54. REGIONAL GEOLOGY OF THE GOBAN SPUR CONTINENTAL MARGIN¹

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ABSTRACT

All available geophysical data have been integrated with the Leg 80 drilling results to produce a regional geological synthesis of the Goban Spur passive continental margin. The thin sediment sequence overlying basement is divided into three seismostratigraphic sequences. The lowest of these, sequence 3, occurs in fault-bounded basins, often wedgeshaped in cross-section. Intraformational reflector geometry shows that faulting and tilting of the underlying basement occurred simultaneously with the deposition of formation 3. The upper two sequences are draped over the basement topography and postdate any large-scale fault movement. The early history of the margin, the rift phase, was characterized by the formation of an "en-échelon" series of margin-parallel normal faults, most of which downthrow to the west, toward the incipient ocean. No evidence of fractures perpendicular to the margin has been found.³ Estimates of upper-crustal thinning, derived from the geometry of fault blocks, are in the range 1.03 to 2.20. They show no relationship with whole-crustal extension values derived from gravity modeling, indicating that the uniform extension model of passive-margin formation cannot be applied to Goban Spur. Evidence from one drill site (549) indicates that the rift phase had a duration of about 15 m.y. However, the age of the rifting-to-spreading transition derived from Site 549 is between 2 and 4 m.y. greater than the age derived from Site 550, drilled on the ocean crust. A possible explanation is that extension within the upper continental crust was taken up by dyke injection, particularly within the rift axis, during the later stages of rifting. Syn-rift volcanic activity, identified from seismic profiles and magnetic-anomaly modeling, may be evidence for this phenomenon. At the end of rifting, a wide zone of fault-bounded basement highs and sedimentary basins, with a relief of ± 1 km relative to sea level, existed east of Pendragon Escarpment. West of the escarpment lay a narrow trough, with water depths up to 2 km. Ocean crust began to accrete in this deep-water area in the middle Albian. The post-Albian structural history of the margin has been dominated by thermal subsidence.

INTRODUCTION

Goban Spur is a classic example of a "starved" passive continental margin, and was chosen for drilling because basement rocks and sediments deposited early in the history of the margin lie within easy reach of riserless drilling techniques. The structural setting of Goban Spur is simpler than that of the adjacent northern margin of the Bay of Biscay, drilled during Leg 48 (Montadert, Roberts, et al., 1979), because it has resulted chiefly from the rifting of Europe from North America, and has thus largely avoided the tectonic effects of the complex movements of Iberia during the Cretaceous and Tertiary.

Bathymetric maps of Goban Spur (Figs. 1 and 2) show a smooth marginal plateau above 2000 m, deepening abruptly to the west along the Pendragon Escarpment to 3500 m (Fig. 2). A further, smaller, and more irregular escarpment, along which the water depth increases to 4400 m, occurs some 35 km west of Pendragon. The southern flank of the Spur is deeply incised by submarine canyons (Fig. 2), whereas to the north, a smooth, gentle slope leads down into the Porcupine Sea Bight, a marginal basin containing thick Tertiary and Mesozoic sediments (Roberts et al., 1981).

Previous geological sampling in the area was restricted to a few dredge hauls. Hercynian granites, Paleozoic (?Carboniferous) sedimentary and metasedimentary rocks, and Cretaceous and Tertiary sediments were recovered from the southern flank of Goban Spur (Auffret et al., 1979), and Devonian sandstone was recovered from Pendragon Escarpment (Dingle and Scrutton, 1979). At best, however, these give only a fragmented insight into the geology of Goban Spur.

Dating of the transition from rifting to seafloor spreading west of Goban Spur using oceanic magnetic anomalies has not been conclusive, mainly because of controversy over the identification of Anomalies 32 to 34 in the North Atlantic (see Cande and Kristoffersen, 1977), but also because of uncertainty in the prediction of spreading rates in the older regions of the North Atlantic. Eastward extrapolation of post-Anomaly 34 spreading rates gives a younger age for the initiation of spreading than that derived from the new drilling results.

A dense network of seismic reflection profiles has now been collected over Goban Spur and over the ocean crust immediately to the west (Montadert, Roberts, et al., 1979; Roberts et al., 1981; Fig. 3). Previous interpretation of parts of this data base has shown the existence of NW-SE-trending tilted and rotated fault blocks, buried beneath a thin sediment cover (Dingle and Scrutton, 1977, 1979; Montadert, Roberts, et al., 1979; Roberts et al., 1981). The depositional style of the sediments al-

¹ Graciansky, P. C. de, Poag, C. W., et al., Init. Repts. DSDP, 80: Washington (U.S. Govt. Printing Office).

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and for new seafloor physiographic and structural features identified by post-Leg 80 surveys.



Figure 1. Generalized bathymetry (meters) of the continental margin southwest of the U.K. and Ireland, showing locations of Sites 548 to 551 and 400 to 402. Heavily outlined box is the area of Figure 2.



Figure 2. Detailed bathymetry (meters) of the Goban Spur region (contour interval 200 m). Short, heavy lines show locations of figures (this chapter), as indicated by the number adjacent to each line. Long, finer lines show locations of the interpreted seismic profiles illustrated in Figure 11.

lowed their division into "syn-rift" and "post-rift" sequences (Montadert et al., 1979), although before Leg 80 the dating of these sequences was based on longrange correlations with the Leg 48 boreholes on the northern margin of the Bay of Biscay.

The seismic reflection data used in this chapter include 3400 km of multichannel and 3300 km of singlechannel profiles held by the Institute of Oceanographic Sciences (U.K.), 1750 km of multichannel profiles held by the Institut Français du Pétrole (France), and 2700 km of single-channel profiles held by the University of Edinburgh (U.K.) (Fig. 3).

SEISMOSTRATIGRAPHY

High-quality multichannel seismic reflection profiles allow the recognition of three seismostratigraphic sequences above acoustic basement. The upper two sequences, which often cannot be differentiated on single-channel profiles, are draped over the basement and lower sediment unit topography, and are, in general, characterized by parallel or subparallel reflecting horizons. They have not been affected by any major tectonic events.

In contrast, the lower sequence occurs in discrete faultbounded sediment basins, usually wedge-shaped in crosssection. The sequence is characterized by divergent reflecting horizons, the dips of which increase with depth in the section, suggesting that deposition took place during rotation of the underlying fault-bounded basement blocks.

The characteristics of the acoustic basement and the three main seismostratigraphic sequences, numbered 1



Figure 3. Seismic reflection profiles used in this chapter.

to 3 from the seabed downward, are described in detail in the discussion following.

Sequence 2

Sequence 1

Sequence 1 is the youngest sequence on the Goban Spur margin, and is characterized by weak layering in its upper part and by a series of discontinuous, almost chaotic reflectors in its lower part. This latter series is highly distinctive, and occurs over the entire margin, but is best developed beneath the upper slope (e.g., profile CM 10) immediately landward of Site 549, Fig. 4). Sequence 1 is of relatively uniform thickness (500-1000 m), except over the seaward tips of the higher tilted blocks (e.g., beneath Sites 549 and 551, Figs. 4 and 5), where it may be notably thinner. A number of minor unconformities appear to be present within sequence 1, but are seen only in areas where the sediments onlap basement highs, or in the eastern, shallow-water part of the study area, where erosion is indicated by truncation of reflectors and by channel formation (Fig. 6).

Over much of the margin, sequence 1 apparently overlies sequence 2 unconformably, although this relationship is not always obvious in areas where both sequences are characterized by parallel, flat-lying reflectors. In other areas, however, the basal strata of sequence 1 clearly onlap or downlap the reflector marking the top of sequence 2 (Figs. 4 and 7); channels cut into sequence 2 (Fig. 6) and the truncation of intra-sequence 2 reflectors at its upper boundary are evidence for a period of erosion.

Sequence 2 is characterized by extreme variability both in its appearance on seismic reflection profiles and in its thickness. West of Goban Spur, where it directly overlies oceanic Layer 2, it consists of a relatively thick (up to 750 m), highly transparent lower unit, overlain by a thinner (100-300 m), strongly layered unit (Fig. 8). In the basin to the east of Site 551 (Fig. 5), a very thin (less than 250 m) transparent unit, restricted to the axis of the basin, is overlain by a relatively thick (up to 800 m), strongly layered unit. Higher on Goban Spur, to the east of Site 549 (Fig. 4), sequence 2 can be divided into three weakly layered to transparent units, separated by unconformities and exhibiting a complex onlap and downlap relationship. On the top of Goban Spur, in the region of Site 548, it is generally thin (100-500 m) and weakly to moderately layered (Fig. 6).

The total thickness of sequence 2 is highly variable, ranging from zero over the crests of certain basement blocks to 1000 m in the axes of the intervening basins. There is no evidence, however, that active faulting controlled its distribution, and the geometry of the intrasequence reflectors indicates post-tectonic deposition. The style of deposition clearly separates sequence 2 from the underlying sequence 3.

Sequence 3

Sequence 3 generally has an unconformable relationship with the overlying sequence 2 (Figs. 4, 6, 7, 9). It



Figure 4. Migrated seismic reflection profile through Site 549 (seismostratigraphic sequences as described in text). Note the change in depositional style between sequence 3 (syn-rift) and sequences 1 and 2 (post-rift). Profile located on Figure 2.



Figure 5. Migrated seismic reflection profile through Site 551 (seismostratigraphic sequences as described in text). The deepest reflector at the northeast end of the profile, which appears to dip southwestward beneath the basaltic volcanics drilled at Site 551, is interpreted as the top of the continental basement. The escarpment to the southwest of Site 551 marks the approximate position of the ocean-continent transition. Profile located on Figure 2.

occurs in discrete, usually wedge-shaped, fault-bounded basins, which may contain up to 4 km of sediment (Figs. 9-12 [Fig. 10 in back pocket]). Sequence 3 is characterized by a series of moderate to strong, divergent reflecting horizons, the dips of which increase with depth in the section, indicating that deposition took place during rotation of the basement blocks. Syn-depositional faulting is common within sequence 3, and may give it an almost chaotic appearance (Figs. 4 and 7). The deepest reflectors within sequence 3 normally appear to parallel the top of the underlying basement block, although the top of the basement is often difficult to recognize beneath the thicker sequence 3 accumulations. Localized high-amplitude reflectors, seen within sequence 3, have previously been identified as the top of carbonate reefs (Masson and Roberts, 1981). Migrated seismic profiles, coupled with new high-resolution lines, show, however, that two types of high-amplitude reflector oc-



Figure 6. Migrated seismic reflection profile through Site 548 (seismostratigraphic sequences as described in text). Note that erosion (dated as pre-Campanian at Site 548) has truncated both the basement horst beneath Site 548 and the syn-rift sediments in the half-graben immediately to the northeast. Profile located on Figure 2.



Figure 7. Unmigrated seismic reflection profile across the southern margin of Porcupine Sea Bight (seismostratigraphic sequences as described in text). Profile located on Figure 2.

cur. One of these usually caps a mound-like buildup of indistinct reflectors, which passes laterally into the normal layered structure of sequence 3. Reflectors of this type are identified as reef bodies (Fig. 9). The other type is flat-lying to irregular, usually of very high amplitude, has no obvious mound-like buildup beneath it, and gives rise to abundant diffractions on unmigrated seismic profiles (Fig. 10). Reflectors of this type are identified as volcanic bodies, possibly sills. On many of the poorerquality seismic lines, however, it is not possible to distinguish between these two phenomena.

Acoustic Basement

The basement is usually distinguished from the overlying sediment cover by its lack of internal structure and by its diffractive character (when seen on unmigrated multichannel or single-channel seismic reflection profiles). However, weak layering within some basement



Figure 8. Migrated seismic reflection profile across the strike of the ocean crust west of Goban Spur. Profile located in Figure 2 (seismostratigraphic sequences as described in text).



Figure 9. Migrated seismic reflection profile across a half-graben on Goban Spur (seismostratigraphic sequences as described in text). Note the divergent reflector pattern within the half-graben and the reef body toward its southeast edge. The minor folding seen within sequence 2 near the northwest margin of the profile resulted from Eocene reactivation of the Lower Cretaceous fault which bounds the half-graben. Profile located on Figure 2.

blocks may correspond to pre-rift sedimentary rocks. Two basement types, believed to represent continental crust and oceanic Layer 2, can be distinguished.

Underlying Goban Spur to the east of Pendragon Escarpment, the top of the continental basement is typically marked by a single strong reflector, indicating a high velocity contrast with the overlying sediments. The basement is broken up into a series of tilted, fault-bounded blocks (horsts) separated by sediment-filled basins (grabens or half-grabens).

Between Pendragon Escarpment and the smaller, unnamed escarpment to the west (Fig. 2), seismic reflection profiles show an acoustic basement surface which superficially resembles the top of the continental crust (Fig. 5); however, other geophysical methods (gravityand magnetic-anomaly modeling) suggest an area of transitional crust (Scrutton, this vol.). This is reinforced by drilling results at Site 551; it will be further discussed later.

West of Site 551, the top of oceanic basement is characterized (on unmigrated seismic reflection profiles) by a highly diffractive surface of considerable relief. Migrated seismic reflection profiles show this surface to consist of a series of eastward-dipping strong reflectors (interpreted as lava flows) offset by faults downthrowing to the west. These faults, some of which have a throw in excess of 1000 m, are the primary cause of the basement relief.

CORRELATION BETWEEN DRILLING RESULTS AND SEISMOSTRATIGRAPHY

Detailed correlations between the drilling results and the seismic reflection profiles through the drill sites are presented in the site chapters (this vol.). The following



Figure 11. Interpreted seismic profiles summarizing the main features of the Goban Spur margin. Profiles are arranged from north (top) to south (bottom, see Fig. 2). The deep sediment basin at the eastern end of profile CM 08 is the Porcupine Sea Bight Basin (Fig. 1).



Figure 12. Thickness of syn-rift sediments in the Goban Spur area (contour interval 500 m, but with 200-m contour added). Main accumulation of syn-rift sediments is in long, narrow, northwesterly-trending basins.

summary is concerned mainly with the age and facies of the three regional seismostratigraphic sequences.

Sequence 1

Sequence 1 was sampled at Sites 548, 549, and 550; the spot-coring undertaken in the upper part of the section at Site 551 gives little information. It ranges from middle Eocene to Recent (Fig. 13).

At Sites 548 and 550, its lower boundary is an unconformity, with lower Eocene sediments below and middle Eocene and Oligocene, respectively, above; at Site 549, it corresponds to a facies change within the middle Eocene, although seismic profile CM 10 indicates the presence of an unconformity, presumably associated with a hiatus shorter than can be detected by biostratigraphy. The sediments of sequence 1 are predominantly calcareous oozes, although more marly intervals occur within the Pliocene and Quaternary.

Sequence 2

Sequence 2 was sampled at all four drill sites. It corresponds to lower Albian to lower Eocene sediments (Fig. 13). Much of this interval, however, is represented by various unconformities and condensed sequences. Sequence 2 therefore consists of a number of discrete sediment units, each deposited in a relatively short period separated by longer periods of erosion, nondeposition, or very slow deposition (Fig. 14). The main periods of

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Figure 13. Summary of seismostratigraphic sequences and chronostratigraphy of equivalent sediments for Sites 400 to 402 (North Biscay margin) and 548 to 551 (Goban Spur margin).

sedimentation within the sequence-2 time interval occurred within the early Eocene and in the Maestrichtian to late Campanian (chalks and marly chalks deposited at all drill sites), in the early Cenomanian to late Albian (calcareous mudstone, Site 550 only), and in the middle to early Albian (calcareous siltstone, Site 549 only). The variability in age, thickness, and facies of the sampled Cretaceous section corresponds well to the variability of the lower part of sequence 2, as seen on seismic reflection profiles.

Sequence 3

Sequence 3 was sampled only at Site 549. Here it consists of sandy calcareous mudstones, underlain by grainstones, which are in turn underlain by interbedded calcareous and noncalcareous sandy mudstones. This 290-m sequence was deposited entirely within the ?late Hauterivian and early to early late Barremian interval. The seismic reflection profile through the site (Fig. 4) shows that sequence 3 is markedly thicker to the east, in the adjacent half-graben. Nevertheless, the configuration of reflectors within sequence 3 in this half-graben suggests that the oldest part of the sequence has been sampled, thus dating the initiation of rifting as ?late Hauterivian or early Barremian. The upper part of sequence 3, however, is clearly absent at Site 549, and may represent Aptian sediment.

Unconformities

Vail et al. (1977) hypothesized a relationship between wide-ranging unconformities and global changes in sea level. However, unconformities on a passive margin such as Goban Spur can equally well relate to the tectonic evolution of the margin, and oceanographic changes in the evolving oceanic basin may also have profound effects (Montadert et al., 1979).

For example, the unconformity encompassing the entire Aptian at Site 549 may be related to a drop in sea level in the late Aptian (Vail et al., 1977). On the other hand, it may be a purely local phenomenon related to the faulting and tilting of the underlying basement block. The latter explanation is supported by the probable occurrence of easterly-thickening Aptian deposits in the half-graben immediately to the east of Site 549, as deduced from seismic profiling.

At the deep-water Site 550, where the entire Turonian-to-Campanian interval (a span of about 25 m.y.) is represented by only 20 m of mudstone and calcareous mudstone, it can be speculated that the low sediment accumulation rate correlates with the high sea-level stand at that time, which may have inhibited the oceanward transport of sediment from the east (Vail et al., 1977). This is likely to be only part of the explanation, however, since studies of calcareous microfossils indicate that



Figure 14. Summary of sediment accumulation rates and distribution of unconformities for Sites 548 to 551. A condensed interval is defined as having a sediment accumulation rate of less than 5 m/m.y. Note that the only unconformities which occur across the entire margin are those in the middle to upper Miocene, middle Oligocene and upper Paleocene, although the middle part of the Upper Cretaceous is absent or highly condensed at all four sites.

the site was below the carbonate compensation depth for much of the Late Cretaceous. Thus, although shallower parts of the margin (and much of western Europe) accumulated considerable thicknesses of nannofossil chalk during this period, dissolution may have prevented carbonate deposition at Site 550, and may have contributed, in part at least, to the deposition of the highly condensed unit.

Even in the Tertiary part of the post-rift sediments, which was laid down in a stable tectonic environment affected only by relatively slow subsidence, the ideas of Vail et al. (1977) must be applied with care to the deeper Leg 80 drill sites. Two sites (549 and 551) were situated at the top of escarpments, and net deposition at these sites was almost certainly reduced relative to the rest of the margin. The occurrence of what appears to be a buried Eocene "escarpment edge" just to the east of Site 551 (Fig. 5) supports this view. At Site 550, carbonate dissolution occurred periodically throughout the Tertiary, complicating the record of sedimentation, just as it did in the Cretaceous.

Site 548, situated in relatively shallow water (1250 m) in an area of subdued topography, should, however, have a "normal" sedimentary record. Here, unconformities in the upper Paleocene, at the top of the lower Eocene, in the middle Oligocene, and at the top of the middle Miocene (Fig. 14) correlate precisely with four of the largest Tertiary sea-level drops postulated by Vail et al. (1977) (Poag et al., this vol.). The tracing of these events to the deeper-water sites (Fig. 14) seems to indicate that, despite the questions asked in the preceding paragraphs, these unconformities do indeed represent marginwide events. It is not clear at present, however, whether these relate to sea-level changes, or other oceanographic events, or a combination of both.

DISCUSSION: GEOLOGICAL EVOLUTION OF THE GOBAN SPUR CONTINENTAL MARGIN

Pre-Rift Geology: The Continental Basement

Continental basement was sampled at two of the Leg 80 drill sites (548 and 549) where low-grade Devonian metasedimentary rocks were recovered (see site chapters). Dredging has also recovered granitic rocks, giving radiometric ages in the range 270 to 290 m.y. from Granite Cliff and Menez Bihan, and Paleozoic sediments and metasediments (Carboniferous and ?older) from King Arthur's Castle, Granite Cliff, and the southern end of Pendragon Escarpment (Auffret et al., 1979).

Permian, Triassic, and Lower and Middle Jurassic rocks have been noticeably absent in both dredge hauls and drill samples from the Goban Spur and north Bay of Biscay margins (Auffret et al., 1979); Montadert, Roberts, et al., 1979; this chapter), although a thick Triassic and Lower and Middle Jurassic succession is known in the Fastnet Basin only 100 km to the east of Site 548 (Robinson et al., 1981). Furthermore, a thin ?Permo-Triassic succession was drilled in IGS borehole 83/24-1 (Fig. 1), 270 km to the east of Site 548, and a thicker Permo-Triassic sequence exists farther east in the English Channel (Evans et al., 1981).

Upper Jurassic basement rocks have not been proven on Goban Spur, although local occurrences of intrabasement reflectors possibly indicate the presence of Mesozoic basins. In contrast, Upper Jurassic and possibly lowermost Cretaceous sedimentary rocks up to 3 km thick cap basement fault-blocks in the northern Bay of Biscay (Montadert et al., 1979). This is in agreement with the ideas of Avedik and Howard (1979), based on seismic refraction work: they suggested that the north Biscay margin crosscuts the Western Approaches Mesozoic Basin, and that the Goban Spur margin crosscuts the structurally high Cornubian Ridge (see Roberts et al., 1981).

Structure within the Continental Crust

Beneath Goban Spur, the continental basement is broken up by a series of (listric?) normal faults trending predominantly northwest to southeast (Figs. 11, 15, 16). Most of these faults downthrow toward the ocean (westward). Fault spacing is typically 5 to 20 km, and the throw of major faults may be in excess of 3 to 4 km. The result of this faulting is a series of tilted basement blocks separated by asymmetric, wedge-shaped basins (half-grabens). Many of the higher tilted blocks are characterized by flat, eroded crests (Figs. 11 and 15), indicating that they lay at or above sea level during the rifting phase.

Detailed studies of the fault patterns have been made possible by the dense coverage of seismic profiles available (Fig. 3). An "*en-échelon*" fault pattern is clearly indicated (Fig. 15). Contrary to the ideas of Dingle and Scrutton (1977, 1979) and Montadert et al. (1979), there is no evidence for the occurrence of NE-SW-trending fractures offsetting the main fault trend, although a group of ENE-WSW normal faults marks the southern edge of the Porcupine Sea Bight and the northern edge of a major half-graben on Goban Spur (Fig. 15; Roberts et al., 1981).

In general, the complexity of the fault pattern observed in the Goban Spur area increases toward the south. In the north, the structure is dominated by a few very large westward-throwing normal faults which can be traced for tens of kilometers subparallel to the margin. Toward the south, the fault patterns become less coherent, the fault direction becomes more variable, and a significant number of eastward-throwing faults occur.



Figure 15. Depth to acoustic basement derived from seismic reflection profiles (Fig. 3). Contour interval 500 m.



Figure 16. Depth to the top of sequence 3 (top of the syn-rift sequence), derived from seismic reflection profiles (Fig. 3). Note that most of the deep northwesterly-trending half-graben basins created by the rifting (Fig. 15) had been completely filled by this time. Contours in meters.

This change in structural style appears to be concentrated along a ENE-WSW-trending zone centered along a line between 48°20'N, 13°W and 49°N, 11°W. This line corresponds to the northern edge of a ENE-WNWtrending magnetic anomaly (Fig. 17), which can be traced ENE into Cornwall (e.g. Guennoc et al., 1978), and thus may mark a southwest extension of the Cornwall Granite trend (Roberts et al., 1981). Further evidence in support of this theory is given by the Carboniferous granites dredged from Menez Bihan and Granite Cliff (Fig. 17), the age of which (270–290 m.y.; Auffret et al., 1979) is similar to that of the Cornubian granites (about 300 m.y.; Raynor, 1981). We suggest that the change in structural style is controlled by a change in the superficial basement geology, and that the more complex fault patterns in the south are related in some way to the presence of granitic intrusions, which are absent to the north.

Extension within the Continental Crust

It is widely accepted that extension and thinning of the continental lithosphere occurs during the formation of intracontinental rifts (McKenzie, 1978); Royden and Keen, 1980; Sclater et al., 1980; Sclater and Christie, 1980; Le Pichon and Sibuet, 1981; Beaumont et al., 1982; Avedik et al., 1982). However, the mechanism by which thinning and extension take place is still a matter



Figure 17. Magnetic anomaly map of Goban Spur and the oceanic area immediately to the west. Note the change in anomaly character in the region of the continent-ocean transition. The two small crosses (bottom center) are the locations of dredged Paleozoic granites.

of controversy. McKenzie (1978) proposed a model based on uniform extension and thinning throughout the lithosphere. Other authors prefer a model in which a lower ductile lithospheric layer is preferentially thinned relative to a brittle upper layer (de Charpal et al., 1978; Montadert et al., 1979; Avedik et al., 1982).

A possible test of these two hypotheses is the comparison of the whole-crustal thinning, as identified from seismic refraction studies or inferred from gravity-anomaly modeling, with the extension observed in the upper crust, which can be calculated from seismic reflection profiles. The latter calculation, however, is critically dependent on (1) correct identification of the pre-rift basement reflector, and (2) an accurate assessment of the dip of the fault plane. Examples illustrating the potential errors, giving extension (β) values which may vary by a factor of 2, are shown in Figure 18. The first example (Fig. 18A), showing the tilted basement block underlying DSDP Site 400 (Montadert, Roberts, et al., 1979). explains why Le Pichon and Sibuet (1981) were able to calculate very high superficial extension ($\beta = 2.6$) for this block, whereas Avedik et al. (1982) calculated a much smaller value ($\beta = 1-1.45$). In this case, the parallel nature of the reflectors within the blocks suggests a pre-rift sediment basin, rather than syn-rift sediments, which would be characterized by a more divergent reflector pattern. The top of basement block 1 (Fig. 18A) therefore approximates to line BB', and the measured superficial extension is small ($\beta < 1.5$). The second example (Fig. 18B) illustrates the dependence of β on the accuracy of estimating the fault-plane's angle of dip.

This dependence becomes even greater when lower-angle faults are under consideration; but low-angle faults are often more easily resolved on seismic reflection profiles.

The superficial extension of several tilted fault blocks on Goban Spur has been calculated (Fig. 19), and can be compared to the total crustal extension as defined by the seismic-refraction and gravity-anomaly modeling studies of Bott and Watts (1970), Scrutton (1979 and this vol.), and Avedik et al. (1982). A pre-rift continental crust thickness of 30 km, based on the crustal thickness beneath southwestern England (Bott and Watts, 1970), is assumed. For each tilted fault block, the total range of possible extension values is given (see calculations in Fig. 18), thus taking into account uncertainties in the interpretation of the seismic reflection profiles.

The total range of superficial extension values is 1.03 to 2.20, with no obvious increase in the amount of extension toward the continent-ocean transition. In contrast, the apparent whole-crustal extension increases from just over 1 some 150 km from the continent-ocean transition to values greater than 4 adjacent to it (Avedik et al., 1982; and see fig. 4 of Bott and Watts, 1970). Indeed, the superficial extension would appear to be independent of the whole-crust extension. These results clearly support lithospheric extension models based on apparent differential thinning between upper and lower lithosphere.

Age and Duration of the Rifting Phase

The dating of the oldest syn-rift and the youngest post-rift sediments at Site 549, coupled with the dating



Figure 18. Possible variations in the superficial extension (β) calculated from faulted basement blocks, related to variations in geological interpretation. A. This shows the importance of picking the correct top of the basement block (redrawn from Montadert et al., 1979). B. This shows variations in β dependent on the choice of dip of the inferred fault plane.

of the oldest sediments above ocean crust at Site 550, allows accurate assessment of the length of the rifting phase. At Site 549, the oldest post-rift sediments were dated latest early Albian (Müller, this vol.; Sigal and Magniez, this vol.) At Site 550, the oldest sediments above ocean crust are dated late Albian (Vraconian) (Müller, this vol.; Sigal and Magniez, this vol.). Absolute dating of these events, however, is complicated by difficulties in dating the stage boundaries within the Cretaceous. Hailwood et al. (1980) dated the latest early Albian at 106 to 107 m.y. ago and the Vraconian at about 101 m.y. ago, giving a discrepancy of 5 or 6 m.y. between the apparent ages of the initiation of spreading derived from Site 550 and the initiation of post-rift sedimentation derived from Site 549. Most of this discrepancy-perhaps up to 3 m.y., given the slow spreading rates deduced for this part of the ocean floor (see later)—can be accounted for by the positioning of Site 550 some 15 km west of the continent-ocean transition. A new assessment of Upper Cretaceous stage dating (Kennedy and Odin, 1982) places the Vraconian at about 96 m.y. ago and the late early Albian at about 102 m.y. ago. This gives a discrepancy of 6 m.y. in the age of the rifting-spreading transition derived from the two sites. This new age-scale also gives higher initial seafloor spreading rates (see later), which means that only 2 m.y. of the 6 m.y. discrepancy can be accounted for in terms of the positioning of Site 550.

It is therefore apparent that the age of the transition from rifting to spreading measured at Site 549 differs by between 3 and 4 m.y. from that deduced from Site 550. This difference may be explained in one of three ways, or by any combination of the three. First, Site 550 was drilled on a basement high which may have remained sediment-free for a short time after its formation. The seismic profile through Site 550 shows draping of the oldest sediment over the basement high, but cannot conclusively exclude the possibility that the oldest part of the sediment section is absent over the high (Fig. 20). Second, older sediment may occur beneath the lava flow on which Site 550 is located. This is supported by the seismic profile through Site 550, which clearly shows a strong subhorizontal reflector some 0.1 s (TWT) below the level to which Site 550 was drilled (Fig. 20). Furthermore, if the basalt flow in which drilling at Site 550 bottomed is traced eastward, it is seen to terminate within the sediment section, the lower part of which disappears beneath it to the west (Fig. 20). Third, it is plausible that faulting and movement of the basement block beneath Site 549 ceased before seafloor spreading began, and that extension of the continental crust during the late stages of the rift phase was accommodated by dyke intrusion, particularly in the area between Pendragon Escarpment and the continent-ocean transition. This is supported by the occurrence of basaltic rocks beneath Site 551, and by magnetic modeling in the continentocean transition area (Scrutton, this vol.).

In conclusion, we are able to date the transition from rifting to spreading between latest early Albian and middle Albian, if allowance is made for the positioning of



Figure 19. Comparison between whole-crustal extension values calculated from gravity models (figures in parentheses) and superficial extension values calculated from seismic reflection profiles (figures without parentheses). Gravity-model information is from Bott and Watts (1970), Scrutton (1979 and this vol.), and IOS unpublished information.



Figure 20. Migrated seismic reflection profile through Site 550 (seismostratigraphic sequences as described in text). Profile located on Figure 2.

Site 550 some 15 km west of the apparent continentocean transition. This gives an absolute age range of 104 to 107 m.y. according to Hailwood et al. (1980), and 98 to 102 m.y. according to Kennedy and Odin (1982).

The oldest syn-rift sediments sampled at Site 549 are probably lower Barremian (Sigal and Magniez, this vol.), although their microfossil content does not rule out their being upper Hauterivian. This gives an absolute age of 120 to 122 m.y. (Hailwood et al., 1980) or 113 to 115 m.y. (Kennedy and Odin, 1982). The length of the rifting phase in the area of Site 549 was therefore on the order of 15 m.y. (± 4 m.y.). It cannot be predicted, however, that all the syn-rift basins on the Goban Spur become inactive at the same time; earlier or later movement on faults other than those controlling the Pendragon Escarpment may extend the duration of the rifting phase.

The Continent-Ocean Transition

In regional terms, the continent-ocean transition west of Goban Spur is most easily recognized on the basis of a change in magnetic-anomaly character (Fig. 17), with anomalies of relatively short wavelength and high amplitude to the west of the transition and those of longer wavelength and lower amplitude to the east. However, the change in magnetic character marks only a boundary between weakly magnetized metasediments of the continental basement to the east and highly magnetized basic igneous rocks, which may be ocean crust or synrift volcanics, to the west (Scrutton, this vol.). For this reason, we have chosen to define the precise position of the continent-ocean transition in terms of a change in basement character as observed on seismic reflection profiles. West of Goban Spur, the transition is also marked by a negative (-20 to -40 mgal) free-air gravity anomaly on its landward side (Fig. 21; and see Scrutton, this vol.).

Within the study area, over thirty seismic reflection profiles have been taken across the continent-ocean transition (Figs. 3, 5, 20, 22 [Fig. 22 in back pocket]). Considerable structural variation can be seen along the length of the transition. Northwest of Goban Spur, profile CM 07 (Fig. 22A) shows a change from highly diffractive basement to the west to smoother basement, containing vague suggestions of tilted fault blocks, to the east. The transition zone is about 10 km wide, and there is no change in basement depth across it (Fig. 15). Immediately to the west of Goban Spur, the apparent continent-ocean transition is marked by a large change in basement depth (Figs. 5, 15, 20). This escarpment gradually diminishes in height both to the north and to the south. Along the western portion of the northern margin of the Bay of Biscay, the structure of the transition is clearly illustrated by profile CM 13 (Fig. 22C). Here, basalts of oceanic Layer 2 are at a slightly higher structural level than the most oceanward continental crust, and appear to have flowed over it. A 15-km-wide transition zone, in which basaltic rocks (in effect oceanic Layer 2) overlie continental basement, is suggested.

Scrutton (this vol.) has constructed a number of gravity- and magnetic-anomaly models across the continentocean transition west of Goban Spur. The magnetic



Figure 21. Free-air gravity anomaly map of Goban Spur and the oceanic area immediately to the west. Note the linear gravity low immediately to the east of the continent-ocean transition west of Goban Spur, and the linear gravity high which mark the volcanic ridge underlying Site 551.

models, in particular, show the variability of the continent-ocean transition zone, and highlight the problem of determining the exact location of the continentocean transition. Along much of the Goban Spur margin, it is clear that basic igneous rocks occur to the east of the continent-ocean transition as defined by seismic reflection profiles (Fig. 23). These may be interpreted either as rift-phase basic igneous rocks or as eastward ex-



Figure 23. Summary of magnetic-anomaly modeling in the Goban Spur area, showing the distribution of inferred syn-rift volcanics (from Scrutton, this vol.).

cursions of an irregular continent-ocean transition. Magnetic modeling does not allow a distinction to be made between these two possibilities (Fig. 24). We prefer, however, an interpretation based syn-rift volcanics (Figs. 23 and 24B), because on certain seismic profiles (e.g., Fig. 5) continental basement can be seen to extend westward beneath a volcanic cover. In this interpretation, true ocean crust occurs only to the west of the escarpment, between Sites 550 and 551 (Fig. 15). Immediately to the east of this, and beneath Site 551, an area of transitional crust, composed of continental basement extensively intruded by, and buried beneath, a cover of basaltic material, is present. This may be the expression of crustal extension by dyke injection during the latter stage of rifting, as was discussed earlier as a possible explanation for the discrepancy between the ages of the rifting-tospreading transition derived from Sites 549 and 550. Such occurrences of volcanic rocks within the rift zone would appear to be intermittent phenomena, since on the basis of magnetic anomalies, rift-phase volcanics are absent along the southern part of the Goban Spur margin (Fig. 23; and see Scrutton, this vol.).

Maury et al. (this vol.) have concluded that the basalts sampled at Site 551 are chemically indistinguishable from normal oceanic tholeiites and have none of the characteristics of rift-phase volcanic rocks. However, they were unable to determine a meaningful K-Ar age for these rocks, probably because of potassium mobility during late low-temperature alteration.

We conclude that the available evidence does not allow an unequivocal assessment of the location of the continent-ocean transition west of Goban Spur. We prefer an interpretation in which the boundary follows the



Figure 24. Possible geological models for the continent-ocean transition in the region of the Site 551, based on seismic-reflection, gravity-anomaly, and magnetic-anomaly data. Model A interprets the basic igneous rocks in Figure 23 as eastward excusions of an irregular continent-ocean transition. Model B, which is preferred, here, interprets them as rift-phase basic volcanics. See text.

escarpment to the west of Site 551, with the volcanic ridge underlying Site 551 being an area of transitional crust formed by extensive intrusion of continental basement by basaltic magma.

The Ocean Crust

Ocean Layer 2 was sampled at Site 550, where 33 m of pillow lavas and thin lava flows were recovered (see site chapter). These are overlain by upper Albian sediments, which gives the basalts a minimum age of about 101 m.y. (Hailwood et al., 1980) or 96 m.y. (Kennedy and Odin, 1982). Owing to alteration, no meaningful K-Ar absolute ages could be determined from basalts recovered at Site 550, despite the remarkably fresh appearance of the core samples in hand specimens (Maury et al., this vol.).

The oldest magnetic anomaly identified west of Goban Spur is now believed to be Anomaly 34 (Fig. 17) (Cande and Kristoffersen, 1977; Scrutton, this vol.). The weak, linear magnetic anomalies observed between Anomaly 33-34 and the proposed continent-ocean boundary (Fig. 17) are not believed to indicate magnetic polarity reversals, and Scrutton (this vol.) has shown that they can be adequately modeled using the basement topography combined with changes in the intensity of magnetization of the basaltic basement rocks.

The rate of accretion of the earliest ocean crust can be calculated knowing the age of the basement at Site 550, the age of the young end of Anomaly 34, and the distance between Site 550 and the young end of Anomaly 34, measured parallel to the spreading direction. The young end of Anomaly 34 has been consistently placed at the Santonian/Campanian boundary (LaBrecque et al., 1977; Lowrie and Alvarez, 1981); this has been assigned an absolute age of between 79 m.y. (LaBrecque et al., 1977; Hailwood et al., 1980) and 84 m.y. (Lowrie and Alvarez, 1981; Kennedy and Odin, 1982). The minimum age of basement at Site 550 corresponds to the late Albian (Vraconian), giving an absolute age of between 97 m.y. (Lowrie and Alvarez, 1981; Kennedy and Odin, 1982) and 101 m.y. (Hailwood et al., 1980). The distance between Site 550 and the young end of Anomaly 34, measured perpendicular to the magnetic anomaly strike, is 115 km. This gives maximum spreading rates of between 0.5 cm/yr. (using the absolute dates of LaBrecque et al., 1977, and Hailwood et al., 1980) and 0.9 cm/ yr. (using the dates of Lowrie and Alvarez, 1981, and Kennedy and Odin, 1982), much lower than has previously been assumed for the Cretaceous quiet-zone evolution of the North Atlantic.

Migrated multichannel seismic reflection profiles across the ocean crust allow a resolution of its topography previously masked by the diffraction patterns commonly associated with unmigrated profiles across oceanic basement. On lines perpendicular to the magnetic anomaly trend (e.g., CM 10 and CM 11), the oceanic basement surface is marked by a strong, eastward-dipping reflector cut by faults which downthrow to the west. The throw of these faults is commonly between 60 and 600 m, although occasionally it may exceed 1 km; fault spacing varies between 1 and 8 km (Figs. 8, 20, 25). These values



Figure 25. Distribution of (A) fault throws and (B) fault spacing in the ocean crust west of Goban Spur.

are almost identical to those deduced from detailed studies of the present-day slow-spreading Mid-Atlantic Ridge (Laughton and Searle, 1979), and seem to confirm the presence of ocean crust beneath Site 550 and the area immediately to the west. They also suggest that a fully developed mid-ocean ridge had been created very shortly after the initiation of seafloor spreading.

In general, few sub-basement events are seen on seismic profiles across ocean crust. However, one line subparallel to the strike of the ocean crust shows a number of strong intrabasement dipping reflections (Fig. 26). These are spaced at 5 to 15 km and have apparent dips on the order of 20°, although the dips are poorly controlled because few velocity data were obtained from the oceanic basement. We believe that these dipping reflections represent highly oblique crossings of faults within the ocean crust; these are the same set of faults that generate the steplike relief of the ocean crust on profiles parallel to the spreading direction, and which were originally formed at the mid-ocean ridge (Fig. 8). They occur only on profiles subparallel to the strike of the ocean crust (i.e., subparallel to the fault direction) because only along this azimuth is the apparent dip of the faults low enough to be detected by seismic reflection methods.

POST-RIFT SUBSIDENCE

Introduction

The subsidence of passive continental margins after continental breakup is a widely known phenomenon believed to be controlled primarily by the cooling of the continental lithosphere (Sleep, 1971; Royden and Keen, 1980; Beaumont et al., 1982). Theory predicts that the thermal subsidence of any point on a margin is dependent on the degree of lithospheric thinning beneath it



Figure 26. Migrated seismic reflection profile along the strike of the ocean crust west of Goban Spur. Arrows mark dipping events within the ocean crust. Profile located in Figure 2.

and on the time elapsed since the thinning occurred (Le Pichon and Sibuet, 1981).

Table 1. Depth constraints for Sites 548 to 551.

The subsidence history of any drill site on a margin can be calculated if sufficient depth-vs.-time control points can be delineated. The present water depth is one such point, and others may be derived from micropaleontology, from former positions of the carbonate compensation depth, or from the occurrence of subaerial erosion surfaces. A subsidence history derived in this way contains, however, two components: one caused by thermal effects, the other caused by sediment loading. It may be further complicated by changes in sea level through time.

An estimate of the subsidence caused by sediment loading can be made using either an Airy-type local loading model or a flexural loading model (Watts and Ryan, 1976; Watts and Steckler, 1981). A flexural loading model is believed to describe most accurately the response of the basement to loading in the post-rift phase of evolution of a passive margin (Watts and Steckler, 1981). It is relatively difficult to apply, however, since it depends on the elastic thickness of the lithosphere, which will change through time with the thermal evolution of the margin (Watts, 1978; Watts et al., 1980).

Calculation of Post-Rift Subsidence

Since the post-rift sediment sequence is thin on Goban Spur, a generalized model of sediment-loading subsidence can be calculated without introducing significant error into the thermal-subsidence calculations (Fig. 27 and Gruvel, pers. comm., 1982).

In an ideal situation, the thermal subsidence of any point on the margin would be estimated by subtracting the subsidence caused by sediment loading from the total subsidence derived from considerations of depth vs. time and sediment accumulation rate (Table 1). Unfortunately, the constraints on the total subsidence profiles of the Leg 80 sites are not sufficiently tight (Table 1) to permit accurate total subsidence curves to be drawn; the

Site	Present water depth (m)	Depth below present seabed (m)	Age (m.y.)	Depositional depth (m) ^a
548	1256	470-530	61-72	?<500 (pal)
			>72	?0 (erosion surface)
549	2533	479-664	104-106	>200, <500 (pal)
		664-674	?108	$? \simeq 0$ (erosion surface)
		>674	118-122	< 200 (pal)
551	3387	?45-130	50-71	>200 (pal)
		138-142	90-95	<1500 (pal)
550	4432	310-465	34-65	>2000 (pal)
	N. 1997 - N	595-686	95-102	<3000 (CCD)

^a pal = derived from micropaleontology; CCD = derived from the estimated depth of the carbonate compensation depth at any moment in geological time.

calculation has therefore only been possible using the basic assumption that the thermal subsidence decays exponentially through time, with a thermal decay constant of 50 m.y. (Sleep, 1971; Royden et al., 1980). The results of these calculations (Gruvel, pers. comm., 1982) are summarized in Figures 27 and 28.

Depth-vs.-Time Constraints from the Leg 80 Sites

The depth constraints which control the subsidence curves for Sites 548 to 551 are summarized in Table 1. The main problem in the generation of these figures is the poor resolution as to depth vs. time that can be achieved for deep water using benthic foraminifers. This restriction, which particularly affects the deep-water Sites 550 and 551, is caused by the wide depth range over which most species of deep-water benthic foraminifers occur.

A more specific problem applies to Site 548. Here, microfossils suggest a depth of deposition of less than 500 m in the Campanian (site chapter), which implies a subaerial relief of over 1500 m in the Albian. Such a relief would have affected much of the shallower part of Goban Spur, and would have been expected to contrib-



Figure 27. Composite subsidence profiles across the Goban Spur margin compiled from the individual subsidence profiles from Sites 548 to 551 (from Gruvel, pers. comm., 1982). A. Subsidence resulting from sediment loading. B. Thermal subsidence. C. Total subsidence. Profiles are illustrated for the early Albian (105 m.y. ago), end of Cenomanian (95 m.y. ago), end of Cretaceous (65 m.y. ago), and of Paleocene (53.5 m.y. ago), end of Eccene (38 m.y ago), and the present. The arrow (OCT) on the upper profile indicates the approximate position of the ocean-continent transition.

ute considerable clastic sediment to the basin to the west. Little such clastic sediment was sampled, however, in the pre-Campanian part of the Upper Cretaceous section at the other three drill sites. We have therefore concluded that the relatively shallow-water fauna of the Campanian at Site 548 has suffered some downslope transport, and that the final depth of deposition was over 500 m. The considerable period of submersion of Site 548 before the late Campanian implied by this conclusion—a period from which no sediments have been preserved—may be reflected in the pre-Campanian hardground sampled at the site (site chapter).

Application of the Subsidence-Curve Data: Post-Rift Evolution of the Margin

Subsidence curves from the individual drill sites have been used, first, to produce a series of subsidence profiles across the margin through time (Fig. 27), and second, in combination with the available geological data, to produce a series of geological cross-sections across the margin through time (Fig. 29). All the data illustrated are along seismic profile CM 10 (Fig. 2).

The first cross-section of Figure 29 shows the topography of the rift system at the beginning of the late Albian, at the time of the rifting-to-spreading transition. East of the Pendragon Escarpment, a relatively wide zone of fault-generated basement highs and sediment basins lay near sea level, with a maximum relief of ± 1 km. To



Figure 28. Thermal subsidence curves for Sites 548 to 551, calculated from the constraints given in Table 1, assuming that the thermal subsidence decays exponentially with a time constant of 50 m.y. (from Gruvel, pers. comm., 1982).

the west, in the axis of the rift system, a narrow deep trough, 60 to 70 km wide, assuming a symmetrical rift, had developed, with water depths up to 2000 m. Despite the relief surrounding this trough, virtually no syn-rift sediments were deposited within it, which suggests that it formed late in the evolution of the rift over a relatively short time. The basalt-capped edifice drilled at Site 551 lies within the central trough, and thus may also be a late feature. Superficially, it resembles an "outer high" in the context of Schuepbach and Vail (1980). These authors argue, however, that "outer highs" result from tectonic rather than volcanic processes, a theory which cannot be supported in the case of the Goban Spur margin.

The late Albian cross-section also indicates that the first ocean crust accreted in a water depth of about 2200 m, assuming that no post-middle Albian movement has occurred on the fault-bounded escarpment west of Site 551. This is confirmed by the subsidence data for Site 550, which indicate that the crust beneath the site accreted at a depth of just over 2200 m (Fig. 28).

An alternative method of calculating the depth of accretion of the first ocean crust involves backtracking along the theoretical oceanic age-depth curve (Parsons and Sclater, 1977; Tucholke and Vogt, 1979). At Site 550, where the minimum age of the ocean crust is 102 m.y., the curve predicts a present basement depth of about 5600 m (Tucholke and Vogt, 1979). But the measured depth at Site 550 is 5100 m, which reduces to 4800 m when sediment loading is taken into account, some



Figure 29. Summary of the post-rift geological history of the Goban Spur, taking into account both subsidence and sedimentation, illustrated using a cross-section along seismic reflection profile CM 10 (Fig. 2).

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800 m shallower than theory predicts. This suggests that the crust under Site 550 formed some 800 m above the "normal" level (2500–2700 m, Parsons and Sclater, 1977; Tucholke and Vogt, 1979), at a depth of 1700 to 1900 m. This is in general agreement with the figure of 2200 m obtained from the subsidence-curve calculation, particularly if the considerable uncertainty in the subsidence calculation is taken into account.

The remainder of the cross-sections in Figure 29 illustrate the geological development of the margin between the end of the Cenomanian and the present. It is notable that the highest basement blocks may not have been completely submerged until the end of the Cenomanian, and that submarine relief related to the exposure of basement rocks at the seabed probably persisted until the end of the Eocene.

POST-RIFT SEDIMENTATION: COMPARISON WITH THE NORTHERN MARGIN OF THE BAY OF BISCAY

In terms of seismostratigraphy, the post-rift sediments of the Bay of Biscay can be divided into three sequences (Montadert et al., 1979), whereas those of the Goban Spur are easily divided into only two (this chapter). Nevertheless, the two areas have a very similar history of post-rift sedimentation (Table 2).

Lower Cretaceous

In the Bay of Biscay, the earliest post-rift sediments are Aptian to Albian "black shales" (de Graciansky et al., 1979), which are clearly recognizable as a discrete unit on seismic reflection profiles (formation 3, Montadert et al., 1979). This facies is much less developed in the Goban Spur area, primarily because the rifting-tospreading transition did not occur until the middle Albian. It is not, therefore, generally recognized as a separate seismostratigraphic unit on Goban Spur. However, lower to middle Albian laminated gray shales, slightly enriched in organic carbon, sampled at Site 549 (site chapter, this vol.), clearly correlate with the "black shales" of the north Biscav margin. At Site 550, a similar facies was also encountered at the base of the postrift sequence, but here it is slightly younger (upper Albian to middle Cenomanian; see site chapter).

Upper Cretaceous

The Upper Cretaceous is represented by highly condensed sequences of nannofossil chalks on both the northern margin of the Bay of Biscay and on Goban Spur. Several hiatuses are encompassed within the Upper Cretaceous section (this chapter, Fig. 13; Montadert et al. 1979); the entire Cenomanian to lowermost Cam-

Northern (Mont	n Bay of Biscay, Sites 400–402 adert, Roberts, et al., 1979)	Chrono- stratigraphy	Goban Spur, Sites 548-551 (this vol.)	
Formation 1	Mainly nannofossil oozes,	Pleistocene	Mainly nannofossil oozes. Alternating marly and carbonate-rich layers in Pliocene and Pleistocene. Very high carbonate content in Miocene.	
	and Oligocene. Alternat-	Pliocene		e. Sequence
	ing marly and carbonate- rich layers in Pliocene and	Miocene		
	Pleistocene. Very high carbonate content in Miocene	Oligocene		
	Marly nannofossil chalks:	Eocene		
Formation 2	relatively high sediment accumulation rate.	Paleocene	Marly nannofossil chalk; high sediment accumulation rate.	
	Mainly chalks; slow sediment		Highly condensed sequence	
	accumulation rate.	Maes- trichtian	(particularly between Campanian and Cenoma- nian) of nannofossil	2
		Campanian	chalks. Numerous uncon- formities, but poor correlation of unconform- ities between sites (see	
	Unconformities at all three	Santonian		
	drill sites	Coniacian	Fig. 14).	
		Turonian		
		Cenomanian	Gray calcareous siltstones, slightly	
Formation 3	"Black shales" enriched in organic matter.	Albian	enriched in organ- ic matter. Facies extends to middle Cenomanian at Site 550 only.	
			Syn-rift sediments	Sequence 3
Formation 4	ormation Syn-rift sediments			

Table 2. Comparison of post-rift seismic sequences of Goban Spur and the northern Bay of Biscay.

panian interval is absent at all three north Biscay sites. Only at Site 549 is a reasonably complete Upper Cretaceous section present, and even there it is only 90 m thick. The scarcity of Upper Cretaceous strata within the study area may also be indicated by the rarity of Upper Cretaceous rocks in the many dredge hauls from the north Biscay margin (Auffret and Pastouret, 1979; Auffret et al., 1979).

Several papers have commented on the apparently reduced thickness of Upper Cretaceous strata on continental margins surrounding the North Atlantic (Rona, 1973; Auffret and Pastouret, 1979; Roberts and Montadert, 1979a, b; Jansa et al., 1979; Sheridan and Enos, 1979). This has, in part, been attributed to erosion (e.g., Sites 400-402, Montadert, Roberts, et al., 1979; Sites 550 and 551 chapters, this vol.), to an elevated carbonate compensation depth, causing dissolution of carbonate sediments (Jansa et al., 1979; Roberts and Montadert, 1979a, b; Site 550 chapter, this vol.), or to an elevated sea level leading to a decrease in sediment transport into the deep ocean (Vail et al., 1977, 1980; Roberts et al., 1981). In reality, various combinations of all these factors probably operated.

Paleocene and Eocene

Early Paleocene sedimentation was characterized by deposition of nannofossil chalks on both Goban Spur and the northern margin of the Bay of Biscay. This period of deposition was terminated by an earliest late Paleocene hiatus, which is recorded at all the Leg 48 and Leg 80 sites. After this break, there was clearly an increase in terrigenous sediment input, coupled with an increase in the sedimentation accumulation rate (Montadert et al., 1979; site chapters, this vol.). This succession of events correlates well with the relatively large drop in sea level 60 m.y. ago (Vail et al., 1977). It may also be related, however, to the Pyrenean tectonism which occurred between the latest Cretaceous and the late Eocene (e.g., Boillot et al., 1979; Grimaud et al., 1982) and which gave rise to localized zones of late Eocene folding and uplift in the southern Porcupine Sea Bight and along the Trevelyan Escarpment (Montadert et al., 1979; Masson and Parson, 1983; Fig. 30).

The upper boundary of the phase of relatively high clastic input is marked by a minor lower to middle Eocene unconformity on Goban Spur, and by a middle Eocene to lower Oligocene unconformity in the northern Bay of Biscay. The diachronous nature of this boundary might be regarded as evidence for a tectonic influence on the lower Eocene clastic-rich facies, since this facies apparently persisted longer in the north Biscay area, where Eocene deformation was most severe.

Oligocene to Quaternary

The Oligocene to Quaternary interval corresponds approximately to seismic formation 1 as defined by both Montadert et al. (1979) and this chapter, although middle and upper Eocene sediments are also included in formation 1 in this chapter. Throughout this interval, nannofossil chalk deposition predominated over the entire study area, and particularly pure carbonates were deposited during the Miocene. Alternating carbonaterich and marly beds characterize the Pliocene and Quaternary, reflecting the climatic fluctuations which occurred during these periods (Pujol et al., this vol.).

CONCLUSIONS

The geological evolution of the Goban Spur continental margin can best be described as two phases, a rifting phase and a seafloor-spreading phase.

The rifting phase was characterized by the formation of a northwesterly-trending "*en-échelon*" fault system, in which the majority of faults downthrow to the southwest, toward the incipient ocean. Sampling of syn-rift sediments at Site 549 has suggested that the main rifting episode was a relatively short-lived event, initiated in the



Figure 30. Migrated seismic reflection profile across the southern Porcupine Sea Bight, showing late Eocene folding (seismostratigraphic sequences as discussed in text). Profile located in Figure 2.

?late Hauterivian or early Barremian and terminated by the initiation of seafloor spreading in the middle Albian. Reconstructions of depth vs. time show that seafloor spreading was initiated within a narrow trough some 2000 m deep, which extended along the axis of the rift system. The remainder of the rift system lay near sea level. A slight discrepancy between the ages of the rifting-spreading transition obtained from Sites 549 and 550 may result from extension in the continental crust having been taken up by dyke injection during the final stages of rifting. Extensive syn-rift volcanism near the rift axis may be related to this phenomenon.

Post-rift evolution of the margin has been characterized by subsidence caused primarily by thermal contraction, with a minor component caused by sediment loading. Faulting is almost unknown in the post-rift section, excepting minor reactivation of some syn-rift faults during the Pyrenean compression. Post-rift sediments are predominantly hemipelagic oozes and chalks characterized by highly condensed sections (e.g., upper Cenomanian-upper Campanian) and numerous unconformities, some of which (e.g., upper Paleocene and middle Miocene) can be traced from the shallowest to the deepest site.

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