1. LEG 173 INTRODUCTION[1]

Shipboard Scientific Party[2]

BACKGROUND

The rifting and breakup of continental lithosphere to form passive or rifted continental margins is one of the most fundamental geological processes on earth. It has many physical similarities with the formation of intraplate sedimentary basins and relates directly to the onset of seafloor spreading, a process that generates the mid-ocean ridges and the crust underlying the ocean basins. Rifted margins exhibit a wide spectrum of characteristics, apparently greater than is seen in continental rifts, probably in response to the large magnitude of extension and different combinations of asthenospheric temperature, lithospheric rheology, strain rate, and stress. The rifting process, through the indirect effects of concurrent greater sedimentation and heat flow and sub-aerial volcanism can also have important environmental and resource implications. Because of the thick to very thick (2 to 15 km) sediments found on many rifted margins and the consequent inaccessibility of basement rocks to scientific drilling, however, rifted margins had, until 1993, largely been studied by remote geophysical means or by the post- and synrift history recorded in the overlying sediments at a few "sediment-starved" margins. In the Northeast Atlantic before 1993, for example, nonvolcanic rifted margins were drilled off Northern Biscay (Montadert et al., 1979b), Vigo Seamount (Sibuet et al., 1979), the Mazagan Escarpment (Hinz et al., 1984), Goban Spur (de Graciansky et al., 1985), and Galicia Bank (Boillot, Winterer, et al., 1988) where crystalline basement was cored for the first time on the west Iberia margin.

Quantitative models have been developed for the postrift isostatic and thermal subsidence of margins and basins (McKenzie, 1978) and, more recently, for the critical effect of asthenospheric temperature at the time of lithospheric breakup on the quantity and composition of melt that may be underplated at the base of the crust and/or erupted as lavas at the surface (Bown and White, 1995; White and McKenzie, 1989). Finite element models have also been developed to model the tectonic evolution of the crust and lithosphere under extensional margins (e.g., Bassi, 1991, 1995; Harry and Sawyer, 1992). Qualitative conceptual models also have been proposed based on geological investigations in extensional terrains such as the Basin and Range province (Wernicke, 1985; Wernicke and Burchfiel, 1982), which have been extrapolated to rifted continental margins (Boillot et al., 1987a, 1988b; Le Pichon and Barbier, 1987; Lister et al., 1986; Reston et al., 1995), or on fossil passive margins trapped in orogens, such as the Ligurian Tethys margins in the Alps (e.g., Froitzheim and Manatschal, 1996; Lemoine et al., 1987). Further, small-scale analogical models have been employed to explore the mechanical behavior of the lithosphere during extension (e.g., Allemand and Brun, 1991; Beslier and Brun, 1991; Brun et al., 1994; Brun and Beslier, 1996). Nevertheless, in spite of the predictive capability of all these models, some uncertainty remains both about the geological processes involved in the crust and mantle and in particular about the nature and location of features such as the ocean/continent transition (OCT).

The North Atlantic Rifted Margins Detailed Planning Group (NARM DPG) was convened by the then Planning Committee of the Ocean Drilling Program (ODP) and met in 1991 to plan a program of drilling to study the problems of the formation and evolution of rifted margins. The Group identified two important classes of rifted margins to be studied: margins in which magmatism dominated the rifting process (volcanic margins) and margins in which magmatism seems to have played a minor role in the rifting process (nonvolcanic margins). The DPG recommended that ODP focus on a transect of holes across each class of margin and that each transect include a conjugate pair of margins. The criteria for selecting the locations of the two transects included (1) the existence of high-quality geophysical data on both conjugate margins, (2) the presence of a relatively thin sediment cover on the conjugate margins so that drilling to basement is possible using the *JOIDES Resolution,* (3) the absence of salt deposits, which could interfere with drilling, and (4) the absence of postrift volcanism, which could have modified the divergent margin.

Non-volcanic margins in particular provide opportunities to investigate and understand the tectonic aspects of rifting for two reasons. First, extensional tectonics, often expressed as normal faults and shear zones, that are seen to penetrate deep into the crust and uppermost mantle on seismic reflection profiles, allows rocks from deeper lithospheric levels to be exposed at the top of acoustic basement, as was demonstrated by Legs 103 and 149. Second, voluminous intrusives/extrusives, which can obscure crustal tectonics, are limited and commonly appear to be absent. Leg 149 (Sawyer, Whitmarsh, Klaus, et al., 1994; Whitmarsh, Sawyer, Klaus, Masson, 1996) and Leg 173 represented part of the program proposed by the DPG for the study of nonvolcanic margins. As originally envisaged, the total program, requiring four two-month legs, included drilling multiple sites in both the southern Iberia Abyssal Plain and the conjugate Newfoundland Basin, and one site on the Galicia Bank margin. Drilling on each of the margins was to include sites that would allow sampling of significant sections of basement with minimum sediment penetration and sites that would sample thicker and stratigraphically more complete sequences of synrift and postrift sediment. Leg 149, followed by Leg 173, drilled sites of the first type off Iberia; a deep hole in the Iberia Abyssal Plain sediments has yet to be planned to complement the Leg 149 and Leg 173 results. Conjugate rifted margins often exhibit some asymmetry in structural style that may relate to the mode of lithospheric rifting (e.g., pure and/or simple shear), or simply to the location of the original break in the continental crust. Although the original program was also designed to allow assessment of the degree of symmetry in the structure and evolution of the conjugate margin, this has not been possible so far. Characterization of basement type (thinned continental crust, transitional basement, or oceanic crust) within a wide zone of the Newfoundland Basin and Iberia Abyssal Plain, the location of the OCT, and the nature and evolution of the corresponding basement rocks on the two margins still remain important scientific objectives. Sites designed to sample synrift sequences will constrain the timing of rifting and breakup, the rift environment, and possibly significant anomalous elevation and/or subsidence asymmetries that are strongly indicated by some seismic data. The subsidence histories of the conjugate margins will help to determine the relative importance of pure and simple shear mechanisms of extension on a lithospheric scale.

THE WEST IBERIA CONTINENTAL MARGIN

The western continental margin of Iberia extends from Cape Finisterre in the north to Cape Saint Vincent in the south (Fig. 1). The

[1]Whitmarsh, R.B., Beslier, M.-O., Wallace, P.J., et al., 1998. *Proc. ODP, Init. Repts.,* 173: College Station, TX (Ocean Drilling Program).
[2]Shipboard Scientific Party is given in the list preceding the Table of Contents.

continental margin has a straight, relatively narrow shelf and a steep continental slope. South of 40°N, the slope is cut by numerous canyons. This simple picture is complicated by several offshore bathymetric features. The largest feature is Galicia Bank, a 200 × 150-km area within which the seafloor shoals to about 600 m water depth. Galicia Bank is characterized by two isolated seamounts on its southern edge (Vasco da Gama and Vigo) and is separated from northwestern Iberia by a broad submarine valley. At 39°N, the Estremadura Spur extends east-west over 100 km offshore and forms a barrier between the Iberia and Tagus Abyssal Plains. Finally, the east-northeast–trending Gorringe Bank forms the southern boundary of the Tagus Abyssal Plain and marks the surface expression of the seismically active Eurasia/Africa plate boundary.

Like many rifted or passive margins, the Iberia margin had a long history of rifting before the separation of Iberia from the Grand Banks of North America (see the review by Pinheiro et al., 1996). Welsink et al. (1989) describe the rift history of the conjugate Newfoundland margin.

Broadly, three main Mesozoic rifting episodes affected the west Iberia margin although all three episodes cannot be convincingly demonstrated together in any one part of the margin. These episodes are variously recorded in the deposits of the onshore Lusitanian Basin, which is probably continuous with the Interior Basin that separates Galicia Bank from northwestern Iberia (Murillas et al., 1990; Wilson et al., 1989), and offshore. Evidence for a Triassic to Early Jurassic (Liassic) continental rifting phase that created graben and half-graben structures in which evaporites were deposited is widespread on North Atlantic margins. The second rifting phase, which consisted of extension in the early Late Jurassic, is indicated solely on onshore and shelf seismic profiles. The last phase of extension occurred in the Early Cretaceous, coincided with the south-to-north breakup of Iberia from the Grand Banks and has been well documented with offshore geological and geophysical data (Boillot, Winterer, et al., 1988, 1989; Pinheiro et al., 1992; Whitmarsh et al., 1990; Wilson et al. 1996). The precise dating and duration of this last episode is arguable (Wilson et al., 1996); in fact, different parts of the margin probably rifted at different times as rifting propagated from south to north.

The rifting phases were accompanied by only minor volcanism (dikes and flows) within Iberia. Two phases of pre-breakup volcanism have been recognized by Ribeiro et al. (1979) and Martins (1991). A tholeiitic phase lasted from 190 to 160 Ma, coeval with Late Jurassic rifting, and a second phase occurred from 135 to 130 Ma in the Lusitanian Basin. This volcanism was relatively minor, and the west Iberia margin has the characteristics of a nonvolcanic margin. For example, tilted fault blocks and half-grabens are clearly observed off Galicia Bank (Mauffret and Montadert, 1987), and in the southern part of the Lusitanian Basin (Wilson et al., 1989), and there is no evidence on any part of the margin of seaward-dipping reflectors or of substantial subcrustal underplating. Nevertheless, Leg 149 penetrated 56 m of gabbro that underwent deformation that ended by 136.4 Ma at Site 900; the gabbro probably represents material emplaced at the base of the crust during rifting (Féraud et al., 1996).

Parts of the west Iberia margin underwent two additional phases of deformation in the Eocene and the Miocene. The Eocene deformation was caused by the Pyrenean orogeny and the abortive subduction of the Bay of Biscay beneath the north Spanish margin; it has been proposed that the main expression of this deformation was the uplift of Galicia Bank and adjacent seamounts (Boillot et al., 1979). The Miocene deformation accompanied tectonism in the Rif-Betic mountains and led to the gentle folding of sediments in the Iberia and northern Tagus Abyssal Plains, as seen on reflection profiles (Masson et al., 1994; Mauffret et al., 1989), and folding, faulting, and inversion of a large part of the Lusitanian Basin (Wilson et al., 1989).

Several plate-tectonic reconstructions have attempted to show the original positions of North America, Iberia, and Europe (Klitgord and Schouten, 1986; Le Pichon et al., 1977; Malod and Mauffet, 1990; Masson and Miles, 1984; Olivet, 1996; Olivet et al., 1984; Sibuet and Srivastava, 1994; Srivastava et al., 1990b; Srivastava and Verhoef, 1992; Srivastava et al., 1988). The along-strike relative positions of the North America and Europe plates are not well constrained because the oceanic crust that lies offshore the Grand Banks and Iberia was formed during the Mesozoic constant magnetic polarity interval and because no large fracture zones occur at this latitude. A useful constraint is provided, however, by the northern termination of the J magnetic anomaly in the Iberia Abyssal Plain, a probable isochron slightly older than M0, which should have been contiguous with the north end of the J anomaly in the Newfoundland Basin. The various reconstructions differ by a few tens of kilometers in the north-south direction. The situation is further complicated by intraplate deformation and "jumping" plate boundaries, which imply that Iberia was alternately attached to Africa or Europe (Olivet et al., 1984; Srivastava et al., 1990a). The reconstruction by Srivastava and Verhoef (1992) is now generally regarded as the most closely constrained (Fig. 2); they also attempted, in another version, to include destretching of the continental crust.

Review of Studies of the West Iberia Margin

Research on the west Iberia margin has been concentrated in three separate segments of the margin; these are the segment that lies west of Galicia Bank (which included Leg 103), the southern Iberia Abyssal Plain segment (Legs 47B, 149, and 173), which extends from the Vasco da Gama Seamount to the Estremadura Spur, and the Tagus Abyssal Plain segment (Fig. 1). These investigations contributed to studies of lithospheric extensional structures in the crust and mantle (shear zones, detachment faults, block faulting), the emplacement and seabed exposure of mantle rocks, synrift magmatism, the onset of seafloor spreading, and the characterization of the OCT (the OCT is here defined as the region between the most seaward, unambiguously continental fault block and oceanic crust). The margin segments exhibit both similarities and differences.

Galicia Bank Segment

The western margin of Galicia Bank, the northernmost of the three segments, has been studied with seismic refraction, seismic reflection, and heat flow profiles (de Charpal et al., 1978; Hoffman and Reston, 1992; Sibuet et al, 1987, 1995; Louden et al., 1989; Mauffret and Montadert, 1987; Mauffret and Montadert, 1988; Montadert et al., 1979a; Reston et al., 1995; Reston et al., 1996; Whitmarsh et al., 1996b), and has also been sampled extensively with dredges (Boillot et al., 1979; Dupeuple et al., 1976, 1987; Groupe Galice, 1979; Mauffret and Montadert, 1987), submersibles (Boillot et al., 1988a, 1995a; Mamet et al., 1991), and by drilling (Boillot, Winterer, et al., 1988). The Galicia margin is the first place where mantle rocks were sampled in the OCT (Boillot et al., 1980), demonstrating that oceanic accretion does not immediately follow continental breakup.

A seismic refraction model across the margin shows a thinned continental crust immediately east of a ~25-km-wide OCT, which in turn lies adjacent to a moderately thin (3 km) oceanic crust, which thickens rapidly to the west (Whitmarsh et al., 1996b). A layer with 7.0–7.6 km/s velocity, which underlies the thinned continental crust and the OCT, may represent either crustal underplating (Horsefield, 1992) or serpentinized upper mantle (Whitmarsh et al., 1996b). The seismic model is consistent with a gravity profile that indicates the margin is in isostatic equilibrium. Because oceanic crust that formed during the Cretaceous constant magnetic polarity interval abuts the OCT at this latitude, conventional seafloor magnetics cannot date the beginning of seafloor spreading. However, Leg 103 scientists were able to recognize an unconformity in their cores that they interpreted as the "breakup unconformity" (Boillot, Winterer, Meyer, et al., 1987). The recognition of the reversed polarity paleomagnetic interval M0 in cores from below this unconformity at Site 641 (Ogg, 1988) indicates that breakup occurred about 120 Ma (Aptian in the

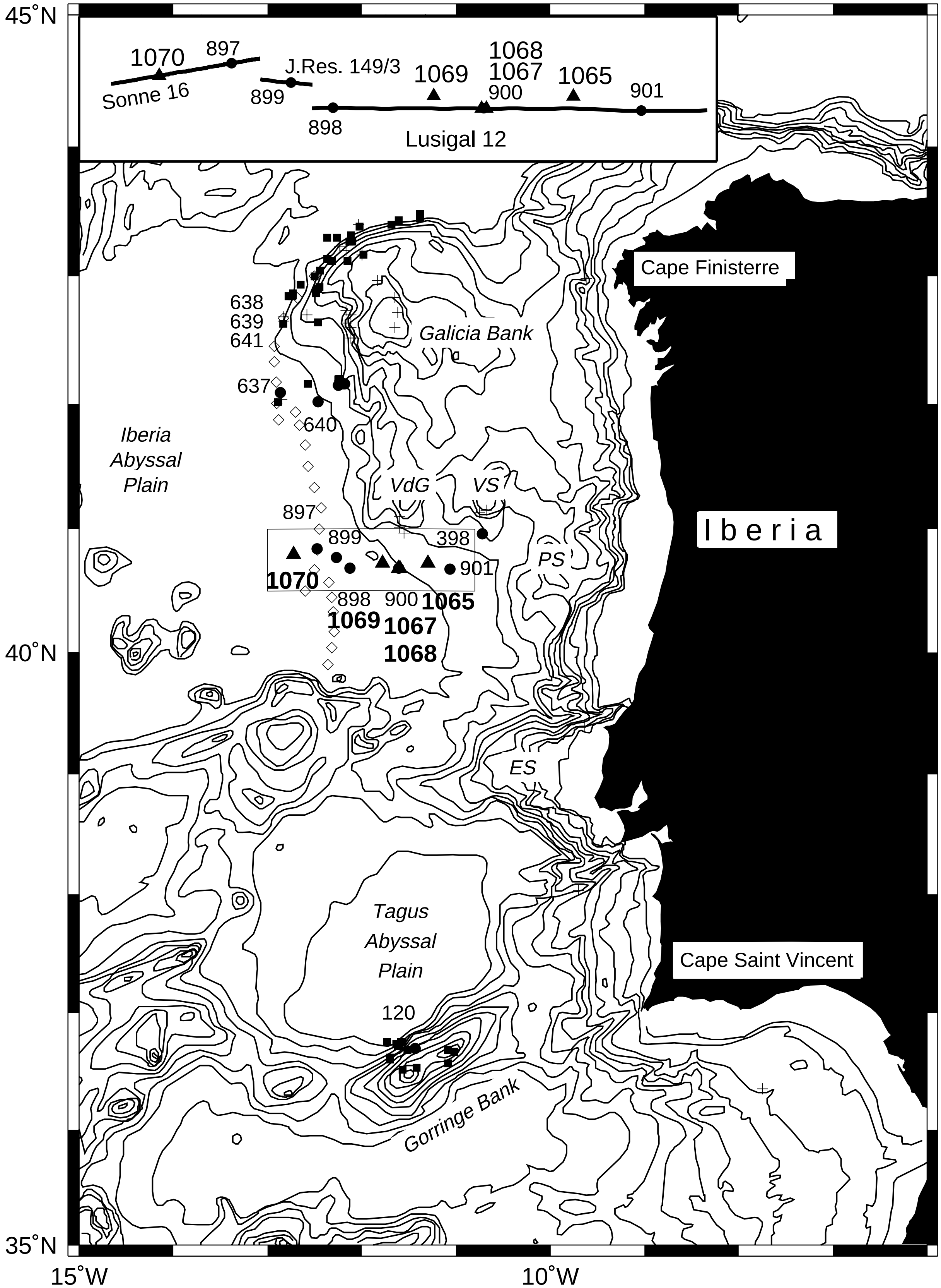

Figure 1. Bathymetric chart of the west Iberia margin (contours at 200, 500, 1000, 1500 through 5500 m). Existing DSDP/ODP sites are shown by solid circles. Sites drilled during Leg 173 are shown by solid triangles. Submersible dives that sampled rock and dredge sites are shown by solid squares and crosses, respectively. Open diamonds trace the peridotite ridge. VdG = Vasco da Gama Seamount; VS = Vigo Seamount; PS = Porto Seamount; ES = Estremadura Spur. Inset shows locations of drill sites relative to three seismic reflection profiles used to create the composite section in Figure 3.

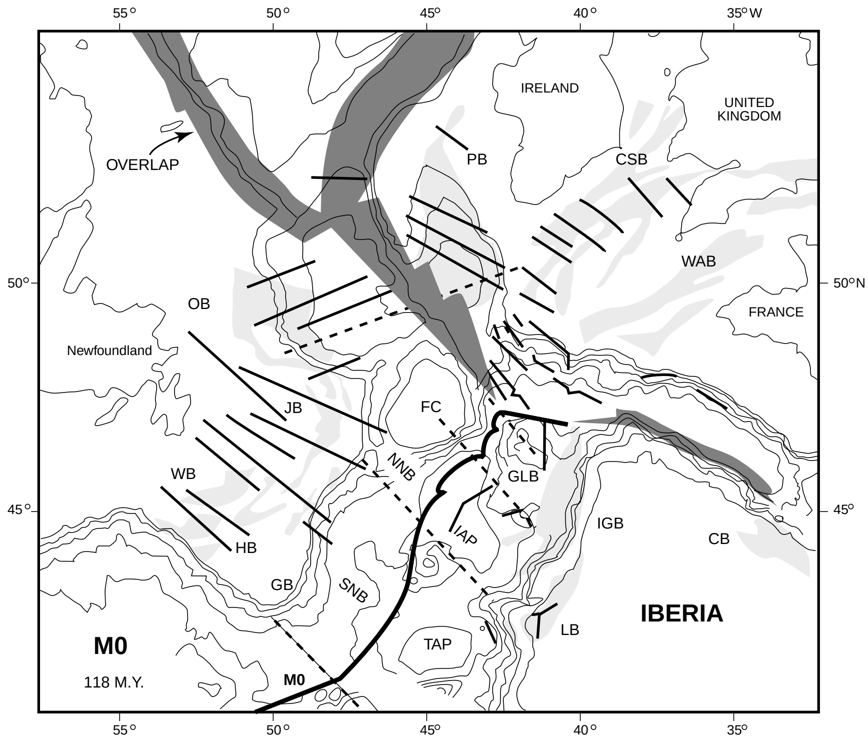

Figure 2. Plate tectonic reconstruction, showing the best estimate of the prerift arrangement of the Newfoundland and west Iberia margins at the time of magnetic anomaly M0 (Srivastava and Verhoef, 1992). The figure shows a simplified bathymetry on each plate, outlines of the sedimentary basins (shaded regions), and their plate tectonic features (continuous lines). Also shown are the direction of plate motion (dashed lines) and the resulting overlap between plate boundaries (dark stippled regions). PB = Porcupine Basin; CSB = Celtic Sea Basin; WAB = Western Approaches Basin; OB = Orphan Basin; JB = Jeanne d'Arc Basin; WB = Whale Basin; HB = Horseshoe Basin; GB = Grand Banks; SNB = South Newfoundland Basin; TAP = Tagus Abyssal Plain; IGB = Inner Galicia Basin; CB = Cantabrian Basin; IAP = Iberia Abyssal Plain; NNB = North Newfoundland Basin; GLB = Galicia Bank; FC = Flemish Cap.

Mesozoic time scale of Gradstein et al. [1995] which is used in this volume). This date is in good agreement with the seismic stratigraphy of the margin, which also suggests a middle or late Aptian age of the "breakup unconformity," as was confirmed at DSDP Site 398 and during ODP Leg 103 (Mauffret and Montadert, 1988; Montadert et al., 1979a; Sibuet et al., 1979).

A strong subhorizontal intrabasement reflector, the so-called S reflector, marks the lower limit of tilted fault blocks on the seismic reflection data (de Charpal et al., 1978). It is a major synrift feature, which was recently interpreted either as a major tectonic contact between the mantle and the crust related to synrift conjugate shear zones in the ductile lower crust and upper mantle (Beslier and Brun, 1991; Brun and Beslier, 1996), or as a major brittle detachment fault that controlled the final breakup of the lithosphere (Reston et al., 1996).

Sampling of the basement has shown that a north-south basement ridge, which marks the oceanward edge of the OCT for more than 120 km off Galicia Bank, is made of serpentinized peridotite and intrusive gabbros (Boillot et al., 1980, 1988a, 1995a; Boillot, Winterer, et al., 1988). Modeling of a deep-towed magnetometer profile strongly suggests that this ridge marks an abrupt change from relatively strongly magnetized oceanic crust to the west to at least ten times more weakly magnetized source rocks to the east (Sibuet et al., 1995). The peridotites are spinel- and plagioclase-bearing harzburgite and lherzolite. They underwent four main events during their high-temperature evolution (Beslier et al., 1990; Evans and Girardeau, 1988; Girardeau et al., 1988): first, high-temperature (900°–1000°C) shear deformation; second, limited partial melting; third, subsolidus reequilibration in the plagioclase stability field; and fourth, mylonitic shear deformation at a lower temperature (850°C), under high deviatoric stress, and at low pressure (i.e., in lithospheric conditions) in a normal shear zone slightly dipping towards the continent. The rocks were then fractured during and after serpentinization under subsurface conditions (Agrinier et al., 1988; Girardeau et al., 1988). This evolution is compatible with uplift beneath a rift zone. Mantle melts occur as gabbro and as a mylonitic Mg-chloritic rock that tentatively is believed to be derived from a Fe-Ti-rich gabbro (Schärer et al., 1995). Dating by ^{40}Ar-^{39}Ar of an intrusive dioritic dike, which displayed a mylonitic foliation subparallel to that observed in the surrounding peridotite, indicates a synrift age of the shear deformation (122 ± 0.3 Ma; Boillot et al. [1989]). Wilson et al. (1996) re-examined published seismic reflection data from the Galicia margin in the light of their seismostratigraphic study of the southern Iberia Abyssal Plain (see below). They found no convincing evidence of reflection divergence into fault footwalls indicative of synrift deposition, and so concluded that "earlier studies did not correctly identify the seismic expression and timing of the rifting episode". They suggested that rifting in the ODP Leg 103 area occurred after the deposition of Tithonian-early Berriasian carbonates and before early Valanginian turbiditic sediments that drape the tilted fault blocks formed during rifting. The basement of the tilted fault blocks is mostly composed of granite and granodiorite (i.e., of upper continental crust), characteristics that suggest they belong to the Ossa Morena Variscan zone known on land in Iberia (Capdevila and Mougenot, 1988). The local presence of a late Devonian–early Carboniferous basin within the basement is in agreement with that conclusion (Mamet et al., 1991). Recently, a thick cataclastic breccia has been sampled at the top of the mylonitized and cataclastically deformed mantle that mixes angular and heterometric pieces from both the mantle and the continental crust. It is interpreted as a cold tectonic melange related to a major shear zone that accommodated the lithospheric thinning at a late stage of the rifting, just before the final breakup (Boillot et al., 1995a). A thin basaltic layer covers the oceanward side of the northernmost part of the peridotite ridge. Dating of the basalts indicates that they were emplaced about 100 ± 5 Ma ago, a date that is younger than the time of breakup (Malod et al., 1993).

Southern Iberia Abyssal Plain Segment

The southern Iberia Abyssal Plain (IAP) margin, which lies between 39° and 41°N, has an unusually wide OCT (up to 130 km), clear crustal block faults, peridotite basement (exposed both along one or more narrow ridges and possibly even more extensively) and evidence of limited synrift magmatism. Here, there is apparently no obvious deep and extensive subhorizontal strong reflector like S, although several intrabasement reflectors have been tentatively interpreted as tectonic contacts, one of which is dissected by higher angle normal faults (Krawczyk et al., 1996). Such reflectors can be interpreted as evidence that the uppermost lithosphere of the OCT was strongly affected by extensional tectonics. Using digitized versions of their interpreted seismic lines, Wilson et al. (1996) produced smoothed total tectonic subsidence (TTS) plots from which extension factors (β) were determined. The plots suggest that a TTS of up to 6 km occurred, which indicates subsidence of either normal thickness oceanic crust or of continental crust thinned by a β of between 5 and 7 at 135 Ma. The results of Leg 149 highlighted the need for more basement drilling, principally within the OCT, in order to understand the rift-to-drift tectonic and magmatic processes at this excellent example of a nonvolcanic rifted margin. Further, independent geophysical and laboratory work since Leg 149 has led to revised tectonic and magmatic models (see below) for the rifting and initial seafloor spreading at this margin (Brun and Beslier, 1996; Krawczyk et al., 1996; Pickup et al., 1996; Sibuet et al., 1995; Whitmarsh and Miles, 1995; Whitmarsh and Sawyer, 1996), which could be tested during Leg 173 by further drilling to basement.

Whitmarsh et al. (1993) studied this segment using seismic refraction and reflection profiles, gravity, and magnetics. They found that the oceanic crust adjacent to the OCT is thin (4 km) and the OCT is underlain by a layer of about 7.6 km/s. More extensive profiles acquired in 1995 are currently being worked on (Dean et al., 1996). Modeling of a single east-west gravity profile across the Iberia Abyssal Plain, which was constrained by a multichannel seismic reflection profile and by the earlier seismic refraction profiles, appeared to confirm the existence of thinned continental crust landward of the unusually thin oceanic crust and the existence of a continuous subcrustal 3.26 Mg/m^3 layer corresponding to the velocity of about 7.6 km/s (Whitmarsh et al., 1993). A regional magnetic anomaly chart and modeling of magnetic anomaly profiles across the Iberia Abyssal Plain (Whitmarsh and Miles, 1995) strongly suggest that seafloor spreading began about the time of anomaly M3 (126 Ma, Barremian) but that crust to the east of M3 was weakly magnetized and included a transitional zone between a peridotite basement ridge and continental crust. This analysis appears to be confirmed by deep-towed magnetometer profiles, which also conclusively indicate that the transitional region locally, in the vicinity of two probably peridotite basement highs, has relatively high magnetization intensities (Whitmarsh et al., 1996a). The basement morphology and associated postrift sediment folding strongly suggest that the peridotite basement ridge represents the counterpart in the Iberia Abyssal Plain of the peridotite ridge drilled off Galicia Bank (Beslier et al., 1993; Boillot et al., 1987b; Masson et al., 1994).

Two possible explanations for the unusual 7.6 km/s layer have been proposed (Whitmarsh et al., 1993). First, it may represent underplated material emplaced at the time of rifting and breakup as in the model of White and McKenzie (1989); although such material is commonplace on volcanic rifted margins, the essential lack of synrift volcanism on the west Iberia margin, the restriction of the layer to only the parts of the OCT where the seismic "crust" is thinnest, and the rather high velocity associated with the layer all strongly suggest this explanation is unlikely to be correct. Second, the layer may represent serpentinized upper mantle peridotite. Serpentinization could have occurred during the late stages of continental rifting, immediately after lithospheric breakup and during the onset of seafloor spreading because of the relatively easy access of seawater (possibly thermally driven) to the upper mantle through the thin overlying basement layer. This is analogous to the serpentinization of the uppermost mantle known to have occurred in oceanic fracture zones. Beslier et al. (1993) also favored this interpretation, by analogy with the existence of serpentinized peridotite and the seismic structure of the adjacent Galicia margin segment (Boillot et al., 1989, 1992). An even more extreme view was recently indicated by Pickup et al.'s (1996) interpretation of a deep multichannel reflection profile across the southern Iberia Abyssal Plain, on which a seismically unreflective uppermost basement layer is interpreted as highly serpentinized peridotite that experienced vigorous, hydrothermally driven convection of seawater to a depth of 1.0–2.5 km.

Wilson et al. (1996) were unable, save for one possible exception, to identify any unequivocal synrift intervals on seismic lines shot across the Iberia Abyssal Plain. They concluded that the rift episode "occurred over a very short interval of time so that synrift sedimentary successions did not accumulate to produce a seismically resolvable thickness". On the basis of seismic profiles and ODP core data obtained from the Iberia Abyssal Plain and a reinterpretation of similar data from the Galicia Margin (summarized above), they concluded that rifting occurred between the late Berriasian and early Valanginian in both areas. This interval between 134 and 140 Ma is consistent with the $^{40}Ar/^{39}Ar$ age of 136.4 $\pm$ 0.3 Ma related to the end of shear deformation reported by Féraud et al. (1996) from Site 900. However, a rift duration as short as ~5 m.y. should have resulted in the generation of substantial quantities of melt (Bown and White, 1995), for which there is no evidence.

Tagus Abyssal Plain Segment

In the Tagus Abyssal Plain, Pinheiro et al. (1992) showed magnetic models that indicate seafloor spreading began about 133 Ma (Valanginian). They used seismic refraction, seismic reflection, and magnetic profiles to show that the oceanic crust adjacent to the OCT is unusually thin (2 km) and that there is a transitional region between the thinned continental crust and the thin oceanic crust, which, although not truly oceanic (for example, it has no seafloor-spreading magnetic anomalies), has a magnetization far stronger than is usually associated with continental crust. This may indicate the presence of intrusive and extrusive material within the basement of the transitional region. The thin oceanic crust is underlain by a 7.6 to 7.9 km/s layer, which is probably serpentinized peridotite.

Gorringe Bank, which marks the southern limit of the Tagus Abyssal Plain, is a 180-km-long and over 4000-m-high east-northeast–west-southwest ridge adjacent to the Azores/Gibraltar plate boundary. The bank is a slab of Mesozoic lithosphere uplifted during the Cenozoic. The bank is capped by two mounts of similar water depth: Mount Gettysburg (24 m) to the southwest and Mount Ormonde (50 m) to the northeast. DSDP Leg 13 demonstrated that the bank is mantled by a cover of Barremian–Aptian pelagic sediments (Shipboard Scientific Party, 1972). Several cruises with dredging (Prichard and Cann, 1982) and diving (Auzende et al., 1979; Auzende et al., 1978; Cyagor Group, 1977; Cyagor II Group, 1984) showed that Mount Gettysburg essentially consists of largely serpentinized peridotites and crosscutting doleritic dikes. The mantle rocks are spinel-bearing harzburgites, with a few dunite and websterite bands; locally, plagioclase and clinopyroxene occur as trace evidence of impregnation by basaltic magmas (Serri et al., 1988). Toward the northeast, a supposed tectonic contact superposes gabbros on the peridotites. Mount Ormonde is mostly composed of various gabbros corresponding to different stages of differentiation (Cyagor II Group, 1984). They are crosscut by numerous dikes, either doleritic or alkaline (Cornen, 1982), which are significantly younger (65–67 Ma; Féraud et al., 1986). Locally, the gabbros display narrow mylonitized bands that developed at high temperature before an episode of hydrothermal metamorphism (Mével, 1988) and low-temperature brecciation. Some preliminary results from recent dives (Girardeau et al., in press) show the occurrence of some lherzolites in the harzburgites and of gabbros on Mount Gettysburg, and suggest that the mylonitization of the gabbros on Mount Ormonde is associated with the original Early Cretaceous opening of the Atlantic Ocean at this latitude. Therefore, Gorringe Bank may represent a cross section of oceanic

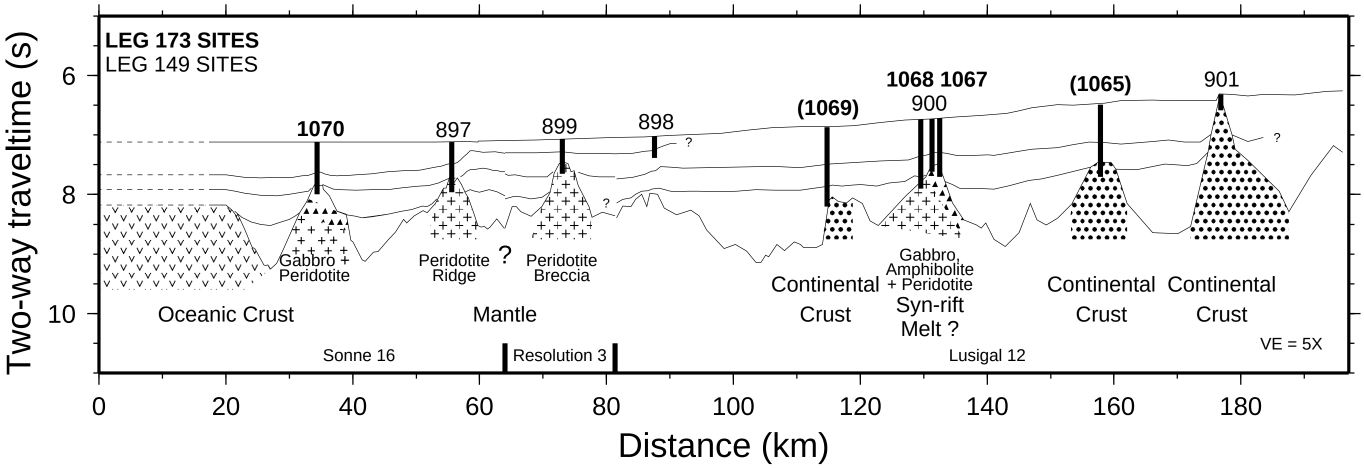

Figure 3. Composite west to east (left to right) cross section through the Leg 149 and Leg 173 drill sites along the tracks shown in Figure 1 (inset). Sites in parentheses are offset a short distance from the profile. Depths and extents of patterns are diagrammatic. Solid triangles = gabbro and amphibolite; + = peridotite.

crust and upper mantle (Cyagor II Group et al., 1984; Féraud et al., 1986; Prichard and Cann, 1982), comparable to oceanic lithosphere formed during the early stages of oceanic spreading (Girardeau, et al., in press), or even a cross section of the OCT itself (Girardeau et al., in press; Whitmarsh et al., 1993).

THE OCEAN/CONTINENT TRANSITION IN THE SOUTHERN IBERIA ABYSSAL PLAIN

Results of Leg 149

During Leg 149 a west-to-east transect of five sites was drilled in the southern Iberia Abyssal Plain, three of which (Sites 897, 899 and 900) reached acoustic basement (Figs. 1, 3). A fourth site (Site 901) enabled a firm prediction to be made that the underlying basement is continental crust. The sites had been chosen in the context of a conceptual model of the location of the OCT previously defined by gravity, magnetic and seismic velocity modeling, and by seismic reflection profiles and with the advantage of information acquired from the adjacent Galicia margin segment. The results obtained during the leg broadly confirmed this model but also produced some surprises (Sawyer, Whitmarsh, Klaus, et al., 1994; Whitmarsh, Sawyer, Klaus, Masson, 1996).

Site 897, situated on the crest of a narrow elongate basement ridge, penetrated up to 153 m of partly brecciated serpentinized peridotites, the nature and evolution of which are comparable to those of the Galicia margin. The peridotites differ in having a larger variety of types, from spinel dunites and harzburgites to spinel-plagioclase-bearing peridotites, including harzburgites, lherzolites, and pyroxenites. They underwent limited partial melting under plagioclase facies conditions that was followed by some impregnation by melt products during the last stages of a high-temperature (~900°C) shear deformation event (Beslier et al., 1996; Cornen et al., 1996a). A subsolidus reequilibration event occurred, during ongoing ductile shear deformation under lithospheric conditions (high deviatoric stress, P<1 GPa), which ceased at a temperature close to 735°C. A complex and intense deformation, which occurred under subsurface conditions during the serpentinization of the rocks (Agrinier et al., 1996; Milliken et al., 1996), documented the late emplacement of the tectonically denuded mantle dome at the rift axis, suggesting that the mantle can be highly stretched in the OCT after continental breakup (Beslier et al., 1996). As Beslier et al. (1993) suggested from seismic reflection data, these results confirm that mantle rocks form a more or less continuous basement ridge at least 300 km long in the OCT of the west Iberia margin. Modeling by Whitmarsh et al. (1996a) of deep-towed magnetometer profiles over the southern Iberia Abyssal Plain, the topography of the top of acoustic basement (Fig. 4), and seismic velocities of Whitmarsh et al. (1990) suggest that the Site 897 ridge lies at the landward edge of thin oceanic crust that formed in late Barremian time at a spreading half-rate of 10 mm/yr (~124 Ma ago).

Site 899, situated on an isolated sub-circular basement high (Fig. 4), encountered an unusual sequence of three serpentinite breccia units with minor fossiliferous claystones containing early Aptian nannofossils. Beneath lies an early Aptian sequence of mass flow deposits consisting of boulder-sized blocks of unbrecciated serpentinized peridotite and other minor clasts and intercalated claystone and siltstone. The peridotites appear to have experienced the same petrostructural evolution as at Site 897. The petrology of the rocks differs only in detail with no websterites and fewer harzburgites and dunites relative to lherzolites; plagioclase-bearing lherzolites have been clearly identified only at Site 899 (Cornen et al., 1996a). It is almost certain that at Site 899 true basement was not reached but a peridotite outcrop may not have been far away at the time of emplacement of the breccia units. It has been suggested that these are either submarine landslide deposits (Gibson et al., 1996) or cataclastic breccias included in an olistostrome derived from a possible, but so far unidentified and more distant, transform fault near the southern margin of Galicia Bank (Comas et al., 1996). However, this site may be atypical of the OCT within which it lies, because a high-resolution magnetic anomaly chart, produced since Leg 149 (Fig. 5), clearly shows the site is associated with an isolated magnetic high whose amplitude and shape are atypical of the rest of the OCT (Miles et al., 1996). Further, modeling of deep-tow magnetic profiles indicates that the bulk magnetization of the crust is unusually high, higher than oceanic crust, in the vicinity of Site 899 (Whitmarsh et al., 1996a); this is supported by measurements on the cores that indicate that the Site 899 peridotites are, on average, five times more strongly magnetized than the Site 897 peridotites (Zhao, 1996).

At both Sites 897 and 899, the clasts in various mass-wasting deposits provide important evidence of the sorts of rocks that were exposed within the OCT during rifting. At Site 897 the conglomerates of Subunit IIIB yield rounded clasts of claystone and chalk/limestone

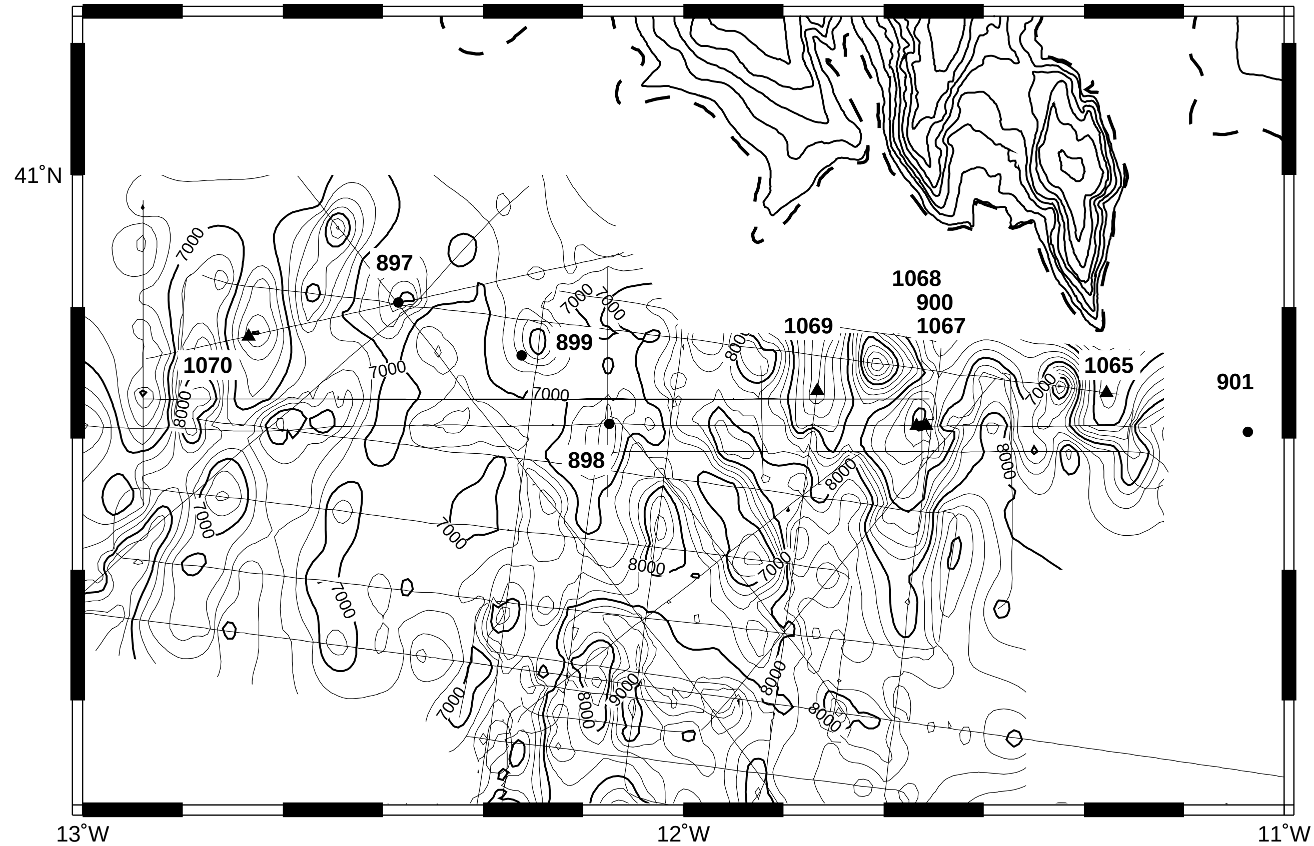

Figure 4. Contour chart of depth below sea level to basement (contour interval 250 m; contouring based on work by C.M. Krawczyk, L.M. Pinheiro, S.M. Russell, and R.B. Whitmarsh) combined with bathymetry of the relief bordering the Iberia Abyssal Plain (contour interval 250 m; courtesy J-C. Sibuet). Dashed line denotes edge of the abyssal plain. Leg 149 and Leg 173 sites are shown by solid circles and solid triangles, respectively. Fine lines are tracks of seismic lines used to contour basement.

with minor dolomite, basalt, arkosic to lithic sandstones, mica schist, basalt, serpentine, and mica. At Site 899 Unit IV contains clasts of metamorphosed or unmetamorphosed ultramafic and mafic rocks (serpentinized peridotite, submarine basalts, microgabbros, chlorite-bearing schists, and amphibolite). Thus, quite a wide variety of igneous, and some metamorphic, rocks can be found as clasts in the coarser fraction of sedimentary units at Sites 897 and 899. Except for the Site 900 gabbro (see below), most mafic rocks display a wide range of composition that overlaps the tholeiitic and transitional fields; some lavas have alkaline affinities (Cornen et al., 1996b; Seifert and Brunotte, 1996). The chlorite-bearing schists may be considered to be former differentiated Fe-Ti leucogabbro or plagiogranite (Cornen et al., 1996b). This does not strongly support the idea of a widespread MORB-like oceanic crust in the OCT. Other clasts suggest the existence of a variety of prerift sedimentary rocks but it is impossible to say whether these cropped out on the continental shelf and slope or closer to the sites.

Site 900 penetrated 56 m of fine- to locally coarse-grained flasered gabbro that underwent three main tectono-metamorphic events during its evolution. First, the gabbro was highly sheared under relatively high-temperature (high amphibolite to granulite facies) and high-pressure (≥0.4 GPa) conditions (Cornen et al., 1996b) that ended around 136 Ma in lower grade metamorphic conditions (Féraud et al., 1996). This deformation generated a clear foliation in the rocks and a granuloblastic to porphyroclastic texture that obviously results from dynamic recrystallization during intense shear deformation (Féraud et al. (1996); Beslier et al., unpubl. data). Because both the neoblasts and the porphyroclasts of plagioclase and clinopyroxene have similar compositions, it is inferred that the high-pressure and high-temperature conditions still existed during the shearing event (Cornen et al., 1996b). Second, retrometamorphism to greenschist facies conditions under static conditions occurred after the ductile shear deformation. Third, a complex and intense extensional deformation under greenschist facies conditions, aided by hydrothermalism, overprinted the previous features. It is mainly expressed by fracturing with local brecciation of the rocks (Beslier et al., unpubl. data). The origin of these rocks is still a matter of debate. Seifert et al. (1996) and Seifert et al. (1997) consider them to be cumulate gabbros because they have high initial ε_{Nd} values (+6.3 to +10.3 at 136.4 Ma) that are normally restricted to MORB rocks. Therefore, they infer that an active mid-ocean ridge magma chamber was present but that subsequently the associated spreading center was abandoned and "jumped" westward. However, the high-grade shearing deformation is not compatible with an evolution at an immature spreading axis. Cornen et al. (1996b) consider that the flaser gabbros have major- and trace-element compositions that are closer to those of transitional magmas formed at the beginning of rifting than to normal MORB-like magmas. They propose that these rocks have crystallized from transitional to tholeiitic synrift magmas that accumulated and slowly cooled at the base of the continental crust as underplated material, which was subsequently sheared during continental stretching. This hypothesis, however, is unable to explain the MORB isotopic signa-

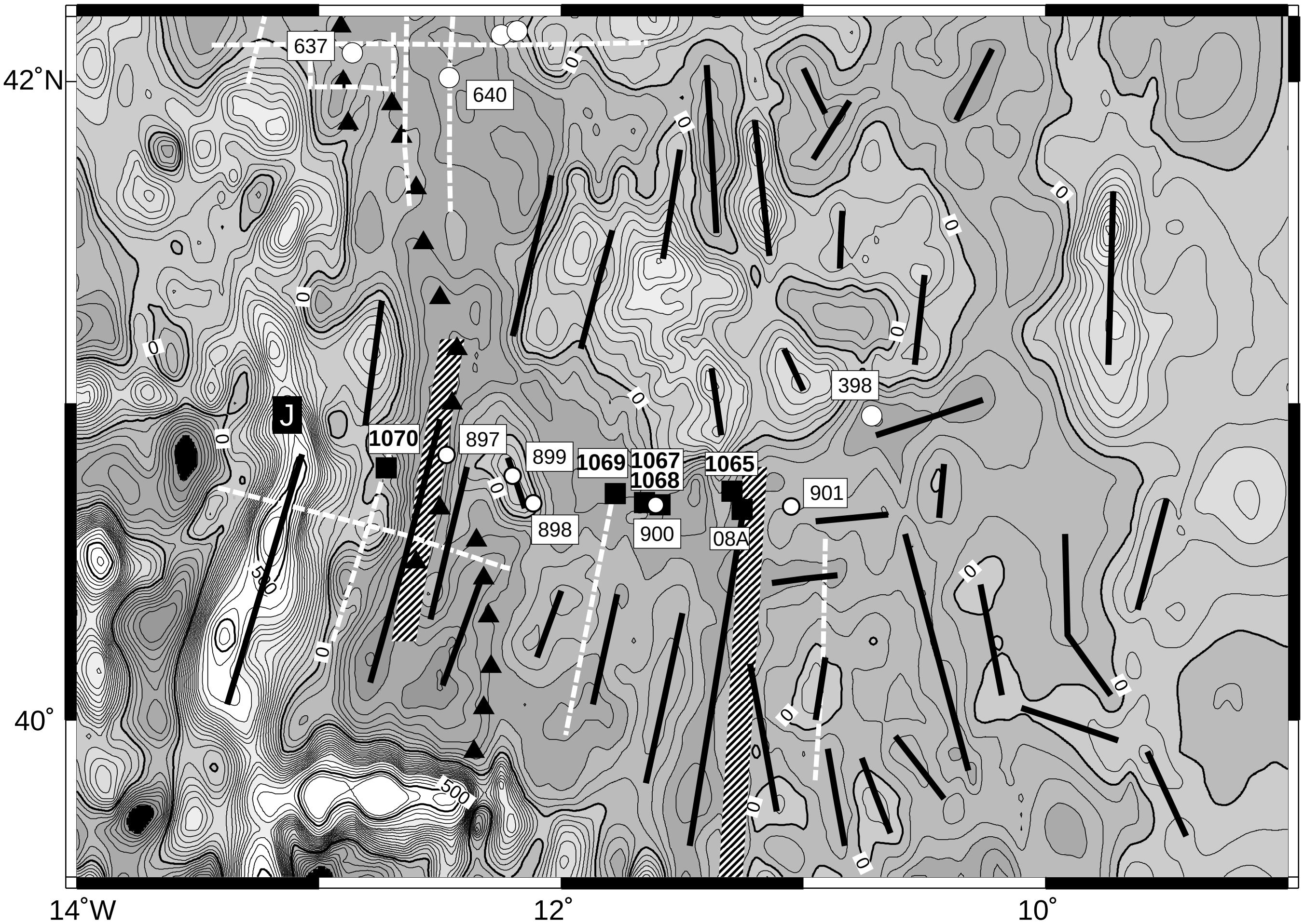

Figure 5. Part of a new reduced-to-the-pole magnetic anomaly chart of the whole west Iberia margin produced in collaboration with the Atlantic Geoscience Centre, Dartmouth, Nova Scotia, Canada (Miles et al., 1996). The chart was made from a 5-km gridded data set and is contoured at 25 nT (zero contours are labeled). The main chart is based on over 400,000 sea-surface observations, which were corrected to remove the effects of secular variation, high geomagnetic activity, spurious tracks, and systematic cross-over errors. Greater confidence in the quality of the resulting data set allowed the use of the small contour interval. The chart shows pre-Leg 173 drill sites in this area (open circles); Leg 173 sites are shown by solid squares (08A = IBERIA-08A which was not drilled). The data were reduced to the pole to clarify many features. Major linear trends in the anomalies are delineated with bold lines. White dashed lines denote seismic refraction lines of Whitmarsh et al. (1990, 1996b). Bands of diagonal lines were used by Whitmarsh and Miles (1995) to separate zones with different characteristic magnetic anomalies. The chart clearly shows the strong positive J anomaly which appears just west of 13°W and south of 41°30′. Between the J anomaly and the continental shelf (~9°15′W), other less strong positive anomalies are associated with the shallow regions of Galicia Bank.

ture of these rocks, at least partly because we are ignorant of the geochemical signature of such underplated material. Additional data, in particular geochemical data, are needed to better constrain the origin of these rocks and to help discriminate between these two hypotheses. If the second hypothesis is finally favored, then the Iberia Abyssal Plain is a unique place to characterize underplating processes at nonvolcanic passive margins.

Site 901 penetrated sediments that conformably overlie a tilted fault block but time constraints caused drilling to be terminated before basement was reached; the hole encountered Late Jurassic (early Tithonian; 149–151 Ma) nannofossils and foraminifers in terrestrial-plant–rich sediments, some 350 m above apparent acoustic basement. These sediments contain neritic (<200 m) benthic foraminifers but no deep-water species. The originally shallow depth and great subsidence (~4500 m) of this site and the sediment age (~24 m.y. older than the best estimate of the onset of seafloor spreading at this latitude) strongly suggests it is underlain by continental crust.

The crestal location on basement highs of late Barremian–Aptian debris flows and rock fall deposits cored at Sites 897 and 899 puzzled Leg 149 scientists. This was because such mass flow deposits would be expected to flow downslope away from the crests and to accumulate in the basins between the highs. Moreover, the deposits are about 1 km higher than the top of seismostratigraphic Unit 6 as defined by Wilson et al. (1996), which is believed to be Hauterivian/possible Valanginian to Aptian in age. As yet there is no commonly agreed-upon explanation for this vertical separation, nor for the occurrence of mass wasting deposits at crestal locations. Using the revised post-cruise biostratigraphy, Whitmarsh and Sawyer (1996) suggested that significant basement relief was created at both sites *after* deposition of these deposits, in early late Aptian (~116 Ma) at Site 897 and in early Aptian (~117–121 Ma) at Site 899, 6–11 Ma after the best estimate of the onset of 10 mm/yr seafloor spreading (approximately the start of anomaly M3, 127 Ma). Wilson et al. (1996) found no evidence for deformation of younger seismostratigraphic units that would be expected if the basement ridges had been uplifted after the Aptian. Following creation of the highs, sedimentation was practically insignificant (0.4 m/Ma at Site 897 and 1.3 m/Ma at Site 899) for over 60 Ma until the sites were eventually inundated by abyssal plain sediments in the early–middle Eocene.

Thus, in summary, the results of Leg 149 combined with the geophysical data proved the existence of a peridotite ridge at the landward edge of the inferred oceanic crust formed by ~10 mm/yr half-rate seafloor spreading. They also showed that between this ridge and Site 901, which is situated on a fault block of almost unequivocal continental crust, there exists a 130-km-wide OCT, much of which is probably underlain by a heterogeneous basement. One indication of the transitional nature of this basement may be the MORB-like to transitional gabbro at Site 900; other indications are the presence of petrogenetically transitional to alkaline mafic clasts in mass wasting deposits at Sites 897 and 899. The magnetic and seismic reflection character and velocity structure of the basement provide additional evidence that has been used to suggest a broad zone of subsediment mantle outcrop (see below). Whether the Site 900 gabbro is pre- or synrift in age, the original granulite metamorphic grade and the ^{40}Ar/^{39}Ar age of the end of the ductile deformation implies that the gabbro was exhumed by important synrift tectonics that accompanied lithospheric extension. Although Site 899 sampled a serpentinite breccia and an underlying serpentinized peridotite mass-flow deposit, magnetic evidence suggests that the basement of this site may be atypical of the rest of the Iberia Abyssal Plain OCT and it may not be correct to infer continuity of peridotite basement between Sites 897 and 899. Models that explain many of these observations are presented below. Leg 173 planned to test such models by focusing on the nature and evolution of the basement itself within this zone and on the possible major tectonic contacts that have been identified there.

MODELS FOR THE FORMATION OF THE OCT IN THE SOUTHERN IBERIA ABYSSAL PLAIN

Based on the Leg 149 results, the latest time- and depth-migrated seismic reflection profiles in the southern Iberia Abyssal Plain, an interpretation of the new magnetic anomaly chart (Fig. 5), and other geophysical observations a series of preliminary tectonic and magmatic models for lithospheric rifting and OCT formation on the west Iberia margin has been produced since Leg 149 (Brun and Beslier, 1996; Krawczyk et al., 1996; Pickup et al., 1996; Whitmarsh and Sawyer, 1996). The models emphasize different processes (e.g., continental stretching and breakup, or the nature and structure of the OCT formed after continental breakup), and differ in the relative importance attributed to the nature of the basement cores relative to the geophysical observations, in the significance attributed to the peridotite ridge, and in whether the east-west distribution of different basement rocks within the OCT is considered to be systematic or just random. Such a variety of approaches was valid after Leg 149 because the number of drill sites that had reached basement remained very small.

The principal feature of the reduced-to-the-pole magnetic anomaly chart is the existence of a region, extending from the peridotite ridge to 11°15′W, with low-amplitude linear anomalies that parallel the seafloor-spreading isochrons (Fig. 5), which have yet to be convincingly modeled by seafloor spreading (Whitmarsh and Miles, 1995; Whitmarsh et al., 1996a). East of 11°15′W, the anomalies are less well organized yet have a predominant trend about 30° different from those immediately to the west. A new model (Whitmarsh and Miles [1995]; Whitmarsh and Sawyer [1996]) explains the isochron-parallel anomalies in the OCT by the intrusion of material into acoustic basement, under the same stress conditions as eventually accom-

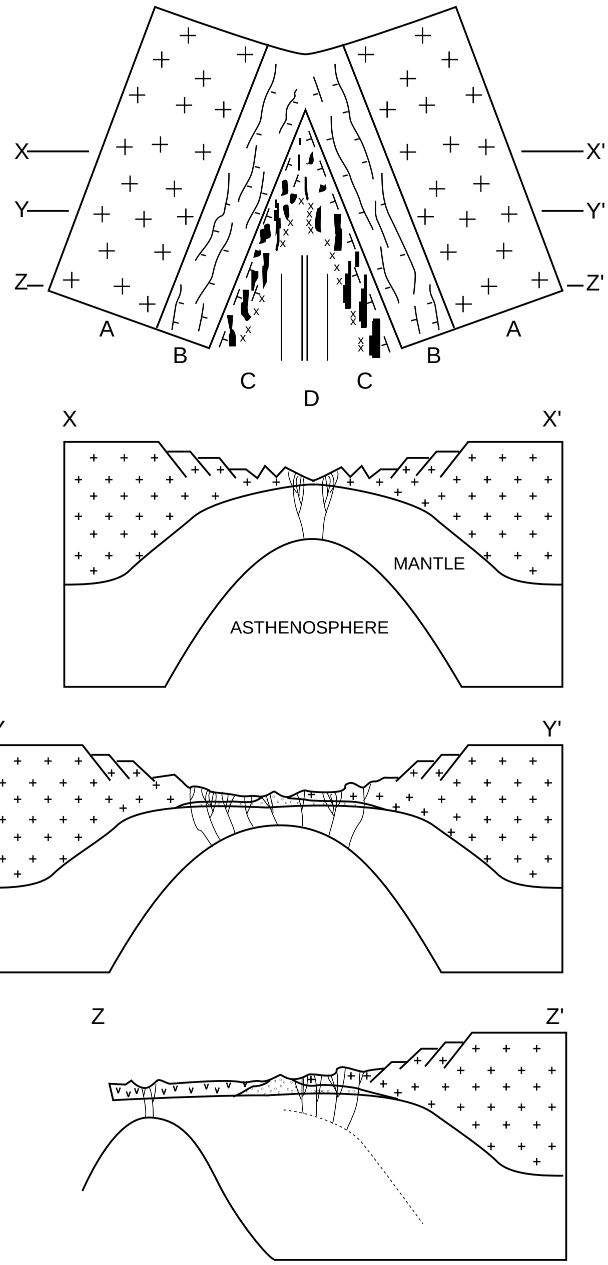

Figure 6. The magnetic, seismic velocity, and seismic reflection contrasts between the central and eastern zones in Figure 5 can be explained by a model featuring a propagating rift (Whitmarsh and Miles, 1995). The plan view (top of figure, not to scale) illustrates this scenario. The model includes A = unthinned continental crust; B = continental crust thinned by tectonic extension with upper crust normal faults; C = a transitional crust consisting of blocks of faulted continental crust and a high density of intrusives (black), which are oriented parallel to the seafloor spreading anomalies; D = oceanic crust produced by seafloor spreading. The three cross-sections (XX′, YY′, ZZ′; not to scale) show schematic views through the plan view. Irregular branching vertical lines = dikes, + = continental crust, v = oceanic crust, × = peridotite ridge, and the stippled area in YY′ and ZZ′ = serpentinized peridotite. Fault symbols (lines with tick marks in B and C) denote rifted continental crust on the top plan view.

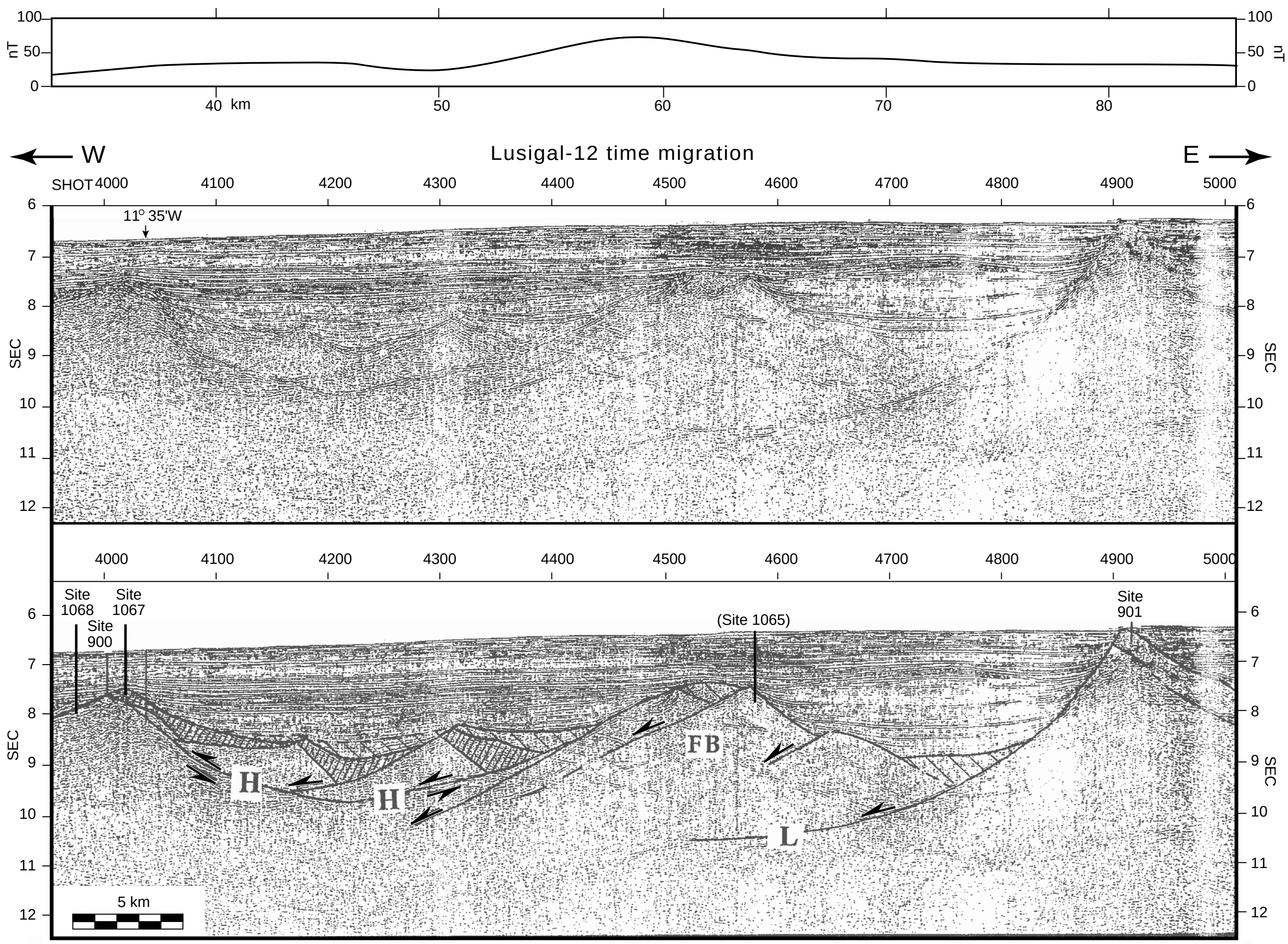

Figure 7. East-west migrated multichannel seismic reflection profile Lusigal-12 through Sites 1068, 1067, 900 and 901 (see Fig. 4 for location). The lower profile is an interpretation of basement reflections seen in the upper profile. Two phases of extension are proposed: one related to block tilting along the H intrabasement reflector, and one related to subsequent block faulting by westward normal faults as the L reflector. Synrift I sediments marked by close diagonal ruling; Synrift II sediments by coarser ruling. FB = fault block. The box at the top of the figure contains a reduced-to-the-pole magnetic anomaly profile.

pany the onset of seafloor spreading (Fig. 6), during the northward propagation of a rift between the proto-Iberia and North America plates. Propagating-rift models of continental breakup are indicated off west Iberia by the fact that the seafloor spreading anomalies “pinch out” against the margin so that progressively younger anomalies are found against the margin in the south-to-north direction in which continental breakup proceeded. Within the central (transitional) zone of the model, continental crust is surrounded and impregnated by intrusive, and possibly extrusive, material formed from passive upwelling of magma created by limited decompression partial melting that occurred during a late stage of continental rifting. The net effect is that the original continental crust becomes substantially extended normal to the margin by the addition of igneous material under the same stress conditions that eventually allow the steady-state creation of oceanic crust by seafloor spreading. The intrusives thus become aligned in a direction parallel to the eventual seafloor-spreading anomalies. Subvertical magnetization or susceptibility contrasts between such materials, and even within contrasting intruded material (e.g., reversals in polarity), could be the explanation for the low-amplitude linear magnetic anomalies, parallel to the seafloor spreading anomalies, which are observed today (see Fig. 5). A comparable situation may be exposed today on the Arabian margin of the central Red Sea (Coleman, 1993). A deep, rather than shallow, crustal magnetic source is qualitatively indicated by the lack of correlation between the reduced-to-the-pole magnetic anomalies and the basement relief along east-west multichannel seismic reflection profiles in this area (Figs. 4, 5, 7). Further evidence is provided by seismic refraction data. North-south Line 2, shot at 11°50′W in the transitional zone, has higher basement velocities, perhaps indicating the presence of mafic intrusives, than Line 1, shot just east of Site 901 (Fig. 5; Whitmarsh et al. [1990]). Thus, some geophysical data suggest that synrift magmatic intrusion may have occurred within the OCT. Gravity modeling based on these refraction data, and a more recent seismic model based on more extensive data collected along profile IAM-9 a few tens of kilometers to the south of the Leg 149 sites (Dean et al., 1996), suggest the widespread occurrence of a serpentinized peridotite layer within a few kilometers of the top of acoustic basement. This model can explain the synrift, apparently MORB-like

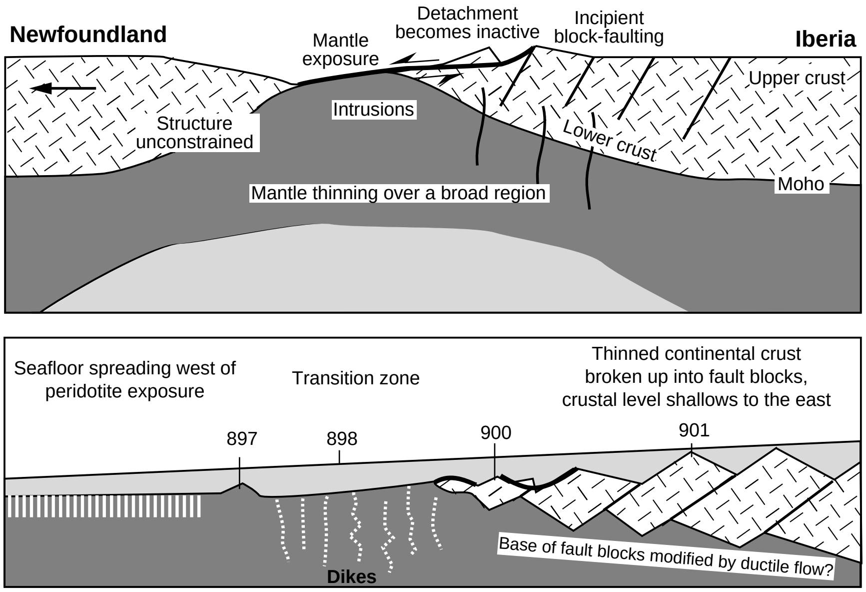

Figure 8. Diagram illustrating an initial, simplified model for extension of the lithosphere controlled by detachment faulting during rifting on the southern Iberia Abyssal Plain margin. In a first phase of rifting, lithospheric extension may have been accommodated by detachment faulting and accompanied by consequent melting and intrusion of gabbro into the lower crust at about 136 Ma (Féraud et al., 1996). Block-faulting subsequently cuts the upper plate into tilted blocks dismembering the detachment system and the lower plate. The resulting crustal cross section shows an oceanward deepening of the lithospheric level consistent with the results of Leg 149 (from Krawczyk et al., 1996).

gabbros at Site 900 as the product of local asthenospheric melting, much as Bonatti (1985) proposed the existence of punctiform initiation of seafloor spreading in the Red Sea.

An alternative interpretation of the region between Sites 897 and 901 is that it has formed by ultraslow (~5 mm/yr) seafloor spreading (Sawyer, 1994; Whitmarsh and Sawyer, 1996). This hypothesis is largely based on the suggestion of a MORB parental source for the Site 900 gabbro and draws strength from some similar characteristics of ultraslow, but generally steady-state, seafloor-spreading crust in the oceans. Although the model is unable to explain either the lineated magnetic anomalies between the peridotite ridge and 11°15′W (attempts to model these anomalies by seafloor spreading have so far been unsuccessful) or the transitional to alkaline mafic clasts in some cores, it predicts a scattered distribution of gabbro and serpentinized peridotite within the region. The model remains valid so long as no continental basement rocks are recovered from the region.

The basement of the eastern part of the OCT also appears to be strongly affected by extensional tectonics and other, more detailed, models have sought to relate features on multichannel seismic reflection profiles to tectonism (Beslier et al., 1995; Brun and Beslier, 1996; Krawczyk et al., 1996; Pickup et al., 1996). Such features are particularly clear on the time-migrated Lusigal-12 seismic reflection profile running east-west along 40°40′N (Fig. 7). Between Site 900 and the next basement high to the east (labeled FB), bright west-dipping to subhorizontal intrabasement reflectors bound landward-thickening wedge-shaped tilted blocks, overlain by landward-thickening wedges of sediments, that may or may not be interpreted as synrift (Wilson et al., 1996). These reflectors end at depths at the level of another bright reflector (H in Fig. 7), ranging between 6 and 10 km. According to Krawczyk et al's. (1996) interpretation, H cuts down from the central high with an oceanward dip, but flattens, turns, and cuts up toward the top of the Site 900 high as a band of strong, landward-dipping reflectors. The easternmost high, where Site 901 is located, appears to be a large tilted fault block capped by a seismically transparent layer of pretilting sediments. Five hundred meters west of Site 901, the seafloor is marked by a 80-m-high scarp, which is the surface expression of recent movement (differential compaction?) across a large fault structure forming the western side of the tilted block. The fault can be followed at depth as a bright reflector (L in Fig. 7) over a distance of almost 25 km. L appears on the time section to be listric, flattening at ~10.2 s TWT (12–14 km) beneath the fault block FB.

Cores from Sites 897 and 900 show that mantle upwelling and limited coeval partial melting, induced by adiabatic decompression, of the peridotites was followed by ductile shearing. The petrostructural evolution of the Site 900 gabbro suggests that it represents melt products that have been intruded/underplated near the base of the crust during extensional shear deformation in the lithosphere. In good agreement with these data, tectonic models support the existence of large-scale shear zones/detachment faults, along which the mantle and the associated granulite facies gabbro of Site 900 were exhumed. This shear deformation and the block faulting led to final breakup. These models predict the exposure of progressively deeper lithospheric levels to the west of Site 900 and possible synrift melt intrusion into the lower crust or uppermost mantle west of Site 901. They also recognize that the geometry of tectonic features, interpreted to exist on seismic profiles, includes the result of intense late stage deformation, in good agreement with the last phase of complex and intense deformation of the rocks. However, the two models proposed below differ in the mechanisms of stretching and breakup of the continental lithosphere, and therefore on the geometry of the lithospheric shear zones/detachment faults.

Mainly from the geometry of the seismic reflectors, Krawczyk et al. (1996) infer two phases of rifting (Fig. 8). The H reflector is interpreted as a large detachment fault that originally cut down to the west and accommodated top-to-the-west simple-shear motion in the upper lithosphere. This movement would have brought lower crustal and mantle rocks close to the surface to the west, similar to the model initially proposed by Wernicke (1981). During progressive extension and exhumation of the lower plate, the unloading of the lower plate to H caused this plate to bow up, as inferred for detachments faults in the western United States (Lister and Davis, 1989). This may have rendered at least part of H inactive. The current attitude of H is inferred to be the result of a second phase of faulting in which the H detachment and the lower plate to the west of it, were dissected by steeper west-dipping faults that accommodated the block faulting and the landward tilting of the fault blocks. Although the L reflector is the most pronounced structure interpreted to have been active during this later phase, other, less clearly imaged faults, which are approximately planar, bound the western flanks of basement highs FB and Site

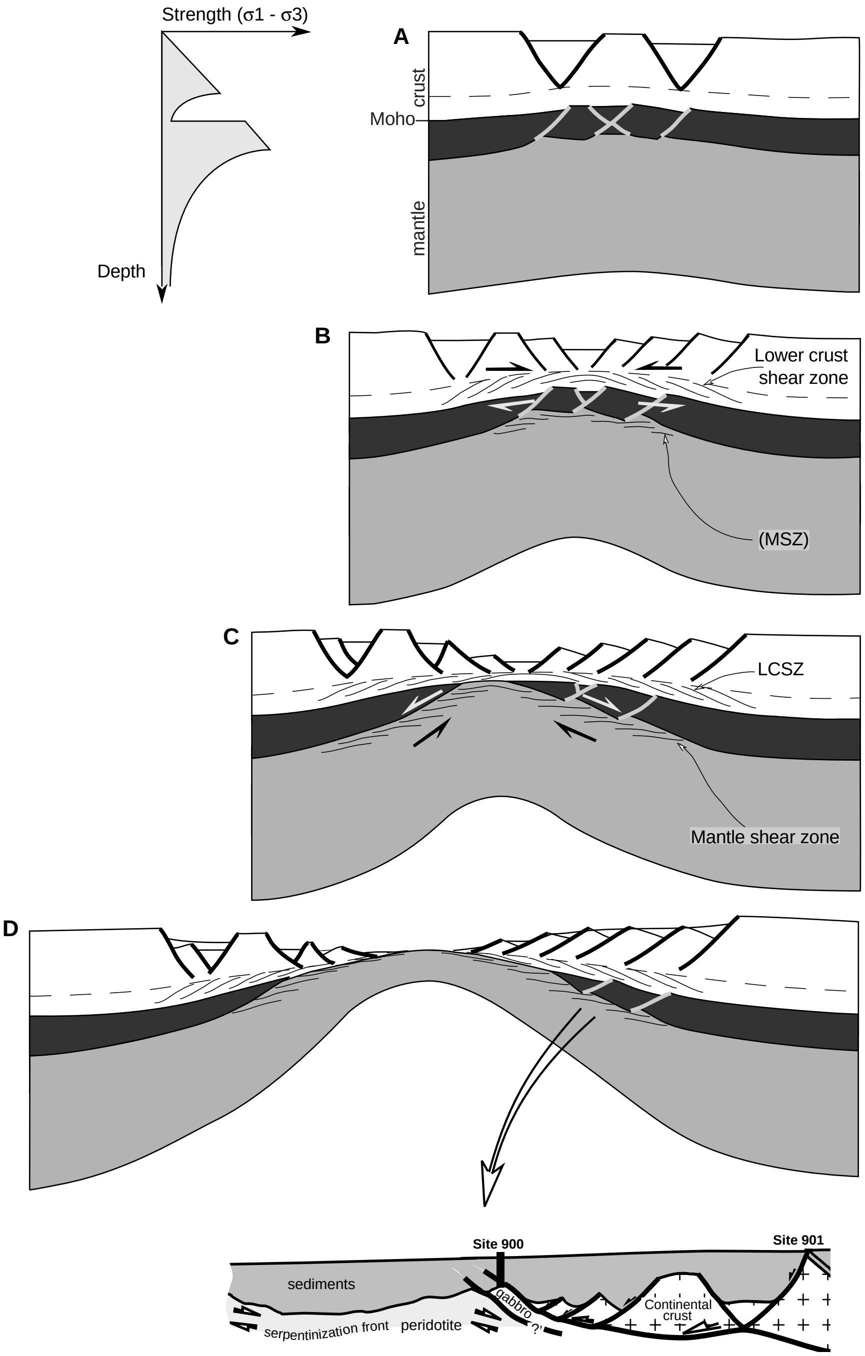

Figure 9. Progressive stretching, from rifting initiation (**A**) to continental breakup (**D**) of a four-layer brittle-ductile continental lithosphere (from Brun and Beslier, 1996). (A) Initial shear-strength profile of the continental lithosphere; intermediate stages (**B** and **C**) illustrate the development of conjugate shear zones in the ductile lower crust (LCSZ) and in the ductile lithosphere mantle (MSZ) and show the breakup of the uppermost brittle mantle (C). According to this model of lithosphere boudinage based on small-scale analogical experiments, the continental breakup leads to mantle exhumation at the conjugate passive margins and to an external symmetry but an internal asymmetry of the rifted zone. An interpretation of the Iberia Abyssal Plain margin structure is proposed in the bottom sketch.

900. The second phase of faulting caused back rotation not only of the fault blocks, but also of detachment H itself, causing it to be present on top of the basement high sampled at Site 900. Thus, Site 900 may have sampled material (the product of synrift or prerift melting) underplated to the lower crust that has been exhumed through detachment faulting and subsequently tilted by block faulting. As the block faulting appears to have dissected the entire detachment system, it is inferred that mantle thinning may have taken place over a broad zone during detachment faulting, rather than being localized down-dip of the detachment, suggesting that no single detachment developed, but rather that extension in the lower lithosphere may have been accommodated ductily or along a system of extensional structures.

Based on an interpretation of the Lusigal-12 line, on the geological data available in the Iberia Abyssal Plain, and on small-scale analogical models of lithosphere necking, Beslier et al. (1995) and Brun and Beslier (1996) proposed another model of formation of the margin and the OCT, which is extrapolated from a model proposed for the formation of the Galicia margin (Beslier and Brun, 1991; Boillot et al., 1995b; Brun and Beslier, 1996). Pickup et al. (1996) developed a model based on deeply penetrating faults and other features observed along IAM-9, a deep multichannel seismic line subparallel to, and a few tens of kilometers south of, the Lusigal-12 line, that involves lithospheric necking and the changing response to extension of the lithosphere as it passes from a plastic to a brittle regime.

In the Brun and Beslier (1996) model, as now widely accepted for a continental lithosphere with a ~35-km-thick crust and a stable geotherm, the initial lithosphere is assumed to present a four-layer type strength profile (with two relatively high-strength zones corresponding to the middle crust and the upper part of the lithosphere mantle; Fig. 9). Stretching of such a brittle-ductile multilayered lithosphere leads to nearly symmetrical necking of the whole lithosphere (pure shear) and also to the internal development of asymmetrical structures (simple shear) because of heterogeneous boudinage and/or faulting of brittle layers (upper crust and uppermost mantle). Relative movement between brittle layers caused by offsets between crust and mantle boudins is accommodated by layer-parallel shear in the ductile layers (lower crust and upper mantle) defining a set of conjugate shear zones. With increasing stretching (Fig. 9), breakup of the brittle upper mantle occurs, and lower crust and mantle shear zones become juxtaposed as the lower ductile mantle layer rises up through the breakup zone. Localization of stretching in the upper part of the lithosphere mantle produces asymmetry in the lithosphere, even if no previously inherited asymmetry was present. A system of tilted blocks develops in the brittle crust over the ductile crustal shear zone. Extreme crustal thinning in the central part of the necked lithosphere results in mantle exhumation. Accordingly, the exhumation of lithospheric mantle is the result of bulk pure shear at lithospheric scale. Shearing of exhumed mantle rocks does not correspond to detachment faults cross-cutting the whole lithosphere at the onset of rifting but results from heterogeneous stretching and boudinage of the high-strength uppermost mantle. Mantle rocks that appear at the extremes of the two conjugate passive margins are those that have been sheared along mantle shear zones. The pre-existing sub-continental Moho passes laterally toward the breakup zone into a newly formed Moho, which vertically juxtaposes the ductile mantle and the lower crust, or even the upper brittle crust (Fig. 9). In the Iberia Abyssal Plain, the H reflector is then interpreted as a major synrift tectonic contact that results from intralithospheric shear zones that brought into contact the (upper?) crust and the underplated gabbros produced by the partial melting of the mantle. It implies that the whole area located between Site 900 and the oceanic crust is a tectonic window opened on deep lithospheric levels (serpentinized peridotite and associated underplated gabbros). The complex late deformation in greenschist facies conditions observed in both peridotite and gabbro and the considerable width (130 km) of the OCT suggest that the mantle and associated mafic rocks can be stretched over a large width after the continental breakup is completed and before a mature accretion axis has become established.

Although currently it is hard to explain the lineated magnetic anomalies in the OCT (Whitmarsh and Miles, 1995) by the above tectonic models, this results more from unfamiliarity with the geophysical characteristics that such an unusual zone would possess than from any specific contrary argument.

Even though one may argue about the details of the above interpretations, and how some of these models could be combined to better explain both continental breakup and OCT formation, it is certainly possible that a major tectonic contact approaches the basement surface near Site 900. Further drilling took place during Leg 173 in order to clarify the nature and petrostructural evolution of the rocks involved in this major synrift extensional structure and the kinematics of the deformation, to investigate the nature of the basement in the transition zone, to seek evidence of synrift magmatism, and to determine the nature and evolution of the adjacent continental and oceanic crusts.

SCIENTIFIC DRILLING OBJECTIVES

As outlined above, Leg 149 largely succeeded in determining the oceanward and landward bounds of the OCT in the southern Iberia Abyssal Plain. The principal problem that remained was to refine knowledge of the nature and evolution of the basement rocks and their relationships within the OCT itself, both for its intrinsic relevance to the general problem of rift-to-drift processes and to test aspects of our best working models of these tectonic and igneous processes at the west Iberia nonvolcanic margin. Sites were therefore chosen for Leg 173 (Figs. 1, 3–5) with the following objectives:

1. To sample acoustic basement, principally within the OCT, to characterize those tectonic and magmatic processes that dominate the transition from continental to oceanic crust in space and time (see Sites 901, IBERIA-07B [which became 1069], IBERIA-08B [1065], IBERIA-09A/09B [1067/1066], IBERIA-09C [1068], and IBERIA-10A [1070]).
2. To determine the role of simple shear deformation in the evolution of the margin. This was to be done by drilling through a major synrift tectonic contact on the east side of the high on which Site 900 has already been drilled. Figure 10, in which the H reflector can be traced almost continuously to the top of the fault block, shows this possibility particularly clearly. This interpretation is supported by the retrograde evolution and intense deformation of the Site 900 gabbros from granulite to greenschist facies conditions. The seismic image at Site 900 suggests that the site could be located very close to the place where H intersects the acoustic basement surface, a result of late normal faulting along the west-dipping steep fault bounding the western flank of the basement high. By offsetting some hundreds of meters to the east of Site 900, however, the possibility existed to drill through the complete shear zone/detachment. This latter site (two alternative sites, IBERIA-09A/09B [1067/1066] were proposed) was also expected to enable the mode and kinematics of the deformation along the tectonic contact to be determined and the lateral extent of the Site 900 mafic rocks to be assessed. Another site (IBERIA-07B [1069]) was drilled on the westernmost basement high associated with a westward-dipping listric normal fault, to test the prediction that continental breakup led to the exposure of progressively deeper lithospheric levels (low level crust or even uppermost mantle) westward in the OCT (see Sites IBERIA-07B [1069], IBERIA-09A/09B [1067/1066], and IBERIA-09C [1068]).
3. To determine the role and extent of synrift magmatism in the OCT basement, which is inferred to exist from the new magnetic anomaly chart and Site 900 cores. Using isotopes, to determine the petrogenetic origin and dates of original crystallization and subsequent metamorphism of igneous rocks (see Sites 901, IBERIA-07B [1069], IBERIA-08B [1065], IBERIA-09A/09B [1067/1066], and IBERIA-09C [1068]).
4. To sample basement beneath Site 901 or Site IBERIA-08B [1065] to confirm predictions of the existence of continental crust there, to determine the approximate level in the crust from which it came, and thereby to set an unequivocal landward limit to the OCT (see Sites 901, IBERIA-08B [1065]).
5. To sample the probably atypical early-formed oceanic crust. This remains unsampled in the Iberia Abyssal Plain and its presence and location are inferred only by geophysical observations. Samples from this site were expected to provide definitive evidence of the oceanic nature of the crust

immediately (20 km) west of the peridotite ridge and, by biostratigraphic or isotopic dating, may enable the seafloor-spreading model to be verified. Samples were also expected to yield the possibly unusual chemistry of the thin crust formed by the earliest magma-starved seafloor spreading and provide valuable petrological information about initial melt production (cf. Site 900 gabbro), following continental breakup at a nonvolcanic margin (see Site IBERIA-10A [1070]).

REFERENCES

Agrinier, P., Cornen, G., and Beslier, M.-O., 1996. Mineralogical and oxygen isotopic features of serpentinites recovered from the ocean/continent transition in the Iberia Abyssal Plain. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 541–552.

Agrinier, P., Mével, C., and Girardeau, J., 1988. Hydrothermal alteration of the peridotites cored at the ocean/continent boundary of the Iberian Margin: petrologic and stable isotope evidence. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 225–234.

Allemand, P., and Brun, J.P., 1991. Width of continental rifts and rheological layering of the lithosphere. *Tectonophysics,* 188:63–69.

Auzende, J.M., Charvet, J., Le Lann, A., Le Pichon, X., Monteiro, J.H., Nicolas, A., Olivet, J.L., and Ribeiro, A., 1979. Le Banc de Gorringe: résultats de la campagne CYAGOR (août 1977). *Bull. Soc. Geol. Fr.,* 21:545–556.

Auzende, J.-M., Olivet, J.-L., Charvet, J., Le Lann, A., Le Pichon, X., Monteiro, J.H., Nicolas, A. and Ribeiro, A., 1978. Sampling and observation of oceanic mantle and crust on Gorringe Bank. *Nature,* 273:45–49.

Bassi, G., 1991. Factors controlling the style of continental rifting: insights from numerical modelling. *Earth Planet. Sci. Lett.,* 105:430–452.

————, 1995. Relative importance of strain rate and rheology for the mode of continental extension. *Geophys. J. Int.,* 122:195–210.

Beslier, M.O., Ask, M., and Boillot, G., 1993. Ocean-continent boundary in the Iberia Abyssal Plain from multichannel seismic data. *Tectonophysics,* 218:383–393.

Beslier, M.O., Bitri, A., and Boillot, G., 1995. Structure de la transition continent-océan d'une marge passive: sismique réflexion multitrace dans la Plaine Abyssale Ibérique (Portugal). *C. R. Acad. Sci. Ser. 2a,* 320:969–976.

Beslier, M.O., and Brun, J.-P., 1991. Boudinage de la lithosphère et formation des marges passives. *C. R. Acad. Sci. Ser. 2,* 313:951–958.

Beslier, M.-O., Cornen, G., and Girardeau, J., 1996. Tectono-metamorphic evolution of peridotites from the ocean/continent transition of the Iberia Abyssal Plain margin. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 397–412.

Beslier, M.O., Girardeau, J., and Boillot, G., 1990. Kinematics of peridote emplacement during North Atlantic continental rifting, Galicia, northwestern Spain. *Tectonophysics,* 184:321–343.

Boillot, G., Agrinier, P., Beslier, M.-O., Cornen, G., Froitzheim, N., Gardien, V., Girardeau, J., Gil-Ibarguchi, J.-I., Kornprobst, J., Moullade, M., Schaerer, U., and Vanney, J.-R., 1995a. A lithospheric syn-rift shear zone at the ocean-continent transition: preliminary results of the GALINAUTE II cruise (Nautile dives on the Galicia Bank, Spain). *C. R. Acad. Sci. Ser 2a,* 321:1171–1178.

Boillot, G., Auxiètre, J., Dunand, J., Dupeuble, P., and Mauffret, A., 1979. The northwestern Iberian margin: a Cretaceous passive margin deformed during Eocene. *In* Talwani, M., Hay, W., and Ryan, W.B.F. (Eds.), *Deep Drilling Results in the Atlantic Ocean: Continental Margin and Paleoenvironment.* Am. Geophys. Union, Maurice Ewing Ser., 3:138–153.

Boillot, G., Beslier, M.O., and Comas, M., 1992. Seismic image of undercrusted serpentinite beneath a rifted margin. *Terra Nova,* 4:25–33.

Boillot, G., Beslier, M.O., Krawczyk, C.M., Rappin, D., and Reston, T.J., 1995b. The formation of passive margins: constraints from the crustal structure and segmentation of the deep Galicia margin (Spain). *In* Scrutton, R.A., Stocker, M.S., Shimmield, G.B., and Tudhope, A.W. (Eds.), *The Tectonics, Sedimentation and Palaeoceanography of the North Atlantic Region.* Geol. Soc. Spec. Publ. London, 90:71–91.

Boillot, G., Comas, M.C., Girardeau, J., Kornprobst, J., Loreau, J.-P., Malod, J., Mougenot, D., and Moullade, M., 1988a. Preliminary results of the Galinaute cruise: dives of the submersible *Nautile* on the Western Galicia Margin, Spain. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 37–51.

Boillot, G., Féraud, G., Recq, M., and Girardeau, J., 1989. "Undercrusting" by serpentinite beneath rifted margins: the example of the west Galicia margin (Spain). *Nature,* 341:523–525.

Boillot, G., Girardeau, J., and Kornprobst, J., 1988b. Rifting of the Galicia Margin: crustal thinning and emplacement of mantle rocks on the seafloor. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 741–756.

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

Boillot, G., Recq, M., Winterer, E.L., Meyer, A.W., Applegate, J., Baltuck, M., Bergen, J.A., Comas, M.C., Davies, T.A., Dunham, K., Evans, C.A., Girardeau, J., Goldberg, D.G., Haggerty, J., Jansa, L.F., Johnson, J.A., Kasahara, J., Loreau, J.-P., Luna-Sierra, E., Moullade, M., Ogg, J., Sarti, M., Thurow, J., and Williamson, M.A., 1987. Tectonic denudation of the upper mantle along passive margins: a model based on drilling results (ODP Leg 103, western Galicia margin, Spain). *Tectonophysics,* 132:335–342.

Boillot, G., Winterer, E.L., et al., 1988. *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program).

Boillot, G., Winterer, E.L., Meyer, A.W., et al., 1987. *Proc. ODP, Init. Repts.,* 103: College Station, TX (Ocean Drilling Program).

Bonatti, E., 1985. Punctiform initiation of seafloor spreading in the Red Sea during transition from a continental to an oceanic rift. *Nature,* 316:33–37.

Bown, J.W., and White, R.S., 1995. The effect of finite extension rate on melt generation at continental rifts. *J. Geophys. Res.,* 100:18011–18030.

Brun, J.P., and Beslier, M.-O., 1996. Mantle exhumation at passive margins. *Earth Planet. Sci. Lett.,* 142:161–173.

Brun, J.-P., Sokoutis, D., and van den Driessche, J., 1994. Analogue modeling of detachment fault systems and core complexes. *Geology,* 22:319–322.

Capdevila, R., and Mougenot, D., 1988. Pre-Mesozoic basement of the western Iberian continental margin and its place in the Variscan Belt. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 3–12.

Coleman, R.G., 1993. *Geological Evolution of the Red Sea:* New York (Oxford Univ. Press).

Comas, M.C., Sánchez-Gómez, M., Cornen, G., and de Kaenel, E., 1996. Serpentinized peridotite breccia and olistrostrome on basement highs of the Iberia Abyssal Plain: implications for tectonic margin evolution. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 577–591.

Cornen, G., 1982. Petrology of the alkaline volcanism of Gorringe Bank (southwest Portugal). *Mar. Geol.,* 47:101–130.

Cornen, G., Beslier M.-O., and Girardeau, J., 1996a. Petrologic characteristics of the ultramafic rocks from the ocean/continent transition in the Iberia Abyssal Plain. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 377–395.

Cornen, G., Beslier, M.-O., and Girardeau, J., 1996b. Petrology of the mafic rocks cored in the Iberia Abyssal Plain. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 449–469.

Cyagor II Group (Auzende, J.M., Ceuleneer, G., Cornen, G., Juteau, T., Lagabrielle, Y., Lensch, G., Mével, C., Nicolas, A., Prichard, H., Ribeiro, A., Ruellan, E., and Vanney, J.R.), 1984. Intraoceanic tectonism on the Gorringe Bank: observations by submersible. *In* Gass, I.G., Lippard S.J., and Shelton, A.W. (Eds.), *Ophiolites and Oceanic Lithosphere.* Geol. Soc. London, 113–120.

Cyagor Group, 1977. Le Banc de Gorringe (S-O du Portugal): un fragment de croute et manteau océanique reconnu par submersible. *C.R. Acad. Sci. Ser. D,* 285:1403–1406.

Dean, S.M., Minshull, T.A., Whitmarsh, R.B., Louden, K., and Chian, D., 1996. A seismic study of the ocean-continent transition in the Iberia Abyssal Plain. *Eos,* 77:F668.

de Charpal, O., Guennoc, P., Montadert, L., and Roberts, D.G., 1978. Rifting, crustal attenuation and subsidence in the Bay of Biscay. *Nature,* 275:706–711.

de Graciansky, P.C., Poag, C.W., et al., 1985. *Init. Repts. DSDP,* 80: Washington (U.S. Govt. Printing Office).

Dupeuble, P.-A., Boillot, G., and Mougenot, D., 1987. Upper Jurassic-lowest Cretaceous limestones dredged from the Western Galicia Margin. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Init. Repts.,* 103: College Station, TX (Ocean Drilling Program), 99–106.

Dupeuble, P.A., Réhault, J.P., Auxiètre, J.L., Dunand, J.P., and Pastouret, L., 1976. Résultats de dragages et essai de stratigraphie des bancs de Galice et des montagnes de Porto et de Vigo (marge occidentale iberique). *Mar. Geol.,* 22:37–49.

Evans, C.A., and Girardeau, J., 1988. Galicia Margin peridotites: undepleted abyssal peridotites from the North Atlantic. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 195–207.

Féraud, G., Beslier, M.-O., and Cornen, G., 1996. ^{40}Ar/^{39}Ar dating of gabbros from the ocean/continent transition of the western Iberia Margin: preliminary results. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 489–495.

Féraud, G., York, D., Mével, C., Cornen, G., Hall, C.M., and Auzende, J.M., 1986. Additional ^{40}Ar-^{39}Ar dating of the basement and the alkaline volcanism of the Gorringe Bank (Atlantic Ocean). *Earth Planet. Sci. Lett.,* 79:255–269.

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

Gibson, I.L., Milliken, K.L., and Morgan, J.K., 1996. Serpentinite-breccia landslide deposits generated during crustal extension at the Iberia Margin. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 571–575.

Girardeau, J., Cornen, G., Agrinier, P., Beslier, M.O., Dubuisson, G., Le Gall, B., Monnier, C., Pinheiro, L., Ribeiro, A., and Whitechurch, H., in press. Résultats des plongées du Nautile sur le Banc de Gorringe. *C.R. Acad. Sci. Ser. 2.*

Girardeau, J., Evans, C.A., and Beslier, M.-O., 1988. Structural analysis of plagioclase-bearing peridotites emplaced at the end of continental rifting: Hole 637A, ODP Leg 103 on the Galicia Margin. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 209–223.

Gradstein, F.M., Agterberg, F.P., Ogg, J.G., Hardenbol, J., van Veen, P., Thierry, J., and Huang, Z., 1995. A Triassic, Jurassic and Cretaceous time scale. *In* Berggren, W.A., Kent, D.V., and Aubry, M.P. (Eds.), *Geochronology, Time Scales and Global Stratigraphic Correlation.* Spec. Publ.—Soc. Econ. Paleontol. Mineral., 54:95–128.

Groupe Galice, 1979. The continental margin off Galicia and Portugal: acoustical stratigraphy, dredge stratigraphy, and structural evolution. *In* Sibuet, J.-C., Ryan, W.B.F., et al., *Init. Repts. DSDP,* 47 (Pt. 2): Washington (U.S. Govt. Printing Office), 633–662.

Harry, D.L., and Sawyer, D.S., 1992. A dynamic model of lithospheric extension in the Baltimore Canyon Trough region. *Tectonics,* 11:420–436.

Hinz, K., Winterer, E.L., et al., 1984. *Init. Repts. DSDP,* 79: Washington (U.S. Govt. Printing Office).

Hoffman, H.-J., and Reston, T.J., 1992. Nature of the S reflector beneath the Galicia Banks rifted margin: preliminary results from pre-stack depth migration. *Geology,* 20:1091–1094.

Horsefield, S.J., 1992. Crustal structure across the continent-ocean boundary [Ph.D. thesis]. Cambridge Univ.

Klitgord, K.D., and Schouten, H., 1986. Plate kinematics of the central Atlantic. *In* Vogt, P.R., and Tucholke, B.E. (Eds.), *The Geology of North America* (Vol. M): *The Western North Atlantic Region.* Geol. Soc. Am., 351–378.

Krawczyk, C.M., Reston, T.J., Beslier, M.-O., and Boillot, G., 1996. Evidence for detachment tectonics on the Iberia Abyssal Plain rifted margin. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 603–615.

Lemoine, M., Tricart, P., and Boillot, G., 1987. Ultramafic and gabbroic ocean floor of the Ligurian Tethys (Alps, Corsica, Apennines): in search of a genetic model. *Geology,* 15:622–625.

Le Pichon, X., and Barbier, F., 1987. Passive margin formation by low-angle faulting within the upper crust: the northern Bay of Biscay margin. *Tectonics,* 6:133–150.

Le Pichon, X., Sibuet, J.C., and Francheteau, J., 1977. The fit of the continents around the North Atlantic Ocean. *Tectonophysics,* 38:169–209.

Lister, G.S., and Davis, G.A., 1989. The origin of metamorphic core complexes and detachment faults formed during Tertiary continental extension in the Northern Colorado River region, U.S.A. *J. Struct. Geol.,* 11:65–94.

Lister, G.S., Etheridge, M.A., and Symonds, P.A., 1986. Detachment faulting and evolution of passive continental margins. *Geology,* 14:246–250.

Louden, K., Sibuet, J.C., and Foucher, J.P., 1989. Heat flow anomalies across Goban Spur and Galicia Bank continental margins (NW Atlantic). *Eos,* 70:1328.

Malod, J.A., and Mauffret, A., 1990. Iberian plate motion during the Mesozoic. *In* Boillot, G., and Fontbote, J.M. (Eds.), *Alpine Evolution of Iberia and its Continental Margins.* Tectonophysics, 184:261–278.

Malod, J.A., Murillas, J., Kornprobst, J., and Boillot, G., 1993. Oceanic lithosphere at the edge of a Cenozoic active continental margin (northwestern slope of Galicia Bank, Spain). *Tectonophysics,* 221:195–206.

Mamet, B., Comas, M.C., and Boillot, G., 1991. Late Paleozoic basin on the West Galicia Atlantic margin. *Geology,* 19:738–741.

Martins, L.F., 1991. Activitade igneo Mesozoica em Portugal [Ph.D. dissert.]. Univ. of Lisbon, Portugal.

Masson, D.G., Cartwright, J.A., Pinheiro, L.M., Whitmarsh, R.B., Beslier, M.-O., and Roeser, H., 1994. Compressional deformation at the ocean-continent transition in the NE Atlantic. *J. Geol. Soc. London,* 151:607–613.

Masson, D.G., and Miles, P.R., 1984. Mesozoic seafloor spreading between Iberia, Europe and North America. *Mar. Geol.,* 56:279–287.

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

————, 1988. Seismic stratigraphy off Galicia. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 13–30.

Mauffret, A., Mougenot, D., Miles, P.R., and Malod, J.A., 1989. Cenozoic deformation and Mesozoic abandoned spreading centre in the Tagus abyssal plain (west of Portugal): results of a multichannel seismic survey. *Can. J. Earth Sci.,* 26:1101–1123.

McKenzie, D., 1978. Some remarks on the development of sedimentary basins. *Earth Planet. Sci. Lett.,* 40:25–32.

Mével, C., 1988. Metamorphism of oceanic layer 3, Gorringe Bank, eastern Atlantic. *Contrib. Mineral. Petrol.,* 100:496–509.

Miles, P.R., Verhoef, J., and Macnab, R., 1996. Compilation of magnetic anomaly chart west of Iberia. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 659–663.

Milliken, K.L., Lynch, F.L., and Seifert, K.E., 1996. Marine weathering of serpentinites and serpentinite breccias, Site 897 and 899, Iberia Abyssal Plain. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 529–540.

Montadert, L., de Charpel, O., Roberts, D., Guennoc, P., and Sibuet, J.-C., 1979a. Northeast Atlantic passive continental margins: rifting and subsidence processes. *In* Talwani, M., Hay, W., and Ryan, W.B.F. (Eds.), *Deep Drilling Results in the Atlantic Ocean: Continental Margins and Paleoenvironment.* Am. Geophys. Union, Maurice Ewing Ser., 3:154–186.

Montadert, L., Roberts, D.G., et al., 1979b. *Init. Repts. DSDP,* 48: Washington (U.S. Govt. Printing Office).

Murillas, J., Mougenot, D., Boillot, G., Comas, M.C., Banda, E., and Mauffret, A., 1990. Structure and evolution of the Galicia Interior basin (Atlantic western Iberian continental margin). *Tectonophysics,* 184:297–319.

Ogg, J.G., 1988. Early Cretaceous and Tithonian magnetostratigraphy of the Galicia Margin (Ocean Drilling Program Leg 103). *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 659–682.

Olivet, J.L., 1996. La cinématique de la plaque ibérique. *Bull. Cent. Rech. Explor.-Prod. Elf-Aquitaine,* 20:131–195.

Olivet, J.L., Bonnin, J., Beuzart, P., and Auzende, J.M., 1984. Cinématique de l'Atlantique nord et central. *Publ. Cent. Natl Exploit. Oceans, Rapp. Sci. Tech.,* 54:108–112.

Pickup, S.L.B., Whitmarsh, R.B., Fowler, C.M.R., and Reston, T.J., 1996. Insight into the nature of the ocean-continent transition off West Iberia from a deep multichannel seismic reflection profile. *Geology,* 24:1079–1082.

Pinheiro, L.M., Whitmarsh, R.B., and Miles, P.R., 1992. The ocean-continent boundary off the western continental margin of Iberia, II. Crustal structure in the Tagus Abyssal Plain. *Geophys. J.,* 109:106–124.

Pinheiro, L.M., Wilson, R.C.L., Pena dos Reis, R., Whitmarsh, R.B., and Ribeiro, A., 1996. The western Iberia Margin: a geophysical and geological overview. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 3–23.

Prichard, H.M., and Cann, J.R., 1982. Petrology and mineralogy of dredged gabbro from Gettysburg Bank, Eastern Atlantic. *Contrib. Mineral. Petrol.,* 79:46–55.

Reston, T.J., Krawczyk, C.M., and Hoffmann, H.J., 1995. Detachment tectonics during Atlantic rifting: analysis and interpretation of the S reflection, the west Galicia margin. *In* Scrutton, R.A. (Ed.), *Tectonics of the North Atlantic Region.* Geol. Soc. Spec. Publ. London.

Reston, T.J., Krawczyk, C.M., and Klaeschen, D., 1996. The S reflector west of Galicia (Spain). Evidence for detachment faulting during continental breakup from prestack depth migration. *J. Geophys. Res.,* 101:8075–8091.

Ribeiro, A., Antunes, M.T., Ferreira, M.P., Rocha, R.B., Soares, A.F., Zbyszewski, G., Moitinho de Almeida, F., Carbalho, D., and Monteiro, J.H., 1979. *Introduction à la Géologie Generale du Portugal:* Lisboa (Ed. Serv. Geol. de Portugal).

Shipboard Scientific Party, 1973. Gorringe Bank: Site 120. *In* Ryan, W.B.F., and Hsü, K.J., et al., *Init. Repts. DSDP.,* 13: Washington (U.S. Govt. Printing Office), 19–41.

Sawyer, D.S., 1994. The case for slow-spreading oceanic crust landward of the peridotite ridge in the Iberia Abyssal Plain. *Eos,* 75:616.

Sawyer, D.S., Whitmarsh, R.B., Klaus, A., et al., 1994. *Proc. ODP, Init. Repts.,* 149: College Station, TX (Ocean Drilling Program).

Schärer, U., Kornprobst, J., Beslier, M.O., Boillot, G., and Girardeau, J., 1995. Gabbro and related rock emplacement beneath rifting continental crust: U-Pb geochronological and geochemical constraints for the Galicia passive margin (Spain). *Earth Planet. Sci. Lett.,* 130:187–200.

Seifert, K., and Brunotte, D., 1996. Geochemistry of weathered mid-ocean ridge basalt and diabase clasts from Hole 899B. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 497–515.

Seifert, K., Gibson, I., Weis, D., and Brunotte, D., 1996. Geochemistry of metamorphosed cumulate gabbros from Hole 900A, Iberia Abyssal Plain. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 471–488.

Seifert, K.E., Cheng-Wen, C., and Brunotte, D.A., 1997. Evidence from ODP Leg 149 mafic igneous rocks for oceanic crust in the Iberia Abyssal Plain ocean-continent transition zone. *J. Geophys. Res.,* 102:7915–7928.

Serri, G., Hébert, R., and Hekinian, R., 1988. Petrology of a plagioclase-bearing olivine websterite from the Gorringe Bank (northeastern Atlantic Ocean). *Can. J. Earth Sci.,* 25:557–569.

Sibuet, J.-C., Louvel, V., Whitmarsh, R.B., White, R.S., Horsefield, S.J., Sichler, B., Leon, P., and Recq, M., 1995. Constraints on rifting processes from refraction and deep-tow magnetic data: the example of the Galicia continental margin (West Iberia). *In* Banda, E., Torné, M., and Talwani, M. (Eds.), *Rifted Ocean-Continent Boundaries:* Amsterdam (Kluwer), 197–218.

Sibuet, J.-C., Mazé, J.-P., Amortila, P., and Le Pichon, X., 1987. Physiography and structure of the Western Iberian Continental Margin off Galicia, from Sea Beam and seismic data. *In* Boillot, G., Winterer, E.L., Meyer, A.W., et al., *Proc. ODP, Init. Repts.,* 103: College Station, TX (Ocean Drilling Program), 77–98.

Sibuet, J.-C., Ryan, W.B.F., et al., 1979. *Init. Repts. DSDP,* 47 (Pt. 2): Washington (U.S. Govt. Printing Office).

Sibuet, J.-C., and Srivastava, S., 1994. Rifting consequences of three plate separation. *Geophys. Res. Lett.,* 21:521–524.

Srivastava, S.P., Roest, W.R., Kovacs, L.C., Oakey, G., Lévesque, S., Verhoef, J., and Macnab, R., 1990a. Motion of Iberia since the Late Jurassic: results from detailed aeromagnetic measurements in the Newfoundland Basin. *Tectonophysics,* 184:229–260.

Srivastava, S.P., Schouten, H., Roest, W.R., Klitgord, K.D., Kovacs, L.C., Verhoef, J., and Macnab, R., 1990b. Iberian plate kinematics: a jumping plate boundary between Eurasia and Africa. *Nature,* 344:756–759.

Srivastava, S.P., and Verhoef, J., 1992. Evolution of Mesozoic sedimentary basins around the North Central Atlantic: a preliminary plate kinematic solution. *In* Parnell, J. (Ed.), *Basins of the Atlantic Seaboard: Petroleum Geology, Sedimentology and Basin Evolution.* Geol. Soc. Spec. Publ. London, 62:397–420.

Srivastava, S.P., Verhoef, J., and Macnab, R., 1988. Results from a detailed aeromagnetic survey across the northeast Newfoundland margin, Part I. Spreading anomalies and relationship between magnetic anomalies and the ocean-continent boundary. *Mar. Pet. Geol.,* 5:306–323.

Welsink, H.J., Srivastava, S.P., and Tankard, A.J., 1989. Basin architecture of the Newfoundland continental margins and its relationship to ocean crust fabric during extension. *In* Tankard, A.J., and Balkwill, H.R. (Eds.), *Extensional Tectonics and Stratigraphy of the North Atlantic Margins.* AAPG Mem., 46:197–213.

Wernicke, B., 1981. Low-angle normal faults in the Basin and Range Province: nappe tectonics in an extending orogen. *Nature,* 291:645–647.

————, 1985. Uniform sense normal simple shear of the continental crust. *Can. J. Earth Sci.,* 22:108–125.

Wernicke, B., and Burchfiel, B.C., 1982. Modes of extensional tectonics. *J. Struct. Geol.,* 4:105–115.

White, R., and McKenzie, D., 1989. Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. *J. Geophys. Res.,* 94:7685–7729.

Whitmarsh, R.B., and Miles, P.R., 1995. Models of the development of the West Iberia rifted continental margin at 40°30′N deduced from surface and deep-tow magnetic anomalies. *J. Geophys. Res.,* 100:3789–3806.

Whitmarsh, R.B., Miles, P.R., and Mauffret, A., 1990. The ocean-continent boundary off the western continental margin of Iberia, I. Crustal structure at 40°30′N. *Geophys. J. Int.,* 103:509–531.

Whitmarsh, R.B., Miles, P.R., Sibuet, J.-C., and Louvel, V., 1996a. Geological and geophysical implications of deep-tow magnetometer observations near Sites 897, 898, 899, 900, and 901 on the west Iberia continental margin. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 665–674.

Whitmarsh, R.B., Pinheiro, L.M., Miles, P.R., Recq, M., and Sibuet, J.C., 1993. Thin crust at the western Iberia ocean-continent transition and ophiolites. *Tectonics,* 12:1230–1239.

Whitmarsh, R.B., and Sawyer, D.S., 1996. The ocean/continent transition beneath the Iberia Abyssal Plain and continental-rifting to seafloor-spreading processes. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 713–733.

Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), 1996. *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program).

Whitmarsh, R.B., White, R.S., Horsefield, S.J., Sibuet, J.-C., Recq, M., and Louvel., V., 1996b. The ocean-continent boundary off the western continental margin of Iberia: crustal structure west of Galicia Bank. *J. Geophys. Res.,* 101:28291–28314.

Wilson, R.C.L., Hiscott, R.N., Willis, M.G., and Gradstein, F.M., 1989. The Lusitanian Basin of west-central Portugal: Mesozoic and Tertiary tectonic, stratigraphic and subsidence history. *In* Tankard, A.J., and Balkwill, H.R. (Eds.), *Extensional Tectonics and Stratigraphy of the North Atlantic Margins.* AAPG Mem., 46:341–361.

Wilson, R.C.L., Sawyer, D.S., Whitmarsh, R.B., Zerong, J., and Carbonell, J., 1996. Seismic stratigraphy and tectonic history of the Iberia Abyssal Plain. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 617–633.

Zhao, X., 1996. Magnetic signatures of peridotite rocks from Sites 897 and 899 and their implications. *In* Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), *Proc. ODP, Sci. Results,* 149: College Station, TX (Ocean Drilling Program), 431–446.

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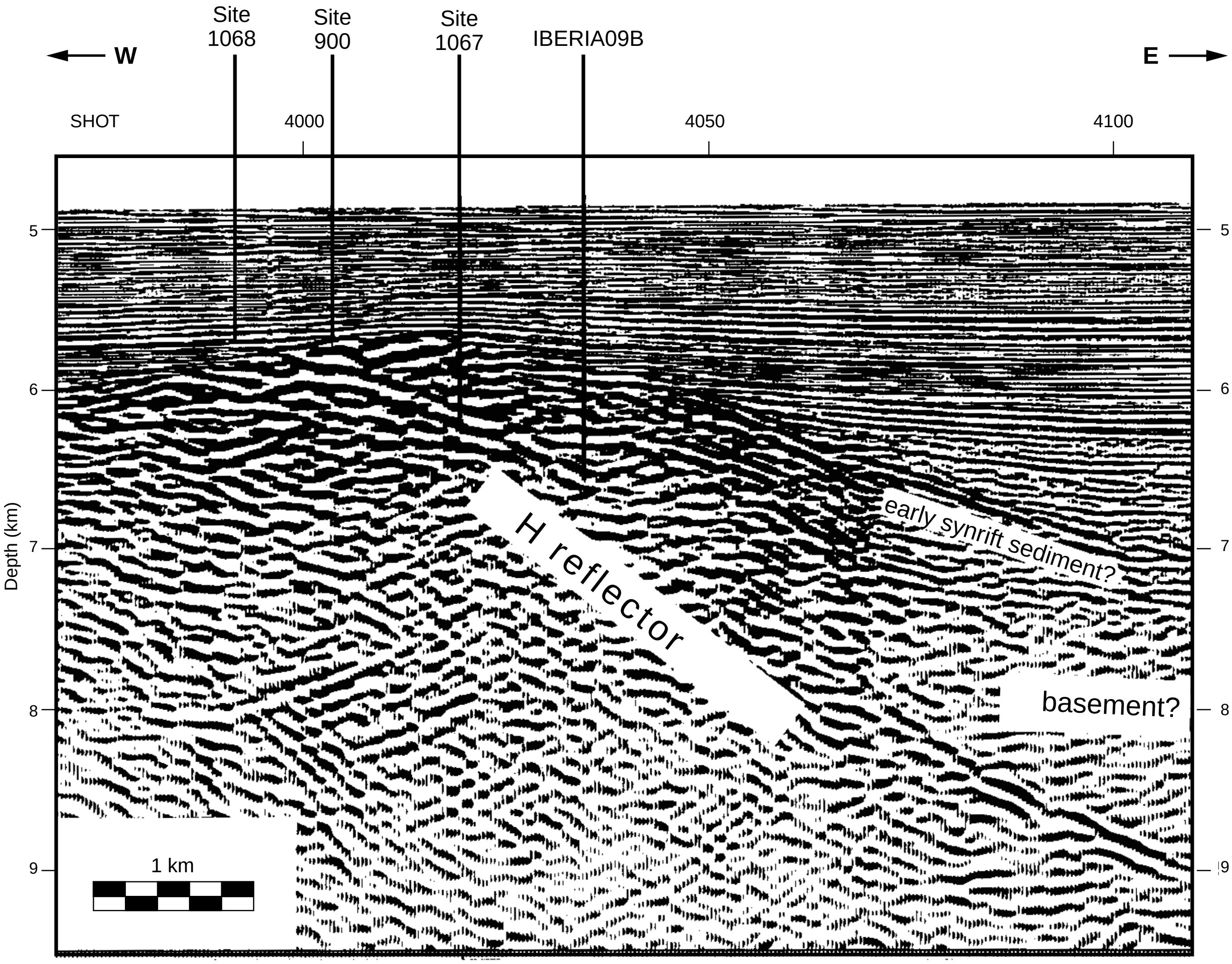

Figure 10. Prestack depth-migrated image of profile Lusigal-12 near Sites 900, 1067, and 1068 (modified from Krawczyk et al, 1996). A probable major synrift tectonic contact (the H reflector) appears to surface here: Site 900 is interpreted, on the evidence of intense ductile shearing of the gabbro under granulite facies conditions (dynamic recrystallization at a depth of ~13 km, which ended during the rifting) to have been drilled at or very close to the contact. Two alternate sites east of Site 900 were proposed, one to penetrate supposed early synrift sediment and into the tectonic contact, and the other to penetrate the same synrift sediment, through the prerift sediment or basement overlying the tectonic contact, and finally through the detachment itself. Eventually only Site 1967 and the new Site 1068 were drilled west of Site 900 during Leg 173.