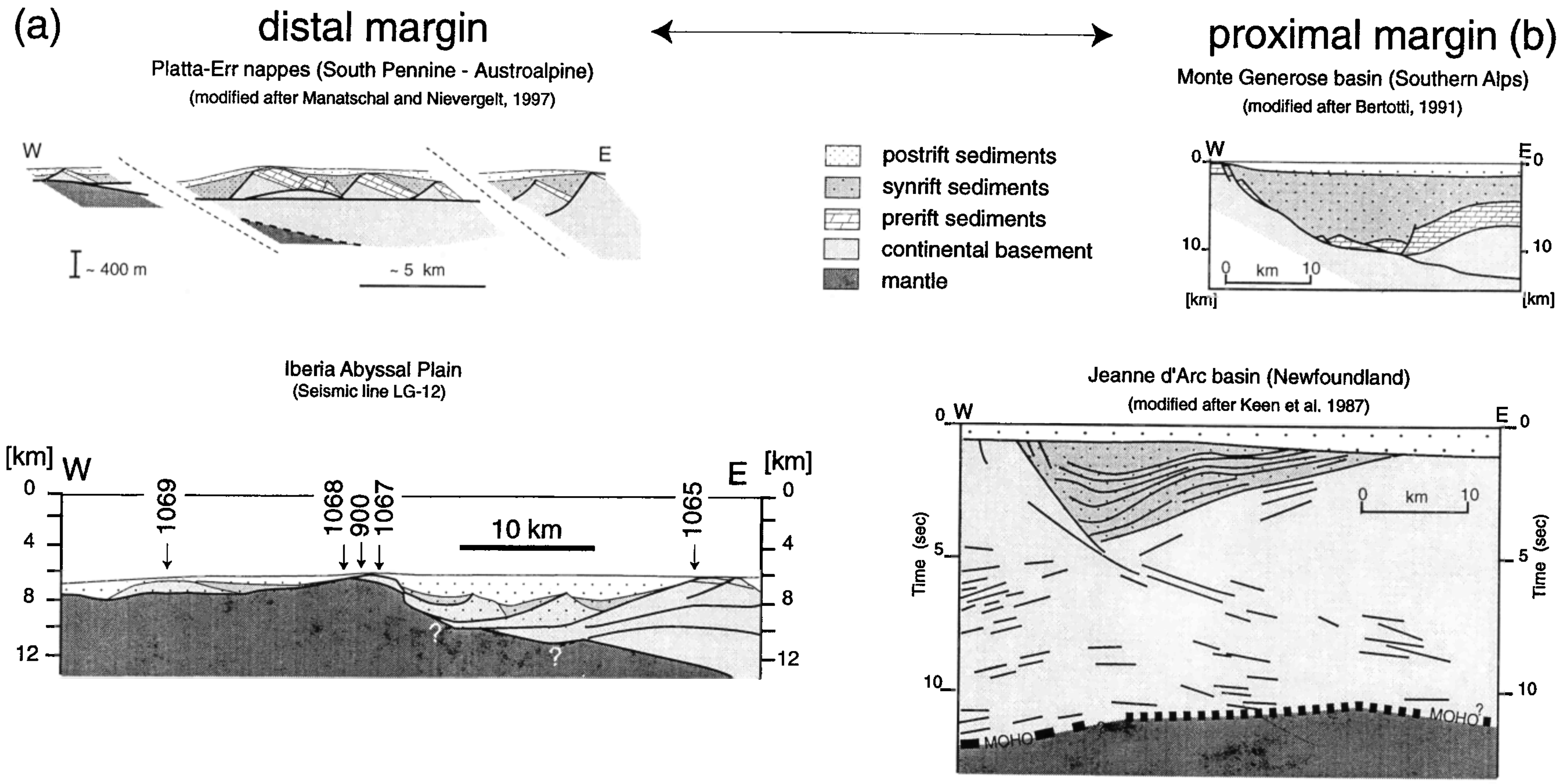


Figure 7. Architecture of rift basins in (a) distal margins and (b) proximal margins. Margins of the Liguria-Piemonte ocean (top cross sections) and the Atlantic (bottom cross sections) are shown, documenting different styles of extensional faulting, basin geometries, and mantle exhumation during early and late rifting. Examples in top cross sections are based on field studies from the Alps, and those in the bottom cross sections are based on seismic profiles from present-day margins.

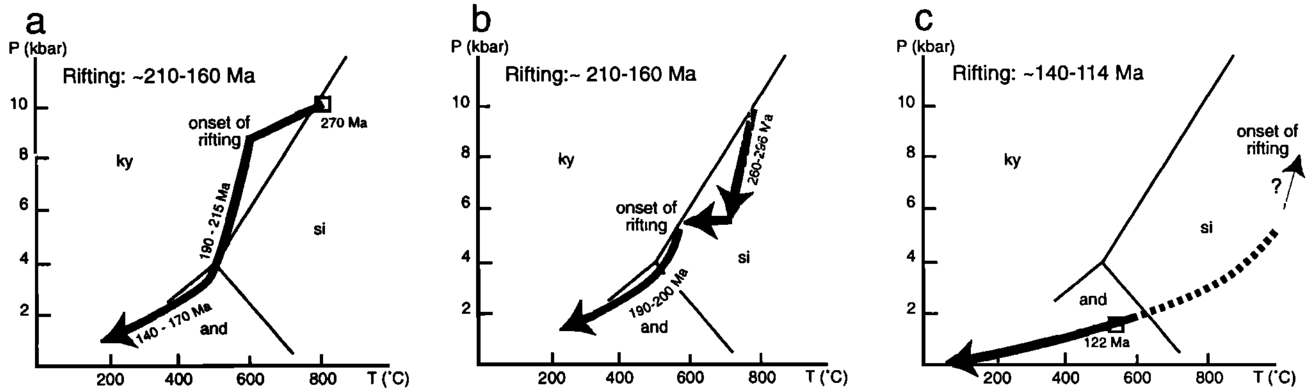


Figure 8. P-T-t paths of lower crustal and mantle rocks from the Adriatic and Galicia margins: (a) Malenco (granulites), (b) Ivrea (granulites), and (c) Galicia (peridotite). The examples of the Malenco and Ivrea granulite terrains have been compiled by *Hermann* [1997]. For locations, see Figures 3 and 4, and for details and further references, see *Müntener et al.* [1999] for Malenco and *Handy and Zingg* [1991, and references therein] for Ivrea. For the Galicia margin, see *Boillot et al.* [1995a, and references therein].

temperature foliation. The ^{39}Ar/^{40}Ar dating of amphiboles from one of these dikes yielded a well-constrained plateau age of 122±0.6 Ma [*Féraud et al.*, 1988]. This age is compatible with the U-Pb ages of 13 different zircon fractions from a gabbro and a sheared chlorite-bearing schist from the same locality, which yielded an identical age of 122±0.3 Ma [*Schärer et al.*, 1995]. Thus dike and gabbro emplacement predates the breakup, generally assumed to occur at 114 Ma along the Galicia margin [*Boillot et al.*, 1988], indicating the occurrence of synrift magmatic activity.

Basalts collected during submersible dives along the northwestern Galicia margin and from the peridotite ridge show a tholeiitic signature free of continental contamination [*Kornprobst et al.*, 1988; *Malod et al.*, 1993; *Charpentier et al.*, 1998]. Rare earth element patterns and isotopic ratios (Nd and Sr) grade from gently enriched to moderately depleted. Two basalt samples from the oceanward slope of the peridotite ridge yielded an ^{39}Ar/^{40}Ar age of 100±5 Ma [*Malod et al.*, 1993], indicating that the emplacement of these basalts postdated the gabbro intrusion and the breakup which occurred at 114 Ma [*Boillot et al.*, 1988].

3.3.2. Transitional ocean floor sequences of the Liguria-Piemonte ocean in Grisons. Relatively complete ocean floor sequences of the Liguria-Piemonte ocean are preserved in the Platta nappe in Grisons. The weak Alpine metamorphic overprint and the strongly localized Alpine deformation led to the preservation of primary contacts between ultramafic and mafic rocks and postrift sediments.

The ultramafic rocks are serpentinized harzburgites and lherzolites. Pyroxenite layers occur within the serpentinites and are commonly subparallel to the locally preserved spinel foliation. In high-temperature mylonitic shear zones, pyroxene shows crystal plastic deformation indicative of temperatures above 700°C, whereas low-temperature mylonitic shear zones are composed of strongly foliated serpentinite (< 350°C) and are usually overprinted by hydrothermal alteration and late brittle deformation. Both types of shear zones are scarce, showing that deformation was localized. Fragments of serpentinite mylonites occur as clasts in cataclastically deformed serpentinite and document progressive deformation under decreasing temperatures. Replacement of serpentine minerals by calcite occurred under still lower temperatures.

Tectonosedimentary breccias, so-called ophicalcites, typically occur along the top of the serpentinites. They include serpentinite clasts embedded in a fine-grained, microsparitic, typically red-stained calcite matrix or white sparry calcite often preserving typical cement fabrics [*Bernoulli and Weissert*, 1985]. The fabric of these breccias varies considerably from serpentinite host rock with fractures filled by red limestone and/or white sparry calcite (Figure 6c), to clast-supported breccias with in situ fragmented serpentinite clasts (ophicalcites I of *Lemoine et al.* [1987]), to coarse, unsorted, matrix-supported breccias with fragments of serpentinite, gabbro, and continent-derived basement rocks and prerift sediments (ophicalcites II of *Lemoine et al.* [1987]). Clasts of basalts are conspicuously absent in these breccias. The amount of matrix within these tectonosedimentary breccias typically increases away from the hostrock and upsection. Geopetal infill of sediment into crevasses and pockets of the mantle rocks indicates that these rocks were exposed at the seafloor (Figure 6c) [*Bernoulli and Weissert*, 1985]. Ophicalcites I are very similar to tectonosedimentary breccias overlying exhumed serpentinites at Site 1070 (Iberia Abyssal Plain) [cf. *Whitmarsh et al.* 1998].

Gabbro bodies occur but are not very common. Part of them are isotropic and show intrusive contacts with the enclosing serpentinites, whereas others have been deformed under high-temperature conditions. Gabbros are also found as clasts in tectonosedimentary breccias and pillow breccias. Two gabbro samples from an intrusion in the Platta nappe were dated by U-Pb on zircons and yielded an age of 161±1 Ma [*Desmurs et al.*, 1999]. This age is almost identical with ages obtained from other gabbros from the Liguria-Piemonte ocean [*Bill et al.*, 1997; *Rubatto et al.*, 1998] and is close to the age of the oldest postrift sediments overlying the serpentinites (late Middle Jurassic) [*Baumgartner*, 1987].

Basaltic dikes cut across the serpentinites (Figure 6b), and the gabbros and basaltic flows overlie stratigraphically the tectonosedimentary breccias and serpentinized mantle rock (Figure 5), however, they do not occur as clasts in tectonosedimentary breccias. The basalts are obviously the youngest rocks within the ultramafic-mafic sequence, and their extrusion post-dated mantle exposure at the seafloor. Thickness and volume of the basalts increase oceanward over 10 - 20 km from zero to a few hundred of meters, whereas the volume of gabbros remains small (< 5%). Nd-isotope data from basalts of the Platta nappe show εNd values ranging from +7 to +9.9, indicating a depleted mantle source for these basalts [*Stille et al.*, 1989].

3.3.3. Similarities between the ocean floor sequences of the Galicia and Adriatic margins. The transitional ocean floor sequences of the two margins show a similar spatial distribution and analogous crosscutting and stratigraphic relationships between serpentinites, ophicalcites, gabbros, and basalts. This sequence is characterized by (1) the exhumation of mantle rocks from deep lithospheric levels to the ocean floor as documented by their deformation under decreasing temperatures indicated by serpentinization, the formation and subsequent cataclastic reworking of serpentinite mylonites, and low-temperature replacement by calcite, (2) the formation of tectonosedimentary breccias reworking and/or overlying serpentinites, (3) scarce magmatic activity before breakup, as indicated by the rare occurrence of gabbros with a prebreakup age, and (4) increasing amounts of basalt toward the ocean.

3.4. Postrift Sediments

Both the Galicia and Adriatic margins are sediment-starved with a discontinuous sedimentary cover of postrift sediments (0-4 km thick). Toward the ocean, the oldest postrift sediments become younger. Along the Galicia margin the oldest postrift sediments are of Hauterivian age and drape the Interior Basin, whereas on the Deep Galicia Margin they are of Albian age. Along the Adriatic margin a similar trend can be observed. In the east the oldest postrift sediments are of Sinemurian age, whereas in the distal margin the first postrift sediments are upper Middle to Upper Jurassic radiolarian cherts (Figures 5 and 6a). In general, the thickness of the postrift sedimentary sequence is strongly variable, and local facies variations or hiatuses occur.

3.5. Detachment Structures in the Distal Margins

3.5.1. The *S* reflector of the Galicia margin. The so-called *S* reflector is a characteristic feature of the nonvolcanic margins associated with the opening of the Iberian Atlantic and the Bay of Biscay. It represents a prominent reflection or a bundle of reflections and was first described in the distal continental margin of the Bay of Biscay by *de Charpal et al.* [1978]. Along the Galicia margin the *S* reflector is well imaged; it is a prominent reflection underlying the tilted fault blocks of the distal margin [*Boillot et al.*, 1980; *Mauffret and Montadert*, 1987]. Toward the peridotite ridge, the *S* reflector shows an upward-convex shape, and at places it appears to have emerged at the seafloor and to be directly overlain by synrift or postrift sediments [*Boillot et al.*, 1988]. Toward the continent, its geometry becomes more complex, and its continuation is unclear. *Reston et al.* [1995, 1996] thought the *S* reflector branches continentward into several reflections whereby the lowest reflection would lead to a breakaway in the east. *Boillot et al.* [1995a] and *Brun and Beslier* [1996] suggested that the *S* reflector belongs to a conjugated fault system with a brittle fault along the base of the tilted blocks showing a top-to-the-ocean sense of shear and a deeper fault zone showing a top-to-the-continent sense of movement.

In seismic profiles, the intensity of the *S* reflector is strongly variable, and its relation to the high-angle faults bounding the tilted blocks is not clear. *Reston et al.* [1995, 1996] were able to demonstrate, on the basis of constructed true depth profiles, that the *S* reflector represents a continuous, locally updomed structure. Thus the high-angle faults and the fault blocks of continental basement rocks and prerift sediments are clearly truncated along the *S* reflector. On the basis of its waveform, Reston et al. interpreted the *S* reflector as a reflection from a discrete interface, i.e., from a fault zone, rather than the result of gradually changing material properties (e.g., downward decreasing serpentinization of ultramafic mantle rocks).

Another controversy related to the *S* reflector concerned the type of material underlying it. *Sibuet* [1992] assumed lower crustal rocks, whereas *Boillot et al.* [1989b] suggested serpentinized peridotite. Since these lithologies are difficult to distinguish by their seismic velocities, drilling will be the only method to obtain a clear answer.

Although most authors agree at present that the *S* reflector images a rift-related detachment structure associated with mantle exhumation, the dynamic significance of this structure is still a matter of debate (see the different models in Figure 3 in the work of *Reston et al.* [1996] and our discussion in section 4). A major reason for the different interpretations is the ambiguity of the kinematic data. The few data available (see Table 1, p. 335, in the work of *Beslier et al.* [1990]) show top-to-the-NE, top-to-the-NW, top-to-the-SE, and top-to-the-west senses of shear, all of them recording deformation under different metamorphic conditions, i.e., at different crustal levels.

3.5.2. The low-angle detachment system in the Err and Platta nappes. In the distal Austroalpine margin, exhumation of the mantle and final emplacement of the tilted fault blocks are connected with the evolution of a low-angle detachment system. The geometry of this detachment system is spectacularly exposed and preserved in the area of Piz d'Err-Piz Bial in the Err nappe (Figures 5 and 6e) [*Froitzheim and Eberli*, 1990]). Except for a few gaps due to Quaternary erosion, the detachment system is exposed over 18 km, parallel to the E-W transport direction. Its hanging wall is formed by fault blocks of continental basement rocks and prerift sediments, tilted toward the east, i.e., toward the continent along west dipping high-angle faults. The fault blocks are of variable size (100 m to a few kilometers across) and become generally smaller toward the ocean. Within the study area the high-angle faults and associated tilted fault blocks are systematically cut by the low-angle detachment faults. An incisement structure is preserved at Piz Jenatsch where a higher detachment fault is cut by a younger lower one (Figure 5). For one of the fault planes a displacement of more than 10 km can be determined [*Manatschal and Nievergelt*, 1997]; however, no mylonites were found to be associated with the detachment system along the whole length of outcrop. The fault rocks associated with the detachment system consist of up to 50 m thick green cataclasites which gradually pass downward into massive, undeformed post-Variscan granite. The detachment planes themselves are sharp and well-defined horizons marked by a characteristic black fault gouge, which accommodated most of the displacement [*Manatschal*, 1999]. Shear sense criteria from fault rocks indicate a top-to-the-west, i.e., a top-to-the-ocean, sense of shear.

Relics of the same fault system, marked by the same characteristic black gouge, can be traced from the Err nappe into the Platta nappe [*Manatschal and Nievergelt*, 1997]. Here the black gouge occurs at the base or within blocks of continental basement and prerift sediments emplaced onto the serpentinized mantle peridotites (Figures 5 and 6d) or as clasts in tectono-sedimentary breccias; however, also here no mylonites were observed. Postrift sediments overlie the allochthonous fault blocks ("extensional allochthons") and the exhumed mantle rocks.

From our observations it becomes clear that the detachment faults are late, shallow crustal structures with a top-to-the-ocean sense of shear. They formed breakaways in the continental crust to the east and penetrated oceanward, i.e., to the west, into deeper crustal levels (Figure 7). This is compatible with the observed top-to-the-ocean sense of shear. The footwall rocks are upper crustal granites and gneisses in the east and serpentinized mantle rocks in the west. Prerift lower crustal rocks were not observed in the footwall.

3.5.3. Similarities between the *S* reflector and the low-angle detachment system. The dynamic interpretation of the *S* reflector and its comparison with the low-angle detachment faults of the Austroalpine margin is hampered by the lack of kinematic data; however, the geometrical similarities of the two sets of structures go, in our opinion, beyond chance. It is undisputed that both kinds of structure (1) are related to rifting, (2) occur and were active at a shallow crustal level, (3) can be traced toward exhumed mantle, and (4) are overlain by continentward rotated tilted fault blocks.

Reston et al. [1996] interpreted the *S* reflector as a continuous fault zone truncating older high-angle faults and associated tilted fault blocks and forming incisement structures. These authors implied a top-to-the-ocean sense of shear on the basis of the continentward tilting of the hanging wall blocks and the inferred occurrence of breakaway structures to the east. Furthermore, they suggested that the *S* reflector forms the base of the synrift sediments at the continent-ocean transition. All these inferences fit exactly with our observations in the Err and Platta nappes. Therefore we think that the detachment system in the Err-Platta nappes and the *S* reflector in the Iberian margin are analogous structures.

The overall observations made along the Err-Platta detachment system fit extremely well with the results of ODP Leg 173 along the Iberia Abyssal Plain (Figures 2a and 2d) [*Whitmarsh et al.*, 1998]. Along this more southern transect, deep-sea drilling penetrated a detachment surface separating exhumed mantle rocks below from tectono-sedimentary breccias above (Site 1068). Further oceanward, an isolated continental block was drilled (Site 1069). This block is soled by a reflection which can be traced eastward toward the exhumed tectonized mantle at Site 1068. This isolated block of continental material therefore represents an extensional allochthon comparable with the analogous "klippen" of the Platta nappe (Figure 7).

4. Discussion

In the following we discuss the temporal and spatial evolution of rifting, in particular the kinematics and the polarity of crustal-scale detachment faulting, and the processes controlling the observed shift of rifting from a wide zone of extension to a localized area in the future distal margin. We also discuss the processes controlling the transition from rifting to seafloor spreading and propose that the thermal evolution and associated changes in the rheology of the lithosphere control the evolution of rifting.

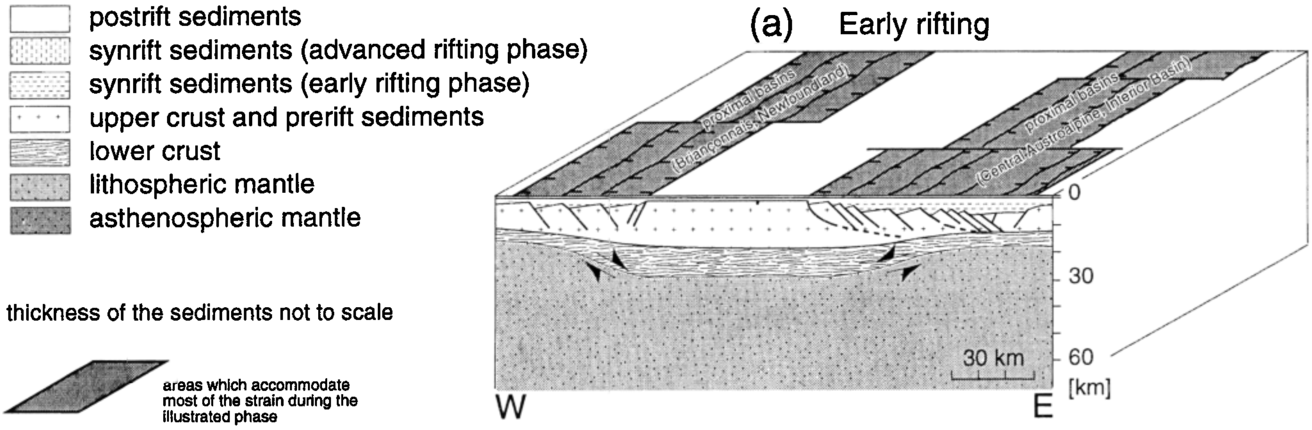
4.1. Early Rifting

In the Tethyan area, basins evolving during the early phase of rifting were distributed over a wide zone which was to become the proximal parts of the future margins (Figure 9a). The basins were bounded by high-angle faults, which had a listric geometry and which can be traced into basement (Figure 7) [e.g., *Bertotti*, 1991]. During this phase, synrift sediments in the hanging wall basins and on the footwall shoulders were marine, and in general, a stratigraphically reduced but deepening-upward sequence from carbonate platform to lithohermal limestones and pelagic deposits is observed on the submarine highs of the footwall blocks [*Bernoulli et al.*, 1990]. No basement rocks were reworked during this phase. This documents that the entire future margin was subsiding. Although the individual basins are half grabens, east and west dipping master faults occur, and the basins are symmetrically arranged along preexisting zones of weakness, showing an overall symmetric structure of the margin (Figure 9a). Pure shear-dominated symmetric rifting on a lithospheric scale during this initial phase is consistent with the subsidence history and the reconstructed stretching factors. For the proximal Adriatic margins, β values of about 1.5 have been estimated by *Froitzheim* [1988] and *Bertotti et al.* [1993]. On the basis of the seismic interpretation of *Murillas et al.* [1990], reproduced in our Figure 2b, a β value of 1.4 can be determined for the Interior and Porto Basins. Thinning of the crust associated with the uplift of mantle material and simultaneous cooling of the crust during this initial rift phase are supported in the Adriatic margin by the evolution of different isotope systems [e.g., *Handy and Zingg*, 1991], by fission track data [*Sanders et al.*, 1996] and by the P-T-t path of a Permian crust-mantle boundary (Figure 8) [*Hermann et al.*, 1997; *Müntener et al.*, 1999].

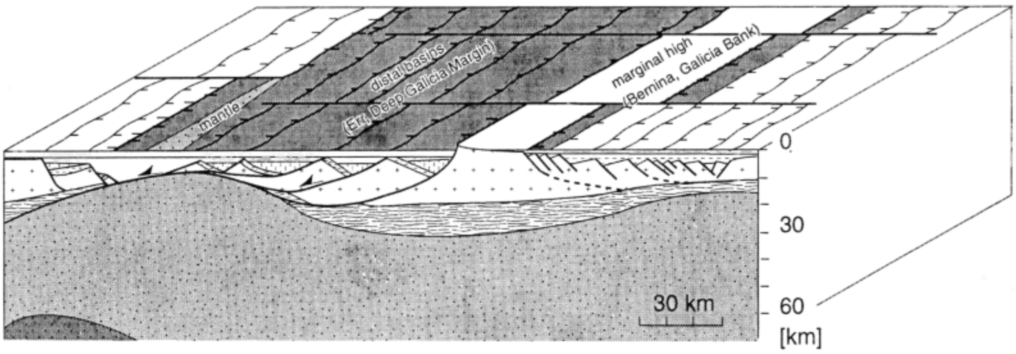
4.2. Advanced Rifting

About 20 - 30 Myr after the initiation of rifting, the rifting site was shifted to the previously not or only weakly extended areas that became the distal margin (i.e., to the Err domain and the Deep Galicia Margin) whereas the proximal basins to the east were draped by postrift sediments. The new basins in the distal margin were distinctly smaller and bounded by west dipping high-angle normal faults (Figures 7 and 9b). The tilted blocks of continental basement and prerift sediments separated by the high-angle faults are underlain by an oceanward dipping low-angle detachment system along which subcontinental mantle rocks eventually were exhumed and exposed to the seafloor (Figures 7 and 9b). Where detachment structures are exposed, like in the Err and Platta nappes, the shear sense indicators along them show a top-to-the-ocean sense of shear. This sense of shear is consistent with the continentward tilting of the hanging wall blocks, with the observation that the detachment system forms a breakaway in the east and cuts downward into mantle rocks toward the west, and finally with the occurrence of continent-derived extensional allochthons overlying the mantle rocks (Figure 5). In the Deep Galicia Margin the shear sense along the inferred detachment fault underlying the continental fault blocks (*S* reflector) is not known; however, a top-to-the-ocean sense of shear is consistent with the overall geometry observed [*Reston et al.*, 1995, 1996].

In contrast to the initial phase of rifting, during which the entire margin subsided, the subsidence/uplift pattern during the advanced stage of rifting was more complex. Subsidence in the distal margins was contemporaneous with uplift and subaerial exposure of small domains continentward of the distal margins which became the local source area of siliciclastic sediments with reworked basement clasts (Galicia Bank [*Winterer et al.*, 1988]; and Bernina domain [*Finger et al.*, 1982]). This uplift is possibly related to the breakaway along the oceanward dipping detachment faults leading to an isostatic "edge effect" caused by the removal of the hanging wall block. The uplift and the unconformity associated with it



(b) Advanced rifting



(c) Transition to seafloor spreading

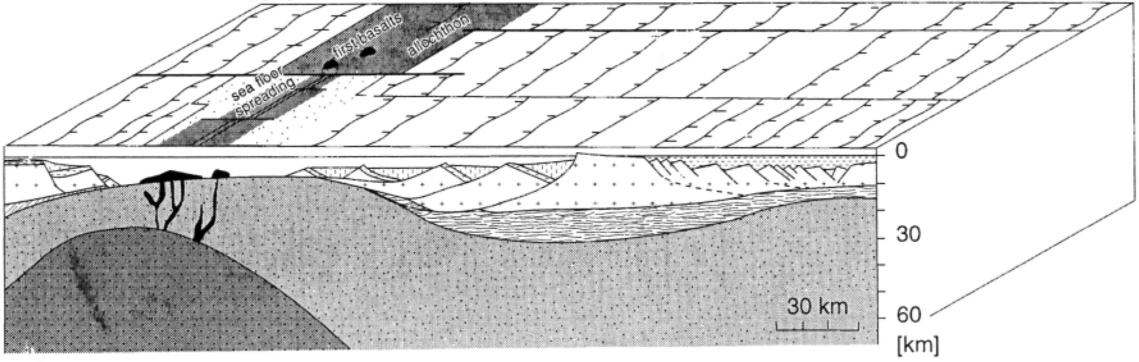


Figure 9. Model presenting the temporal and spatial evolution from (a) initial rifting to (b) advanced rifting to (c) final seafloor spreading.

and the small amount of internal extension allow one to distinguish these highs as characteristic segments of the margins, separating their proximal and distal parts. Along the Galicia Margin the high isolated the distal margin from the continent, which may explain why the distal margin became sediment-starved; in the Tethyan realm a much wider area of the Adria microplate was submerged during Liassic rifting and subsided to subphotic depth [*Bernoulli and Jenkyns*, 1974].

4.3. Mantle Exhumation

The p-T-t data of mantle rocks from the Adriatic and the Galicia margins document a similar exhumation history. In the south Pennine-Austroalpine Malenco complex, a large volume of gabbroic rocks intruded along the crust-mantle boundary during Permian time [*Hermann et al.*, 1997], documenting that the mantle rocks already occupied a shallow position in the subcontinental mantle lithosphere at that time [*Müntener et al.*, 1999]. After this the mantle and lower crustal rocks cooled more or less isobarically, until decompression accompanied by increased cooling occurred during initial rifting in latest Triassic to Early Jurassic times (Figure 8). Cooling of lower crustal rocks below 300°C is documented for this period in the southern Alps [*Handy and Zingg*, 1991] and in the Malenco complex [*Muntener et al.*, 1999].

In the south Pennine Platta nappe the widespread preservation of a high-temperature spinel foliation in many of the mantle rocks finally exposed on the seafloor shows that deformation during uplift was not pervasive but localized along fault zones. Decreasing temperatures during exhumation are documented by (1) serpentinization of mantle peridotites and their subsequent mylonitization, (2) brecciation of the former serpentine mylonites leading to serpentinite breccias with a serpentine matrix, and (3) low-temperature replacement of serpentine minerals by calcite. All these transformations predate final exposure at the seafloor documented by the different types of tectonosedimentary breccias ("ophicalcites") described above. The occurrence of clasts of continental basement rocks, of Triassic prerift sediments, and of fault rocks typically associated with the low-angle detachment faults in these breccias shows that mantle exhumation was closely associated with the emplacement of the extensional allochthons [*Froitzheim and Manatschal*, 1996].

On the Deep Galicia Margin, breccias resembling ophicalcites were interpreted as extensional mélanges following the trace of a brittle detachment fault [*Boillot et al.*, 1995b]. To the south, along the Iberian Abyssal Plain, breccias with clasts of continental basement rocks overlying exhumed mantle rocks were drilled at Site 1068, ODP Leg 173 [*Whitmarsh et al.*, 1998]. Seaward of this window of exhumed subcontinental mantle, continental basement rocks were drilled again at Site 1069. These continental rocks are underlain by a prominent reflector which we interpret as the trace of a low-angle detachment fault underlying an extensional allochthon, forming an isolated klippe on the exhumed mantle rocks (Figure 7). Like *Reston et al.* [1995, 1996], we interpret the *S* reflector as an oceanward dipping crustal-scale detachment fault, forming a breakaway in the continental crust to the east and cutting across mantle rocks, exhuming and exposing them on the seafloor. In our interpretation the *S* reflector is thus analogous to the detachment fault system in the south Pennine-Austroalpine margin.

4.4. Magmatic Evolution and Transition to Seafloor Spreading

The large volumes of gabbroic rocks occurring in Alpine sections along the mantle-crust boundary are associated with magmatic underplating in Permian times [*Quick et al.*, 1994; *Hermann et al.*, 1997] and clearly not related to Mesozoic rifting. During late rifting and mantle exhumation the production of melt was apparently very limited along the continent-ocean transition zone. In this transitional part of the margin the crust is formed mainly by serpentinized mantle rocks containing only small portions of magmatic rocks. Apparently, a diffuse magmatic activity penetrated and overprinted the exhumed subcontinental mantle during the rise of the asthenosphere before seafloor spreading initiated along a mid-ocean ridge. Along the Deep Galicia Margin the intrusion of rift-related gabbros (121 Ma) [*Schärer et al.*, 1995] predates the formation of true oceanic crust to the west (114 Ma [*Boillot et al.*, 1995a]) by several millions of years.

In the Platta nappe, basalts with a mid-ocean ridge (MOR) signature locally overlie the serpentinites adjacent to the thinned continental crust. Their volume increases oceanward, but the amount of gabbro remains small, suggesting a gradational transition to a true oceanic crust. As along the Iberian margins the intrusion of gabbros into the Platta serpentinites predates the extrusion of basaltic pillow lavas and flows. For the pillow lavas no geochronological data are available, but their relative age with respect to the gabbros (161±1 Ma [*Desmurs et al.*, 1999]) becomes clear from stratigraphic and crosscutting relationships. The gabbros intruded the already serpentinized mantle rocks (L. Desmurs, personal communication, 1998) and were locally deformed under high-temperature conditions. In contrast to the basalts, the gabbros occur also as clasts in tectonosedimentary breccias and were exposed on the seafloor, where they were covered by pillow lavas and breccias. Locally, undeformed basaltic dikes cut across gabbros deformed under high temperatures. A combination of the geochronological data from the Deep Galicia Margin and of the stratigraphic relationships in the Platta nappe suggests that along both margins the gabbros were emplaced during the transition from rifting to the formation of new oceanic crust. The similar age of all gabbros determined so far from south Pennine ophiolites in the Alps (165-161 Ma [*Bill et al.*, 1997; *Rubatto et al.*, 1998]) further suggests that in the Alps the oldest parts of the ocean were preferentially preserved and that if a slow-spreading ridge eventually developed, it was largely subducted. In any case, a classical oceanic crust with a sheeted dike complex and a substantial gabbroic layer has not been observed in the Alps, and many of the preserved Alpine ophiolites may therefore represent transitional rather than "true" oceanic crust accreted along a spreading ridge.

4.5. Shallow and Deep Detachment Structures

Whereas the shallow structures of the Galicia margin and of the south Pennine-Austroalpine margins can be reconstructed with some confidence, the deep structure of the margins and their kinematic evolution is still a matter of controversy. Indeed, many different models which are incompatible with each other exist for the polarity and kinematics of the low-angle detachment faults of the continent-ocean transition (for Galicia see Figure 3 in the work of *Reston et al.* [1996]; for the Adriatic margins, compare, e.g., the work of *Lemoine et al.* [1987], *Trommsdorff et al.* [1993], and *Frotzheim and Manatschal* [1996]).

Where, like in the Err-Platta nappes, shallow detachment structures are exposed, the shear sense indicators show a top-to-the-ocean sense of shear. This sense of shear is compatible with the observation of continentward tilting of the hanging wall blocks, the occurrence of normal faults forming a sequence of breakaways in the east, and of continent-derived extensional allochthons overlying the detachment which oceanward cuts down into mantle rocks. However, this sense of shear records only late deformation occurring at shallow crustal levels. In the Alps, kinematic data recording deformation deeper in the crust (>10 km) associated with mantle exhumation are not available either because the deep crustal levels are not exposed (Err, Bernina, and other Austroalpine nappes) or because the structures were strongly overprinted by Alpine deformation and metamorphism and their interpretation is therefore ambiguous (Malenco complex).

For the Deep Galicia Margin most of the authors accept a late top-to-the-ocean sense of shear for the detachment fault(s) underlying the tilted blocks at a shallow crustal level. This top-to-the-ocean sense of shear is in line with kinematic indicators observed in chloritized and strongly altered gabbroic schists which yielded a protolith age of 122 Ma [e.g., *Boillot et al.*, 1995a] and the continentward tilting of the fault blocks overlying the *S* reflector. However, locally, a top-to-the-continent, i.e., top-to-the-east, sense of shear was determined in peridotite mylonites [*Beslier et al.*, 1990]. These mylonites were formed at 1000° - 850°C and are therefore older than 122 Ma, the time when the $^{39}Ar/^{40}Ar$ system in amphibole crossed its blocking temperature, which is about 550°C [*Féraud et al.*, 1988]. The movements associated with the two opposite shear sense indicators are therefore of different age.

In an analog experiment, *Brun and Beslier* [1996] produced a conjugate set of faults with opposite shear senses as considered by some authors for the Deep Galicia Margin. In their experiment they obtained a top-to-the-ocean movement in the upper crust and a conjugate top-to-the-continent one in the lower crust. On the basis of the analogy between their experiment and the shear sense distribution in the Galicia margin, they concluded that mantle exhumation was controlled by boudinage of the lithosphere. However, in contrast to the analog model, the shear senses established along the Deep Galicia Margin are not of the same age and therefore are not necessarily part of the same kinematic system. Moreover, in the analog experiment the change during exhumation of the rheological properties of, for example, the lower crust, cannot be taken into account during the experiment, which leads to unrealistic conditions during the final stages of the experiment. In fact, cooling of the lower crust during early rifting is documented along the Adriatic margin (Figure 8).

A model including an east, i.e., continentward, dipping detachment has been proposed for the Adriatic margin by *Trommsdorff et al.* [1993]; this model was later modified by *Hermann and Müntener* [1996]. These authors assumed that a first, top-to-the-east directed detachment fault system became inactive and further extension was accommodated by a younger, west dipping one. This model is mainly based on the observation that a crust-to-mantle boundary, i.e., part of the former Adriatic Moho, was exposed along the Adriatic margin. This observation may, in our opinion, be explained in a much simpler and coherent way by only one west dipping detachment fault system (see Figure 4b). The most important argument against the model of Hermann and Müntener is its incompatibility with the overall geometrical and uplift subsidence patterns observed along the asymmetrical Tethyan margins [see *Manatschal and Bernoulli*, 1999, Figure 3). Moreover, extension along one single low-angle detachment fault cutting across the entire crust during early rifting (1) is in conflict with the observation that early faults are listric and sole out at midcrustal levels (Figure 7) and (2) is also not realistic from a mechanical point of view as long as the lower crust was hot and behaved in a ductile way (see below).

An indirect way to determine the polarity of detachment systems along nonvolcanic margins is to consider the asymmetry of pairs of margins. *Lister et al.* [1986, 1991] showed that depending on the polarity of the detachment system, an upper plate margin can be distinguished from a lower plate margin on the basis of their architecture and subsidence/uplift patterns. Lower plate margins are characterized by basement rocks exhumed from deeper crustal levels and overlain by highly faulted remnants of the upper plate, the so-called extensional allochthons. These rotated blocks and half grabens of the upper plate, left behind on the lower plate, are bounded by oceanward dipping high-angle faults. The faults are truncated along their base by one or several low-angle detachment faults with a complex geometry. The uplift/subsidence pattern of lower plate margins is characterized by a small and localized shoulder uplift continentward of the subsiding basins of the distal margin (Figure 9b). In contrast, the upper plate margins tend to be relatively narrow and structurally simpler and show pronounced regional uplift during late rifting. Using these criteria, the Adriatic and Galicia margins represent typical lower plate margins characterized by low-angle detachment faults at a shallow crustal level, the widespread occurrence of tilted blocks, the exhumation of serpentinized mantle rocks at the seafloor, and subsidence of the distal basins bounded by a small rift shoulder. Likewise, the Briançonnais and Newfoundland margins are interpreted as upper plate margins [*Manatschal and Bernoulli*, 1999, and references therein].

A lower plate position of the Galicia and the Adriatic margins implies a west dipping detachment system with a top-to-the-ocean sense of shear as proposed by *Reston et al.* [1996] for Galicia and by *Froitzheim and Manatschal* [1996] for the south Pennine-Austroalpine margin. These authors suggested that over the entire lithosphere, extension was accommodated by a detachment system with a top-to-the-west sense of movement resulting in a strong asymmetry of the pairs of margins [*Manatschal and Bernoulli*, 1999]. However, the (low) continentward dip of the *S* reflector seems to be at odds with a top-to-the-ocean sense of shear during extension and exhumation of mantle rocks; indeed, a top-to-the-ocean sense of shear seems to imply that the extensional allochthons were moved uphill during extension. To explain this apparent paradox, a comparison with the evolution of metamorphic core complexes might be helpful; these complexes show many geometrical similarities with the extensional systems of our continental margins [*Davis and Lister*, 1988], although their tectonic setting is very different. Metamorphic core complexes show a progressive warping of the initial detachment system and the inactivation of older breakaways during this process. To generate this type of structural geometry, a strong vertical force driven by the buoyancy of the footwall rocks must lead to the upwarping of the originally planar or listric detachment surface. Upwarping of the detachment surface leads to a geometry mechanically not suitable for further movement along the original fault, favoring the stepwise initiation of new faults and an oceanward shift of the breakaways (compare, Figure 11 in the work of *Lister and Davis* [1989]). The oceanward migration of detachment faulting, together with simultaneous uplift of the footwall rocks, can explain many of the complex geometries observed in seismics profiles or in the field, such as incisement structures, tilted blocks truncated by detachment faults, the oceanward decreasing size of fault blocks, and the extensional allochthons of continental crust overlying exhumed mantle rocks (Figure 4b). We also suggest that with the evolution of new faults, the lower crustal rocks of the continent-ocean transition zone were covered and hidden at depth by later emplaced blocks of upper crustal rocks.

Isostatic uplift of mantle rocks may be related to the removal of overlying crustal material during extension, to serpentinization of mantle rocks leading to a strong increase in volume and consequently to lower densities, or to the uplift of a hot, less dense asthenospheric mantle. Although all three processes appear to be involved during the latest stages of rifting, the rates of extension and the driving forces are not yet determined.

The occurrence of the high-temperature, top-to-the-continent directed shear zones in the mantle rocks of the Deep Galicia Margin is not explained by our model. A possible explanation for these shear zones and their direction of movement could be that they formed during updoming of the mantle during initial stages of symmetrical rifting (Figure 9a).

4.6. Extensional Decoupling During Rifting: The Role of the Lower Crust

One of the most enigmatic features of the Galicia and Adriatic margins is that the site of final continental breakup is not located within the zone of initial rifting (Figure 9). Riftactivity shifted from the future proximal margins and localized in the previously not or only weakly extended future distal margins. Associated with this shift of rifting is a change in the tectonic style from symmetric pure shear to asymmetric and localized simple shear extension. Most of the models used to explain the formation of rifted margins do not take this change into account, although it appears to be a common feature along nonvolcanic margins.

The shift of the site of rifting shows that the response of the lithosphere to extension changed through time. We think that these changes were strongly controlled by the thermal state of the lithosphere during extension, a view also supported by numerical modeling [*Hopper and Buck*, 1996]. The thermal history of the Adriatic margins clearly shows that cooling of the lithosphere started immediately after the onset of rifting. This is indicated by the p-T path of lower crustal rocks and their cooling ages (Figure 8) and the cooling history of upper crustal rocks below the Lombardian basin [*Sanders et al.*, 1996]. Fault zones developing during this initial stage of rifting sole out at middle- to lower crustal levels (see Figure 7b), and reflection seismic profiles across early rift basins do not show a displacement of the Moho (see, e.g., Jeanne d'Arc basin of the Newfoundland margin) [*Keen et al.*, 1987]). Obviously, the lower crust was weak, and extension in the upper crust decoupled from pure shear extension in the lower crust and in the upper mantle.

Cooling of the lithosphere changed its rheological properties and led to strengthening, which eventually forced rifting to shift into previously not or weakly extended areas. Also, the geometry of the fault zones changed: Fault systems developing during late stages of rifting in the distal margin are low-angle detachment faults which cut across the lower crust and into mantle rocks (Figures 5 and 7). During this stage, extension was no longer decoupled between the upper and lower crust, and the faults could exhume mantle rocks directly to the seafloor. Whether the low-angle detachment faults cut across the entire lithosphere, as suggested by most of the simple shear models, or merged with a network of ductile shear zones in the upper mantle is beyond observation. Finally, ongoing extension together with heat advected by melts derived from the rising asthenosphere might have eliminated the remaining yield strength of the subcontinental lithosphere, resulting in final breakup and seafloor spreading.

5. Conclusions

Although the Galicia and Adriatic margins are of different age and ultimately had a different fate, the spatial and temporal evolution of rifting and the duration of its different stages are very similar. The similar tectonic evolution resulted in a similar architecture of the margins with different segments showing specific basin geometries and a characteristic sedimentary evolution. The tectonic evolution is characterized by a shift of the site of rifting from a broad zone of deep-seated pure shear extension below the future proximal margins to a localized area in the future distal margin, which previously had undergone distinctly lesser extension. During this later phase, extension was accommodated by low-angle detachment faults with a top-to-the-ocean sense of movement that is in line with a lower plate position of the Galicia and the Adriatic margins, likewise compatible with the subsidence-uplift pattern observed along the margins and their conjugate counterparts (Newfoundland and Briançonnais). Seafloor spreading initiated only after exhumation of the subcontinental mantle to the ocean floor requires, on a lithosphereic scale, a mechanism of simple shear extension in the lower crust and uppermost mantle. During rifting, the style of large-scale deformation thus changed from symmetric and homogeneous (pure shear) to asymmetric and localized (simple shear).

An important factor controlling the evolution of the margin appears to be the thermal state of the lithosphere. Cooling and strengthening of the lower crust may control localization of deformation and the changing style of deformation. Therefore models in which the rheological properties do not change as a function of cooling may not be appropriate for the study of processes leading to the formation of passive continental margins.

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