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47. THE OCEAN/CONTINENT TRANSITION BENEATH THE IBERIA ABYSSAL PLAIN AND CONTINENTAL-RIFTING TO SEAFLOOR-SPREADING PROCESSES[1]

R.B. Whitmarsh[2] and Dale S. Sawyer[3]

ABSTRACT

The west Iberia continental margin is a nonvolcanic rifted margin which, following rifting that began in the Late Triassic, broke away from Newfoundland in the Early Cretaceous as rifting propagated from south to north. The ocean/continent transition off Newfoundland seems to occur ~50 km seaward of the shelf edge near the base of the continental slope. Off west Iberia, the ocean/continent transition is defined by an 80- to 130-km-wide region beneath the Iberia Abyssal Plain between the most oceanward-tilted basement fault blocks (continental crust?) and a 300-km-long narrow peridotite ridge parallel to the margin. Leg 149 drilled a west-to-east transect of holes across the ocean/continent transition, and three of these reached acoustic basement, beginning at the peridotite ridge.

Sites 897 (over the peridotite ridge) and 899 (19 km further east) sampled serpentinized peridotites with similar, but not identical, petrologies that experienced a similar history of exhumation from the deep mantle to the surface. They crystallized at 1170°-1230°C, and this was followed by ductile shear deformation at 880°-1000°C. After limited partial melting, secondary minerals crystallized at about 30 km depth. Mylonitization, and then low-temperature deformation and serpentinization, followed as the rocks were exhumed at the seabed. One possibly important, but unexplained, difference between the two sites is the fivefold greater strength of remanent magnetization of the Site 899 cores, which is reflected in the amplitudes of magnetic anomalies observed over the sites. The peridotites are neither clearly subcontinental nor suboceanic. Further landward, Site 900 sampled a flasered cumulate gabbro basement. The rare-earth element (REE) patterns of the gabbros are ambiguous and have been matched by different authors to both island arc and N-MORB basalts. The light REE patterns fit a transitional MORB parent magma whereas Nd and Pb isotope ratios strongly suggest a MORB parent. The gabbro therefore likely formed in a magma chamber from a melt that was little contaminated, if at all, by continental crust. Traces of the primary mineralogy indicate dynamic crystallization at depths of at least 13 km and temperatures typical of granulite facies conditions. During exhumation to the seabed the rocks underwent retrograde metamorphism at 280 ± 20°C followed by crystallization of some plagioclases at 136.4 ± 0.3 Ma. Late Barremian- to late Aptian-age debris-flow and mass-flow deposits were encountered above acoustic basement at Sites 897 and 899, both of which are now situated on substantial basement elevations hundreds of meters above basement of the adjacent basins. They are overlain by poorly fossiliferous uppermost Maastrichtian to Eocene thin claystones and conglomerates (lag deposits) of similar age to the sediments at the same level in the flanking basins. We explain these deposits by rapid Aptian uplift (or relative uplift), about 10 m.y. after the onset of 10 mm/yr seafloor spreading west of the peridotite ridge, followed in Eocene time by blanketing by the sediments of the Iberia Abyssal Plain. This also explains the long history of seawater alteration of serpentinized peridotite at both sites.

The above drilling results are explained here by two possible hypotheses, that also take account of geophysical constraints on the development of the west Iberia ocean/continent transition, neither of which explains all the observations. The first hypothesis is that the ocean/continent transition is underlain by crust formed by ultraslow (~5 mm/yr half-rate) seafloor spreading which, by analogy with the slow-spreading Mid-Atlantic Ridge, leads to the seabed exposure of peridotite and gabbro by extensive faulting. This hypothesis explains the drilling results but has difficulty in explaining the magnetic anomalies and the minute volume of basalt encountered in the cores. The second hypothesis envisages an ocean/continent transition of tectonically and magmatically disrupted continental crust. It explains the Site 900 MORB gabbro as material underplated beneath thinned continental crust or as an aborted point-of-initiation of seafloor spreading, and the alkaline to transitional characters of igneous clasts in Site 897 and 899 deposits as evidence of continental rifting. Its main problem is the lack of unambiguous continental basement samples although there is evidence of re-worked continental sediments in the cores. Both hypotheses predict a highly heterogeneous crust, which will be hard to characterize without further basement sampling.

INTRODUCTION

Rifted continental margins are one of the primary topographic features on the Earth's surface. They provide evidence of one of the most significant geological processes, namely the rifting and breakup of continental lithosphere, which ultimately leads to the creation of new oceanic crust by seafloor spreading and the eventual accretion of the underlying mantle of the oceanic lithosphere. The rifting and breakup of continental lithosphere involves both tectonism on a crustal and lithospheric scale and magmatism that has its roots in adiabatic decompression melting within the asthenosphere. Rifted margins at which magmatism dominates are often described as volcanic, and those where it plays an apparently insignificant role are called nonvolcanic, even though a spectrum of magmatic intensity probably exists between these two extremes.

Although the initial stages of rifting, and the sedimentation that accompanies them, can be examined on land or even offshore on continental shelves, as has been done in the many basins drilled during hydrocarbon exploration, the final stages of the process invariably take place under oceanic water depths. Therefore, their study involves marine geological and geophysical investigations and, ideally, deep drilling. Nevertheless, the thick sediments commonly found on rifted margins hinder investigation, by deep seismic reflection profiling, of the pre- and synrift sediments and the acoustic basement, and presently prevent the use of deep scientific drilling to sample these rocks. Studies have therefore been restricted to those margins that, as

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[2]Institute of Oceanographic Sciences Deacon Laboratory, Brook Road, Wormley, Surrey GU8 5UB, United Kingdom. Present address: Challenger Division for Seafloor Processes, Southampton Oceanography Centre, Empress Dock, European Way, Southampton SO14 3ZH, United Kingdom. Bob.Whitmarsh@soc.soton.ac.uk

[3]Department of Geology and Geophysics MS-126, Rice University, 6100 South Main St., Houston, TX 77005-1892, U.S.A. dale@geophysics.rice.edu

a result of particular geographical or geological circumstances, have been relatively starved of sediment. Even so, the whole gamut of geological and geophysical techniques has not often been applied to any one such margin. Further, by its nature, continental rifting and breakup is a bilateral process involving a pair of conjugate margins so that to obtain the most complete picture, such margins should be studied in pairs. Given the logistical and funding problems of studying pairs of margins, now located on different continents, such conjugate pairs of margins have been studied even more rarely. Between 1988 and 1991, a dozen proposals for scientific drilling on rifted margins in the North Atlantic Ocean were put forward. The JOIDES Planning Committee (PCOM), recognizing the importance of studying pairs of conjugate margins and the rather piecemeal approach of the past, set up a North Atlantic Rifted Margins Detailed Planning Group (DPG) to recommend the way forward. The DPG identified two drilling transects, one designed to investigate a nonvolcanic margin and the other a volcanic margin. Consequently, Ocean Drilling Program (ODP) Leg 149 was programmed in the Iberia Abyssal Plain as the first of a series of legs to address a transect across the Iberia/Newfoundland pair of conjugate nonvolcanic margins.

SCIENTIFIC BACKGROUND

The first quantitatively rigorous fit of North America to Europe was performed by Bullard et al. (1965). They confirmed that, once the Bay of Biscay itself had been closed by a clockwise rotation of Iberia against Europe, the southeast Grand Banks margin could be matched with the west Iberia Margin. Bullard et al. fitted the 500-fathom contour. More refined fits were obtained subsequently as seafloor spreading magnetic anomalies were identified between the two margins and in the Bay of Biscay, and as fracture zones were traced in the region between the Azores-Gibraltar plate boundary and the Charlie-Gibbs Fracture Zone (Masson and Miles, 1984; Klitgord and Schouten, 1986; Srivastava and Tapscott, 1986; Srivastava et al., 1988a,b; Srivastava et al., 1990a,b; Srivastava and Verhoef, 1992; Sibuet and Srivastava, 1994). Srivastava and Verhoef (1992) attempted a kinematic reconstruction that even included the closure of rift basins (Fig. 1). Mapping of marine magnetic anomalies at first identified only the enigmatic J anomaly offshore west Iberia and the Grand Banks. This anomaly, although not a seafloor spreading isochron, has been shown to lie between anomalies M0 and M2 in the Central Atlantic (Rabinowitz et al, 1978; Tucholke and Ludwig, 1982), thereby restricting its age to earliest Aptian-Barremian (120-125 Ma; note that the timescale of Gradstein et al. [1994] is used in this paper). The J anomaly exists only off the southern parts of the Grand Banks and west Iberia Margins, thereby implying a south-to-north propagation of continental separation. More recent detailed studies of the magnetic anomalies offshore the west Iberia Margin have confirmed this interpretation by identifying anomalies as far back as M11 (late Valanginian) in the Tagus Abyssal Plain (Pinheiro et al., 1992) and M3 (early Barremian) in the Iberia Abyssal Plain (Whitmarsh et al., 1990; Whitmarsh and Miles, 1994; Whitmarsh et al., this volume). No seafloor spreading magnetic anomalies appear to exist immediately west of Galicia Bank because seafloor spreading began here during the Late Cretaceous constant polarity interval (Ogg, 1988). In view of the difficulty of fitting relatively short magnetic profiles to anomalies computed from the geomagnetic reversal timescale, the above magnetostratigraphic dates are tentative. Further, lacking a full understanding of the nature and origin of the seaward part of the ocean/continent transition zone, these dates should be regarded as the latest dates at which seafloor spreading began offshore west Iberia. (We use the same definition of the ocean/continent transition as Whitmarsh and Miles [1995], that is, that part of the lithosphere which includes the crust between the thinned continental crust characterized by tilted fault blocks, and the first oceanic crust formed by seafloor spreading.)

Rifting within the extensively investigated basins of the continental crust of west Iberia and the Grand Banks had a long but relatively well-defined history (Wilson, 1975; Wilson et al., 1989; Hiscott et al., 1990; Pinheiro et al., this volume). Three principal phases of extension have been identified, (1) Late Triassic (2) Kimmeridgian (Late Jurassic) and finally (3) an Aptian or earlier (Early Cretaceous) phase that led to continental breakup. These phases were separated by periods of subsidence. The principal basin in western Iberia is the Lusitanian Basin and its probable north-northwest offshore continuation, the Interior Basin, which separates Galicia Bank from mainland Spain (Murillas et al., 1990). The Lusitanian Basin formed by the extensional reactivation of Hercynian basement faults beginning in the Late Triassic (Montenat et al., 1988; Wilson et al., 1989).

The west Iberia Margin (Fig. 2) is unusual in that it has experienced two periods of postrift tectonism. The first had a component of north-south compression contemporary with the partial closure of the Bay of Biscay and incipient subduction along the North Spanish margin at the same time as the Late Cretaceous to Eocene Pyrenean orogeny. This affected the northern part of the margin and led to the uplift of Galicia Bank and, further south, of Vigo, Porto, and Vasco da Gama Seamounts (Boillot et al., 1979) but is not expressed in seismic reflection profiles across the southern Iberia Abyssal Plain (Wilson et al., this volume). In Miocene times, the southern part of the margin also underwent northwest-southeast compression contemporary with a phase of compression in the Betic Range of southern Spain. Locally, this latter episode gently folded the sediments of the Iberia and Tagus Abyssal Plains (Groupe Galice, 1979; Mauffret et al., 1989; Masson et al., 1994; Wilson et al., this volume), which are now overlain by horizontally bedded turbidites.

The Ocean/Continent Transition Off West Iberia

The principal objective of Leg 149 was to investigate the ocean/continent transition zone off western Iberia by drilling into the basement rocks (see below). The sites were chosen, and the drilling was conducted, in the light of the information outlined below; more details are given in the site chapters of the *Initial Reports* volume (Sawyer, Whitmarsh, Klaus, et al., 1994).

A bathymetric chart of the margin west of Iberia (Fig. 2) suggests that it can be divided into at least three segments; these are the Tagus Abyssal Plain segment, the southern Iberia Abyssal Plain segment, and the Galicia Bank segment. The abyssal plains are separated by the Estremadura Spur, a feature of unknown age, which is possibly postrift in age (Mougenot, 1988, reports a dredged sample of alkaline basalt with an $^{39}Ar/^{40}Ar$ date of 74.0 ± 0.7 Ma). The southern Iberia Abyssal Plain segment and Galicia Bank are separated by a marked change in basement depth just south of Vigo and Porto Seamounts. These two features have tended to be chosen as the practical limits of past investigations and will be used here to describe the three segments of the margin.

The Tagus Abyssal Plain segment is the least known and has been described by Mauffret et al. (1989) and Pinheiro et al. (1992). The postrift history of this basin may have been complicated by the proximity of the active transpressional Eurasia-Africa plate boundary along its southwestern and southern margin (Sartori et al., 1994). The lack of basement samples, the very limited number, length, and poor distribution of seismic refraction profiles, and the apparently weak seafloor-spreading magnetic anomalies east of anomaly J all make it difficult to differentiate crustal types within the plain (Pinheiro, 1993; we ignore for this purpose samples obtained from Gorringe Bank, a probable fragment of uplifted oceanic crust on the southern margin of the Tagus Abyssal Plain; see Auzende et al., 1978; Whitmarsh et al., 1993).

The western and northern margins of Galicia Bank have been intensively investigated over many years by G. Boillot and his associates. The eastern margin of Galicia Bank is poorly known (Murillas et al., 1990) and has been little studied since the paper of Black et al.

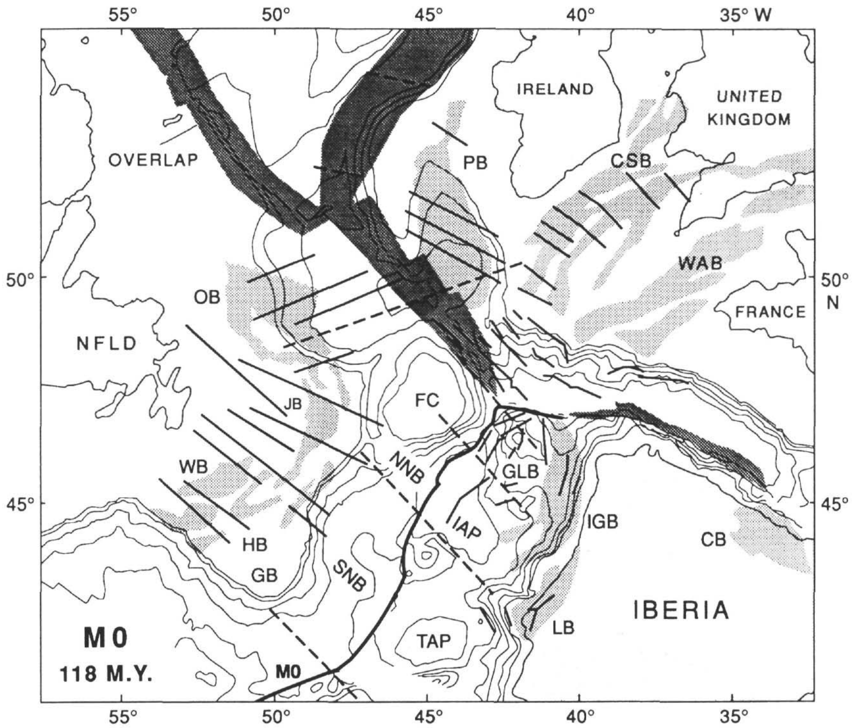

Figure 1. Reconstruction of the North Central Atlantic at chron M0 showing a simplified bathymetry on each plate, outlines of the sedimentary basins (shaded regions), and their plate tectonic features (continuous lines) (fig. 2 of Srivastava and Verhoef, 1992). Also shown are the direction of plate motion (dashed lines) and the resulting overlap between plate boundaries (dark stippled regions). JB = Jeanne d'Arc Basin, WB = Whale Basin, HB = Horseshoe Basin, OB = Orphan Basin, FC = Flemish Cap, CSB = Celtic Sea Basin, WAB = Western Approaches Basin, PB = Porcupine Basin, GLB = Galicia Bank, IGB = Inner Galicia Basin, LB = Lusitanian Basin, CB = Cantabrian Basin, NNB = North Newfoundland Basin, SNB = South Newfoundland Basin, IAP = Iberia Abyssal Plain, TAP = Tagus Abyssal Plain. The Nazaré Canyon is to the northwest of the Lusitanian Basin.

(1964). As a result of its elevated location, Galicia Bank has long been starved of sediment and this has helped in the study of the sediment seismostratigraphy and basement tectonics. The elevated location of Galicia Bank has been explained by 3000 m of Eocene uplift during the Pyrenean compressional episode (Boillot et al., 1979), but the existence of shallow-water uppermost Cretaceous limestones near the top of Galicia Bank (Black et al., 1964) seems to make this a questionable hypothesis. Further, there is no suggestion that uplift has affected the similarly shallow, but conjugate, Flemish Cap. On the western flank of Galicia Bank, multichannel seismic reflection profiles clearly show landward-tilted fault blocks offset down to the west (Groupe Galice, 1979; Montadert et al., 1979; Mauffret and Montadert, 1987, 1988), and these have been shown to possess a regular north-south trend over many tens of km (Thommeret et al., 1988). Some of these blocks outcrop, and sampling by dredging, coring, and diving has revealed a broad area of Paleozoic sediments, continental metamorphic, and granitic basement rocks north of 41°N (Capdevila and Mougenot, 1988) and as far west as 12°14'W at 42°11'M, (Mamet et al., 1991) and 12°27'W at 43°04'N (Boillot et al., 1979). Drilling by ODP Leg 103 either side of, and to the west of, the tilted fault block at 12°14'W further revealed the nature, age, and origin of this margin (Boillot and Winterer, 1988; Boillot, Winterer, et al., 1988). The drill cores revealed a "prerift" sequence of uppermost Jurassic (Tithonian) shallow-water carbonate rocks overlain by about 250 m of dolomite. This sequence is overlain by synrift sediments of Valanginian-Aptian age estimated to be at least 1 km thick in the deepest part of the half-graben (but see Wilson et al., this volume, who question the existence of synrift sediments here). Continental breakup on this margin is estimated to have occurred at about the Aptian-Albian boundary (112 Ma). This agrees with the results of drilling at Site 398 immediately south of Vigo Seamount (Fig. 2; Sibuet and Ryan, 1979) which bottomed in Hauterivian pelagic limestone. Sibuet and Ryan (1979) and Sibuet et al. (1980) concluded that rifting ceased in the latest Aptian.

A feature of possibly great significance to the development of the whole west Iberia Margin is the outcrop, west and north of Galicia Bank, of serpentinized peridotite. Samples of serpentinized peridotite were first reported by Boillot et al. (1980) from a basement ridge that outcrops just above the abyssal plain at 42°00'N, 12°53'W. Subsequently, this basement ridge was drilled at Site 637 during ODP Leg 103 (Girardeau et al., 1988; Evans and Girardeau, 1988) and further samples of serpentinized peridotite were obtained from the northern and western margins of Galicia Bank by submersible (Boillot et al., 1988a). The peridotite rocks appear to have been emplaced 122 Ma ago (i.e., they are synrift in age; Féraud et al., 1988; Schärer et al., 1995). The mechanism of emplacement has been proposed to be both diapiric (Boillot et al., 1980) and tectonic, involving the S reflector (Boillot et al., 1989c) and unroofing (seafloor exposure) of upper mantle rocks (Boillot et al., 1989a,b). There is strong evidence from modeling an east-west deep-towed magnetic anomaly profile that, as one crosses the peridotite ridge from west to east at 42°08'N, there is a strong-to-weak (~10-fold) change in the bulk magnetization of the crust (Sibuet et al., 1995), and therefore the ridge probably marks the landward edge of oceanic crust at this latitude. The velocities in a seismic refraction model that crosses the peridotite ridge are consistent with this interpretation (Whitmarsh et al., unpubl. data; see Fig. 3A). Boillot et al. (1989a) proposed that the peridotite samples represent a continuous belt off Galicia Bank and Beslier et al. (1993) proposed an en echelon, but roughly margin-parallel, ridge extending over at least 300 km as far south as 40°N.

Along the southern Iberia Abyssal Plain segment of the west Iberia Margin, all inferences about the nature of the ocean/continent transition were based on geophysical observations (Fig. 3B); no basement samples existed before Leg 149. The principal evidence is provided by four widely spaced ocean-bottom seismograph (OBS) seismic refraction profiles shot parallel and normal to the margin (Fig. 2; Whitmarsh et al., 1990). The interpretation of these profiles demonstrated that the deep margin possesses an unusually thin (3-4 km) crust and that the structure varies with distance offshore. Whitmarsh et al. (1990) interpreted the seismic velocity structure in terms of thinned continental crust passing westwards into a higher velocity crust underlain by 7.5 km/s material. The westernmost line was shot over 6-km-thick crust, interpreted to be oceanic, underlain by 7.55-7.6 km/s material. A gravity model, constrained by the seismic velocities, is consistent with the high velocity layer at the base of the crust being a laterally continuous layer, probably of serpentinized peridotite (Whitmarsh et al., 1993). Underplating, as an explanation of these relatively high velocities, seems unlikely because it implies unrealistically high asthenospheric temperatures. Modeling of sea-surface magnetic anomaly profiles on this margin segment strongly suggests that seafloor spreading began at about the beginning of chron M3 (127 Ma; Whitmarsh et al., 1990). This was subsequently confirmed by deep-towed profiles (Whitmarsh et al., this volume). The first seafloor spreading crustal block in such models lies in a location consistent with the interpretation of the seismic refraction lines and just west of a narrow elongated basement ridge across which there is poor transmission of seismic energy; on the North Biscay Margin a similar zone has been associated with the landward edge of oceanic crust (Whitmarsh et al., 1986). Beslier et al. (1993) predicted that this ridge was the southward continuation of the peridotite ridge off Galicia Bank.

The Ocean/Continent Transition Off Newfoundland

Off the southeast Grand Banks Margin, the enigmatic J magnetic anomaly occupies a symmetrical position to that off west Iberia and dies out north of 44°30'N, just south of Flemish Cap (Srivastava et al., 1990a). In other respects, the magnetic anomalies here have been investigated less than off west Iberia and, other than J, no Early Cretaceous anomalies have been detected (Srivastava et al., 1990a). Two transects of OBS seismic refraction lines have been shot off the southeast Grand Banks margin. Off Flemish Cap, approximately conjugate to the deep Galicia Bank margin, Todd and Reid (1989) suggested, on the basis of seismic velocities and gravity modeling, that the thinned continental crust extends only 25 km seaward of the edge of the continental shelf. However, the constraints are weak because mantle arrivals were seen on only one of the refraction lines. Nevertheless, it is interesting that one line (HU-6), supposedly shot over oceanic crust, exhibited a crustal thickness of only 4 km. There is also widespread evidence of arrivals with a velocity of about 7.3 km/s (possibly highly altered mantle, according to Reid, 1994) at the base of the crust in water depths of more than 4000 m. The second transect was shot over the southern part of the margin, conjugate to the Tagus Abyssal Plain, and was also supported by gravity modeling (Reid, 1994). Again, this has been interpreted in terms of a rapid thinning of continental crust over a distance of 30-40 km beneath the continental slope. Seaward, there is a 50-km-wide zone of 7.2-7.6 km/s material at the base of a thin (<3 km) crust. Although the oceanic crust thickens seaward, it does not exceed 4 km on this transect.

Keen and de Voogd (1988) have described two deep (20-s two-way traveltime) multichannel seismic reflection profiles across the southeast Grand Banks margin either side of the refraction line shot by Reid (1994). The northern line (Line 85-4) exhibits fault blocks beneath the continental slope that die out seaward, in a water depth of about 3000 m, and are superseded by a rougher acoustic basement that suggests the presence of oceanic crust (Keen and de Voogd, 1988). Further south, Line 85-2 exhibits a steeper continental slope and a landward-dipping reflector (also seen on four adjacent multichannel profiles) said to mark the continent-ocean boundary. Thus, this interpretation is consistent with that based on the above refraction profiles.

An alternative view was presented by Tucholke et al. (1989). These authors emphasized the rough/smooth magnetic anomaly boundary coincident with the J anomaly, the smooth acoustic basement landward of the J anomaly and a seismostratigraphic unconformity, which they correlate with a breakup unconformity on the adjacent shelf, which is underlain by synrift (?) sediments. They therefore interpret the J anomaly as marking the continent-ocean boundary.

DRILLING OBJECTIVES

The principal objective of Leg 149 was to sample the upper crust within the ocean/continent transition of the Iberia Abyssal Plain to establish its nature and to test predictions based on geophysical observations. Naturally, this bold objective had to be tempered by the accessibility of the crust using current technology. To achieve significant progress within a single leg, four proposed sites (IAP-2, 3C, 4, and 5) were chosen. These sites lie on basement highs situated at critical points within the ocean/continent transition. We expected, and were able, to drill three of these sites within the time allotted to Leg 149: IAP-4 became Site 897, IAP-2 became Site 898, and IAP-5 became Site 900 (Fig. 2). Two other sites were also drilled: Site 899 was considered an alternate for IAP-2, and Site 901 is east of IAP-5 (Fig. 2). Proposed site IAP-3C was not drilled (Fig. 2). We planned to penetrate the upper acoustic basement to a depth of several hundred meters and, using cores and downhole logs, to determine its origin and history. This was to be done by petrological and chemical analyses of the cores, by microstructural examination of the cores, by examination of the mineralogy of the cores, by apatite fission track analysis and/or isotope dating of suitable core material, by velocity and magnetic measurements in cores, by analysis of geochemical logs, by interpretation of the Formation MicroScanner (FMS) and other logs, and by whatever other means seemed appropriate. The leg results related to these primary objectives are synthesized in this paper.

Secondary objectives related to the sediments themselves. One aim was to determine the history of turbidite sedimentation in the Ibe-

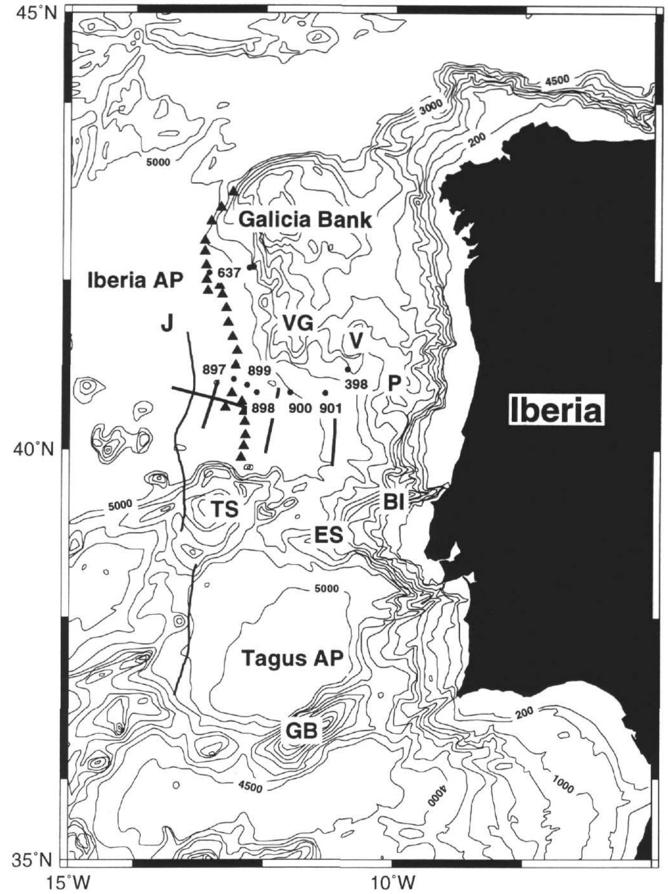

Figure 2. Bathymetric chart of west Iberia (contour interval 200, 500, 1000, 1500, etc., m). Triangles denote the position of the peridotite ridge (from Beslier et al., 1993); solid circles are positions of DSDP/ODP sites; open circle is position of undrilled Site IAP-3C; bold line is the magnetic J anomaly picked from the reduced-to-the-pole chart of Miles et al., (this volume, Fig.1 [foldout in back pocket]). VG = Vasco da Gama Seamount, V = Vigo Seamount, P = Porto Seamount, TS = Tore Seamount, ES = Estremadura Spur, BI = Berlenga Islands, GB = Gorringe Bank.

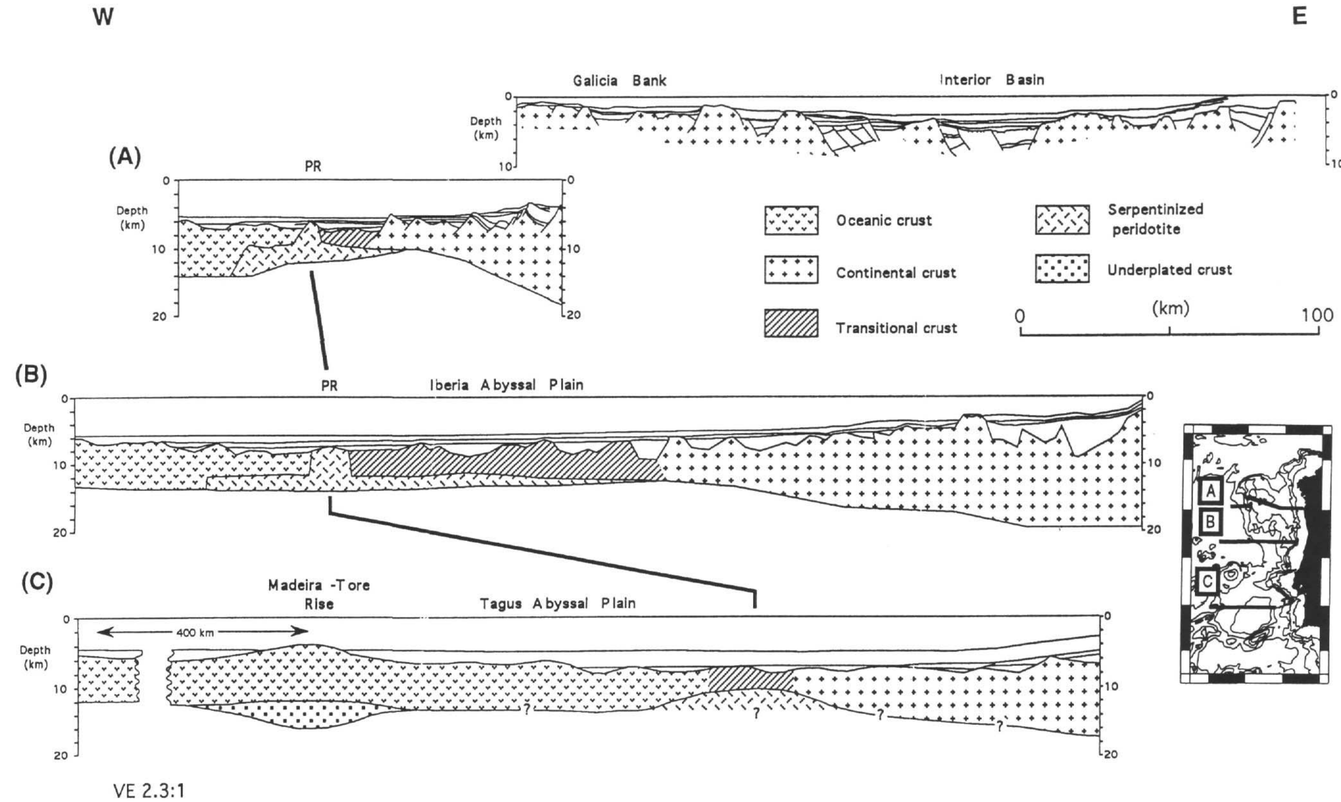

Figure 3. Summary east-west cross sections along three transects off west Iberia (see inset for locations) PR = peridotite ridge. **A.** Across Galicia Bank, based on multichannel seismic reflection profiles of Murillas et al. (1990) and seismic refraction results of Whitmarsh et al. (work in progress) and Horsefield (1992). **B**. Across the southern Iberia Abyssal Plain, based on seismic refraction lines of Whitmarsh et al. (1990), gravity model of Whitmarsh et al. (1993) and on multichannel seismic reflection profile LG-12 (Beslier, this volume). **C.** Across the Tagus Abyssal Plain, based on seismic refraction models of Pinheiro et al. (1992) and Peirce and Barton (1991). A peridotite ridge has not yet been recognized beneath the Tagus Abyssal Plain.

ria Abyssal Plain. Work done in the Madeira Abyssal Plain indicates that, in general, a single turbidite was deposited each time sea level changed between a glacial and an interglacial period or vice versa (Weaver et al., 1986). We also expected to determine to what extent the age and frequency of turbidites related to past climatic change. Another objective was to date the deformation of the sediments and to relate this to the Paleogene and Miocene deformation in Europe. The results of the leg related to these secondary objectives are synthesized by Milkert et al. (this volume) and Wilson et al. (this volume).

We planned to estimate heat flow at each of the Leg 149 sites through measurements of thermal conductivities and thermal gradients. Thermal conductivity of the core samples was to be measured routinely on the ship. The thermal gradient was to be determined by measuring in situ temperatures in relatively shallow sediments at various depths (approximately the upper 300 m below seafloor) using the ADARA temperature and Wireline Sampling, Temperature and Pressure (WSTP) tools. Temperatures in open holes were to be measured as part of most logging runs. We planned to correct temperature logs, on the basis of the results of successive runs in the same hole, for disturbances caused by drilling and circulation. The results of the leg related to these objectives are reported by Louden (this volume).

We expected to acquire data that could be used to estimate the late postrift subsidence history of the Iberia Abyssal Plain. We planned to observe the paleodepth, age, environment of deposition, and physical properties of each sedimentary unit. We did not expect to be able to deduce the synrift and early postrift subsidence histories because we did not expect to encounter a continuous sequence of sediments of that age. The subsidence history was unlikely to be precise at these sites because the basin was relatively starved of sediments. Estimates of depths of deposition of continental slope sediments from paleoenvironmental observations are, at best, accurate to 500 m. The results of the leg related to these objectives are reported by Wilson et al. (this volume).

DRILLING RESULTS THAT CONTRIBUTE TO INVESTIGATIONS OF THE ORIGIN OF THE OCEAN/CONTINENT TRANSITION

This section summarizes the results from the transect of Leg 149 drill sites that pertain to the continental-rifting to seafloor-spreading history of the west Iberia Margin (Fig. 4).

Site 897

Site 897 (Shipboard Scientific Party, 1994a) was drilled on the crest of a narrow elongate basement ridge predicted, apparently correctly, by Beslier et al. (1993) to represent the southward continuation of the peridotite ridge previously sampled on the deep Galicia Bank margin. This ridge is also coincident with the landward edge of oceanic crust in models of seafloor spreading magnetic anomalies in the vicinity of the site (Whitmarsh et al., this volume). The basement was penetrated to a maximum depth of 143 m and consists of peridotite that is almost completely serpentinized and was partially brecciated during and after serpentinization. Petrological, geochemical, and microstructural studies of the cores all point to the same sequence of

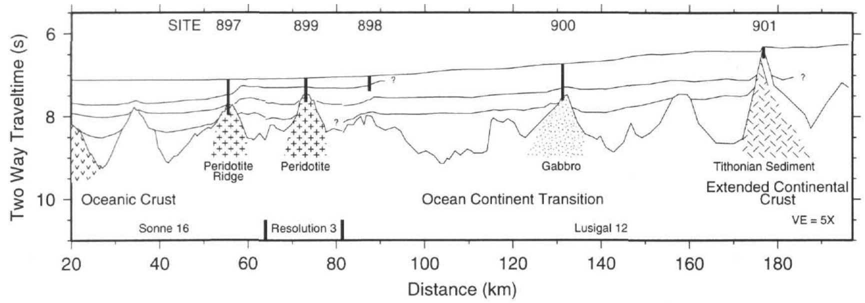

Figure 4. Composite sediment and basement section through the Leg 149 sites. The composite section was assembled from parts of three roughly east-west seismic sections. Major sediment reflectors are indicated as is the type of acoustic basement at each drill site.

events which accompanied the exhumation of the peridotite from the deep mantle to the seabed.

1. The first event for which there is evidence in the rocks was prolonged crystallization of pyroxene under equilibrium conditions of 1170°-1230°C (based on two-pyroxene geothermometer calculations by Seifert and Brunotte, chapter 23, this volume). At a depth of 30 km this corresponds to a mantle potential temperature of 1156°-1216°C.
2. The next event was a high temperature ductile shear deformation, as evidenced by a porphyroclastic texture of the peridotites and websterites, the occurrence of kink-bands in olivine and orthopyroxene porphyroclasts, the distortion of clinopyroxene exsolution lamellae, and weak recrystallization of websterite, olivine, and pyroxenes. Petrofabric analysis of olivine suggests deformation at temperatures of 900°-1000°C (Beslier et al., this volume) whereas two-pyroxene geothermometry indicates temperatures of 880°-965°C (Cornen et al., this volume). There is some evidence for the existence of plagioclase melt during this phase (Cornen et al., this volume).
3. Following the ductile shear, the peridotite experienced limited partial melting, as demonstrated by the abundance of websterites, the modal compositions of the peridotites, the microtextural relationships, the residual character of the peridotites, the poikilitic character of many of the pyroxene crystals, and the presence, locally, of up to 35% plagioclase (Cornen et al., this volume; Beslier et al., this volume). REE spidergrams, normalized to primitive mantle, similarly suggest that some of the peridotites are still fertile for further melting (Seifert and Brunotte, chapter 23, this volume). At about the same time, and certainly after Event 2, websterite underwent a subsolidus reequilibration under plagioclase facies conditions (less than 1 GPa, ~30 km depth) as evidenced by the crystallization of secondary olivine and orthopyroxene between, or in, the exsolution lamellae of former pyroxenes (Beslier et al., this volume) at a two-pyroxene geothermometer temperature of around 780°C (Cornen et al., this volume).
4. In a few shear zones in the cores, there is evidence of a mylonitic and ultramylonitic plastic shear deformation that clearly succeeded the websterite reequilibration and took place in the plagioclase stability field at around 700°C (735°C from pyroxene neoblasts [Cornen et al., this volume]). The deformation was coaxial with the previous high temperature ductile shear deformation (Beslier et al., this volume).
5. The next event of note was extensive low-temperature deformation under (near?) surface conditions coeval with serpentinization of the peridotites. It is expressed as brittle fracturing and brecciation over thick rock intervals or as plastic shearing in serpentinite-rich zones. There are local shear zones, some of which are low-dipping (~30°) and may have accommodated large displacements in the surrounding rock (Beslier et al., this volume). Oxygen isotope measurements suggest that serpentinization occurred below 100°C and was accompanied by a voluminous throughput of seawater, probably close to the seafloor (Agrinier et al., this volume).
6. Finally, calcite was precipitated from seawater at 10°-20°C according to stable isotopes and trace-element analyses (Milliken and Morgan, this volume) and at 13°-19°C according to oxygen isotopes (Agrinier et al., this volume). The highly oxidized and carbonated zone, which forms the upper part of the basement section, is thought to have resulted from prolonged exposure on the seafloor prior to burial by sediment (Gibson et al., this volume).

Although in some circumstances it is possible to distinguish between subcontinental and suboceanic peridotites, this was not possible at Site 897. The peridotites have a low Na-content, like oceanic peridotites, but also display clear subsolidus reequilibrium features, like subcontinental ones (Cornen et al., this volume).

Louden (this volume) measured radiogenic heat production on six samples of the peridotite from Site 897 and found an average of <0.01 microwatts/m^3. This value is consistent with the low concentrations of radionuclides that are typically found in mantle peridotites and with leaching of radionuclides during hydration.

Site 899

Site 899 (Shipboard Scientific Party, 1994b) was drilled on an isolated subcircular basement high within the ocean/continent transition. Acoustic basement was penetrated to a maximum of 180 m in Hole 899B and yielded cores of brecciated serpentinite over a very mixed assemblage of unbrecciated ultramafic and mafic rocks. There is no evidence that in situ basement rocks were sampled at this site; the cores represent debris flows and mass wasting deposits. In spite of this qualification, however, the cores from the lower part of Hole 899B reveal very useful information about the nature of the contemporary outcropping rocks that provided the source of cored material.

The peridotites in the Site 899 cores appear to have experienced the same phases of high-temperature ductile shear deformation, limited partial melting, subsolidus reequilibration under plagioclase facies conditions, and mylonitic and ultramylonitic plastic shear deformation as at Site 897 (Cornen et al., this volume). The petrology of the rocks differs only in detail; at Site 899 harzburgites and dunites are scarce relative to lherzolites (which suggests less partial melting and is unusual for peridotites collected away from fracture zones), websterites are absent, and the harzburgites are richer in orthopyroxenes (which are quite isotropic in shape and display no traces of deformation or resorption); plagioclase-bearing lherzolites have been clearly identified only at Site 899. The final phase of calcite precipitation from seawater also occurred at the same low-temperatures as at Site 897.

The serpentinized breccia unit (Subunit IVA) and the underlying deposit (Subunit IVB), of boulder-sized (0.3 to >3.0 m) polymict clasts separated by sedimentary intervals, are of interest in that they provide evidence of late-stage tectonism (or, at least, erosional processes) on a nonvolcanic margin as well as of the range of lithologies exposed in such a situation. The clasts in Subunit IVA consist of 90% serpentinized peridotite and <10% low-grade metamorphosed Mg- and Ca-rich igneous rocks, that fall into five mineral assemblages characteristic of prehnite-pumpellyite facies (Shipboard Scientific Party, 1994b), and rare microgabbros (diabases). Subunit IVB contains a great variety of clasts. These have been classified as basalt lavas, undeformed microgabbros, sheared amphibolite, mylonites (mylonitized gabbro?), and sediments (claystones, and one example each of limestone, laminated siltstone and chert). The basalts and microgabbros have similar petrological characteristics, phase chemistry and REE patterns; their parental magmas had alkaline affinities. The microgabbros are rather transitional with a REE pattern similar to E-MORB (Cornen et al., this volume; Seifert and Brunotte, chapter 29, this volume) but the basalt REE pattern is close, if not closer, to continental flood basalts (CFB) when a standard is used for normalization which is based on CFBs (from the Columbia River, Keweenawan, Noril'sk and Mid-Continent Rift provinces) that are relatively free of contamination by continental lithosphere, probably because their rapid ascent precluded reaction with continental material (Seifert and Brunotte, chapter 29, this volume). The metabasite clasts have relict pyroxene and oxides with mineralogy and bulk composition close to that of the unaltered alkaline microgabbro; other pyroxenes, however, have anomalous Na similar to the flaser gabbros sampled at Site 900. One ripidolite (chlorite)-bearing clast contains the same mineralogy and REE content as a chlorite-bearing schist sampled above peridotite outcrops on the deep Galicia Bank margin and interpreted as an underplated Fe-Ti gabbro (Schärer et al., 1995; Cornen et al., this volume). Thus, most of the mafic rock clasts (basalts, microgabbros, metabasites, and amphibolite) display transitional MORB to alkaline features that suggest a subcontinental source of melt. However, the clasts are clearly not in situ and the distance of their source(s) from the site remains unknown.

Tectonic Implications of Site 897 and Site 899 Mass-flow and Debris-flow Deposits

As mentioned above, the sediments towards the bottom of the holes at Sites 897 and 899 have the character of debris flows and mass-flow deposits. It is interesting to consider the nature, age, and origin of these deposits and the tectonic implications of encountering them on basement highs (Fig. 4).

Lithological Subunit IIIB is found in Holes 897C, 897D, and 899B (Fig. 5A). At Site 897, it consists of four sequences of claystones, sandstones, and poorly sorted conglomerates, where each sequence becomes finer grained upwards. The sequences have been clearly dated in both Holes 897C and 897D as early late Aptian by using nannofossils (de Kaenel and Bergen, this volume) and have been interpreted as turbidites and debris flows (Shipboard Scientific Party, 1994a). Subunit IIIB at Site 899 is rather different; it is noncalcareous and consists of sandstone, sandy silty and laminated claystone, silty claystone with black (probable manganese) concretions, volcanic ash, and a polymictic clayey conglomerate with highly altered basalt clasts, black metasediments, and claystone. Subunit IIIB has been dated by planktonic foraminifers to be Campanian to early Maastrichtian in age (Gervais, this volume); a sample of autochthonous hemipelagic Subunit IIB material from 4.5 m higher up Hole 899B, that is barren of planktonic foraminifers, contains deep-water benthic foraminifers of middle/early Eocene age (Kuhnt and Collins, this volume). Subunit IIIB has been interpreted as a mixture of turbidites and debris flows with the basal conglomerate representing a "weathering" deposit (i.e., one exposed on the seafloor for a long period; Shipboard Scientific Party, 1994b). Therefore, it is incorrect to assume that Subunits IIIB at the two sites are equivalent because of their substantial difference in age.

At Site 899, lithological Subunit IVA is 114m thick and consists of three serpentinized peridotite breccias (Fig. 5A). The bottom 9.5-19 m are of early Aptian age (de Kaenel and Bergen, this volume); the remainder is unfossiliferous, and we assume it is of early Aptian age, too (de Kaenel and Bergen, this volume; Kuhnt and Collins, this volume). This Subunit has been interpreted as a subaqueous mass-flow deposit because of its texture (unsorted and angular fragments, etc.) and other features (Shipboard Scientific Party, 1994b). A similar origin has been attributed to the 74-m-thick Subunit IVB, already described, which directly underlies Subunit IVA. Comas et al. (this volume) describe the whole of Unit IV as an olistostrome (by analogy with similar deposits exposed in the Alps) whereas Gibson et al. (this volume) describe Subunit IVA as a submarine landslide deposit (by analogy with subaerial landslides). Obviously, such deposits depend on gravitational potential energy to generate the kinetic energy required for their lateral displacement from a source region. Comas et al. favor a fault-scarp source a few tens of kilometers to the north whereas Gibson et al. argue for a more local source within a few kilometers. We lack detailed knowledge of the topography, now and in the past, of the acoustic basement in the immediate vicinity of Site 899. Although the site is situated on a basement high, as seen on two seismic cross sections, it is important to remember that a locally shallower high may exist out of the plane of these sections.

At Site 897, lithological Unit IV is 30-40 m thick and consists of intervals of serpentinized peridotite sandwiched between claystones and other lithologies including breccias (Fig. 5A). The sediments young upwards and range in age from late Barremian to early Aptian (de Kaenel and Bergen, this volume). This unit has also been interpreted as a subaqueous mass-flow deposit, and it is tempting to correlate it with Subunit IVA 20 km away at Site 899 (Comas et al., this volume), but strictly the age equivalence is uncertain (only the bottom 9.5-19 m of Subunit IVA at Site 899 are definitely of equivalent age); this deposit could also have a more local origin.

A much-discussed problem is how the debris-flow and mass-wasting deposits described above were emplaced on what today is a basement high. Did they travel uphill or were they uplifted after deposition? The crux of the problem is that on the seismic sections through the area seismostratigraphic unit 6 of Wilson et al. (this volume), of comparable age to the drilled debris flows and mass-wasting deposits, is found in the deep basins that flank the Site 897 and Site 899 basement highs (Fig. 3 of Wilson et al., this volume). Wilson et al. show, by linking to seismic profiles across Deep Sea Drilling Project Site 398 (Sibuet and Ryan, 1979; Sibuet et al., 1980), that unit 6 includes sediments of probable (Valanginian?) Hauterivian-Aptian age. The relief of acoustic basement, from the crest of the high to the mean level of the adjacent basin, is computed to be about 450 m at Site 897 and about 750 m at Site 899, using sediment velocities of Whitmarsh et al. (1990). At first sight, it might therefore seem that very significant vertical motion of the seabed was required to explain these observations (mass flows have only a limited ability to travel uphill, and certainly less than 450 m). It important to recall, however, that some of the sediments were drilled within acoustic basement itself and that the resolution of the seismic sections we have available is at best a few tens of meters.

Thus, at Site 899, all of Unit IV lies within acoustic basement of early Aptian age and the basement is capped by the Campanian to lower Maastrichtian conglomerate of Subunit IIIB (Fig. 5 A). The debris-flow deposit of Subunit IVB contains a mixed set of lithologies and therefore is more likely to have had a variety of nonlocal sources. We propose that Subunit IVB was laid down before uplift of the present high (Fig. 5C). This was followed by rapid uplift in the early Aptian during which nearby outcrops of peridotite supplied the brecciated material of Subunit IVA possibly in the form of submarine landslides (Gibson et al., this volume); the concurrent effect of this uplift on the deposition of Seismostratigraphic Unit 6 in the flanking basins is hard to detect on seismic profiles. The unit is recognized only in the deepest parts (it tapers out towards the highs), but the Unit 5/6 interface is often more steeply dipping than the overlying beds. Thin (seismically unresolvable) patches of Unit 6 could well lie within the top of acoustic basement. The Site 899 basement, once elevated, was exposed for many m.y. to seawater and only in early Eocene time was it eventually blanketed by sediments of uppermost Seismostratigraphic Unit 4 (Fig. 5B).

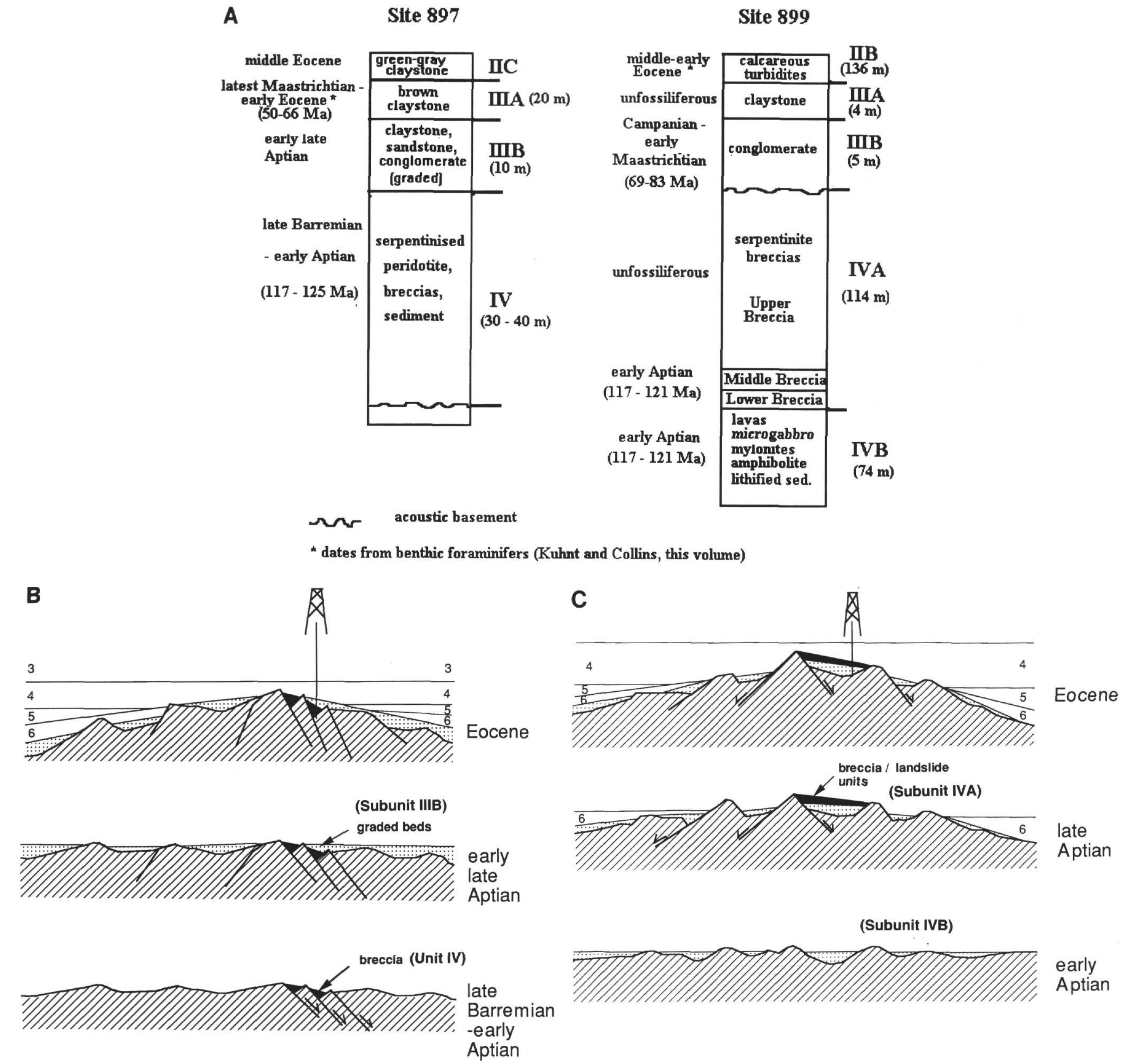

Figure 5. **A.** Summary of the ages, thicknesses, and sequences of lithostratigraphic units just above and within acoustic basement at Sites 897 and 899. **B.** Sketches of the development of the Site 897 basement high illustrating how early late Aptian age sediments are now found on top of the high. **C.** Sketches of the development of the Site 899 basement high illustrating how early Aptian age sediments are now found on top of the highs.

At Site 897 different arguments are employed. Here, acoustic basement coincides with the base of Unit IV; there is 30-40 m (Unit IV) of serpentinized peridotite and upper Barremian-lower Aptian sediment (some unfossiliferous), 10 m (Subunit IIIB) of lower upper Aptian claystone, sandstone, and conglomerate, and 20 m (Subunit IIIA) of uppermost Maastrichtian-lower Eocene (Kuhnt and Collins, this volume) brown claystone between basement and the middle Eocene sediments of Subunit IIC (Fig. 5A). Such a 60-70-m interval corresponds to about 0.06 s two-way traveltime on the seismic sections and is at the limit of resolution. We propose that the Site 897 basement rocks were brecciated while being tectonically emplaced at the seabed in late Barremian-early Aptian time to generate the rocks of Unit IV (Fig. 5B). At this site we attribute the absence of an equivalent of the lower Aptian debris flow Subunit IVB of Site 899 to either the complexities of contemporary basement relief (i.e., no pathway existed for the debris flow to reach Site 897) or to Site 897 having a higher elevation than Site 899 at the time. Basement relief was sufficiently low to allow the accumulation of the four lower upper Aptian graded turbidites and debris flows, of unknown provenance, of Unit IIIB. Only then did uplift accelerate in a similar manner to that which occurred at Site 899, albeit a few million years later. The high was exposed at the seabed until it too was eventually blanketed, in the early middle Eocene, by sediments of the top of seismostratigraphic unit 4. Gibson et al. (this volume) have also proposed prolonged exposure (they estimate 40 m.y.) of the peridotite basement to circulating seawater at this site from a study of peridotite alteration. seismostratigraphic units 4 and 5 are not resolved on the top of the high; the equivalent beds are only 20 m thick (~0.02 s two-way traveltime) and are likely to be a different more clay-rich lithology than their equivalents in the deeper flanking basins.

Thus, the observation of gravity-driven deposits now found on top of basement highs can be explained by the above arguments. The tectonic implications are simple and not surprising. Significant uplift occurred at both sites in early late Aptian (~116 Ma) at Site 897 and in early Aptian (~117-121 Ma) at Site 899 a few m.y. after the best estimate of the onset of seafloor spreading (approximate start of anomaly M3, 127 Ma; Whitmarsh et al., this volume). This was almost certainly a time of rapid tectonic adjustment and differential vertical motion of the basement driven by the normal faulting (at both steep and low angles), which accompanied continental breakup. Although we have discussed the relatively elevated positions of the basement highs in terms of uplift, this could equally have resulted during regional subsidence involving greater subsidence of the flanking basins, perhaps accompanying thermal subsidence, which is greatest immediately after continental breakup. Following uplift, sedimentation on the highs was practically insignificant (0.4 m/m.y. at Site 897 and 1.3 m/m.y. at Site 899) for over 60 m.y. until the sites were inundated by sediment in the early middle Eocene.

Site 900

Site 900 is situated over an angular basement high that had been interpreted using geophysical data to be a tilted continental basement block (Fig.4; Shipboard Scientific Party, 1994c). We cored 748.9 m of Pleistocene to Paleocene sediment and 56.1 m of basement composed of metagabbro.

The origin of the gabbro is debatable. According to Cornen et al. (this volume) the entire basement core consists of massive flasered gabbro. The samples of the metagabbro that they analyzed display a light rare-earth element (LREE) enrichment up to 2.7 times chondrite and REE enrichment of 2.5-6.5 times chondrite. The latter enrichment is weak relative to the unfoliated gabbros and lavas from Site 899. The REE pattern is relatively flat with a positive europium (Eu) anomaly. The petrologic characteristics of the metagabbro strongly resemble those of basalts, dolerites, and gabbros from the Galicia Bank margin. Regarding the origin of the gabbro, Cornen et al. (this volume) note that its mineralogical features and geochemical characteristics differ from those of the extensively investigated tholeiitic gabbros from the Southwest Indian Ridge. The high and relatively flat REE pattern is unusual for gabbro at mid-ocean ridges (N-MORB), but more closely resembles that of island arc gabbro. The LREE enrichment is unusual and better fits that of transitional magma such as forms at the beginning of rifting. The rocks of the Galicia Bank margin, which Cornen et al. argue are similar to the Site 900 gabbro, are considered to have formed at the end of continental rifting by the interaction of a MORB source and an enriched continental mantle.

Seifert et al. (this volume) report that the metagabbro shows an extreme depletion of incompatible trace elements, a lack of oxide minerals, and an Fe/Mg ratio below that of primitive basalt. They argue that this suggests a cumulate origin for the metagabbro. Consistent with the results of Cornen et al. (this volume) they find REE patterns are flat at 2-4 times chondrite with a small positive Eu anomaly. Seifert et al. suggest that the REE pattern for all but one sample is similar to that of cumulate gabbro from 26°N on the Mid-Atlantic Ridge. The other sample is similar to N-MORB. Low concentrations of incompatible elements indicate little retention of magma. A loosely constrained parental magma calculation yields an REE pattern similar to transitional MORB. Isotopic analyses of the metagabbro (Seifert et al.) yield large positive epsilon neodymium (Nd) values (6.3-10.3). Epsilon Nd for MORB is about +8, and it is significantly reduced, usually to negative values, when there has been continental contamination of asthenospheric melt. Neodymium isotopes are particularly valuable for indicating origin in rocks like these because they are relatively unaffected by late-stage seawater alteration. Thus, the Site 900 Nd isotope data are consistent with formation of the metagabbro at a mid-ocean ridge. Lead isotopic analyses also indicate that the gabbro was formed in an oceanic environment.

The petrologic observations suggest the following general sequence of events for these rocks.

1. Dynamic recrystallization during shear deformation developed the foliation and the porphyroclastic or granuloblastic texture of these rocks. Cornen et al. (this volume) report that some of the flasered gabbros contain primary phases, which are indicative of metamorphism (intense shearing) at pressure of at least 0.4 GPa (about 13 km) and temperatures typical of granulite facies conditions. These conditions are evidenced by jadeite content of the clinopyroxene and from the presence of spinel within plagioclase neoblasts. Seifert et al. (this volume) report that the oceanic cumulate gabbros have a poorly developed and discontinuous foliation that texturally resembles metamorphosed oceanic cumulate gabbros from ODP Site 735 on the Southwest Indian Ridge. Alternating discontinuous felsic and mafic bands define the poorly developed foliation. They argue that this is characteristic of high-grade metamorphism in oceanic crust and quite distinct from that in continental crust.
2. Cornen et al. (this volume) report that some of the gabbro bears a secondary mineralogy that points to reequilibration in static lower amphibolite to greenschist facies conditions. Agrinier et al. (this volume) used oxygen isotope data to show that Event 2 seems to have included hydrothermal addition of water and to have occurred at about 280 ± 20°C. This temperature is above the closure temperature of plagioclase, indicating that this retrograde metamorphic phase preceded Event 3 described below.
3. $^{39}Ar/^{40}Ar$ dating suggests that the gabbro cooled through the closure temperature of plagioclase (200°-250°C) at about 136.4 ± 0.3 Ma. $^{39}Ar/^{40}Ar$ dating was attempted on two bulk samples of plagioclase extracted from the metagabbro (Féraud et al., this volume). Both age spectra are disturbed by alteration processes and possibly excess argon. The two plagioclase samples came from a zone preserved from the late low-temperature deformation event accompanied by hydrothermal fluid circulation. One sample displays a plateau with a weighted mean age of 136.4 ± 0.3 Ma (this value slightly violates the conventional test of consistency at the 95% confidence interval within the plateau). The other sample displays a plateau with a weighted mean age of 124 ± 1 Ma (this value slightly violates the conventional test of 50% plateau width). Féraud et al. (this volume) consider the first date, 136.4 ± 0.3 Ma, to be the more reasonable estimate. It is not clear whether their preference is based on geochemical observations or on perceived consistency with other geological interpretations in the region but both dates indicate cooling (emplacement?) of the gabbro at about the time of continental breakup.
4. The gabbro was exposed at the seafloor for an extended period, probably tens of millions of years, prior to burial by sediments. During this period the rocks experienced alteration by mass-wasting and pervasive penetration of seawater (Agrinier et al., Cornen et al., and Seifert et al., all in this volume). Fractures were filled with chlorite, as well as other minerals, and finally calcite. This deformation is unevenly distributed through the cores and locally it brecciated the rocks.

Louden (this volume) measured radiogenic heat production on six samples of the metagabbro from Site 900 and found an average of 0.21 ±0.19 $\mu W/m^3$. This value is higher than typical values for tholeiitic basalts (0.08 $\mu W/m^3$) and somewhat lower than values for the lower continental crust (0.4 $\mu W/m^3$). Some reduction was probably caused by radionuclide mobility during weathering. The recalculated heat flow at Site 900 is 63 ± 9 mW/m^2. This higher heat flow relative to Site 897 is difficult to explain given the low value of radiogenic heat production measured in the gabbro cores. It implies that either the gabbro cores are unrepresentative of the crustal rocks and/or that radioactive nuclides have been leached from the cores (and the upper hundreds, rather than thousands, of meters of acoustic basement) by prolonged hydrothermal circulation.

Site 901

Site 901 overlies the crest of an angular basement high that appears to be a tilted basement fault block covered by pre- and post-faulting sediments (Shipboard Scientific Party, 1994d). Site 901 was drilled quickly during the last 48 hours of the leg. We drilled and washed down to 182.0 mbsf and then cored intermittently down to 247.8 mbsf. Coring was terminated when we had to depart for Lisbon. The apparent prefaulting sediment at Site 901 is 500-600 m thick (Shipboard Scientific Party, 1994d). Therefore, we estimate that there are 250-350 m of presumably older prefaulting sediment below the base of Hole 901 A.

The deepest sediments recovered in Hole 901A were dark carbonaceous claystones (black shales) containing late and early Tithonian nannofossils (de Kaenel and Bergen, this volume). Late and probably early Tithonian foraminifers were also found (Collins et al., this volume). Tithonian benthic foraminifers assemblages were found that indicate a neritic (< 200 m) depositional environment and probably dysaerobic (poorly oxygenated) bottom-water conditions. Deeper water benthic foraminifers were not found.

We take the benthic foraminiferal observations as strong evidence that the basement at Site 901 is extended continental crust. The sediments were probably deposited on continental crust prior to significant extension. If we take the maximum thickness of sediment below the terminal depth of the hole (350 m), and the 200 m water depth, we estimate that the sediment-unloaded depth-to-basement at Site 901 in the Tithonian was about 320 m. If this is interpreted as initial subsidence caused by extension (McKenzie, 1978), of basement originally at sea level, we estimate a beta factor of about 1.15 at that time.

HYPOTHESES FOR THE ORIGIN OF THE OCEAN/CONTINENT TRANSITION IN THE SOUTHERN IBERIA ABYSSAL PLAIN

Two contrasting hypotheses for the origin of the ocean/continent transition beneath the southern Iberia Abyssal Plain are proposed in this paper. The first, presented by Sawyer (1994) and as hypothesis 1 in this paper, suggests that the ocean/continent transition is composed largely of very slow-spreading oceanic crust. The second, presented by Whitmarsh and Miles (1994, 1995) and as hypothesis 2 in this paper, suggests that the ocean/continent transition was formed by tectonic and magmatic disruption of continental crust. Two further papers in this volume, which discuss extensional processes during the formation of the ocean/continent transition, are also relevant here. One, presented by Krawczyk and Reston (in Krawczyk et al., this volume), suggests that the ocean/continent transition exposes progressively from east to west, upper continental crust, lower crust, and upper mantle adjacent to a major detachment fault. The other, presented by Beslier et al. (this volume), suggests that the ocean/continent transition was formed by widespread exposure of the upper mantle due to detachment faulting during rifting, followed by mechanical extension of the exposed mantle at the end of rifting.

The Ocean/Continent Transition as a Zone of Very Slow Seafloor Spreading

Hypothesis 1. The Leg 149 transect west of the Site 901 basement high is underlain by very slow spreading oceanic crust.

An important result of the leg, which is not explicitly mentioned in the results of individual sites, is that in-place basement rocks of unequivocal continental crustal character were not recovered during the leg. Whereas there are a few small examples of arkose and mica schist (suggesting a continental crustal source) in debris-flow deposits over basement, the rocks recovered during Leg 149 that bear on the issue of crustal type are almost entirely serpentinized peridotite, gabbro, and bits of basalt. A few of the gabbro and basalt clasts in debris flows at Site 899 are described as having transitional to alkaline affinities, but the vast majority of the mafic rocks recovered show no trace of a continental origin. The simplest interpretation of this observation is that there is little, if any, continental basement around. Starting with this premise, we have tried to develop the hypothesis that all or most of the Leg 149 transect is underlain by oceanic crust. This hypothesis is contrary to the geophysical model upon which Leg 149 was based. We will deal first with the petrological data from basement rocks at Sites 900, 897, and 899. We will then look at the existing geophysical data. We find that all of the existing data, with the possible exception of magnetic anomaly data, are consistent with this hypothesis.

The Site 900 metagabbro seems to have been formed as a cumulate rock from a MORB melt (Seifert et al., this volume). Rocks like these form most commonly as middle to lower oceanic crust at mid-ocean ridges. The Nd isotope evidence for a seafloor spreading origin for the rocks is very strong (Seifert et al., this volume). This makes their emplacement through a screen of extended continental crust or as a continental crust underplate possible, but perhaps less likely. Cornen et al. (this volume) emphasize slight deviations of the rock from that which would be emplaced at a mid-ocean ridge, mentioning slight transitional and alkaline affinities and comparing the rock to those of the Galicia Bank margin that have been interpreted to have been produced by a MORB source interacting with continental mantle. The latter comparison seems at odds with the Nd isotope results from Site 900 (Seifert et al., this volume). The Site 900 rock is basically oceanic in its petrology.

The drilling at Sites 897 and 899 indicates that we are dealing with multiple exposures of peridotite at the seafloor. Although basement at and around Site 899 has not been adequately imaged, the existing seismic data (Sawyer, Whitmarsh, Klaus, et al., 1994) do suggest that Site 899 is located over an isolated basement high. Although we did not drill in-place peridotite basement, we did drill through landslide or mass-flow deposits of virtually 100% peridotite containing large boulders. Gibson et al. (this volume) interpret this to mean an outcrop of peridotite basement must have been nearby. That Site 899 is located at the edge of an isolated positive magnetic anomaly (Miles et al., this volume), probably signifies that the peridotite exposure near Site 899 is local and does not parallel the margin. It does not, however, change the conclusion that, contrary to prior interpretations, peridotite was exposed at the seafloor away from the peridotite ridge, and an adequate model for the formation of this margin must take it into account. It is also significant that the peridotites recovered at both sites are petrologically similar (Cornen et al., this volume), suggesting that they were derived and emplaced at the seafloor from similar sources and by a similar sequence of pressure-temperature conditions.

If the Leg 149 transect is underlain by oceanic crust, then the spreading rate can be constrained to average $\leq$ 6.8 mm/yr half rate. This is based on (1) a seafloor-spreading interpretation of magnetic anomalies that dates the oceanic crust immediately west of the peridotite ridge at about 127 Ma (Whitmarsh et al., this volume), (2) the Tithonian (about 146 Ma) paleontological date for the deepest sediments drilled at Site 901, and (3) the distance between the two locations, 130 km. The rate is an upper bound because the crust at Site 901 had to have been formed prior to the age of the deepest sediment drilled. How much before is unknown because we were unable to core to basement at Site 901. Given the thickness of sediment below the terminal depth of Hole 901 A, we suspect that a rate of about 5 mm/yr is probably more likely. Seafloor spreading at half rates under 10 mm/yr is considered very slow. Seafloor spreading at rates between 10 and about 70 mm/yr is considered slow. Seafloor spreading at rates above about 70 mm/yr is considered fast.

In the following paragraphs we will summarize the characteristics of very slow spreading oceanic crust. This summary is based primarily on Cannat (1993), who studied the emplacement of mantle rocks at the seafloor at mid-ocean ridges globally, and Srivastava and Keen (1995), who studied the extinct mid-ocean ridge in the Labrador Sea. These environments are unambiguously oceanic and are at the slow-spreading end of the spectrum.

Slow-spreading oceanic crust is characterized by exposures of mantle peridotite and gabbro, in addition to basalt, at the seafloor (Cannat, 1993). Such exposures have been identified at slow or very slow spreading ridges in the North, Equatorial, and South Atlantic spreading centers, and at the Southwest Indian Ridge, the Cayman Trough, and the Galapagos rift. Cannat (1993) identifies the critical characteristic to be the lack of adequate magma rising from the asthenosphere to fill the gap between the spreading plates. Bown and White (1994) have shown that when seafloor spreading occurs faster than about 10 mm/yr half rate, there is adequate magma produced by decompression melting of the asthenosphere to form a normal thickness oceanic crust. In this spreading regime, the crustal thickness is remarkably uniform and independent of spreading rate, because the amount of melt is proportional to the space created by spreading. When spreading slows below about 10 mm/yr half rate, significant cooling of the rising magma can occur and the amount of melt becomes less than proportional to the amount of spreading. This creates the magma-starved seafloor spreading environment described by Cannat (1993; Fig. 6). Some of the divergent relative motion between the spreading oceanic plates is accommodated by the intrusion and extrusion of rising melt. The melt cools to form gabbro pods and a thin basalt upper crust. In normal oceanic spreading, the gabbro would be intruded into a screen of older gabbro. In the magma starved ridge model, the gabbro is often intruded into a screen of mantle peridotite. The remaining divergent relative motion between the spreading oceanic plates is accommodated by mechanical extension of the oceanic lithosphere. This means that there will be a regional tendency to passively uplift gabbro and upper mantle rocks. This extension process also faults and refaults the portion of the magma starved lithosphere above the brittle-ductile transition. In Cannat's (1993) model, this produces rotated fault blocks reminiscent of continental rifting. This faulting will often exhume gabbro and upper mantle peridotite exposing it at the seafloor. These rocks will show indications of brittle and/or ductile shearing that occurred during their uplift and faulting induced exhumation. Because of the slow spreading rate, there will be significant conductive cooling of the lithosphere and upwelling asthenosphere. This means that there will be a thicker axial lithosphere than at a faster spreading center and the brittle/ductile transition will be deeper. Thus we would expect to see evidence for brittle shearing at depths normally found more commonly in continental rifts. We also would expect to find that rising melt could cool and solidify at depths greater than typical for normal spreading rate ridges. The mantle rocks exposed at a slow spreading ridge are expected to be relatively undepleted because less melt was extracted during their decompression than at a normal ridge.

Many of the Leg 149 drilling observations are consistent with this model. In this model, it is no surprise that we would observe serpentinized peridotite exposures at Site 899 as well as Site 897. If the transect to the west of Site 901 were underlain by slow spreading oceanic crust, then we might expect a rather random distribution of peridotite, gabbro and basalt exposures in the basement. Some exposures might take the form of margin-parallel ridges, while others might not. In this hypothesis the peridotite ridge has little or no special significance. It no longer necessarily forms a boundary between oceanic crust and extended continental crust, but is just one of perhaps many seafloor peridotite exposures within slow spreading oceanic crust. The peridotite ridge continues to be an interesting feature, however, because no similar features have been observed in slow spreading oceanic crust elsewhere. The peridotites at Sites 897 and 899 were emplaced at shallow levels in a relatively cool thermal field that did not allow significant partial melting (Cornen et al., this volume). This is quite consistent with the slow seafloor spreading hypothesis. The gabbro at Site 900 is also consistent with this model. If formed at a very slow spreading center, the Site 900 rocks could have been emplaced as a gabbro body in a country rock of peridotite and gabbro at a depth of about 13 km. It was subsequently brought upward in a ductile shear zone (Event 1) producing the "flaser" texture and remained at a shallower depth to acquire (Event 2) its lower amphibolite to greenschist facies metamorphic grade, and then brought to the seafloor by block faulting (Event 3) at 136.4 ± 0.3 Ma. It remained exposed at the seafloor, subject to mass wasting (Event 4), for tens of millions of years before it was buried by continent-derived sediment.

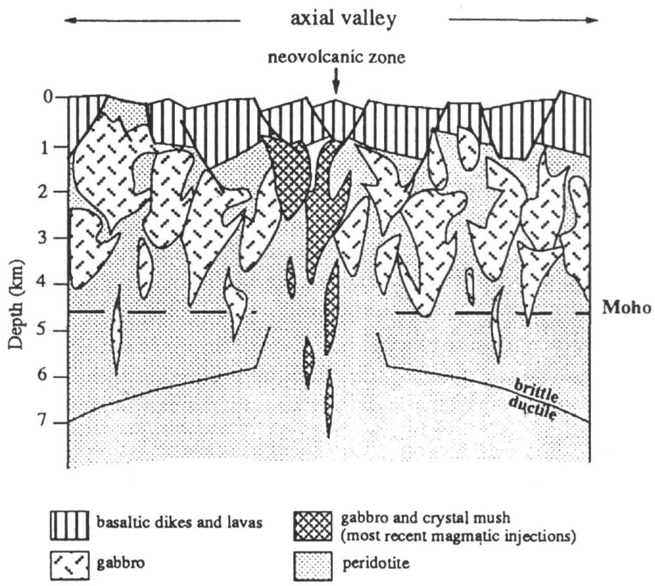

Figure 6. Sketch showing an interpretive geological cross section of the axial valley floor at a magma-starved, rifted mid-ocean ridge. Moho depth and brittle ductile transition are constrained by seismic refraction (Purdy and Detrick, 1986), gravimetric (Morris and Detrick, 1991), and microseismicity (Toomey et al., 1985; 1988) data from the MARK area. Peridotite screens surrounding the gabbro intrusions may be partially serpentinized. This sketch does not show deformation features, except for a few normal faults in the dikes and lavas. Ductile and brittle shear zones due to the extension of the axial lithosphere may be expected to significantly alter this simplified picture, adding structural discontinuities to the magmatically discontinuous crust represented here. The figure and caption are from figure 4 of Cannat (1993).

Very slow spreading oceanic basement is typically rougher than normal oceanic crust (Cannat, 1993; Srivastava and Keen, 1995). It is rough because it has been mechanically extended and the extension has been accommodated by normal faulting and block rotation. Srivastava and Keen (1995) interpreted a seismic reflection profile across the now extinct spreading center in the central Labrador Sea. The youngest oceanic crust (Fig. 7) was formed at a minimum half rate of about 3 mm/yr, has a highly faulted basement surface, appears to have been mechanically extended by about 70%, and is about 3 km thick (Srivastava and Roest, 1995; Osier and Louden, 1992, 1995). The older oceanic crust at each end of the line shown in Figure 7 was formed at about 10 mm/yr half rate, has a relatively unfaulted basement surface, appears to have been mechanically extended by only about 15%, and is about 6 km thick. Since the hypothetical spreading rate for the oceanic crust under the Iberia Abyssal Plain transect is about 5 mm/yr, the seismic profile of the 3 mm/yr unequivocal oceanic crust of the central Labrador Sea should give a good idea of the kind of structures that should be observed.

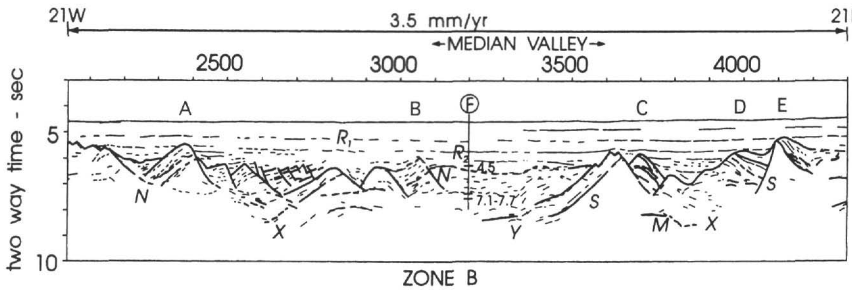

Figure 7. Line drawing of seismic reflection profile crossing the extinct spreading center in the Labrador Sea (Srivastava and Keen, 1995). This crust is interpreted to have formed at 3.5 mm/yr half rate. The structures within this crust may be similar to those seen in the ocean/continent transition of the Iberia Abyssal Plain. F is the extinct spreading center. R_1 and R_2 are sedimentary reflectors. N, S, X, and Y are intracrustal or upper mantle reflectors. M is a possible Moho reflector. Locations A-E are tilted fault-bounded blocks capped with prefaulting layered material originally deposited as flat-lying sequences.

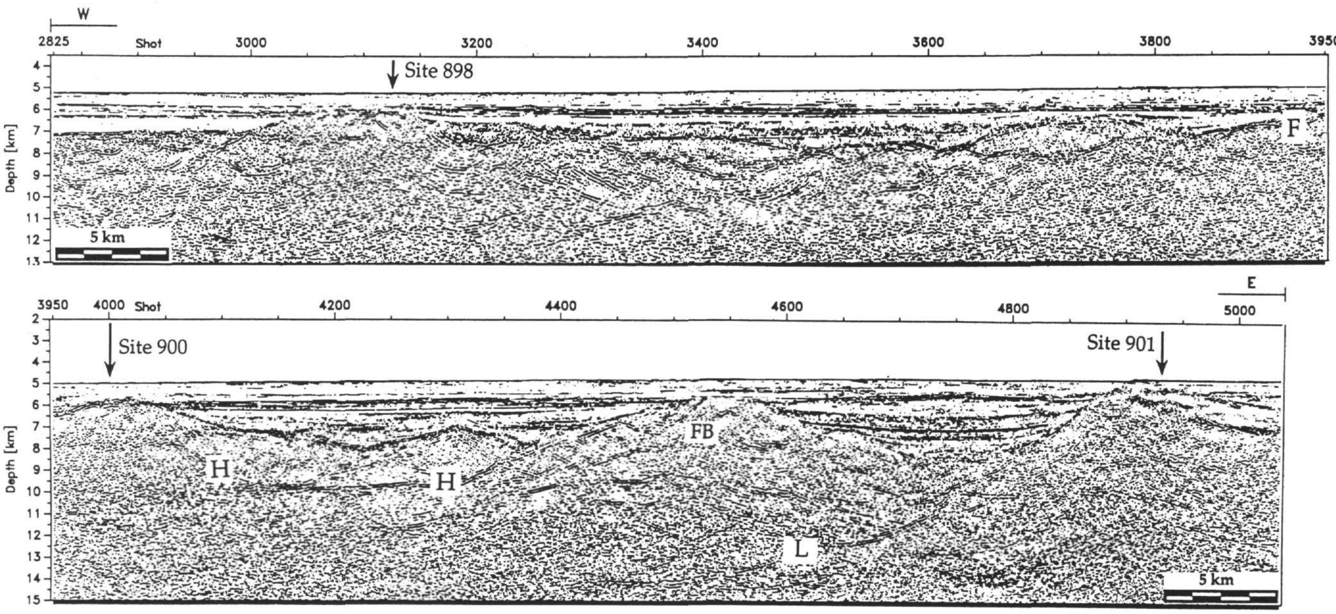

Figure 8. Depth-migrated section of profile LG-12 between Sites 898 and 901 (Krawczyk et al., this volume; 50-m shot spacing; no vertical exaggeration). Significant fault structures, H, L, and F are labeled. L bounds the Site 901 basement high on the west. Krawczyk et al. (this volume) interpret H to be a low-angle detachment fault that was active late in the rifting process.

We find that many features of the Iberia Abyssal Plain transect (Figs. 3, 8; Krawczyk et al., this volume; Whitmarsh et al., this volume), particularly the region between Site 901 and Site 898 along Profile LG-12 (Beslier, this volume), are similar to those seen in Figure 7. The basement in both areas consists of normal-fault-bounded, rotated, blocks of crust. The basement relief in both areas is about 1.0-1.5 s two-way traveltime. The larger blocks in both areas have a typical width of 8-12 km. Most of the faults dip toward the extinct spreading axis in the Labrador Sea and, on Profile LG-12 in the Iberia Abyssal Plain, dip toward the west. We initially argued that the presence of rotated fault blocks was strong evidence for extended continental crust. The Labrador Sea analog draws that argument into question by showing that rotated fault blocks are also a characteristic of very slow spreading oceanic crust. There are also regions of the Iberia ocean/continent transition, particularly to the east of the peridotite ridge, where block faulting is not obvious because the basement is relatively smooth (Fig. 8). In the context of very slow seafloor spreading, these regions are probably areas that have been affected by less mechanical extension and more vigorous magmatism.

Very slow spreading oceanic crust is thinner than normal oceanic crust, may not have normal oceanic crust seismic velocities, and may not have a well-developed Moho (Cannat, 1993). The characteristic seismic velocity structure of oceanic crust is produced by the layering of basalt over gabbro over the upper mantle. At magma-starved ridges, this layering does not develop in the normal way. Because upper mantle rocks are brought to the surface in some places, and form a screen for the basalt and gabbro elsewhere, there are not likely to be clear layers 2 and 3. Seismic velocities in the crust may be laterally variable and are unlikely to match those of normal oceanic crust. There will probably not be a clear Moho because the base of the crust is characterized, not by a fairly sharp boundary between gabbro and peridotite layers, but by a gradual upward increase in the ratio of gabbro to peridotite (Fig. 6). Because the magma supply is low and much divergent motion is accommodated by rifting and thinning the already formed oceanic crust, where you can define a Moho, the crustal thickness is likely to be less than that of normal oceanic crust (Bown and White, 1994).

Osier and Louden (1992) reported seismic refraction results from the region of the reflection data shown in Figure 7. Their refraction data in the region of 10 mm/yr seafloor spreading (the crust just older than that shown in Fig. 7) show crust 4-6 km thick, only slightly thinner than normal oceanic crust (7 km; White et al., 1992). In this crust there are clear layers 2 and 3 with velocities of 5.1-6.2 km/s and 6.9-7.6 km/s respectively. Their refraction data from the region of 3 mm/yr spreading (Fig. 7) suggest the presence of thin (3-4 km), low-velocity crust (3.6-4.5 km/s) and a low-velocity upper mantle (with an average velocity of 7.7 km/s). Srivastava and Keen (1995) noted that the layer thicknesses and velocities are highly variable and that upper mantle velocities as low as 7.1 km/s were observed in the slowest spreading region. This kind of variability is quite consistent with the description of very slow spreading oceanic crust by Cannat (1993).

The seismic refraction data along the Leg 149 transect (Whitmarsh et al., 1990; Fig. 3B) show many of the characteristics of very slow spreading oceanic crust. The crust in Figure 3B identified as transitional crust varies from 2.5 to 4 km in thickness. This is thinner than normal oceanic crust and thinner than most unequivocal extended continental crust. Conversely, it is in the range of thicknesses observed in very slow spreading oceanic crust. There is a layer of low velocity upper mantle (7.4-7.55 km/s along Whitmarsh et al.'s Line 2) under the area of disputed crust type interpreted to be serpentinized peridotite (Whitmarsh et al., 1990; Fig. 3B). A similar layer is seen in the refraction data from the very slow spreading oceanic crust in the Labrador Sea. The refraction data west of the peridotite ridge (Whitmarsh et al., 1990; Fig. 3B) also show a layer of low velocity upper mantle. The crust there is 3 km thick, the crust has velocity 4.0-5.5 km/s, and the upper mantle, also interpreted to be serpentinized peridotite, has velocity 7.0-7.6 km/s. On the basis of these refraction data, the crust in both areas seems to have the thickness predicted for, and observed in, very slow spreading oceanic crust.

The magnetic observations in the Iberia Abyssal Plain are not as neatly explained by the very slow spreading oceanic crust hypothesis as by the second hypothesis presented below. Effective magnetic stripes are probably not created during very slow (less than 5 mm/yr, see below) seafloor spreading because the addition of magma to the crust is discontinuous and irregular in space. That the anomalies in the region between Sites 901 and 897 are of low amplitude and lineated is generally consistent with the very slow spreading oceanic crust hypothesis. However, Whitmarsh and Miles (1995) were unable to match observations within the ocean/continent transition using a magnetic model using very slow spreading rates. This is a problem for this hypothesis because very slow spreading anomalies have been successfully modeled elsewhere (southern Australia margin at 5 mm/yr, Mutter et al., 1985; southern Red Sea at 5-11 mm/yr, Roeser, 1975; Miller et al., 1985; southern Aegir Ridge in the Norwegian Sea at 5 mm/yr, Nunns, 1983).

Very slow spreading oceanic crust in the Iberia Abyssal Plain is probably bounded in the east by extended continental crust under Site 901. The Site 901 basement block is constrained to be continental because we do not think that very slow spreading ocean crust would form at 320 m water depth. We do not see equally strong evidence that any basement west of the Site 901 high is continental. We have not yet explored the plate tectonic reconstruction implications of a >130-km-wide region of ultraslow oceanic crust under the Iberia Abyssal Plain.

Very slow spreading oceanic crust in the Iberia Abyssal Plain is probably bounded in the west by a spreading rate change several tens of km west of the J anomaly. The peridotite ridge is often interpreted to be an important crust type boundary and would be a natural place to argue for an increase in oceanic spreading rate. Whitmarsh et al. (this volume) have interpreted magnetic anomaly data between the peridotite ridge and the J anomaly to indicate seafloor spreading at 10 mm/yr half rate. Before accepting this conclusion, we should look carefully at the oceanic crust to the west of the peridotite ridge. Constraint on the structure of the oceanic crust to the west of the peridotite ridge comes from seismic refraction data (Whitmarsh et al., 1990) and seismic reflection profiles. Seismic refraction line L3 was the basis of the portion of Figure 3B just west of the peridotite ridge. Whitmarsh et al. (1990) found anomalously thin oceanic crust with low velocity upper mantle that they interpreted as serpentinized peridotite. This could also be interpreted as very slow spreading oceanic crust (thin crust with a low velocity upper mantle; Srivastava and Keen, 1995). In Figure 3B the thin crust abuts normal thickness crust to the west of the refraction profile, but the location is constrained using only gravity data (Whitmarsh et al. 1993). The seismic reflection profile Sonne-16 extends westward from the peridotite ridge at Site 897. There it crosses two more basement highs with relief of about 1-1.5 s two-way traveltime and spacing of 15-20 km peak to peak (Fig. 9). This is greater relief and a shorter wavelength than observed by Srivastava and Keen (1995) in the Labrador Sea in areas where spreading rate was faster (10 mm/yr half rate). We suggest that there may be a strip of oceanic crust west of the peridotite ridge that was produced by very slow spreading. The rate may have increased gradually to normal Atlantic spreading rates, or there may be an abrupt boundary. A good possibility for the timing of this acceleration is the time at which continental breakup is estimated to have occurred, and hence seafloor spreading to have begun, on the Galicia Bank margin to the north (about 112 Ma).

Following the logic of this hypothesis, we find no equivalently compelling reasons to assert that any crust to the west of the Site 901 basement high is extended continental crust. Krawczyk et al. (this volume) argue that the seismic reflection data show evidence for detachment faulting in the Site 900 high. We believe that this observation is equally consistent with the very slow seafloor-spreading hypothesis in that there was undoubtedly a large amount of mechanical extension of the newly formed oceanic crust. The extension was probably accommodated by the same kinds of faulting/shearing as is found in continental crust.

Beslier et al. (this volume) present a model for the formation of the Iberia Abyssal Plain margin that is somewhat similar to our very slow seafloor-spreading hypothesis. As in this hypothesis, they argue that the Site 901 high is probably the western edge of extended continental crust. In their model, simple shearing late in the rifting process exposed a broad region of ultramafic and mafic rocks at the seafloor to the west of the Site 901 basement high. The region of denuded upper mantle was then mechanically extended to its current width of about 150 km. They suggest that the denuded mantle was stretched during the last stages of continental rifting offshore Galicia Bank. Our hypothesis differs only (1) in explaining how the mantle is denuded in the first place (we do not invoke regional simple shear) and (2) in the degree of melt generation (we call for some melt to make the Site 900 gabbro and the various gabbros and basalts found in mass-flow deposits). The hypotheses agree as to the location of the edge of extended continental crust, the presence of peridotite in a region rather than confined to a single ridge, the importance of mechanical extension and faulting within the lithosphere to the west of the edge of continental crust, and the possible temporal association of the extension landward of the continental edge in the Iberia Abyssal Plain with the continuing rifting on Galicia Bank.

The Ocean/Continent Transition as a Zone of Tectonically and Magmatically Disrupted Continental Crust

Hypothesis 2. Most of the region between Site 897 and Site 901 is underlain by tectonically and magmatically disrupted continental crust.

The evidence that the ocean/continent transition is a zone of tectonically and magmatically disrupted continental crust stems from surface and near-seafloor magnetic anomalies, from our interpretation of a margin-parallel peridotite basement ridge, and from the interpretation of multichannel seismic reflection profiles. A discussion of the Leg 149 results, in the context of this hypothesis, is presented after describing the evidence for it.

Whitmarsh and Miles (1995) proposed a model to explain geophysical observations over the west Iberia ocean/continent transition that involves a propagating rift which, by means of faulting of the continental crust followed by magmatism (intrusive and possibly extrusive) and eventual breakup, leads to the ocean/continent transition as we find it today. They recognized three zones of magnetic anomalies on a chart of the west Iberia Margin (Fig. 10). West of the peri-

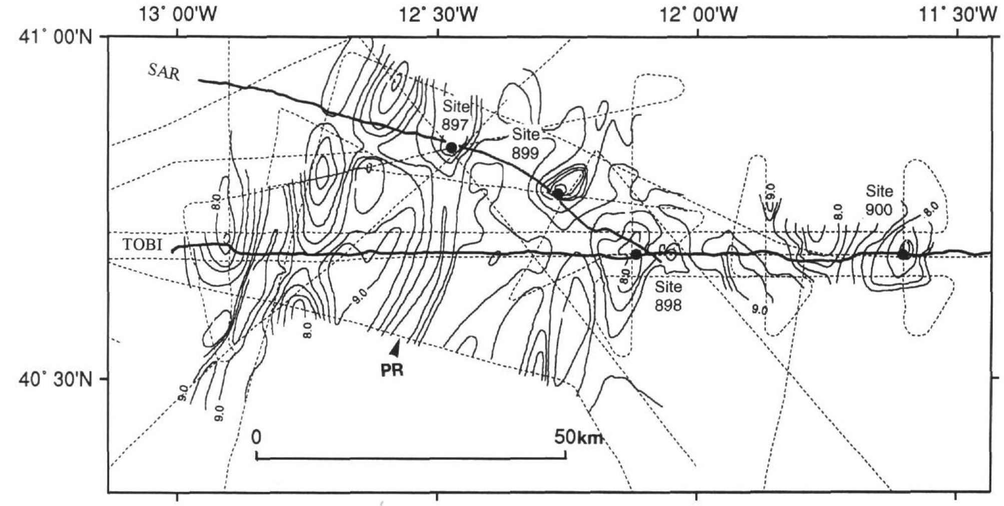

Figure 9. Basement relief in the Iberia Abyssal Plain contoured in two-way traveltime (s) from all available seismic reflection profiles (dotted lines; from C. Krawczyk, P.R. Miles, L.M. Pinheiro, R.B. Whitmarsh, unpublished data). SAR and TOBI tracks are shown by the bold east-west and west-northwest-east-southeast lines, respectively. ODP sites are indicated by solid circles. PR = peridotite ridge.

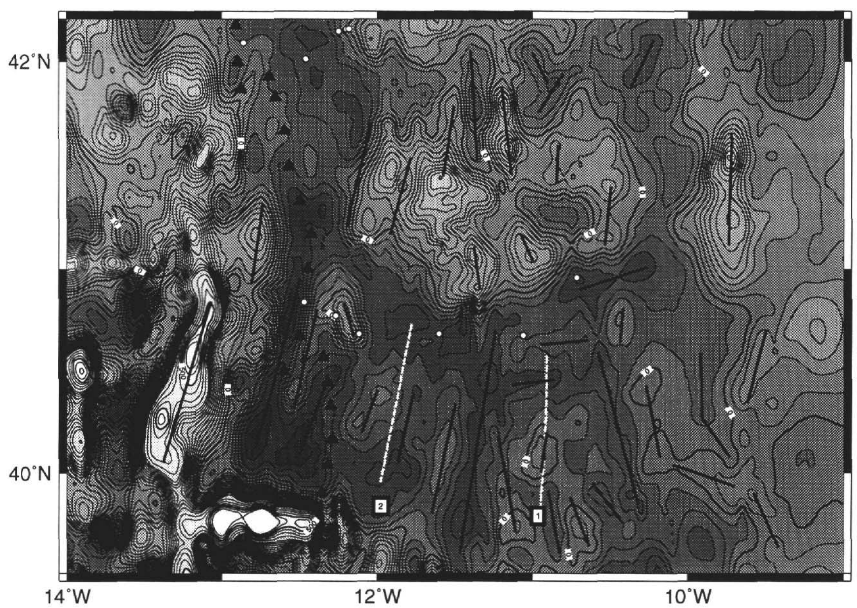

Figure 10. Reduced-to-the-pole magnetic anomaly chart of west Iberia (25 nT contours between +300 and -300 nT, otherwise 50 nT) from Miles et al., this volume. Magnetic highs/lows are light gray/dark gray, respectively. Shading scheme and contours have been chosen to emphasize the smaller amplitude anomalies. Thick lines = principal linear features. White dots = DSDP/ODP sites. Triangles = predicted basement peridotite ridge (Beslier et al., 1993). White lines = seismic refraction lines of Whitmarsh et al. (1990). From Whitmarsh and Miles (1995).

dotite ridge (evidence for the existence and continuity of this ridge is presented below) they recognized, and modeled, M-series seafloor spreading anomalies. East of the ridge they found linear 015°-trending magnetic anomalies in the ocean/continent transition; these anomalies parallel the seafloor-spreading anomalies further west but could not be explained using a constant spreading rate and the Mesozoic geomagnetic reversal timescale. A major premise of their model is that these anomalies are the result of synrift intrusions in the lower continental crust under the same stress regime that later determined the direction and nature of faulting within the oceanic crust. Further east, east of 11°15'W, the more elongate linear anomalies trend roughly east of south (about 165°) but in addition there are other shorter features with a variety of trends. The anomalies in this easternmost zone, which includes Site 901 and other tilted fault blocks that extend landwards beneath the continental slope, and which is assumed to consist of stretched continental crust, are less well organized and the east-of-south trends differ from those in the central zone by about 30°. The Whitmarsh and Miles (1995) model in Figures 11 and 12 is presented in the form of a sketch that stresses the propagation of a continental rift and the implied transition from the almost amagmatic rifting of a nonvolcanic margin to the predominantly magmatic process of seafloor spreading itself at a half-rate of about 10 mm/yr. In the model the semicontinuous peridotite ridge found off the northwest part of the west Iberia Margin is attributed to tectonic exposure, by a so-far unexplained mechanism, of upper mantle material along the line of continental breakup.

We emphasize the evidence for an almost continuous linear peridotite ridge off west Iberia and its relationship to the ocean/continent

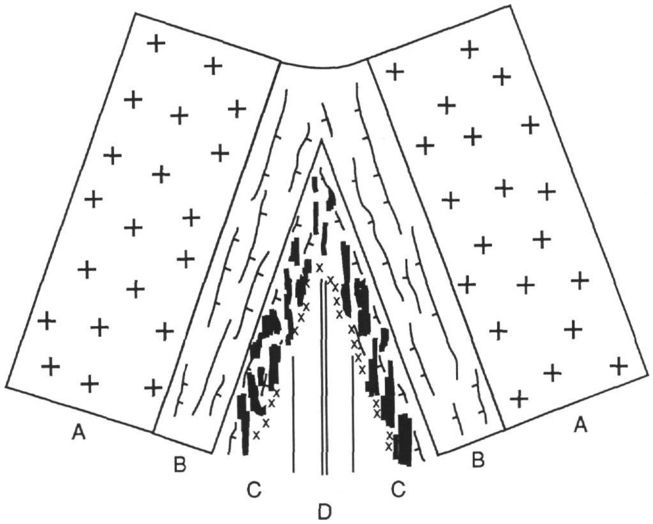

Figure 11. Sketch (not to scale) of a propagating rift model from Whitmarsh and Miles (1995) to illustrate Hypothesis 2 presented here. The model includes (A) unthinned continental crust; (B) continental crust thinned by tectonic extension with upper crust normal faults; (C) a transitional crust consisting 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. Crosses indicate outcrops of mantle ultramafic rocks (peridotite ridge).

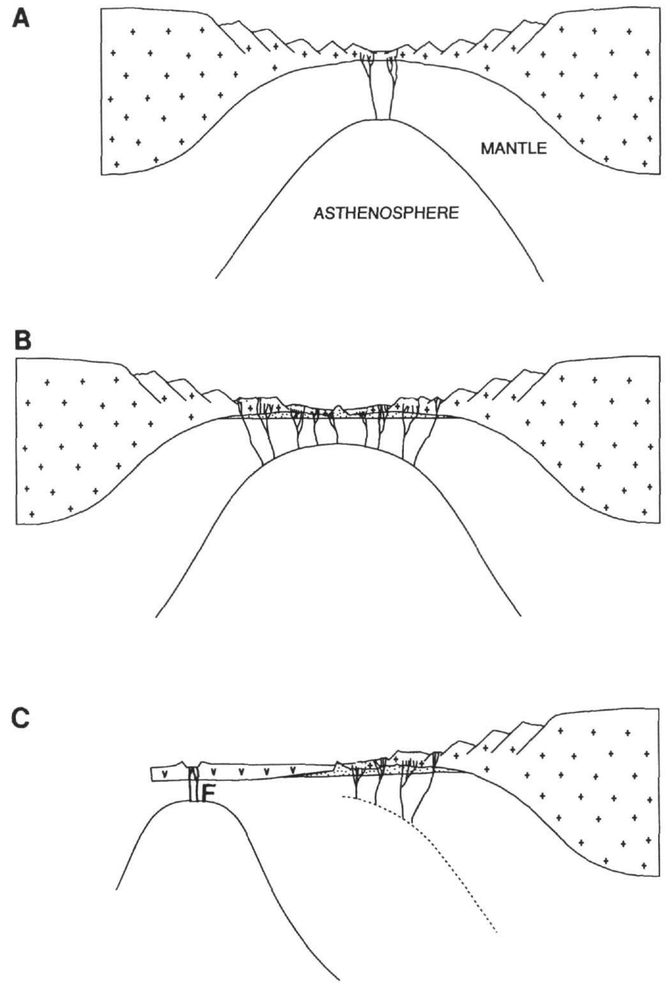

Figure 12. Schematic cross sections (not to scale) illustrating the development of the model shown in Figure 11 (from Whitmarsh and Miles, 1995). **A.** Just prior to continental breakup; crosses indicate continental crust. The first products of asthenospheric decompressional melting penetrate the upper crust. **B.** At the point just after continental breakup where the top of the asthenosphere is shallow enough to allow igneous intrusion of the continental crust to occur over a broad zone. Dotted region represents serpentinized peridotite. **C.** After the point where igneous activity has become sharply focused and steady-state seafloor spreading has been reached, after initially producing an abnormally thin oceanic crust (v symbol).

transition, oceanic crust, and continental crust. North of Gorringe Bank (Fig. 2), serpentinized peridotite has been dredged, sampled by submersible or drilled at seven locations off west Iberia between latitudes 40°46' and 43°N (Boillot et al., 1988a,b; Sawyer, Whitmarsh, Klaus, et al., 1994). Off Galicia Bank, the samples come from the relatively steep west and northwest flanks of the bank and from a narrow (10-20 km wide) ~000°-trending basement ridge traced for over 115 km in a north-south direction on seismic reflection profiles that cross the ridge about every 5 km (Thommeret et al., 1988).

Beslier et al. (1993) used other, less closely spaced, seismic profiles to propose that the above two ridges were part of a 300-km-long feature offset, principally sinistrally and in en echelon fashion, into three or four separate segments. In the southern Iberia Abyssal Plain, the ridge has been traced for over 40 km on seismic reflection profiles (Fig. 9). Here, even though the ridge is directly sampled only at Site 897, its recognition is assisted by its general association with overlying folded sediments; it appears that the peridotite ridge acted as a weak zone during the Miocene northwest-southeast compressional episode (Masson et al., 1994).

Added evidence for the significance of the west Iberia peridotite ridge comes from other geophysical observations and modeling. First, based on modeling deep-tow magnetic profiles a very clear contrast exists between the bulk magnetization of the crust either side of the peridotite ridge both off Galicia Bank (Sibuet et al., 1995) and in the southern Iberia Abyssal Plain, although less abruptly there because the deep-tow profile crosses the strongly magnetic, and possibly unique, Site 899 basement high (Whitmarsh et al., this volume). In both cases the landward crust possesses a significantly weaker magnetization. Second, seismic velocity structures either side of the peridotite ridge also appear to differ significantly (Whitmarsh et al., 1990; Horsefield, 1992; Whitmarsh et al., work in progress) and have been interpreted as thin oceanic crust and ocean/continent transition crust. The thin oceanic crust is explained by a poor magma supply caused by the cooling effect of continental lithosphere (Whitmarsh et al., 1993) or by seafloor spreading that has immediately followed prolonged continental rifting, which has the same effect (Bown and White, 1995). Once continental rifting has ceased, seafloor spreading rapidly (in a few m.y.) reaches a steady state and, provided it exceeds 7.5 mm/yr half-rate (Bown and White, 1994), normal thickness oceanic crust is produced. Last, in the southern Iberia Abyssal Plain there is a distinct change in the form of acoustic basement either side of the peridotite ridge. To the west the basement consists of parallel ridges and troughs trending slightly east of north whereas to the east it consists of more or less isolated highs that have a weak northerly trend and that rarely have a length-to-width ratio of more than 3:1 (Fig. 9 and unpubl. data collected summer, 1995).

Irrespective of the origin of the ocean/continent transition crust, there is compelling evidence from independent observations of its magnetization, seismic velocity structure, and upper surface that it differs significantly from the thin oceanic crust further west and that, wherever measurements have been made off west Iberia, the peridotite ridge itself is closely associated with the boundary between the two crustal types.

Although there is evidence for a more or less continuous margin-parallel linear peridotite ridge from about 40° to 43°N, we note that the west Iberia Margin is possibly unique in possessing such a ridge of peridotite; no such ridge has yet been convincingly demonstrated on the conjugate Newfoundland margin. Peridotite has been sampled off other rifted margins in the approximate vicinity of the ocean/continent transition (Nicholls et al., 1981; Bonatti et al., 1986) but only rarely in circumstances that could be described as similar to west Iberia (Site 651 in the Tyrrhenian Sea, Kastens et al., 1988). This may simply be because on no other rifted margin is the ridge so accessible and/or the margin so thoroughly surveyed.

In spite of the strong evidence for the peridotite ridge, it is difficult to explain how it was emplaced. An explanation in terms of a major low-angle detachment fault, such as has been postulated beneath Galicia Bank (Boillot et al., 1987), has yet to receive support from reflection profiles across Site 897. Whitmarsh and Miles (1995) invoked tectonic emplacement at continental breakup by an "as yet unknown mechanism." S. Pickup (pers. comm., 1995) has recognized a 30° landward-dipping reflector beneath the southern Iberia Abyssal Plain, south of the Leg 149 transect, which intersects the east side of the peridotite ridge and can be traced to a depth of at least 12 km. It is possible that the peridotite ridge is associated with the final breakup of continental crust immediately preceding the onset of seafloor spreading but we cannot at present explain its continuity or long narrow ridge-like character. The evidence of limited melting of the peridotite implies that emplacement was slow (to inhibit significant adiabatic melting) and/or affected by the adjacent cold continental lithosphere.

We now consider how the results of Leg 149 fit hypothesis 2. Site 897 was drilled on the crest of the peridotite ridge and provided the first direct evidence that this ridge exists within the southern Iberia Abyssal Plain. The peridotite cores from this site, and the cores from Sites 637 (drilled on the peridotite ridge off Galicia Bank), all indicate a very similar early pressure-temperature history (crystallization at up to 30 km depth and 1200°C and ductile deformation at around 1000°C) and hence, presumably, point to a similar mode of emplacement of these rocks. Beslier et al. (this volume) describe the emplacement as happening during lithospheric stretching as adiabatic uplift of a mantle dome which was subsequently tectonically exposed at the seafloor. This view is entirely consistent with the Whitmarsh and Miles (1995) model (Fig. 12). The greater mylonitization of the Site 637 cores may simply be the chance result of the borehole intersecting a zone which had experienced extensive shearing while Site 897 did not.

Site 899 was unusual in that it penetrated a subcircular basement high that, on the basis of a surface magnetic anomaly chart of the west Iberia Margin (Miles et al., this volume, Fig.1 [foldout in back pocket]), may be in a location that is unique within the Iberia Abyssal Plain ocean/continent transition; the chart shows the site to lie on the west flank of a north-northwest–south-southeast isolated steep-sided positive magnetic anomaly of over 125 nT amplitude, quite different from the generally lower amplitude and more gently sloping anomalies elsewhere in the ocean/continent transition. A similar conclusion emerges when a deep-towed magnetometer profile across the site is studied; this shows an 8-km-wide ~600 nT positive anomaly over Site 899, which is far larger than the weaker anomalies immediately to the east (Whitmarsh et al., this volume). Modeling of the deep-tow profile indicates that the bulk magnetization of the crust is unusually high, higher than oceanic crust, in the vicinity of Site 899 and this is supported by measurements on the cores (Zhao, this volume; Whitmarsh et al., this volume) which indicate that the Site 899 peridotites are on average five times more strongly magnetized than the Site 897 peridotites. Since the amplitudes of magnetic anomalies are strongly correlated with the iron oxide mineralogy of the (near surface?) crustal rocks, which in this case appear to be ultramafic, we caution against extrapolating the results from this site to the ocean/continent transition in general and, in particular, against assuming that there is a continuous peridotite basement between Sites 897 and 899, which are 20 km apart. The question whether in fact the Site 899 cores are representative of the underlying basement, or of more distant location(s), is addressed later. If the underlying basement is ultramafic, one explanation could be that the site is located at the northern end of a peridotite ridge segment that overlaps, en echelon fashion, the segment on which Site 897 is situated (Beslier et al., 1993). However, we are unable to explain why the Site 899 basement should be more strongly magnetized and the strong magnetization may have an independent origin. For example, the greater lherzolite content implies more Fe and hence, possibly, more magnetic minerals. Nevertheless, a similar mode of peridotite emplacement to Sites 637 and 897 is indicated by a very similar early pressure-temperature history (crystallization at up to 30 km depth and 1200°C and ductile deformation at around 1000°C) and hence, presumably, to a similar mode of emplacement of these rocks.

The basement we drilled at Site 900 is massive flasered (sheared) cumulate gabbro with a primary mineralogy indicating metamorphism at at least 0.4 GPa (~13 km depth) and REE patterns characteristic of a transitional MORB parent magma (Cornen et al., this volume; Seifert et al., this volume) and Nd and Pb isotope ratios indicative of a MORB parent magma (Seifert et al., this volume). The gabbro has many compositional and textural similarities to oceanic cumulate gabbros (Seifert et al., this volume). While the Site 900 Nd isotope data are consistent with formation of the gabbro from such a parent magma, on the other hand, rocks generated from a single source generally have a narrow range, say 0.5, of epsilon Nd; the larger range found here (6.3-10.3) may be indicative of some mixing of sources (e.g., MORB and continental crust). Further, it is interesting that Schärer et al. (1995) studied a gabbro and an associated chlorite rock (dated as synrift) from the western margin of Galicia Bank that had small positive epsilon Nd values in the range 3.6 to 5.6. They concluded that the gabbro crystallized beneath stretched continental crust prior to continental breakup between Iberia and Newfoundland and was emplaced as synrift underplated material. Thus, a range of epsilon Nd values from 3.6 to 10.3 has been observed off west Iberia, suggesting perhaps a varying degree of continental contamination of the synrift melt along the margin. An absolute age of 134.6 Ma for the last thermal event experienced by the Site 900 gabbro indicates its tectonic deformation, and probable crystallization, during synrift time (Féraud et al., this volume). We therefore conclude that the Site 900 gabbro may also have been the product of synrift underplating.

Given that Site 897 was drilled on the peridotite ridge, and that Site 899, for reasons already described, appears to lie over crust that is probably atypical of the ocean/continent transition, only Site 900 can be considered to have sampled the basement of the ocean/continent transition itself. Therefore we have to look to Site 900 cores and to the allochthonous clasts found in debris-flow deposits at Site 897 (Subunit IIIB) and at Site 899 (Unit IV) to make some estimate of the nature of the crust exposed within the ocean/continent transition towards the end of rifting.

At Site 897 the conglomerates of Subunit IIIB yield rounded clasts of different lithologies, ages, and degrees of lithification, suggesting that many different lithologies cropped out on the seabed. The lithologies include claystone and chalk/limestone with minor dolomite, basalt, arkosic to lithic sandstones, and a sandstone with fragments of shallow-water fossils, mica schist, basalt, serpentine and mica (Shipboard Scientific Party, 1994a; Comas et al,, this volume). At Site 899, Unit IV contains clasts of mafic rocks (submarine basalts, microgabbros, mafic clasts and amphibolite) which display transitional MORB to alkaline features.

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, the igneous rocks have transitional MORB to alkaline affinities and do not strongly support the idea of a widespread MORB-like oceanic crust in the ocean/continent transition. REE patterns of some basalt clasts are close to CFB, relatively free of contamination by continental lithosphere. 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.

The lithologies present from the southern Iberia Abyssal Plain are rather different from the metamorphosed sediments and granitic rocks that have been sampled around Galicia Bank and the granitic rocks that crop out in the Berlenga-Farilhoes Islands near the head of the Nazaré Canyon (Fig. 1). Black et al. (1964) reported dredged limestones from Vigo Seamount and Galicia Bank but, on the basis of their shape and smooth striated surfaces, attributed many plutonic and metamorphic rocks to glacial erratics. Capdevila and Mougenot (1988) report dredged samples from around Galicia Bank of predominantly metamorphosed sediment (mica schist, gneiss, granulite, Paleozoic sediments(?), phyllite, and metaarkose) and a variety of granitic rocks (granodiorite, granite, tonalite, granophyre), but it is interesting that they do not mention a single sample of gabbro. Although it is not clear what criteria were used by Capdevila and Mougenot to confirm that these samples were not glacial erratics (Davies and Laughton, 1972 show that erratics occur as far south as 30°N in the northeast Atlantic) granite and granodiorite were also sampled, presumably in situ, by submersible at about 12°30'W (Boillot et al., 1988a). Therefore, it seems that the difference between the southern Iberia Abyssal Plain and Galicia Bank (principally a lack of granitic rocks) is real and is probably attributable to the different outcrop geology of the adjacent onshore regions where granite outcrops only north of about 41°N. We therefore conclude that the lack of granitic rocks in cores from the southern Iberia Abyssal Plain part of the west Iberia Margin does not necessarily signify a lack of continental crust there.

A final new piece of evidence is afforded by seismic reflection images of the crust and upper mantle of the west Iberia ocean/continent transition (Krawczyk et al., this volume). Prestack depth migration of the east-west LG-12 profile, which passes over or close to almost all the Leg 149 drill sites, enabled the identification of steep- and low-angle normal faults (labeled H and L in Fig. 8) between Sites 900 and 901. Such signs of extensional tectonics are consistent with, but do not prove, tectonic disruption of continental crust. Of particular significance is the apparent location of Site 900 on the lower plate to an important detachment fault; this provides a means to explain the close to synrift exhumation and retrograde metamorphism of the Site 900 gabbro from around 13 km depth, where it underwent dynamic recrystallization, to the seafloor. Further west, as far as the peridotite ridge., the acoustic basement has a much smoother appearance, on this and other reflection profiles, with relief of under 1 s two-way traveltime. Here we take the smoother basement to be evidence of the growing oceanward importance of extrusive magmatism which lead to the eventual onset of 10-mm/yr seafloor spreading west of the peridotite ridge.

Hypothesis 2 can now be refined in the light of the new ODP data. The most striking new result is the discovery of an apparently synrift MORB-like cumulate gabbro within the ocean/continent transition in a location where continental crust was predicted. Although the original model predicted widespread intrusive and even extrusive material, in or over the continental crust, respectively, it now seems that underplating at the base of the continental crust, or even the emplacement of the gabbro as a layer trapped within the uppermost mantle, is also possible (Beslier et al., this volume). We do not know whether underplating is widespread within the ocean/continent transition but the sort of gabbro drilled at Site 900 could form in local pockets, akin to an aborted form of the punctiform initiation of seafloor spreading proposed by Bonatti (1985) for the northern Red Sea, just as well as in a widespread sheet. In other words, the first synrift magmatism may begin as a series of isolated patches of MORB-like products at the base of the continental crust. Eventually such patches may merge to form a continuous belt which is the forerunner of oceanic crust produced by seafloor spreading. The exhumation of cumulate gabbro from 13 km depth to the top of basement is simply explained by late-stage low-angle detachment faulting (implying significant horizontal extension), which acts at a smaller scale than envisaged in the original model. Why has underplating not been detected by seismic refraction lines? First, it may be discontinuous on the scale of ~80-km-long refraction lines, as suggested above, and therefore hard to detect. Second, even if the underplated gabbros should form a continuous sheet, the seismic refraction method is unlikely to be able to resolve layers at the base of the crust, or within the uppermost mantle, that are thinner than a few hundred meters, and we have no evidence at present that the gabbro is thicker than this.

The presence, within mass-wasting deposits, of basaltic and microgabbroic clasts with transitional to alkaline affinities is to be expected from melts that have passed through, or been contaminated by, continental lithosphere. The minor amounts of mica schist and arkose in the same deposits is evidence of continentally derived material. The absence of granitic rocks in the cores is inconclusive and may have a simple geological explanation linked to the distribution of onshore outcrops.

Whitmarsh and Miles (1995) supported their case for this type of model by reference to analogous, but better exposed, nonvolcanic rifted margin locations. The best active analogue is probably the northern Red Sea which is at least 1500 km from the Afar hotspot. Here Bonatti (1985) envisages a thinned and stretched continental crust injected by diffuse basaltic intrusions and he quotes examples of samples of "...basaltic/gabbro rocks...showing transitional or alkaline affinities" from the Brothers Islands at 26°15'N. Bonatti and Seyler (1987) report that on Zabargad Island at 23°40'N there is a peridotite-silicic gneiss-gabbroic association which is a sample of continental upper mantle/lower crust. The pressure-temperature history of the Zabargad gabbros and gneisses is very similar to that deduced from the Site 900 gabbros. Similarly Pautot et al. (1984) report a basalt from Charcot Deep at 25°15'N with alkaline/transitional affinities. The comparison with the results of Leg 149 is striking and suggests a fundamental similarity between the evolution of the northern Red Sea and the west Iberia Margins.

Finally, a principal factor which led to the original model was the apparent absence of seafloor-spreading anomalies within the west Iberia ocean/continent transition and the presence there of relatively low-amplitude and long wavelength isochron-parallel anomalies. The origin of these anomalies is still debatable but cumulate gabbros, if distributed with a predominant margin-parallel trend, could make an important contribution to such anomalies. The discovery of cumulate gabbro within the ocean/continent transition does not preclude the original model but has allowed for its refinement; the model will receive strongest support, however, should future basement drilling discover evidence of unequivocal in situ continental rocks within the ocean/continent transition.

DISCUSSION

The fundamental problem of the west Iberia/Newfoundland conjugate pair of nonvolcanic rifted margins is one of plate geometry. How can we explain plate tectonic reconstructions that restore the Flemish Cap against Galicia Bank that at the same time further south leave an area hundreds of km wide of unusually thin crust (Fig. 1) which, according to finite element modeling, cannot be explained by any physically reasonable extension of continental crust (Bassi et al., 1993)? The problem cannot be solved by a simple geometrical translation of an independent Galicia Bank or Flemish Cap microplate.

We have presented two hypotheses, neither of which we find to be fully satisfactory, which attempt to solve the problem in the light of all available tectonic, magmatic, and other geophysical information. Hypothesis 1 emphasizes an idealized vertical cross section for oceanic crust formed by ultraslow seafloor spreading and the correspondence of the geological and geophysical evidence with such a cross section. It can explain the existence of gabbro and serpentinized peridotite at the top of acoustic basement, but the basalt that the model predicts should also exist there is found in only minute quantities in the cores. Hypothesis 2 emphasizes the plan view development of rifting along the margin. It can explain the geophysical observations and the existence of gabbro and the peridotite ridge, but has to explain the Site 899 peridotites as atypical of, and even unique to, the southern Iberia Abyssal Plain margin. In the final analysis, both hypotheses, although they approach the problem from different standpoints, tend to converge on a concept of some sort of highly heterogeneous crust that may in fact turn out to be hard to distinguish by remote (geophysical) observations. Any successful model must eventually take all the above-mentioned factors into account as well as appearing to be physically and Theologically feasible.

Some of the evidence we have is ambiguous or neutral. For example, the total tectonic subsidence (Sawyer, 1985) of the Iberia Abyssal Plain crust west of Site 901 and east of the peridotite ridge (Wilson et al., this volume) is consistent with the Sclater et al. (1980) prediction for 130-150 Ma oceanic crust. This is not a strong discriminant between the two hypotheses, however, because the predicted subsidence of very highly extended continental crust is about the same. Evidence of crustal extension on seismic reflection profiles likewise is not a strong discriminant; both hypotheses expect strong tectonic extension. Although it is generally believed that, on the basis of seismic velocity structure, normal oceanic crust can be distinguished from continental crust, which has been stretched by factors of say 3 or less, it becomes increasing clear that the situation may be far more difficult to decipher when comparison is made between crust generated by seafloor-spreading at less than 7.5 mm/yr half-rate and highly evolved disrupted and intruded continental crust. We also find that some geochemical evidence can also be ambiguous and difficult to decipher.

In spite of the apparent differences between the two hypotheses we would like to emphasize that Leg 149 drilling, and particularly the basement cores, have provided some important new and original information about the ocean/continent transition of the southern Iberia Abyssal Plain and presumably about ocean/continent transition zones of nonvolcanic margins in general. Sites 897 and 899 showed that serpentinized peridotite outcrops can exist not only at, or close to, the oceanward edge of the ocean/continent transition but also within the ocean/continent transition. Both peridotites experienced very similar melting and exhumation histories from the deep mantle, underwent important hydrothermal alteration during prolonged exposure on a seafloor elevation, and were uplifted shortly after the onset of seafloor spreading. Site 900 provided important evidence of a cumulate gabbro, of an almost certainly synrift age, that cooled at about 13 km depth and was exhumed tectonically. The gabbro was emplaced either in the upper mantle beneath oceanic crust formed by ultraslow seafloor spreading or was underplated beneath thinned continental crust.

The differences between the Galicia Bank and southern Iberia Abyssal Plain segments of the west Iberia Margin are also important and deserve further study. It is clear that the ocean/continent transition, however formed, is much narrower off Galicia Bank. Sampling there has revealed widespread granitic rocks and only rare igneous (basaltic or gabbroic) rocks whereas in the southern Iberia Abyssal Plain segment gabbroic and basaltic material, either in situ or as clasts, features at Sites 899 and 900, and granitic material is completely absent (although the latter is probably a function of a lack of granitic outcrops on the adjacent land).

In future, all aspects of the problem deserve further study. There is a need for more extensive and detailed geophysical observations, as well as for further sampling of acoustic basement that extends, and fills gaps in, the transect of drill sites begun during Leg 149. Only by this means will we obtain a clearer view of the temporal and spatial development of the continent-to-ocean transition. It is salutary to realize that, although significant progress has been made from the cores obtained so far, basement has been sampled at only three sites in the southern Iberia Abyssal Plain. Until samples of in situ lava flows are recovered from beneath the southern Iberia Abyssal Plain, the first hypothesis must remain tentative, and, until samples of unequivocal continental basement material are obtained there, the second hypothesis must also remain tentative. We also suggest that coordinated studies of analogous active nonvolcanic rifted margins, such as the northern Red Sea, might help to progress studies of the difficult problem we have presented here.

SUMMARY AND CONCLUSIONS

The west Iberia Margin is a nonvolcanic rifted margin which, following rifting that began in the Late Triassic, broke away from Newfoundland in the Early Cretaceous by a process of rifting which propagated from south-to-north.

The following summarizes the principal conclusions of this chapter.

1. Off west Iberia the ocean/continent transition is defined by an 80-130-km-wide region between the most oceanward tilted basement fault blocks (continental crust?) and a 300-km-long narrow margin-parallel peridotite ridge. Leg 149 drilled a west-to-east transect of holes across the ocean/continent transition, three of which reached acoustic basement, beginning at the peridotite ridge.
2. Sites 897 and 899 sampled serpentinized peridotites with similar, but not identical, petrologies that experienced a similar history of exhumation from the deep mantle to the surface. They crystallized at 1170°-1230°C, and this was followed by ductile shear deformation at 880°-1000°C. After limited partial melting, secondary minerals crystallized at around 30 km depth. Mylonitization, then low-temperature deformation and serpentinization, followed as the rocks were exhumed at the seabed. One important difference between the two sites is the fivefold greater strength of remanent magnetization of the Site 899 cores, which is reflected in the amplitudes of magnetic anomalies observed over the sites. The peridotites are neither clearly subcontinental nor suboceanic.
3. Site 900 sampled a flasered cumulate gabbro basement. The rare-earth element patterns are ambiguous and have been matched by different authors to both island arc and N-MORB basalts. The light rare-earth element pattern fits a transitional MORB parent magma. Nd and Pb isotope ratios also strongly suggest a MORB parent. The gabbro therefore likely formed in a magma chamber from a melt that was little contaminated, if at all, by continental crust. Traces of the primary mineralogy indicate dynamic crystallization at depths of at least 13 km and temperatures typical of granulite facies conditions. We think this happened either in the upper mantle beneath ultraslow spreading oceanic crust or by underplating under thinned continental crust. During exhumation to the seabed the rocks underwent retrograde metamorphism at 280 ± 20°C, followed by crystallization of some plagioclases at 136.4 ± 0.3 Ma.
4. Upper Barremian to upper Aptian debris-flow and mass-flow deposits were encountered above acoustic basement in holes at Sites 897 and 899, both of which are now situated on substantial basement elevations hundreds of meters above the adjacent basins. They are overlain by poorly fossiliferous latest Maastrichtian to Eocene thin claystones and conglomerates (lag deposits) of similar age to the sediments at the same level in the flanking basins. We explain these deposits by rapid Aptian uplift (or at least less subsidence than in the flanking basins), ~10 m.y. after the onset of 10 mm/yr seafloor spreading west of the peridotite ridge., followed in Eocene time by even-

tual blanketing by the sediments of the Iberia Abyssal Plain. This also explains the long history of seawater alteration of serpentinized peridotite at both sites.

5. The above results are explained here by two possible hypotheses for the development of the west Iberia ocean/continent transition, neither of which fully explains all the observations. The first hypothesis is that the ocean/continent transition is underlain by crust formed by ultraslow (~5 mm/yr) seafloor spreading which, by analogy with the slow-spreading Mid-Atlantic Ridge, allows the seabed exposure of peridotite and gabbro by extensive faulting. This hypothesis explains the drilling results but has difficulty in explaining the magnetic anomalies or the minute volume of basalt in the cores. The second hypothesis envisages an ocean/continent transition of tectonically and magmatically disrupted continental crust. It explains the Site 900 MORB gabbro as material underplated under thinned continental crust or as an aborted point-of-initiation of seafloor spreading and the alkaline-to-transitional characters of igneous clasts in Site 897 and 899 deposits. Its main problem is the lack of unequivocal continental basement samples, although there is evidence of reworked continental sediments in the cores. Both hypotheses tend to expect a highly heterogeneous crust in the ocean/continent transition that may be hard to characterize other than by further basement sampling.

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