

# THE NORTH AND NORTH-WEST SPANISH CONTINENTAL MARGIN: A REVIEW<sup>(1)</sup>

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## ABSTRACT

The west Galicia margin (Spain) provides a model for passive continental margin evolution. During the Mesozoic, the continental crust of the margin experienced several episodes of extension. The main stage was during the Berriasian-late Aptian interval. The stretching of the lithosphere resulted in a) the thinning of the continental crust, testified by superficial extensional structures (normal faults, tilted fault blocks) and rapid subsidence, and b) the emplacement of mantle rocks (peridotite) at the continental rift axis. The crustal thinning and the final emplacement of peridotite onto the seafloor are explained by uniform, normal, simple shear of the continental lithosphere, following asthenosphere diapirism. The sedimentary response to the rifting process was the rapid deposition of coarse detrital sediments in the half grabens of the rift. The pre-rift sediments are platform carbonates, and the post-rift sediments are distal turbidites followed by pelagic sediments.

To the south of the Bay of Biscay, the north Galicia and Asturias margin evolved as the Galicia passive margin until latest Cretaceous time. During the Paleocene-Eocene interval, it was converted into an active margin as a result of the Eurasian-Iberian plate convergence and related southward subduction of the European plate beneath the Iberian plate. A marginal trench and a tectonic accretionary prism developed at the plate boundary and the margin was shortened, deformed and eroded, becoming a subduction-related fold belt. Today it is the westward prolongation of the Pyrenean and Cantabrian collisional chain. Therefore models accounting for the emerged part of the Pyrenees must also be constrained by offshore data.

**Key words:** Rifting, Extension, Passive continental margin, Peridotite, North Atlantic, Pyrenees.

## RESUMEN

El margen occidental de Galicia es un ejemplo de un nuevo modelo de la evolución de los márgenes continentales pasivos. Durante el Mesozoico se sucedieron diversas etapas de extensión de la corteza continental de este margen. El episodio expansivo principal acaeció en el intervalo comprendido entre el Berriasiense y el Aptiense superior. El estiramiento de la litosfera dio lugar a: 1) al adelgazamiento de la corteza continental, puesto de manifiesto por estructuras superficiales extensivas (fallas normales, bloques basculados) y 2) al emplazamiento de rocas procedentes del manto (peridotitas) en el eje del *rift* continental. El adelgazamiento de la corteza y el emplazamiento final de las peridotitas en los fondos marinos se explican por una cizalla simple uniforme normal de la corteza litosférica causada por el diapiroismo de la astenosfera. La respuesta de la sedimentación al fenómeno del *rifting* es el depósito rápido de sedimentos detriticos gruesos en las semifosas del *rift*. Los sedimentos anteriores al *rift* están constituidos por turbiditas distales seguidos por sedimentos pelágicos.

El margen septentrional de Galicia y Asturias, localizado al sur del Golfo de Vizcaya, se comportó como el margen pasivo de Galicia hasta el Cretácico terminal. Durante el intervalo Paleoceno-Eoceno se convirtió en un margen activo como resultado de la convergencia de las placas Europea e Ibérica, y la subducción hacia el sur de la Placa Europea bajo la Placa Ibérica. En el límite de placas se desarrolló una fosa marginal y un prisma de acreción tectónica, sufriendo el margen un acortamiento, en relación con la subducción. Actualmente constituye la continuación, hacia el oeste, de la cadena de colisión de los Pirineos y la Cantábrica. Por esta causa los modelos evolutivos de la parte emergida de los Pirineos deben de tener en cuenta los datos de la plataforma submarina.

**Palabras clave:** Extensión, Margen continental pasivo, Peridotitas, Atlántico Norte, Pirineos.

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## 1. INTRODUCTION

The continental margins bounding the Galicia, Asturias and Basque areas are among the best known in the world (fig. 1). Offshore studies began 20 years ago, and since that time, an exceptional amount of geological and geophysical data was collected. Recently, significant new progresses were accomplished by using the drilling vessel JOIDES Resolution (Leg. 103 of the Ocean Drilling Program, West Galicia margin) and the french submersibles Cyana and Nautile (Asturias and Galicia margins). All these studies were designed to elucidate two major geodynamic processes: a) crustal thinning and tectonic denudation of mantle rocks during the rifting stage of the west Galicia passive margin, and b) crustal shortening and thickening of the north Spanish passive margin, when it was turned into an active margin during Paleogene time. This paper summarizes the new data and compares the west Galicia passive margin, the north Galicia and Asturias margins deformed by the subduction of the oceanic lithosphere of the Bay of Biscay, and the Basque and Pyrenean margins now folded as a result of the Cenozoic collision between Eurasian and Iberian plates.

We will address the geodynamic evolution in two ways. First, we consider the rifting processes as exemplified by the west Galicia margin, and, second we concentrate on the north Galician, Asturian and Basque margins, in order to document the effects of subduction and/or collision on the former passive margins. In each section, we focus on structure and structural evolution, with the sedimentary record being considered in the context of the successive tectonic stages. Most of the data, ideas and figures gathered in this paper come from papers collected in the Initial Reports of the DSDP, Leg 103, parts A and B (Boillot, Winterer, Meyer *et al.*, 1987, in press), and from the scientific reports of the french diving campaigns (Malod *et al.*, 1982; Groupe Cybère, 1984; Boillot *et al.*, 1986, 1987b in press a, b, c).

## 2. THE WEST GALICIA MARGIN: A MODEL PASSIVE MARGIN.

The deep Galicia margin, northwest of the Iberian Peninsula (Fig. 1 and 2) is a starved passive margin with only a thin (0-4 km) sedimentary cover above the acoustic basement. Rift structures, apparent on seismic pro-

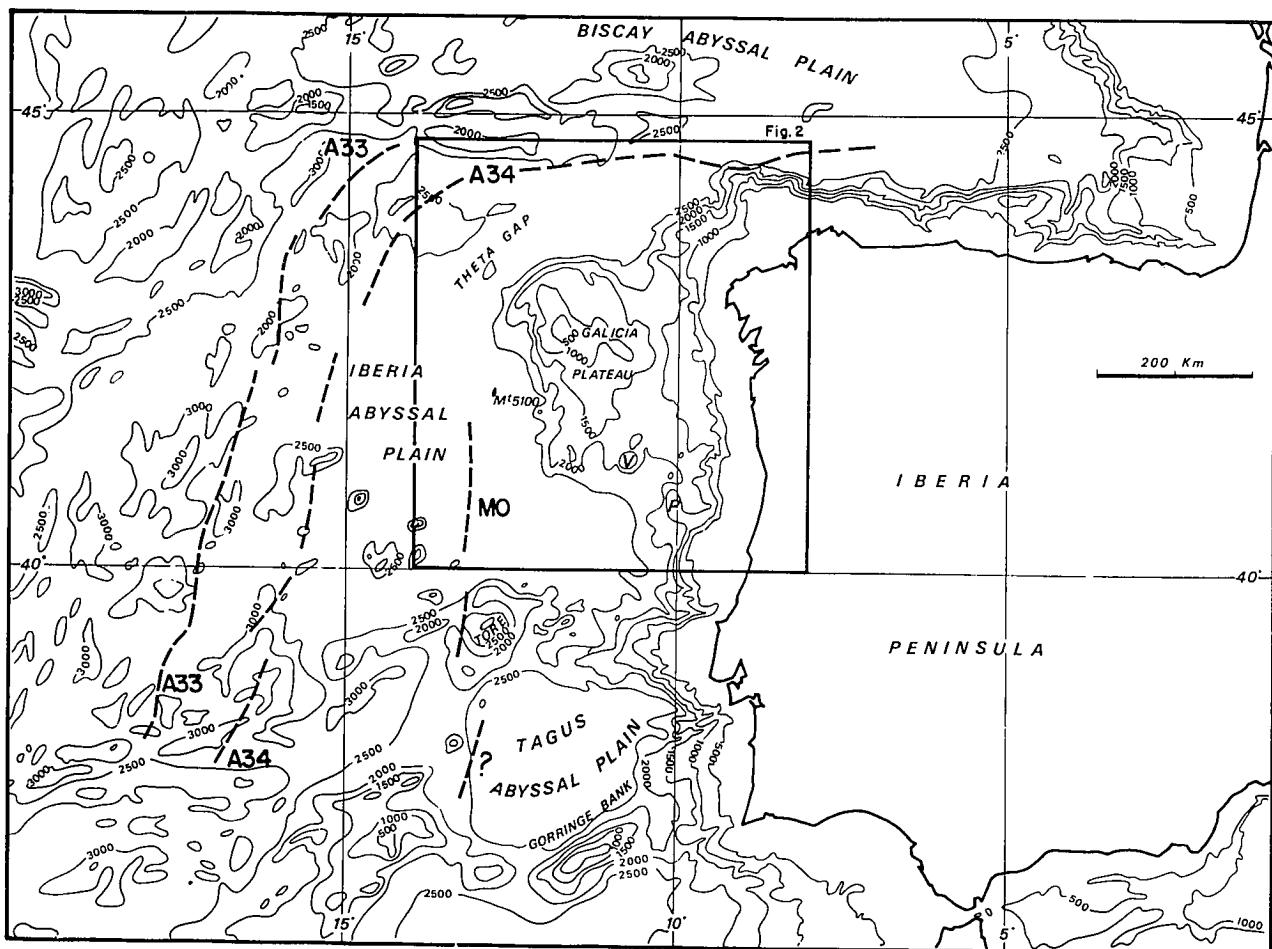


Fig. 1.—The Atlantic margins of the Iberian Peninsula. Magnetic anomalies after Gennoc *et al.*, (1979). Bathymetry in fathoms, after Laughon *et al.*, (1975). The frame is the location of Fig. 2 V: Vigo seamount; P: Porto seamount.  
Fig. 1.—Los márgenes atlánticos de la Península Ibérica. Anomalías magnéticas según Guennoc *et al.*, (1979). Batimetría en brazas, según Laughon *et al.*, (1975). El recuadro corresponde al área representada en la fig. 2. V: Seamount de Vigo. P.: Seamount de Oporto.

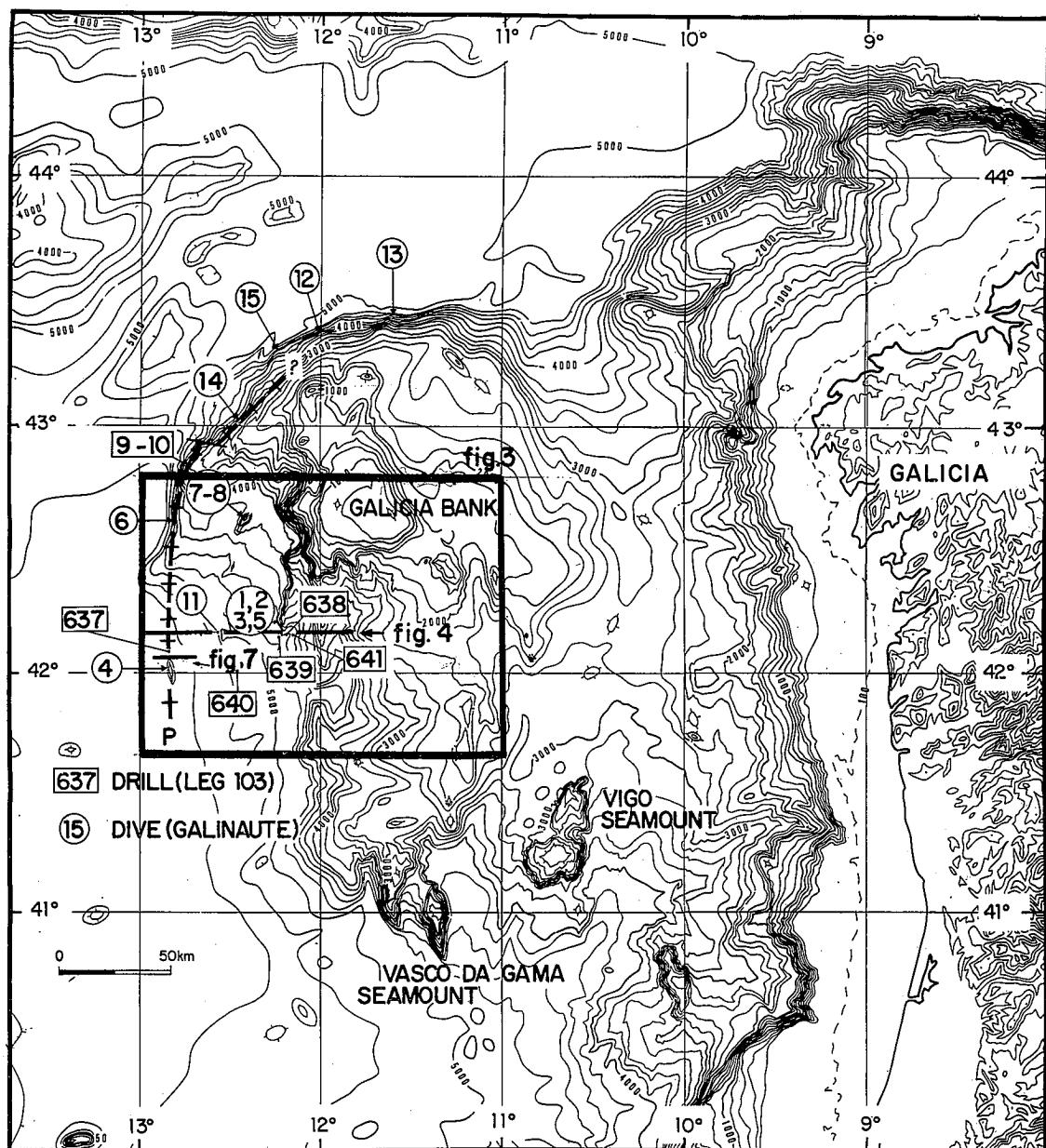


Fig. 2.—Location of drill Sites (Leg. 103) and Galinaute dive Sites on the west Galicia margin. For location see Fig. 1. Bathymetry in meters after Lallemand *et al.*, 1985. The Fig. 3, 4 and 7 are located. P. Peridotite ridge.

Fig. 2.—Localización de los sondeos (Leg. 103) y de los puntos de inmersión del Galinaute en el margen occidental de Galicia. Para la localización ver figura 1. Batimetría en metros según Lallemand *et al.*, (1985). Se sitúan el área incluida en la fig. 3 y los cortes de las figs. 4 y 7. p: cresta de peridotitas.

files, control the present-day morphology (Montadert *et al.*, 1979; Groupe Galice, 1979; Boillot *et al.*, 1979, 1980; Sibuet *et al.*, 1987; Mauffret and Montadert, 1987). The continental basement is broken by normal faults into narrow (10-20 km), elongated (60-100 km) tilted blocks trending north and dipping gently east, forming a series of half graben. Tilted blocks and normal faults are disrupted and/or offset by a discrete pattern of transverse faults, orientated N55-70° and N115-135° (Boillot, 1986; Thommeret *et al.*, in press; Fig. 3 and 4).

The Galicia margin is morphologically complex and anomalous. The province of Galicia is bounded to the west by a narrow (30 km) shelf adjacent to a wide (100 km) and deep (3-4 km) offshore basin (fig. 2). Part-

her seawards, several plateaus form a discontinuous barrier between this basin and the Iberian plain, including from north to south Galicia, Vigo, Vasco de Gama and Porto seamounts (Berthois *et al.*, 1965; Laughton *et al.*, 1975; Vanney *et al.*, 1979; Roberts and Kidd, 1984; Lallemand *et al.*, 1985). These plateaus are Mesozoic tilted blocks that were reactivated and uplifted by early Cenozoic Pyrenean tectonics, except Vigo seamount, which is an uplifted basin (Boillot *et al.*, 1979; Mougenot *et al.*, 1984, 1986). In contrast, the south-western Galicia margin was relatively unaffected by Cenozoic deformation. The sites chosen for drilling during ODP Leg 103 are located on this part of the margin, which is close to the ocean-continent crustal boundary. In ad-

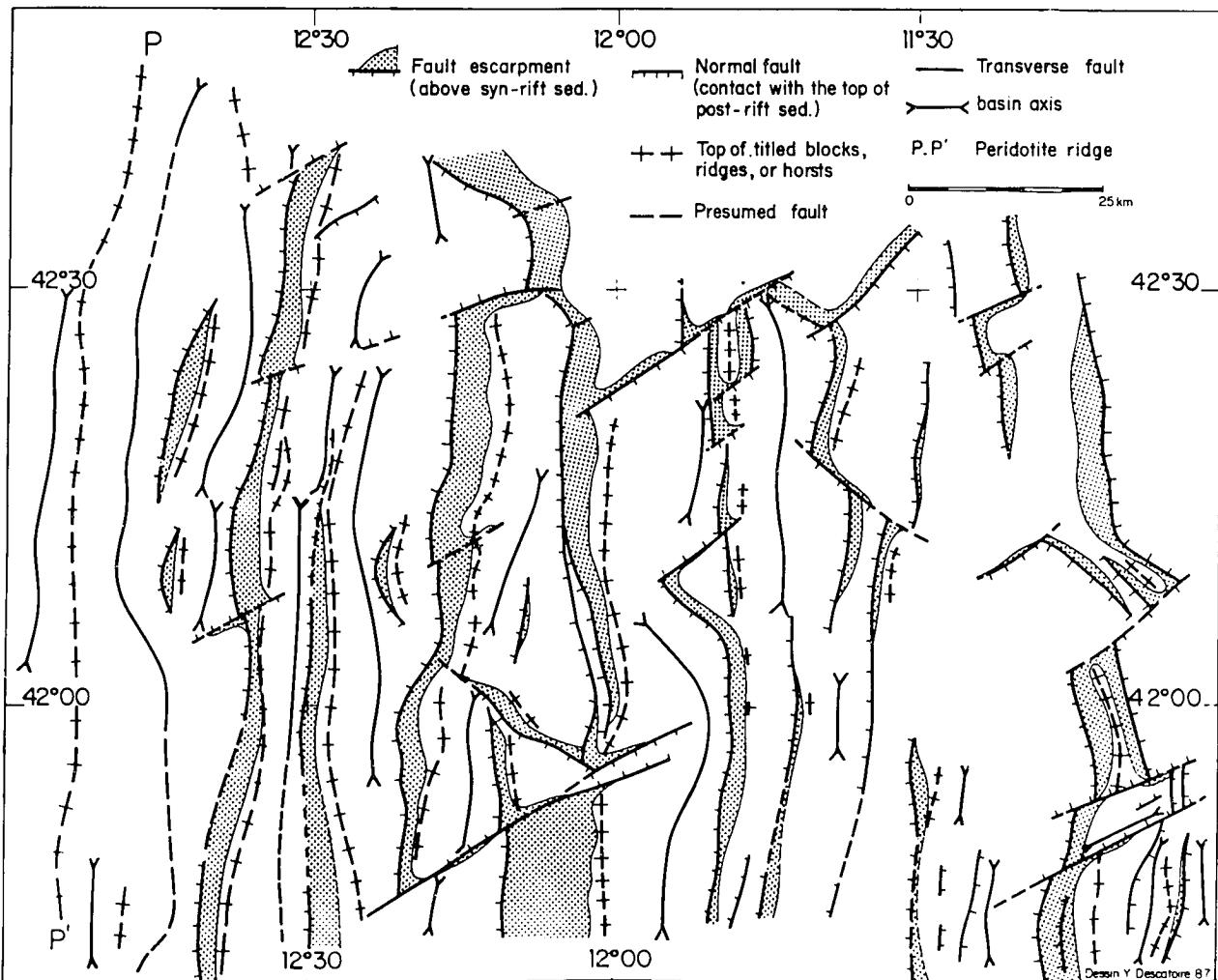


Fig. 3.—Structural map of the western part of the Galicia margin, after Thommeret *et al.*, (in press). For location, see fig. 2. p-p. peridotite ridge.  
Fig. 3.—Esquema estructural de la parte occidental del margen de Galicia, según Thommeret *et al.*, (en prensa). Localización en fig. 2. p-p': cresta de peridotitas.

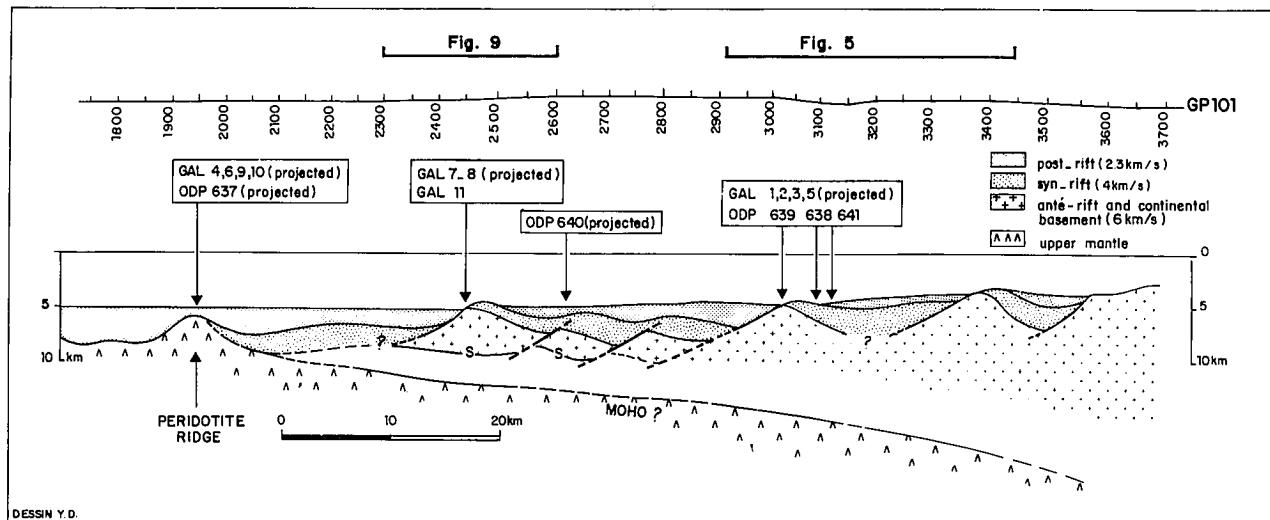


Fig. 4.—Synthetic cross section of the west Galicia passive margin, after Boillot *et al.*, (1986). The Moho is located assuming isostatic equilibrium, densities of 2.8 and 3.3 for the thinned continental crust and upper mantle, respectively. S = Seismic reflector. The drill and dive sites and the seismic line of Fig. 5 and 9 are located. For location, see fig. 2.  
Fig. 4.—Corte sintético del margen pasivo del oeste del Galicia, según Boillot *et al.*, (1986). Posición de la Moho suponiendo un equilibrio isostático, con densidades de 2,8 y 3,3 para la corteza continental adelgazada y el manto superior, respectivamente. S = reflector sísmico. Se sitúan los sondeos y las inmersiones, así como los perfiles de las fig. 5 y 9.

dition to the drilling data, we will also report the results of several dives carried out with the submersible Nautile during the Galinaute cruise (fig. 2), in a region located northward along the western slope of the Galicia Bank.

### 2.1. Timing of events: rifting and drifting.

The fundamental processes associated with the evolution of a passive margin, such as crustal thinning, faulting and subsidence, cannot be clearly understood without a knowledge of the timing of events. Obtaining such knowledge was the broad objective of drilling at Sites 638, 639 and 641 during Leg 103 of ODP (fig. 5). A summary compilation of the stratigraphic results of drilling at these three sites is shown in fig. 6. The sedimentary sequence is described in stratigraphic order from the base up.

1. A layer of conglomerate or breccia that may lie between the Tithonian unit (see following text) and basement rock. This layer contains a great variety of low-grade metamorphic rocks of sedimentary origin, and silicic volcanic or hypabyssal rocks of rhyolitic or rhyodacitic composition. The pebbles of the conglomerate originated from a thick underlying sedimentary and volcaniclastic sequence (800 m), sampled by the Nautile (Boillot *et al.*, 1986, in press a). As yet we have no age determination for this sequence, which could be Paleozoic, Triassic or Jurassic (pre-Tithonian) in age. Whatever the age or the sediments, they indicate that tectonic extension and associated volcanism occurred before the main stage of rifting on the Galicia margin.

2. A layer (400 m?) of Upper Jurassic (Tithonian)

and possibly lowest Cretaceous limestone and dolomite, with lesser amounts of sandstones and claystones, deposited in relatively shallow water before the main stage of rifting. This layer seems to be ubiquitous on the west Galicia margin, as it was sampled by dredging in many other places (Dupeuble *et al.*, 1987).

3. About 40 m of lower Valanginian calpionellid-bearing marlstone, probably deposited at moderate depth not long after the onset of rifting and rapid subsidence.

4. 400 m of upper Valanginian and Hauterivian marlstone interbedded with turbiditic sandstones and claystones rich in terrestrial plant debris, and Barremian and Aptian alternating clays, calcareous clays, marl/marlstones and clayey limestones, including thin turbidites and debris flows, deposited in deep water during rifting.

5. Up to 1500 m of sediment deposited after rifting ceased and oceanic spreading began between Iberia and Newfoundland. These sediments include Albian-Cenomanian black shale (thin turbidites with plant debris) and upper Cretaceous-Cenozoic calcareous oozes and thin turbidites (Leg 47b of IPOD; Sibuet, Ryan *et al.*, 1979).

This sedimentary record indicates that tectonic extension and deepening of the margin occurred in several stages. Several hundred meters of subsidence accommodated the pre-Tithonian sedimentary and volcanic sequence, and the Tithonian shallow-water carbonate beds. Then, the carbonate platform was faulted and drowned sometime during the latest Tithonian to earliest Valanginian interval. Subsequently, the platform was partly buried by syn-rift clastic sediments, which

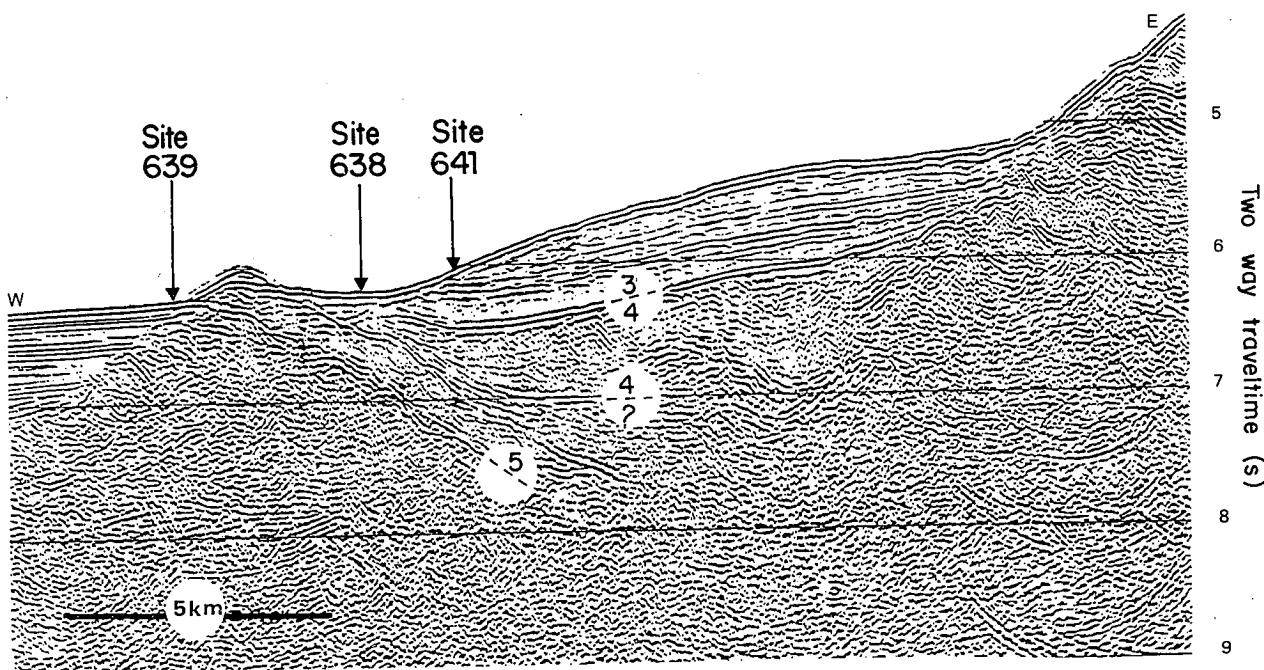


Fig. 5.—Section of the multichannel seismic profile GP-101, showing the location of drill Site 638, 639 and 641. 3 = post-rift strata; 4 and 5 syn-rift strata. After Boillot, Winterer, Meyer *et al.*, 1987. For location, see fig. 4.  
Fig. 5.—Perfil sísmico de multicanal GP-101, en el que se muestra la localización de los sondeos 638, 639 y 641. 3: sedimentos *postrift*; 4 y 5 sedimentos *sin rift*. Según Boillot, Winterer, Meyer *et al.*, (1987). Localización en la fig. 4.

progressively filled half grabens formed by tilting of fault blocks along normal faults. Seismic and drilling data clearly reveal that faulting and associated subsidence continued episodically for about 25 m.y. from early Valanginian through late Aptian times.

The regional post-rift (break-up) unconformity, dating the onset of seafloor spreading (drifting), is late Aptian in age (Sibuet, Ryan *et al.*, 1979; see also the Initial Report of Leg 103, part A, Site 641). Magnetic anomaly Mo occurs within the lower Aptian at Site 641,

about 30 m below the base of the overlying post-rift sedimentary sequence (see Site chapter 638, in Boillot, Winterer, Meyer *et al.*, 1987). This helps to explain why the Mo magnetic anomaly cannot be recognized west of the Galicia margin (fig. 1), a place where seafloor spreading began later. Anomaly Mo is identified to the west of the Portuguese margin, suggesting that seafloor spreading propagated northward along the west Iberian rift when Newfoundland separated from Iberia during early Cretaceous time.

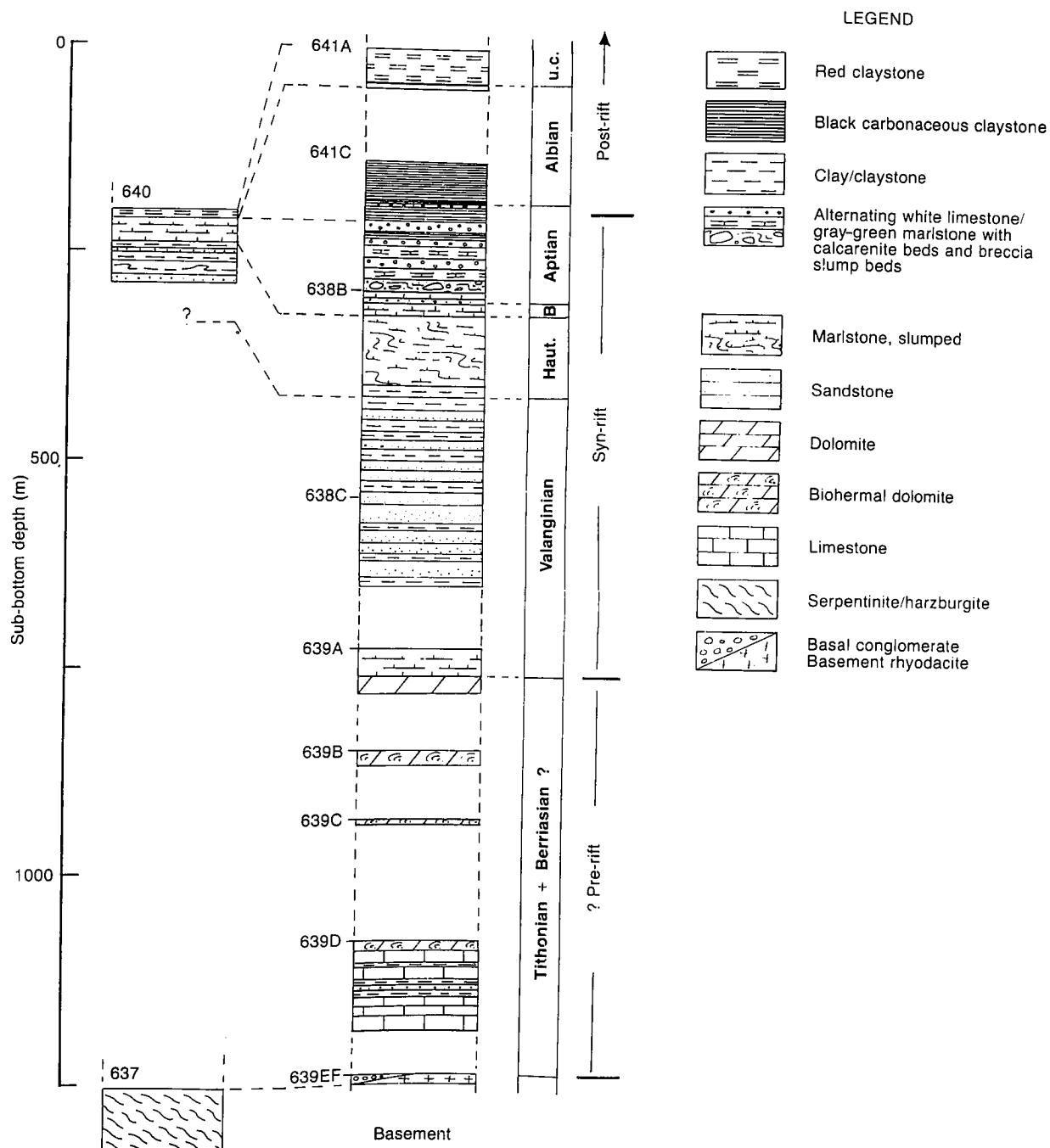


Fig. 6.—Schematic stratigraphic column of the west Galicia margin, assembled after results of drilling a Sites 638, 639 and 641 (see fig. 5 for location). U.C.: Upper Cretaceous; B: Barremian. After Boillot, Winterer, Meyer *et al.*, 1987.  
Fig. 6.—Columna estratigráfica esquemática del margen occidental de Galicia, elaborada a partir de los datos de los sondeos 638, 639 y 641 del ODP (ver fig. 5 para localización). U.C.: Cretácico superior; B: Barremiense. Segundo Boillot, Winterer, Meyer *et al.* (1987).

## 2.2. The peridotite ridge.

At the boundary between the thinned continental crust of the margin and the oceanic crust of the Atlantic, the basement beneath the sediments forms a ridge (fig. 2 and 7), which can be followed on seismic reflection profiles for about 130 km. The ridge was sampled by means of dredge hauls, drilling at Site 637, and dives of the submersible Nautilus. It consists of serpentized peridotite (Boillot *et al.*, 1980, 1986; Boillot, Winterer, Meyer *et al.*, 1987). In its southern part, the ridge strikes N-S, parallel to the trend of the tilted fault blocks on the margin (fig. 3). North of 42°8' latitude north, in the region uplifted during the Eocene Pyrenean tectonics, the ridge turns east and follows the northwest edge of Galicia bank (fig. 2). The ridge crest is 2 or 3 km shallower than the flanking sedimentary basins, and its width, taken as the distance between the axes of these basins, is about 10 to 12 km (fig. 7). On its eastern side, the ridge is partly buried by early Cretaceous synrift sediments. Therefore, the ridge seems to have been emplaced along the margin prior to the beginning of oceanic accretion in this part of the Atlantic, probably at the end of the rifting stage.

The succession of structural and metamorphic events that affected the serpentized peridotite is shown in Table 1 (Evans and Girardeau, in press; Girardeau *et al.*, in press; Kornprobst and Tabit, in press; Agnirier *et al.*, in press; Kimball and Evans, in press; Evans and Baltuck, in press). As yet, it has not been possible to date these events, but their succession clearly shows the rise of the peridotite from a depth of several tens of kms in the upper mantle up to the earth's surface, where the rock was serpentized and fractured. Therefore, the evolution of the peridotite, ending in the early Cretaceous when the basement ridge bounding the margin was emplaced, seems to be a direct result of the horizontal stretching of the lithosphere during the rifting stage of the margin.

## 2.3. Models.

Structural studies (Girardeau *et al.*, in press) show that the peridotite were deformed in a rotational regime, probably by simple shear. This fact does not fit well with models of homogeneous deformation of the lithosphere, which instead implies pure shear (McKenzie, 1978; Le Pichon and Sibuet, 1981). On the other hand, petrological studies (Evans and Girardeau, in press) show that the peridotite were partly melted (5-10% of the rock) at the beginning of their evolution, yet the products of this melting (gabbro and basalt) are not found.

To account for these facts, the Leg 103 scientific team proposed a new model, derived from that of Wernicke (1981; 1985). In this model, the thinning of continental crust under a passive margin and the exposure of the upper mantle result from normal simple shear along a gently-dipping detachment fault rooted in the mantle. On the Galicia margin, the direction and sense of movements for both the ductile and brittle deformation of the mantle rocks, as determined in cores orientated by paleomagnetism, suggest that the detachment fault dips east, towards the continent (Boillot *et al.*, 1987a). The model also explains the fact that, on the Galicia margin, the amount of crustal thinning as indicated by the probable depth of the Moho is everywhere higher than the amount of horizontal stretching implied from the geometry of normal faults between tilted blocks (Chenet *et al.*, 1982). For example, the thinning coefficient is more than 3 at the drill Site 638, where the crust is about 10 km thick, when the superficial extension coefficient is less than 2 at the same place. This difference cannot be explained by the uniform stretching model, which implies the same amount of thinning and stretching for the whole lithosphere. But the detachment fault model does not require there to be a relation between the amount of superficial stretching and the amount of crustal thinning (fig. 8).

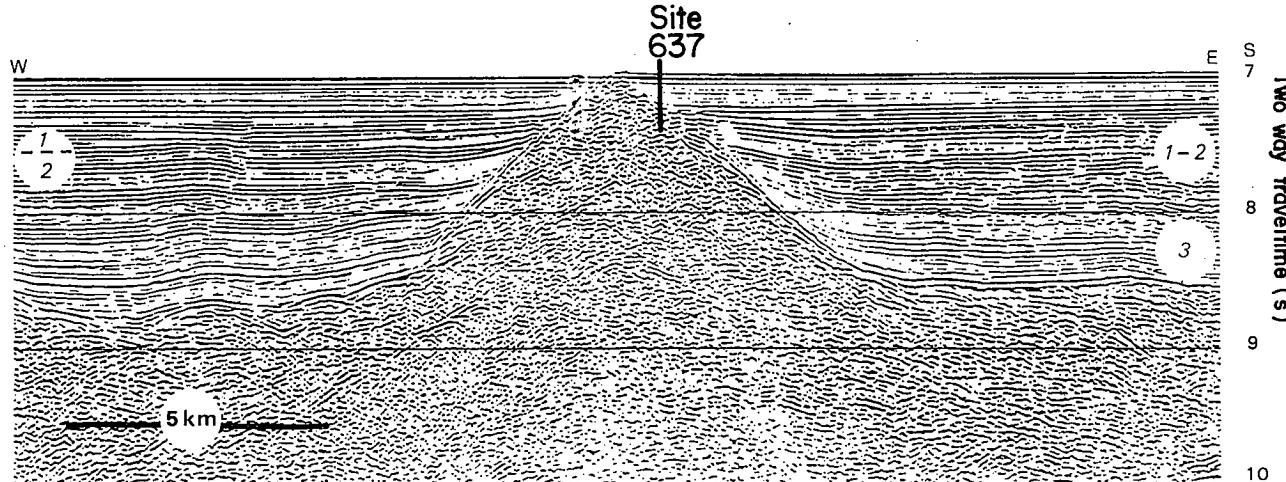


Fig. 7.—Section of the seismic line GP12, showing the peridotite ridge and the location of the drill Site 637 (after Boillot, Winterer, Meyer *et al.*, 1987). For location, see fig. 2.

Fig. 7.—Sección de la linea sismica GP12, mostrando la creta de peridotitas y la localización del sondeo 637 del ODP (Según Boillot, Winterer, Meyer, *et al.* 1987). Localización en la fig. 2.

The detachment fault model was recently elaborated to account for the partial melting of the peridotite at shallow depths in an early stage of its evolution (table 1 and fig. 8; Boillot *et al.*, in press b). The initial stage consists of the rise of an asthenospheric diapir beneath continental crust of normal thickness (Nicolas *et al.*, 1987; Bonatti and Seyler, 1987). The products of partial melting (gabbro) then accumulate along the summit of the diapir. In a second stage, the gabbro and some of the peridotites are stripped off tectonically by movements along a detachment fault, and these rocks then form the basement for sediments between the two conjugate margins. "True" oceanic accretion then begins in the zone where the lithosphere was most thinned during the preceding stages. This model accounts for all the available data along the Galicia margin, assuming a former detachment fault dipping east, beneath the present margin. More precisely, it explains:

1. The discrepancy between the low amount of superficial extension by comparison with the amount of crustal thinning.

2. The thinning of the crust mainly at the expense of its lower part, with the basement of the present margin belonging to the upper continental crust (Capdevila and Mougenot, in press). Note that this implication of the model is in agreement with seismic refraction data recorded on the Galicia mainland, showing the westward thinning of the lower crust beneath the Galicia coast and shelf (Córdoba *et al.*, 1987).

3. The partial melting of the peridotite at high temperature ( $1250^{\circ}$ ) during the first stage of its uplift and its behaviour as asthenosphere, and the mylonitic deformation of the rock at decreasing temperature ( $1000^{\circ}$ - $850^{\circ}$ ) and depth (30-7 km) (table 1).

4. The final tectonic denudation of the upper mantle at the ocean-continent boundary, and its hydrother-

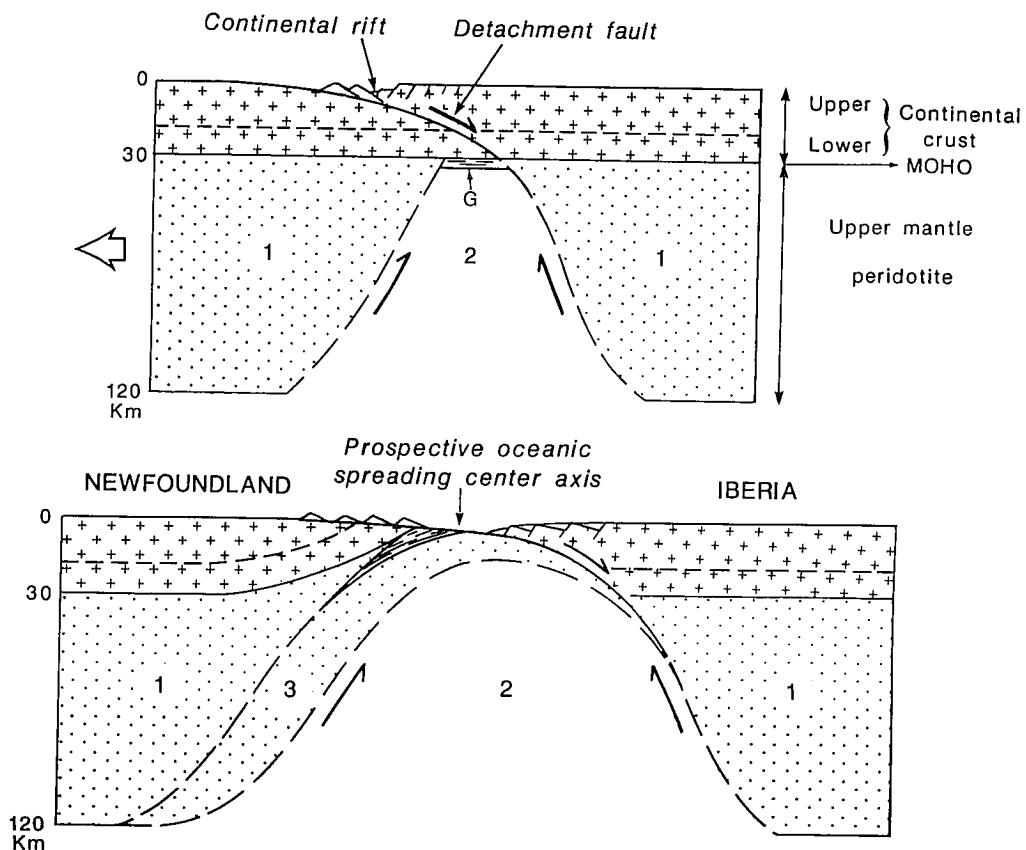


Fig. 8.—A composite model for the thinning of the continental crust and the emplacement of peridotites onto the seafloor along the Galicia margin. A: bulge of asthenosphere, resulting in low partial melting of peridotites and crystallization of the partial melting products into gabbros (G). This stage of evolution is derived from the models of Nicolas *et al.*, (1987) and Bonatti and Seyler (1987). B: Tectonic denudation of gabbro along the Newfoundland margin and of cooled, poorly depleted peridotites along the west Galicia margin. This stage of evolution was described by the model of Boillot *et al.*, (1987a). 1. lithosphere subcrustal peridotite; 2. peridotite experiencing low partial melting, and behaving as an asthenosphere bulge; 3. cooled, poorly depleted peridotite; G. Gabbro. The last stage of the evolution would be the beginning of classic seafloor spreading.

Fig. 8.—Modelo en el que se refleja el adelgazamiento de la corteza continental y el emplazamiento de las peridotitas en el fondo marino en el margen de Galicia. A: Protuberancia de la astenosfera producida por la fusión parcial de las peridotitas y cristalización de los productos de fusión parcial en gabros (G). Esta etapa de la evolución se basa en los modelos de Nicolas *et al.*, (1987) y Bonatti y Seyler (1987). B: Denudación tectónica de los gabros a lo largo del margen de Terranova y enfriamiento de las peridotitas ligeramente empobrecidas a lo largo del margen de Galicia. Esta etapa de evolución ha sido descrita en el modelo de Boillot *et al.* (1987). a) Leyenda: 1. Peridotitas subcorticales litosféricas. 2. Peridotitas con ligera fusión parcial. 3. Peridotitas enfriadas y ligeramente empobrecidas al comienzo de la expansión oceánica clásica.

mal serpentization by reaction with sea-water, the fracturing of the serpentine and finally the emplacement of the peridotite ridge just before the initiation of sea-floor spreading between Galicia and Newfoundland.

5. The existence of a strong seismic reflector or bundle of reflectors labelled S, that appear within or beneath the continental basement in the deepest parts of the Galicia margin (fig. 9; de Charpal *et al.*, 1978; Montadert *et al.*, 1979; Mauffret and Montadert, 1987; Sibuet *et al.*, 1987). This reflector can be interpreted as the seismic signature of the detachment fault, as proposed by Wernicke and Burchfield (1982). According to the model proposed in fig. 8, the detachment fault separates the serpentinite from the upper continental crust, within the basement of the deep margin. These two kinds of rocks have very different densities and so-

nic velocities, and this contrast may explain the geophysical characters of the S reflector.

Taking the West Galicia margin as a model for passive margins, we shall now proceed eastward to the North Spanish margin, where the syn-rift structures were remobilized by Eocene tectonics.

### 3. THE NORTH SPANISH MARGIN

During the Mesozoic, the geological evolution of the North Spanish margin closely resembled that of the West Galicia margin. Rifting of the continental crust occurred in several stages, mainly during the Berriasian-Aptian interval. Later, the sinistral transcurrent motion of Iberia with respect to Eurasia resulted in the ope-

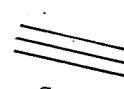
Succession of events	Assumed temperature	Assumed depth	
Event 5 Fracturing, then infilling of fractures by calcite (and serpentine)	$\leq 10^\circ$	0	
Event 4 Ubiquitous serpentization	$110-55^\circ$ ( $< 300^\circ$ )	5 km	
Event 3 Beginning of serpentization ?  Crystallization of amphiboles in static conditions: – hornblende, tremolite – pargasite	350°  750° 900-800°		
Event 2 Stretching and mylonitization at decreasing temperature  New foliation $S_1$	850°  1000°	7 km	 $S_1$
Event 1 End of the partial melting in the plagioclase facies.  Beginning of stretching Beginning of partial melting	970° 1100° 1250°	$\leq 30$ km	
Event 0 Initial peridotite: coarse grained spinel lherzolite; initial foliation $S_0$	?	?	 $S_0$

Table 1: Succession of structural and metamorphic events experienced by the Galicia margin peridotite, according to the studies by Evans and Girardeau, Girardeau *et al.*, Kornprobst and Tabit, Agrinier *et al.*, Kimball and Evans and Evans and Baltuck, all papers in press.

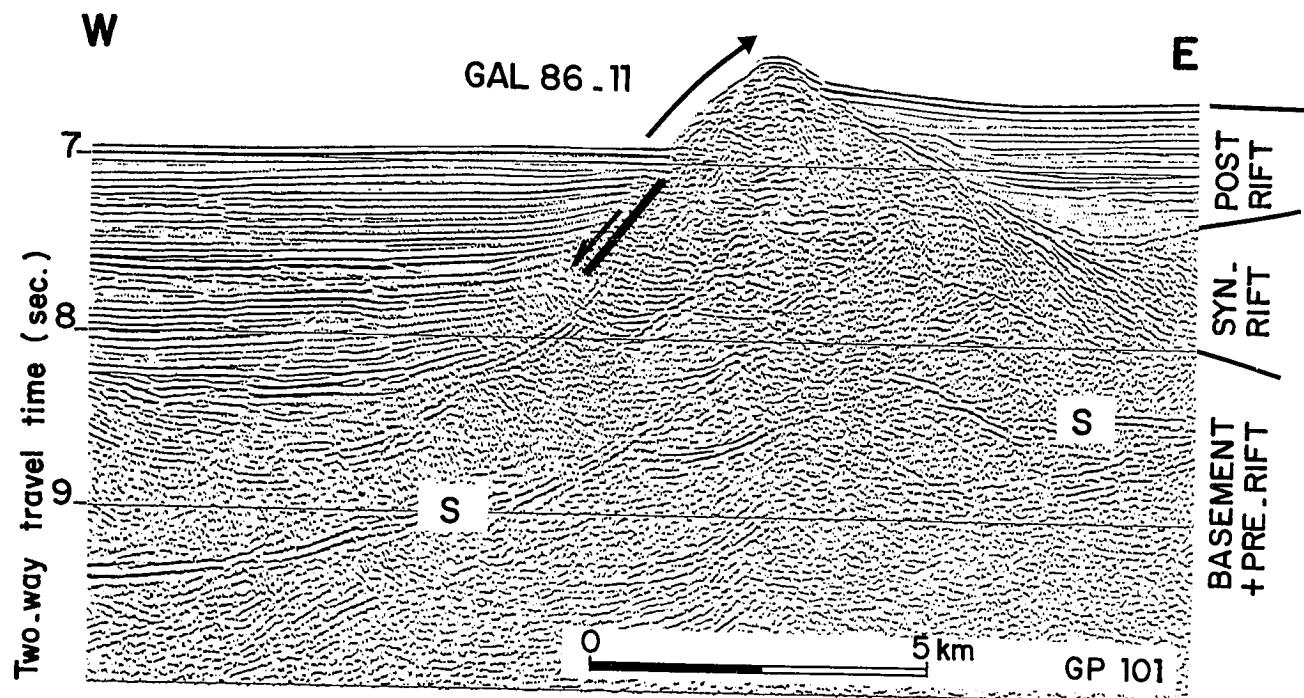


Fig. 9.—Section of the seismic line GP 101, showing the S reflector. The acoustic basement underlying the S reflector consists of granodioritic rocks, sampled on the normal fault scarp by the submersible Nautile during the dive GAL 86-11 (after Boillot *et al.*, *in press. a.* For location, see fig. 4).

Fig. 9.—Sección de la línea sismica GP101, mostrando reflector S. El basamento acústico bajo el reflector S está constituido por rocas granodioríticas, muestreadas en el escarpe de falla normal por el sumergible Nautile durante la inmersión GAL 85-11 (según Boillot *et al.* *in press. a.* Localización en la fig. 4).

ning of the Bay of Biscay by seafloor spreading (Le Pichon *et al.*, 1971; Olivet *et al.*, 1984). Finally, during the latest Cretaceous to the late Eocene, the Iberian plate moved northwestwards with respect to the Eurasian plate (Olivet *et al.*, 1984; Srivastava and Tapscott, 1987). The motion (about 150 km) resulted in the subduction of the Bay of Biscay oceanic lithosphere beneath Iberia, with related deformation of the former passive margin, which was turned into an active margin for a few million years (Grimaud *et al.*, 1982; fig. 10). In the following sections, we consider both the structures inherited from the rifting stage of the Bay of Biscay, and the deformation resulting from the active stage of the margin.

### 3.1. The Mesozoic continental rift.

The total thickness of the Mesozoic sedimentary cover of the northern Spanish margin ranges from 1000 m to 1500 m (fig. 11 and 12). On the basis of observations and sampling data from coring, dredge hauls and dives using the french submersible Cyana (Boillot *et al.*, 1971, 1973, 1976, 1985, 1987b; Lamboy and Dupeuble, 1975; Malod and Boillot, 1980; Malod *et al.*, 1982; Groupe Cybère, 1984), this cover can be divided into pre-, syn- and post-rift sequences, separated by unconformities and/or distinguished by changes in sedimentary facies. In this section, we address only the pre- and syn-rift sequences.

**Middle Jurassic-Berriasian carbonate (700-400 m).** This sequence consists of platform limestones, locally sandy or dolomitic, which were deposited at shallow depths, sometimes in the tidal zone. The formation is ubiquitous on the whole north Spanish margin, and was deposited prior to the main rifting event.

**Berriasian-Aptian (300-600 m).** This sequence is different on Ortegal Spur (Fig. 11), on the present Asturian and Basque shelf, and on Le Danois Bank to the east (fig. 12). a) At Ortegal Spur, the rocks are principally sandy limestone, siltstone and sandstone. On the Asturian and Basque shelf, they are mainly limestone and marl. The benthic foraminiferal assemblage, as well as plant debris, suggest a shallow-water depositional environment (0-200 m). b) On the contrary, Hauterivian turbidite-like sandstones rich in plant debris and Barremian-Aptian black shale sampled on the northern slope of the Le Danois Bank were probably deposited in deep water. They resemble the lower Cretaceous sequence drilled during Leg 103 on the west Galicia margin. In fact, Ortegal Spur and the present Spanish shelf were located on the shoulder of the rift during that time, when the Le Danois Bank slope was probably located close to the rift axis, in an area of rapid subsidence.

The arrangement of the Mesozoic rift at the northern boundary of Iberia during the Mesozoic is complex. Clearly, the normal faults and tilted blocs of the west Galicia margin trend N-S. They are offset by N70°

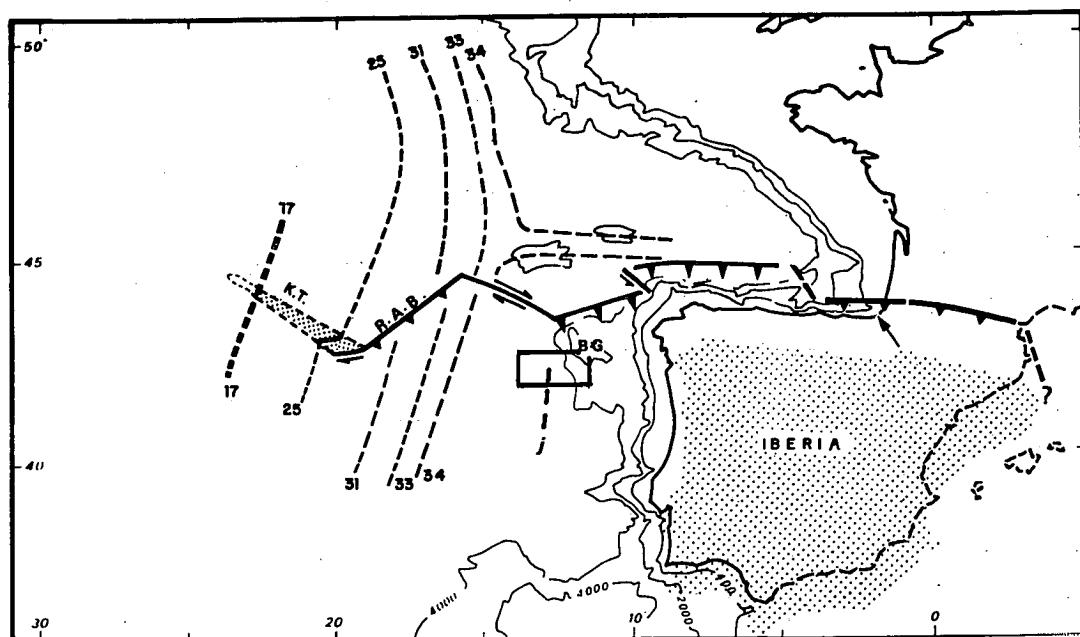


Fig. 10.—Eocene plate boundaries and magnetic anomalies in the north-east Atlantic. The shaded area represents the reconstructed position of Iberia when anomaly 33 was formed. Arrow in northern Spain indicates direction of motion of Iberia with respect to Europe during Paleocene and Eocene times. GB: Galicia Bank; KT: King's Trough; ABR: Açores-Biscay Rise. Leg. 103 drilling area is outlined. Bathymetric contours in meters. Magnetic anomalies from Kristoffersen (1978). Modified from Grimaud *et al.*, (1982).

Fig. 10.—Límite de placas durante el Eocene y anomalías magnéticas en el noreste del Atlántico. El área sombreada marca la posición de Iberia cuando se produjo la anomalía 33. Las flechas en el norte de España indican la dirección del movimiento de Iberia con respecto a Europa durante el Paleoceno y Eocene. GB: Banco de Galicia; KT: King's Trough; ABR: Cresta Azores-Vizcaya. El recuadro limita el área de los sondeos del Leg. 103. Batimetría en metros. Anomalías magnéticas según Kristoffersen (1978). Modificado de Grimaud *et al.*, (1982).

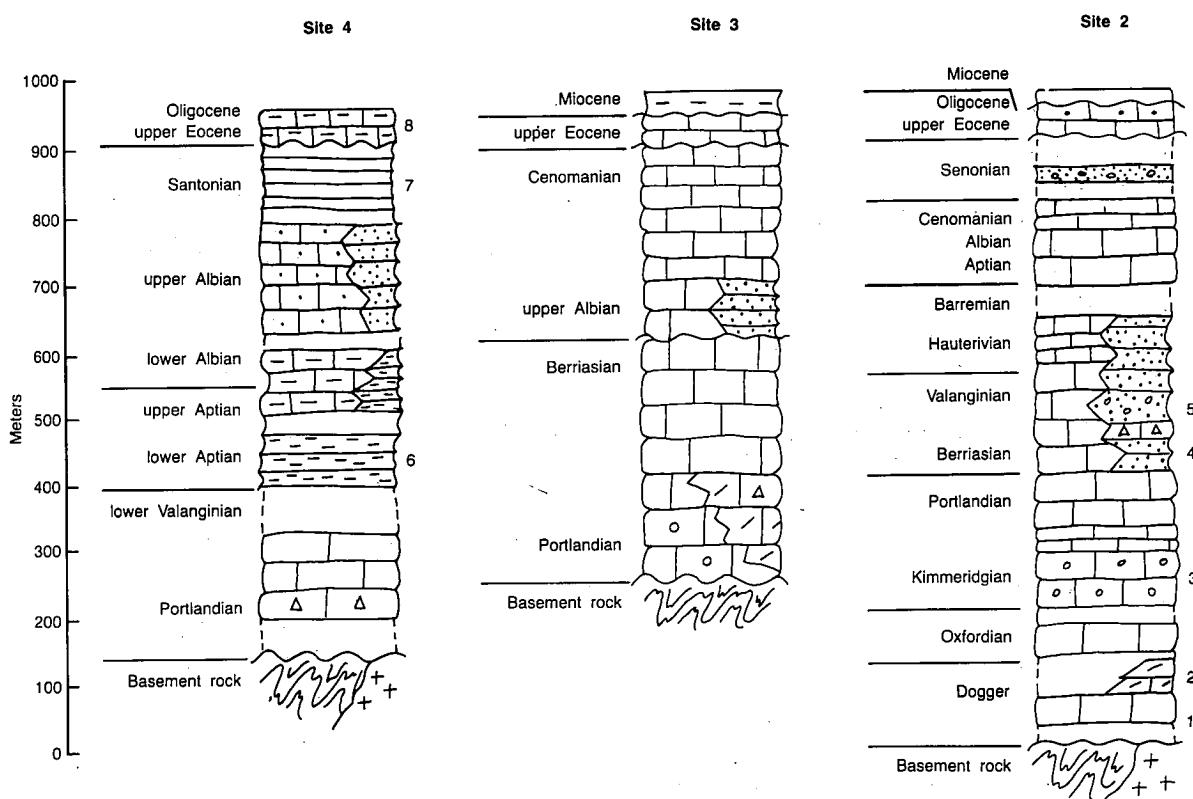


Fig. 11.—Stratigraphic log at Ortegal Spur, derived from samples collected during dives and dredges. 1. massive limestone; 2. dolomitic limestone; 3. brecciated and gravelly limestones; 4. sandstone; 5. coarse sandstone and conglomerate; 6. silt; 7. shale; 8. marly limestone. For location of the dive Sites, see fig. 14. (After Boillot *et al.*, 1987 b).

Fig. 11.—Perfil estratigráfico en el Ortegal Spur, a partir de muestras tomadas con dragas e inmersiones. Leyenda: 1.—Calizas masivas; 2. Calizas dolomíticas; 3. Calizas brechificadas y calcareitas; 4. Limos; 5. Conglomerados y areniscas de gran grueso; 6. Limos; 7. Lutitas; 8. Calizas margosas. Localización de inmersiones en la fig. 14 (según Boillot *et al.*, 1987).

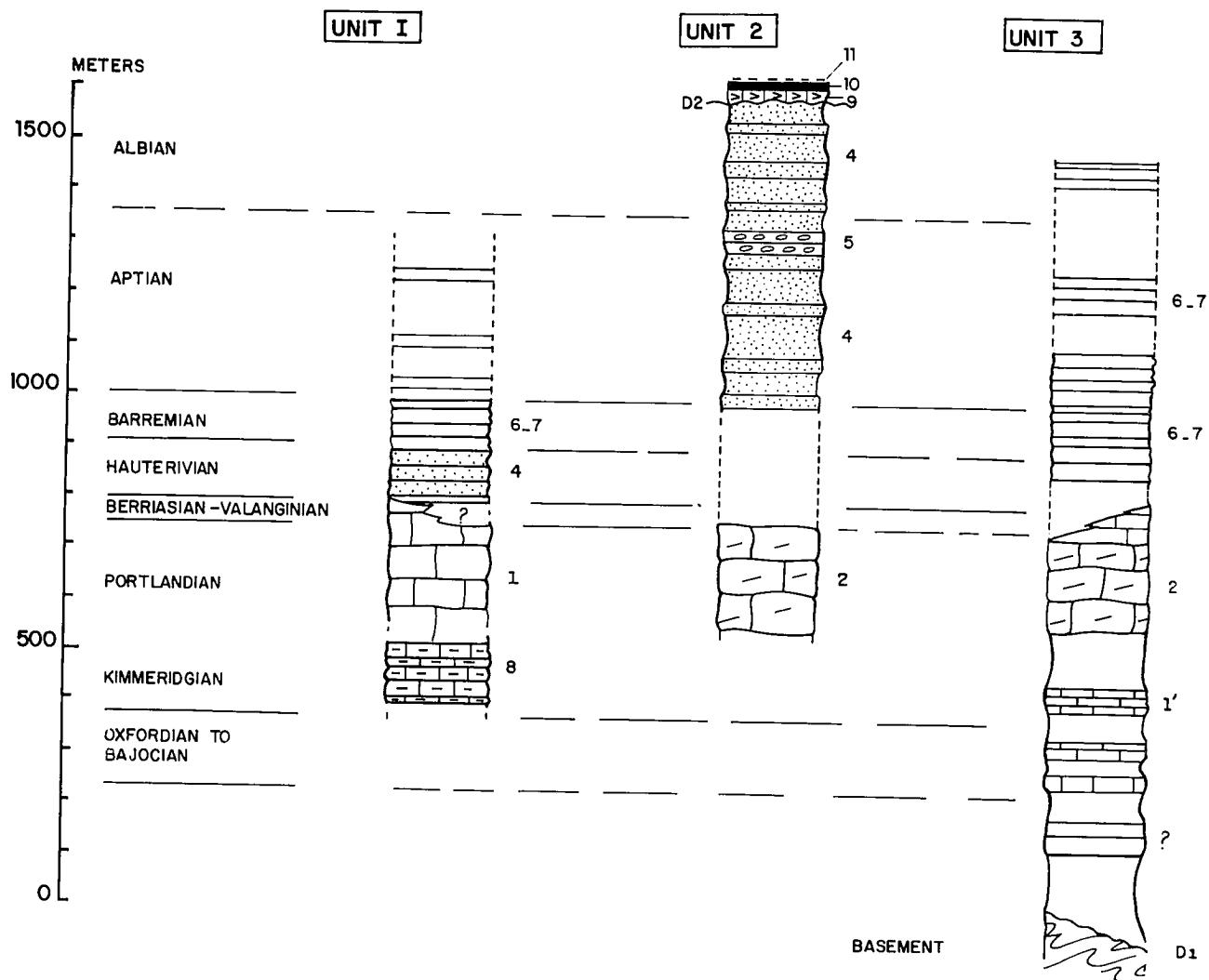


Fig. 12.—Stratigraphic log on the Le Danois Banck (north Spanish slope). Same legend as on fig. 11, plus: 9. Early Tertiary rerefal limestone; 10. Aquitanian limestone and Pliocene marl. After Malod *et al.*, 1982.

Fig. 12.—Perfil estratigráfico del Banco de Danois (talud del norte de España). Leyenda como en el fig. 11, y además: 9. Calizas arrecifales del Terciario inferior; 10. Calizas del Aquitaniense y margas del Plioceno. Según Malod *et al.*, (1982).

transverse faults (fig. 3). On the slope of Ortegal Spur, above the 2500 m isobath, main escarpments follow the N50°-N70° direction (Vanney *et al.*, 1984). These escarpments correspond to outcrops of basement rock and Jurassic limestone separated by faults and locally covered by Albian-Cenomanian post-rift sediments (fig. 13). Consequently, we interpret the N50°-N70° fractures as normal faults of the rift, reactivated to varying degrees during the Cenozoic (fig. 14). The N120°-150° faults also depicted in fig. 14 are hypothetical transverse faults, inferred as having produced the offset of N50°-N70° structures.

To the north of Asturias and Basque areas, the direction of the Cretaceous passive margin is about N120° (see below). Rift structures following that direction are partly preserved in the Basque region (Rat *et al.*, 1983; Wiedman *et al.*, 1983). Such large faults as the Vidio fault northeast of Galicia (Lamboy and Dupeuble, 1975), the Ventaniella fault in Asturias (Martínez-Alvarez, 1968) and the Urbenia fault in the Cantabrian chain (Pujalte, 1981) also strike in this direction, and

can be either normal or strike-slip faults of the rift, reactivated later by Cenozoic tectonics. Conversely, N70° structures recognized in the same regions can be interpreted as transverse or normal faults.

The origin of the fault geometry is still under active discussion. According to Boillot (1986), the N120° fractures are normal faults resulting from NE-SW extension during the rifting stage (fig. 15). On the contrary, Olivet *et al.*, (1984) and Malod (in press.) consider that the SE motion of Iberia with respect to Eurasia started during the rifting stage, i.e. prior to Albian. They interpret the N120° fractures as strike-slip faults connecting pull-apart basins of a segmented rift (fig. 16). In fact, it is possible that NE-SW extension prevailed in early phases of the rifting, while NW-SE extension preceded the drifting in the Bay of Biscay (Malod, in press.) The present North Spanish margin probably experienced several stages of crustal extension, during Triassic, Late Jurassic and Early Cretaceous times. Unfortunately, the first stages are poorly documented, and even for the Early Cretaceous stage, the kinematics of

the extensional process remains a topic for further research.

### 3.2. The Mesozoic passive margin.

Seafloor spreading started during late Aptian or early Albian times in this part of the North Atlantic (Montadert *et al.*, 1979). The Galicia and Asturias margins evolved as passive margins during the rest of the Cretaceous, while the Basque and Pyrenean regions were locally deformed by strike-slip motion along the north-Pyrenean transform zone (Choukroune and Matrauer, 1978). There, E-W Albian-Cenomanian basins are superimposed on the former rift structure, and can be interpreted as pull-apart basins that experienced rapid subsidence, high heat flow and associated thermal (Pyrenean) metamorphism (see Banda and Wickham, 1986).

The post-rift sequence of the north Spanish margin is comparable to that of the deep margin of Galicia Bank. It is separated from the underlying layers by the break-up (post-rift) unconformity. Albian fine-grained sandstones, rich in organic matter and deposited in deep water, have been sampled on the Le Danois Bank slope. Closer to the continent, the Upper Cretaceous consists of flysch deposited along the Cretaceous continental slope. Today this domain (the Cretaceous slope) corresponds to the Basque basin, the Asturian outer continental shelf and probably to the deepest part of the north Galicia margin (Ortegal Spur). The flysch has not been found on the Le Danois Bank, which was probably located beyond the influence of turbidity flows originating mainly from the east and the south. Landwards, the Upper Cretaceous shallow water limestones were deposited on the internal part of the present shelf. Clearly, the Cretaceous paleogeographic boundaries follow the N120° direction, which is also the general trend

of the former passive margin in this area. As discussed in the preceding section, this margin can either be interpreted a transform or a divergent passive margin. In any case, it is oriented obliquely to the Cenozoic E-W North Spanish margin.

### 3.3. The Cenozoic active margin.

Short term latest Mesozoic-early Cenozoic convergence between the Iberian and European plates transformed the Cretaceous passive North Spanish margin into an active margin with southward subduction of the ocean lithosphere of the Bay of Biscay. This geological event resulted in the formation of a variety of structures in the North Spanish marginal trench and margin (fig. 17):

#### 1. The North Spanish marginal trench

The Paleocene-Eocene marginal trench bounding the north-Spanish slope in characterized by gravity and seismic data.

a) A belt of negative anomalies in the gravity field (-100 to -150 milligals) follows the margin from the Galicia Bank to the Basque country (Sibuet and Le Pichon, 1971; Lalaut *et al.*, 1981; fig. 18). To the west, off Asturias and Galicia, the trench is about 100 km wide and 700 km long; it is cut by the Thetha Gap transform fault (Grimaud *et al.*, 1982; fig. 10). To the east, it splits into two branches, bounding the northern and southern sides of the Landes Plateau. The southern branch follows the Gouf de Cap Breton submarine canyon and then corresponds to the north Pyrenean frontal thrust.

b) A sedimentary basin is superimposed on the belt of gravity anomalies. The northern part of the trench is buried by undeformed sediments deposited before (Late Cretaceous) or during (Paleocene-Eocene) the subduction process (Derégnacourt and Boillot, 1982a

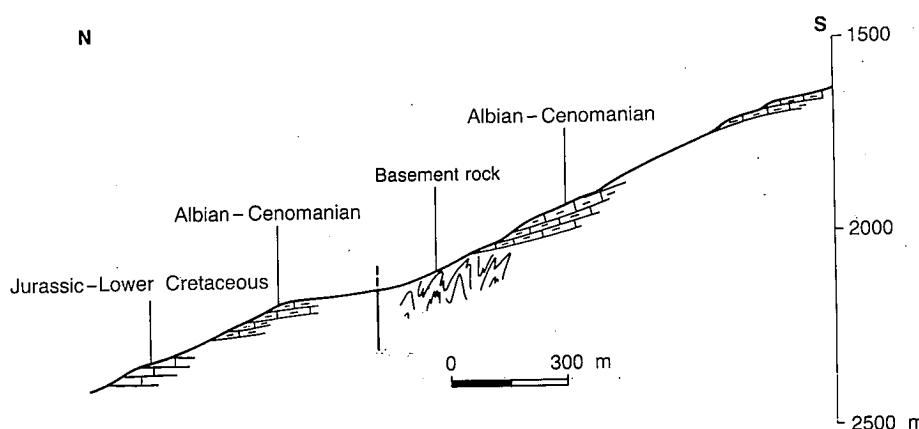


Fig. 13.—Geological cross section at dive Site 3, Ortegal Spur. Albian-Cenomanian sediments rest unconformable on the basement in the upper part of the cross section, and on the Jurassic and early Cretaceous in the lower part (post-rift or break-up unconformity, late Aptian or early Albian in age at that place). After Boillot *et al.*, (1987b). For location, see fig. 14.

Fig. 13.—Corte geológico en la inmersión número 3 en el Espolón de Ortegal. Los sedimentos albienses y cenomanienses descansan discordantemente sobre el basamento en la parte alta del corte y sobre los materiales jurásicos y cretácicos en la parte baja (la discordancia postrifting se localiza en el Aptiense terminal o Albiense basal). Según Boillot *et al.* (1987b). Localización en la fig. 14.

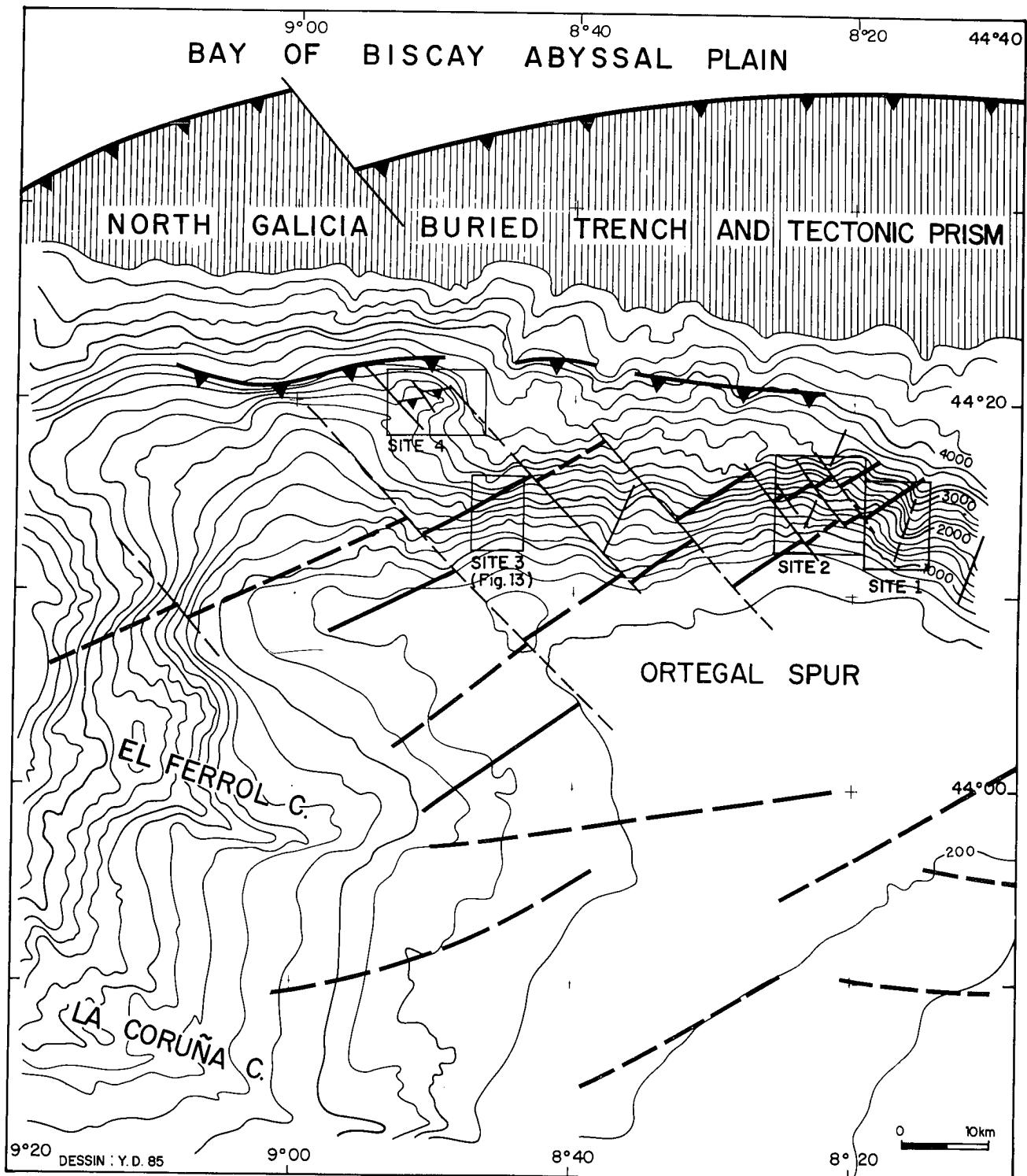


Fig. 14.—Tectonic sketch of Ortegal Spur. Main apparent west-southwest structures are linked to the rifting, and east-west structures of the deep margin are probably related to the extinct north Galicia subduction trench. Dives sites of the Cybere cruise are located. Fault locations on the shelf after Lamboy and Dupeuble (1975), and bathymetric contours after Vanney *et al.*, (1985). Bathymetric contours in meters; contour interval: 200m. From Boillot *et al.*, 1987b.

Fig. 14.—Esquema tectónico del Espolón de Ortegal. Las principales estructuras oeste-suroeste están vinculadas a la fase de *rifting*, mientras que las estructuras este-oeste lo están a la fosa con subducción que se situaba al norte de Galicia. La localización de las fallas en la plataforma se basa en Lamboy y Dupeuble (1975), y las curvas batimétricas en Vanney *et al.*, (1985). Las curvas batimétricas se expresan en metros: equidistancia: 200 metros. Según Boillot *et al.*, (1987b).

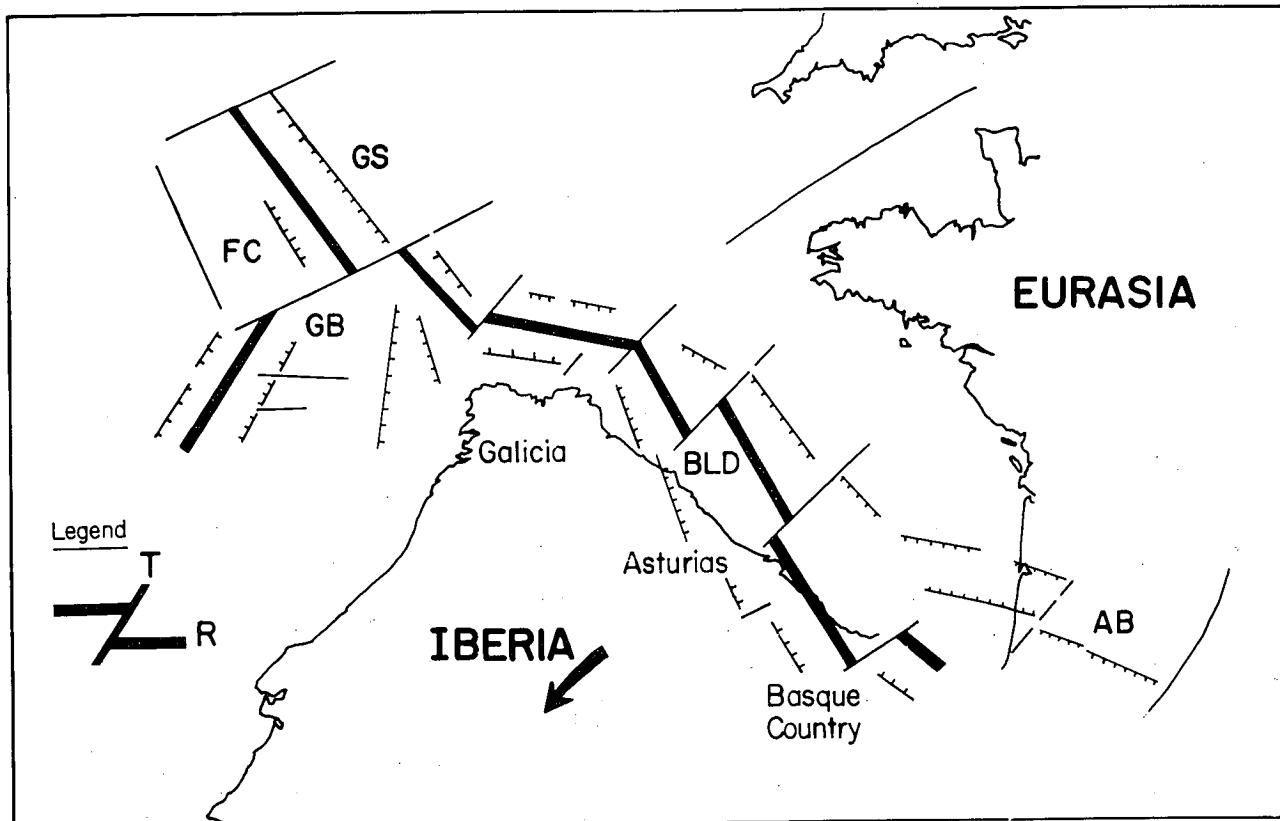


Fig. 15.—Diagram of the Mesozoic continental rift north and west of Iberian plate. At least during the first stage of the rifting, Iberia moved south-west with respect to Europe, and the lithosphere stretched along a northeast direction. Arrow indicates the motion of Iberia with respect to Europe. AB: Aquitaine basin; BLD: Le Danois Bank; FC: Flemish Cape; GB: Galicia Bank; GS: Goban Spur; OS: Ortegal Spur; R: Axis of the continental rift; T: Transverse fault; After Boillot *et al.*, (1987b).

Fig. 15.—Esquema del rift continental al norte y oeste de la Placa Ibérica. Al menos durante la primera etapa de *rifting*, Iberia se movió hacia el suroeste con respecto a Europa, y la litosfera se expandió hacia el nordeste. La flecha indica el movimiento de Iberia con respecto a Europa. AB: Cuenca de Aquitania; BLD: Le Danois Bank; CF: Flemish Cape; GB: Banco de Galicia; GS: Espolón de Goban; OS: Espolón de Ortegal; R: eje del *rift* Continental; T: Falla transformante; Según Boillot *et al.*, (1987b).

and b). The southern part, along the north-Spanish slope, consists of a continuous belt of pre-Oligocene deformed sediments (Montadert *et al.*, 1971), now interpreted as a tectonic accretionary prism of an active margin (fig. 19).

c) The tectonic contact between the thrusted margin and the accretionary prism cannot be recognized on seismic records. It is assumed to follow the bottom of the continental slope. To the east, this major thrust zone divides into two branches corresponding with those defined on the gravity anomalies (fig. 17).

The timing of the subduction (pre-, syn-, and post subduction stages) is based on seismic correlations from holes DSDP 118 and 119 (Laughton, Berggren *et al.*, 1972). It is not specifically defined, because faults prevent direct correlations between seismic reflectors. Thus, the subduction process may have continued episodically after the Eocene time, during the Oligocene and even during the Neogene.

## 2. Folds and possible reverse or strike-slip faults on the North Spanish margin.

A major Cenozoic unconformity occurs on the As-

turian shelf, separating Cretaceous folded and eroded sediments (Lower Cretaceous carbonates; Upper Cretaceous flyschs) from tabular, neritic Upper Eocene limestone and sandstone. On the other hand, reverse and possible strike-slip Cenozoic faults are recognized or postulated on the margin, and especially on the slope. By comparison with the pre-late Eocene structures of the Asturian shelf, the Cenozoic structures of the rest of the margin are assumed to be mainly Paleocene and/or Eocene in age. Nevertheless, later tectonics also occurred on the margin, especially during late Oligocene. On the shelf, it is possible to separate the effects of these different tectonic events, but it is difficult to do so on the slope because of the lack of Eocene deposits.

Thus, the north Galicia and Asturias margin is clearly an orogenic belt related to subduction. The deformation was probably enhanced by the strike slip component of the relative motion (NW-SE) of the converging plates (fig. 10). A significant thickening of the crust was achieved through this process since the Cretaceous margin is now considerable uplifted. This folded belt is the clear westward continuation of the Pyrenean belt. The Galicia and Asturias segment of the north Spanish margin were unaffected by collision be-

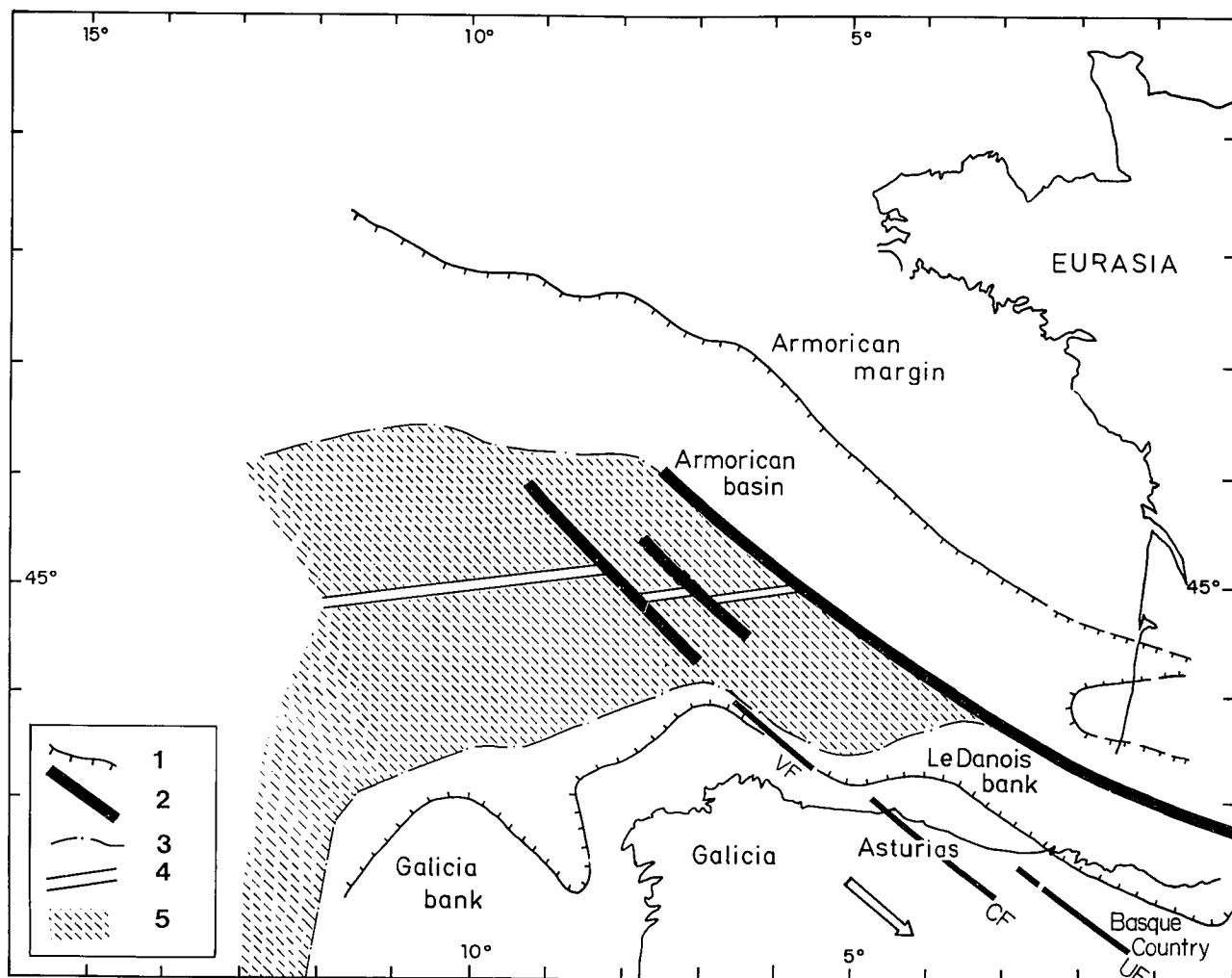


Fig. 16.—Diagram of the regional paleogeography 75 m.y. ago, during Senonian time, after the Southeastward drift of Iberia. Location of Iberia with respect to Europe after Olivet *et al.*, (1984). The margin pattern is partly inherited from the rift geometry as depicted on fig. 15. 1: Upper limit of the continental slope; 2: transform direction; 3: Ocean-Continent boundary; 4: axis of oceanic spreading; 5: oceanic crust. White arrow shows the direction of Iberian motion with respect to stable Europe. C.F.: Cantabrica fault; V.F.: Urbenia fault; VF: Vidio fault.

Fig. 16.—Esquema paleogeográfico regional para hace 75 millones de años, durante el Senoniense. Localización de Iberia con respecto a Europa según Olivet *et al.*, (1984). Los rasgos generales del margen están parcialmente heredados de la geometría del *rift* representada en la fig. 15. Leyenda: 1: Límite superior del talud continental. 2: Dirección de las transformantes. 3: Límite océano-continente. 4: Eje de expansión oceánica. 5: Corteza oceánica. La flecha blanca muestra la dirección de movimiento de Iberia con respecto a la Europa estable. C.F. Falla cantábrica; UF: Falla de Urbenia; VF: Falla de Vidio.

cause the Iberia-Eurasia convergence stopped before the complete subduction of the Bay of Biscay oceanic lithosphere. However, the Basque area and the Pyrenees are a collisional chain, where the two margins of the Cretaceous flysch basin are involved in the tectonics. Nevertheless, a clear continuity exists between the subduction chain to the west and the collision chain to the east, and this fact of major importance must be taken into account when interpreting the entire fold belt (Boillot, 1984).

The fig. 20 is a diagrammatic picture of the offshore Cenozoic structures on the north Spanish margin, including the Basque country. We believe that those structures resulted mainly from reactivation of former rift structures, except those located on the deep margin

which were related to the marginal trench and probably created by Cenozoic subduction and/or collision.

#### 4. CONCLUSION: SOME OUTSTANDING PROBLEMS

The Pyrenees and their relationship to the north Spanish margin are beyond the scope of this paper. Nevertheless, several questions remain open for further discussion:

1. *The Pyrenees, subduction and collision?* By extending to the east the structural pattern of the north-Spanish margin, Boillot and Capdevila (1977) suggested that the Pyrenean collision followed a southward

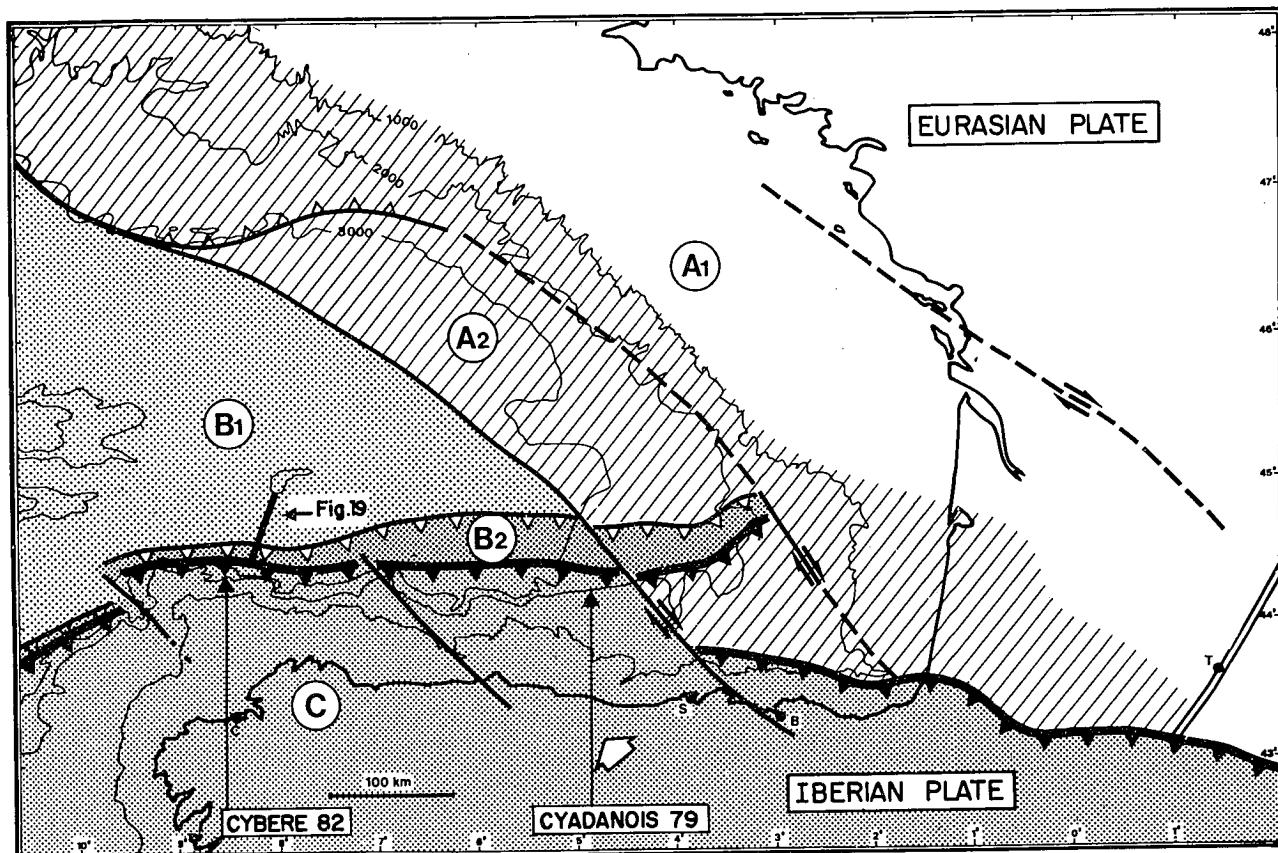


Fig. 17.—Structural diagram of the Bay of Biscay and location of areas studied by diving (Cyadanois and Cybere cruises). A1: normal continental crust; A2: thinned continental crust; B1: oceanic crust; B2: tectonic accretionary prism, built up during Paleocene and Eocene at the Iberian-Europea converging plate boundary; C: thickened continental crust, resulting from the plate convergence and from the southward subduction of the oceanic lithosphere beneath Iberia. White arrow shows the direction of Iberian motion with respect to stable Europe (After Derégnaucourt and Boillot, 1982b).

Fig. 17.—Esquema estructural del Golfo de Vizcaya en el que se sitúan las áreas estudiadas con sumergibles (cruceros Cyadanois y Cybere). Leyenda: A1: corteza continental normal; A2: corteza continental adelgazada; B1: corteza oceánica; B2: prisma de acreción tectónica, formado en el límite convergente de las placas europea e ibérica durante el Paleoceno y Eoceno; C: corteza continental engrosada por efecto de la convergencia de placas y de la subducción hacia el sur de la litosfera oceánica bajo Iberia. La flecha blanca muestra la dirección de movimiento de Iberia con respecto a la Europa estable. (Según Derégnaucourt y Boillot, 1982b).

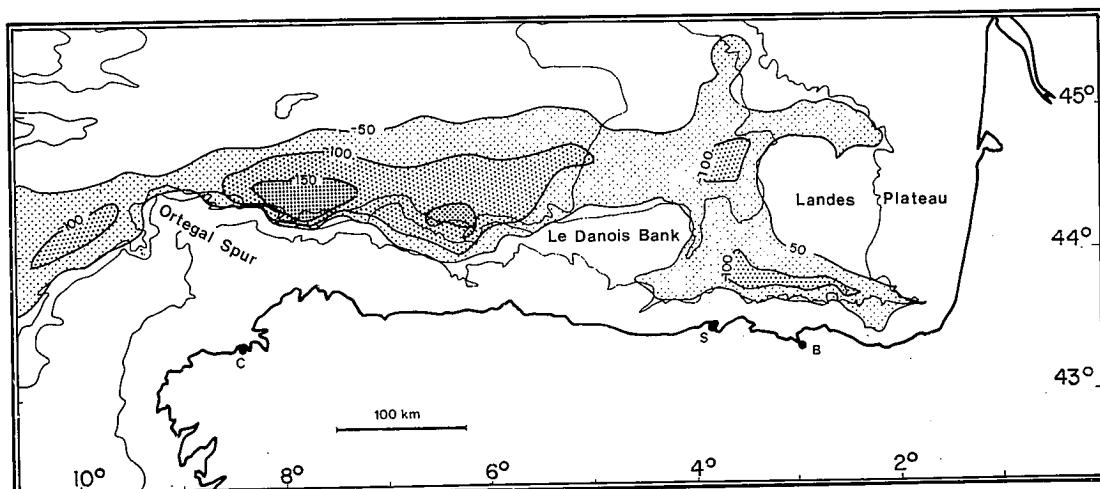


Fig. 18.—Negative gravimetric anomalies along the north Spanish margin, superimposed on the Paleocene-Eocene marginal trench (After Lalaut *et al.*, 1981).

Fig. 18.—Anomalías gravimétricas negativas a lo largo del margen del norte de España, sobreimpuestas a la fosa marginal Paleoceno-Eoceno (según Lalaut *et al.*, 1981).

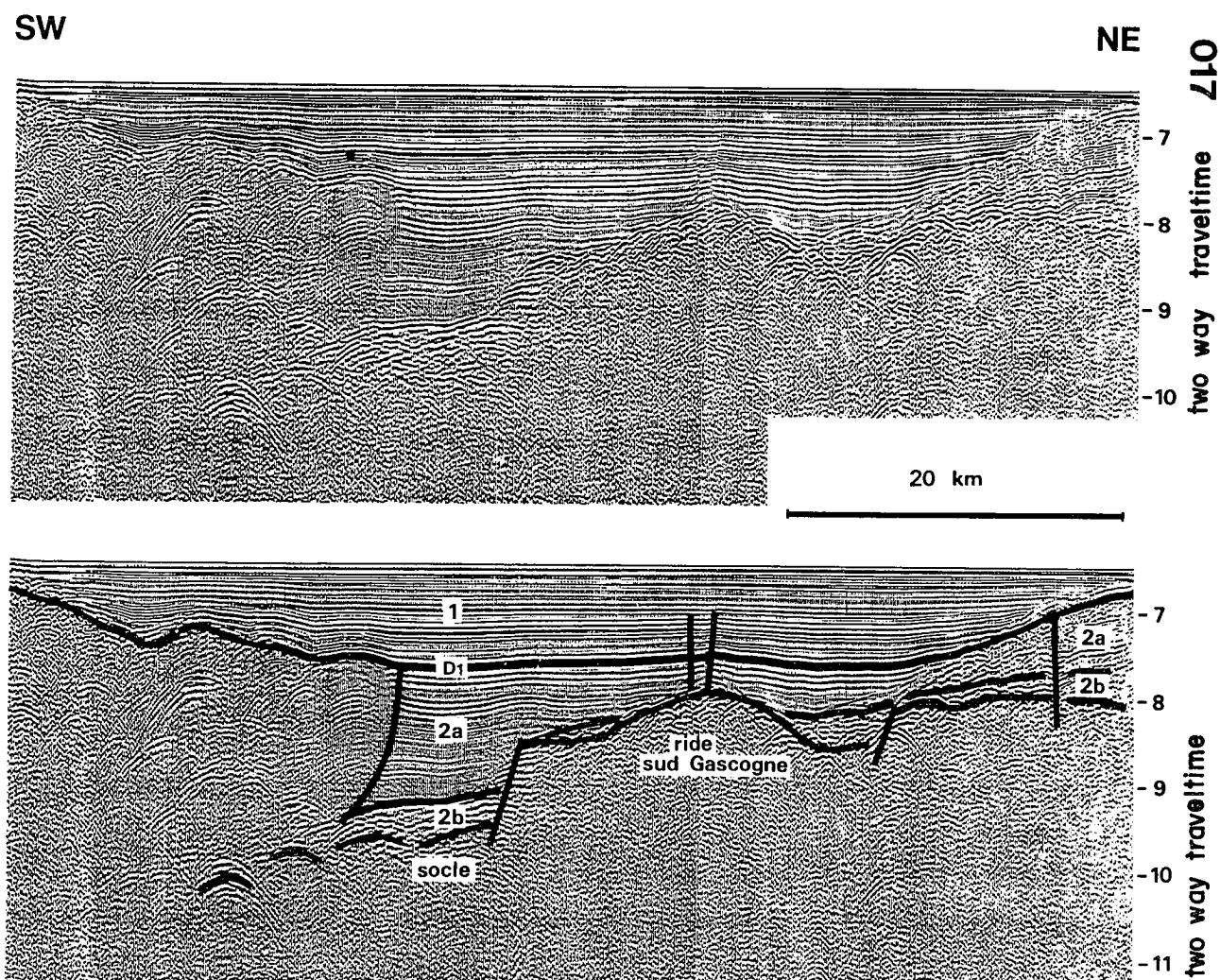


Fig. 19.—Seismic profile crossing the north Spanish marginal trench to the north of Ortegal Spur (For location, see fig. 17).  
1: upper Eocene-Present sediments; D1: pre-upper Eocene unconformity; 2a: Paleocene-lower Eocene sediments; 2b: upper Cretaceous sediments (After Derégnacourt and Boillot, 1982b).

Fig. 19.—Perfil sismico a través de la fosa marginal del norte de España, al norte del promontorio de Ortegal (localización en la fig. 17). Leyenda: 1: sedimentos del Eocene superior a actuales. D1: disconformidad anterior al Eocene superior. 2a: sedimentos del Paleoceno-Eocene inferior. 2b: sedimentos del Cretácico superior (según Derégnacourt y Boillot, 1982b).

subduction of the thinned continental crust of the deep basin ("flysch trough") separating the Aquitaine and Iberian margins during upper Cretaceous time. According to the ECORS Pyrenees Team (1988), this hypothesis is now difficult to support, because of the northward dipping of the deeper seismic reflectors beneath the Pyrenean axial zone. However, these deep reflectors are related to south-Pyrenean thrusting, i.e. late structures by comparison with the north-Pyrenean reverse faults and thrusting. Did a "flip" occur at early Eocene time, and result in a reversal of the subduction direction from south or north? Or did this change occur in space, along a transform zone located in the Basque Pyrenees as proposed by Engeser and Schwentke (1986)?

2. *What happens to the north Iberian Cretaceous margin in the Pyrenees?* Structural, paleomagnetic and even sedimentological data suggest that most of the North Pyrenean zone belonged to the Aquitaine mar-

gin before the Cenozoic collision (see for example Puigdefabregas and Souquet, 1986). It leaves only a very small place to accommodate the Iberian margin to the south, because the axial zone of the Pyrenees was a shelf at that time (Upper Cretaceous). In fact, some elements of the Iberian margin have been recognized in the Western Pyrenees (Camerot *et al.*, 1978; Camerot, 1987; Boirie and Souquet, 1982). They are thrusted southward, as would be expected if they are squeezed tilted fault blocks belonging formerly to the Iberian passive margin. The Igouane and Mendibelza massifs are also thrusted southward, and these may have the same origin. Nevertheless, the Iberian margin is very narrow by comparison with the Aquitaine margin. Have Albian and/or Cenozoic tectonics resulted in partial or total destruction of this margin? Or was the pre-Albian rift originally asymmetrical (Malod, 1987)?

3. *Are there Mesozoic detachment faults beneath*

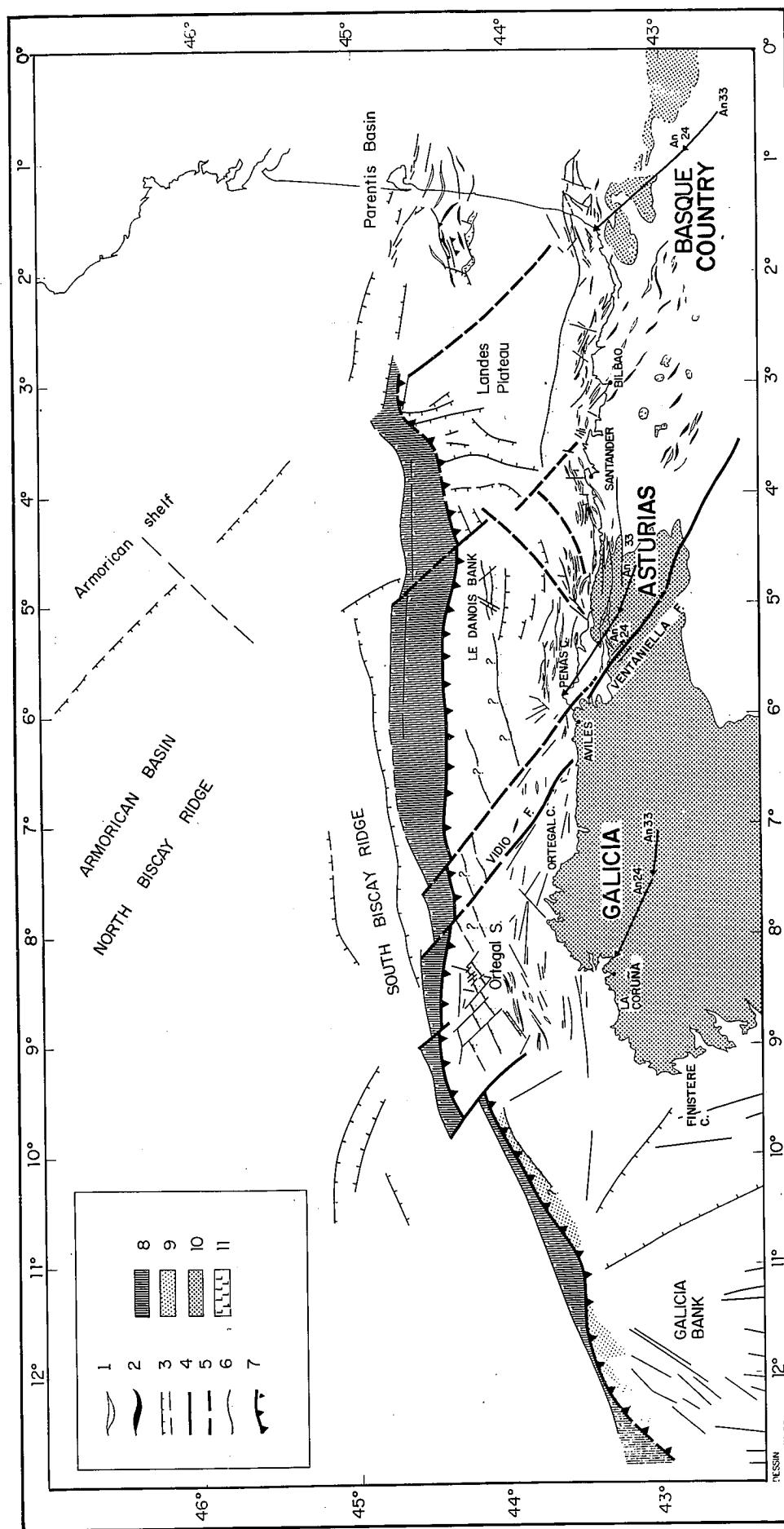


Fig. 20.—Diagrammatic structural map of the North Spanish margin. 1: syncline; 2: anticline; 3: Mesozoic normal fault; 4-5: strike-slip fault; 6: Cenozoic reverse fault; 7: Cenozoic tectonic accretionary prism; 10: Pre-Mesozoic basement (on land); 11: salt diapir.

Fig. 20.—Esquema estructural del margen septentrional de España. 1: sinclinal. 2: anticlinal. 3: falla normal mesozoica; 4-5: fallas de desgarre; 7: Fallas nuevas cenozoicas; 8-9: prisma de acreción tectónico. 10: basamento premesozoico (en tierra). 11: Diapiro salino.

**the Bay of Biscay and Pyrenean margins?** The present west Galicia margin can be used as a model to interpret the Bay of Biscay Cretaceous passive margins. The Armorican margin closely resembles the west Galicia margin (Montadert *et al.*, 1979; Montadert, 1984; Barbier *et al.*, 1986). The timing of events is the same in the two areas, and the rift structures also are similar (normal listric faults, tilted crustal blocks, and deep seismic reflectors S). To explain these features, Le Pichon and Barbier (1987) postulated a syn-rift detachment fault dipping northeast, beneath the Armorican margin, comparable to the eastward dipping detachment fault we postulate beneath the west Galicia margin (see the first part of this paper). On the other hand, the Aquitaine Basin clearly extends eastward along the Armorican margin. Does a remnant detachment fault exist also beneath this basin? Or did the Mesozoic detachment fault dip south, beneath the north Iberian margin (including the Pyrenees), as proposed by Malod (1987)?

To answer these questions, we need new data, especially concerning the deep crustal and lithospheric structure. However, we have also to re-consider available geological and geophysical data in the light of new concepts and models of lithospheric stretching, and by systematic comparisons of present continental margins with former margins now folded in orogenic belts such as the Pyrenees. In this way, we shall certainly achieve new insights into the geology of Iberian continental margins, both offshore and onshore.

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