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45. DRILLING ON THE GALICIA MARGIN: RETROSPECT AND PROSPECT[1]

Gilbert Boillot, Laboratoire de Géodynamique Sous-Marine du CEROV, Villefranche-sur-Mer, France
and
Edward L. Winterer, Scripps Institution of Oceanography, La Jolla, California

ABSTRACT

The principal findings of drilling on the Galicia margin during ODP Leg 103, supplemented by sampling from the submersible *Nautile* and calibration of seismic-reflection profiles, are:

1. Peridotite, exposed in a ridge at the foot of the margin, is clinopyroxene-bearing spinel harzburgite, more than 90% serpentinized and cut by veins of calcite. During the rifting and lithospheric stretching stage of margin formation, these rocks ascended to the seafloor from a depth of about 30 km, where the temperature was about 1250°C. The rocks record the successive effects of partial melting, stretching, serpentinization, and fracturing.
2. The sequence of sedimentary strata in a typical tilted fault block of the margin comprises, in ascending order:
a. An unknown thickness (probably >500 m) of undated sandstone containing volcanic detritus and interbedded with shelly dolomite. (Known only from *Nautile* samples.)
b. A few meters thickness of conglomerate of low-grade clastic metasedimentary clasts, resting on fragments of altered rhyolite. This rhyolite may not be part of a conglomerate but part of the Hercynian basement.
c. About 400 m of Jurassic (Tithonian) shallow-water carbonate rocks, including about 100 m of limestone interbedded with sandstone and claystone, overlain by about 250 m of intensely fractured dolomite. The strata up to this level are classified as "pre-rift," but may have been deposited during an episode of pre-Cretaceous faulting.
d. A syn-rift sequence, more than 1 km thick in the deepest part of the half-graben, consisting of about 40 m of Valanginian calpionellid marlstone overlain by Valanginian-Aptian turbidite sandstone, claystone, and hemipelagic limestone interbedded with debris-flow beds rich in shallow-water bioclasts. At least one angular unconformity internally divides the syn-rift sequence.
e. Albian-Recent post-rift strata, thickest in the half-grabens and thin to absent over the upper edges of the fault blocks. These strata were not systematically explored during Leg 103.
3. The prominent deep seismic reflector, "S," seen on profiles on the western part of the margin, lies within or beneath continental crust.

In spite of the significant advances made possible by the drilling during Leg 103, a number of unanswered important questions remain as obstacles to our building a comprehensive geodynamic model for the evolution of the Galicia margin:

1. The mechanism of emplacement of the ridge of mantle peridotite at the foot of the margin and the respective roles of asthenospheric diapirism and detachment faulting in the uplift and unroofing of the peridotite.
2. The extent on the seafloor of peridotite west of the peridotite ridge bounding the margin.
3. The nature of the regional deep seismic reflector termed horizon S, interpreted as the seismic signature of a detachment fault.
4. The stratigraphy of pre-Valanginian ("pre-rift") strata on the margin and a possible episode of Jurassic or older rifting and crustal thinning.
5. The possible role of hydrothermal fluids rising along the rift faults in the diagenesis (dolomitization and silicification) of pre- and syn-rift sediments.
6. The regional tectonic significance of widespread unconformities within the rift-stage sediments and their relation to the inception of seafloor spreading along adjacent parts of the North Atlantic margins.

INTRODUCTION

Traditionally, each *Scientific Results* volume contains a synthesis chapter written by the Co-Chief Scientists, in which the main results of the leg, as set forth in the other chapters, are summarized. We have elected to depart somewhat from this tradition and, following a brief résumé of the principal results of drilling, to focus mainly on unresolved problems of significant topical or regional interest. Finally, we suggest possible new strategies to solve some of these problems. We have emphasized the bearing of our drilling results on tectonic problems because these are not treated systematically elsewhere in the volume. For overviews and detailed discussions of the petrologic, geochemical, biostratigraphic and geophysical results of Leg 103, the reader is referred to the many special chapters on these topics in this volume. A more lengthly overview of the regional background, drilling objectives, and the drilling results is contained in the introductory chapter in the *Initial Results*.

Following the résumé of main results, we take up these unresolved problems in stratigraphic order, beginning with basement rocks and proceeding upward through the sedimentary cover. Tectonic problems are addressed in the context of the rocks that are most closely affected by the tectonism. Most of the data from which we assembled this chapter comes from site chapters of the *Initial Results* (Boillot, Winterer, et at., 1987) and from the specialized chapters in this volume. Although some background data are included in our discussions, we assume the

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

reader has access to these other chapters. Because they include extensive references to the pertinent literature, we have generally referenced these chapters rather than the outside literature, and the reader is referred to them for further detailed information and references.

There is a wide diversity of views among the chapter authors of the *Scientific Results*, and where this pertains to the problems under discussion, we try to present the contrasting interpretations, including our own. It is not our intention to pontificate, but rather to point out the differing interpretations. The initials after the section headings indicate the main author.

Leg 103, without doubt, advanced our knowledge and ideas about the formation of passive margins. Especially, it opened new perspectives into the geodynamic processes responsible for crustal thinning beneath passive margins and for the emplacement of mantle peridotites along the ocean/continent boundary. Some of the outstanding problems now seem to be well framed, and their solution through further work along the Galicia margin appears possible for the same reasons that made Ocean Drilling Program (ODP) Leg 103 a success: the presence of only a thin and discontinuous sedimentary cover and hence, unusually easy access to continental basement, peridotite, and oceanic crust. In keeping with the recommendations of the July 1987 Conference on Scientific Ocean Drilling (COSOD II), which affirmed that passive margins and the dynamics of their formation are among the most important scientific questions for the next decade, we believe that exploitation of the exceptional conditions along the Galicia margin should continue.

RÉSUMÉ OF MAIN RESULTS OF ODP LEG 103

Objectives of Drilling

The Galicia margin of Iberia (Figs. 1–3) is an exceptionally attractive target for drilling. The part of the margin seaward from Galicia Bank is a passive margin that has been but slightly affected by post-rift tectonism and is now generally buried under only a thin cover of post-rift sediments. The structures and rocks revealed by seismic-reflection profiles (Fig. 4) and by dredging are typical of many passive margins, with continental basement broken by normal faults into long, narrow tilted blocks (Thommeret et al., this volume). Wedges of sediment partly fill the half-grabens, and crystalline basement and sedimentary strata crop out on the exposed upper edges of the blocks. Strata recording the early part of the evolution of an Atlantic passive margin are thus accessible to the drilling technology available on *JOIDES Resolution*.

To extend the work begun on Deep Sea Drilling Project (DSDP) Leg 47B, at Site 398 (Sibuet, Ryan, et al., 1979) (Fig. 2), which explored most of the post-rift (post-Aptian) sediments but only scratched the uppermost part of the syn-rift sequence, various proponents of Galicia drilling and the JOIDES Atlantic Regional Panel and the Planning Committee formulated a drilling program designed mainly

1. To core the entire syn-rift and pre-rift sedimentary sequence and the upper part of the basement, preferably at a single site. This program was begun at our Site 638, where we failed to penetrate the entire section, and was extended deeper at Site 639 (Leg 103 site locations are shown on Fig. 2). The aim was to date the beginning of rifting and to date the various unconformites seen within the syn-rift sequence on seismic profiles. We also planned to explore the Jurassic carbonate platform rocks, known from dredging and believed to be pre-rift, and to sample any pre-platform strata. Time permitting, we planned to drill another nearby site where the syn-rift/post-rift boundary could be dated. This became our Site 641.
2. To elucidate the nature of the peridotite known from dredging to crop out on a ridge at the foot of the margin, close to the ocean/continent boundary. A site was chosen a few kilometers north of where highly altered peridotite had been dredged, where seismic profiles showed the ridge to be buried by Cenozoic sediments. This became our Site 637. Fresh samples were needed to constrain the hypotheses of origin: is this a mantle diapir, exposed by tectonic unroofing during extension or perhaps a part of the Hercynian crystalline basement?
3. To sample the summit of a tilted block close to the ocean/continent boundary. Beneath the block, the prominent regional S reflector is conspicuous on seismic profiles (Fig. 4), and we hoped to constrain theories for its origin. This reflector, as well an another reflector similarly situated beneath the Armorican margin, had been ascribed by de Charpal et al. (1978), Montadert et al. (1979), and Chenet et al. (1982) to the boundary between brittle and ductile continental crust. This became our Site 640.

Drilling Results

Site 637

Drilling results on the east side of the ridge at Site 637 (Fig. 5) show that the ridge consists, at least in part, of serpentinized peridotite, confirming the previous dredge results. A total of 74 m of peridotite was penetrated, beneath a cover of about 32 m of Paleogene clay and 180 m of Neogene marl, clay, and turbidite sands.

The peridotite is clinopyroxene-bearing spinel harzburgite, which is more than 90% serpentinized and is cut by veins of calcite and serpentinite. Locally, the peridotite contains rare plagioclase rimming the spinel crystals. Thin sections of the peridotite show both porphyroblastic and mylonitic textures, the latter being restricted to thin shear zones. The pervasive foliation dips about 30°–45° in the upper 50 m of the section, increasing to about 65° in the lower 20 m. Lineation formed by elongate pyroxene and spinel crystals is mostly downdip. Shipboard magnetic data show that the foliation and the lineation dip east. Analysis of mylonitic texture in the shear zones consistently shows a top-to-the-east sense of shear on surfaces that dip east. After shearing, the peridotite was intensively serpentinized and then fractured. During the late stages of deformation, calcite and serpentinite replaced and infilled fractures.

The Cenozoic sediments overlying the peridotite form three lithologic units. From oldest to youngest, these include (1) lower Paleogene pelagic and hemipelagic brown and reddish clay (180–212 m below seafloor [mbsf]), (2) upper Miocene to middle Pliocene interbedded marl and calcareous turbidites and lesser amounts of brown clay (135–180 mbsf), and (3) middle Pliocene to upper Pleistocene turbidites, probably derived from the Iberian mainland (0–135 mbsf) (Comas and Maldonado, this volume).

Sites 638, 639, and 641

These three sites sampled most of the Cretaceous and Upper Jurassic sediments on one of the tilted fault blocks of the Galicia margin (Figs. 6 and 7). The composite sedimentary sequence, constructed from data from the three sites, comprises the following units (Fig. 8):

1. Hydrothermally altered rhyolitic rocks, which may be part of Hercynian basement, overlain by a thin layer of conglomerate composed of a great variety of low-grade metasedimentary rocks.
2. About 400 m of Upper Jurassic (Tithonian) shallow-water carbonates with lesser amounts of sandstone and claystone. The uppermost 250 m (?) is thoroughly dolomitized and cut by several generations of veins and fractures.
3. About 40 m of lower Valanginian calpionellid-bearing marlstone, probably deposited at moderate depth. We associate

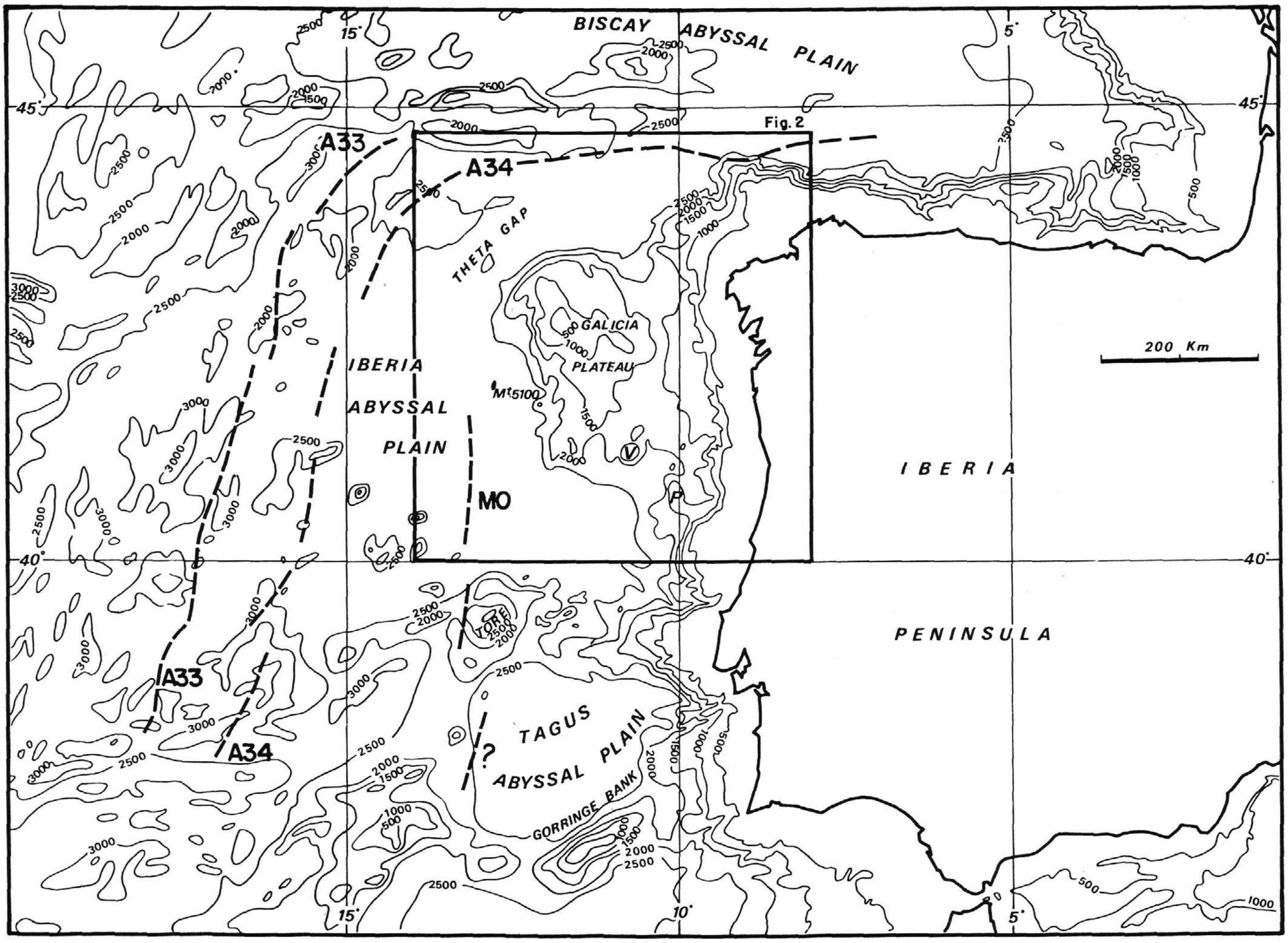

Figure 1. Atlantic margin of Iberia and adjacent abyssal plains. Contours in fathoms; contour interval, 500 fm. V = Vigo Seamount; P = Porto Seamount. From Laughton et al. (1975). Magnetic anomalies after Guennoc et al. (1979). The area shown in Figure 2 is outlined.

this deepening with the beginning of the cycle of rifting that created the half-grabens.

4. About 400 m of upper Valanginian and Hauterivian interbedded turbidite sandstone and claystone rich in terrestrial plant debris, and Barremian and Aptian alternating clay, calcareous clay, marl/marlstone, and clayey limestone, including thin turbidites and debris flows, deposited in relatively deep water during rifting. We place the age of the boundary between syn- and post-rift sediments at the top of the Aptian.

5. About 200 m of post-rift Albian black shale and Upper Cretaceous marl and brown clay, including 25 cm of black zeolitic clay of an age close to the Cenomanian/Turonian boundary (see Thurow et al., Dunham et al., and Stein et al., this volume).

6. About 184 m of upper Miocene–lower Pleistocene nannofossil ooze, deposited as fill in a broad valley eroded into Lower Cretaceous sediments.

Site 640

Site 640 was designed to sample the upper part of the rocks seen overlying the S reflector on seismic-reflection profiles (Fig. 9). Although the S reflector lies nearly 2 s (two-way traveltime [twt]) beneath the site, sampling of the acoustically confused rocks above it would constrain the origin of the reflector: are the rocks sedimentary or crystalline, oceanic or continental, pre-rift or syn-rift?

Coring showed that the uppermost strata beneath 157 m of Neogene ooze and Paleogene and Upper Cretaceous brown clay are syn-rift sediments, including Hauterivian interbedded turbidite sandstone, mudstone, and marlstone, Barremian ooze and clay, and Aptian-Albian ooze (Fig. 8). Only 62 m of these strata had been cored when drilling was terminated for lack of time. The results indicate that the original deposition was in a syn-rift basin on continental crust, but the more fundamental question of the material significance of the S reflector remains unanswered. More recently, diving with the submersible *Nautile* along a fault scarp close to Site 640 demonstrated that the S reflector, apparent on seismic lines at the dive site, is located within or beneath the continental crust (Boillot et al., this volume).

DISCUSSION

In this section we focus on some of these scientific results and also on the major problems left incompletely resolved by the studies on the Galicia margin, including Leg 103 drilling and the 1986 dives of the submersible *Nautile*.

Basement Across the Ocean-Continent Boundary (GB)

In our opinion, one of the main contributions of Leg 103 to the understanding of passive margins comes from the results at Site 637, on the peridotite ridge along the west edge of the Galicia margin.

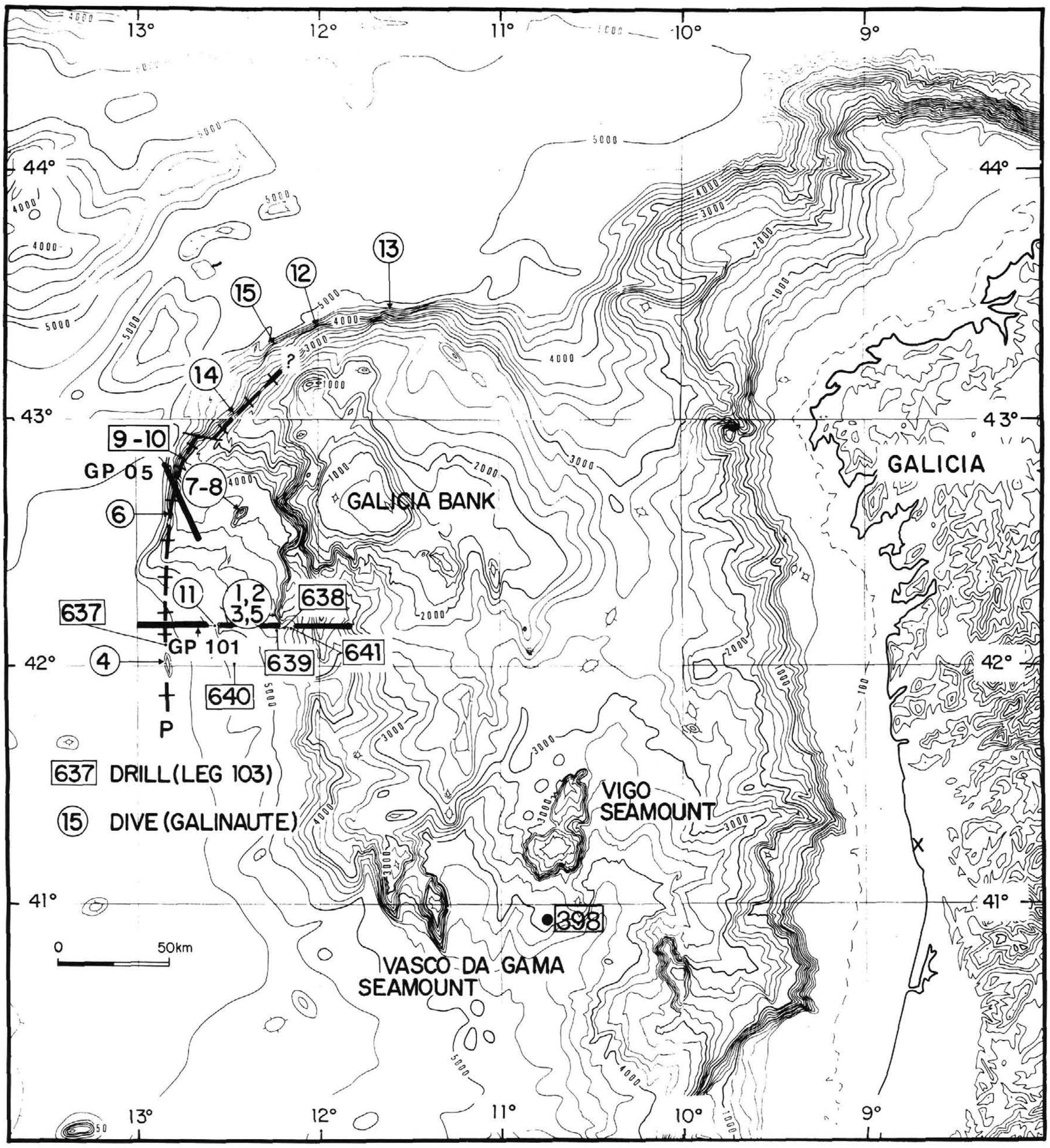

Figure 2. Location of ODP Leg 103 and DSDP Leg 47B drill sites and of *Nautile* dive sites (Boillot et al., this volume) on the Galicia margin. The locations of seismic lines GP-5 (reproduced on Fig. 14) and GP-101 (partly reproduced in Fig. 6) are shown. Bathymetry after Lallemand et al. (1985). Contour interval, 250 m. P = peridotite ridge.

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 is a structural high in the form of a ridge (Fig. 5), which can be followed on seismic-reflection profiles for about 130 km. The data from Site 637 and the results of the dives of the submersible *Nautile* (Boillot et al., this volume) demonstrate that the ridge consists of serpentinized peridotite. In its southern part, the ridge is oriented north-south, parallel to the trend of the tilted fault blocks on the margin (Fig. 10). North of 42.8°N, in the region elevated during Eocene Pyrenean movements (Grimaud et al., 1982), the ridge turns east and follows the northwest edge of Galicia Bank. The ridge crest is 2–3 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.

On the east side, next to the margin, the ridge is partly buried by syn-rift sediments (Mauffret and Montadert, this volume). The ridge was therefore emplaced prior to the beginning of oceanic accretion in this part of the Atlantic, probably at the end of the rifting stage along the margin. The form and dimensions of the ridge, which resemble those of the tilted fault blocks farther east (Mauffret and Montadert, 1987), also suggest that structurally it belongs more to the margin than to the adjacent ocean.

Petrologic and Structural Evolution of Serpentinized Peridotite Drilled at Site 637

The succession of structural and metamorphic events that affected the serpentinized peridotite drilled at Site 637 is compiled in Table 1 from investigations in this volume (Evans and Girardeau; Girardeau et al.; Kornprobst and Tabit; Agrinier et al.;

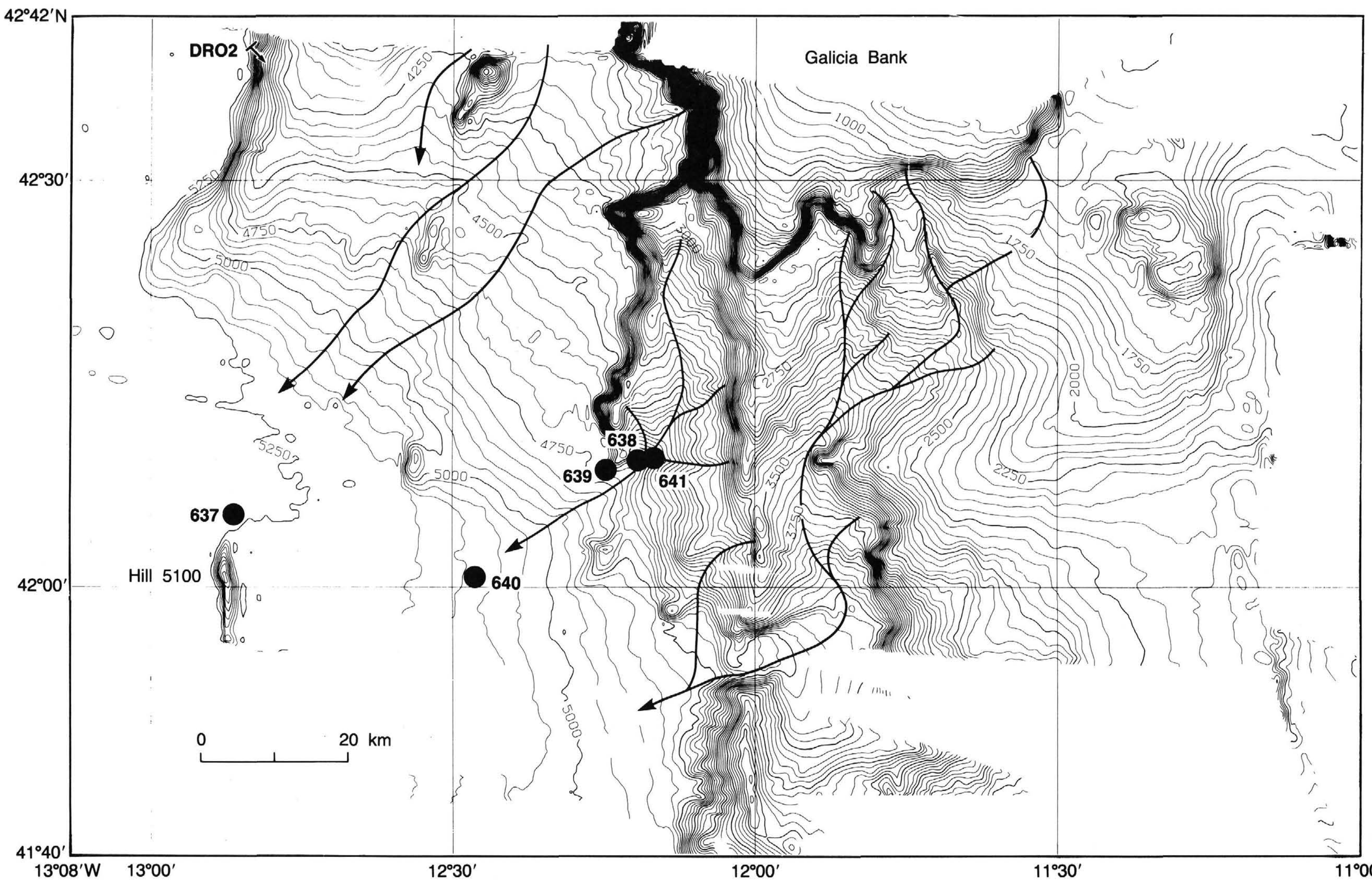

Figure 3. Location of drill sites and detailed bathymetry of the west central part of the Galicia margin, constructed from Sea Beam data collected during Seagal cruise of *Jean Charcot*. After Sibuet et al. (1987). Main canyon trends are shown with bold lines. Depths in uncorrected meters (water velocity, 1500 m/s). Contour interval, 50 m.

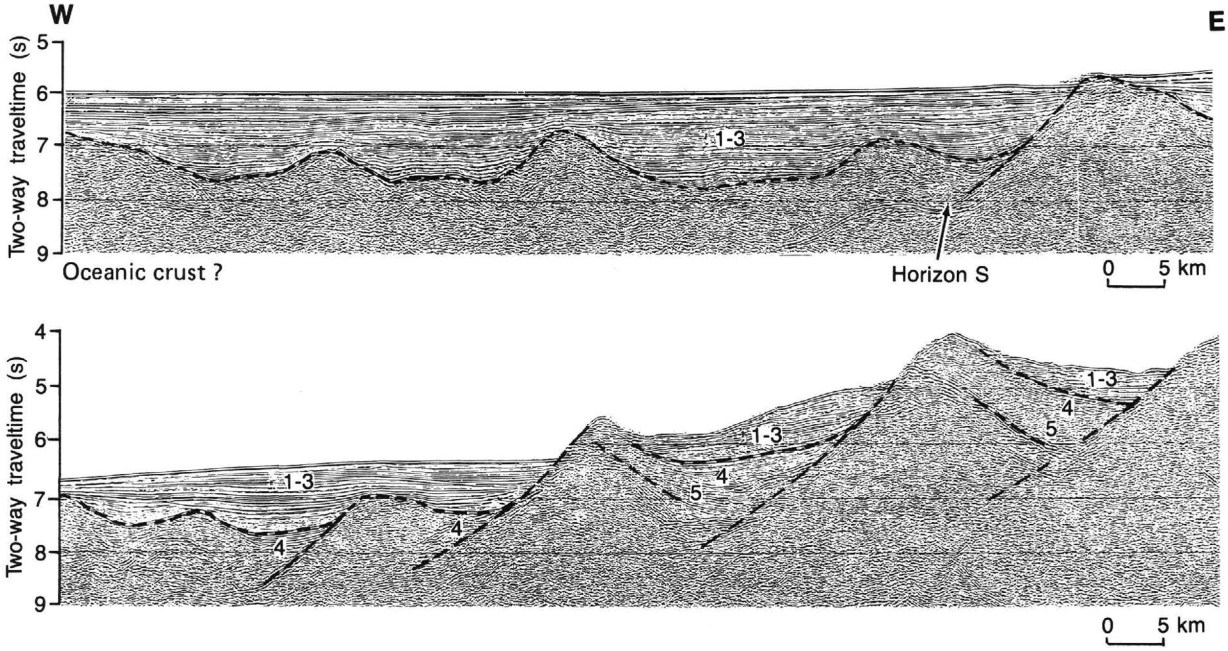

Figure 4. Seismic profile line GP-11 (24-fold, migrated) showing a typical section across the Galicia margin. The lower panel is the eastward continuation of the upper panel. A ridge of peridotite, in the center of the upper panel, separates possible oceanic crust on the west from continental crust, broken into tilted fault blocks, on the east. Seismic Units 1-3 are post-rift sediments; Units 4 and 5 are syn-rift sediments. Horizon S is a prominent deep reflector seen on most profiles beneath the western part of the margin. Vertical exaggeration, 3.3×. Modified from Mauffret and Montadert, 1987.

Kimball and Evans; Evans and Baltuck). At this time we have no evidence to date these events, but their succession clearly shows the rise of the peridotite from a region at a depth of several tens of kilometers in the upper mantle to the Earth's surface, where the rock was serpentinized and fractured.

1. Event 1: Partial melting, which began at temperatures of at least 1200°-1250°C. The partially melted peridotite was part of the asthenosphere. The relatively shallow depth at which melting occurred (30 km or less) shows that the lithosphere was already considerably thinned during this first phase.
2. Event 2: Stretching and mylonitization at temperatures decreasing from about 1000° to 850°C. Plastic deformation took place in a rotational regime. The resulting foliation dips east at 30°-70°, and the lineation is generally oriented down-dip. Analysis of microstructures in the mylonite zones shows the sense of shear to be top-to-the-east. All these data show an oblique movement of the peridotite toward the surface, with a component toward the west.
3. Events 3-5: Crystallization of amphiboles under static conditions, followed by serpentinization and then fracturing and filling of the fractures with calcite. This last set of events took place close to the surface (7-0 km), at temperatures declining from 900° to near 0°C as a result of intense hydrothermal alteration. The final fracturing of the serpentinite is associated with movements along east-dipping normal faults superimposed on the earlier foliation. The kinematics of this last stage of deformation are therefore the same as those during the ductile deformation of event 2.

The uplift and partial melting of the upper mantle and the subsequent deformation of the peridotite and final emplacement of the serpentinite onto the seafloor during the Early Cretaceous are a result, in our opinion, of the horizontal stretching of the lithosphere during the rifting of the margin.

Models

The physical conditions and the petrologic and structural evolution of the drilled rocks thus strongly constrain models for the formation of passive margins.

1. Structural studies (Girardeau et al., this volume) show that the peridotite was deformed by simple shear. This finding does not fit with models invoking homogeneous deformation of the lithosphere, which instead implies pure shear (McKenzie, 1978; Le Pichon and Sibuet, 1981).
2. Petrologic studies (Evans and Girardeau, this volume) show that the peridotite was partly melted (5%-10% of the rock), yet the products of this melting (gabbro and basalt) predicted by models invoking vertical diapirism of the asthenosphere (e.g., Nicolas et al., 1987; Bonatti and Seyler, 1987) were not found in the Site 637 cores or in dredge samples. On the other hand, it should be kept in mind that we do not know what volume of rock was partially melted, and thus, whether it was sufficient to form dikes or a magma chamber or if it was merely trapped interstitially or in small bodies in the peridotite. Nor do we know the original geometry of the rising mantle material, which would indicate the most logical location of bodies of coalesced melt.

To account for the known facts, the Leg 103 scientific team proposed a model (Boillot et al., 1987), 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 sense of movement for both the ductile and brittle deformation and their orientation as given by remanent magnetism in core samples suggest that the detachment fault dips east, toward the continent.

This preliminary model was recently elaborated by entailing the partial melting of the peridotite at shallow depths in an

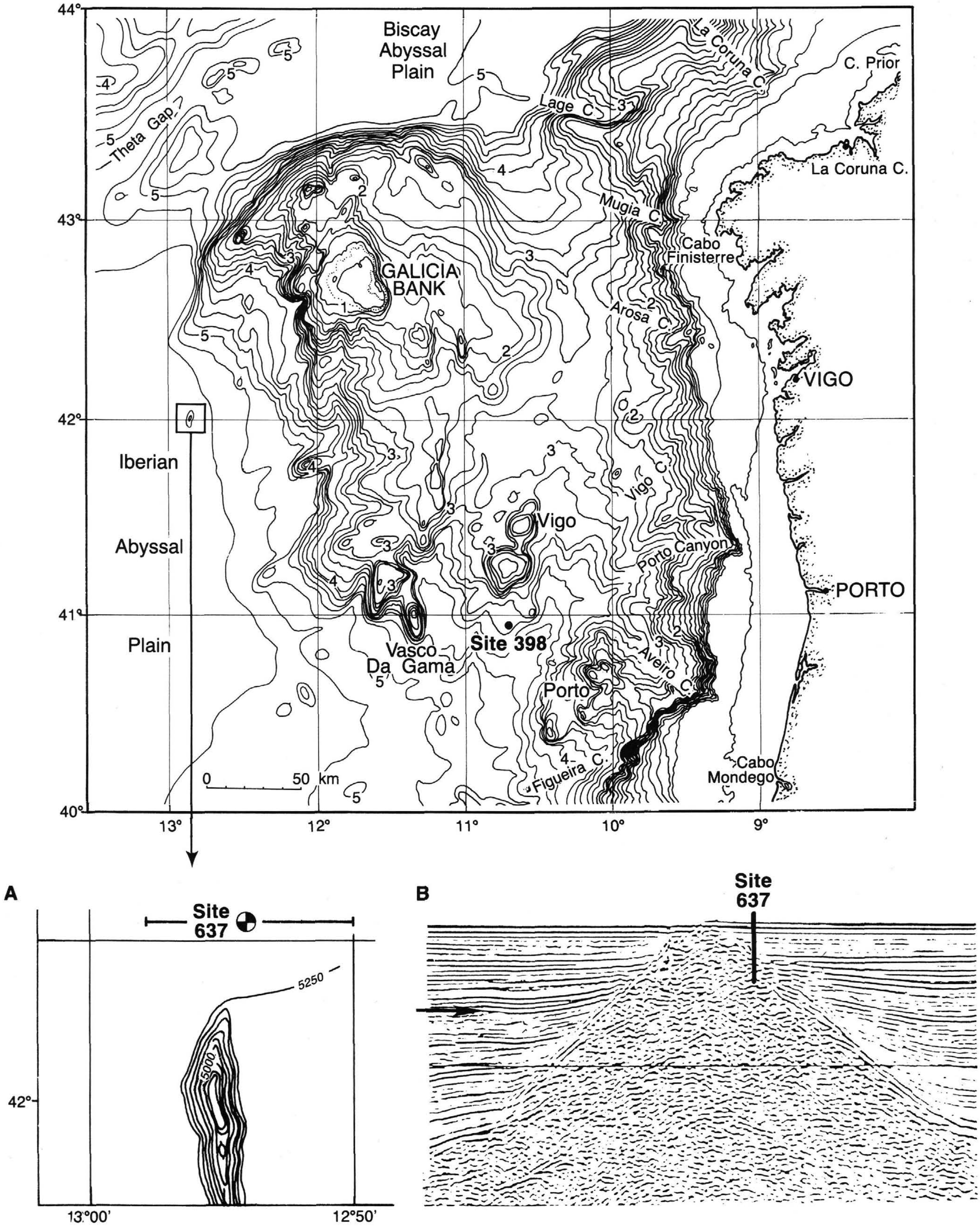

Figure 5. **A.** Location of Site 637 on Sea Beam map, showing relation to peridotite ridge. **B.** Multichannel seismic line GP-12 across Site 637. Courtesy of L. Montadert.

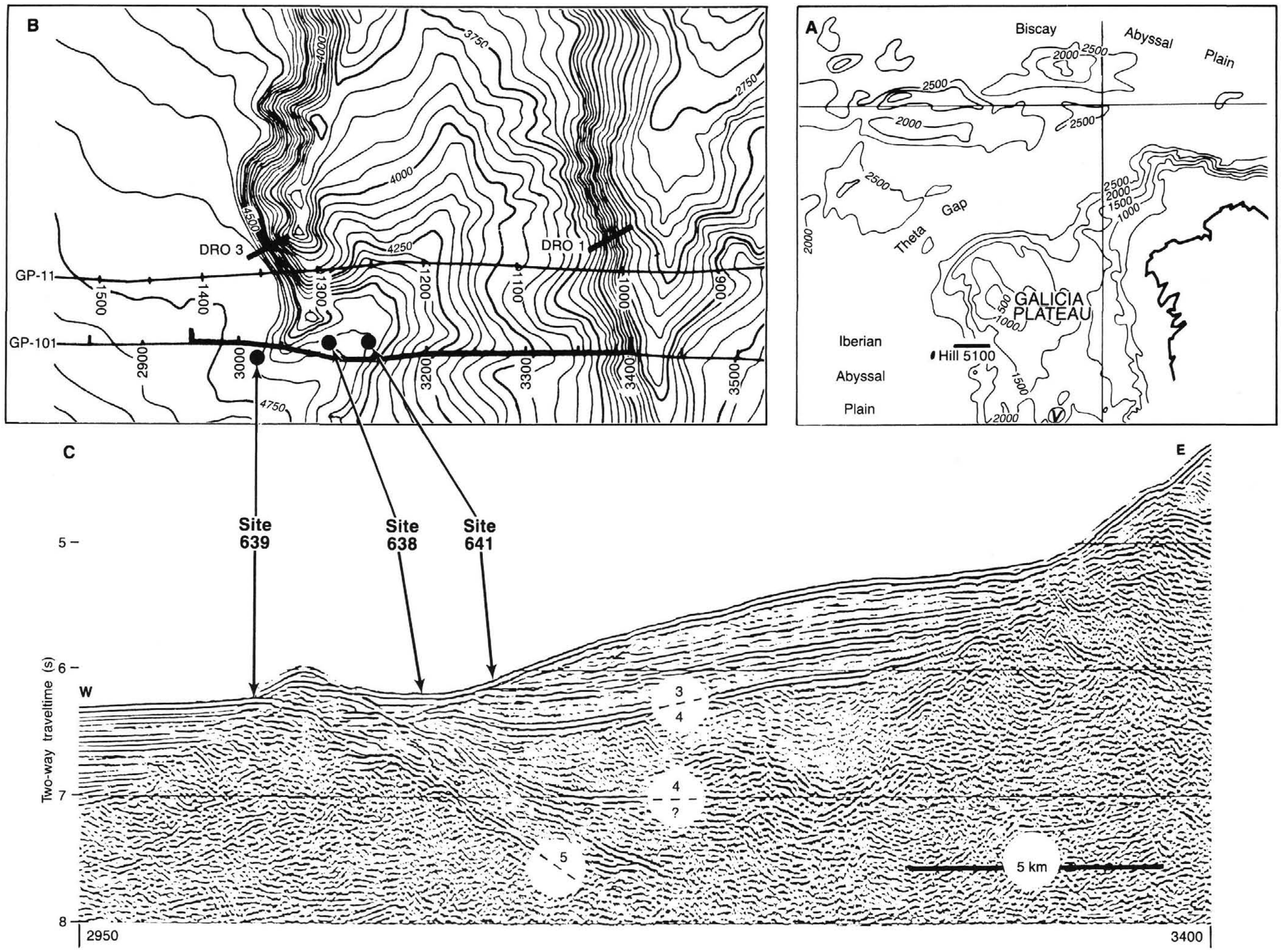

Figure 6. **A.** Galicia margin, with heavy line showing location of part of seismic profile GP-101 depicted in Figure 6C. **B.** Sea Beam map of the area near Sites 638, 639, and 641, showing location of dredge localities DR01 and DR03 (Mougenot et al., 1985). After Sibuet et al. (1987). **C.** Multichannel seismic profile GP-101. Seismic Unit 3 = post-rift sediments; Units 4 and 5 = syn-rift strata. Courtesy of L. Montadert.

early stage of margin formation (Fig. 11) (Boillot et al., this volume). The initial stage of the emplacement is the rise of an asthenospheric diapir under continental crust of normal thickness. The products of partial melting then accumulate along the summit of the diapir (Bonatti and Seyler, 1987). In a second stage, the gabbros and some of the peridotite are stripped off tectonically by movements along a detachment fault, thus, these rocks 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.

Although this model accounts for the currently available data along the Galicia margin, it cannot be accepted as proven until the detachment fault itself has been identified by geological/geophysical studies or by drilling.

An alternate model, favored by one of us (ELW), proposes that the sense of shear indicated by the structural studies of the core samples does not necessarily prove an eastward dip of the regional master detachment fault; instead, the top-to-the-east shearing may be the result of more localized shear deformation along the east flank of the rising diapiric ridge of peridotite. During diapiric rise, the "missing" gabbro and basalt are sheared away to a (buried) flank position(s); it is thus not surprising that these rocks remain undiscovered. In this model, the regional detachment surface dips west, parallel to and perhaps merging with the S reflector (Fig. 12). This orientation fits with the known contrasts in fault block geometry between the Newfoundland and Galicia margins (Tankard and Welsink, 1987; Grant et al., this volume; Winterer et al., this volume).

Seismic Horizon S (GB)

Horizon S is a strong reflector or bundle of reflectors that appears within or beneath continental basement in the deepest parts of the Galicia margin (de Charpal et al., 1978; Montadert et al., 1979; Boillot et al., 1980; Mauffret and Montadert, 1987). This reflector has been interpreted in different ways:

1. According to de Charpal et al. (1978), it is at the brittle/ductile boundary within the continental crust, which was stretched and thinned during rifting of the margin.
2. According to Wernicke and Burchfield (1982), horizon S is the trace of the intracrustal detachment fault that was active during the rifting of the margin. This is essentially the interpretation of Winterer et al. (this volume), as shown in Figure 12.
3. According to Boillot et al. (this volume), horizon S is the trace of the detachment fault for Early Cretaceous rifting. However, the fault is not consistently intracrustal, because at some places it puts the thinned continental crust of the margin in contact with the serpentine resulting from hydrothermal alteration of mantle peridotite (Figs. 11 and 13B).

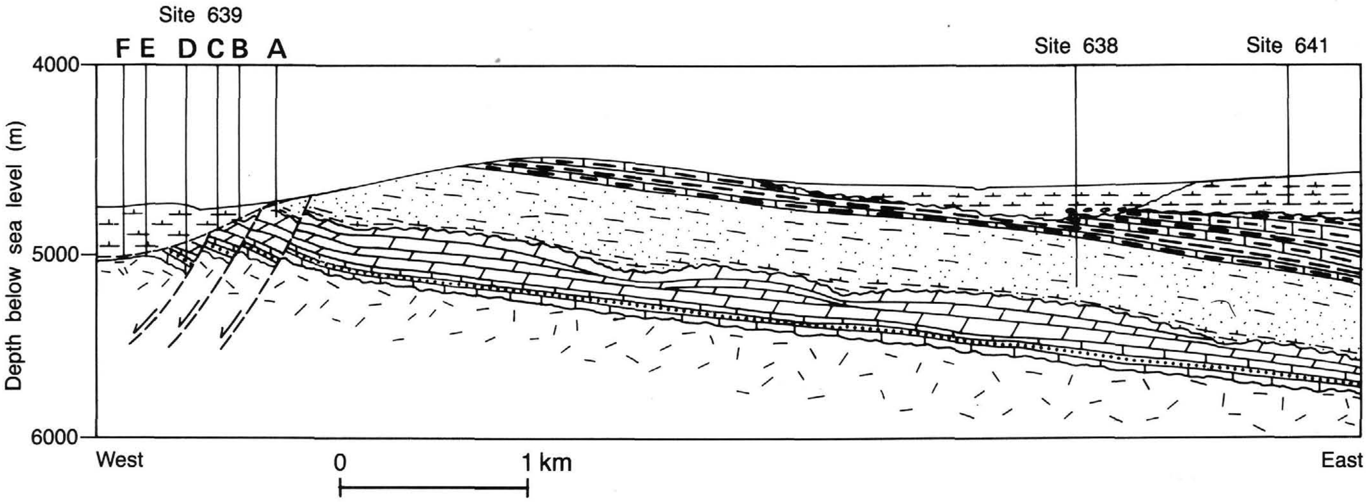

Figure 7. Cross section through Sites 639, 638, and 641, as interpreted from drilling and seismic data. The faulting shown at Site 639 is schematic, and the irregular upper surface of the dolomite can also be interpreted as resulting from offsets by normal faults.

An argument in support of this last interpretation comes from study of seismic profile GP-05, northwest of Galicia Bank (Figs. 2 and 14), on which there is a strong reflector at 2.5 s twt beneath the seafloor. This reflector has many of the characteristics defined for horizon S in the region farther south, around Site 640. It is overlain by a relatively transparent layer quite comparable to the layer overlying horizon S at the place where it was first described. The major uncertainty in correlating the deep reflector in the northern area with horizon S farther south is the lack of a connecting seismic line which would make an unambiguous connection between the two areas possible.

The deep reflector on seimsic line GP-05 (Fig. 14) shallows toward the northwest and intersects the seafloor just above the peridotite ridge identified there by the *Nautile* dives (Boillot et al., this volume). At a somewhat shallower depth, about 30 km to the north, on the slopes of Galicia Bank, dredging recovered *in situ* a sample of granodiorite (Capdevila and Mougenot, this volume). The deep reflector seen on profile GP-05, tentatively correlated with horizon S, could therefore separate continental crust from the serpentine observed on the dives. If this interpretation is correct, the region around profile GP-05 presents an exceptional opportunity to sample across the continent/serpentine contact. In the model shown in Figure 11, this contact is the detachment fault that juxtaposed mantle and continental crust during rifting.

Pre-Valanginian Sediments of the Deep Margin (ELW)

Stratigraphy of Pre-Valanginian Strata

The stratigraphy—lithology, thickness, and age—of the sedimentary strata on the Galicia margin older than the Valanginian calpionellid marlstone cored at Hole 639A is ambiguous, yet a correct knowledge of this sequence is prerequisite to a sound interpretation of the regional history of crustal thinning and rifting. At Site 639, the data from the six holes drilled as a transect can be assembled in many plausible arrangements, and when the data from nearby *Nautile* dives (Boillot et al., this volume) are taken into account, the admissible cross sections and stratigraphic columns grow in number.

The simplest possible way to reconstruct the section at Site 639 is to assume a simple homoclinal sequence, dipping east, as shown in Figure 15. The reconstructed sequence has many large gaps. Although the regional dip, from the seismic data, is about 10° (Figs. 7 and 16), indications in some of the cores from Site 639 suggest the local dip may be close to 30°. The sequence, which was only partly sampled by drilling, begins with basal fragments of hydrothermally altered rhyolite (Evans, this volume), which may be clasts in a conglomerate or breccia or perhaps *in-situ* crystalline basement, associated with beds (or pebbles) of quartzite, claystone, and metaconglomerate. Another layer of rhyolite (pebbles?) occurs about 60 m above the lower rhyolite pieces and is overlain by fractured, vuggy dolomite, followed by about 90 m of upper Tithonian shallow-water limestone interbedded with claystone and sandstone (Moullade et al., this volume; Jansa et al., this volume). Above this is a 400-m-thick section of beds that was unsampled except for three intervals of fractured dolomite. The entire inferred sequence has a thickness of about 550 m, assuming a 30° dip.

An important question is whether the fragments of rhyolite and of metasedimentary rocks are clasts in unmetamorphosed Mesozoic sediments or are part of *in-situ* crystalline basement, of Paleozoic or Precambrian age. The various igneous and metamorphic rock fragments were recovered in mixed order in Holes 639E and 639F, and although many of the pieces are small enough to have moved past one another during the coring process, others are too large, and we must admit that they are of an original mixed order. Nor do any of the fragments consist of more than one distinct rock type; no cored contact is preserved. These facts suggest that the crystalline rock fragments come from conglomerate or breccia layers rather than from *in-situ* basement. If the crystalline rocks are not basement, then an unknown thickness of strata may lie beneath the Tithonian strata sampled by drilling. Regionally, there are thick sequences of pre-Tithonian sediments on adjacent and conjugate parts of the Atlantic margin (Montenat et al., this volume; Réhault and Mauffret, 1979; Winterer et al., this volume; Grant et al., this volume; Meador and Austin, this volume).

A precisely controlled set of samples from only about 6 km north of Site 639, from the same escarpment, comes from the post-Leg 103 *Nautile* dives. If we compare the simplest possible stratigraphic column for Site 639 with the simple unfaulted column deduced for the site of the *Nautile* dives (Boillot et al., this volume) (Fig. 17), there are both similarities and differences. Along the submersible traverse, the exposed thickness of the pre-Valanginian sequence is about 850 m and consists of interbedded hydrothermally altered dacitic volcanoclastic sandstone, graywacke, subgraywacke, and dolomite containing ghosts of echinoderms and bivalves. The uppermost unit, about 30 m

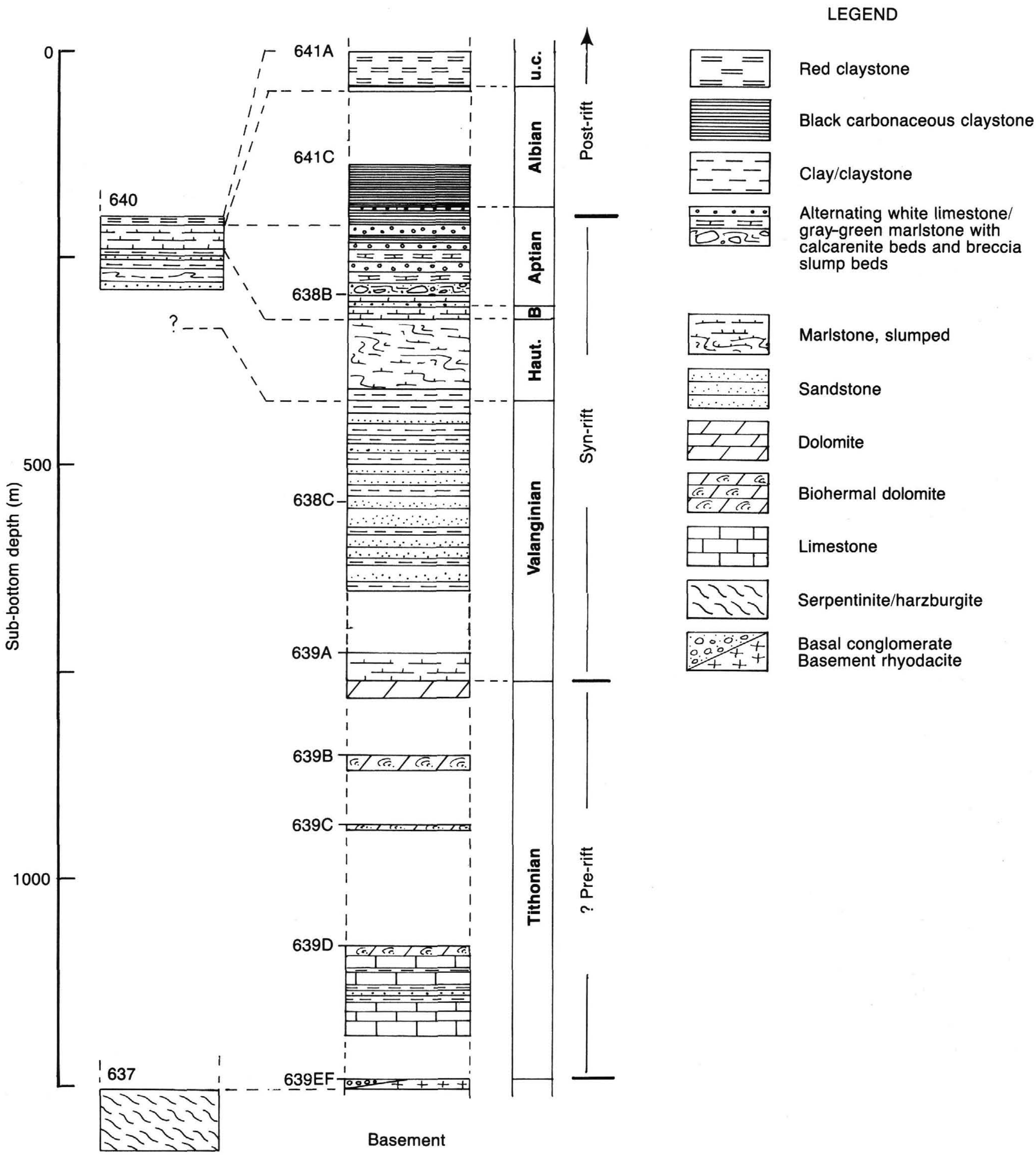

Figure 8. Composite stratigraphic column of Mesozoic strata at Sites 638, 639, and 641 (main column) and at Sites 637 and 640 (sections at left). u.c. = Upper Cretaceous; B = Barremian.

thick, is a fractured vuggy dolomite identical to that at the top of the pre-Valanginian section at Site 639. Boillot et al. (this volume) assigned it to the Tithonian by analogy with the similarly positioned dolomites at Site 639. A notable difference is that no fossiliferous Tithonian limestone like that at Hole 639D was sampled by the *Nautile*, but because the submersible traversed thick (100-m) intervals with no outcrop, it is not safe to assume that this limestone is absent.

Another traverse by *Nautile* across an escarpment about 25 km west of Site 639 (Boillot et al., this volume) recovered several samples of coarse-grained sandstone cropping out from 100 to 300 m above *in-situ* granodiorite basement (assuming no faults). Whether this sandstone correlates with the Valanginian sequence cored at Site 638 or whether it is part of the pre-Tithonian section revealed at the dive site close to Site 639 is not established. If it is Valanginian, then a complicated pre-Valanginian geology

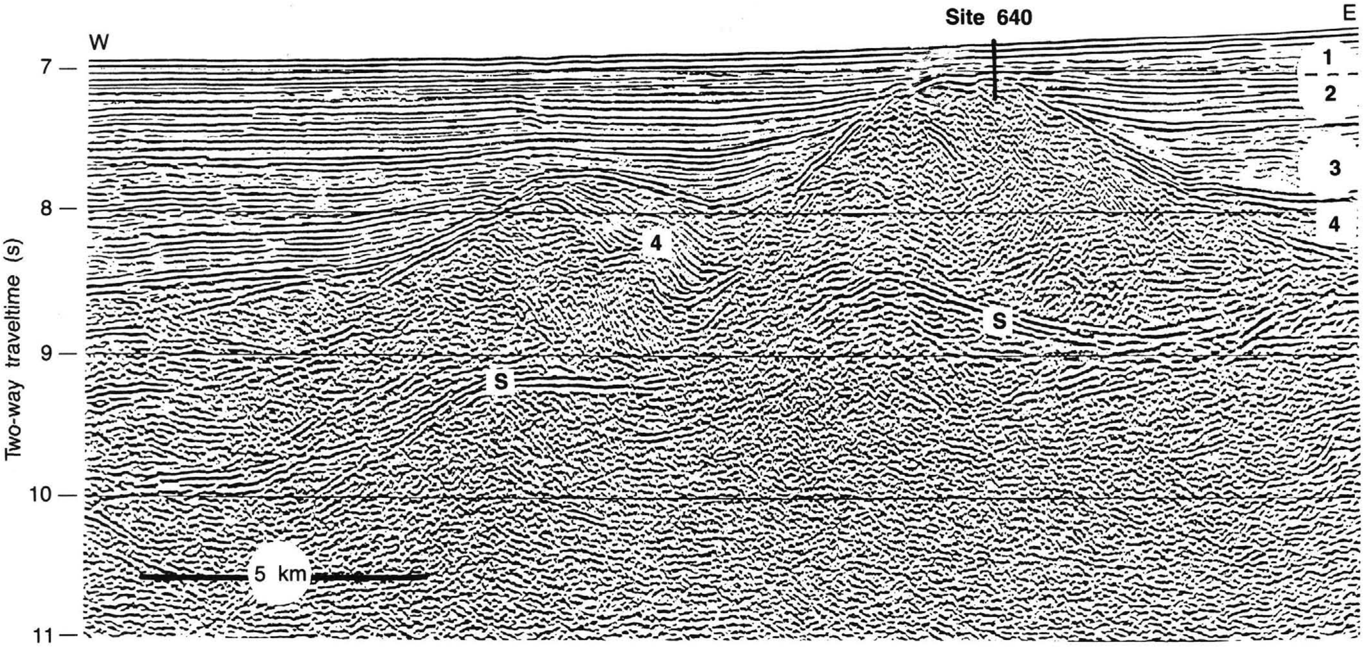

Figure 9. Multichannel seismic line GP-102B1, showing location of Site 640 and the S reflector. Seismic Units 1-3 = post-rift sediments; Unit 4 = syn-rift sediments. Courtesy of L. Montadert.

is implied, with an incomplete cover of Tithonian strata on basement; if it is related to the sequence at the other dive site, it would plausibly fit below the lowest strata shown in Figure 17.

Possible Faulting Near Site 639

More complicated cross sections at Site 639, involving faulting or rock slides, result if the aim is to simplify the stratigraphic sequence, for example, by having all of the brecciated dolomite lie stratigraphically above the Tithonian limestone. The cross section shown in Figure 18 does this by assigning the brecciated dolomite west of limestone (and hence possibly underlying it) to a rock slide, which is extended to include the upper brecciated limestone and dolomite at Hole 639D. Given the similar fractured nature of the dolomite recovered at Site 639, the rock slide hypothesis could be extended to relegate all of the Site 639 dolomite to slide masses.

A different type of complication enters with the introduction of faulting to the cross sections. The aims here are (1) to explain the abnormally steep dips measured in marlstone cores at Hole 639A and in limestone at Hole 639D and (2) to minimize the thickness (>220 m) of the upper fractured dolomite implied in the no-fault schemes. The fact that Holes 639A, 639B, and 639C were arbitrarily sited and yet recovered nothing but fractured dolomite below the Valanginian suggests that this type of dolomite may be common in the upper part of the slope at Site 639. In contrast, little dolomite was recovered in dredges from the escarpments that expose Upper Jurassic carbonates near Site 639 (Mougenot et al., 1985). Even in the *Nautile* dives, the observed thickness of the upper fractured dolomite is only a few tens of meters. Unless the dolomites are actually in the form of a local thick lens at Site 639, then repetition by normal faulting is required. This is the scheme shown in Figure 7, which was also adopted with minor modifications by Jansa et al. (this volume). The faults can be drawn either as steep faults, as in Figures 7 and 13B, or as very gently dipping faults, nearly parallel to the buried fault-line escarpment that connects Site 639 to the adjacent half-graben to the west (Fig. 13A).

Pre-Valanginian Tectonics

Whether there is a substantial thickness of pre-Valanginian strata on the west part of the Galicia margin is important in reconstructing the tectonic history. The carbonate and clastic rocks sampled at Site 639 are "pre-rift" with respect to the overlying Valanginian-Aptian sediments, which were laid down in subsiding half-grabens bounded by episodically active faults. Nevertheless, it is possible that the pre-Valanginian beds are related to a previous cycle of rifting. In many other areas in this part of the North Atlantic, rifting began much earlier than the Valanginian. A Late Triassic–Liassic episode is recorded in the Grand Banks region (Grant et al., this volume), in the Lusitanian Basin (Montenat et al., this volume), and probably in the Interior Basin between Galicia Bank and the mainland of Iberia (Boillot et al., 1979). A second phase of rifting, during Oxfordian and Kimmeridgian time, is well documented in the Lusitanian Basin (Montenat et al., this volume). In the Jeanne d'Arc Basin on the Grand Banks, renewed strong subsidence and clastic sedimentation, suggesting tectonism, began in the Kimmeridgian, but this event is not well expressed in the more seaward Carson Basin, which contains about 700 m of Kimmeridgian and Tithonian interbedded mudstone and limestone (Grant et al., this volume).

Given the regional evidence for pre-Cretaceous rifting, the coarse pre-Tithonian clastics recovered by the *Nautile*, and possibly from Holes 639E and 639F, may be the result of a Late Jurassic or older rifting episode.

Origin of Fractured Dolomite at Site 639

The fractured dolomite at Site 639 presents a host of unresolved problems, both as to the environment of dolomitization and the origin of the intensive fracturing. The problems have strong implications for the tectonic and subsidence history of the margin. While there is general agreement among those who have studied these rocks in this volume (Loreau and Cros; Haggerty and Smith; Jansa et al.) concerning most of the descriptive data, there is some disagreement about interpretation.

Everyone agrees on the following:

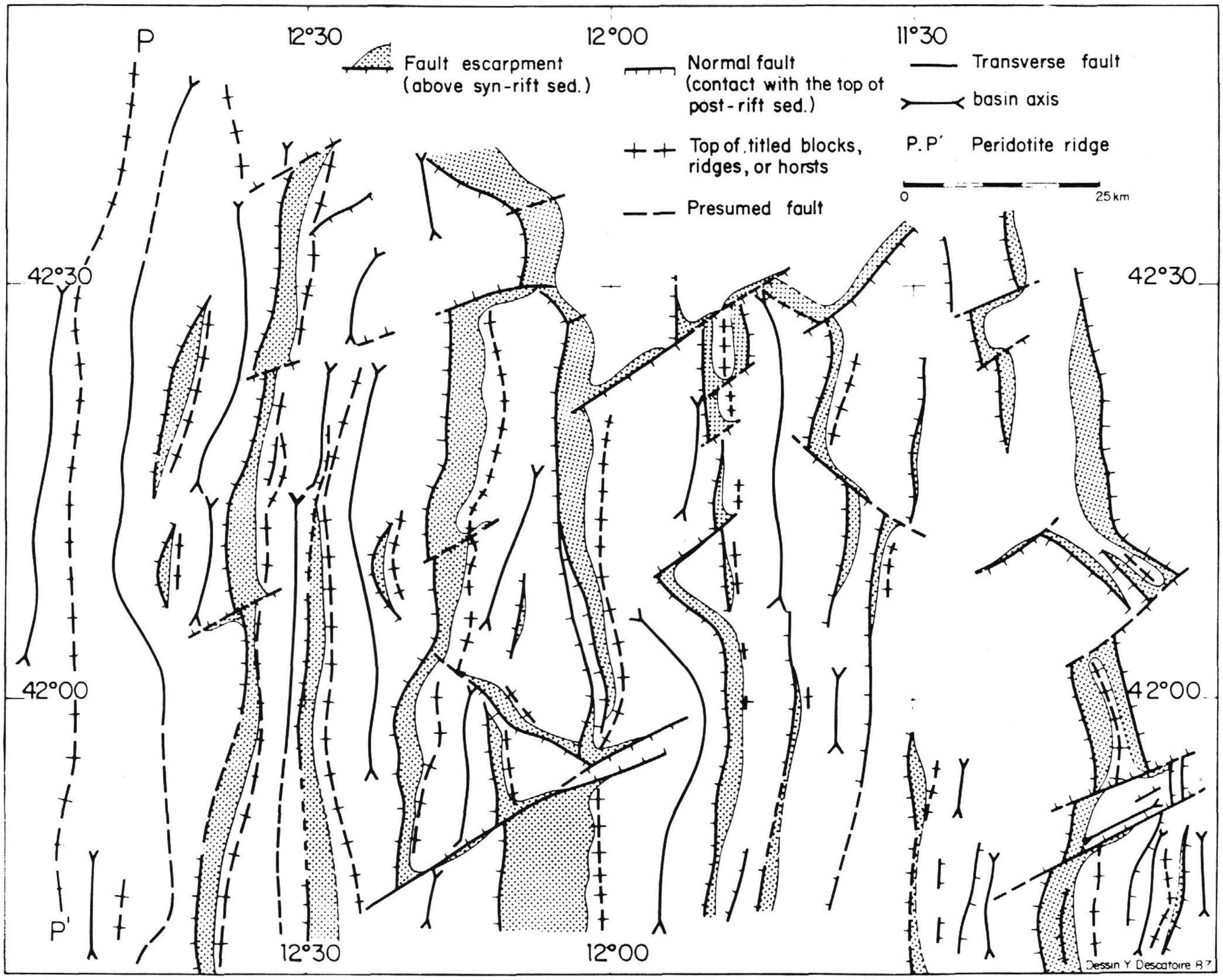

Figure 10. Structural map of the western part of the Galicia margin. P-P = peridotite ridge. After Thommeret et al. (this volume).

1. The rocks were originally limestone, deposited in a shallow-water marine environment.
2. Some dolomitization began before fracturing.
3. A long history of progressive fracturing and dolomitization ensued, with the dolomite both replacing limestone and precipitating in open fractures. Late-stage internal sediments of dolomite partly fill some fractures and vugs.
4. The fracturing is at least partly related to extensional tectonics during dolomitization.

Dolomitization

There is some disagreement about the environment and mechanisms of dolomitization:

1. Loreau and Cros (this volume) conclude that the first stage of dolomitization took place under the influence of fresh water in the "mixing zone". They cited replacement of aragonite bioclasts by sparite, dissolution linked to dedolomitization, and negative $\delta^{18}O$ values in support. Subsequent dolomitization resulting in saddle dolomite that occurs in the upper section of the Tithonian is tentatively explained by hydrothermalism related to fracturing. Emersion is also postulated to account for the vadose filling of cavities and fractures by silt.

2. Jansa et al. (with Haggerty demurring; this volume) conclude that dolomitization occurred in mixed waters, at the base of a freshwater lens developed on an emergent platform, but, based on work by Haggerty and Smith, they accept the possibility of dolomitization by hypersaline brines migrating from an evaporitic environment created by emergence of the platform. They also mention the possibility of hydrothermal hypersaline brines, moving up along faults, as agents of dolomitization in burial conditions. The fracturing and brecciation are attributed to tectonism and possibly to collapse resulting from dissolution of evaporites, evidenced by cavities now floored by baroque dolomite.

3. Haggerty and Smith (this volume) and Daniel and Haggerty (this volume) conclude from fluid inclusion, dolomite pyrolysis, trace element, and stable isotope data that initial dolomitization was from warm hypersaline brines derived by lateral and/or downward migration from a restricted basin and explain their very light oxygen isotope signatures as the result of somewhat elevated water temperatures and of an "evaporative loop" in which oxygen isotopic values become progressively more negative as seawater is evaporated. They calculated from their isotopic and pyrolysis data that the dolomitizing solutions had concentrations about 45 times that of normal seawater. They

Table 1. Succession of structural and metamorphic events experienced by the Galicia margin peridotite, from studies in this volume by Evans and Girardeau, Girardeau et al., Kornprobst and Tabit, Agrinier et al., Kimball and Evans, and Evans and Baltuck.

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

measured fairly high concentrations of strontium. The hydrothermal model for initial dolomitization was rejected by Haggerty and Smith (this volume) because the underlying limestones are not dolomitized and because the chemistry of the dolomites shows that they formed from oxidized solutions rather than reduced solutions that are more typical of hydrothermal brines.

In our opinion, the hydrothermal model provides an explanation for the hypersaline dolomitizing fluids, derived from interstitial brines deep below the seafloor, and for the relatively high temperatures, which suggest derivation (ignoring any mixing with cool seawater) from depths of at least 1 or 2 km, in light of the possible high temperature gradients on a thin, actively extending margin. The model also accounts for the closely linked timing of the dolomitization and fracturing. It is possible that hydrothermal activity accounts for the authigenic chloritic matrix reported in some of the sandstone samples collected during the *Nautile* dives (Boillot et al., this volume) and for the alteration of the rhyolite fragments at Site 639 (Evans, this volume). Nevertheless, the hydrothermal model does not account for the occurrence of limestone in close association with the dolomite.

Fracturing

The intensive fracturing of the dolomite is most plausibly associated with movements on nearby faults. The dolomite was the least buried formation in the sequence and, if made brittle by early cementation and partial dolomitization, would have been most subject to fracturing during faulting. The underlying

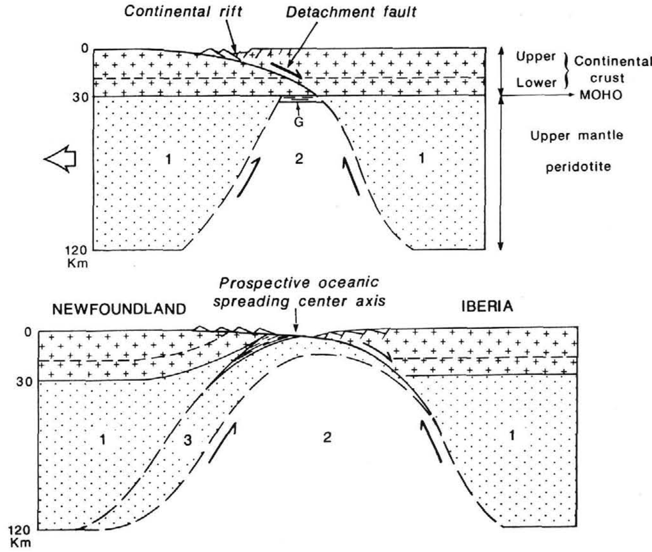

Figure 11. A composite model for the thinning of the continental crust and the emplacement of peridotite at the seafloor along the Galicia margin. 1 = subcrustal peridotite; 2 = peridotite in bulge that is partially melted; 3 = cooled, depleted peridotite. **A.** Bulge of asthenosphere, resulting in small amount of partial melting of peridotite and cyrstallization of gabbro (G). This stage is derived from Bonatti and Seyler's (1987) model. **B.** Tectonic denudation of gabbro along the left margin and of cooled, slightly depleted peridotite along the right margin (Galicia). This stage follows the model of Boillot et al. (1987). The next stage would be the beginning of normal ocean spreading.

limestone is also somewhat fractured, in some places intensively (for details see "Site 639" chapter, Boillot, Winterer, et al., 1987; Loreau and Cros, this volume), but it is also somewhat clayey, in contrast to the virtually clay-free dolomite, and may have responded to stress without extensive shattering.

As suggested in the "Site 639" chapter (Boillot, Winterer, et al., 1987), some of the breccia may be part of a pre-Valanginian gravitational slide of previously fractured and dolomitized rock (Fig. 18). The occurrence of a few meters of jumbled pieces of limestone, marlstone, and sandstone above the dolomite breccia in Hole 639E suggests downslope transport.

Possible Emergence and Drowning of the Carbonate Platform

Several papers in this volume (Loreau and Cros; Haggerty and Smith; Jansa et al.) invoke an emergence to allow erosional features and diagenesis by meteoric or mixed waters or by hypersaline waters in a nearby lagoon formed by the emergent edge of the faulted carbonate platform. On the other hand, the immediately overlying Valanginian calpionellid marls (Figs. 8 and 18) indicate a water depth of several hundred meters (Moullade et al., this volume). The actual contact was not recovered, but any missing interval cannot be more than 7 m thick ("Site 639" chapter, Boillot, Winterer, et al., 1987). We infer that the final drowning was rapid, too rapid in fact for shallow-water organisms to keep up, and without any trace of a transitional facies.

The proposed emergence interrupts the normal subsidence of the margin, expected during crustal thinning, and requires either a eustatic drop in sea level (Loreau and Cros, this volume) or tectonic upward tilting at Site 639 or a combination of the two (Jansa et al., this volume). The timing of the hypothesized eustatic drop is constrained no more precisely than to post-middle Tithonian and pre-late Valanginian. The possibility of tectonic tilting is consistent with the history of progressive listric faulting and tilting revealed in the analysis of seismic-reflection

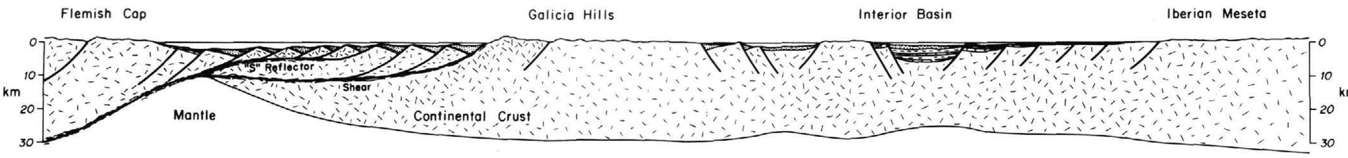

Figure 12. Schematic cross section across the Galicia margin at about 42°10′N during the Valanginian, prior to the inception of seafloor spreading between this margin and Newfoundland. Horizon S is shown as a detachment fault dipping west. The Galicia Bank region is shown as elevated by isostatic rebound from the unloading by detachment faulting to the west. No vertical exaggeration. From Winterer et al. (this volume).

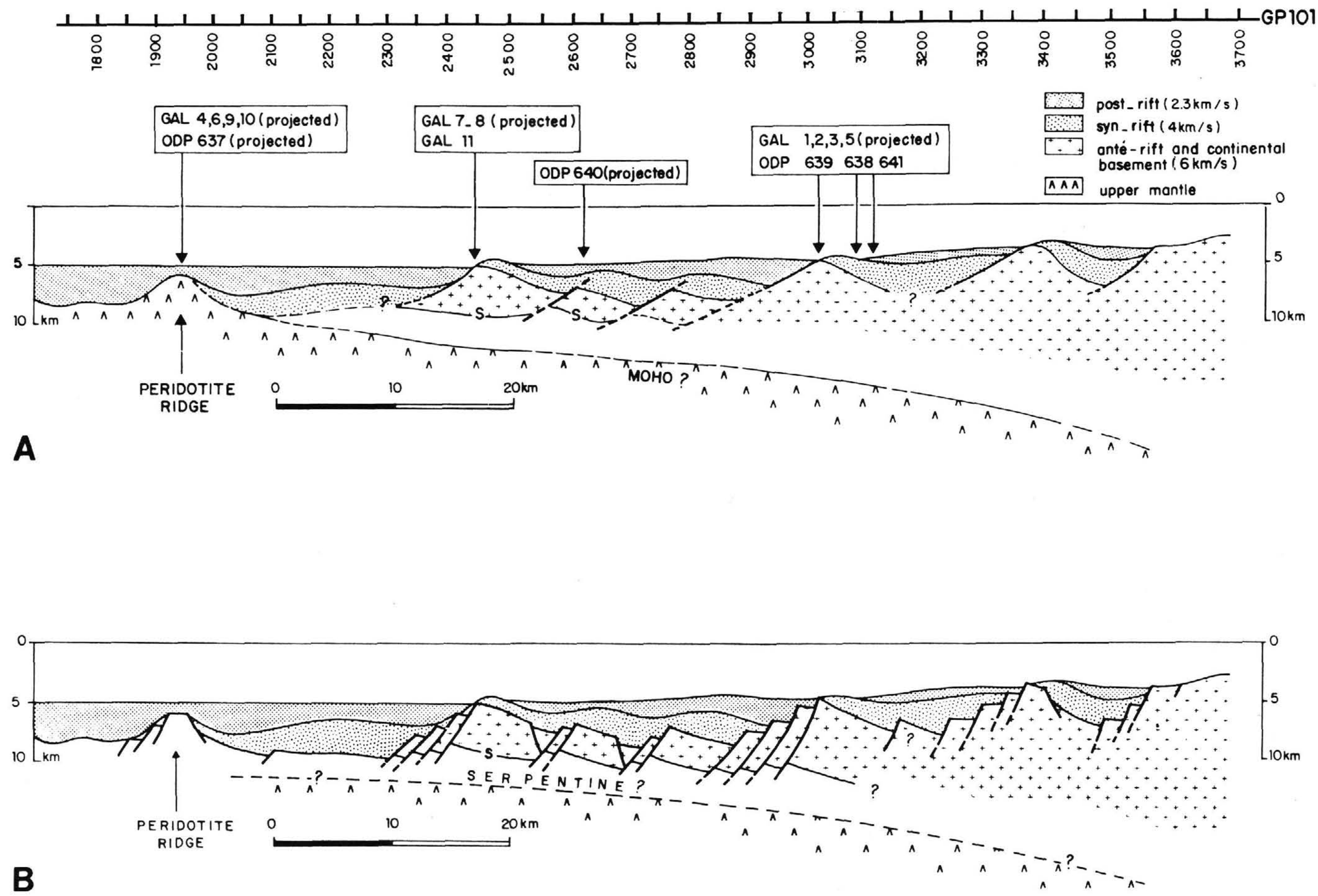

Figure 13. Interpretive cross sections of the Galicia margin, along seismic line GP-101 (see Fig. 2 for location). **A.** Interpreted as cut by low-angle listric faults. **B.** Interpreted as cut by high-angle normal faults, as discussed in Boillot et al. (this volume). The Moho is positioned assuming isostatic equilibrium, using densities of 2.8 and 3.3 g/cm^3 for thinned continental crust and mantle, respectively. S = S seismic reflector.

profile data near Site 639 (Boillot et al., 1986). Whether this inferred tilting also resulted in actual uplift of the front edge of the tilted block is not revealed in the analysis. Wernicke's (1985) analysis of the isostatic effects of crustal thinning by low-angle listric faulting showed that isostatic rebound of the footwall block after unloading by low-angle faulting can produce real uplift. Alternatively, thermal uplift of the hanging-wall block during simple-shear extension of the lithosphere is predicted in the model of Buck et al. (1988).

The next stage of the evolution implies increased slip rates along extensional faults following most of the dolomitization (the calpionellid marlstone is slightly dolomitized). The platform quickly drowned to depths below the zone where benthic organisms could accumulate apace with the relative subsidence rates, into a realm where sedimentation of hemipelagic marl then began.

Syn-Rift Sediments and Tectonics (ELW)

Valanginian-Hauterivian Turbidites: Tectonics and Provenance

The marl grades upward, about 20 m above its base, into alternating marl, claystone, and sandy turbidites, which herald the beginning of a flood of clastics (Fig. 8). The 250-m-thick upper Valanginian sandy turbidite section sampled at Sites 638 and 639 is part of wedge that thickens eastward into a half-graben typical of those on the Galicia margin (Fig. 6C). Active tectonism is indicated not only by the great volume of these coarse sediments in the region seaward of Galicia Bank, estimated to be about 7000 km^3, with an average thickness of about 600 m (Winterer et al., this volume), but also by the geometry seen on seismic profiles. The long history of episodic faulting and rotation inferred from analysis of the profiles (Boillot et al., this volume) is accompanied by rapid subsidence, documented by

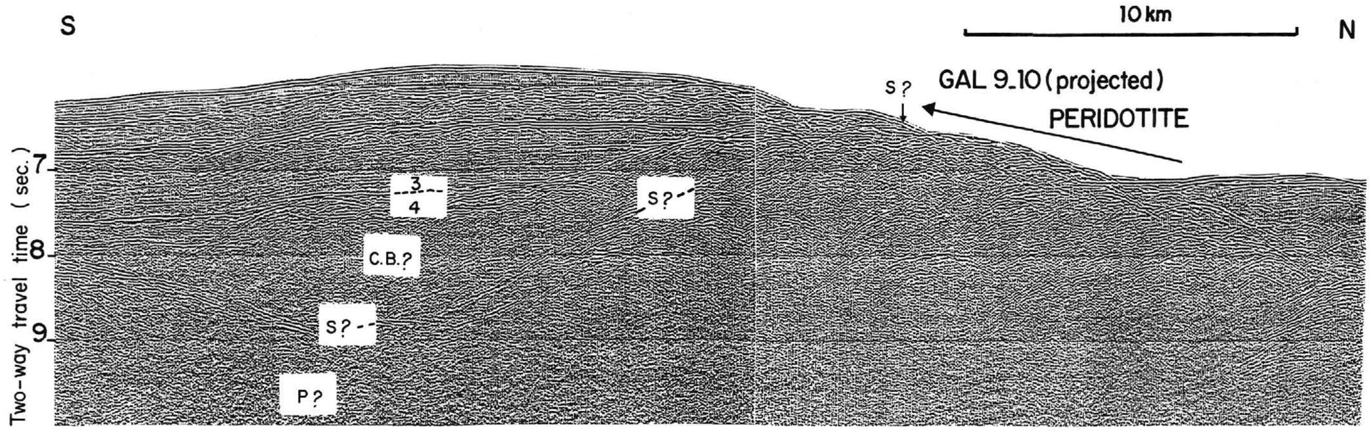

Figure 14. Seismic profile GP-05, recorded on a line across the northwest slope of Galicia Bank (see Fig. 2 for location). A strong reflector, labeled S, rises northwestward from about 2.5 s twt below the seafloor and crops out at the position shown by a small arrow. The escarpment to the north of the outcrop of the reflector exposes serpentinite, sampled by the *Nautile* (Boillot et al., this volume). Profile courtesy of L. Montadert.

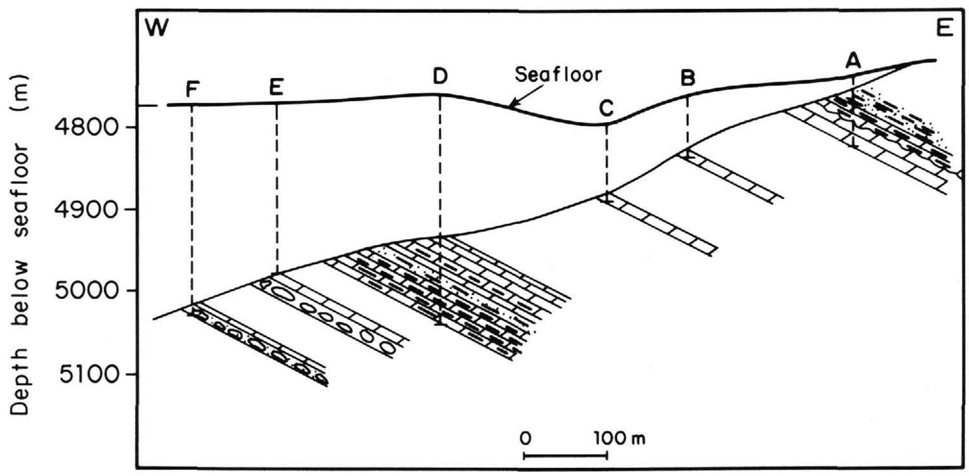

Figure 15. Cross section through Site 639, interpreted as having no faults or rock slides, and treating the rhyolite and crystalline rocks in Holes 639E and 639F as clasts in conglomerate or breccia beds. Compare with Figures 7 and 18.

changes in the benthic biotic content of the sediments (Moullade et al., this volume). This history is consistent with models of episodic crustal thinning.

Analysis of the detrital mineralogy of the sandy turbidites (Johnson, this volume; Winterer et al., this volume) and of seismic-reflection profiles across the region (Winterer et al., this volume) point toward Galicia Bank and its southern extension as the likely source area and furthermore suggest active uplift of the source area.

Hauterivian-Aptian Tectonics, Sedimentation, and Diagenesis

Following the rapid (about 50 m/m.y.) deposition of the Valanginian sandy turbidite sequence, accumulation rates slowed gradually through the Hauterivian, Barremian, and Aptian, probably in response to gradual lowering of the source area of the sandy sediments and perhaps to diminishing slip rates on the faults that bound the half-grabens in which the sediments accumulated. Seismic profiles (Figs. 4 and 13) show that these sediments thicken markedly downdip into the half-grabens.

As imaged on seismic profiles (Figs. 4, 6, and 9), a prominent regional unconformity divides this upper part of the syn-rift sequence (seismic Unit 4) from the lower part (seismic Unit 5) (Mauffret and Montadert, 1987, this volume), but the significance of this boundary in the tectonic history of the margin is not yet clear. Unit 4 shows much more continuous layering than Unit 5 and likely corresponds to the sequence of marlstone and limestone overlying the sandy turbidites at Site 638. Although no unconformity is documented from paleontological data, biostratigraphic resolution in the upper Valanginian and Hauterivian at Site 638 may be inadequate in this interval to detect a short pause in accumulation (see "Site 638" chapter, Boillot, Winterer, et al., 1987; in this volume: Applegate and Bergen; Taugourdeau-Lantz; Masure; Drugg and Habib; Moullade et al.), and there may be an undetected hiatus at the lithologic change within the upper Valanginian. In fact, the only possible hiatus in the syn-rift sequence at Site 638 occurs a little higher up, within the marlstone-limestone sequence, between Hauterivian and upper Barremian beds, and this may be a case of strong condensation rather than an unconformity (Moullade et al., this volume).

In the Hauterivian-Apian sequence, evidence of tectonically oversteepened slopes is given by the many intervals with slump structures and by debris flow units. The latter are as much as 2 m thick and contain bent fragments of the various *in-situ* sedi-

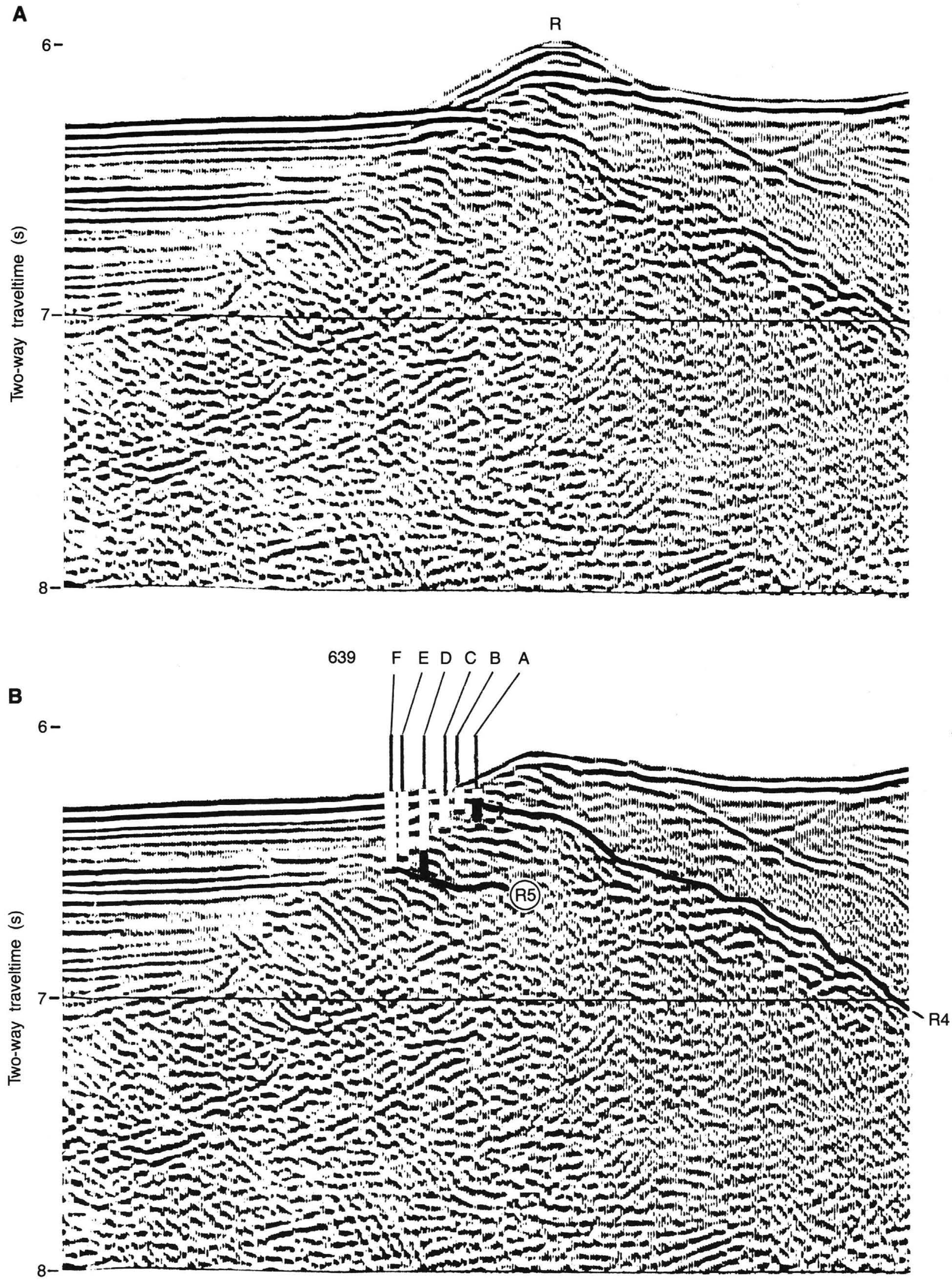

Figure 16. **A.** Section of seismic line GP-101 near Site 639. R = ridge top. **B.** The same section, but with diffractions removed from ridge (R) and showing the projected positions of Holes 639A–639F. For each hole, the Neogene section is shown in white and the Mesozoic rocks in black. R4 = seismic reflector correlated with the top of the dolomite; R5 = top of beds (basement?) with rhyolite and other crystalline rock fragments.

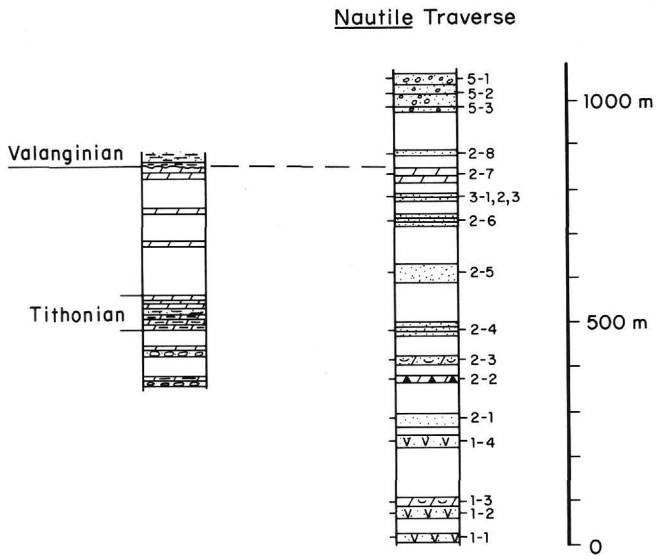

Figure 17. Reconstructed stratigraphic column at Site 639, using the assumptions of Figure 7 and at the *Nautile* traverse about 6 km north of Site 639. At the *Nautile* traverse, no rocks are paleontologically dated, but Sample 2-7 is from dolomite identical to that sampled in the upper part of the sequence at Site 639. Numbers beside *Nautile* column are sample numbers (Boillot et al., this volume). Shell symbols in dolomite indicate mollusks and echinoderms; V symbols in sandstone indicate volcanic detritus.

ments and, beginning in the upper Barremian, fragments of shallow-water limestone.

In addition to the debris flows and slumps, beds of redeposited shallow-water limestone punctuate the Barremian-Aptian sequence, giving evidence of carbonate banks in the area of Galicia Bank and its southern extension. These beds also show that slopes remained steep enough to initiate turbidity currents that carried pebble-sized limestone fragments. A special feature of the redeposited limestone beds, studied in detail by Haggerty and Germann (this volume), is their cementation by chalcedony and megaquartz, the oxygen-isotopic composition of which suggests deposition from warm fluids, most likely channeled along faults.

Age of the Break-Up Unconformity

Although the strategy on Leg 103 was to concentrate mainly on sampling rocks older than the prominent unconformity that separates strongly faulted sequences below from nearly unfaulted strata above (i.e., on syn- and pre-rift rocks), we were successful in coring across the break-up, or sealing, unconformity, and thus, we can date the time when important extensional faulting ceased on the margin. Most models for continental margin evolution associate the cessation of faulting on the margin with the inception of seafloor spreading in the adjacent ocean.

Cores from Site 641 show an unconformable contact between marlstone with interbeds of redeposited shallow-water limestone below and carbonate-free black claystone above, close to the Albian-Aptian boundary ("Site 641" chapter, Boillot, Winterer, et al., 1987). We correlate this with the regional break-up unconformity seen on seismic records (Figs. 4 and 6C). The hiatus at the contact is small—less than 1 m.y.—in the stratigraphic scheme adopted by the Leg 103 scientists ("Introduction, Objectives, and Principal Results," Boillot, Winterer, et al., 1987), but the lithologic change marks not only the cessation of turbidity currents bearing coarse bioclastic debris but also a shallowing of the compensation depth for calcium carbonate relative to the seafloor. At DSDP Site 398, about 150 km southwest of Site 641, a similar lithologic change at the same biostratigraphic level is also interpreted as coincident with the regional seismic boundary (Sibuet, Ryan, et al., 1979). At Site 641, the

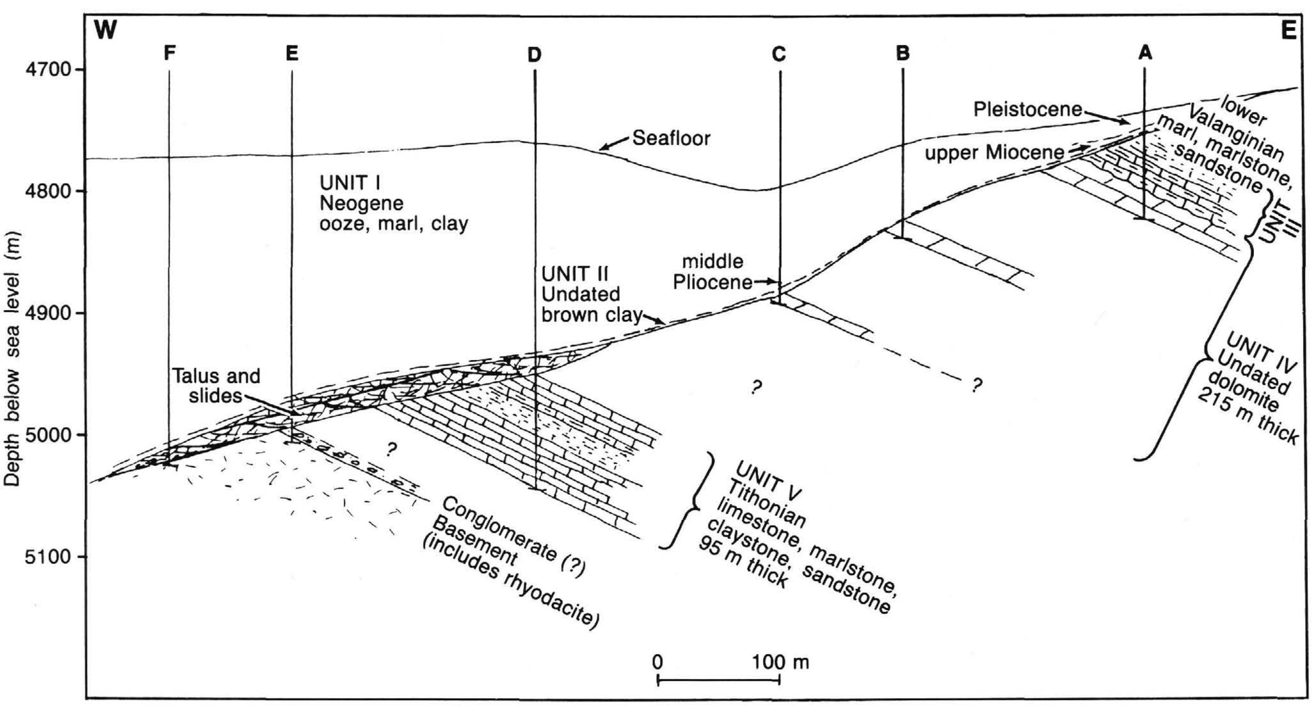

Figure 18. Geologic cross section through Site 639 (from "Site 639" chapter, Boillot, Winterer, et al., 1987), interpreted as having no faults but with buried rock slide(s) to explain dolomite above limestone in Hole 639D and limestone and dolomite above conglomerate(?) in Hole 639E. The rhyolite is treated as part of the basement.

unconformity lies about 40 m above the magnetic reversal referred to M0 by Ogg (this volume). Anomaly M0 is in the lower Aptian, about 4 m.y. older than the beds just beneath the unconformity, using the time scale of Palmer (1983).

We have no direct evidence on the date of inception of seafloor spreading in the adjacent Atlantic, because the planned site was not drilled for lack of time during Leg 103. The seafloor seaward of the Galicia margin is within the mid-Cretaceous magnetic quiet zone, because there is no sign of anomaly M0. On the other hand, as shown on the magnetic anomaly map (Fig. 1), anomaly M0 is present close to the ocean/continent boundary in the region immediately south of the Galicia margin. A possible explanation is that a fracture zone separates ocean crust just seaward of Galicia margin from crust in the area south of 41.5°N, where M0 is recognizable. The age discontinuity across this zone, if we accept the break-up unconformity as a signal of the beginning of spreading west of the Galicia sector, is at least 4 m.y. The beginning of seafloor spreading in the southern region, where M0 can be mapped, would be near the beginning of the Aptian or slightly earlier, depending on how much oceanic crust is present beneath the Iberian Abyssal Plain, east of anomaly M0.

Data on Mn and Fe concentrations in pelagic carbonates from Holes 638B and 641C are interpreted by Clauser et al. (this volume) as suggesting regional hydrothermal activity in early Hauterivian time, and again from late Barremian to early Aptian time. These data are consistent with northward propagation of seafloor spreading in the North Atlantic during Early Cretaceous time.

If the beginning of spreading in this southern region is also associated with its own break-up unconformity, older than the break-up unconformity on the Galicia margin, that older unconformity may be expressed on the nearby Galicia margin. This may be the prominent regional unconformity at the base of seismic Unit 4 (Fig. 6C), which corresponds to the unconformity or condensed sequence at the Barremian/Hauterivian boundary or to the contact between turbidites and marlstone in the upper Valanginian at Site 638 ("Site 638" chapter, Boillot, Winterer, et al., 1987), that is, about 2–8 m.y. prior to the time of the field reversal marked by anomaly M0.

SUGGESTIONS FOR ADDITIONAL WORK TO ADDRESS OUTSTANDING PROBLEMS ON THE GALICIA MARGIN

Several of the major outstanding tectonic problems remaining after Leg 103 are solvable by additional work on this margin, and we present concrete suggestions here on how and where the work might best be done. We believe that the solution of these problems is vital to the understanding required for building and testing geodynamic models of passive margin evolution, and some of these problems are ripe for attack within the limits of existing technology.

Seismic Horizon S (GB)

In our opinion, direct sampling across horizon S is a major objective for the future. To test the detachment fault hypothesis directly, it is necessary to obtain a controlled set of samples across the supposed normal shear that cuts the lithosphere. On the Galicia margin, the best candidate for this shear zone is the strong reflector horizon S, which we assume merges near the foot of the margin with the tectonic contact between thinned continental crust and serpentinized upper mantle. Unfortunately, in the region where horizon S was defined by de Charpal et al. (1978), the reflector is more than 1.5 s twt beneath the seafloor. Assuming a minimum velocity of sound of 5 km/s in the continental crust, horizon S would lie at a sub-bottom depth of more than 3 km, a depth not within our reach using present technology, especially in consideration of the difficulties in drilling through basement rocks. We must therefore find another site, where horizon S is more accessible.

We have already shown that a deep reflector comparable to horizon S can be identified on a seismic profile taken northwest of Galicia Bank (Fig. 14), about 40 km north of the area where horizon S is well known on many profiles. If a detailed site survey using a multichannel reflection system shows an unambiguous correlation between the northern deep reflector and horizon S, we then would have a realistic chance of reaching the supposed detachment fault with the drill. The northern deep reflector rises northwestward to crop out at the seafloor. Drilling to a depth of 1 to 2 km along profile GP-05 a short distance southeast of the outcrop should enable us to sample the rocks associated with the strong reflector.

If this reflector is verified to be the detachment fault, we could progress in our understanding of the kinematics and dynamics of shearing in the lithosphere. The samples would provide quantitative data on the history of pressure, temperature, and stress conditions in the shear zone. Dating of the time of crystallization of new minerals associated with the deformation should furnish the necessary data for a chronology of events. Finally, study of the physical properties of the drilled rocks, including acoustic impedance, would provide a basis for a geophysical interpretation of horizon S as a strong seismic reflector.

Horizon S is one of the remaining major enigmas of passive margins. Resolution of this enigma is without question a major objective for the coming years. Galicia margin offers a unique opportunity for attaining this ambitious objective without a fundamental technological change.

Basement West of the Peridotite Ridge (GB)

In 1985, when *JOIDES Resolution* set out on Leg 103, the basement west of the peridotite ridge (Fig. 4) was included in the drilling objectives (see the "Introduction, Objectives, and Principal Results" chapter, Boillot, Winterer, et al., 1987). At that time, no one doubted the oceanic character of this western region, which is within the Cretaceous magnetic quiet zone, and the drilling proposal was therefore based on determination of the age of the oldest oceanic crust formed between the Galicia and Newfoundland margins. After considerable discussion, JOIDES panels gave this objective a secondary priority, which resulted in our giving it up for lack of time.

Now, in light of drilling results at Site 637, the nature of the basement bordering the peridotite ridge on the west remains as an important unanswered question. We do not know the western limit, beneath the thick sediment cover, of the ultramafic rocks that crop out on the peridotite ridge. Assuming the applicability of the model shown in Figure 11, it is possible that peridotite could underlie at least part of the magnetic quiet zone between the Galicia margin and anomaly 34. At this time, we have no data to distinguish "true" oceanic crust from a layer of serpentine of the same thickness. It is doubtful that geophysical techniques, including seismic refraction, can discriminate satisfactorily between these two possibilities, especially because there are no magnetic anomalies. Samples are required.

If the basement consists of basalt and gabbro crystallized soon after the emplacement of the ridge peridotite, then geochemical and petrologic studies can shed light on any genetic relations between the basalts and gabbros and the peridotites cored at Site 637. If, on the other hand, the basement is serpentine, then the question of mechanisms for the emplacement of mantle rocks at the seafloor is wide open. The peridotite ridge is now considered to be a structure formed by processes that operated at what became the foot of the margin. If it is established that the adjacent ocean floor is also formed of serpentine, the same reasons giving the ridge the particular significance we have assigned it in these *Proceedings* would no longer be valid. Instead, it would evidence oceanic opening without eruption of lavas at the surface.

Pre-Valanginian Stratigraphy and Possible Jurassic Rifting (ELW)

The evidence from drilling at Site 639 and from samples collected during *Nautile* dives (Boillot et al., this volume) strongly suggest the presence of locally thick sequences of pre-Valanginian ("pre-rift") sedimentary rocks on the Galicia margin, but we cannot yet construct a reliable stratigraphic column for these strata. We know the age of only one interval, of Tithonian limestone. The available evidence points toward great thicknesses—as much as 1000 m—of coarse clastics interbedded with shallow-water carbonate rocks. We must know the age range and environments of deposition of these strata if we are to understand the complete history of crustal thinning and rifting on this margin and develop reliable geodynamic models. Judging from seismic profiles, the strata crop out extensively on the high, north-trending fault-line scarps southwest of Galicia Bank, and can be sampled with dredge, submersible, or drill, in increasing order of reliability.

Seismic Unit 4: Age and Tectonic Significance (ELW)

Because neither the drilling at Site 398 nor at Site 638 unambiguously pinned down the age of the base of seismic Unit 4, we are still left with a significant gap in the reconstruction of the tectonic history of the Iberian margin and its relation to seafloor spreading events in the adjacent nascent ocean. We have suggested that the unconformity may be due to changes in the tectonic regime—rates of faulting—associated with the beginning of seafloor spreading in the region immediately south of the Galicia sector of the Iberian margin. What is required is a closely spaced set of samples across the seismic unit boundary where seismic Unit 4 is well developed. The section in the half-graben shown in Figure 19 would be ideal.

CONCLUSIONS

Although drilling during Leg 103 left certain fundamental questions unanswered, especially those relating to the pre-Late Jurassic history of the margin and to the true nature of the deep reflectors seen on seismic profiles, the results add significantly to our understanding of the early stages of evolution of the Galicia margin, which is now among the best known in the world.

In particular, cores from the long ridge of peridotite along the foot of the margin display the effects of partial melting, stretching and serpentinization during cooling, and ascent of the original peridotite from depths of about 30 km to the seafloor during the rifting stage of the margin. Rifting occurred along a set of north-trending, west-dipping normal faults that most likely soled out in a regional décollement surface, imaged in seismic records as the S reflector. The drill cores demonstrate that the rifting fragmented an Upper Jurassic carbonate platform, which may in turn rest on an older sedimentary terrane, the existence of which is hinted at by samples from dredge hauls and *Nautile* dives. During the Valanginian-Aptian, clastic sediments, largely derived from uplifted crystalline basement and from contemporaneous carbonate platforms on Galicia Bank, poured into the half-grabens as faulting continued episodically. Faulting virtually ceased in the latest Aptian, as seafloor spreading began in the adjacent Atlantic, and Albian black muds were deposited unconformably over the old structures.

The drilling, dredging, diving, and seismic results, taken together, now frame many of the remaining important questions into hypotheses that can be tested with the drill.

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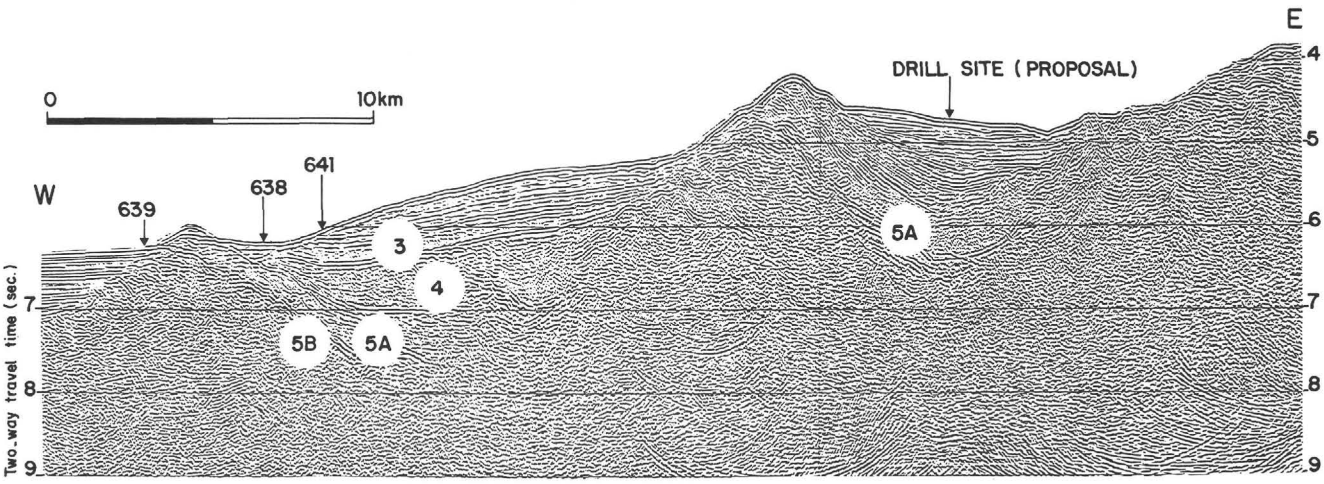

Figure 19. Seismic-reflection profile GP-101 through a location where it is convenient to date the base of seismic Unit 4. At Site 638, only the wedge edge of Unit 4 was sampled, and at Site 641, drilling terminated in the uppermost part of Unit 4. 3 = Albian post-rift sequence; 4 and 5A = Valanginian–late Aptian syn-rift sequence; 5B = Tithonian carbonate platform. Courtesy of L. Montadert.