Whitmarsh, R.B., Sawyer, D.S., Klaus, A., and Masson, D.G. (Eds.), 1996
Proceedings of the Ocean Drilling Program, Scientific Results, Vol. 149

1. THE WESTERN IBERIA MARGIN: A GEOPHYSICAL AND GEOLOGICAL OVERVIEW[1]

L.M. Pinheiro,[2] R.C.L. Wilson,[3] R. Pena dos Reis,[4] R.B. Whitmarsh,[5] A. Ribeiro[6]

ABSTRACT

This paper presents a general overview of the geology and geophysics of western Iberia, and in particular of the western Portuguese Margin. The links between the onshore and offshore geology and geophysics are especially emphasized. The west Iberia Margin is an example of a nonvolcanic rifted margin. The Variscan basement exposed on land in Iberia exhibits strike-slip faults and other structural trends, which had an important effect on the development, in time and space, of subsequent rifting of the continental margin and even perhaps influences the present-day offshore seismicity. The margin has had a long tectonic and magmatic history from the Late Triassic until the present day. Rifting first began in the Late Triassic; after about 70 Ma, continental separation began in the Tagus Abyssal Plain. Continental breakup then appears to have progressively migrated northwards, eventually reaching the Galicia Bank segment of the margin about 112 Ma. Although there is onshore evidence of magmatism throughout the period from the Late Triassic until 130 Ma and even later, this was sporadic and of insignificant volume. Important onshore rift basins were formed during this period. Offshore, the record is complex and fragmentary. An ocean/continent transition, over 150 km wide, lies beyond the shelf edge and is marked on its western side by a peridotite ridge and thin oceanic crust characterized by seafloor spreading anomalies. Rifted fault blocks are recognized within the ocean/continent transition along the whole margin. Mostly, they merge westwards into a transitional zone where the basement often has low relief of unknown origin, and linear magnetic anomalies parallel the seafloor spreading anomalies. There is indirect geophysical evidence that this zone is underlain by intrusions in the lower crust. The most plausible, but not the only, explanation seems to be that this part of the ocean/continent transition consists of fragments of magmatically disrupted and intruded thinned continental crust. The margin also underwent important postrift compression in Eocene and Miocene time, as demonstrated by folding and nondeposition or erosion of abyssal plain sediments. The Eocene deformation is clearly visible off Galicia but the Miocene deformation is dominant in the rest of the margin, where it may have overprinted the former compressional episode. The margin is still seismically active in its southern part at the present day, mainly because of its proximity to the Azores-Gibraltar plate boundary.

INTRODUCTION

The purpose of this paper is to provide a general overview of the geology and geophysics of western Iberia, and in particular of the western Portuguese Margin, in order to help the reader to integrate the other papers in this volume into a more general geological, tectonic, and magmatic framework. Thus, we review the geoscientific literature that describes both the onshore and offshore history and development of the margin. No attempt is made here to address the issue of the conjugate margin off the Grand Banks of Canada, the history of which, until continental breakup occurred, must have been very similar to what is described here. The nature of the Grand Banks shelf and margin has been extensively described elsewhere (e.g., Keen and de Voogd, 1988; Tankard and Welsink, 1989; Welsink et al., 1989; Tucholke et al., 1989; Srivastava et al., 1990a; Srivastava and Verhoef, 1992), particularly in view of the very extensive exploration for hydrocarbons on the Grand Banks shelf.

We lay particular emphasis on bringing together the onshore and offshore geology and geophysics, an aspect of this margin that in the past has not received as much attention as it deserves. In spite of the different nature and density of the information available from observations on land and at sea, a coherent and consistent picture emerges of the history of this margin, which is perhaps unique among nonvolcanic margins in the degree to which both the onshore and offshore geology and geophysics are now understood.

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

[2]Departamento de Geociências, Universidade de Aveiro, 3800 Aveiro, Portugal. lmp@zeus.ci.ua.pt

[3]Department of Earth Sciences, The Open University, Milton Keynes, MK7 6AA, United Kingdom.

[4]Departamento de Ciências da Terra, Universidade de Coimbra, 3049 Coimbra codex, Portugal.

[5]Challenger Division for Seafloor Processes, Southampton Oceanography Centre, Empress Dock, European Way, Southampton SO14 3ZH, United Kingdom.

[6]Departamento de Geologia, Faculdade de Ciências da Universidade de Lisboa, Rua da Escola Politécnica, 58, 1294 Lisboa codex, Portugal.

PLATE TECTONIC SETTING AND SEISMICITY

The Iberian Peninsula is located on the Eurasian Plate, just north of the present-day Africa/Eurasia plate boundary (Fig. 1G). This boundary is located along the Azores-Gibraltar Fracture Zone (AGFZ), just south of Iberia, and continues eastwards into the Mediterranean (Srivastava et al., 1990a, b).

Analysis of seafloor-spreading magnetic anomalies, aided by paleomagnetic studies (Van der Voo and Zijderveld, 1971; Galdeano et al., 1989), has shown that the plate tectonic setting of the Iberian Peninsula (Fig. 1) changed significantly throughout its postrifting history (Olivet et al., 1984; Klitgord and Schouten, 1986; Malod and Mauffret, 1990; Srivastava et al., 1990a,b; Roest and Srivastava, 1991; Sibuet and Collette, 1991). It acted either as an independent plate or as an accreted terrane of the Eurasian or the African Plate (Klitgord and Schouten, 1986; Srivastava et al., 1990a,b; Roest and Srivastava, 1991). For the major part of the Cretaceous Magnetic Quiet Period, when the Bay of Biscay spreading ridge was active (Sibuet and Collette, 1991), Iberia appears to have moved as an independent plate (Fig. IB). From the Late Cretaceous (sometime before 84 Ma, Chron 34) to the middle/late Eocene (Chron 18), it was probably attached to the African plate, with the Eurasian/African plate boundary extending westward from the Bay of Biscay (Srivastava et al., 1990a,b; Figs. 1C and 1D). Sometime between Chrons 21 and 13 (possibly at Chron 19, middle Eocene) the plate boundary jumped to the south, extending from King's Trough to the Pyrenees, along the Azores-Biscay Rise and the North Spanish Trough (Fig. 1E). Since the early

Oligocene (Chron 13), Iberia has been part of the Eurasian plate, with the Eurasian/African plate boundary located in its present position along the AGFZ (Figs. 1F and 1G; Klitgord and Shouten, 1986; Srivastava et al., 1990a,b; Roest and Srivastava, 1991).

The western segment of the plate boundary, between the Azores Triple Junction and the Madeira-Tore Rise, is clearly marked by seismic activity along the AGFZ (Fig. 2A; Udias et al., 1976; Udias, 1982; Grimison and Chen, 1986, 1988; Madeira and Ribeiro, 1990). Focal-mechanism solutions indicate a mainly northeast-southwest to north-northeast-south-southwest extension near the Azores, changing to dextral strike-slip motion along the AGFZ (Fig. 2B; Udias et al., 1976; Udias, 1982; Grimison and Chen, 1986; 1988). Further to the east, between the Madeira-Tore Rise, Gibraltar, and the western Mediterranean, the seismicity becomes more diffuse and the definition of the plate boundary is more problematical (Fig. 2A; Grimison and Chen, 1988; Bergeron and Bonnin, 1991). East of 12°W, and in particular near Gorringe Bank, the focal-mechanism solutions indicate a combination of strike-slip and northwest-southeast thrust faulting related to the Africa/Eurasia convergence that began in the Late Cretaceous (late Campanian?) (Fig. 2B; McKenzie, 1972; Fukao, 1973; Grimison and Chen, 1988).

Today, the main seismic activity in western Iberia is located at or close to the Eurasia/Africa boundary, particularly in the Gorringe Bank region (Fig. 2). There is also moderate intraplate seismicity associated with the Tore Seamount and along the offshore extensions of late Variscan basement faults (Fig. 2C). Onshore, the most active faults are the Vilariça Fault, the Seia-Lousã Fault and its probable extension along the Nazaré Fault and the Lower Valley of the Tagus Fault (Fig. 2C). Other faults with associated seismicity include the Messejana Fault (also known as Odemira-Avila Fault) and the Loulé Fault (Fig. 2C; Cabral, 1983; Moreira, 1984; 1985; Oliveira, 1986).

PHYSIOGRAPHIC FEATURES OF THE CONTINENTAL MARGIN

Excellent reviews of the main physiographic and geomorphological features of the Portuguese part of the west Iberia Margin can be found in Vanney and Mougenot (1981), Brum Ferreira (1981), Sibuet et al. (1987) and Mougenot (1988). These features are clearly visible on the bathymetric charts of the northeast Atlantic published by Laughton et al. (1975) and Lallemand et al. (1985). The summary presented here is largely based on the references presented above, to which the reader is referred for further details.

The width of the western Iberian continental shelf varies between 10 and 65 km along its approximate 700 km length (Fig. 3). It can be divided into three distinct segments, separated by the major canyons that cut across the shelf and descend into the abyssal domains (Vanney and Mougenot, 1981; Mougenot, 1988): (1) north of the Nazaré Canyon, (2) between the Nazaré and the Setúbal canyons, and (3) between the Setúbal and the Saint Vincent canyons (Fig. 3). The segment north of the Nazaré Canyon includes the Galicia Margin of northwest Spain and the Portuguese geographical provinces of Minho and Beira Litoral. Between Porto and Cape Finisterra the continental shelf is only about 30 km wide and is oriented roughly north-south. To the south of Porto, the continental shelf widens a little to about 40 km, and south of Aveiro its width increases still further to 50-55 km. In this region, the continental slope is significantly steeper than it is to the north, and its orientation changes to approximately northeast-southwest along the postulated Aveiro fault zone. Beyond the continental slope adjacent to Aveiro, a marginal plateau extends westward, widening to nearly 200 km in the north (Fig. 3). This comprises Galicia Bank and three major seamounts: Porto, Vigo, and Vasco da Gama. These shallow regions are separated from the continental slope by the Galicia Interior Basin. The D. Carlos Valley, in the southeast limit of Galicia Bank, is probably one of the main conduits of sediments from the continental shelf into the Iberia Abyssal Plain (Gardner and Kidd, 1987).The segment between the Nazaré and the Setúbal canyons is characterized by a relatively wide continental shelf (Fig. 3; Mougenot, 1988). In this area, a ridge projects out from the continental slope, forming one of the most important physiographic features of the margin: the Estremadura Spur. This spur separates the Iberia Abyssal Plain in the north from the Tagus Abyssal Plain in the south, and extends nearly as far west as the Tore Seamount, at the northern end of the Madeira-Tore Rise. The origin of the spur is not clear, but it coincides partly with an alignment of small volcanic features, at least as far west as the 2000-m contour (Mougenot, 1988). Besides the Nazaré and the Setúbal canyons, the only other major canyons that dissect this segment of the margin are the Lisboa and Cascais canyons, both situated southwest of Lisboa (Van-

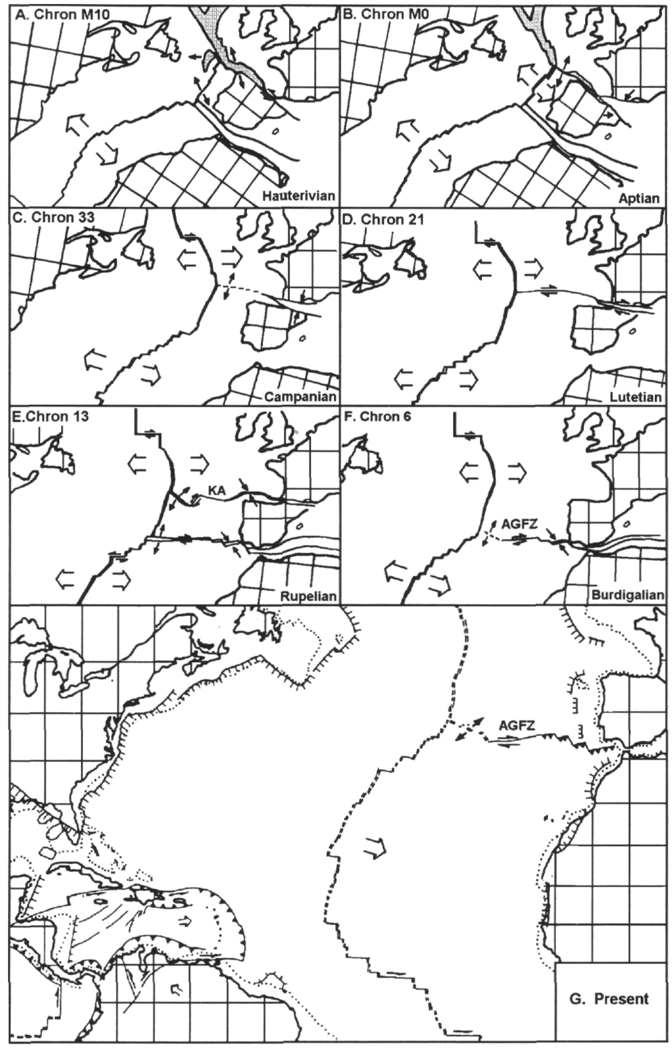

Figure 1. The changing plate tectonic setting of Iberia. **A-F:** Plate tectonic setting between Chron M10 (Hauterivian) and Chron 6 (Burdigalian), after Srivastava et al., 1990a (with permission). AGFZ = Azores-Gibraltar Fracture Zone; KA = Kings Trough-Azores Biscay Rise-North Spanish Trough-plate boundary. Shaded areas represent regions of overlaps between plates, implying later extension in those regions. The regions with gaps between plates imply later compression. **G.** Present-day plate tectonic setting, after Klitgord and Shouten (1986). Dotted lines denote present-day 200-m isobath (edge of the continental shelf)- Lines with small perpendicular dashes denote the approximate continental edge. Large arrow indicates plate motions relative to North America (after Klitgord and Shouten [1986], with slight modifications off Iberia).

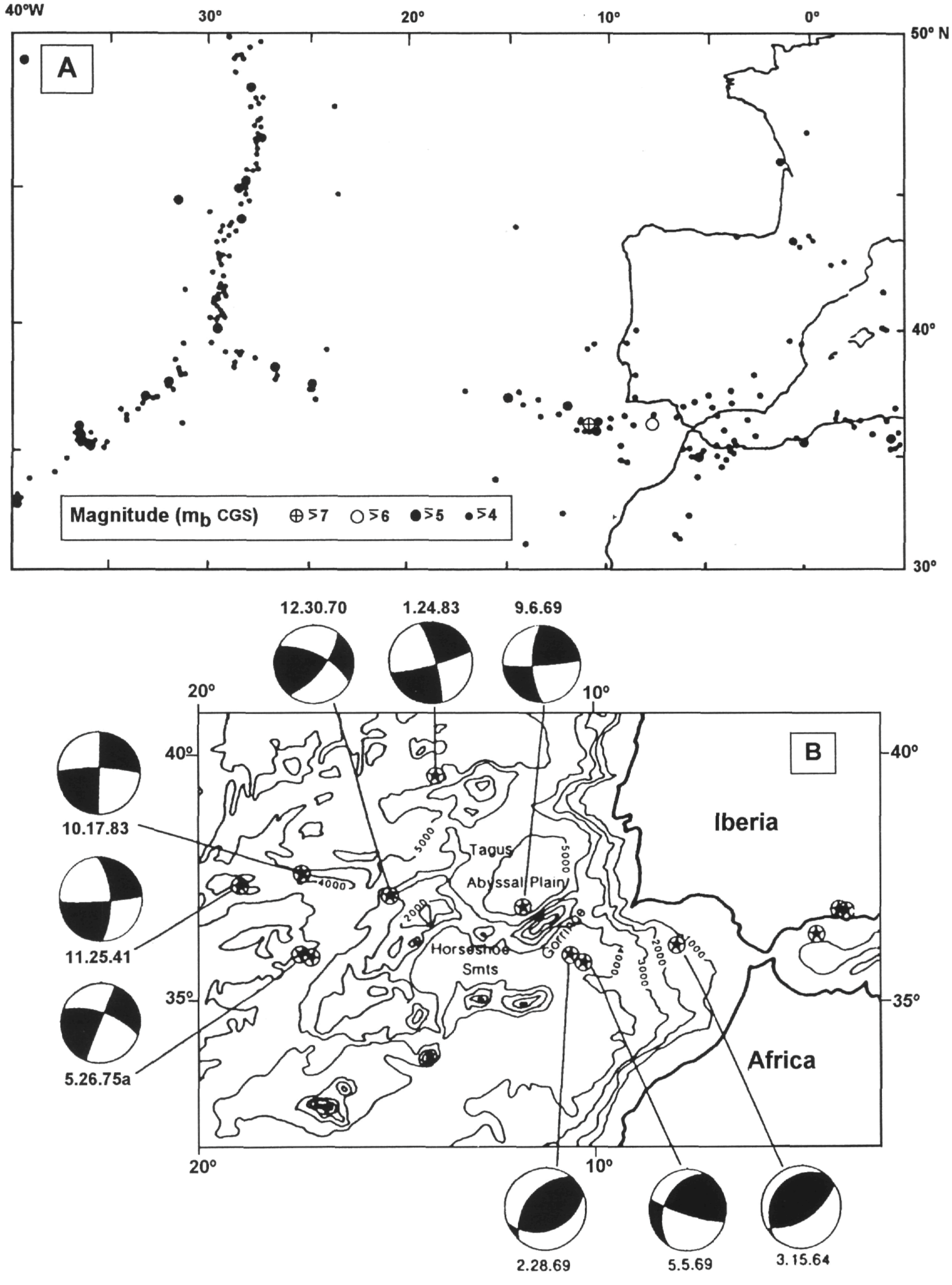

Figure 2. **A.** Seismicity map for the area between the Azores triple junction and the western Mediterranean for the period 1962-1972, showing location of epicenters of events with body-wave magnitude ≥4. From Udias et al. (1976). **B.** Earthquake focal mechanisms in the area of the eastern end of the Azores-Gibraltar plate boundary (after Grimison and Chen, 1988). Bathymetry in meters. A mixture of strike-slip and thrust faulting characterizes the area. **C.** Simplified neotectonic map of Portugal (after Cabral, 1989) showing the main active structures and the magnitudes and epicentral location of the main historical and present-day seismic activity (after Moreira, 1985). Open symbols denote historical events; solid symbols denote present-day instrumental earthquakes. 1 = active fault; 2 = suspected active fault trace; 3 = normal fault or indeterminate vertical movement component (barbs on downthrown side); 4 = reverse fault; 5 = strike-slip fault. GO = Gorringe Bank; LF = Loulé Fault; LVT = Lower Valley of the Tagus Fault; MF = Messejana Fault; NF = Nazaré Fault; SL = Seia-Lousã Fault; VF = Vilariça Fault. Simplified bathymetry in meters, after Lallemand et al., 1985).

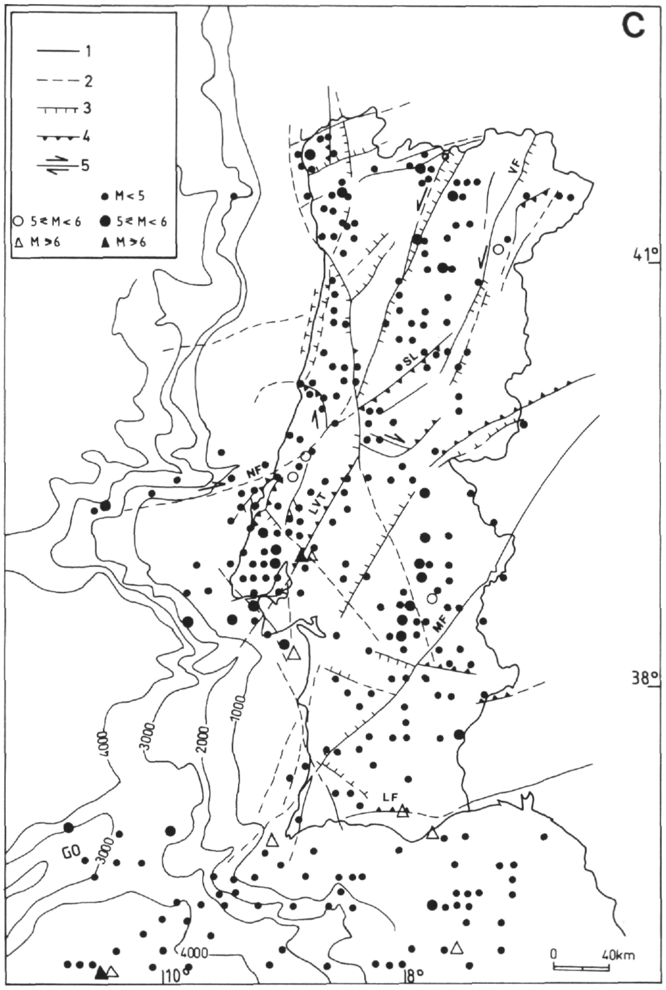

Figure 2 (continued).

ney and Mougenot, 1981). Small islands (Farilhões and Berlengas) occur close to the coast (10-15 km) to the northwest of Peniche (see Fig. 4A for location). The segment between the Setúbal and the Saint Vincent canyons is characterized, in general, by a fairly narrow (10-20 km) continental shelf, although between 37°25'N and 38°N it is extended to the west by a platform (Vanney and Mougenot, 1981). The coastline in this segment consists of prominent scarps cut in Paleozoic schists.

Two large abyssal plains occur off western Iberia: the Tagus Abyssal Plain in the south, and the Iberia Abyssal Plain in the north. The Tagus Abyssal Plain is an enclosed, extremely flat-floored basin, bounded to the east by the irregular continental margin of Portugal, onto which sediments have been channelled through the Lisboa and Setúbal canyons (Fig. 3). Piston cores show that the upper few meters of the sedimentary cover consist of thick turbidites, separated by pelagic layers (P. Weaver, pers. comm., 1989). The abyssal plain is surrounded by three major ridges: the Madeira-Tore Rise in the west, the Estremadura Spur in the north, and Gorringe Bank in the south. The Madeira-Tore Rise shoals to less than 1000 m in places, and it was formed very near the Mid-Atlantic Ridge by excess volcanism, contemporaneously with the adjacent oceanic lithosphere (Peirce and Barton, 1991). Gorringe Bank separates the Tagus Abyssal Plain from the Horseshoe Abyssal Plain. This ridge has been interpreted as an uplifted and tilted block of oceanic crust, resulting from the Africa-Europe collision (Purdy, 1975). At DSDP Site 120, on the northern flank of Gorringe Bank (see location in Fig. 3), the oldest sediments drilled were Lower Cretaceous (Barremian), resting on top of

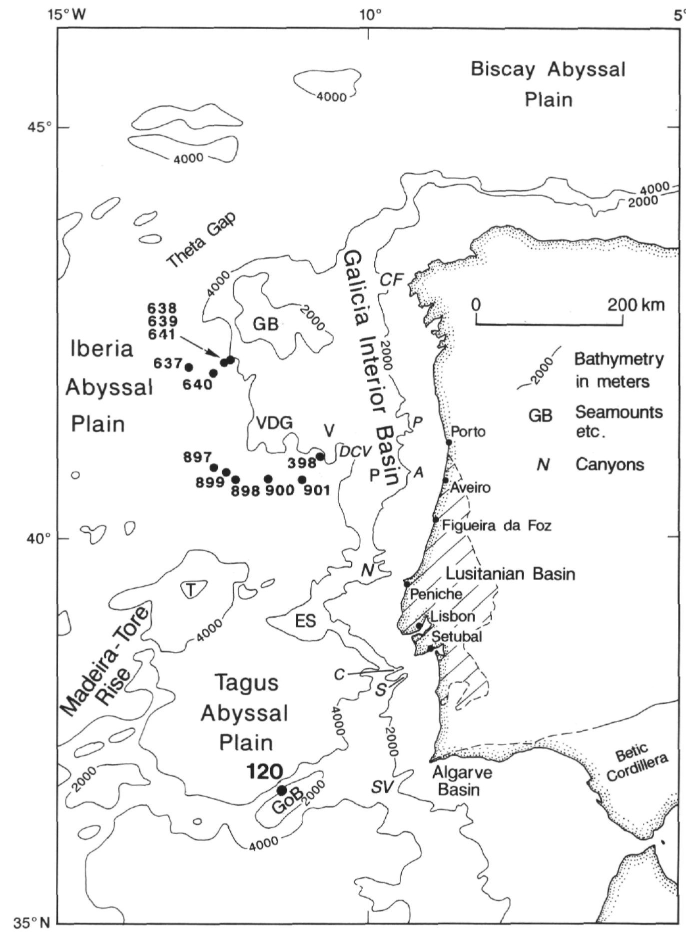

Figure 3. Physiographic features of the west Iberia Margin. Bathymetry is shown in meters, simplified after Lallemand et al. (1985). Submarine canyons (in italic): P = Porto; A = Aveiro; DCV = Dom Carlos Valley; C = Cascais (note that the Lisboa canyon branches shoreward to the northwest from this canyon and merges to the south with the Setúbal canyon; the contours do not show its location); S = Setúbal; N = Nazaré; CF = Cape Finisterra; SV = Cape Saint Vincent. Seamounts and other features: GB = Galicia Bank; P = Porto Seamount; T = Tore Seamount; V = Vigo Seamount; VDG = Vasco de Gama Seamount; ES = Estremadura Spur; GoB = Gorringe Bank.

an oceanic basement consisting of spilitic basalts, serpentinites and meta-gabbros (Ryan, Hsü et al., 1973). The Gettysburg Seamount (western peak of Gorringe Bank) appears to be essentially formed of serpentinite, whereas the basement of the Ormonde Seamount (eastern peak of Gorringe Bank) appears to consist of an oceanic section with gabbros of Berriasian age (140-143 Ma), locally covered by alkaline volcanic rocks of Paleocene age (Auzende et al., 1978; Cornen, 1982; Féraud et al., 1986).

The Iberia Abyssal Plain is located off central and northwestern Iberia. To the south, the Iberia Abyssal Plain is separated from the Tagus Abyssal Plain by the Estremadura Spur and the Tore Seamount. To the west and northwest, the limits of the abyssal plain are roughly defined by the 4800-m isobath, which approximately marks the transition from the flat topography characteristic of the abyssal plain to the irregular morphology typical of oceanic basement further west. The sedimentary cover beneath the plain consists mainly of thick turbidites separated by pelagic layers. Major turbidite flows were probably channelled to the abyssal plain down the Porto, Aveiro, and Nazaré Canyons, but Galicia Bank and other elevated areas to the east and northeast of the abyssal plain may also have contributed sediment. The margin shows no signs of significant subma-

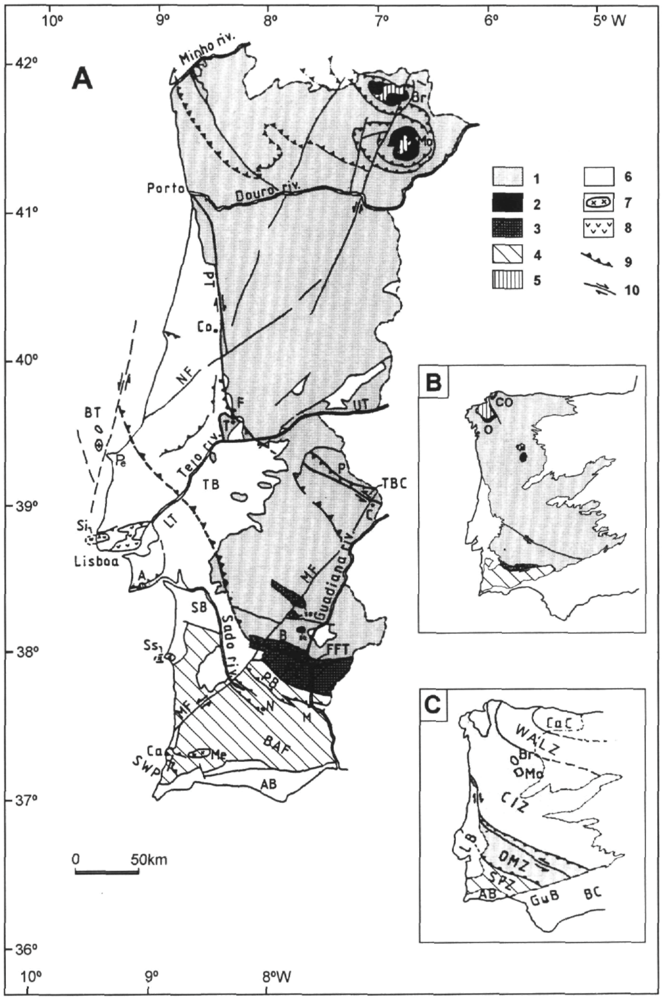

Figure 4. Zones and terranes of the Variscan orogen in the Hesperic Massif of Iberia. **A.** Detailed map of the main terranes in the Portuguese part of the Hesperic Massif adjacent to the study area (modified from Ribeiro and Silva, in press). 1 = Iberian Terrane; 2 = Northern Ophiolite; 3 = South Oceanic Terrane, which includes the southern ophiolite (so = Beja-Acebuches ophiolite and ophiolite klippen) and the Pulo do Lobo Unit (PL); 4 = South Portuguese Terrane; 5 = Northern Continental Terrane; 6 = Mesozoic-Cenozoic Borderland: includes the Lusitanian Basin (SB = Sado Basin; TB = Tagus Basin; LT = Lower Tagus; UT = Upper Tagus) and the Algarve Basin (AB); 7 = Upper Cretaceous intrusives (Si = Sintra; Ss = Sines; Me = Monchique); 8 = Lisboa Basaltic complex; 9 = reverse fault (dashed = inferred); 10 = strike-slip fault; A = Arrábida; B = Beja; BAF = Southern Alentejo Flysch Group; Br = Bragança Massif; BT = Berlengas Suspect Terrane; C = Campo Maior; Ca = Carrapateira; Co = Coimbra; F = Ferreira do Zêzere; FFT = Ferreira do Alentejo-Ficalho Thrust; M = Mértola; MF = Messejana Fault; Mo = Morais Massif; N = Neves-Corvo; NF = Nazaré Fault; P = Portalegre; PB = Pyrite belt; Pe = Peniche; PT = Porto-Tomar Fault; SWP = Southwest Portuguese Domain; T = Tomar; TBC = Tomar-Badajoz-Córdoba shear zone. The Ferreira do Zêzere-Portalegre thrust runs between F and P. The Ferreira do Alentejo-Ficalho thrust separates the northern end of the Pulo do Lobo Unit from the Beja-Acebuches ophiolite. A is a detailed version of the portion of B adjacent to the study area. **B.** Terranes in the Hesperic Massif (modified from Quesada et al., 1994). Symbols as in A. CO = Cabo Ortegal complex; O = Ordenes complex. **C.** Classical subdivision of the Hesperic Massif. CIZ = Central Iberian Zone; OMZ = Ossa Morena Zone; SPZ = South Portuguese Zone; WALZ = West Asturian Leonese Zone; CaC = Cantabrian Zone; LB = Lusitanian Basin; AB = Algarve Basin; GuB = Guadalquivir Basin; BC = Betic Cordillera; Br = Bragança Massif; Mo = Morais Massif. (Modified from Ribeiro and Silva, in press).

rine fans or sediment slumping, unlike the conjugate margin (Gardner and Kidd, 1987). This has been interpreted by Gardner and Kidd (1987) as indicating that bottom currents must play an important role in redistributing the sediments transported downslope from the adjacent continental shelf and offshore highs.

GEOLOGICAL FRAMEWORK OF WESTERN IBERIA

Variscan Basement

The western margin of Iberia consists of a fragment of the Variscan orogen (often referred to as the Hesperic massif), above which occur several Mesozoic-Cenozoic basins that were deformed to varying degrees during two Cenozoic compressional episodes (Capdevila and Mougenot, 1988). The evolution of these basins was related to the reactivation of structures within the Variscan basement, in particular of those associated with the late Variscan deformation phases.

The Variscan structures in Iberia, together with those of the Armorican Massif, define the so-called Ibero-Armorican Arc, which was subsequently disrupted by the opening of the Bay of Biscay. In the Iberian Peninsula, the Variscan orogeny started in the middle Devonian and ended in the Carboniferous (Ribeiro et al., 1979). In very broad terms, the Variscan Massif consists mainly of folded, thrusted and metamorphosed rocks of Precambrian and Paleozoic age, which were extensively intruded by large granitoid batholiths during, and after, the Variscan continent-continent collision. The massif has been divided into five main zones, each characterized by distinct paleogeography, structural style, metamorphism, and magmatism (Julivert et al., 1980; Ribeiro et al., 1979) (Fig. 4C). The Central Iberian Zone, the Ossa Morena Zone, and the South Portuguese Zone underlie the west Iberia Margin. The thrust separating the Central Iberian Zone from the Ossa Morena Zone (the Ferreira do Zêzere-Portalegre thrust) is inclined to the southwest, whereas the thrust that marks the northern boundary of the South Portuguese Zone (The Ferreira do Alentejo-Ficalho thrust) dips north or northeast (Ribeiro et al., 1979; Ribeiro and Silva, in press). In the Central Iberian Zone, both synorogenic (350-300 Ma) and post erogenic granitoids (280 Ma) are observed (Ribeiro and Silva, in press). In the Ossa Morena zone, however, the synorogenic magmatism consists of a suite of gabbros, diorites, and granodiorites, followed by intermediate calc-alkaline volcanics of lower Carboniferous age (Ribeiro et al., 1979; Ribeiro and Silva, in press). The South Portuguese Zone consists of low-grade upper Paleozoic metasediments, with an ophiolite marking its northern boundary. First-phase fold axes in all the zones trend predominantly north-northwest to south-southeast, but second-phase structures show a northwest to southeast trend in the Ossa Morena and South Portuguese Zones.

Recently, the Hesperic Massif has been interpreted as consisting of several major tectonostratigraphic terranes (Figs. 4A, B) that were accreted together at different times during the late Paleozoic Variscan orogeny, which resulted from the closure of the Paleotethys Ocean (Ribeiro et al., 1987; 1990; Quesada, 1990; 1991; Ribeiro and Silva, in press). In Portugal, these include: (1) the Iberian terrane, which occupies most of the Massif, shows Gondwanian affinities at least until the late Proterozoic and is considered the Iberian autochthon (it includes the Central Iberian and Ossa-Morena zones, cf. Fig. 4C); (2) the northern ophiolite, which is an exotic terrane or terranes overlying the Iberian Terrane, representing Paleozoic ocean floor and overlying sedimentary sequences that form an ophiolitic thrust sheet obducted eastward onto the Iberian terrane during the Variscan convergence (possible remnants of a Paleozoic ocean lying west and south of the Iberian Terrane) (Ribeiro et al., 1990; Quesada, 1990); (3) the northern continental terrane, another exotic terrane overlying the Iberian Terrane and consisting of several allochthonous units (Precambrian basement rocks overlain by Paleozoic clastic cover sequences) of unknown palinspastic origin, composing thrust sheets that sit on top of the ophiolitic nappes in NW Iberia (Bragança and Morais Massifs in Portugal [Fig. 4A] and Cabo Ortegal and Ordenes in Spain [Fig. 4B]); (4) The South Portuguese terrane, which is a continental block that includes the pyrite belt (quartzites, a volcanic sedimentary complex, and turbidites), the Baixo Alentejo Flysch Group, and the southwest Portugal domain (sedimentary units that represent time equivalents of the Pyrite belt, in condensed fácies); and (5) the South Oceanic Terrane, which represents a major suture within the Variscan fold belt and separates the South Portuguese Terrane from the Iberian Terrane. The South Oceanic Terrane includes the southern ophiolite (Beja-Acebuches ophiolite and ophiolite klippen overlying the Iberian Terrane) and the Pulo do Lobo unit. The latter is essentially a metasedimentary unit with minor representation of ophiolite melanges and volcanic rocks with a typical N-MORB chemistry (from the base upwards: serpentinites, flaser gabbros, massive and banded gabbros, sheeted dikes, and pillow basalts) (Quesada et al., 1994; Ribeiro and Silva, in press). The area of basement exposed in and on the seafloor around Berlengas and Farilhões islands, off Peniche (Fig. 4A), has been recently interpreted as a suspect terrane, corresponding to a sliver of the Iberian Terrane transported southwards by a late Variscan movement along a north-south oriented sinistral strike-slip fault, and subsequently thrust eastwards during the Permian (Ribeiro and Silva, in press).

The late Variscan deformation in Iberia was marked by the development of brittle fault systems (Ribeiro et al., 1990). During this phase, the stress field was complex. The Porto-Tomar Fault was reactivated as a dip-slip, high-angle reverse structure, suggesting an essentially east-west maximum compressive direction. At approximately the same time, however, conjugate strike-slip faults were also generated: a sinistral north-northeast-south-southwest to east-northeast-west-southwest set (which is predominant) and its dextral north-northwest-south-southeast to northwest-southeast conjugate set (Ribeiro et al., 1979; Dias, 1986); these indicate a north-south maximum compressive stress (Ribeiro et al., 1990). This complex regime has been interpreted as a corner effect within the second-order Iberian block, which was compressed within a diffuse, large-scale right-lateral shear zone extending between the Urals and the southern Appalachians (Arthaud and Matte, 1975; Ribeiro et al., 1990).

A detailed study of plutonic and metamorphic rock samples from dredge hauls and cores from the west Iberia Margin has led the recognition of the offshore extension of the main zones of the Hesperic Massif observed under Iberia (Capdevila and Mougenot, 1988). The Iberian terrane (which includes the Central Iberian Zone and the Ossa Morena Zone) seems to extend from northernmost Iberia to south of the Nazaré canyon. Low-pressure amphibolite-facies metamorphic rocks of sedimentary origin, a variety of granites, granodiorites, tonalites, and low-to-intermediate pressure granulites have been recovered from Galicia Bank and from the northern Portuguese continental shelf. Of these, only the granulites (1500 m.y. old; Postaire, 1983) apparently have no onshore equivalents in the Iberian terrane and probably form part of the pre-Variscan complex (the granulites from the NW Iberian nappes are of high-pressure type and probably Paleozoic). The tonalites and granodiorites commonly found in the Ossa Morena Zone are well represented on the northwestern Iberian continental slope and in the Galicia Bank area but not on the shelf. Similar basement rocks have also been recovered from the continental shelf to the west of the Nazaré canyon and as far south as the Sines area. In contrast, the abundant peraluminous leucogranites found in Galicia (Central Iberian Zone) are rarely found on the slope but are common on the Galicia and northern Portuguese continental shelf. Based on this evidence, Capdevila and Mougenot (1988) interpreted that rocks from the Central Iberian Zone form the basement of the continental shelf of Galicia and northern Portugal, whereas rocks from the Ossa Morena zone form the basement in the area between the continental slope off Galicia and the continental shelf west of the Nazaré canyon. Between the Cape of Sines and Cape Saint Vincent, coring yielded flysch sequences, volcano-sedimentary rock sequences, and quartzites typical of the South Portuguese Zone (Capdevila and Mougenot, 1988), suggesting that the offshore basement to the south of Lisbon consists essentially of rocks from the South Portuguese terrane.

Mesozoic-Tertiary Basins and Structural Highs

Many of the tectonic trends within the Mesozoic-Tertiary basins of western Iberia follow the prominent Variscan features (Fig. 5; cf. Fig. 2C). The late Variscan strike-slip faults, in particular the predominant north-northeast-south-southwest to east-northeast-west-southwest set (see Fig. 2C), have played an important role since the Trias sic. They behaved as normal faults during the pre-Alpine extension and as reversed or thrust faults, sometimes with a strike-slip component, during the Alpine compression (Boillot et al., 1979; Alvarado, 1983; Masson et al., 1994). These late Variscan faults appear therefore to constitute the main structural control on the Mesozoic rifting geometry. The major canyons that cut across the continental shelf also exhibit a mainly northeast-southwest trend, roughly parallel to the dominant set of late Variscan strike-slip faults in western Iberia, and appear to lie on the offshore continuation of major late Variscan faults observed on land (e.g., the Nazaré Canyon, on the continuation of the Nazaré Fault, and the Saint Vincent Canyon on the continuation of the Odemira-Ávila Fault, also known as the Messejana Fault; Boillot et al., 1974; Lallemand and Sibuet, 1986).

The main basins and structural highs that can be observed along the west Iberia Margin are (1) the deep Galicia Margin, (2) the Western Banks, (3) the Galicia Interior Basin, (4) the Porto Basin, (5) the Lusitanian Basin, and (6) the Alentejo Basin. The geometry of these features shows a strong relationship to the orientation of Variscan basement structures described above, particularly the late orogenic strike-slip faults. Their location is shown in Figure 5, cross sections across them are presented in Figure 6, and a brief summary of the main features of each of them is given below.

1. Deep Galicia Margin. The fault spacing across the series of half-grabens in this region ranges between 10 and 25 km and the basin-bounding down-to-the-ocean normal faults trend predominantly north-south, with transfer faults oriented northeast-southwest (Figs. 5, 6A; Montadert et al., 1979; Thommeret et al., 1988; Boillot et al., 1989b).
2. Western Banks. This area is a zone trending north-northwest—south-southeast along which occur a series of elevated areas, including Galicia Bank, and Vasco da Gama, Vigo, and Porto Seamounts (Figs. 5, 6B; Mougenot et al., 1984; Murillas et al., 1990). These features have been interpreted as horsts that formed initially during Mesozoic rifting and were later reactivated and uplifted during Cenozoic compressive events (Boillot et al., 1979; Mougenot, 1988). Murillas et al. (1990) suggested that the alignment of elevated areas extends to the south under the Portuguese continental shelf as the Berlenga horst that forms the western boundary of the Lusitanian Basin. However, this structure does not form a southward continuation of the north-northwest-south-southeast trend, and in any case is separated from it by two major faults (Aveiro and Nazaré).
3. Galicia Interior Basin. This basin extends over 350 km north-northwest-south-southeast beneath the Galicia Trough (Figs. 5, 6B-D). It is bounded by down-to-the-basin normal faults with trends varying between north-northeast-south-southwest and northwest-southeast. The associated transfer faults show northeast-southwest to east-northeast-west-southwest orientations (Figs. 5, 6B, D; Murillas et al., 1990).
4. Porto Basin. This is a relatively narrow basin (~50 km wide) extending beneath the outer continental shelf and slope oppo-

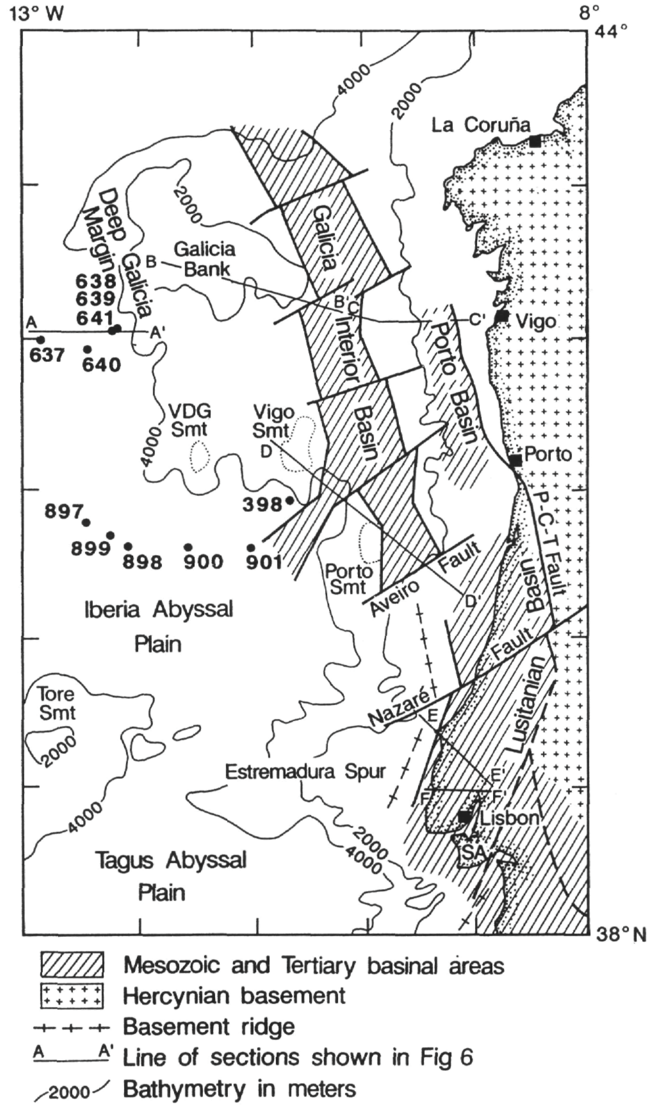

Figure 5. Sketch map showing the distribution of marginal sedimentary basins along the west Iberia Margin (from Murillas et al., 1990). S A = Serra da Arrábida; P-C-T = Porto-Coimbra-Tomar Fault; VDG = Vasco da Gama Seamount. The lines labeled A-F show the location of structural cross sections shown in Figure 6. The Berlenga horst is located at the northwest end of the line E-E'.

site Porto (Figs. 5, 6C). It may be the northward continuation of the Lusitanian Basin (or the northernmost sub-basin within the Lusitanian Basin). Its eastern margin is formed by the offshore continuation of the Porto-Coimbra-Tomar Fault (Murillas et al., 1990). A horst block on its western side separates it from the Galicia Interior Basin.

5. Lusitanian Basin. This basin extends northwards for over 300 km from the Serra da Arrábida, south of Lisbon (Figs. 5, 6D). The greater part of it is situated onshore, with its eastern boundary coinciding with the Porto-Coimbra-Tomar Fault. South of the Nazaré Fault, its western margin is formed by the Berlenga horst, but, north of this fault, the tectonic character of the western edge is not known. Constituent sub-basins are bordered today either by salt structures or by down-to-the-ocean normal faults (Montenat et al., 1988; Wilson et al., 1989). The former show a dominant north-northeast-south-southwest trend, and the latter show a north-south trend.
6. Alentejo Basin. This is a north-south-trending basin situated offshore to the south of the Serra da Arrábida (Mougenot, 1988); its relationships to the Lusitanian Basin are not known.

Crustal Structure Under Western Iberia

In general terms, the continental crust under western Iberia is approximately 30-35 km thick. Recent studies of new data collected during the ILIHA deep seismic sounding experiment (ILIHA DSS Group, 1993), together with a re-interpretation of the preexisting seismic refraction lines (Mendes Victor et al., in press) have resulted in a well constrained model of the crustal structure beneath Iberia. This reveals a fairly heterogeneous upper crust and a reasonably homogeneous lower crust throughout Iberia. The middle crust in the internal zones of the Variscan orogen is thicker than in the external zones. In broad terms, the seismic-velocity/depth structure beneath western Iberia can be summarized as follows: (1) In places, the sedimentary cover is quite thick (5-6 km in the Lusitanian Basin and in the Algarve Basin, respectively) and the observed seismic velocities reach 5.0 to 5.3 km/s; (2) the upper crust is heterogeneous but it is normally characterized by P-wave velocities of 5.9-6.1 km/s; (3) the middle crust generally exhibits P-wave velocities between 6.2 and 6.4 km/s; and (4) the lower crust generally exhibits velocities in the interval 6.6-7.1 km/s (Mendes Victor et al., in press). The P-wave velocities in the top of the upper mantle are generally 8.0-8.1 km/s, but unusually high velocities of 8.3 km/s are observed under the Iberian Massif in northwest Iberia.

GEOLOGICAL HISTORY OF THE WESTERN IBERIAN CONTINENTAL MARGIN

Introduction

This section focuses on the timing of rifting, regional subsidence, and compressional events that contributed to the formation of the present-day structure of the west Iberia Margin.

Two problems must be faced in compiling information to compare the timing of tectonic episodes along the margin. The first is the fact that comparable data sets are not available from different sectors (Table 1): all have varying degrees of seismic profile coverage, but it is only in the Lusitanian Basin that profiles across the rift-basins have been calibrated using borehole data. The other problem concerns the criteria used by previous researchers to identify synrift packages on seismic sections. As shown by Wilson et al. (this volume) nearly all previously identified synrift packages in the offshore areas are probably postrift in origin.

Figure 7 is a summary of the Mesozoic and Tertiary stratigraphy of the west Iberia Margin. It shows in the right-hand column thirteen unconformity-bound sequences (UBS) defined by Wilson et al. (1989), Cunha (1992), and Pena dos Reis et al. (1992) that are related to significant events in the geological history of the margin. The discussion below summarizes the timing of tectonic events that can be inferred from this stratigraphic record.

Mesozoic Rifting and Subsidence

Three main rifting episodes have affected the west Iberia Margin: (a) a Late Triassic to Early Jurassic (Hettangian) rifting event, (b) a late Oxfordian/early Kimmeridgian rifting event well documented in the Lusitanian Basin (Wilson et al., 1989), and (c) a Valanginian/Hauterivian to Aptian rifting episode that culminated with the formation of oceanic crust and the separation of Iberia from the Grand Banks of Canada. These rifting episodes are well recorded in the stratigraphic record. Wilson et al. (1989) identified four Mesozoic unconformity-bound sequences in the Lusitanian Basin, which they related to events in the early evolution of the North Atlantic. Their paper reviewed in detail the tectonic implications of the three oldest

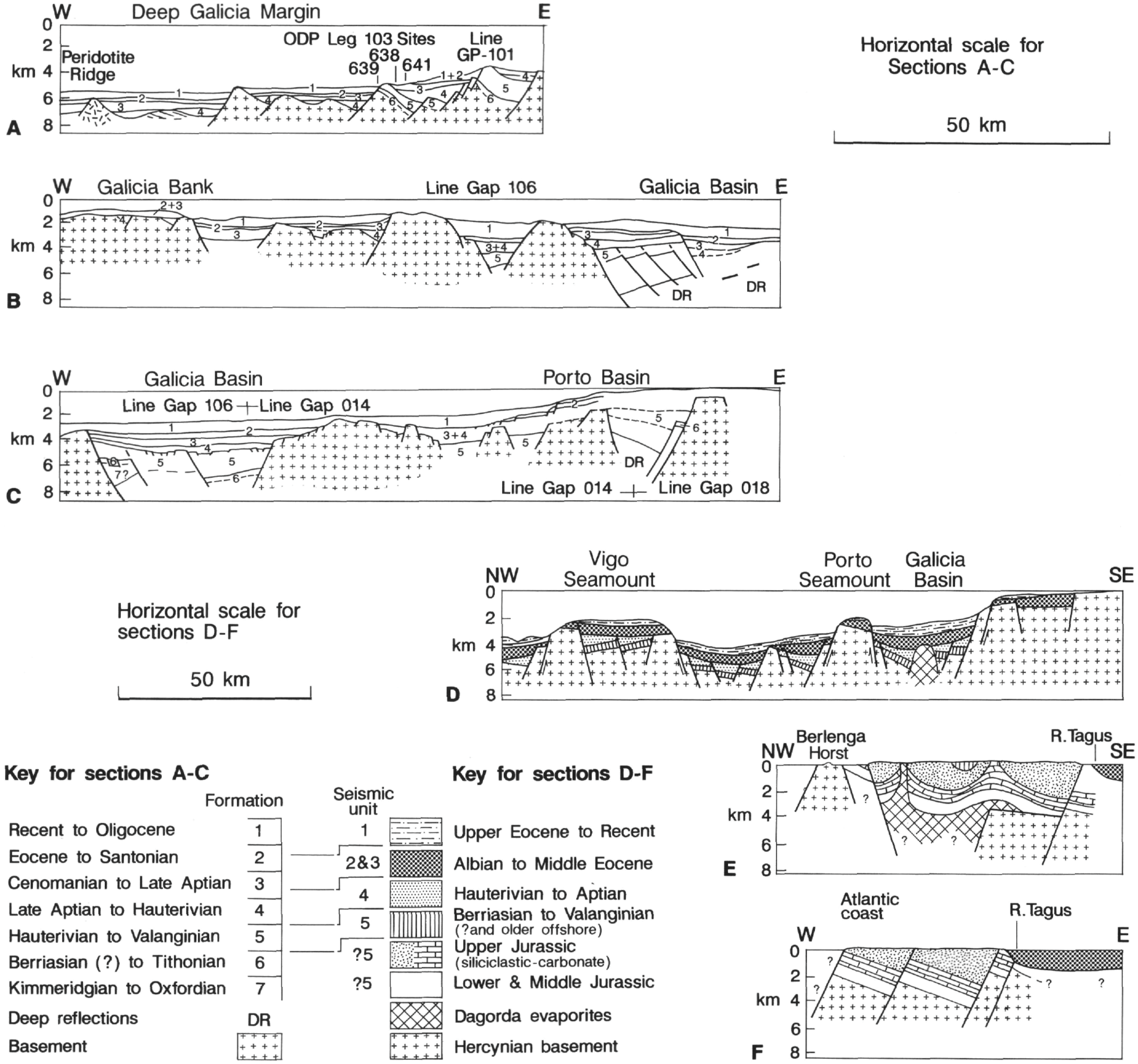

Figure 6. Sketch of structural sections along the west Iberia Margin (for locations, see Figure 5. **A-C** from Murillas et al. (1990); **D-F** from Wilson et al. (1989). The seismic units referred to in the key for D-F are those of Groupe Galice (1979). Note that A-C and D-F are drawn at different horizontal scales.

of these sequences, and so only a brief summary of their significance is given below. This is followed by additional information concerning the Aptian-Campanian sequence.

1. Triassic-Callovian. This sequence is typical of the early rift-sag successions encountered in most North Atlantic Margin basins. In the Lusitanian and Porto Basins this sequence is one to two kilometers thick. Triassic red fluvial siliciclastics are capped by Hettangian evaporites that subsequently influenced the manner in which extensional and compressional movements along Variscan basement faults affected the cover of younger sediments. Where the evaporites were thick, halokinetic structures formed above basement faults, but where they were thin, the faults propagated through younger formations. The Triassic and Hettangian sediment probably accumulated in grabens and half grabens. The later Lower and Middle Jurassic shales and carbonates exhibit simple fácies geometries indicating a westward-dipping ramp, with some localized indications of contemporaneous faulting related to the Berlenga ridge. The Late Triassic-Early Jurassic rifting event resulted in thick salt deposition along the Portuguese and Grand Banks Margins (Pautot et al., 1970; Tankard and Welsink, 1987; Grant et al., 1988; Mougenot, 1988; Austin et al., 1989; Mauffret et al., 1989a,b). Soares et al. (1993) divided the Triassic-Callovian of the northern part of the Lusitanian Basin into eight megasequences bounded by discontinuities. They sug-

Table 1. Summary of data sets available from sectors of the Western Iberian Margin.

	Seismic	Boreholes	Seafloor	Outcrop	Key references
Deep Galicia Margin	SC, MC (u,m)	ODP Leg 103	D, S, (s)	—	Mauffret and Montadert, 1988 Boillot and Winterer, 1988
Iberia Abyssal Plain	SC, MC (m)	ODP Leg 149	—	—	Wilson et al., this volume
Western Banks	SC, MC (u)	DSDP Site 398 (Vigo Seamount)	D	—	Murillas et al., 1990; Groupe Galice, 1979
Galicia Interior Basin	SC, MC (u)	—	—	—	Murillas et al., 1990; Mougenot, 1988; Groupe Galice, 1979
Porto Basin	SC, MC (u)	8 offshore	D	—	—
Lusitanian Basin	MC (m)	38 onshore* 14 offshore	D	Yes	Wilson et al., 1989

Note: SC = single channel; MC = multichannel; (m) = migrated; (u) = unmigrated; D = dredging; S = sampling; (s) = submersible; — = not available; * = deeper than 500 m.

gested that rifting continued into the early Liassic, whereas Wilson et al. (1989) indicated that thermal subsidence began at the beginning of the Jurassic.

2. Oxfordian-Tithonian. In the Lusitanian Basin, the base of this sequence is marked by a basin-wide late Callovian-early Oxfordian hiatus, associated with karst features. The middle Oxfordian is characterized by lacustrine and marginal marine carbonates and clastics, and some evaporites. Fully marine carbonate deposition returned to the Lusitanian Basin in the late Oxfordian, during which time apparent basement subsidence rates increased up to 200 m/m.y. (Hiscott et al., 1990). A sudden influx of coarse siliciclastic sediments at the end of the Oxfordian was accompanied by a further increase in subsidence rates (to 270 m/m.y. in one sub-basin), and the first significant diapirism occurred at this time. In the southern part of the Lusitanian Basin, half-graben basins developed in which Kimmeridgian siliciclastics clearly indicate the development of syn- and immediate postrift basin fill (Leinfelder and Wilson, 1989; in press). Rapid basement subsidence rates and a complex distribution pattern of early Kimmeridgian fácies support the conclusion that this interval marks a significant transtensional rifting phase (Wilson et al., 1989). Younger sediments exhibit simpler fácies distributions characteristic of a postrift episode, and carbonate shelf systems developed over Galicia Bank (Boillot, Winterer, Meyer, et al., 1987; Jansa et al., 1988), in the Porto Basin and in the southern part of the Lusitanian Basin. The complete sequence is 2-4 km thick in the Lusitanian Basin (1-3 km synrift and approximately 1 km postrift), but less than 2 km in the Porto Basin, which suggests that rift-related subsidence was less significant in the north than it was to the south.
3. Valanginian-Early Aptian. On shore and in the Porto basin, this interval shows a relatively simple fácies geometry, with largely fluvial siliciclastic sands and conglomerates interfingering with shallow water carbonates. It is relatively thin (up to 500 m), indicating relatively low subsidence rates. On the Galicia Margin, the sediments of this age are interpreted as synrift sediments that heralded ocean opening in late Aptian times between Iberia and the Grand Banks. The source of the Lower Cretaceous sediments was probably the Galicia Bank, since sediments eroded from the Iberian mainland were trapped in the Interior Basin (Winterer et al., 1988). This episode is also well documented on the Galicia Bank margin (Boillot et al., 1985, 1989b; Boillot and Winterer, 1988; Moullade et al., 1988; Murillas et al., 1990; Malod and Mauffret, 1990). However, Wilson et al. (this volume) conclude that the intervals previously recognized as synrift beneath the deep Galicia Margin are in fact immediate postrift, as defined by Prosser (1993), as no convincing reflection divergence towards fault planes can be observed on published seismic sections from this area. They suggest that depositional rates during the synrift episode were extremely low, so that sequences thick enough to image by reflection seismic methods did not accumulate. In the light of this conclusion, they suggest that the synrift episode occurred after the deposition of the Tithonian-Berriasian carbonate platform on the Galicia Margin, but before the Valanginian siliciclastic sediments previously identified as synrift. This interpretation indicates a very short-lived period of rifting off the deep Galicia Margin, spanning a few million years from the latest Berriasian to earliest Valanginian.
4. Late Aptian-Early Campanian. In the Lusitanian Basin the base of this sequence is marked by the abrupt and widespread onset of coarse-grained siliciclastic sediments. Sediment was transported southwestward from the Hesperic Massif across coalescent alluvial fans that developed in a humid environment; these merge in the same direction into coastal marine sediments, including carbonates (Berthou, 1973; Soares, 1980; Dinis and Pena dos Reis, 1989). During the Turonian, localized movement of diapiric structures is indicated by an unconformity, karstification, and locally derived sediments (Pena dos Reis, 1993). Berthou (1973), Berthou and Lauverjat (1979), and Boillot et al. (1972; 1975) suggested that reactivation of the Nazaré-Lousã Fault at the end of the Cenomanian resulted in uplift of areas to the south of the structure, causing a northward displacement of marine sedimentation. However, we believe that the main movement was later, probably during the middle Campanian, since the upper Campanian fluvial sediments show a northerly drainage pattern in contrast to the earlier southwesterly direction. The top of the upper Aptian-Campanian sequence in the northwestern margin of the Lusitanian Basin is characterized by a thick silcrete, which indicates a long period of nondeposition during the Santonian?-early Campanian (Pena dos Reis and Cunha, 1989a; b; Cunha et al., 1992).

Kinematics and Timing of the Deformation

The overall tectonic style of the continental margin deformation acquired during the three main rifting episodes is essentially characterized by a series of tilted fault blocks bounded by normal faults (mostly dipping to the west), which delineate a series of half-graben structures (Mauffret and Montadert, 1987; Thommeret et al., 1988; Sibuet, 1992). Off Galicia, these half-graben structures trend predominantly north-south, as shown by the available detailed seismic and seabeam coverage (Sibuet et al., 1987). In the southern Iberia Abyssal Plain and in the Tagus Abyssal Plain, however, the trend of the continental blocks is more variable (Whitmarsh et al., 1990; Pinheiro et al., 1992; Pinheiro, 1994; Whitmarsh et al., 1995). These tilted fault blocks can be continuous for a few tens of kilometers off Galicia, but they appear to be more intensely disrupted further south, being generally offset by northeast-southwest transverse faults that

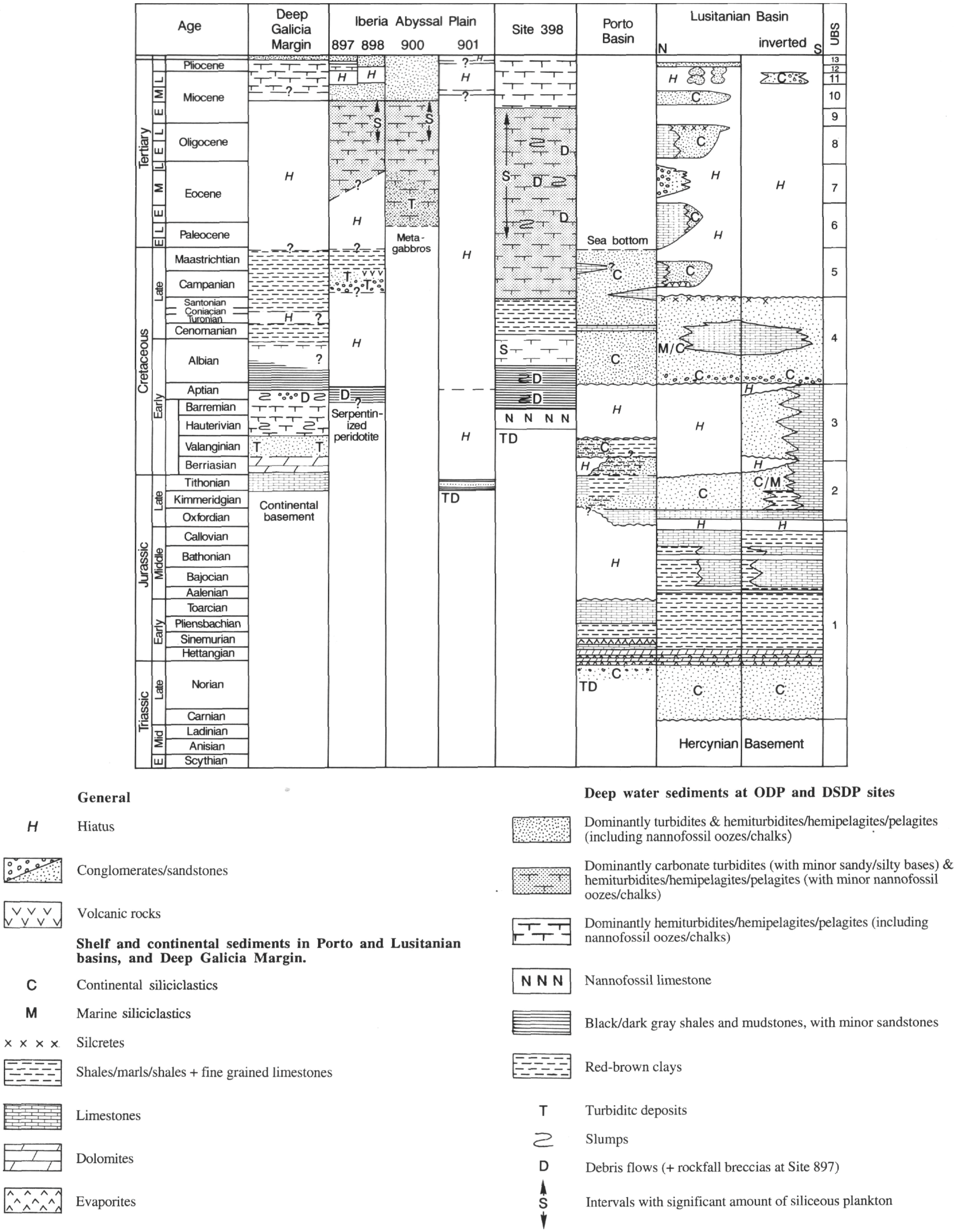

Figure 7. Simplified summaries of stratigraphic successions along the west Iberia Margin.

probably correspond to late Variscan continental strike-slip faults reactivated during rifting (Ribeiro et ai., 1979; Sibuet, 1992; Ribeiro and Silva, in press). The faulting associated with rifting was accompanied by large-scale mass movements from the crests of the tilted blocks.

Off Galicia, the end of continental extension has been dated at approximately the Aptian-Albian boundary (112 Ma; Boillot et al., 1989b). However, Wilson et al. (this volume) argue that synrift sediments cannot be identified on the available seismic data and suggest that the continental extension in this segment of the margin may have ended in the late Berriasian-early Valanginian.

The mechanism of lithospheric extension is still not clear, but it seems that a composite model involving pure-shear affecting the whole lithosphere and simple-shear affecting only the upper brittle crustal levels along a low-angle detachment that soles out at the brittle/ductile transition can best explain the observations (Sibuet, 1992). A very strong seismic reflector imaged on the available seismic sections west of Galicia Bank—the S-reflector—is a good candidate for such a detachment (Boillot et al., 1989b; Sibuet, 1992; Reston et al., 1994; Sibuet et al., in press). A similar reflector was first detected in the Bay of Biscay, both on the Armorican and the Galician Margins and was called the S-reflector by de Charpal et al. (1978), who interpreted it as representing the brittle/ductile transition. Later, Boillot et al. (1989b) applied the simple shear model, initially proposed by Wernicke (1981) to explain the Basin and Range province, to the Galicia area and postulated that the S-reflector represents the contact between the thinned continental crust and serpentinized peridotite. More recently, Hoffman and Reston (1992) and Reston et al. (1994) confirmed the detachment hypothesis and Sibuet et al. (in press) have shown that this supposedly low-angle normal fault is not rooted in the mantle, although it possibly coincides with the top of the serpentinized peridotite layer in the lowermost part of the continental margin, close to the peridotite ridge. Further to the east, however, it appears to sole out in the middle crust (Sibuet et al., in press). In the Iberia Abyssal Plain, Beslier et al. (1993) have shown that a fairly strong S-like reflector, or set of reflectors, also occur locally at approximately the same depth as the S-reflector off Galicia, but their expression on the seismic sections is not as dramatic and they are far more discontinuous. No evidence of an S-like reflector has yet been found in the Tagus Abyssal Plain.

As regards the peridotite ridge, detailed structural and microstructural studies have shown that it was emplaced at seafloor level at the end of continental rifting or at the very beginning of oceanic accretion (Beslier et al., 1990). It successively experienced (1) a low degree of partial melting and strong ductile high-temperature-low-stress deformation under asthenospheric conditions during adiabatic uplift; (2) strong, but heterogeneous ductile deformation (mylonitization and ultramylonitization) under lithospheric conditions (high but decreasing temperature of 1000°-850°C and high deviatoric stress of 180 MPa); (3) brittle deformation; and (4) serpentinization by hydrothermal alteration by seawater (Girardeau et al., 1988; Beslier et al., 1990; Boillot et al., 1992; Beslier et al., 1993). The age of emplacement of the peridotite ridge off Galicia has been dated at 118-122 ± 0.9 Ma (early Aptian). This corresponds to $^{39}Ar/^{40}Ar$ dating of amphibole crystals and plagioclase neoblasts within a syntectonic dike that crosses the peridotite and which has undergone ductile deformation under a similar stress regime and physical conditions as the peridotite (Féraud et al., 1988; Boillot et al., 1992).

Cenozoic Compressional Episodes

The Pyrenean and Alpine orogenies in Iberia were the result of collisions between Iberia and Europe, and Africa and Iberia, respectively. In Iberia and along the west Iberia Margin, the most important compressive deformation occurred during the Miocene, although Eocene deformation occurred, particularly in the Hesperic massif (Boillot et al., 1979; Ribeiro, 1988, Ribeiro et al., 1990; Malod et al., 1993; Masson et al., 1994). The Eocene deformation episode is essentially related to the formation of the Pyrenean-Cantabrian Chain and to the incipient subduction developed in the southern Bay of Biscay (Sibuet and Le Pichon, 1971; Grimaud et al., 1982; Malod et al., 1993). The Miocene event is fundamentally related to the deformation of the Betic chain in southeastern Iberia (Sibuet and Le Pichon, 1971;Murillas et al., 1990).

Evidence Onshore

In the Lusitanian and Algarve Basins, the Miocene tectonic style is "thin-skinned," with the Hettangian evaporites acting as a décollement (Ribeiro, 1988; Ribeiro et al., 1990). In contrast, structures affecting basement rocks probably have deeper roots, possibly at the base of the crust. The Central Cordillera, for example, has been interpreted as a basement pop-up structure (Ribeiro, 1988; Ribeiro et al., 1990), the trend of which continues southwestward into the major zone of inverted Mesozoic rocks in the Lusitanian Basin (see Fig. 8). In this basin, the Miocene compression was directed north-northwest-south-southeast, oblique to the predominant north-northeast-south-southwest trend of Mesozoic extensional faults. This resulted in basin inversion with a strong transpressional component (Wilson et al., 1989).

The main tectonic events that affected the west Iberia Margin in the Cenozoic are well documented by several unconformity-bound sequences (UBSs; 5-13 in Fig. 7; Cunha, 1992, Pena dos Reis et al., 1992). These are discussed briefly below.

Late Campanian-Early Lutetian (UBSs 5 and 6)

This interval consists of two major sequences: (1) an upper Campanian-Maastrichtian sequence and (2) a Paleocene-lower Lutetian sequence; these are separated by an unconformity with a stratigraphic hiatus offshore and, probably, also onshore. This interval may have been influenced by the changing movement of Iberia relative to Europe, with the beginning of the compression at the northern border of Iberia (Grafe and Wiedmann, 1993). The reactivation of the Nazaré-Lousã Fault is signalled by a change in fluvial paleocurrent directions from southwesterly drainage during the earlier late Aptian-Campanian, to northwesterly during this interval. Significant volcanic activity occurs south of the Nazaré Fault, particularly in the Lisbon and adjacent areas. Here, subvolcanic complexes, basaltic extrusives, and associated dikes occur. Kullberg (1985), suggested that the alkaline ring complexes were emplaced along a deep north-northwest-south-southeast strike-slip zone. According to Ribeiro et al. (1985, 1990) rift migration elsewhere in the North Atlantic may have changed the stress field and deeply fractured the previously thinned continental margin.

Onshore, Maastrichtian yellowish quartzarenites and red mudstones (Antunes, 1979), interpreted as deposited by a meandering fluvial system draining to the northwest, change distally to transitional and marine environments (Pena dos Reis, 1983). Correlative diapiric reactivation built up coalescent peri-diapiric alluvial fans, transverse to the north-south salt structures. Offshore, very shallow marine fine siliciclastics and dolostones are dominant.

Late Lutetian-Chattian (UBSs 7 and 8)

During this interval compression intensified in the Pyrenees (Srivastava et al., 1990a,b). On the western border of the Hesperic Massif, extensional reactivation of northeast-southwest faults caused the formation of the Mondego and lower sub-basins (Cunha, 1992). In some areas, the lower boundary corresponds to an angular unconformity. A prograding conglomerate sandstone at the top of this interval is related to the uplift of the Hesperic Massif (Pena dos Reis, 1983). The sands and conglomerates show a broad northeast to southwest drainage pattern, which contrasts with the underlying Upper

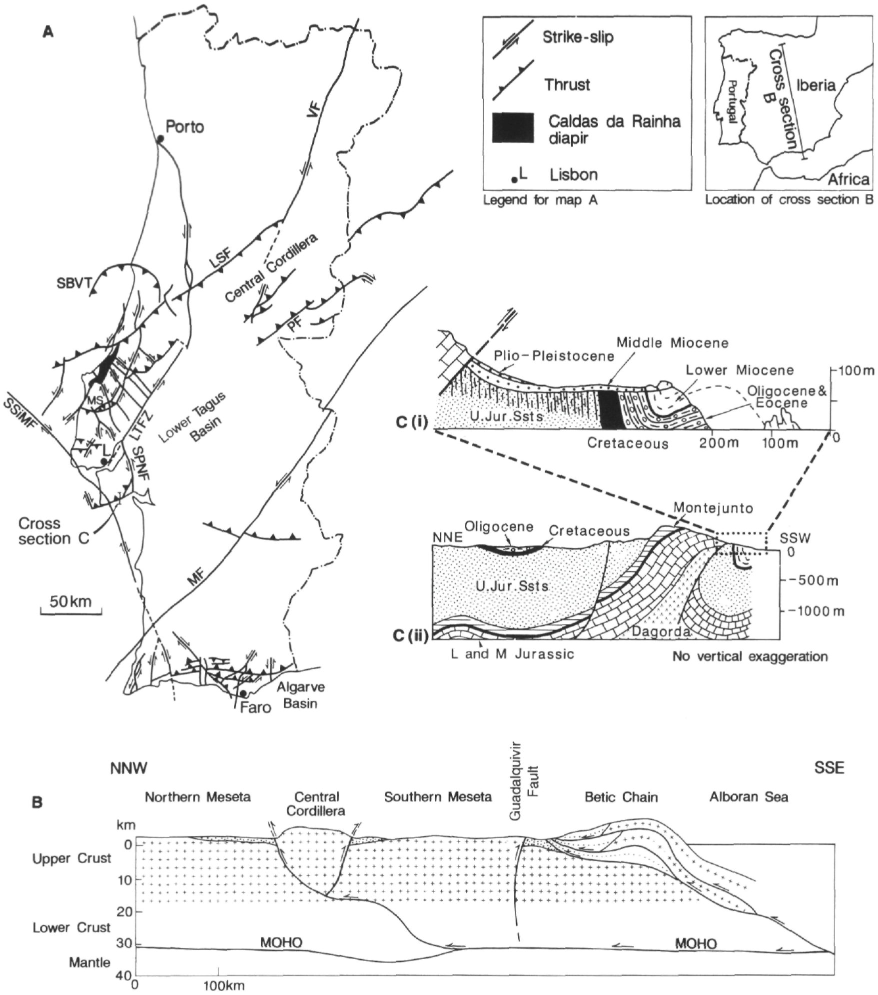

Figure 8. Neogene tectonic structures in Portugal. **A.** Map of structures active in Portugal during the Neogene. The inset map of Iberia shows the location of B (from Ribeiro et al., 1990). LSF = Lousã-Seia fault; LTFZ = Low Tagus Fault Zone; MF = Messejana Fault; MS = Montejunto Structure; PF = Ponsul Fault; SBVT = Serra de Boa-Viagem Thrust; SPNF = Setúbal-Pinhal Novo Fault; SSiMF = Sintra-Sines-Monchique Fault; VF = Vilariça Fault. **B.** Sketch section showing the possible origin of the Central Cordillera as a pop-up structure linked to a detachment at the level of the Moho (from Ribeiro et ai., 1990). **C.** Sketch sections showing the structure at the eastern end of the Arrábida hills. C(ii) shows the development of an asymmetric anticline above a reverse fault; the structure may have developed initially as a Mesozoic diapiric structure that was subjected to transpressional movements during the Miocene. C(i) is a detail of the eastern portion of C(ii) and shows stratigraphic relationships that indicate significant deformation around the early middle Miocene boundary. Dagorda: Late Triassic and Hettangian evaporites; Montejunto: Oxfordian carbonates. From Wilson et al. (1989).

Cretaceous paleocurrent trend. The upper boundary is very well defined by a thick silcrete (Meyer and Pena dos Reis, 1985).

Late Chattian-Early Tortonian (UBSs 9 and 10)

During this period, Iberia joined the Eurasian Plate and the plate boundary with Africa migrated to the Azores-Gibraltar Fracture Zone (Fig. 1F). The sedimentary record can be divided into two unconformity-bounded sequences: (1) an upper Chattian-upper Burdigalian sequence, and (2) an upper Burdigalian-lower Tortonian sequence. The lower boundaries of these sequences correspond to the Castillan and Neocastillan tectonic phases, respectively (Pena dos Reis and Cunha, 1989b). The lower sequence is only preserved beneath the continental shelf in the Tagus sub-basin, and consists of marine, brackish, and continental fácies (Antunes et al., 1987). The upper sequence occurs in both the Tagus and Mondego sub-basins, and includes fine sands, plus lacustrine and shallow marine carbonates.

Late Tortonian-Lower Calabrian (UBSs 11-13)

During this interval, three alluvial fan sequences in the Mondego sub-basin are interpreted to result from distinct pulses of uplift (UBS 11: *late Tortonian-early Messinian;* UBS 12: *late Messinian-Zandian;* UBS13; *Piazencian-early Calabrian*?). Carvalho et al. (1983) and Ribeiro et al. (1990) suggested that the main uplift of the Central Cordillera (Fig. 8C) occurred during the Tortonian, although stratigraphic relationships in the Serra da Arrábida indicate that the principal deformation occurred around the early middle Miocene boundary. UBSs 11-13 are affected by neotectonic structures, and precede Quaternary fluvial incision.

Evidence Offshore

Offshore, there is strong evidence of Cenozoic reactivation of late Variscan faults as reverse faults or transpressive strike-slip structures, particularly during the Eocene and the Miocene (Boillot et al., 1979; Mauffret et al., 1989a; Masson et al., 1994; Pinheiro, 1994). This deformation appears to continue today, as suggested by localized presence of seafloor scarps above compressional structures within the ocean/continent transition, accompanied by deformation of recent sediments (Pinheiro, 1994). Off Galicia Bank, the peak of the deformation occurred in the Eocene (Boillot et al., 1979). In the Tagus and in the southern Iberia abyssal plains, however, the main episode of deformation occurred in the middle Miocene, and it may have overprinted the former Eocene deformation (Mougenot, 1988; Mauffret et al., 1989a,b; Masson et al., 1994; Pinheiro, 1994). The Cenozoic deformation in the abyssal domain is essentially characterized by a broad oceanward-facing monoclinal sequence that can be several tens of kilometers wide, and which invariably terminates in a tight asymmetric fold, generally associated with a steeply dipping fault or faults (Masson et al., 1994). In general, such faults do not seem to be directly rooted in any major basement faults. In both the Iberia and the northern Tagus abyssal plains, the overall pattern of the observed deformation is compressional (or transpressional). However, there is evidence of a later extensional or transtensional regime, that could be local, but which has caused a tectonic subsidence/collapse, in places, of the original compressional structures (Pinheiro, 1994). This latter episode apparently preceded the deposition of the seismic stratigraphic Unit 1A (late Miocene to Holocene) defined by Mauffret and Montadert (1988), and therefore also occurred during the middle Miocene.

As shown by Masson et al. (1994) and Pinheiro (1994), there is a close spatial link between the western termination of the deformation zone and the ocean/continent transition. In the southern Iberia Abyssal Plain, the deformation zone generally ends abruptly at the presumed contact between the transitional crust that characterizes the ocean/continent transition and the thin oceanic crust. This coincides with the location of the serpentinized peridotite ridge that marks the ocean/continent transition in this area and which was drilled during this leg at Site 897. There are some seismic lines on the Tagus and on the Iberia Abyssal Plains, however, in which the deformation zone appears to terminate just west of the presumed location of the serpentinized peridotite ridge, over what has been interpreted as an area of thin oceanic crust underlain by serpentinized upper mantle (Pinheiro, 1994). Smaller, but similar, compressional (transpressional?) structures also occur much further to the east, in an area of the ocean/continent transition interpreted as thinned continental or transitional crust underlain by serpentinized upper mantle (Pinheiro, 1994). This strongly suggests that the contrasts in lithologies and rheologies that take place across the ocean/continent transition (as inferred from the interpretation of the geophysical data: see Whitmarsh et al., 1990; Pinheiro et al., 1992; Whitmarsh et al., 1993; Masson et al., 1994; Pinheiro, 1994), together with the fact that the crustal section is very thin and it is underlain by serpentinized peridotite, have probably played an important role in controlling the fold propagation and the pattern and lateral extent of the deformation zone. The existence of large volumes of serpentinized peridotite at shallow depth, in particular, appears to have been the crucial factor on the control of the fold propagation, since it is well known that these rocks, under stress, tend to deform in a fairly ductile fashion and to accommodate a significant part of the deformation (Murton, 1986; Reinen et al., 1991).

The observations discussed above suggest the reactivation, in an essentially transpressional strike-slip regime related to the stress field generated by the oblique northwest-southeast Africa-Iberia-Eurasia convergence during the Cenozoic, of the north-northeast-south-southwest normal faults that controlled the Mesozoic rifting. The change from transpression to transtension observed on some of the available seismic lines could be local and it may have been caused by small adjustments in the plate motion at the end of the middle Miocene. Other possible concurrent causes could be phenomena of local flow within the underlying serpentinite, or the formation of localized pull-apart zones in the areas of relay (or transfer) between the overlapping adjacent fault segments, caused by the interaction between the slip movement on the faults (Pinheiro, 1994).

Magmatic Activity in the Mesozoic

Introduction

Magmatic activity in western Iberia during the Mesozoic is present both in the Hesperic Massif (Central Iberian Zone) and in the Lusitanian Basin. The main areas of such activity in Portugal are shown in Figure 9; they include dikes of lamprophyres, dolerites, and basalts (alkaline, transitional, and tholeiitic) as well as minor hypovolcanic massifs (Ribeiro et al., 1979; Martins, 1991). The occurrences, however, are volumetrically insignificant in comparison to volcanism that preceded and accompanied continental rifting and separation between northwest Europe and Greenland.

Mesozoic igneous rocks are largely confined to the west Iberia Margin south of the Nazaré Fault, and none have been found north of Figueira da Foz. Four main phases of Mesozoic magmatic activity have been identified:

1. A Late Triassic-Early Jurassic alkaline basaltic cycle (203-224 Ma)
2. An Early Middle Jurassic tholeiitic cycle (190-160 Ma)
3. An Early Cretaceous transitional cycle (135-130 Ma)
4. A Late Cretaceous alkaline cycle (100-70 Ma)

The first three cycles are associated with the three main rifting episodes (Ribeiro et al., 1979; Martins, 1991).

Late Triassic-Early Jurassic Alkaline Cycle

The oldest magmatic phase is restricted to the Central Iberian Zone and is represented by dolerite and porphyry dikes associated with the Variscan granites, as well as by several occurrences of lamprophyres and dolerite-basaltic dikes with K-Ar ages of 224 ± 11 Ma (Late Triassic). These suggest an alkaline basaltic activity with a shoshonitic tendency (Ribeiro et al., 1979), probably related to the last phases of the Variscan orogeny in Iberia. The dikes appear to intrude along fractures possibly related to postorogenic extension (Ribeiro et al., 1979). Basaltic dikes with ages of 203 ± 3 Ma (Hettangian/Sinemurian) have also been reported in this area (Ribeiro et al., 1979; Martins, 1991). All the ages referred to in this section ("Magmatic Activity in the Mesozoic") have been taken from syntheses in Ribeiro et al. (1979) and Martins (1991).

Early Middle Jurassic Tholeiitic Cycle

This episode is represented in western Iberia by the intrusion of basaltic and doleritic dikes within the Hettangian-Sinemurian sedimentary formations of the Mesozoic-Cenozoic Borderland (Algarve, Bordeira, Santiago do Cacem and Sesimbra) and by the Messejana dikes along the Odemira-Ávila Fault (Fig. 9). There is also clear evidence of Early Jurassic tholeiitic magmatism in the Algarve (190 to 180 Ma; Ribeiro et al., 1979; Ferreira and Macedo, 1979). These tholeiitic volcanics characterize an area with similar sedimentological and volcanic character to the margins of the Central Atlantic (i.e., the northwest African Margin from Senegal to Morocco and the northeast United States from the Bahamas to Nova Scotia; May, 1971; Ribeiro et al., 1979; Martins, 1991). Based on the ages obtained for the intrusions along the Odemira-Ávila Fault, Martins (1991) proposed that the tholeiitic activity could extend from 190 to 160 Ma (early Middle Jurassic), with most of the magmatic activity concentrated in the period 190-180 Ma. Judging from stratigraphic relationships, however, it is possible that the magmatic activity might have started somewhat earlier, at about 205 Ma (Rhaetian-Hettangian boundary; Martins, 1991).

Early Cretaceous Transitional Cycle

Late Jurassic-Early Cretaceous intrusives occur mainly in the Lusitanian Basin (Ribeiro et al., 1979; Martins, 1991) as north-northeast-south-southwest and east-southeast-west-northwest dikes (consisting of dolerites, gabbros, and diorites) and also as minor hypovolcanic massifs in the Leiria area. The intrusions are mostly confined to an area between two major tectonic lineaments: the Nazaré Fault (NF) and the Serra d'Aire-Montejunto lineament (SAML) (Fig. 9; Ribeiro et al., 1979). Evidence of such activity between the Oxfordian and the Hauterivian has been recognized in Soure, Codiceira, Vermoil, Leiria, and Rio Maior (Fig. 9). The basaltic dike of Caceia (126 Ma) in the Algarve could also belong to this cycle (Ribeiro et al., 1979). With the exception of Soure (159 ± 3 Ma) and Codiceira (165 ± 3 Ma), all the K-Ar ages for these intrusions range between 130 ± 3 and 144 ± 2 Ma, and concentrate mainly in the interval 135-130 Ma. Martins (1991) and Ferreira and Macedo (1987) considered the ages of 159 and 163 Ma obtained for Soure and Codiceira, respec-

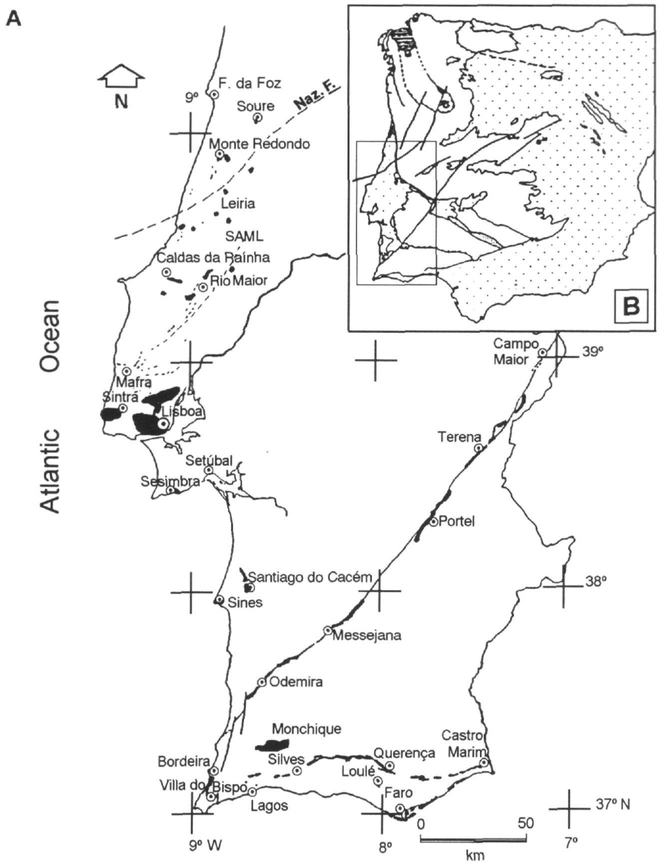

Figure 9. A. Areas of occurrence of Mesozoic igneous rocks (in black) in western Iberia. SAML = Serra d'Aire-Montejunto Lineament; Naz. F. = Nazaré Fault. B. Inset showing the location of A in the Iberian Peninsula.

tively, as not reliable, due to alteration of the samples. Therefore, according to these authors, the magmatic activity in this period is concentrated in the interval 135-130 Ma (Valanginian/Hauterivian), although Willis (1988) reports a 140 Ma age (Berriasian) for east-southeast-west-northwest dikes to the east of Caldas da Rainha.

Late Cretaceous Alkaline Cycle

The main onshore occurrences related to this episode include the radial dike complex near Mafra (ca. 100 Ma), the Basaltic Complex of Lisboa (72.5±3 Ma), and the Sines, Sintra, and Monchique massifs (80 to 70 Ma) (Fig. 9). The subvolcanic massifs of Sintra, Sines, and Monchique show ages that decrease to the south (Ribeiro et al., 1979) and a similar suggestion of a decrease in age of the alkaline magmatism towards the south is also suggested by the relative ages of the basaltic complex of Lisboa (K-Ar isochron of 72.5±3 Ma), compared with the Mafra complex (about 100 Ma; Fig. 9; Ribeiro et al., 1979). Other volcanic manifestations of the same period include a few outcrops NW of the Serra d'Aire-Montejunto lineament and a few volcanic plugs (e.g., in the Nazaré area; Ribeiro et al., 1979). Evidence of alkaline volcanism is also found in the western Algarve, where K-Ar whole rock ages between 72 and 77 Ma have been reported (Martins, 1991). In the offshore, there is also evidence of alkaline volcanism. Alkaline basalts with an $^{40}Ar/^{39}Ar$ age of 74 ± 0.7 Ma were recovered from the Estremadura Spur and alkaline rocks of Maastrichtian-Paleocene age (60-74 Ma), with a similar petrology to those at the Sines and Monchique, overlie oceanic gabbros in Gorringe Bank (Féraud et al., 1977; Mougenot, 1988).

In summary, it seems clear that the Mesozoic volcanic igneous activity in western Iberia is scattered and relatively minor (Fig. 9). In fact, the synrift volcanism associated with the final rifting event that culminated with seafloor spreading is confined to a small area between the Nazaré Fault and the Serra d'Aire-Montejunto Lineament and appears to have occurred essentially in the period 135-130 Ma (Valanginian/Hauterivian). As such, the west Iberia Margin can be considered a nonvolcanic margin.

THE OCEAN/CONTINENT TRANSITION OFF WESTERN IBERIA

Crustal and Upper Mantle Structure Across the Ocean/continent Transition

The integration of all the available geophysical data and modeling results, constrained locally with data from deep ocean drilling (DSDP Leg 47B: Groupe Galice, 1979; Sibuet and Ryan, 1979; ODP Leg 103: Boillot, Winterer, Meyer, et al., 1987; ODP Leg 149: Sawyer, Whitmarsh, Klaus, et al., 1994), allows the definition of a reasonably well-constrained picture of the crustal and upper mantle structure across the ocean/continent transition in the three main segments of the west Iberia Margin (Boillot et al., 1980; 1986; 1989a,b; 1992; Whitmarsh et al., 1990; Pinheiro et al., 1992; Horsefield, 1992; Whitmarsh et al., 1993; Pinheiro, 1994; Whitmarsh and Miles, 1995). The designation ocean/continent transition is used here, instead of the term ocean/continent boundary (OCB) often referred to in the literature, because it now seems clear that the transition between continental and oceanic crust at nonvolcanic passive margins does not generally occur at a sharp boundary but, instead, it occurs over a transition zone, in which the crustal section has characteristics intermediate between continental and oceanic crust, and beneath which serpentinized upper mantle rocks may occur.

In broad terms, the ocean/continent transition off western Iberia (Figs. 10A, B) is characterized by an exceptionally thin crust (less than 6 km thick, typically 2-4 km thick, and, in places, virtually nonexistent), underlain by a fairly thick layer (normally 2-3 km but up to 7 km) of a material with a high *P*-wave velocity ($7.2 \le V_P \le 7.9$ km/s). These high seismic velocities are intermediate between those generally observed in the lower continental or oceanic crust (normally lower than 7.1-7.2 km/s in the absence of crustal underplating) and those that characterize the uppermost "normal" mantle (8.0-8.1 km/s). Such material has been interpreted as serpentinized upper mantle (Pinheiro et al., 1992; Whitmarsh et al., 1993; Pinheiro, 1994). The idea of serpentinized peridotite beneath the thinned continental crust of passive margins was initially proposed by Boillot et al. (1987; 1989a). Off Galicia Bank, serpentinized upper mantle rocks outcrop on the ocean floor at the western edge of the ocean/continent transition, and form a basement ridge that can be followed almost continuously by sampling and by seismic reflection profiles for more that 130 km (Boillot et al., 1989a,b). Based on the interpretation of seismic reflection profiles, Beslier et al. (1993) postulated the continuation of this ridge into the southern Iberia Abyssal Plain; the proposed location of the ridge coincided with the inferred landward edge of oceanic crust formed by seafloor spreading, based on geophysical criteria (Whitmarsh et al., 1990; 1993). Drilling results from Leg 149 have shown this interpretation to be correct (Sawyer, Whitmarsh, Klaus et al., 1994). The existence of a serpentinized peridotite ridge in an analogous location in the Tagus Abyssal Plain, although possible, has not been yet demonstrated.

The width of the area of the crust less than 6 km thick underlain by high velocity material (7.4 km/s) that characterizes the ocean/continent transition off west Iberia may reach 110-150 km in the Tagus and southern Iberia Abyssal Plains but does not exceed more than a few tens of kilometers in the Galicia segment (Fig. 10B). The nature of this thin crust underlain by serpentinized upper mantle is not clear but its geophysical signature (in particular the magnetic character and the seismic-velocity profile) indicates that a major change in crustal structure occurs across its width.

The western portion of the thin crust underlain by high seismic-velocity material (i.e., that which is situated to the west of the serpentinized peridotite ridge in the south Iberia Abyssal Plain and off Galicia Bank) exhibits an acoustic basement topography and character typical of oceanic crust and its seismic crustal structure is remarkably similar to those that characterize oceanic fracture zones (very thin crust with seismic velocities in the crustal section compatible with oceanic layer 2 but absence of layer 3). In particular, the basement relief and associated magnetic anomalies are generally elongated and show a fairly well-defined north-northeast-south-southwest trend, similar to those observed on the oceanic crust further to the west. Also, in the Iberia Abyssal Plain and the Tagus Abyssal Plain, the observed magnetic anomalies can be modeled by seafloor spreading, which clearly indicates that the underlying crust is thin oceanic crust. This thin oceanic crust was formed during the first stages of oceanic accretion. Although the basement to the west of the peridotite ridge has not been drilled on the west Iberia Margin, oceanic basalts were found covering the western side of the peridotite ridge on the northwestern slope of Galicia Bank (Malod et al., 1993).

On the contrary, the transitional crust east of the peridotite ridge shows a basement topography which is often smoother but more irregular than that observed in either typical oceanic or continental domains, and it exhibits seismic velocities in the lower crust that are different from those observed in oceanic layer 3, and which are more compatible with the seismic velocities that characterize the lower continental crust further to the east (Line 1, from Whitmarsh et al., 1990, and onshore). The associated magnetic anomalies exhibit a similar trend to that observed on the oceanic crust to the west of the peridotite ridge (Whitmarsh and Miles, 1995) but their amplitude is generally much lower, although, locally, they may reach several hundred nT. The basement blocks tend to be more equidimensional and exhibit more variable trends that range from north-south to northwest-southeast (Whitmarsh et al., 1990; Pinheiro et al., 1992; Pinheiro, 1994; Whitmarsh and Miles, 1995). Such a crust is interpreted as transitional, and it most probably consists largely of extremely thinned and highly intruded continental crust (similar to the Jizan area, in the Red Sea; Voggenreiter et al., 1988), within which local areas of serpentinized upper mantle rocks may occur, given the extreme degree of thinning of the crustal section (Pinheiro, 1994; Whitmarsh and Miles, 1995). Alternative interpretations are that significant portions of this crust represent oceanic crust formed in a very slow-spreading environment or that a significant portion of the supposedly very thin crust represents serpentinized upper mantle (Boillot et al., 1992; Whitmarsh and Miles, 1995). As shown by the drilling results from ODP Legs 103 and 149, the seismic velocities at the top of the serpentinized peridotite ridge can be very low (3.5-4 km/s), but increase strongly with depth, passing through values within the range normally observed in the crustal section and reaching at depth the very high values intermediate between lower crust and upper mantle. In addition, although no significant magnetic anomalies are generally associated with the peridotite ridge, locally, some prominent anomalies are associated with serpentinite breccias, as shown by drilling at Site 899 (Sawyer, Whitmarsh, Klaus et al., 1994; Whitmarsh et al., this volume). Therefore, it is not easy to distinguish a zone of serpentinized upper mantle from the adjacent transitional or thin oceanic crust based only on geophysical criteria.

Figure 10A shows the main features of the ocean/continent transition along the west Iberia Margin, as determined from an integration of all the available geophysical data, and Figure 10B shows typical cross sections of the crustal and upper mantle structure across the ocean/continent transition in the three main segments of the margin (Whitmarsh et al., 1990; Horsefield, 1992; Pinheiro et al., 1992; Pinheiro, 1994).

The main features of the ocean/continent transition show similar trends in all three segments, but, northwards along the margin, the location of the ocean/continent transition is progressively offset toward the west in each consecutive segment. Such offsets could indicate hiatuses in the northward propagation of the rifting, probably caused by the intersection of the propagating tip of the rift with a major preexisting lithospheric weakness, such as a late Variscan fault. One such offset is observed in the area of the Estremadura Spur, which is located just south of the Nazaré Fault. The other major offset occurs at the southern edge of Galicia Bank, and it may also be related to an old late Variscan fault zone.

Timing of the Onset of Seafloor Spreading

Tagus Abyssal Plain

Mauffret et al. (1989a,b) postulated the existence of a fossil Late Jurassic to Early Cretaceous spreading center in the Tagus Abyssal Plain (magnetic anomalies M-21 to M-16), abandoned prior to a ridge jump towards the Grand Banks of Canada, sometime before magnetic anomaly M-10. More recently, however, Pinheiro et al. (1992) have shown, through the integration of all the available geophysical data and magnetic modeling of selected profiles, that such an hypothesis is unlikely. These later authors have shown that the magnetic anomalies predicted by the model of Mauffret et al. (1989a,b) did not fit the observed data. Furthermore, they showed that the oldest magnetic anomaly in this area that could be modeled with seafloor spreading was anomaly M11, which gives an age for the onset of seafloor spreading in this area of 133 Ma, with a half-spreading rate of 10 mm/yr. Recently, these results have been refined by Whitmarsh and Miles (1995; see also Whitmarsh et al., this volume), who modeled eight magnetic profiles across the ocean/continent transition in this area, using a new magnetic chart with a slightly better data coverage than the one used by Pinheiro et al. (1992). The new results basically confirm the Pinheiro et al. (1992) interpretation and show that the most landward magnetic anomaly that can be modeled with seafloor spreading is anomaly M-10NN, which gives a timing for the onset of seafloor spreading of 132 ± 1.9 Ma (late Valanginian/early Hauterivian, according to Gradstein et al., 1994), with a half-spreading rate of 10 mm/yr.

Southern Iberia Abyssal Plain

In the southern Iberia Abyssal Plain, recent results of modeling sea-surface magnetic profiles (Whitmarsh et al., 1990; Whitmarsh and Miles, 1995) have shown that the most landward magnetic anomaly that could be modeled with seafloor spreading was anomaly M-8, which gives an age of ca. 129.5 Ma (middle Hauterivian) for the onset of seafloor spreading in this area, with a half-spreading rate of 9 mm/yr. However, more detailed modeling of one deep-towed magnetic profile across the same area (Whitmarsh and Miles, 1995) suggested that the onset of seafloor spreading in this area took place approximately at 126 Ma (beginning of anomaly M3; early Barremian) at a half-spreading rate of 10 mm/yr.

Galicia Bank

Off Galicia Bank, no magnetic modeling has been attempted, since the J-anomaly is absent in this area and therefore the oceanic crust was formed in the Cretaceous Magnetic Quiet Interval (Ogg, 1988). As such, neither the age of the onset of seafloor spreading nor the half-spreading rate are known for this segment of the margin. Drilling results have dated the interpreted breakup unconformity at the late Aptian, near the Aptian/Albian boundary (112 ± 1.1 Ma; Boillot et al., 1989b). This coincides with the boundary between UBSs 3 and 4 (see Fig. 7). However, in the light of the Wilson et al. (this volume) conclusion that rifting in this area is late Berriasian-early Valanginian, the age of the so-called breakup unconformity may significantly postdate the commencement of seafloor spreading in this area.

CONCLUSIONS

From the above review of the literature of the west Iberia Margin, several important conclusions can be drawn:

1. The west Iberia Margin is an example of a nonvolcanic rifted continental margin. Onshore, there is evidence of only very limited intrusive and extrusive activity accompanying the rifting. Offshore, there is no direct evidence of synrift volcanism but there is indirect evidence of important volumes of lower crustal intrusives within part of the ocean/continent transition.
2. Late Variscan strike-slip faults and trends played a key role in influencing the development and evolution of the west Iberia Margin, not only onshore, in the formation of sedimentary basins, but also offshore, in the way in which rifting propagated, possibly intermittently, along the margin.
3. There has been a long history of rifting on this margin beginning in the Late Triassic. Continental breakup occurred from south to north in the Early Cretaceous between 134 and possibly 112 Ma, beginning about 70 Ma after rifting first began.
4. The ocean/continent transition, that region between the edge of the continental shelf and the landward edge of thin oceanic crust generated at the onset of seafloor spreading, is marked by a peridotite basement ridge and remains a key area of research. The oceanward part of the ocean/continent transition is least understood yet probably retains the most critical information regarding the tectonic and magmatic processes that controlled the temporal and spatial transition from continental to oceanic crust. The most plausible, but not the only, explanation for it seems to be that this part of the ocean/continent transition consists of fragments of magmatically disrupted and intruded thinned continental crust.
5. This margin has twice, in Eocene and Miocene times, suffered a compressional episode since rifting ended. The effect of these two compressional episodes on the margin has been to fold, reactivate old Variscan structures, some of which had been previously reactivated during the rifting episodes, and cause minor faulting within the abyssal plain sediments. Onshore, many of the main late Variscan structures were also reactivated.

ACKNOWLEDGMENTS

We thank G. Boillot, J.-C. Sibuet, and D. Masson, who were referees of this paper, for their constructive criticisms that helped significantly to improve this work. Thanks also to A.S. Andrade, J. F. Santos, H. A. Mendes (UA) and P. Terrinha (UL), B. Barbosa (IGM), P. Cunha, J. Dinis, and H. Henriques (UC), and S. Srivastava (GSCA) for helpful discussions and comments. Many thanks to E. Martins for her help with the revision of this paper. Many thanks to V. Hervalejo, G. Marques, M. Pessoa (UA), John Taylor (OU) and Rose Edwards (IODSL) for their work on some of the figures.

REFERENCES

Alvarado, M.M., 1983. Evolucion de Ia Placa Ibérica. *In Geologia de España—Libro Jubilar J.M. Rios-Tomo 1*. Inst. Geol. Min. Esp., 21-55.

Antunes, M.T., 1979. Ensaio de síntese crítica acerca do Cretácico terminal e do Paleogénico de Portugal. *Ciênc. Terra,* 5:145-174.

Antunes, M.T., Calvo, J.P, Hoyos, M., Morales, J., Ordonez, S., Pais, J., and Sese, C., 1987. Ensayo de correlacion entre el Neogeno de Madrid y Lisboa. *Comm. Serv. Geol. Portugal,* 73:85-102.

Arthaud, F., and Matte, P., 1975. Les décrochements tardihercyniens du sud-ouest de l' Europe. Géometrie et essay de réconstituition des conditions de la deformation. *Tectonophysics,* 25:139-171.

Austin, J.A., Jr., Tucholke, B.E., and Uchupi, E., 1989. Upper Triassic-Lower Jurassic salt basin southeast of the Grand Banks. *Earth Planet. Sci. Lett.,* 92:357-370.

Auzende, J.-M., Olivet, J.-L., Charvet, J., Le Lann, A., Le Pichon, X., Monteiro, J.H., Nicolas, A., and Ribeiro, A., 1978. Sampling and observation of oceanic mantle and crust on Gorringe Bank. *Nature,* 273:45-48.

Bergeron, A., and Bonnin, J., 1991. The deep structure of Gorringe Bank (NE Atlantic) and its surrounding area. *Geophys. J. Int.,* 105:491-502.

Berthou, P.Y., 1973. *Le Cénomanien de l' Estramadure Portugaise.* Mem. Serv. Geol. Portugal, 23.

Berthou, P.Y., and Lauverjat, J., 1979. Essai de synthèse paléogéographique et paléobiostratigraphique du bassin occidental Portugais au cours du Crétacé supérieur. *Cienc. Terra,* 5:121-144.

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

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

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

Boillot, G., Berthou, P.Y., Dupeuble, P.A., and Muselec, P., 1972. Géologie du plateau continental Portugais au nord du Cap Carvoeiro. La série stratigraphique. *C. R. Acad. Sci. Ser. 2,* 274:2748-2751.

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

Boillot, G., Comas, M.C., Girardeau, J., Kornprobst, J., Loreau, J.P., Malod, J., Mougenot, D., and Moullade, M., 1986. Fonds sous-marins basaltiques et ultramafiques au pied d'une marge stable. Résultats préliminaires de la campagne Galinaute (plongées du submersible Nautile à l'Ouest de 1'Espagne). *C. R. Acad. Sci. Ser.* 2, 303:1719-1724.

Boillot, G., Dupeuble, P.A., Hannequin-Marchand, I., Lamboy, M., Lepretre, J.P., and Musellec, P., 1974. Le rôle des décrochements "tardi-hercyniens" dans 1'évolution structurale de la marge continentale et dans la localization des grands canyons sous-marins à l'Ouest et au Nord de la Péninsule Ibérique. *Rev. Geogr. Phys. Geol. Dyn.,* 16:75—86.

Boillot, G., Dupeuble, P.A., and Musellec, P., 1975. Carte géologique du plateau continental nord-portugais. *Bull. Soc. Geol. Fr.,* 17:462-480.

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

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

Boillot, G., and the Leg 103 Shipboard Scientific Party, 1985. Evolution of a passive margin. *Nature,* 317:115—116.

Boillot, G., Mougenot, D., Girardeau, J., and Winterer, E.L., 1989b. Rifting processes of the West Galicia margin, Spain. *In* Tankard, A.J., and Balkwill, H.R. (Ed.), *Extensional Tectonics and Stratigraphy of the North Atlantic Margins.* AAPG Mem., 46:363-377.

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

Boillot, G., and Winterer, E.L., 1988. Drilling on the Galicia Margin: retrospect and prospect. *In* Boillot, G., Winterer, E.L., et al., *Proc, ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 809-828.

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

Brum Ferreira, D., 1981. Carte géomorphologique du Portugal. Deux cartes en couleur au 1/500.000 et notice explicative. *Mem. Cent. Estud. Geogr.,* 6:55.

Cabral, J., 1983. Factores sismotectónicos para selecção de sítios para centrais nucleares (Seismotectonic factors in the selection of sites for nuclear power stations). *Geonovas,* 1:37-45. [in Portuguese]

————, 1989. An example of intraplate neotectonic activity, Vilariça Basin, northeast Portugal. *Tectonics,* 8:285-303.

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

Carvalho, A.M., Ribeiro, A., and Cabral, J., 1983. Evolução paleogeográfica da Bacia Cenozoica do Tejo-Sado. *Bol. Soc. Geol. Portugal,* 24:209-212.

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

Cunha, P.M. Proença, 1992. Estratigrafia e sedimentologia dos depósitos do Cretácico Superior e Terciário de Portugal Central, a leste de Coimbra [Ph.D. thesis]. Univ. Coimbra.

Cunha, P.M. Proença, Pena dos Reis, R., and Dinis, J., 1992. A importância de um silcreto bacinal como marcador do final da etapa sedimentar Aptiano superior-Campaniano inferior na bacia Lusitânica: perspectivas de generalização deste modelo. *Act. III Congr. Geol. Esp. VII Congr. Latinoam. Geol.,* 1:102-106.

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

Dias, R.M.S., 1986. Estudo de um sector do autóctone de Trás-os-Montes oriental a ENE de Torre de Moncorvo [Ms.C. thesis]. Univ. Lisboa.

Dinis, J.M.L., Pena dos Reis, R., 1989. Litostratigrafia e modelos deposicionais nos "grés belasianos" (Cretácico inferior e médio) a leste de Leiria (Portugal Central). *Geociencias,* 4:75-96.

Féraud, G., Beslier, M.-O., and Boillot, G., 1988. Datation ^{40}Ar-^{39}Ar de Ia mise en place des peridotites bordant la marge de la Galice (Espagne). *C. R. Acad. Sci. Ser.* 2, 307:49-55.

Féraud, G., Bonnin, J., Olivet, J.L., Auzende, J.M., and Bougault, H., 1977. Sur quelques datations du volcanisme alcalin de la ligne Açores-Gibraltar et leur contexte géodynamique. C. *R. Acad. Sci. Ser.* 2, 285:1203-1206.

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

Ferreira, M.P, and Macedo, C.A.R., 1979. Actividade magmática durante o Mesozóico. I—Achega para a datação K-Ar das rochas filoneanas básicas intrusivas na zona Centro Ibérica (Portugal). *Mem. Not. Mus. Lab. Min. Geo. Univ Coimbra,* 87:29-49. [in Portuguese]

————, 1987. Idade radiométrica K-Ar do filão das Gaeiras (Óbidos).
Proc. VII Semana de Geoq. Resumos de Comunic., U. Aveiro, [in Portuguese]

Fukao, Y., 1973. Thrust faulting at a lithospheric plate boundary, the Portugal earthquake of 1969. *Earth Planet. Sci. Lett.,* 18:205-216.

Galdeano, A., Moreau, M.G., Pozzi, J.P., Berthou, P.Y., and Malod, J.A., 1989. New paleomagnetic results from Cretaceous sediments near Lisboa (Portugal) and implications for the rotation of Iberia. *Earth Planet. Sci. Lett.,* 92:95-106.

Gardner, J.V., and Kidd, R.B., 1987. Sedimentary processes on the northwestern Iberian continental margin viewed by long-range side-scan sonar and seismic data. *J. Sediment. Petrol.,* 57:397-407.

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

Gradstein, P.M., Agterberg, F.P., Ogg, J.G., Hardenbol, J., van Veen, P., Thierry, J., and Huang, Z., 1994. A Mesozoic time scale. *J. Geophys. Res.,* 99:24051-24074.

Gräfe, K.-U., and Wiedmann, J., 1993. Sequence stratigraphy in the Upper Cretaceous of the Basco-Cantabrian Basin (northern Spain). *Geol. Rundsch.,* 82:327-361.

Grant, A.C., Jansa, L.F., McAlpine, K.D., and Edwards, A., 1988. Mesozoic-Cenozoic geology of the eastern margin of the Grand Banks and its relation to the Galicia Bank. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 787-808.

Grimaud, S., Boillot, G., Collette, B.J., Mauffret, A., Miles, P.R., and Roberts, D.B., 1982. Western extension of the Iberian-European boundary during the early Cenozoic (Pyrenean) convergence: a new model. *Mar. Geol.,* 45:63-67.

Grimison, N.L., and Chen, W.-P., 1986. The Azores-Gibraltar plate boundary: focal mechanisms, depths of earthquakes, and their tectonic implications. *J. Geophys. Res.,* 91:2029-2047.

————, 1988. Source mechanisms of four recent earthquakes along the Azores-Gibraltar plate boundary. *Geophys. J.,* 92:391-401.

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

Hiscott, R.N., Wilson, R.C.L., Gradstein, P.M., Pujalte, V., García-Mondéjar, J., Boudreau, R.R., and Wishart, H.A., 1990. Comparative stratigraphy and subsidence history of Mesozoic rift basins of North Atlantic. *AAPG Bull.,* 74:60-76.

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

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

ILIHA DSS Group, 1993. A deep seismic sounding investigation of lithospheric heterogeneity and anisotropy beneath the Iberian Peninsula. *Tectonophysics,* 221:35—51.

Jansa, L.F., Comas, M.C., Sarti, M., and Haggerty, J.A., 1988. Late Jurassic carbonate platform of the Galicia Margin. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 171-192.

Julivert, M., Fontbote, J.M., Ribeiro, A., and Conde, L., 1980. Mapa tectónico de la Península Ibérica y Baleares (Memória explicativa) (scala: 1:1,000,000). *Publ. Inst. Geol. Min. Esp.* [in Spanish]

Keen, C.E., and de Voogd, B., 1988. The continent-ocean boundary at the rifted margin off eastern Canada: new results from deep seismic reflection studies. *Tectonics,* 7:107-124.

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

Kullberg, M.C., 1985. Controlo estrutural na instalação do Maciço de Sintra. *Bol. Soc. Geol. Portugal,* 24:219-223.

Lallemand, S., Mazé, J.P., Monti, S., and Sibuet, J.C., 1985. Présentation d'une carte bathymétrique de l'Atlantique Nord-Est. *C. R. Acad. Sci. Ser. 2,* 300:145-149.

Lallemand, S., and Sibuet, J.-C., 1986. Tectonic implications of canyon directions over the northeast Atlantic continental margins. *Tectonics,* 5:1125-1143.

Laughton, A.S., Roberts, D.G., and Graves, R., 1975. Bathymetry of the northeast Atlantic: Mid-Atlantic Ridge to southwest Europe. *Deep Sea Res. Oceanogr. Abstr.,* 22:791-810.

Leinfelder, R.R., and Wilson, R.C.L., 1989. Seismic and sedimentologic features of Oxfordian-Kimmeridgian syn-rift sediments on the eastern margin of the Lusitanian Basin. *Geol. Rundsch.,* 78:81-104.

————, in press. Third order sequences in an Upper Jurassic rift-related second order sequence, central Lusitanian basin, Portugal. *In* Graciansky, P.-C, Hardenbol, J., Jaquin, T., Vail, P.R., and Farley, M.B. (Eds.), *Mesozoic and Cenozoic Sequence Stratigraphy of European Basins*. Spec. Publ.—Soc. Econ. Paleontol. Mineral.

Madeira, J., and Ribeiro, A., 1990. Geodynamic models for the Azores triple junction: a contribution from tectonics. *Tectonophysics,* 184:405-415.

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

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

Martins, L.S., 1991. Actividade ígnea Mesozoica em Portugal (Contribuição Petrológica e Geoquímica). (Mesozoic Igneous Activity in Portugal: Contribution from Petrological and Geochemical studies) [*Ph.D. thesis*]. Univ. Lisboa, Portugal, [in Portuguese]

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

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

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

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

————, 1989b. Results from multichannel reflection profiling of the Tagus Abyssal Plain (Portugal)—comparison with the Canadian margin. *In* Tankard, A.J., and Balkwill, H.R. (Eds.), *Extensional Tectonics and Stratigraphy of the North Atlantic Margins.* AAPG Mem., 46:379-393.

May, P.R., 1971. Pattern of Triassic-Jurassic diabase dikes around the North Atlantic in the context of the predrift position of the continents. *Geol. Soc. Am. Bull.,* 82:1285-1292.

McKenzie, D.P., 1972. Active tectonics of the Mediterranean region. *Geophys. J. R. Astron. Soc.,* 30:109-185.

Mendes Victor, L.A., Miranda, J.M., and Matias, L.M., in press. Crustal structure of Western Iberia from geophysical studies. *ILIHA,* Madrid.

Meyer, R., and Pena dos Reis, R., 1985. Paleosols and alunite silcretes in continental Cenozoic Western Portugal. *J. Sediment. Petrol,* 55:76-85.

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

Montenat, C., Guery, F., Jamet, M., and Berthou, Y.B., 1988. Mesozoic evolution of the Lusitanian Basin: comparison with the adjacent margin. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 757-775.

Moreira, V.S., 1984. Sismicidade histórica de Portugal Continental. *Rev. Inst. Nac. Meteorol. Geol.* (separata), [in Portuguese]

————, 1985. Seismotectonics of Portugal and its adjacent area in the Atlantic. *Tectonophysics,* 117:85-96.

Mougenot, D., 1988. Géologie de la Marge Portugaise [Thèse de Doctorat d'Etat és Sciences Naturelles]. Univ. Pierre et Marie Curie, Paris VI.

Mougenot, D., Kidd, R.B., Mauffret, A., Regnauld, H., Rothwell, R.G., and Vanney, J.R., 1984. Geological interpretation of combined SeaBeam, Gloria and seismic data from Porto and Vigo Seamounts, Iberian continental margin. *Mar. Geophys. Res.,* 6:329-363.

Moullade, M., Brunet, M.-F., and Boillot, G., 1988. Subsidence and deepening of the Galicia Margin: the paleoenvironmental control. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 733-740.

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

Murlon, B.J., 1986. The tectonic evolution of the Western Limassol Forest Complex, Cyprus [Ph.D. thesis]. Open University, UK.

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

Oliveira, C.S., 1986. A sismicidade histórica e a revisão do Calálogo Sísmico. *Int. Rep.,* L.N.E.C., Lisboa, [in Portuguese]

Olivet, J.L., Bonnin, J., Beuzart, P., and Auzende, J.M., 1984. Cinématique de 1'Atlantique nord et central. *Publ. Cent. Natl Exploit. Oceans, Rapp. Sci. Tech.,* 54:108-112.

Pautot, G., Auzende, J.M., and Le Pichon, X., 1970. Continuous deep salt layer along North Atlantic margins related to early phase of rifting. *Nature,* 227:351-354.

Peirce, C., and Barton, P.J., 1991. Crustal structure of the Madeira-Tore Rise, eastern North Atlantic—results from a DOBS wide-angle and normal incidence seismic experiment in lhe Josephine Seamount region. *Geophys. J. Int.,* 106:357-378.

Pena dos Reis, R., 1983. A sedimenlologia de depósitos continentais. Dois exemplos do Cretácico Superior-Miocénico de Portugal [Ph.D. thesis]. Univ. of Coimbra, Portugal.

————, 1993. Evidences of probably synsedimentary faulting on the upper boundary surface of lhe Lousões Fine Sandstones (Upper Cretaceous) (Nazaré Lighthouse outcrop, Lusitanian Basin, Portugal). *Abstr. 1st General Meeting IGCP Proj.,* 362:60-61.

Pena dos Reis, R., Corrochano, A., Bernardes, C.A., Cunha, P. Proença, and Dinis, J.M., 1992. O Meso-Cenozóico da margem Atlântica Portuguesa. *III Congr. Geol. Esp. Excurs.,* 115—138.

Pena dos Reis, R., and Cunha, P.M. Proença, 1989a. Comparación de los rellenos terciarios en dos regiones del borde occidenlal del Macizo Hespérico (Portugal Central). *In* Dabrio, C.J. (Ed.), *Studia Geol. Salmanticensia, Paleogeografia de Ia Meseta norte durante el Terciario.* Univ. Salamanca, 5:253-272.

————, 1989b. A definição litostratigráfica do grupo do Buçaco na região da Lousã, Arganil e Mortágua (Portugal). *Com. Serv. Geol. Portugal,* 75:99-109.

Pinheiro, L.M., 1994. The crustal structure under the Tagus Abyssal Plain and the ocean-continent transition off western Iberia [Ph.D. thesis]. Univ. London.

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

Postaire, B., 1983. Sistématique Pb et U-Pb sur zircons. Applications aux roches de haut grade métamorphique impliquées dans la chaine Hercynienne (Europe de 1'Ouest) et aux granulites de Laponie (Finlande). *Bull. Soc. Geol. Mineral. Bretagne,* C15:29-72.

Prosser, S., 1993. Rift-related linked depositional systems and their seismic expression. *In* Williams, G.D., and Dobb, A. (Eds.), *Tectonics and Seismic Stratigraphy.* Geol. Soc. Spec. Publ. London, 71:35-66.

Purdy, G.M., 1975. The eastern end of the Azores-Gibraltar plate boundary. *Geophys. J. R. Astron. Soc.,* 43:973-1000.

Quesada, C., 1990. Precambrian terranes in the Iberian Variscan Foldbelt. *In* Strachan, R.A., and Taylor, G.K. (Eds.), *Avalonian and Cadomian Geology of the North Atlantic:* London (Blackie and Son), 109-133.

————, 1991. Geological constraints on the Paleozoic tectonic evolution of tectonostratigraphic terranes in the Iberian Massif. *Tectonophysics,* 185:225-245.

Quesada, C., Fonseca, P.E., Munhã, J., Oliveira, J.T., and Ribeiro, A., 1994. The Beja-Acebuches Ophiolite (Southern Iberia Variscan fold belt): geological characterization and geodynamic significance. *Bol. Geol. Miner.,* 105:3-49.

Reinen, L.A., Weeks, J.D., and Tullis, T.E., 1991. The frictional behaviour of serpentinite: crustal faults. *Geophys. Res. Lett.,* 18:1921-1924.

Reston, T.I., Krawczyk, C.M., and Wilkens, D., 1994. Detachment faulting and continental breakup: the S reflector offshore Galicia. *Ann. Geophys.,* 1 (Suppl. 1):C43.

Ribeiro, A., 1988. A tectónica Alpina em Portugal. *Geonovas,* 10:9-11. [in Portuguese]

Ribeiro, A., Antunes, M.T., Ferreira, M.P., Rocha, R.B., Soares, A.F., Zbyszewski, G., Moitinho de Almeida, F., Carvalho, D., and Monteiro, J.H., 1979. Introduction à la Géologie generale du Portugal. *Ed. Serv. Geol. Portugal,* 114.

Ribeiro, A., Possolo, A., Brandão Silva, J., Kullberg, M.C., Cabral, J., and Dias, R., 1985. Modelos tectónicos: ensaio de aplicação a Portugal. *Bol. Soc. Geol. Portugal,* 24:145-151.

Ribeiro, A., Quesada, C., and Dallmeyer, R.D., 1987. Tectonostratigraphic terranes and the geodynamic evolution of the Iberian Variscan fold belt. *Conf. on Deformation and Plate Tectonics,* Gijon, Spain, Abstract Vol., 60-61.

————, 1990. Geodynamic evolution of the Iberian Massif. *In* Dallmeyer, R.D., and Martinez Garcia, E. (Eds.). *Pre-Mesozoic Geology of the Ibe- rian Peninsula:* Berlin (Springer-Verlag), 399-409.

Ribeiro, A., and Silva, J.B., in press. Portugal. *In* Moores, E., and Fairbridge, R. (Eds.), *Encyclopedia of World Regional Geology.*

Roest, W.R., and Srivastava, S.P., 1991. Kinematics of the plate boundaries between Eurasia, Iberia and Africa in the North Atlantic from the late Cretaceous to present. *Geology,* 19:613-616.

Ryan, W.B.F., Hsü, K.J., el al., 1973. *Init. Repts. DSDP,* 13 (Pts. 1 and 2): Washington (U.S. Govt. Printing Office).

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

Sibuet, J.-C., 1992. New constraints on the formation of the non-volcanic continental Galicia-Flemish Cap conjugate margins. J. *Geol. Soc. London,* 149:829-840.

Sibuet, J.-C., and Collette, B.J., 1991. Triple junctions of Bay of Biscay and North Atlantic: new constraints on the kinematic evolution. *Geology,* 19:522-525.

Sibuet, J.-C., and Le Pichon, X., 1971. Structure gravimétrique du Golfe de Gascogne et le fossé marginal nord-Espagnol [Gravimetric structure of the Gulf of Gascony and the north Spanish marginal trough]. In *Histoire Structurale du Golfe de Gascogne* (Vol. 2): Paris (Ed. Technip), VI.9-1-VI.9-17.

Sibuet, J.-C., Louvel, V., Whitmarsh, R.B., White, R.S., Horsefield, S.J., Sichler, B., Léon, P., and Recq, M., in press. Constraints on rifting processes from refraction and deep-tow magnetic data: the example of the Galicia continental margin (West Iberia). *Proc. 1994 NATO-ARW Workshop Rifted Ocean-Continent Boundaries.*

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

Sibuet, J.-C., and Ryan, W.B.F., 1979. Site 398: evolution of the West Iberian passive continental margin in the framework of the early evolution of the North Atlantic Ocean. *In* Sibuet, J.-C., Ryan, W.B.F., et al., *Init. Repts. DSDP,* 47 (Pt. 2): Washington (U.S. Govt. Printing Office), 761-775.

Soares, A.F., 1980. A "Formação Carbonatada" Cenomano-Turoniana na região do Baixo Mondego. *Com. Serv. Geol. Portugal,* 66:99-109.

Soares, A.F., Rocha, R.B., Elmi, S., Henriques, M.H., Mouterde, R., Almeras, Y., Ruget, C., Marques, J., Duarte, L., Carapito, M.C., and Kullberg, J.C., 1993. Le sous-bassin nord-lusitanien (Portugal) du Trias au Jurassique moyen: histoire d'un "rift avorté." *C. R.. Acad. Sci. Ser.* 2, 317:1659-1666.

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

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

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

Tankard, A.J., and Welsink, H.J., 1987. Extensional tectonics and stratigraphy of Hibernia oil field, Grand Banks, Newfoundland. *AAPG Bull,* 71:1210-1232.

————, 1989. Mesozoic extension and styles of basin formation in Atlantic Canada. *In* Tankard, A.J., and Balkwill, H.R., *Extensional Tectonics and Stratigraphy of the North Atlantic Margins.* AAPG Mem., 46:175-195.

Thommeret, M., Boillot, G., and Sibuet, J.-C., 1988. Structural map of the Galicia Margin. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 31-36.

Tucholke, B.E., Austin, J.A., and Uchupi, E., 1989. Crustal structure and rift-drift evolution of the Newfoundland Basin. *In* Tankard, A.J., and Balkwell, H.R. (Eds.), *Extensional Tectonics and Stratigraphy of the North Atlantic Margins.* AAPG Mem., 46:247-263.

Udias, A., 1982. Seismicity and seismotectonic stress field in the Alpine-Mediterranean region. *In* Berckhemer, H., and Hsu, K. (Eds.), *Alpine-Mediterranean Geodynamics.* AGU Geodyn. Ser., 7:75-82.

Udias, A., López Arroyo, A., and Mezcua, J., 1976. Seismotectonic of the Azores-Alboran region. *Tectonophysics,* 31:259-289.

Van der Voo, R., and Zijderveld, J.D.A., 1971. Renewed paleomagnetic study of the Lisbon volcanics and implications for the rotation of the Iberian Peninsula. *J. Geophys. Res.,* 76:3913-3921.

Vanney, J.-R., and Mougenot, D., 1981. La plate-forme continentale du Portugal et les provinces adjacentes: analyse geomorphologique. *Mem. Serv. Geol. Portugal,* 28.

Voggenreiter, W., Hötzl, H., and Jado, A.R., 1988. Red Sea related history of extension and magmatism in the Jizan area (southwest of Saudi Arabia): indication for simple-shear during early Red Sea rifting. *Geol. Rundsch.,* 77:257-274.

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

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

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

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

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

Willis, M.G., 1988. The tectonic history of the Lusitanian Basin of Portugal [Ph.D. thesis]. The Open University, Milton Keynes, UK.

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

Winterer, E.L., Gee, J.S., and Van Waasbergen, R.J., 1988. The source area for Lower Cretaceous clastic sediments of the Galicia Margin: geology and tectonic and erosional history. *In* Boillot, G., Winterer, E.L., et al., *Proc. ODP, Sci. Results,* 103: College Station, TX (Ocean Drilling Program), 697-732.

Date of initial receipt: 8 December 1994
Date of acceptance: **8 September 1995**
Ms 149SR-246

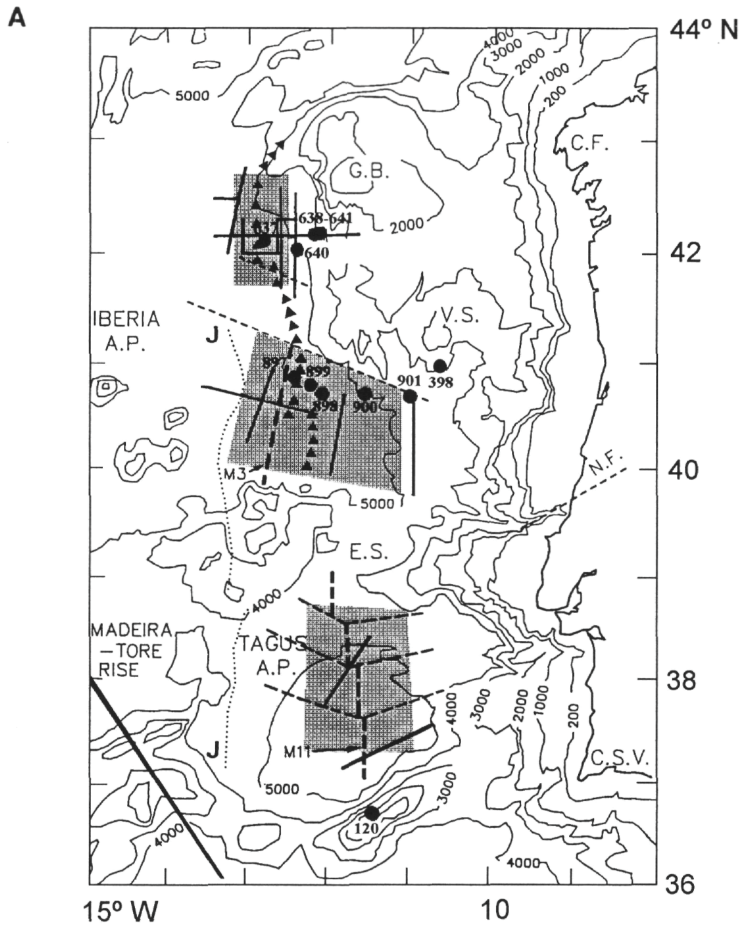

Figure 10. **A.** Bathymetric chart of the west Iberia Margin (after Lallemand et al., 1985; contours in meters) showing the main features in the region of the ocean-continent transition. Stippled area shows distribution of probable serpentinized peridotite at the base of the crust, based on observations along seismic refraction lines (bold lines; after Whitmarsh et al., 1993; unpubl. data). The triangular symbols in the central and northern part of the margin denote the trend of a peridotite ridge, based on sampling and the interpretation of seismic reflection profiles (Beslier et al., 1993). Bold dashed lines denote seafloor spreading magnetic anomalies M3, M1l. Thin dotted line shows the J-anomaly (ca. MO/MI). Bold dashed lines in the Tagus Abyssal Plain: (1) vertical lines denote Chron M1l, which approximately marks the easternmost extent of oceanic crust; (2) other dashed lines show the location and direction of continental faults and fracture zones in the vicinity of the ocean/continent transition, interpreted on the basis of the magnetic, gravity and bathymetric data (after Pinheiro, 1994). Deep Sea Drilling Project and Ocean Drilling Program drill sites are shown by closed circles. GB = Galicia Bank; V.S. = Vigo Seamount; E.S. = Estremadura Spur; C.F. = Cape Finisterra; N.F. = Nazaré Fault; C.S.V. = Cape Saint Vincent. **B.** Three east-west cross-sections across the ocean-continent transition zone off western Iberia, based on seismic reflection profiles and seismic refraction and gravity models. The profiles have been aligned along known basement exposures of a peridotite ridge (PR) and the probable landward edge of the oceanic crust in the Tagus Abyssal Plain, (a) Composite section of profiles GP101, GAP-106, and GAP-014 from Murillas et al. (1990). GP101 has also been enhanced with the crustal structure from Whitmarsh et al. (pers. comm., 1995) based on wide-angle seismic profiling, (b) Section across the Iberia Abyssal Plain, constrained by seismic refraction profiles in the western half and by gravity modeling elsewhere from Whitmarsh et al. (1990; 1993). Basement surface from the Lusigal-12 profile (see Beslier, this volume), (c) Section across the Tagus Abyssal Plain based partly on seismic refraction modeling of Pinheiro et al. (1992) and Peirce and Barton (1991).

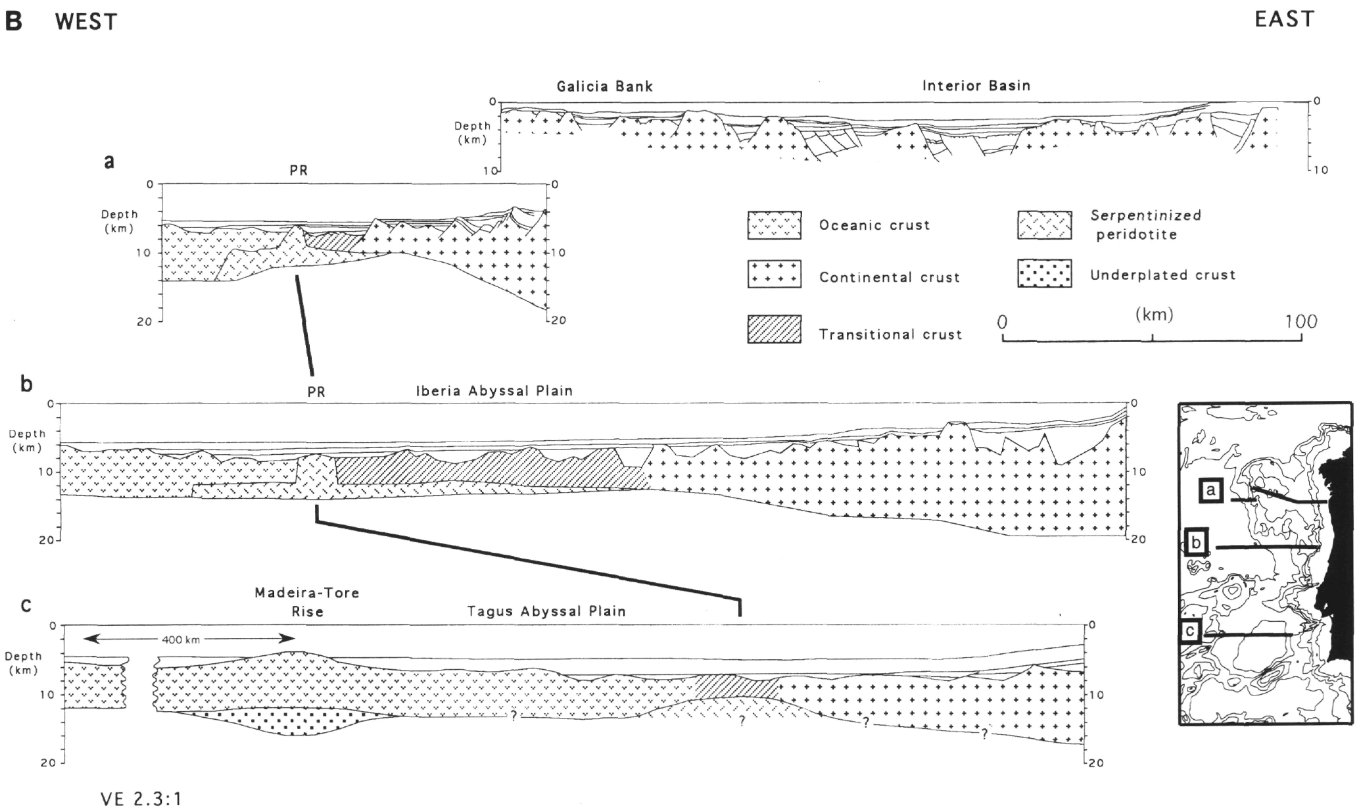

Figure 10 (continued).