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# 1. Leg 210 Synthesis: Tectonic, Magmatic, and Sedimentary Evolution of the Newfoundland-Iberia Rift[1]

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## ABSTRACT

The Newfoundland-Iberia rift is a type example of a nonvolcanic rift. Rifting occurred in two phases: Late Triassic into Early Jurassic, and Late Jurassic through Early Cretaceous. During the first phase, extension occurred in a wide rift mode without continental separation. During the second phase, extension initially was spatially and temporally variable but eventually focused at future distal margins where it was succeeded by seafloor spreading. Second-phase spreading appears to have been concentrated in three episodes: (1) Late Jurassic–Berriasian rifting that culminated in separation of continental crust in the southern half of the rift, (2) Valanginian–Hauterivian rifting of continental crust in the northern rift that culminated in separation of continental crust there and was coincident with rifting of subcontinental mantle lithosphere in the southern rift, and (3) Barremian–Aptian rifting of mantle lithosphere that was at least partially subcontinental. We suggest that rising asthenosphere breached the mantle lithosphere and led to seafloor spreading near the Aptian/Albian boundary. This event is marked by a prominent seismic stratigraphic horizon ("Aptian event") in the sedimentary record of both margins. Extension and exhumation of mantle lithosphere in episodes 2 and 3 is not easily characterized as either continental rifting or seafloor spreading, and we propose that it be referred to as "transitional extension."

Transitional extension during episode 3 (and possibly also during episode 2) was accompanied by intraplate extension of exhumed mantle and by minor magmatism, probably because a well defined plate boundary characterized by significant magmatic accretion had not been


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established and horizontal in-plane tensile stress was elevated throughout the plates. However, during the later part of episode 3 there appears to have been increasing magmatism and even formation of relatively normal ocean crust, at least locally. During the transition between episodes 2 and 3, the southern part of the rift was affected by magmatism related to a mantle plume that formed the Southeast Newfoundland Ridge, J Anomaly Ridge, Madeira-Tore Rise, and probably part of Gorringe Bank at the southern edge of the rift.

M-series magnetic anomalies are observed over the peridotite basement exhumed during episodes 2 and 3. Compared to their counterparts in ocean crust of the main North Atlantic Basin to the south, anomalies older than ~M1 are poorly developed, and our modeling suggests that they are best explained by magnetization of serpentinites rather than igneous rocks. The high-amplitude J Anomaly (~M1–M0) is strongest at the southern edge of the rift, but the amplitude decreases northward to near the southern margin of Galicia Bank and is relatively constant farther to the north. The character of the anomaly most likely reflects the spatial distribution of magma that emanated from the plume at the southern margin of the rift.

Significant postrift magmatism on the Newfoundland margin was expressed at Site 1276 as two diabase sills that intruded Aptian–Albian sediments. The upper sill dates to ~105 Ma (middle Albian) and the lower sill is ~98 Ma (early Cenomanian). The upper sill is intruded at the stratigraphic level of the Aptian event (U reflection in the Newfoundland Basin). Widespread distribution and high amplitude of the U reflection in the basin suggest that sills may be pervasive at this stratigraphic level, which may have been a level of hydrostatic equilibrium for intruding magmas. There is no known magmatism of comparable age and extent on the Iberia margin. The source of the Newfoundland magmas is postulated to be the Madeira and Canary hotspots. The Newfoundland Basin is thought to have passed over these hotspots at ~100–90 Ma before they were crossed by the Mid-Atlantic Ridge and subsequently affected the southernmost part of the Iberia plate and northwestern edge of the African plate.

Key features of the sedimentary record in the Newfoundland-Iberia rift are similar to those of the main North Atlantic Basin to the south, although there are variations at individual drill sites that are related to, for example, input of shallow-water clastic debris and seafloor paleodepth. The sedimentary record cored to date reflects the following:

1. Tithonian–Berriasian shallow-water carbonate platforms, together with rift basins that accumulated detrital and hemipelagic sediments;
2. Valanginian–Hauterivian deposition of carbonate-rich sediments in rift basins above the calcite compensation depth (CCD);
3. Barremian–Cenomanian hemipelagic deposition below the CCD and under dysoxic to anoxic conditions, with the occurrence of at least six oceanic anoxic events;
4. Turonian–Paleocene deposition of reddish to multicolored sediments on a generally well oxygenated seafloor below the CCD; winnowing of fines from sediments at Site 1276 suggests strong abyssal circulation during Turonian to earliest Campanian time; recurrence of dark to black shales in the lower to middle Paleocene section indicates stagnant circulation and low-oxygen conditions in the deepest basins; and

5. Late Paleocene to Oligocene deposition of hemipelagic and pelagic (calcareous and siliceous) sediments, with the CCD at intermediate levels; initiation of strong abyssal circulation that has persisted to the present appears to have begun near the Eocene/Oligocene boundary.

# INTRODUCTION

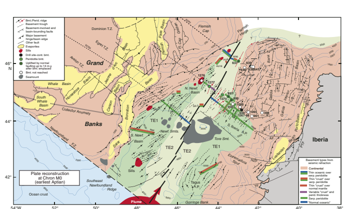
Continental rifting that leads to seafloor spreading commonly is classified as either volcanic or nonvolcanic (Mutter et al., 1988; White and McKenzie, 1989; White et al., 1987), although variations in tectonic stress, lithospheric strength, and mantle conditions likely create a range of margin types between these end-members (Mutter, 1993). At nonvolcanic margins, lithospheric extension has been proposed to occur either symmetrically by pure shear (McKenzie, 1978) or asymmetrically by simple shear (e.g., Lister et al., 1986, 1991; Rosendahl, 1987; Wernicke, 1985). Pure-shear extension is envisaged to thin crust relatively uniformly, with brittle deformation occurring in the upper crust and ductile deformation in the lower crust, and it leads to similar crustal thickness, structure, composition, and subsidence history on conjugate margins. In contrast, conjugate nonvolcanic margins formed by simple shear exhibit marked differences in these features, and any attendant magmatism probably is minimal compared to a pure-shear environment (Buck, 1991; Latin and White, 1990).

To understand how rifts develop, it is necessary to obtain and interpret data from a combination of geophysical surveys, geological sampling (e.g., drilling), and modeling of conjugate margins. During the early phases of the Ocean Drilling Program (ODP), a North Atlantic Rifted Margins Detailed Planning Group (NARM-DPG) was formed by the Joint Oceanographic Institutions for Deep Earth Sampling (JOIDES) Planning Committee to study options and make recommendations for drilling of conjugate-margin transects on volcanic and nonvolcanic margins. For nonvolcanic margins, the NARM-DPG recommended the Newfoundland and Iberia conjugates as high-priority targets. Perceived advantages of drilling these margins included the following:

1. They are considered to be representative of nonvolcanic rifting.
2. Rifting is complete and the entire rift history can be studied.
3. Along-strike spatial relations of crustal conjugates are well constrained in plate reconstructions.
4. Sediments are comparatively thin so that important basement targets can be seismically imaged and many are accessible by drilling.
5. Locations are logistically convenient and readily accessible.

The earliest drilling in the Newfoundland-Iberia rift preceded the NARM-DPG report and was conducted on Galicia Bank (Fig. F1), first during Deep Sea Drilling Project (DSDP) Leg 47 at Site 398 south of Vigo Seamount (Sibuet and Ryan, 1979) and subsequently on the western Galicia margin during ODP Leg 103, Sites 637–641 (Boillot, Winterer, et al., 1988). Drilling designed to follow the concept of conjugate transects was initiated during ODP Leg 149 (Sites 897–901) (Whitmarsh, Sawyer, Klaus, and Masson, 1996) and continued during ODP Leg 173 (Sites 1065–1070) (Beslier, Whitmarsh, Wallace, and Girardeau, 2001). During both of these legs, sediments and basement were sampled along the

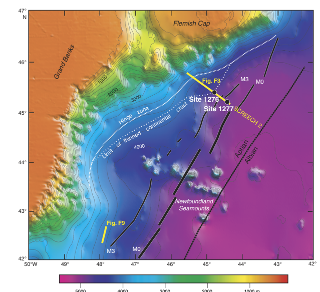
**F1.** Reconstruction of the Newfoundland-Iberia rift at Chron M0, p. 45.



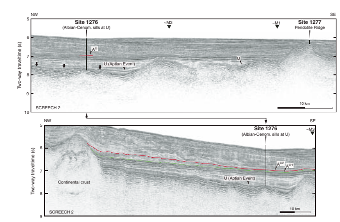
southern perimeter of the Galicia crustal block beneath the edge of the southern Iberia Abyssal Plain. These ODP expeditions were supported and complemented by extensive geophysical studies in this area (e.g., Whitmarsh et al., 1990, 1996b; Pinheiro et al., 1992; Reston et al., 1995; Whitmarsh and Miles, 1995; Reston, 1996; Pickup et al., 1996; Discovery 215 Working Group, 1998; Dean et al., 2000).

In contrast, during this period no scientific drilling was done and little geophysical work was conducted on the conjugate Newfoundland margin. In 2000, the SCREECH program (Study of Continental Rifting and Extension on the Eastern Canadian sHelf) acquired a set of three major seismic reflection/refraction transects that crossed the Grand Banks and Flemish Cap into the Newfoundland Basin (e.g., Funck et al., 2003; Hopper et al., 2004; Lau et al., 2006a, 2006b; Shillington et al., 2006; Van Avendonk et al., 2006). The central transect, SCREECH Line 2 (Shillington et al., 2004, 2006; Van Avendonk et al., 2006), was conjugate to Leg 149/173 drilling off Iberia and it provided the survey data necessary to select and drill the first holes in the Newfoundland margin. During ODP Leg 210, two holes were drilled along this transect (Figs. F1, F2, F3). The first hole, at Site 1276, was located ~15 km west of magnetic anomaly M3 and penetrated to 1736.9 meters below seafloor (mbsf) before hole conditions forced abandonment. The hole terminated in the lower of two diabase sills that were intruded into lowermost Albian to uppermost Aptian sediments (Tucholke, Sibuet, Klaus, et al., 2004), and it is estimated to have bottomed only ~100–150 m above basement. Site 1277 was drilled ~40 km to the southeast on a prominent, shallowly buried basement ridge a few kilometers seaward of anomaly M1 (Fig. F3). In this hole serpentinized peridotite basement beneath a thin cover of basalt flows and basaltic, gabbroic, and serpentinite debris flows was recovered (Tucholke, Sibuet, Klaus, et al., 2004).

F2. Bathymetry of the Newfoundland Basin, p. 47.



F3. SCREECH Line 2 MCS west and east of Site 1276, p. 48.



Drilling on both sides of the Newfoundland-Iberia rift has produced surprising insights into nonvolcanic rifting but at the same time has raised important new questions. Perhaps the single most important observation has been that extensive tracts of mantle ("transition zones") were exhumed seaward of the edges of thinned continental crust and landward of oceanic crust formed by seafloor spreading (we use the term seafloor spreading here to mean generation of new seafloor at a discrete plate boundary and with a major, if not dominant, component of magmatic accretion). Drilling at Site 1277 seaward of anomaly M1 on the Newfoundland margin, and at Sites 897, 899, and 1070 in the area of anomalies M5 to M1 on the Iberia margin, demonstrated that basement there is serpentinized mantle with only very minor indications of magmatism, despite the fact that apparently normal magnetic anomalies are developed in this young part of the M-series reversals. Thus there are major questions not only about the source of the magnetic anomalies but also about whether the prolonged mantle exhumation should be considered to be "seafloor spreading" or just continued rifting, for example, of subcontinental mantle lithosphere. In addition, there are important questions about how expected tectonic events such as separation of continental crust or initiation of seafloor spreading might be manifested in the structural and stratigraphic record along the margins. Finally, drilling at Site 1276 provided the surprising result that substantial and possibly widespread diabase sills were intruded into postrift sediments of the Newfoundland Basin, whereas there is no evidence of similar magmatism on the Iberia margin. Was this magmatic pattern "inherited" because of asymmetries in mantle fertility that developed during rifting, or was it caused by another process, such as plate migration across a mantle plume?

In the following pages, we consider these issues as we summarize the tectonic development of the Newfoundland-Iberia rift through time. We then outline the postrift sedimentary history, and we conclude with thoughts about important issues to be addressed by future drilling.

# TECTONIC DEVELOPMENT OF THE NEWFOUNDLAND-IBERIA RIFT

## Prerift Crust

Continental crust forming the Newfoundland and Iberia basement consists of Precambrian and Paleozoic rocks accreted during Paleozoic closure of the Iapetus and Rheic oceans. The Taconian, Salinic, Acadian, and Alleghenian orogenies (Early to Middle Ordovician, early Silurian, Silurian–Devonian, and Carboniferous–Permian, respectively) resulted in accretion of a series of northeast-southwest–oriented terranes on the present Canadian Atlantic margin (van Staal et al., 1998; Waldron and van Staal, 2001; Percival et al., 2004). Iberia basement accreted during the mid-Devonian to Carboniferous (i.e., in the early part of the Carboniferous–Permian Hercynian [Variscan] orogeny). Easternmost Newfoundland and most of the Grand Banks platform consist of Avalon Terrane (Silurian–Devonian), although a sliver of Meguma Terrane (Carboniferous–Permian) is present at the southern edge of the Grand Banks, separated from the Avalon by a prominent magnetic anomaly (Collector anomaly) (Fig. F1).

The accreted terranes in Iberia are oriented northwest–southeast, but offshore they may have curved northeast and east as part of the Ibero-Armorican arc (Ziegler, 1982; Capdevila and Mougenot, 1988; Silva et al., 2000). Tonalites were intruded into the continental crust of the southern Galicia block (ODP Site 1067) in the late Proterozoic and were later deformed during the Hercynian orogeny (Table T1) (Rubenach, 1999). Permian-age zircons from a surrounding amphibolite also indicate late Hercynian magmatism at this site (Manatschal et al., 2001). Large granitoid batholiths intruded the onshore terranes during and following the Hercynian (Pinheiro et al., 1996). Granulites sampled from the northwest Galicia margin appear to have formed in the lower to middle crust, but they equilibrated at upper-crustal conditions in Late Carboniferous to Permian time (Table T1) (Fuegenschuh et al., 1998); this suggests that they may have been exhumed by a late phase of extension toward the end of the Hercynian orogeny.

T1. Igneous and metamorphic rocks in the Newfoundland-Iberia rift, p. 55.

Large-scale, original structural fabrics within the Paleozoic terranes are well defined by gravity and magnetic anomalies (Welsink et al., 1989; Srivastava et al., 1990a; Silva et al., 2000). A general concordance between these fabrics and subsequent Mesozoic rift patterns suggests that the Paleozoic structures exerted significant control on rift development (Wilson, 1988; Welsink et al., 1989; Pinheiro et al., 1996). Mesozoic rifting occurred during two phases, separated by a period of epeirogenic subsidence, as described below.

## Rift Phase 1: Late Triassic into Early Jurassic

In Late Triassic to earliest Jurassic time, extension operated in a wide-rift mode and created large rift basins over a distance of at least 500 km across the Grand Banks-Iberia platform. The Jeanne d'Arc, Whale, Horseshoe, Carson, Salar-Bonnition, and Flemish Pass basins were formed in the present Grand Banks, and the Lusitanian, Porto, and possibly Galicia Interior basins opened on the Iberia margin (Fig. **F1**) (Wilson, 1988; Murillas et al., 1990). These basins accumulated red siliciclastic sediments during the Late Triassic (Carnian–Norian) and then were filled with widespread evaporite deposits during the earliest Jurassic (Hettangian–Sinemurian) (Jansa and Wade, 1975; Wilson, 1988; Rasmussen et al., 1998; Alves et al., 2003). This rifting was synchronous with rifting between Africa and North America farther south, but it did not lead directly to opening of an ocean basin between Newfoundland and Iberia.

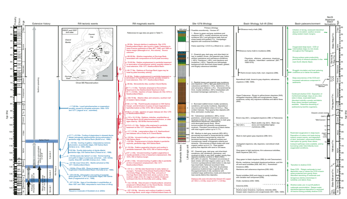
The rifting was associated with relatively minor magmatic activity during the Carnian–Hettangian (Table **T1**). Most of the magmatism appears to have been limited to Iberia, where intrusions and flows affected the southwestern Iberia margin and the Central Iberian Zone (see Pinheiro et al., 1996, for a summary). At the southern edge of the rift, offshore Iberia, a Sinemurian diorite from Gorringe Bank (Carpena, 1984) probably was emplaced during this rift phase. There is only one occurrence of magmatism documented on the Newfoundland side, consisting of diabase intrusions in the Avalon Peninsula (Hodych and Hayatsu, 1980).

During the remainder of the Early Jurassic and through the Middle Jurassic, most of the Grand Banks-Iberia platform experienced epeirogenic subsidence without significant extension (Tankard and Welsink, 1987). An exception was in the Lusitanian Basin south of the Nazaré fault where an episode of extension and subsidence allowed deposition of a thick sequence of Sinemurian–Callovian carbonates (Rasmussen et al., 1998).

## Rift Phase 2: Middle Jurassic through Early Cretaceous

A second phase of rifting that began in the Middle Jurassic became more intense in the Late Jurassic and finally led to continental breakup and seafloor spreading late in the Early Cretaceous (Tucholke et al., 2007; Peron-Pinvidic et al., 2007). During this period, patterns of extension shifted from a wide-rift mode and eventually focused at future distal margins where continental crust separated. Based on analysis of seismic sequences and the timing and patterns of faulting on both sides of the rift, Tucholke et al. (2007) proposed three primary episodes of extension during this rift phase. The first began in the Middle to Late Jurassic and culminated near the end of Berriasian time, the second occurred during the Valanginian to Hauterivian, and the third ended near the Aptian/Albian boundary (Fig. **F4**). Each of these is discussed below.

**F4.** Rift-event summary from latest Jurassic to Paleogene, p. 49.



In this paper, we use the timescale of (Gradstein et al., 2004), which the Leg 210 Scientific Party decided to standardize on for the present *Scientific Results* volume. We point out that this timescale is different from the Gradstein et al. (1995) and Channell et al. (1995) timescales used in the Leg 210 *Initial Reports* volume and also used by Tucholke et al. (2007). In particular, the bases of the Aptian and older stages in the Early Cretaceous are some 3–4 m.y. older in the Gradstein et al. (2004) timescale. This changes the timing of some events discussed by Tucholke et al. (2007); thus, some interpretations in the present text may differ from those in Tucholke et al. (2007).

## Rift Phase 2, Episode 1: Middle Jurassic to Berriasian Continental Extension

Although the Middle Jurassic to Berriasian was a period of general extension across the Grand Banks-Iberia platform, the location, timing, and intensity of rifting varied considerably. In the Jeanne d'Arc Basin on the central Grand Banks, rifting is documented from late in the Middle Jurassic (late Callovian) into the Aptian, but it was most intense during the late Kimmeridgian to early Valanginian (Tankard et al., 1989). Along the southern Grand Banks in the Whale Basin there appears to have been weak Early to early Middle Jurassic extension, followed by increased rifting in the Bajocian–early Bathonian to late Kimmeridgian and then by limited extension through the late Tithonian (Balkwill and Legall, 1989).

On the present Iberia margin, rifting in the Lusitanian Basin occurred primarily in the Late Jurassic (Oxfordian to early Kimmeridgian), with minor extension continuing through the Tithonian to Barremian (Wilson, 1988; Wilson et al., 1989; Rasmussen et al., 1998). Along strike to the north, interpreted Tithonian to Berriasian shallow-water carbonates in the Galicia Interior Basin are highly faulted on seismic reflection sections (Murillas et al., 1990). On thinned continental crust at the deep western margin of Galicia Bank, faulted, shallow-water Tithonian to (?)Berriasian carbonates are succeeded by deeper-water carbonates and clastics (Boillot, Winterer, Meyer, et al., 1987). Deeper-water Tithonian claystones are present over crust along the southern margin of the Galicia continental block (Whitmarsh and Sawyer, 1996; Whitmarsh and Wallace, 2001). Thus major extension appears to have affected both the Galicia Interior Basin and the western and southern Galicia margins during the Tithonian and Berriasian.

Even though the above data for the two margins indicate that strong extension affected the northern part of the rift in the Tithonian to Berriasian, they seem to suggest little coeval extension in the southern rift. However, no precise information is available about timing and intensity of rifting in the Salar-Bonnition Basin or on the conjugate southern Iberia margin seaward and south of the Lusitanian Basin (Fig. **F1**). Thus, it is possible that this was a period of major extension in the central to southern part of the rift. As we discuss below, this extension probably focused at the future distal margins and resulted in separation of continental crust in the southern rift.

Our current understanding of the seaward limits of thinned continental crust along the Newfoundland and Iberia margins is shown in Figure **F1.** In the central Newfoundland Basin and under the Iberia Abyssal Plain along the southern margin of the Galicia block, these limits are reasonably well defined by seismic refraction results and by drilling. On the Newfoundland margin, SCREECH Line 3 refraction data suggest a 60-km-wide zone of very thin (~4 km) continental crust followed seaward by an 80-km-wide zone of exhumed, serpentinized peridotite (Lau et al., 2006b). Similarly, on SCREECH Line 2 about 60 km of thinned continental crust is succeeded seaward by exhumed mantle, with the transition probably occurring in the vicinity of Site

1276 (Van Avendonk et al., 2006). On the conjugate Iberia margin, a change from thinned continental crust to exhumed mantle is well defined by drilling results and by refraction data along CAM lines and the IAM9 line (Chian et al., 1999; Discovery 215 Working Group, 1998; Dean et al., 2000).

South of the Newfoundland and Tore seamounts, however, the crust-mantle transition is poorly constrained. On the Newfoundland margin, diapiric evaporites, most likely of Hettangian–Sinemurian age, in the Salar-Bonnition Basin (Austin et al., 1989) indicate that the seaward hinge line of this basin probably marks the minimum seaward extent of thinned continental crust. A single refraction profile in this area (Reid, 1994) suggests thin "crust" of uncertain (but possibly serpentinite) composition overlying serpentinized mantle seaward of the hinge line, consistent with a crust/mantle boundary near the hinge line (Fig. F1). Under the conjugate central Tagus Abyssal Plain, similar velocity structure may also indicate the presence of exhumed mantle, although Pinheiro et al. (1992) interpreted the basement as oceanic. The edge of thinned continental crust to the east is poorly defined; Pinheiro et al. (1992) suggested that it may be in the eastern part of their refraction line and within the limits of a refraction line described by Purdy (1975) to the south, although it could lie farther east.

Assuming that the limits of thinned continental crust shown in Figure F1 for the southern part of the rift are approximately correct, we can estimate the age of the boundary there based on arguments from seafloor magnetic anomalies. Srivastava et al. (2000) modeled magnetic anomalies within the rift and proposed identifications of the M-series reversals. There are some significant discrepancies between their identifications of older anomalies and the limits of thinned continental crust shown in Figure F1. For example, they identify anomalies M17 and M20 over probable continental crust beneath the Salar-Bonnition Basin and over the thin continental crust interpreted along SCREECH Line 3. In addition, most of the M-series anomalies appear to be over exhumed mantle, so there are major questions about the source of the magnetic anomalies. The source and identification of the anomalies, as well the extension rates that they imply, are discussed in detail in **"Magnetic Anomalies in Zones of Transitional Extension: Character, Origin, and Implications,"** p. 15. In that discussion, we conclude that there is large uncertainty about the identification of anomalies older than ~M8, although the implied extension rates are reasonable. If we take the magnetic anomaly identifications of Srivastava et al. (2000) at face value, the edge of thinned continental crust on the Newfoundland margin falls at about anomaly M16 (~141 Ma; late Berriasian), and on the Iberia side it is at about anomaly M17 (~143 Ma; middle Berriasian). This suggests that extension of continental crust in the southern part of the rift probably proceeded through Tithonian to Berriasian time and culminated near the end of the Berriasian.

The timing of extension discussed above is consistent with ages of known magmatic activity, metamorphism, and crustal cooling apparently associated with rifting (Table T1). Hornblende ages in amphibolites at ODP Site 1067 and 1068 on the southern Galicia margin in the central part of the rift suggest rifting and cooling of these rocks at ages as early as ~167 Ma (Bathonian) to ~153 Ma (Kimmeridgian) and continuing to ~132 Ma (Hauterivian) (Table T1). Igneous rocks in the younger part of this time range (<145 Ma; Berriasian–Hauterivian) are scattered through the central to southern part of the rift, most notably in Gorringe Bank, the Lusitanian Basin, and the Whale Basin; however, igneous rocks of this age are uncommon in the northern part of the rift. At the southern edge of the Galicia block (ODP Site 900), a $^{40}Ar/^{39}Ar$ plagioclase age of 136.4 ± 0.3 Ma (late Valanginian) has been interpreted to mark retrograde metamorphism of gabbros as they were exhumed by rifting (Féraud et al., 1996).

From all the above observations, we conclude that scattered but locally strong rifting extended the continental crust in the Newfoundland-Iberia rift from as early as Middle Jurassic time (Bajocian–Bathonian) to the Berriasian and Valanginian. Rifting was most intense in the southern to central part of the rift during the Middle Jurassic to middle Late Jurassic, followed by strong extension that probably affected the entire length of the rift in Tithonian–Berriasian to early Valanginian time. The later stage of rifting caused widespread extension of continental crust across the Galicia block in the northern rift, but in the southern half of the rift it appears to have culminated in separation of the crust (Figs. **F1, F4**).

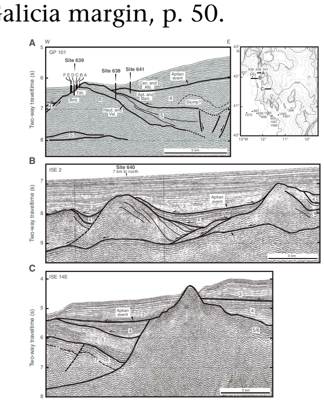
## Rift Phase 2, Episode 2: Valanginian to Early Barremian Continental Extension and Mantle Exhumation

### *Northern Rift Segment: Continental Extension*

In the northern, Flemish Cap-Galicia Bank segment of the rift during this period, only minor extension occurred in the Galicia Interior Basin. Interpreted Valanginian–Hauterivian sedimentary sequences there show some faulting, primarily toward the center of the basin, but the faults are scattered throughout the basin and most have small offsets (Murillas et al., 1990; Pérez-Gussinyé et al., 2003). The southern Galicia margin also shows signs of fault slip along a few major fault blocks in continental crust, whereas other blocks were inactive (Tucholke et al., 2007). In conjugate crust of the Grand Banks, intense fault-controlled subsidence was sharply reduced in the Jeanne d'Arc Basin compared to the preceding rift episode, and basin deposition changed from restricted marine or nonmarine to more open-marine conditions (Tankard et al., 1989).

In sharp contrast to the above, strong extension occurred along the western margin of Galicia Bank and necessarily along the conjugate seaward edge of Flemish Cap. Valanginian–Hauterivian sedimentary sequences on western Galicia Bank were faulted and rotated, primarily toward the end of this period (Fig. **F5**) (Mauffret and Montadert, 1987; Tucholke et al., 2007). Fission-track ages of ~129–126 Ma (Barremian) on apatites in granulites sampled from northwest Galicia Bank are interpreted to record the extensional exhumation of continental rocks on this margin (Table **T1**) (Fuegenschuh et al., 1998).

**F5.** MCS profiles across thinned continental crust of the western Galicia margin, p. 50.



At the latitude of the ODP Leg 103 drilling transect (Fig. **F1**), the boundary between the western, feather edge of thin continental crust and exhumed mantle lies between Site 640 and Site 637 (drilled on a peridotite ridge), ~50–60 km east of anomaly M0 as identified by Srivastava et al. (2000). At an extension half-rate of 9–13 mm/yr (see **"Implied Extension Rates,"** p. 19), this boundary would date to ~129–131 Ma (Hauterivian/Barremian boundary). Farther south along the southwest margin of the Galicia continental crust, the crust/mantle boundary is about the same age. ODP Sites 899, 897, and 1070 were drilled on peridotite ridges approximately on anomalies M5 (130 Ma), M3 (129 Ma), and M1 (127 Ma), respectively, whereas Site 1069 penetrated Tithonian and older (?Paleozoic) sediments over presumed continental crust west of peridotite basement cored at Site 1068 (Fig. F1) (Sawyer, Whitmarsh, Klaus, et al., 1994; Whitmarsh, Beslier, Wallace, et al., 1998). Depending on where the boundary lies between Sites 899 and 1069, it could predate 130 Ma (late Hauterivian) by as much as several million years. Taken together, the above age data suggest that separation of continental crust between Galicia Bank and Flemish Cap occurred from near the Hauterivian/Barremian boundary into the early Barremian. This pattern is consistent with the time at which the last major normal faulting occurred in continental crust along the west-central margin of Galicia Bank (Fig. F5).

### *Central to Southern Rift Segment: Mantle Exhumation*

During the Valanginian–Hauterivian little to no extension occurred in the continental crust of the southern Grand Banks and the southern Iberia margin (Balkwill and Legall, 1989; Wilson, 1988; Wilson et al., 1989; Rasmussen et al., 1998). Instead, extension concentrated within the deep basin, starting from the southern limit of continental crust on Galicia Bank and reaching to the southern edge of the rift. It appears that mantle was being exhumed in that area, and we consider this period, up to near the time of Chron M3 (~128 Ma), to be one of transitional extension (TE1) (Fig. F1). Normal seafloor spreading was not yet established at this time.

Evidence for mantle exhumation during this period comes from both drilling results and interpretations of seismic refraction data. Drilling at Sites 897, 899, and 1068 (Fig. F1) established that basement in those locations consists of serpentinized peridotite with very minor amounts of igneous rocks (Sawyer, Whitmarsh, Klaus, et al., 1994; Whitmarsh, Beslier, Wallace, et al., 1998). A large set of seismic refraction and reflection studies around the southwest Galicia margin (e.g., CAM, IAM9 in Fig. F1) documents thin "crust" (2–4 km) that has low seismic velocities of ~4–4.5 km/s (Chian et al., 1999; Whitmarsh et al., 1990; Discovery 215 Working Group, 1998; Dean et al., 2000). On the basis of seismic velocities, velocity gradients, and the general absence of a Mohorovicic (Moho) reflection, this basement has been interpreted as intensely serpentinized mantle peridotite, with serpentinization progressively reduced with depth (~7.2 km/s increasing to ~7.9 km/s) (Discovery 215 Working Group, 1998; Chian et al., 1999). The upper basement layer on the IAM9 reflection profile published by Pickup et al. (1996) also has a markedly nonreflective character that they interpreted to be the result of intense serpentinization. From refraction data on conjugate SCREECH Line 2 across the Newfoundland margin, Van Avendonk et al. (2006) interpreted a narrow zone of basement between Sites 1276 and 1277 as serpentinized mantle (Fig. F1).

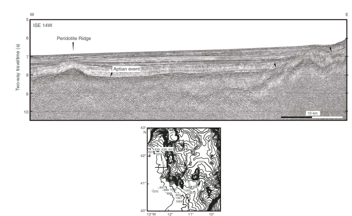
Farther south in the Newfoundland Basin, velocity modeling along SCREECH Line 3 (Fig. F1) suggests an ~80-km-wide zone of exhumed, serpentinized mantle reaching seaward from highly thinned continental crust to about anomaly M3 (Lau et al., 2006b). Three other refraction studies in the conjugate TE1 zones (SR1, R94, and P92; Fig. F1) show a thin (1–3 km), low-velocity "crust" overlying a ~7.1–7.9 km/s layer that is 2–6 km thick and is presumed to be partially serpentinized mantle. The similarity of this velocity structure to that of serpentinized basement along the CAM and IAM9 profiles suggests that TE1 basement in the southern part of the rift is also serpentinized mantle.

As previously noted, Srivastava et al. (2000) proposed identifications of M-series magnetic anomalies within the rift. These anomalies span the exhumed mantle in TE1 and also reach into the older part of TE2, discussed in the next section. The source and identification of the anomalies, and the implied extension rates, are discussed in **"Magnetic Anomalies in Zones of Transitional Extension: Character, Origin, and Implications,"** p. 15.

## Rift Phase 2, Episode 3: Barremian to Late Aptian/Early Albian Extension

Separation of Flemish Cap from Galicia Bank appears to have been completed by about early Barremian time, with mantle subsequently exhumed between these continental edifices (see discussion of peridotite ridges, below). Although separation of continental crust is often considered to mark the cessation of continental rifting and the coincident start of seafloor spreading (e.g., Falvey, 1974), there is good evidence in the Newfoundland-Iberia rift that significant intraplate extension continued well away from the plate boundary throughout Barremian and Aptian time (i.e., from about Chron M3 to younger than M0). Furthermore, a major seismic stratigraphic marker in the sedimentary record on both margins dates to the Aptian/Albian boundary and in the past was interpreted as the "break-up unconformity," presumed to mark the onset of seafloor spreading. This marker is the "U reflection" in the Newfoundland Basin (Fig. **F3**) (Tucholke et al., 1989), the "orange reflection" around DSDP Site 398, and the lateral equivalent of the orange reflection around the Galicia margin (Figs. **F5, F6**) (Groupe Galice, 1979; Mauffret and Montadert, 1988) (see **"Sedimentary Record in the Deep Basins,"** p. 25). These seismic stratigraphic markers are collectively referred to here as the "Aptian event." The Aptian event is the most prominent Mesozoic sedimentary horizon on either margin, and it records a major tectonic and/or sedimentary event. We consider the Barremian and Aptian up to this marker to constitute a second period of transitional extension (TE2) (Figs. **F1, F4**). Below, we summarize the evidence for intraplate extension during this period and offer a possible explanation for its cause and for the cause of the Aptian event.

**F6.** MCS profile showing Aptian event reflection that merges with peridotite ridge, p. 51.



Evidence for Barremian–Aptian extension in the Jeanne d'Arc Basin was presented by Driscoll et al. (1995), who related late Barremian–early Aptian and late Aptian unconformities developed there to rift-onset episodes. Magmatism that may have been related to extension is represented by Barremian to Aptian basaltic flows and sills on the southern Grand Banks and by Aptian lamprophyre dikes that intruded Paleozoic rocks in northern Newfoundland (Table **T1**).

On a broader scale, a major unconformity termed the Avalon unconformity is developed across the Grand Banks, and it appears to mark a major change in stress regime that may coincide with the Aptian event. The unconformity generally separates faulted and folded, pre-Albian sequences below from flat-lying Albian and younger sediments above. Grant et al. (1988) associated this unconformity with "a decline in stress regimes as the continental crust between Flemish Cap and Galicia Bank separated and seafloor spreading began in the late Aptian." On the conjugate margin, Wilson (1988) also discussed an Aptian unconformity in the Lusitanian Basin; he did not specifically relate it to an extensional episode, but he suggested that it might have developed at the onset of seafloor spreading.

In contrast to the above, the deep-basin parts of the rift show relatively little seismic stratigraphic evidence for faulting and tectonic rotation of Barremian–Aptian sedimentary sequences below the Aptian event. A few places in the Newfoundland Basin show this deformation (e.g., Fig. F3, just west of anomaly M1), but in most areas the Aptian event is a very high amplitude reflection that masks the deeper section, and any extensional structures that might be present there are difficult to resolve. Over the continental crust of Galicia Bank, the Barremian–Aptian sequence usually is relatively level between basin-bounding blocks (Fig. F5); where the sequence laps high onto bounding blocks its configuration appears to reflect sediment input from local sources rather than tectonic movement.

In the exhumed mantle south and west of Galicia Bank, age data on igneous intrusions and other geological data on faulting and uplift appear to present a different picture. At the peridotite ridge where ODP Site 1070 was drilled, ~127-Ma basement (anomaly M1) was intruded by gabbroic pegmatites at 127 ± 4 Ma, and then again at ~124.2 ± 0.7 to 123.9 ± 1.2 Ma as indicated by ages of hornblendes with blocking temperatures of ~500°C (Table T1). However, plagioclases in the samples with the latter two ages did not cool below their blocking temperature (~150°–250°C) until ~117–101 Ma. There are two possible explanations for these age offsets. One is that the gabbros were intruded at depth, had a long cooling history, and later were quenched as they were exhumed by local, late-stage faulting (Whitmarsh and Wallace, 2001). The other is that the gabbros were originally intruded near the basement surface and thus cooled quickly, but the plagioclases were recrystallized by continuing hydrothermal activity and their ages thus reflect the end of retrograde metamorphism (Jagoutz et al., submitted [N1]). Presently it is not possible to distinguish between these alternatives. Plagioclase ages of 115.7 ± 0.3 and 111.0 ± 0.3 Ma were also determined for two other gabbro pegmatites intruded in the peridotite basement at Site 1070 (Table T1), and these could reflect ages of original magmatism or late-stage cooling as noted above. In any of these cases, the 117- to 101-Ma activity at Site 1070 would have occurred 10–26 m.y. after basement was first emplaced and ~90–340 km away from the plate boundary, assuming an extension rate of 9–13 mm/yr (see "Implied Extension Rates," p. 19). The ~124-Ma magmatism may have been associated with local extension, and it occurred ~30–40 km from the plate boundary.

There is some evidence that the nearby peridotite massifs drilled at Sites 897 and 899 (Fig. F1) were uplifted by late-stage normal faulting. As noted earlier, magnetic anomaly identifications indicate that basement was emplaced at ~129 Ma at Site 897 and ~130 Ma at Site 899. At these two sites, serpentinized peridotite olistostromes and debris flows covered the peridotite basement during early Aptian and early late Aptian time, respectively. The deposits have been interpreted as being derived either locally at the crests of then-existing basement highs (Gibson et al., 1996; Whitmarsh et al., 2001) or as being emplaced by long-distance flows in a basin-plain setting before normal faulting uplifted the massifs (Comas et al., 1996). If the Comas et al. (1996) interpretation is correct, the flows were uplifted no earlier than early late Aptian time (~120 Ma), some 9–10 m.y. after the basement was first emplaced and ~80–130 km from the plate boundary, assuming that the extension rate was 9–13 mm/yr. The occurrence of such intraplate extension would lend support to the idea that the age data at Site 1070, noted above, can be explained by late-stage normal faulting.

To the north, the peridotite ridge drilled at ODP Site 637 is ~40 km east of anomaly M0 as identified by Srivastava et al. (2000), and at an extension rate of 9–13 mm/yr it should date to ~128–129 Ma. High-temperature shearing of a mylonitized dikelet in peridotite sampled from this ridge ~90 km north of Site 637 predates 122.0 ± 0.6 Ma (Féraud et al., 1988), but plagioclase in the same sample cooled through its blocking temperature about 5 m.y. later at 117 ± 0.9 Ma (Boillot et al., 1989) (Table **T1**). As at Site 1070, the late plagioclase cooling might be due to late faulting or it might mark the end of hydrothermal retrograde metamorphism. In the same location, however, gabbros were emplaced in the peridotite ridge at ~122 Ma and then later sheared (Table **T1**) (Schärer et al., 1995, 2000), indicating that normal faulting occurred >6–7 m.y. after the basement was emplaced and at a distance at least ~50–90 km from the plate boundary.

A sequence of minor magmatic events, much like those observed at Site 1070 off Iberia, occurred on the Newfoundland margin at Site 1277. Peridotite basement there was emplaced at ~127 Ma and coincidentally was intruded by an alkaline gabbro dike (128 ± 3 Ma) and then intruded again at 113.2 ± 2.1 Ma (Table **T1**) (Jagoutz et al., submitted [**N1**]). Gabbro cobbles overlying the peridotite have much younger $^{40}Ar/^{39}Ar$ plagioclase ages of ~92–69 Ma (Table **T1**). The 113-Ma intrusion may have been related to extension up to 130–180 km distant from the plate boundary (assuming 9–13 mm/yr extension rates). However, it is unclear whether the younger ages define original magmatic ages that could have been associated with intraplate extension or whether they mark the end of retrograde metamorphism associated with hydrothermal activity. Either is possible; as discussed in **"Postrift Magmatism,"** p. 21, the Newfoundland Basin contains clear evidence for postrift magmatism.

To explain the observed and inferred examples of intraplate extension noted above, Tucholke et al. (2007) proposed that the entire rift continued to experience elevated in-plane tensile stress throughout Barremian and Aptian time. They suggested that this occurred because strong, probably subcontinental mantle lithosphere was being exhumed, which strengthened the plate boundary and thus distributed strain across the rift. Intraplate extension most likely would have concentrated in weak, serpentinized areas, consistent with the above observations that normal faulting appears to have been more common in the exhumed mantle than in the continental crust of Galicia Bank. Tucholke et al. (2007) also proposed that the rising asthenosphere finally breached the probably subcontinental mantle lithosphere near the Aptian/Albian boundary. This released in-plane tensile stress and led to magmatic seafloor spreading. The plate-boundary release could have resulted in a pulse of relative in-plane compression, causing local and regional lithospheric deformation (Braun and Beaumont, 1989; Kooi and Cloetingh, 1992). If this deformation stimulated mass wasting and mobilized turbidity currents it could account for the reflective sedimentary layers that define the Aptian event in seismic reflection records (Figs. **F3, F5, F6**). According to this hypothesis, the Aptian event is not a "break-up unconformity" in the sense that it marked the separation of continental crust. Instead, it coincides with much later separation of the probably subcontinental lithospheric mantle.

The above interpretation implies that substantial amounts of mantle lithosphere were exhumed within the rift during TE2. As shown in Figure **F1,** peridotite basement was drilled in TE2 at both Sites 1070 and 1277, and basement in the CAM refraction data and part of the IAM9 line south and east of Site 1070 has been interpreted as serpentinized mantle. Refraction data at HU18, W96, and SR2 also suggest thin "crust" (possibly serpentinite) overlying serpentinized mantle, and the seaward end of SCREECH Line 2 has a velocity signature that may represent mixed serpentinized mantle and oceanic crust.

Elsewhere, the SCREECH Line 1 refraction data in TE2 indicate that very thin oceanic crust overlies a thick layer of serpentinized mantle. The seaward portions of SCREECH Line 3, IAM9, and W96 have been interpreted as normal-thickness ocean crust. However, the IAM9 line ends among seamounts north of the Tore seamount complex and the end of SCREECH Line 3 is near the Newfoundland seamounts, so it is possible that crustal characteristics at the seaward ends of these lines were affected by seamount volcanism (Tucholke et al., 2007). In addition, they may have been affected by magma emanating from a plume at the southern margin of the rift, as discussed in the next section. The overall picture is that there was significant mantle exhumation in TE2, but there appears to have been an increasing magmatic component with time that may have included at least local creation of oceanic crust.

## Mantle Exhumation and Magmatism during Transitional Extension Period 2

Five drill sites (637, 897, 899, 1070, and 1277) have sampled basement at either the transition between TE1 and TE2 or farther seaward in TE2 (Fig. **F1**). The basement is dominantly exhumed mantle with a richly variable composition, and it includes websterites, harzburgites (± spinel), lherzolites (± spinel or plagioclase), and generally small amounts of dunite. Igneous components are minor. Mafic rocks include basalt, diabase, and gabbro. Minor alkaline veins are present in the peridotites at Site 1277 (Müntener and Manatschal, 2006), and weakly alkaline rocks are also documented at Site 899 (Cornen et al., 1996b).

Most of the peridotites show broadly similar thermal and alteration histories, although there are local differences (e.g., Agrinier et al., 1988; Evans and Girardeau, 1988; Girardeau et al., 1988; Beslier et al., 1996; Cornen et al., 1996a). In general, these peridotites experienced high-temperature deformation at ~900°–1100°C and pressures of ~7–10 kbar, followed by multiphase metamorphism at decreasing temperatures in the presence of hot fluids. Samples were often mylonitized and then partially reequilibrated at temperatures below ~850°–700°C. At lower temperatures of ~50° to ~300° they were serpentinized in the presence of infiltrating seawater; calcite impregnated and/or filled veins in the brittlely deformed rocks at temperatures as low as near-seafloor conditions (Skelton and Valley, 2000). These features are consistent with the expected history if the mantle were exhumed along one or more major shear zones.

Serpentinized spinel harzburgites recovered at Site 1277 on the Newfoundland margin are among the most depleted abyssal peridotites observed worldwide, probably having experienced 14%–25% melting in the spinel peridotite field (Müntener and Manatschal, 2006). Intervals of normal mid-ocean-ridge basalt (N-MORB) were recovered above the peridotites (**Robertson,** this volume), but the combined total thickness of these rocks is small (~9 m), which is highly inconsistent with the degree of melting documented in the underlying mantle. Based on mineral compositions, Müntener and Manatschal (2006) proposed that the depleted peridotites are subarc mantle, from which melt was extracted probably during the Paleozoic closure of the Iapetus and Rheic oceans in this region. If this is correct, the Site 1277 peridotites were part of the mantle lithosphere beneath the prerift continental crust before they were exhumed in approximately late Barremian time. Their prior history of extensive melting precluded generation of much magma when the peridotites were exhumed.

Compared to the Newfoundland samples, spinel peridotites on the Iberia margin are more fertile but show a large variation in degree of melting, and locally they exhibit evidence of equilibration in the plagioclase stability field (Müntener and Manatschal, 2006). On the basis of whole-rock major and platinum-group element (PGE) geochemical data, Hébert et al. (2001) argued that the rocks at Site 1070 (and also to the east at Site 1068) represent subcontinental mantle. Chemical heterogeneity of the peridotites might be explained by local synrift melt percolation and equilibration in the plagioclase stability field (e.g., Hébert et al., 2001), by subsolidus deformation related to extension (Beslier et al., 1996), or by both these processes (Cornen et al., 1996a). In any case, the amount of melt generated appears to have been insufficient to produce significant intrusions or flows, minor occurrences of which have been sampled only at Sites 899 and 1070 in TE1 and TE2. Basalt and diabase clasts in breccias overlying peridotite basement in Hole 899B have transitional to enriched mid-ocean-ridge basalt (E-MORB) characteristics (Seifert and Brunotte, 1996). At Site 1070, a 2.7-m-thick gabbro pegmatite lies at the top of the peridotites; the isotopes and modeled bulk composition of this gabbro also are similar to E-MORB in most respects (Beard et al., 2002).

The rift axis at the southern edge of the Newfoundland-Iberia rift (Fig. **F1**) was affected by plume magmatism from about the time of Chron M4 (latest Hauterivian; near the transition from TE1 to TE2) until some time after Chron M0 (Aptian). This magmatism formed the Southeast Newfoundland Ridge and probably part of the conjugate Gorringe Bank. It also created volcanic ridges that extended both to the south (forming the J Anomaly Ridge and the conjugate Madeira-Tore Rise) and to the north (Tucholke and Ludwig, 1982; Tucholke et al., 1989). It is uncertain how far the melt from the plume may have reached northward into the rift, but the amplitude characteristics of the magnetic J Anomaly (see **"Magnetic Anomalies in Zones of Transitional Extension: Character, Origin, and Implications,"** below) suggest that it could have had an effect all the way to the southern Galicia margin. This may be recorded at Site 899 where a plume contribution to the E-MORB basalts has been suggested by Seifert and Brunotte (1996). Such magmatism might help to explain the oceanic-type crust indicated by refraction data at the seaward ends of the SCREECH and IAM9 lines (Fig. **F1**).

# MAGNETIC ANOMALIES IN ZONES OF TRANSITIONAL EXTENSION: CHARACTER, ORIGIN, AND IMPLICATIONS

## Observed Magnetic Anomalies

The most prominent magnetic anomaly in the Mesozoic crust of the Newfoundland-Iberia rift is the so-called "J Anomaly." This anomaly occurs over crust from just seaward of anomaly M1 to just seaward of M0 in TE1, and it has been attributed to a thickened magnetic layer, to unusually high magnetization, or to both (Rabinowitz et al., 1978). The anomaly amplitude is highest (~1000 nT) over the Southeast Newfoundland Ridge and conjugate parts of the Madeira-Tore Rise at the southern edge of the rift, and it decays to both the north and south (Tucholke and Ludwig, 1982; Tucholke et al., 1989; Miles et al., 1996). Within the Newfoundland-Iberia rift, the anomaly amplitude reaches much lower, relatively constant levels (~200 nT) just north of the southern margin of Galicia Bank. Thus, if the unusual magnetization of the J Anomaly is associated with melt distributed from the plume under the Southeast Newfoundland Ridge, the plume magmatism may have affected the rift as far north as Galicia Bank and Flemish Cap.

Landward of the J Anomaly, the broad negative magnetic anomaly M3 is readily recognized. This anomaly has similar amplitude on both sides of the rift, but the amplitude is half that observed in the Central Atlantic Ocean (Srivastava et al., 2000). Magnetic anomalies older than M3 have very low amplitudes (~100 nT or less) (Miles et al., 1996) compared to their Central Atlantic counterparts. In addition, the amplitudes are higher on the Newfoundland side than on the Iberia side, where anomalies are barely visible in surface-ship data. Although pre-M3 magnetic anomalies are weak, they are parallel to subparallel to the margins and appear to be symmetrical across the rift. Thus they bear some of the characteristics of seafloor-spreading anomalies, which led Srivastava et al. (2000) to identify them as part of the M-series sequence.

**F7.** Magnetic anomalies in the Newfoundland-Iberia rift, p. 52.

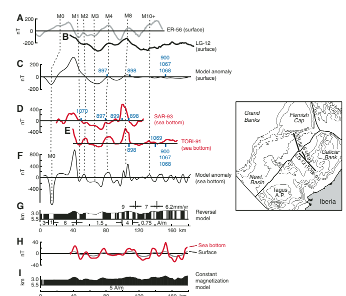


Figure **F7** shows examples of magnetic profiles from the conjugate margins of the rift near the southern margin of Flemish Cap and Galicia Bank. The conjugate sea-surface profiles in Figure **F7A** and **F7B** show general symmetry with one another. Deep-tow profiles along the western part of the Leg 149/173 drilling transect (Fig. **F7D, F7E**) show higher-amplitude anomalies. Comparison of the sea-surface and deep-tow magnetic anomalies with model calculations shows good correlation for anomalies M1 to ~M8 but no clear correlation for older anomalies. However, the amplitudes of even these observed anomalies can be fit only by using model magnetization values that vary widely (1–6 A/m) (Fig. **F7G**); this requirement raises questions about whether the anomalies are true seafloor-spreading lineations (Whitmarsh et al., 2001; Russell and Whitmarsh, 2003). For this reason, we reexamine the source of the magnetic anomalies in the next section. There we conclude that the anomalies are caused largely by serpentinization of exhumed mantle rocks and not by magnetization of igneous rocks emplaced during seafloor spreading.

## Source of Magnetic Anomalies

### The Role of Magmatism

As discussed previously, the composition of peridotites recovered at drill sites in zones TE1 and TE2 indicates that the mantle was strongly to moderately refractory and thus was unlikely to have produced significant volumes of melt as mantle was exhumed. Nonetheless, the drilled peridotites also show significant heterogeneity, and it is possible that relatively fertile mantle is present in the rift but remains unsampled. If we consider decompression melting of normal mantle, gabbroic and basaltic crust ~2 km thick might be produced (Bown and White, 1994), and we thus could expect substantial magmatic products to be present somewhere within the TE zones. For several reasons, however, it seems highly unlikely that these igneous rocks are present at shallow levels in the basement. First, it is clear from drilling results that there is a dearth of volcanism and magmatic intrusions in the uppermost basement, and this makes it doubtful that there are large underlying magmatic bodies at shallow levels. Second, seismic refraction results, summarized earlier, show little evidence for the presence of igneous rocks in the upper basement except in limited (younger) parts of TE2. Finally, persistent shallow magmatic intrusions should produce a "layer" with relatively uniform magnetization, counter to observations that magnetization intensity appears to be highly variable (Fig. F7).

An alternate possibility is that magnetic anomalies are caused by magnetic bodies located deep in the basement (~5–8 km), as has been proposed for the southern Iberia Abyssal Plain (Russell and Whitmarsh, 2003). To test this idea, Sibuet et al. (2007) conducted a combined inversion for the structural index and the source location using Euler deconvolution, which only uses derivatives of magnetic anomalies (Hsu, 2002). The primary result of this analysis is that the main magnetic sources are at depths <3 km within the basement and structural indexes are ~2, which means that the magnetic sources approximate cylinders that are probably horizontal and parallel to the magnetic lineations. Considering the above arguments that there probably is not a significant, shallow igneous source layer generating the magnetic anomalies, we must to look for other ways to account for the magnetization. As discussed below, magnetization of serpentinites in the upper basement provides a likely explanation.

## The Role of Serpentinization

The contribution of peridotites to magnetic anomalies becomes significant when the peridotites are strongly serpentinized (>75%) (Oufi et al., 2002). This degree of serpentinization may be widespread in the upper 1–2 km of exhumed mantle in the zones of transitional extension, where observed low seismic velocities would correlate with >75% serpentinization (e.g., Christensen, 2004; Chian et al., 1999; Lau et al., 2006b).

Direct evidence that serpentinites can contribute significantly to the magnetic field has been presented by Zhao et al. (Zhao, 1996, 2001; Zhao et al., this volume, 2006), who studied heating and cooling curves for serpentinized peridotite samples from both the Newfoundland and Iberia margins. In general, they found that yellow and brown peridotites have Curie temperatures near 420°C; this indicates the presence of maghemite and appears to account for low natural remanent magnetization (~1 A/m). Curie temperatures of gray and green peridotite samples are ~570°C, indicating presence of magnetite and accounting for high remanent magnetization (up to ~9 A/m). These values of remanent magnetization are comparable to or higher than those of oceanic basalts. All samples show inclinations close to those predicted for the paleolatitude of the drill sites, indicating that natural remanent magnetization (NRM) intensities probably represent primary magnetization.

Oxygen isotope profiles of serpentinized peridotites at Sites 1068 and 1070 show evidence for two phases of serpentinization (Skelton and Valley, 2000), and these may explain the differences between the gray-green and yellow-brown serpentinites. The first phase (>175°C) involved pervasive infiltration of water and bulk serpentinization that is interpreted to have occurred before the mantle was exhumed at the seafloor. The second phase (<50°–150°C) occurred at or close to the seafloor. Consequently, strong magnetization was first acquired at depth during an initial phase of serpentinization when gray-green serpentinites recorded the polarity of the ambient magnetic field. The second phase affected only the uppermost basement at the seafloor, giving rise to maghemitized yellow-brown serpentinites. Paleomagnetic results show that the yellow-brown "altered" serpentinites acquired their final magnetization later than the gray-green "fresh" serpentinites. Zhao (2001) suggested that the magnetization in the yellow-brown serpentinites zone was an overprint imposed during the Cretaceous Long Normal Superchron.

Serpentinites from Holes 897D, 899B, and 1070A on the Iberian margin yielded mean NRM intensities of 0.35, 1.8, and 1.6 A/m and mean magnetic susceptibility values of $2.1 \times 10^{-2}$, $2.9 \times 10^{-2}$, and $2.9 \times 10^{-2}$ SI units, respectively (Zhao, 2001), with variations in magnetic susceptibility mimicking those of the NRM intensity. The higher magnetization intensity in Hole 899B is consistent with the presence of the large magnetic anomaly recorded on the SAR-93 deep-tow profile (Fig. **F7**). Furthermore, several samples from Hole 899B retained nearly half their NRM intensity after 400°C demagnetization, suggesting that remanent magnetization in magnetites can contribute significantly to the strong 800-nT magnetic anomaly (Whitmarsh et al., 1996a). At all three sites a 40- to 75-m-thick yellow-brown "altered" zone with predominantly normal polarity and probably with maghemite overlies a gray-green "fresh" zone where magnetite is likely present (Zhao, 2001).

In yellow-brown and gray-green serpentinized peridotites at Site 1277 on the Newfoundland margin, measured Curie temperatures of 420°C and 550°–580°C indicate the presence of maghemite and magnetite, respectively. High NRM intensities (up to 9 A/m) in the gray-green serpentinized peridotites may help to explain the amplitude of observed sea-surface magnetic anomalies (**Zhao et al.,** this volume).

It is important to note that magnetic properties of extensively serpentinized peridotites can vary significantly depending on whether the peridotites have been maghemitized (either at the seafloor or in a permeable fault zone), whether the conditions of serpentinization allowed for the crystallization of iron-bearing silicates together with serpentine and magnetite, and whether conditions favored formation of a serpentine meshwork with thin veinlike concentrations or small and/or highly elongated magnetite grains (Oufi et al., 2002). Consequently, there is no reason to expect laterally uniform magnetization in serpentinized basement. From this perspective, the highly variable magnetization values used in the magnetic model of Figure **F7** are not unreasonable.

Based on the above discussion, we suggest that serpentinization may explain the magnetization responsible for the magnetic anomalies in the TE zones, exclusive of the interval around anomalies M1–M0 where plume magmatism from the Southeast Newfoundland Ridge may have had an effect. In this model, fluids penetrated deeply along margin-parallel faults and fractures caused by plate bending of the exhuming mantle, which allowed penetrative serpentinization to propagate as deep as 2–3 km. Pseudo-single domain magnetite grains in gray-green serpentinites oriented along the magnetic field below the Curie temperature (~570°C), recording the ambient magnetic field as the mantle cooled at crustal levels (Sibuet et al., 2007). Strong spatial variability in serpentinization is likely to have occurred, and this would have created vertically and horizontally heterogeneous crustal magnetization, consistent with forward modeling where the mean magnetization varies from 0.75 to 6 A/m (Fig. **F7**). As the serpentinized mantle was exhumed at the seafloor, the upper several tens of meters were altered by further reaction with seawater (Fisher-Trop reaction); this increased the degree of serpentinization, and magnetite was replaced by maghemite.

According to this process, seafloor age indicated by a magnetic anomaly would be the same as the age at which the serpentinized peridotite became exposed at the seafloor (i.e., the same relative timing as for anomalies generated from magnetization of igneous bodies), or it would somewhat predate the time of exposure (taking into account the time required for the magnetized serpentinite to be exhumed from some depth). At Site 1070, the ~128-Ma magnetic anomaly age is the same as the U/Pb age of 127 ± 4 Ma obtained for a gabbroic intrusion in the mantle peridotite (Table **T1**) (Beard et al., 2002). This is also true at Site 1277, where the magnetic anomaly age (~127 Ma) is the same as a 128 ± 3 Ma alkaline gabbro dike in peridotite (Jagoutz et al., submitted [**N1**]). The coincidence of magnetic and intrusion ages suggests that the intrusions were emplaced when the serpentinized mantle first acquired its magnetization. However, the error bars on the intrusion ages allow for the possibility that the magnetic ages are somewhat older (up to ~3–4 m.y.), as noted above. **Zhao et al.** (this volume) determined that the magnetic inclinations in the serpentinites at Site 1277 imply ~35° of counterclockwise tectonic rotation (viewed facing toward N010°E) after the magnetization was acquired. Considering all of the above, the following evolution of the basement ridge where Site 1277 was drilled seems likely:

1. Mantle peridotite was first serpentinized and thereby acquired its magnetization at subseafloor depths of up to ~2–3 km; this was accomplished as fluids penetrated along the master normal fault (and perhaps subsidiary faults) that uplifted the ridge.
2. Subsequent exhumation uplifted the ridge more than a kilometer above the surrounding seafloor. If the exhumation was along a normal fault on the east side of the ridge (as required to account for counterclockwise rotation) the basement morphology (Fig. **F3**) suggests minimum fault heave of ~5 km; if we add exhumation from subseafloor depths of ~2–3 km, the total horizontal extension probably was at least 7–8 km.
3. Minor magmatism, possibly enhanced by decompression melting, occurred during the exhumation.
4. Upon exposure at the seafloor, a thin “skin” of the exhumed mantle was further altered and maghemitized by low-temperature serpentinization.

If magnetized serpentinites account for the observed magnetic anomalies, it is conceivable that basement topography alone generates the anomalies. We investigated this possibility with the forward model shown in Figure **F7I.** We assumed a magnetized basement ~3 km thick with a flat bottom and a constant magnetization of 5 A/m, a value close to or somewhat larger than the mean value measured in serpentinized peridotite samples. Model results were computed for the sea surface and sea bottom (Fig. **F7H**). The results show that the contribution of basement topography is only ~10% of the amplitude of observed magnetic anomalies. A similar result was obtained by Russell and Whitmarsh (2003). We conclude that basement topography alone cannot explain the amplitude of magnetic anomalies.

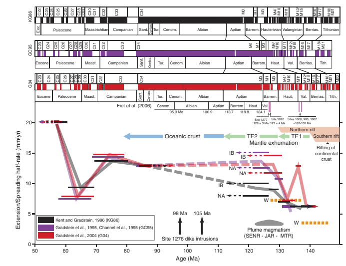
## Implied Extension Rates

If we assume that the magnetic anomaly picks by Srivastava et al. (2000) are correct (but taking note that anomaly picks older than M8 are highly debatable), we can calculate extension/spreading half rates as function of time as shown in Figure F8. The upper part of the figure shows three different timescales used in the calculations: (1) the Kent and Gradstein (1986) timescale (KG86), which was used by Srivastava et al. (2000); (2) the Gradstein et al. (1995) and Channel et al. (1995) timescales (GC95); and (3) the most recent Gradstein et al. (2004) timescale (G04). For comparison, we also show the Fiet et al. (2006) timescale, which is based on ages of glauconites sampled in the Vocontian Basin (southeast France) together with cyclostratigraphy previously obtained in southeast France and basins in central and southern Italy. The large differences between these timescales significantly affect calculated extension rates as shown at the bottom of the figure.

Extension half-rates calculated for Chron M8 (Hauterivian) and older are generally consistent at 6–7 mm/yr (Fig. F8). A similar rate (7.2 mm/yr) for the younger part of this period was derived by Whitmarsh et al. (2001) from geometrical reconstruction of faults (dotted line labeled W in Fig. F8), but they derived a lower rate (>3.5 mm/yr) for the older part. The major inconsistency in extension rate is for the G04 timescale, which shows a large excursion at M11 that probably is not real. Whether this reflects a problem with the timescale or with the anomaly identifications is unknown. In the Barremian after M8 (i.e., in the early part of TE2), rates increase to ~9–13 mm/yr, depending on the timescale used. Then from M0 to C34, rates are ~13 mm/yr or increase slowly to this value. Note, however, that there are no constraints within the long Cretaceous Magnetic Quiet Zone between M0 and C34, and there may be unknown rate variations in that interval.

The 6- to 7-mm/yr rates characterize TE1 and they are ultraslow, comparable to the slowest rates on Earth (i.e., Gakkel Ridge in the Arctic Ocean [Cochran et al., 2003]), and the ~9- to 13-mm/yr rates in TE2 are comparable to those at the slow end of slow-spreading mid-ocean ridges. It is natural to look for lithologic and morphologic analogs at mid-ocean ridges that spread at these rates. For example, on the Southwest Indian Ridge where the spreading half-rate is ~7 mm/yr, ~40% of the crust has a smooth or corrugated surface that may be formed by long-lived normal faults (Cannat et al., 2006) and it is composed of peridotites ± gabbros and basalts (Seyler et al., 2003); the remaining 60% corresponds to apparently more volcanic and normally faulted seafloor that is more like that of slow-spreading ridges. However, it must be kept in mind that the critical control on morphology and composition of newly emplaced seafloor is not extension rate but melt supply. This is dramatically demonstrated, for example, in the Australia-Antarctic Discordance where the half-spreading rate is 75 mm/yr (i.e., intermediate rate), yet because of very low melt supply the morphology is like that of slow- to ultraslow-spreading ridges (Okino et al., 2004). Furthermore, for some uncertain distance seaward into the TE zones (but apparently seaward past anomaly M1, as discussed earlier), subcontinental mantle lithosphere was being exhumed and its properties may have influenced morphology and mantle melting patterns in much different ways than asthenospheric mantle beneath mid-ocean ridges. Thus, there are good reasons to be cautious about making comparisons solely on the basis of extension/spreading rate.

F8. Half-rates for extension/spreading calculated using three different timescales, p. 53.



# POSTRIFT MAGMATISM

As discussed earlier, rifting of strong, possibly subcontinental lithospheric mantle may have persisted until latest Aptian to earliest Albian time (ca. 112–110 Ma). We here consider this to be the end of the second period of transitional extension (TE2). In the following postrift period, extension localized at a well defined spreading axis and magmatic seafloor spreading ensued.

## Sills at Site 1276

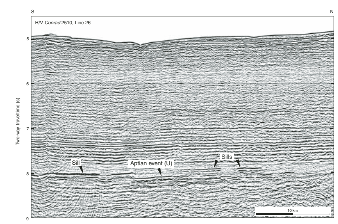
The most significant postrift magmatism, other than seafloor spreading, encountered in the Newfoundland Basin is in the form of two alkaline diabase sills drilled at Site 1276, which is located in TE1 probably near the seaward edge of thinned continental crust (Van Avendonk et al., 2006) (Figs. **F1, F4**). The main body of the lower sill, in which the hole bottomed, is >18 m thick (1719.2 to >1736.9 mbsf). It is accompanied by at least four other thin (3–31 cm) sills that were cored 7–14 m above the main sill and may be apophyses of the sill (Tucholke, Sibuet, Klaus, et al., 2004). The upper sill is ~10 m thick (1612.7–1623 mbsf). Both sills were intruded into uppermost Aptian to lowermost Albian sediments at an estimated ~90–160 m and 200–270 m above basement, respectively, based on interpretation of the seismic reflection record (Fig. **F3**) and seismic velocities at the drill site (see **Shillington et al.,** this volume). Whole-rock $^{40}Ar/^{39}Ar$ radiometric dates indicate that the upper sill was intruded first, at 104.7 ± 1.7 to 105.9 ± 1.8 Ma (mid-Albian on the Gradstein et al., 2004, timescale), and the lower sill was subsequently intruded at 95.9 ± 2.0 to 99.7 ± 1.8 Ma (early Cenomanian) (Table **T1**) (Hart and Blusztajn, 2006). Considering these ages in the context of the plot of sediment age vs. subbottom depth for Site 1276 (Tucholke, Sibuet, Klaus, et al., 2004), the upper sill was intruded beneath a minimum of ~260 m of overburden, and the lower sill was intruded beneath at least ~590 m of overburden (neither value accounts for postintrusion compaction). Karner and Shillington (2005) estimated intrusion depth of 0–556 mbsf based on reconstruction of porosity-depth relations, which is similar to these values.

Geochemical studies of the sills (Hart and Blusztajn, 2006) indicate that they are alkali basalt-hawaiite (differentiated basanite) and that they have experienced significant alteration, as evidenced for example by extremely high Ba contents (up to 4900 ppm) in abundant acicular secondary apatite, particularly in the upper sill. Isotopic and trace element data show that the two sills are not strictly co-genetic but originated from separate magmatic events, consistent with their ~7.5-m.y. age difference. The source of the magma is uncertain, although the sills clearly were not derived from MORB-type asthenospheric upper mantle. Hart and Blusztajn (2006) concluded from the isotope geochemistry that the magma probably was derived from an enriched plume source in the mantle, possibly contaminated by a component of continental material.

The age of the lower sill is not significantly different from that of a trachyte from Scruncheon Seamount (97.7 ± 1.5 Ma) within the Newfoundland Seamounts 200 km south of the drill site (Table **T1**) (Sullivan and Keen, 1977). Trace element geochemistry of the sills is also similar to that of basalts from the Newfoundland Seamounts (Hart and Blusztajn, 2006). These commonalities suggest a shared source for the igneous rocks. Duncan (1984) proposed that at ~100–90 Ma the Newfoundland Seamounts and the southern Newfoundland Basin passed over the Madeira and Canary hotspots, respectively, with the southern Newfoundland Basin also crossing the Azores hotspot some 20 m.y. later. Thus, both available geochemical data and plate-kinematic data are consistent with the idea that plume-related magmatism was responsible for intrusion of the sills.

The extent to which other parts of the Newfoundland Basin were affected by sill injection is unclear, but it appears that this magmatism may have been a basin-wide phenomenon. At Site 1276, the depth of the upper sill is coincident with the U reflection or Aptian event (Fig. F3), and the lower sill also coincides with a strong, deeper reflection (**Shillington et al.,** this volume). Few features in the seismic reflection record are clearly diagnostic of sills. Minor disruptions of the U reflection 2–3 km to either side of the drill site might represent locations where melt was injected into the basal sedimentary column (Fig. F3). Such disruptions are not uncommon elsewhere in the basin. In other cases, short segments of the U reflection show increased amplitude (Fig. F3, left side) and could mark sill locations. At a few other places such short, high-amplitude reflection segments occur near the U reflection but are separated from it (Fig. F9), and these very likely are sills (Tucholke et al., 1989).

**F9.** MCS profile showing sills at the level of the U reflection, p. 54.



On a broader scale, the U reflection itself has unusually high amplitude throughout most of the basin, to the degree that it commonly masks most deeper structure (Tucholke et al., 1989); it also is generally much stronger than the correlative "orange reflection" (Aptian event) on the conjugate Iberia margin. The strength of the U reflection is something of a puzzle because at Site 1276 there is little evidence for unusually high impedance in the sediments at that level (excluding sediments that have been hydrothermally altered by the sill injection) (Tucholke, Sibuet, Klaus, et al., 2004). We would expect to see such sediments represented at Site 1276 if they give rise to the reflection throughout the basin. On the other hand, if sills were consistently injected at this level, they could explain the strength of the reflection. Laterally extensive sills are well known in the geological record. For example, the Carboniferous Great Whin Sill in northern Britain extends over 5000 km$^2$ (Johnson and Dunham, 2001). This sill mainly follows bedding planes, with occasional splitting; it has been suggested that once magma has reached a level of hydrostatic equilibrium at shallow subbottom levels, it can flow laterally with little frictional resistance, much like subaerial flood basalts (Francis, 1982). Considering this mechanism, multiple sills could have been emplaced at a relatively uniform depth in the level-bedded sediments of the Newfoundland Basin. If this was the situation for the older event (the shallower sill at Site 1276), then younger sills might be mostly restricted to the underlying section, as observed at Site 1276. In the future, detailed analysis of lateral changes in amplitude of the U reflection and the underyling section, combined with structural interpretations, may help to resolve the occurrence and distribution of sills.

If there was widespread intrusion of sills, and their emplacement was related to passage of the Newfoundland Basin over one or more mantle plumes as discussed above, it may help to explain why the U reflection is much stronger than the equivalent reflection event on the Iberia margin. According to Duncan (1984), the axis of the Mid-Atlantic Ridge did not pass over the hotspots until ~70 Ma, so the Iberia plate was isolated from their possible magmatic effects until ~40 m.y. after the

Aptian event. Even then, the hotspots were located under the southernmost edge of the Iberia plate and the northwestern African plate, so they would have had little effect on the Iberia side of the rift.

## Hydrothermal Metamorphism around the Sills

The older, upper sill was injected into calcareous, turbiditic grainstones, siltstones, and claystones (see Tucholke, Sibuet, Klaus, et al., 2004, for details). Its upper contact is preserved as a chilled margin juxtaposed against a thin (<1 cm) layer of baked sediment. Prominent megascopic effects of hydrothermal alteration are observed for ~66 cm above the contact, in the form of porphyroblasts composed of pure stoichiometric calcite (T. Pletsch, pers. comm., 2007; note that the porphyroblasts were incorrectly reported in the Leg 210 *Initial Reports* as consisting of albite, quartz, alkali feldspar, and magnesian chlorite, which actually comprise the surrounding sediment). The porphyroblasts decrease in size and frequency with distance from the contact. A subvertical, crenulated vein filled with calcite and pyrite in this interval is interpreted to have precipitated from fluids circulating through fractures as the sill was injected. The crenulation, together with compaction of sedimentary laminae around the porphyroblasts, provides evidence that the sill was injected into relatively unconsolidated sediments. Later compaction shortened the vein by ~21% (Karner and Shillington, 2005). The lower contact of the sill was not recovered; although the sill shows a chilled margin, the subjacent sediments show no thermal overprint in smear slides and it is likely that a section of the sedimentary record was not recovered in the cores.

Pross et al. (2007) used sporomorph colors to assess thermal alteration above the sill. They determined that thermal effects extend upward ~20 m, but the strongest alteration occurs in an aureole within ~4.2 m of the sill. They also observed reduced thermal alteration in an interval ~5–6 m from the sill, which suggests small-scale convection in the circulating hydrothermal fluids.

Near the bottom of Site 1276, the upper contact of the younger, lower sill was not recovered, and most contacts of the overlying thin sills are missing or poorly preserved because of drilling disturbance. The principal evidence for hydrothermal metamorphism is observed in Sections 210-1276A-98R-2 to 98R-CC between a pair of thin sills at ~1711 mbsf and the top of the main sill at ~1719 mbsf. Here, a ~2-m section of claystone, mudstone, and minor calcareous sandstone was affected by contact metamorphism. The alteration produced a variety of effects in different beds, including (1) hornfels textures with porphyroblasts that are visually similar to those overlying the upper sill but that have not been chemically analyzed; (2) nearly total replacement by albite, sudoite, nonstoichiometric calcite, pyrite, and zeolite; (3) color alteration; and (4) thermal alteration of organic matter (Tucholke, Sibuet, Klaus, et al., 2004). Veins in the sediment contain stoichiometric calcite with pyrite, and extremely light carbon isotope composition of the carbonates suggests a carbon source in the adjacent organic-rich sediments (T. Pletsch, pers. comm., 2007).

Thus overall, strong hydrothermal metamorphic effects associated with the sills appear to have been very restricted. They occurred only within a few meters of the two major sills and within a few centimeters to decimeters of the minor, thin sills. However, lesser effects may have extended 10–20 m from the sills.

## Other Postrift Magmatism

Other postrift magmatism, unrelated to seafloor spreading, is present within the Newfoundland-Iberia rift, but it is widely scattered and very limited in volume (Table T1). On the Newfoundland margin, a 96-Ma porphyritic monzodiorite dike was penetrated in the Emerillon C-56 well (Jansa and Pe-Piper, 1986), and alkaline dikes and volcanic rocks dating to ~100–70 Ma are documented on the southern Iberia margin (Pinheiro et al., 1996). In the deep basin, minor igneous rocks associated with basement peridotites have a similar wide range of apparent ages at ODP Sites 1070 and 1277 and along the northwest Galicia margin (Table T1). However, ages on all these samples were determined on plagioclases, and it is possible that the ages do not represent original magmatism but were reset by late-stage hydrothermal alteration.

Jansa and Pe-Piper (1988) attributed the 96-Ma dike in the Emerillon well to intraplate fracture or fault reactivation related to opening of the Labrador Sea, and they suggested that most such igneous activity occurred at times of significant plate-motion changes. However, if the widely scattered (>30 m.y.) igneous ages in the deep Newfoundland-Iberia rift (Table T1) represent true ages of magmatic events, then minor magmatism would appear to be relatively common and not necessarily linked to major plate-tectonic events. Furthermore, these ages could indicate that intraplate stress locally reactivates off-axis faults more often than heretofore recognized, particularly in areas of exhumed peridotite where weak, serpentinized faults can fail easily. Where the faulting was combined with a thermally thin plate or underlying plume activity, widely distributed minor magmatism might result. It is highly unlikely that this would ever be recognized on a normal mid-ocean ridge where igneous rocks are abundant, whereas these rocks stand out uniquely in peridotite basement and thus tend to be extensively studied and documented.

A separate group of postrift igneous rocks has been documented on Gorringe Bank (Table T1). This feature lies along the Iberia/Africa plate boundary, which is the former Newfoundland-Gibraltar Fracture Zone. According to plate-kinematic arguments, this boundary was locked from Late Cretaceous to middle Eocene time as Iberia moved with Africa, but subsequently it has been active and notably compressive in its eastern (Gorringe) part (Srivastava et al., 1990b). On the other hand, tilted sedimentary strata around Gorringe Bank and related highs suggest compression across the boundary since Late Cretaceous time (Féraud et al., 1982). The Late Cretaceous magmatic episodes indicated by isotopic ages of igneous rocks from this feature may reflect intermittent effects of plate-boundary tectonics. Féraud et al. (1986) argued that a primary alkaline magmatic event at ~65–67 Ma may have reset older rocks to ~111 Ma and ~75–84 Ma ages. However, the 77-Ma U-Pb zircon age on ferrogabbro reported by Schärer et al. (2000) (Table T1) appears to represent a primary igneous event. The ~65- to 67-Ma alkaline event suggested by Féraud et al. (1986) could reflect the influence of the Madeira and Canary hotspots, over which the region probably passed beginning at ~70 Ma (Duncan, 1984).

# SEDIMENTARY RECORD IN THE DEEP BASINS

## Prerift Sedimentation

Prerift sediments have been cored in Hole 639D on the western margin of Galicia Bank, where Tithonian shallow-water limestone and marlstone with lesser sandstones and claystones are capped by Tithonian to ?Berriasian dolomites (Boillot, Winterer, Meyer, et al., 1987). These were interpreted by Boillot et al. to have been deposited under shelf to lagoonal conditions prior to rifting of the underlying crust.

## Synrift Sedimentation

### Tithonian

In contrast to Hole 639D, Tithonian sediments on the southern edge of Galicia Bank at Sites 1065 and 1069 under the southern Iberia Abyssal Plain are dominated by gray turbiditic to hemipelagic clays and claystones, with thin conglomerates and clasts of shallow-water limestones and ?Paleozoic metasediments (Whitmarsh, Beslier, Wallace, et al., 1998). At Site 901 the Tithonian sediments are olive-black clays that are locally silty, dolomitic, and rich in plant debris (Sawyer, Whitmarsh, Klaus, et al., 1994). These sediments differ markedly from the coeval red to gray-green limestones of the Cat Gap Formation, which was deposited farther south in the deep western North Atlantic Basin (Fig. **F4**). The dearth of carbonate and presence of turbidites in the Newfoundland-Iberia rift suggest that deposition was at least at moderate depths, rather than in shallow water. Studies of benthic foraminifers at Site 901 and calcareous nannofossils at Sites 1065 and 1069, together with evidence from the lithofacies, indicate that deposition was under neritic, dysoxic conditions in a marginal shelf basin at Site 901 (Collins et al., 1996) and in a restricted interior basin with little communication with the open ocean at Sites 1065 and 1069 (Concheryo and Wise, 2001). Thus, the crust along the southern Galicia margin probably was rifting by Tithonian time, creating a series of extensional graben that were relatively isolated from the main North Atlantic Basin. Thin laminae of nannofossil ooze appeared at Sites 901 and 1065 in the latest Tithonian, suggesting increasing connections with the open ocean, although dysaerobic conditions still persisted at the seafloor (Concheryo and Wise, 2001).

### Berriasian to Early Barremian

The Early Cretaceous (Berriasian to early Barremian) in the main North Atlantic Basin was a period characterized by a deep calcite compensation depth (CCD) and by deepwater deposition of calcareous sediments comprising the Blake-Bahama Formation (Fig. **F4**) (Jansa et al., 1979). The same conditions appear to have been present within the Newfoundland-Iberia rift. Upper Berriasian to lower Valanginian nannofossil chalks were deposited at Site 1069 and Valanginian–Barremian breccias containing basement clasts in a chalk matrix were deposited at Site 1068 to the east (Whitmarsh, Beslier, Wallace, et al., 1998), while upper Hauterivian–lower Barremian nannofossil limestones with interbedded mudstones were deposited at Site 398 (Sibuet, Ryan, et al., 1979). On the western Galicia margin to the north, ?Berriasian, presumably shallow-water dolomites were rifted and succeeded by deposition of Valanginian–Barremian marlstones in deepening rift basins (Boillot, Winterer, Meyer, et al., 1987). These were augmented by input of silty and sandy turbidites (in places arkosic) during the Valanginian–Hauterivian as the rift topography was eroded.

## Barremian–Aptian

A marked change in basin conditions occurred during the Barremian as the CCD shoaled and the deep basin became less oxygenated. Sediment facies changed to predominantly gray-green to black hemipelagic claystones (locally carbonate or silica rich) that commonly were interrupted by coarser-grained turbidites from the continental margins. These sediments are ubiquitous on both the Newfoundland and Iberia margins, even up through the Cenomanian, and they are equivalent to the Hatteras Formation in the main North Atlantic Basin to the south (Fig. **F4**) (Tucholke and Vogt, 1979; Jansa et al., 1979).

A notable facies consisting of breccias occurs atop the peridotite basement highs drilled at Sites 897, 899, 1070, and 1277 (Sawyer, Whitmarsh, Klaus, et al., 1994; Whitmarsh, Beslier, Wallace, et al., 1998; Tucholke, Sibuet, Klaus, et al., 2004). These breccias contain varying proportions of sedimentary, serpentinite, gabbro, and basalt clasts and appear to have been deposited by mass flows. At least some of these deposits probably were sourced from local topography (Gibson et al., 1996), although it has been argued that at Sites 897 and 899 they were derived from more distal sources before the basement blocks were faulted and uplifted (Comas et al., 1996). Ages of the deposits are late Barremian to late Aptian (Site 897), early Aptian (Site 899), and late Aptian (Site 1070). No age-diagnostic fossils were found in the mass flows at Site 1277, so it is uncertain when they were deposited (Tucholke, Sibuet, Klaus, et al., 2004).

The widespread reflection marking the Aptian event and the interpreted cessation of rifting near the Aptian/Albian boundary has been drilled in only three holes where it is intact (Sites 398, 641, and 1276). In other holes it correlates with a significant hiatus or it is not present, primarily because the holes were drilled on basement highs. At Site 398, Aptian sediments below the reflection are black to green-gray mudstones interbedded with debris flows, mud-supported conglomerates, and turbidites, contrasting with overlying Albian claystone and mudstone that is dominantly laminated or burrowed (Sibuet, Ryan, et al., 1979). Tucholke et al. (2007) suggested that this sharp reduction in mass flows was caused by subsidence of source areas when in-plane stresses relaxed at the end of rifting. At Site 641, Aptian sediments are dominated by marlstones and limestones, and they are succeeded by Albian black to gray-green claystones (Boillot, Winterer, Meyer, et al., 1987). Because Aptian sediments elsewhere in the rift are dominated by dark claystones, this change probably reflects subsidence of the seafloor, possibly coupled with a further rise in the CCD. At Site 1276 there appears to be no significant change in lithology that can be associated with the Aptian event (U reflection) (Tucholke, Sibuet, Klaus, et al., 2004). As discussed above (see **"Sills at Site 1276,"** p. 21), the strong reflection at this site is due to the presence of the upper diabase sill.

It is noteworthy that the reflectivity of the Aptian event and the underlying sediments around many of the peridotite ridges on the Galicia margin is exceptionally strong adjacent to the ridges but fades with distance from the ridges (Fig. **F6**) (Tucholke et al., 2007). Tucholke et al. (2007) proposed that the highly reflective sediments were deposited by extensive mass wasting from the weak, serpentinized peridotites; the mass wasting was strongly curtailed when in-plane stresses were relaxed at the end of rifting and the peridotite ridges subsided. Overall, the reduction in mass wasting at the end of rifting (leading to deposition of finer-grained sediments), combined with subsidence (and thus a change from deposition of calcareous to noncalcareous sediments), may largely account for the pronounced reflectivity and rift wide distribution of the reflection that marks the Aptian event.

## Postrift Sedimentation

### Albian to Cenomanian

Gray-green to black mudstones and claystones with lighter, commonly marly intervals continued to accumulate until the end of Cenomanian to early Turonian time, as they did in the North Atlantic Basin farther to the south (Fig. **F4**). In the Newfoundland-Iberia rift, these sediments are represented primarily at Sites 398, 641, and 1276.

At Site 1276 the lower (1502–1719 mbsf) and upper (1067–1130 mbsf) parts of this succession are dominated by muddy gravity-flow deposits with a wide variety of sedimentary structures, whereas the central part is largely burrowed hemipelagic mudstone (Tucholke, Sibuet, Klaus, et al., 2004; **Shirai et al,** this volume) (see **Sawyer and Fackler,** this volume, for an automated method used to associate core lithology to measured properties). Sediments in the middle and upper parts of the succession accumulated at rates of ~18–23 m/m.y. whereas the lower part was deposited at much faster rates of ~69 m/m.y. to as much as 105 m/m.y. (Tucholke, Sibuet, Klaus, et al., 2004; **Ladner,** this volume). Overall dark color of the sediments and lack of burrowing in the upper and lower intervals indicates low-oxygen conditions (dysoxic to anoxic) at and/or below the seafloor, and this is supported by elevated concentrations of redox-sensitive trace metals in the sediments (**Robertson,** this volume; **Arnaboldi and Meyers,** this volume). Reducing conditions in the central, more burrowed part of the section were either less severe or restricted mostly to the subsurface, or both.

Total organic carbon (TOC) in the Albian–Cenomanian sequence at Site 1276 varies from 0 to 11.7 wt%, with generally low values (~1 wt% or less) in the upper part of the sequence increasing to normally >1 wt% in the lower, Albian part (Tucholke, Sibuet, Klaus, et al., 2004; **Arnaboldi and Meyers,** this volume). Most TOC values >2 wt% occur in black shales represented by oceanic anoxic events (OAEs). Five OAEs have been recognized: (1) the Cenomanian–Turonian OAE 2 ("Bonarelli" event); (2) the "Mid-Cenomanian Event;" and (3) to (5) the Albian OAE 1b, OAE 1c, and OAE 1d events (Fig. **F4**). In addition, another possible OAE with total organic carbon as high as 3.2 wt% was detected in the middle Albian section. **Arnaboldi and Meyers** (this volume) report that the organic carbon throughout the section was derived from both terrestrial and marine sources, although there appears to be a dominance of marine carbon in OAE 2 and OAE 1b. In addition, $\delta^{15}N$ is inversely correlated with total organic carbon in OAEs, suggesting that the layers contain marine carbon deposited when nitrogen fixation was abnormally elevated in the surface waters.

In gray-green to black claystones of the same age on the Iberia margin, TOC values are predominantly low at Site 398 (<1 wt%) (Sibuet, Ryan, et al., 1979), but higher average values (2–4 wt%) are present at Site 641 (e.g., Boillot, Winterer, Meyer, et al., 1987; Dunham et al., 1988). As at Site 1276, the TOC is a mixture of terrestrial and marine carbon, with higher values generally occurring in black layers. OAE 2 is present at both Sites 398 and 641, and it is dominated by marine carbon, with TOC values as high as ~10–13 wt% (Dunham et al., 1988; Thurow et al., 1988).

Terrestrial sources provided most of the sediment deposited in the Newfoundland Basin during this period (**Robertson,** this volume), and the primary source was probably the adjacent Grand Banks. Studies of fluid inclusions in detrital quartz grains from this and the overlying sedimentary sections (**Shryane and Feely,** this volume) show that the fluid signatures are typical of those in detrital quartz derived from granitic sources. **Hiscott** (this volume) studied paleoflow directions determined from ripples and grain fabric in turbidites and deduced that the currents probably flowed primarily to the north-northeast. This suggests a source area on the southeastern Grand Banks in the area of the Avalon uplift. This area was elevated from at least Middle Jurassic to Cenomanian time (Jansa and Wade, 1975; Grant et al., 1988), as evidenced by the long-ranging hiatus represented by the Avalon unconformity in this area. Surprisingly, there is little evidence for flows from Flemish Cap to the north, perhaps because of the much smaller area of this feature or because it was submerged at the time. Sandstone compositions (**Marsaglia et al.,** this volume) are consistent with one major terrigenous source on the Grand Banks, and relatively low potassium feldspar contents distinguish the Site 1276 sandstones from those deposited off Iberia at the same time. The composition resembles a "recycled orogen" provenance (Dickinson, 1985), suggesting that the sandstones were recycled from foreland basin sediments originally deposited during the Paleozoic (**Marsaglia et al.,** this volume). White micas from turbidites in the Site 1276 Albian sequence have $^{40}Ar/^{39}Ar$ ages between 250 and 340 Ma (**Wilson and Hiscott,** this volume). Wilson and Hiscott interpret the direct source of these micas to be the Meguma Terrane at the southeast end of the Grand Banks, possibly originally emplaced during the Acadian orogeny (~375–415 Ma) but possibly later reactivated and metamorphosed during the Alleghenian orogeny (~260–350 Ma). However, they also note that the original provenance could have been Iberia if Iberia sediments were shed westward into foreland basins (now under the easternmost Grand Banks) during the Alleghenian orogeny.

## Turonian to Late Paleocene

Deep-basin conditions in the rift changed dramatically near the Cenomanian/Turonian boundary from dysoxic/anoxic to well oxygenated, and this oxygenated state mostly persisted through the late Paleocene. The deep basin remained well below the CCD. The correlative facies in the main North Atlantic Basin is the dominantly reddish to multicolored shales of the Plantagenet Formation (Jansa et al., 1979). The cause of the change to well ventilated deep basins and presumably increased deep circulation is uncertain, but it has been suggested that it relates to a deep-basin connection that became established between the North and South Atlantic at about this time (Tucholke and Vogt, 1979).

At Site 1276, change to a well oxygenated seafloor occurs in the middle to upper Turonian. It is marked by a change to deposition of reddish, highly bioturbated, siliciclastic muddy sandstones and sandstones, both of which have very low TOC values (0–0.4 wt%). In addition to the low TOC values, strongly oxidizing conditions are indicated by locally high concentrations of MnO (**Robertson,** this volume). These sediments were deposited at very low rates (<2 m/m.y.) until earliest Campanian time (Tucholke, Sibuet, Klaus, et al., 2004); there is poor biostratigraphic control in this interval and hiatuses may be present. There was no significant change in sandstone detrital modes in the Upper Cretaceous section compared to the deeper Albian–Cenomanian sediments (**Marsaglia et al.,** this volume), and white mica in the Turonian sediments shows the same age distribution and presumed source area as the underlying Albian section (**Wilson and Hiscott,** this volume). Thus, the sandstones continued to be deposited from gravity flows from the Grand Banks, but the extremely low sedimentation rates indicate that this source was highly attenuated, probably because of elevated eustatic sea level in the Late Cretaceous (Haq et al., 1988). The general scarcity of fines also indicates that the sediments were strongly winnowed by abyssal currents.

Reddish to multicolored sediments were also deposited on the conjugate Iberia margin at this time, although they are mostly pelagic claystones to hemipelagic silty claystones (Sibuet, Ryan, et al., 1979; Boillot, Winterer, Meyer, et al., 1987; Whitmarsh, Beslier, Wallace, et al., 1998). In contrast to Site 1276, these sediments show little indication of modification by deep currents. Thus, the currents appear to have been westward intensified, primarily affecting only the Newfoundland margin.

The lowermost Campanian to upper Paleocene sediments at Site 1276 consist of low-TOC, largely unburrowed reddish to multicolored claystones and mudstones that accumulated at rates of ~4–6 m/m.y.; they contain a low proportion (~20%) of coarser, more calcareous beds that were introduced mostly by mass flows and are concentrated in the lower part of the section (Tucholke, Sibuet, Klaus, et al., 2004). The carbonates herald increasing input of calcareous sediments that predominated in the late Paleocene and Eocene. The shift away from deposition of coarse siliciclastic sediments suggests declining input of terrigenous sediment from the adjacent Grand Banks, even though sedimentation rates increased. This apparent paradox may be explained if we consider the abundant fines in the sediments. These were not winnowed away as they were in the underlying sandstones, suggesting a marked reduction or cessation of vigorous abyssal circulation beginning in the early Campanian. The lower and particularly the middle Paleocene part of the section contain numerous dark gray to black beds (Fig. **F4**), most of which are associated with gravity flows. Although these sediments are low in TOC (mostly <1 wt%), the dark colors indicate reducing conditions, at least below the sediment surface, and they also support the idea that deep circulation was sluggish and the basin was not well oxygenated during this time. A similar observation was made at DSDP Sites 386 and 387 on the Bermuda Rise (Tucholke and Vogt, 1979), which indicates that this phenomenon may have extended throughout the deep basins of the North Atlantic.

Campanian to Paleocene sediments on the Iberia margin at Sites 1068 and 1069 are broadly similar to those at Site 1276 (Whitmarsh, Beslier, Wallace, et al., 1998), but at the ~1-km shallower seafloor of Site 398 the sediments are dominated by red-brown marly chalk (Sibuet, Ryan, et al., 1979). Although dark gray-green beds associated with gravity flows are present at these sites, they are not as dark as the Paleocene beds at Site 1276 and thus appear not to be as reduced. All these sites were drilled on basement highs, and thus the sediments were deposited on shallower, possibly more oxygenated seafloor than that at Site 1276.

At many drill sites in the main North Atlantic Basin the upper Maastrichtian is represented by the carbonate-rich Crescent Peaks Member of the Plantagenet Formation (Fig. F4), which is interpreted to have been deposited during a deep excursion of the CCD (Tucholke and Vogt, 1979). This facies is represented on the Iberia margin by the marly nannofossil chalks at Site 398 and by nannofossil chalks at Site 1068, but it is not present at Sites 1069 and 1070 (Sibuet, Ryan, et al., 1979; Whitmarsh, Beslier, Wallace, et al., 1998). The last three sites all penetrated this stratigraphic interval at ~5900 m below present sea level, so the same facies might be expected at each site. However, because the sites were drilled on basement highs it's possible that highly calcareous sections originally deposited at Sites 1069 and 1070 were later removed by mass wasting. Maastrichtian sediments at Site 1276, penetrated at ~5550 m below present sea level, are dominantly mudstones and claystones similar to underlying and overlying beds. Sedimentation rates did not change during this period (Tucholke, Sibuet, Klaus, et al., 2004), so a pulse of pelagic carbonate deposition would not have been diluted and masked by increased sediment input from shallow water. Thus, it appears that there was significant asymmetry in CCD depth between the two margins of the rift, with Site 1276 below the CCD while deeper seafloor on the Iberia margin was above it.

One unusual sedimentary component, consisting of mafic and felsic volcaniclastic debris within turbidites in Core 210-1276A-15R, was recognized in the middle Paleocene (~56 Ma) (**Urquhart et al.,** this volume) section at Site 1276. Mica $^{40}Ar/^{39}Ar$ ages from the turbidites range from 55 to 74 Ma (plus one 186-Ma sample), with 8 of the 13 ages dating to 55–61 Ma (**Wilson and Hiscott,** this volume). **Marsaglia et al.** (this volume) also detected minor volcanic debris in samples ranging from Santonian to Eocene age at Site 1276, and they found hitherto unrecognized mafic and felsic(?) volcanic detritus in Iberia cores at Site 897 (Upper Cretaceous?), Site 1068 (lower Eocene), and Site 1069 (upper Paleocene). There was scattered magmatic activity around the rift during the Late Cretaceous to Paleocene (Table T1), but there is little evidence for subaerial volcanism that could have dispersed significant amounts of ash over the basins. Thus, it seems likely that most of these occurrences were from local sources. A significant exception may have been related to the opening of the Norwegian-Greenland Sea, which produced voluminous amounts of volcanic rocks (Sinton and Duncan, 1998; Hopper et al., 2003) and might have been a source of subaerial volcaniclastic ash at ~55–56 Ma (Storey et al., 2007). Given the indications for weak bottom circulation in the middle Paleocene noted above, however, it seems unlikely that volcaniclastic debris would have been transported into the Newfoundland-Iberia rift by deep currents.

In contrast to the findings of **Marsaglia et al.** (this volume), **Robertson** (this volume) conducted geochemical studies of the fine-grained "background" sediments enclosing the turbidites at Site 1276 and found that volcaniclastic debris appears to be isolated in the turbidites, indicating derivation from a local source rather than from more widely dispersed volcanism. Concentration of micas in the turbidites and their relatively limited age grouping within the Paleocene (**Wilson and Hiscott,** this volume) also are consistent with a local provenance. One possible source area is the Newfoundland Seamounts to the south, which would be consistent with transport directions of turbidity currents (albeit in the Albian section) that were determined by **Hiscott** (this volume). Although the only known volcanism at the seamounts occurred in the Cenomanian (Table T1), it is possible that volcanism continued there or recurred at a later time. It is also possible that the volcaniclastic debris was derived from undocumented sources on the Grand Banks. Previously unsuspected igneous rocks have been penetrated in exploration wells there (e.g., Cenomanian dikes in the Emerillon C-56 well) (Table T1), and there could be other, younger intrusive or volcanic rocks present that are yet to be discovered.

## Late Paleocene to Middle Eocene

In the main North Atlantic Basin, upper Paleocene to middle Eocene sediments are characteristically biosiliceous, cherty claystones that characterize the Bermuda Rise Formation (Fig. F4). The time equivalent sequence at Site 1276, however, contains only a minor biosiliceous component. It consists of carbonate grainstones and marlstones with subordinate mudrock and represents a record dominated by deposition from mass flows, largely below the CCD (Tucholke, Sibuet, Klaus, et al., 2004). The lower part of the sequence is dominated by siliciclastic turbidites, and the proportions and types of siliciclastic grains indicate sources similar to those of the underlying section (**Marsaglia et al.,** this volume). In the middle Eocene section, white-mica ages continue to group at 250–340 Ma, but a new group (402–428 Ma) of white micas appears, creating a bimodal mica-age distribution (**Wilson and Hiscott,** this volume). This distribution suggests addition of another source of siliclastic detritus, either because of shifting sediment supply routes or because of erosional exposure of older rocks. The upper part of the upper Paleocene to middle Eocene sequence shows a strong shift to largely calcareous (grainstone) turbidites (**Marsaglia et al.,** this volume), indicating progressive development of carbonate-bank deposits in source areas on the adjacent shallow continental margin. Mineralogy of the smectites in the fine-grained sediments suggests that the source areas were relatively warm and humid (Tucholke, Sibuet, Klaus, et al., 2004).

A very diverse and abundant assemblage of mostly reworked, large benthic foraminifers was identified in the uppermost Paleocene to lowermost middle Eocene part of the section (Georgescu et al., submitted [N2]). This assemblage indicates the presence of a carbonate shelf in the adjacent source area of the Grand Banks, consistent with the above interpretations of source area based on sediment composition. Most of the benthic taxa are representative of the Caribbean bioprovince, and Georgescu et al. (submitted [N2]) propose that the warm climate of the Paleocene/Eocene Thermal Maximum, together with biodispersal by the proto-Gulf Stream, accounts for their presence that this high latitude.

The conjugate Iberia margin (Sites 398, 897, 1068, and 1069) similarly shifted to accumulation of carbonate-rich sediments during this period (Sibuet, Ryan, et al., 1979; Sawyer, Whitmarsh, Klaus, et al., 1994; Whitmarsh, Beslier, Wallace, et al., 1998), although the bulk of the carbonate consists of nannofossils. Only Site 398 shows a significant biosiliceous component, with siliceous marly chalk and radiolarian mudstone deposited during the late middle to early late Eocene. Thus, overall, the sedimentary record within the widening rift demonstrates productive surface waters, but the plankton were dominantly calcareous, in contrast to the rich siliceous fauna and flora farther south in the North Atlantic.

## Middle Eocene to Early Oligocene

Sediments deposited during the middle Eocene to early Oligocene were generally less calcareous than those in the underlying section. Site 1276 accumulated claystones and mudstones below the CCD, interrupted by occasional coarser turbidites (Tucholke, Sibuet, Klaus, et al., 2004). Benthic foraminifers in the nonturbidite sediments of this section are primarily agglutinated forms comprising faunal associations similar to those in Eocene sediments previously cored in the Labrador Sea (ODP Leg 105) and off Iberia (ODP Legs 149 and 173) (**Takata,** this volume). The dearth of calcareous forms in the "background" sediments is consistent with deposition on a seafloor below the CCD. Claystones, silty claystones, and nannofossil claystones with interbedded turbidites also were deposited on the Iberia margin, with only Site 398 accumulating largely marly nannofossil chalk with lesser mudstone (Sibuet, Ryan, et al., 1979).

The middle Eocene to lower Oligocene section at Site 1276 should contain the event that correlates with the Horizon $A^U$ reflection farther south in the western North Atlantic Basin. Along that continental margin Horizon $A^U$ commonly shows unconformable relations with underlying reflections and it correlates with a major hiatus near the Eocene/Oligocene boundary (Tucholke and Mountain, 1979; Miller and Tucholke, 1983). It has been interpreted to record the initiation of strong abyssal circulation in the North Atlantic Ocean, probably correlating with the first significant flow of cold bottom waters over the Greenland-Scotland Ridge from the Norwegian-Greenland Sea (Miller and Tucholke, 1983). One candidate for the Horizon $A^U$ equivalent at Site 1276 is at ~7.02 s two-way traveltime ($A^{U1}$, Fig. **F3**), one of several coherent reflections below the top of a sequence of strong, relatively level reflections at Site 1276. Traced landward, it becomes the top of this reflection sequence, and the overlying reflections become less coherent and splay into a landward-thickening wedge (Fig. **F3**). A second likely candidate ($A^{U2}$) is at the top of this wedge. This reflection separates the wedge from overlying beds that show thickening–thinning patterns suggestive of sediment waves; the wedge itself nearly pinches out at Site 1276, where the $A^{U2}$ reflection is at ~6.96 s reflection time.

**Shillington et al.** (this volume) constructed synthetic seismograms using core physical property data from Site 1276, and their correlations to the seismic reflection record suggest that $A^{U1}$ falls at 864.7 mbsf, matching the lithologic Unit 1/2 boundary at the site (Tucholke, Sibuet, Klaus, et al., 2004). Wood et al. (submitted [**N3**]) examined the nannofossil biostratigraphy of this interval in detail. They concluded that the Unit 1/2 boundary is represented by a condensed interval or hiatus that spans between 1.2 and 6.9 m.y. and is contained within the middle Eocene (47.3–40.4 Ma). They also noted that cooler-water taxa are more abundant above this level than below, which could indicate a change in circulation patterns, although the depth interval they analyzed is very restricted (<3 m) and may not be representative. Wood et al. (submitted [**N3**]) compared their stratigraphic results with Site 398 on the Iberia margin and noted a comparable break in the record there at the same age. Because of the nearly identical stratigraphic gaps at these widely separated locations, they suggested that the condensed intervals/hiatuses on the two margins were caused by high eustatic sea level that restricted sediment supply to the deep basins. This is consistent with Site 1276 shipboard sediment descriptions that define an increased hemipelagic component in lithologic Unit 1 compared to Unit 2 and that note an absence of any obvious structures that could be related to abyssal current effects (Tucholke, Sibuet, Klaus, et al., 2004). Considered together, these results suggest that $A^{U1}$ does not mark the initiation of strong abyssal circulation and therefore is not equivalent to Horizon $A^{U}$. In this case the sedimentary wedge between $A^{U1}$ and $A^{U2}$ probably was formed by downslope sediment movement rather than by current-controlled deposition. Its restriction to a near-slope position may reflect limited sediment input associated with elevated sea level.

The $A^{U2}$ reflection, which underlies wavelike bedforms in the seismic reflection record (Fig. F3), appears to be a better match to Horizon $A^{U}$, and this is our preferred interpretation. According to the seismic/borehole correlation of **Shillington et al.** (this volume), this reflection falls at the very top of the cored sediment section at 800 mbsf. Sediments in samples from Sections 210-1276A-1W-4 and 1W-CC are dated by calcareous nannofossils and dinocysts as uppermost Eocene to lower Oligocene (Tucholke, Sibuet, Klaus, et al., 2004). Although these samples were from a wash core (753–800 mbsf), it seems likely that the core catcher did recover sediments from the 800-m level; the next core (210-1276A-2R) is also upper Eocene to lowermost Oligocene (palynomorphs) or upper Eocene (nannofossils). Thus, the best available evidence is that the interpreted Horizon $A^{U}$ and the initiation of strong abyssal circulation in the North Atlantic dates very close to the Eocene/Oligocene boundary, as first suggested by Miller and Tucholke (1983).

# LOOKING TO THE FUTURE

Drilling in the Newfoundland-Iberia rift has been tremendously successful in advancing our understanding of the evolution of nonvolcanic rifted margins, but major questions remain to be answered. In the Newfoundland Basin, the nature of the basement below the bottom of Site 1276 remains unknown. Seismic velocity models from refraction data (Van Avendonk et al., 2006) suggest that the drill site is very close to a boundary between highly thinned continental crust and exhumed upper mantle. Deepening this hole into basement will provide significant new insights into the nature of these rocks; their tectonic, igneous and metamorphic evolution; and the potential that they may have had for generating synrift melts and/or postrift magmatism that formed the overlying diabase sills.

The nature of basement near the continental margins and its lateral variability remain very poorly understood, particularly in the outer transition zone (TE2), where rifting of apparently subcontinental mantle lithosphere eventually gave way to seafloor spreading. A suite of drill sites along the transect of ODP Legs 149, 173, and 210 and extending seaward onto Albian oceanic crust (i.e., beyond the time of the Aptian event), would much more clearly define how the transition from nonvolcanic rifting to seafloor spreading occurs, as well as the role that mantle composition (e.g., subcontinental or asthenospheric) and heterogeneity may play in this evolution.

Finally, this rift should also be studied and drilled along another transect, ideally one between Flemish Cap and the Galicia margin where ODP Leg 103 drilled. The timing of breakup and the extent of TE2 (TE1 is not present there) are much different in this location than in the central to southern part of the rift. The phases of rifting along this corridor are well defined by basement structure and sedimentary stratigraphy in seismic reflection records, and the continental crust is highly thinned. Furthermore, the spatial relations between continental crust and the subcontinental mantle are constrained both vertically (as defined by the “S reflection”) and horizontally (between the seaward-most continental blocks and the adjacent peridotite ridge).

With a few exceptions (e.g., Sites 398, 638, and 1276) the stratigraphic record associated with opening of the Newfoundland-Iberia rift is poorly sampled. Future drilling along either of the above transects should sample the most complete pre-, syn-, and postrift sedimentary sections that are technically feasible. Drilling should also include judicious sampling of basement highs to study the transition from exhumed mantle to oceanic crust.

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**Figure F1.** Reconstruction of the Newfoundland-Iberia rift at Chron M0 (~125 Ma, earliest Aptian), modified from Shipboard Scientific Party (2004). Present-day bathymetric contours (blue) are in meters. Continental crust is pale red, and yellow shows basins within continental crust that contain significant uppermost Triassic to Lower Jurassic evaporite deposits. DSDP and ODP drill sites are shown and coded according to whether they reached apparent continental basement (black), peridotite basement (green), or no basement (white); sites with good evidence that peridotite basement was faulted and uplifted 3–14 m.y. after it was emplaced have black dots at symbol centers. Peridotite ridges are marked by red lines with diamond symbols. Seismic refraction lines that constrain the distribution of basement types are shown by dotted lines, and colored bars and are explained in the box at lower right. The grid of lines at the southwest corner of the Galicia margin are CAM data. Thin "crust" shown by orange bars could be either magmatic ocean crust or serpentinized mantle, whereas blue bars show where velocity, layer thickness, and/or reflection characteristics suggest the presence of magmatic ocean crust. Data sources: SCR1 (SCREECH Line 1, Funck et al., 2003; Hopper et al., 2004), HU (Hudson profiles, Todd and Reid, 1989), W96 (Whitmarsh et al., 1996b), SCR2 (SCREECH Line 2, Van Avendonk et al., 2006; Shillington et al., 2006), CAM (Chian et al., 1999), IAM9 (Dean et al., 2000; Pickup et al., 1996), SR (Srivastava et al., 2000), R94 (Reid, 1994), P92 (Pinheiro et al., 1992). TE1 (darker green) shows conjugate zones of transitional extension reaching up to about anomaly M4 (Hauterivian/Barremian boundary); we interpret this as an interval of mantle exhumation. TE2 (lighter green) shows conjugate zones of subsequent transitional extension. We interpret this interval also to have been dominated by mantle exhumation, probably with a variable magmatic component, that reached to near the Aptian/Albian boundary (~112–110 Ma; i.e., to limits younger than shown in the reconstruction). See text for discussion. Red area at the southern edge of the rift schematically indicates a mantle plume that is interpreted to have constructed the Southeast Newfoundland Ridge, conjugate basement highs now incorporated in the western part of Gorringe Bank, and basement ridges reaching north and south along the plate boundary (e.g., J Anomaly Ridge, Madeira-Tore Rise) (Tucholke and Ludwig, 1982; Tucholke et al., 1989). This magmatism appears to have occurred from about the time of Chron M4 (latest Hauterivian) to younger than Chron M0, peaking at M1–M0 (late Barremian–earliest Aptian) (Tucholke and Ludwig, 1982). Red areas in the Newfoundland Basin indicate postrift sills intruded in ~Aptian–Albian sediments; these are interpreted from seismic reflection data and have been drilled at Site 1276. A.P. = abyssal plain, T.Z. = transfer zone, F.Z. = fracture zone. NZF = Nazaré Fault, PTF = Porto-Tomar Fault, AF = Aveiro Fault. (**Figure shown on next page.**)

**Figure F1 (continued).** (Caption shown on previous page.)

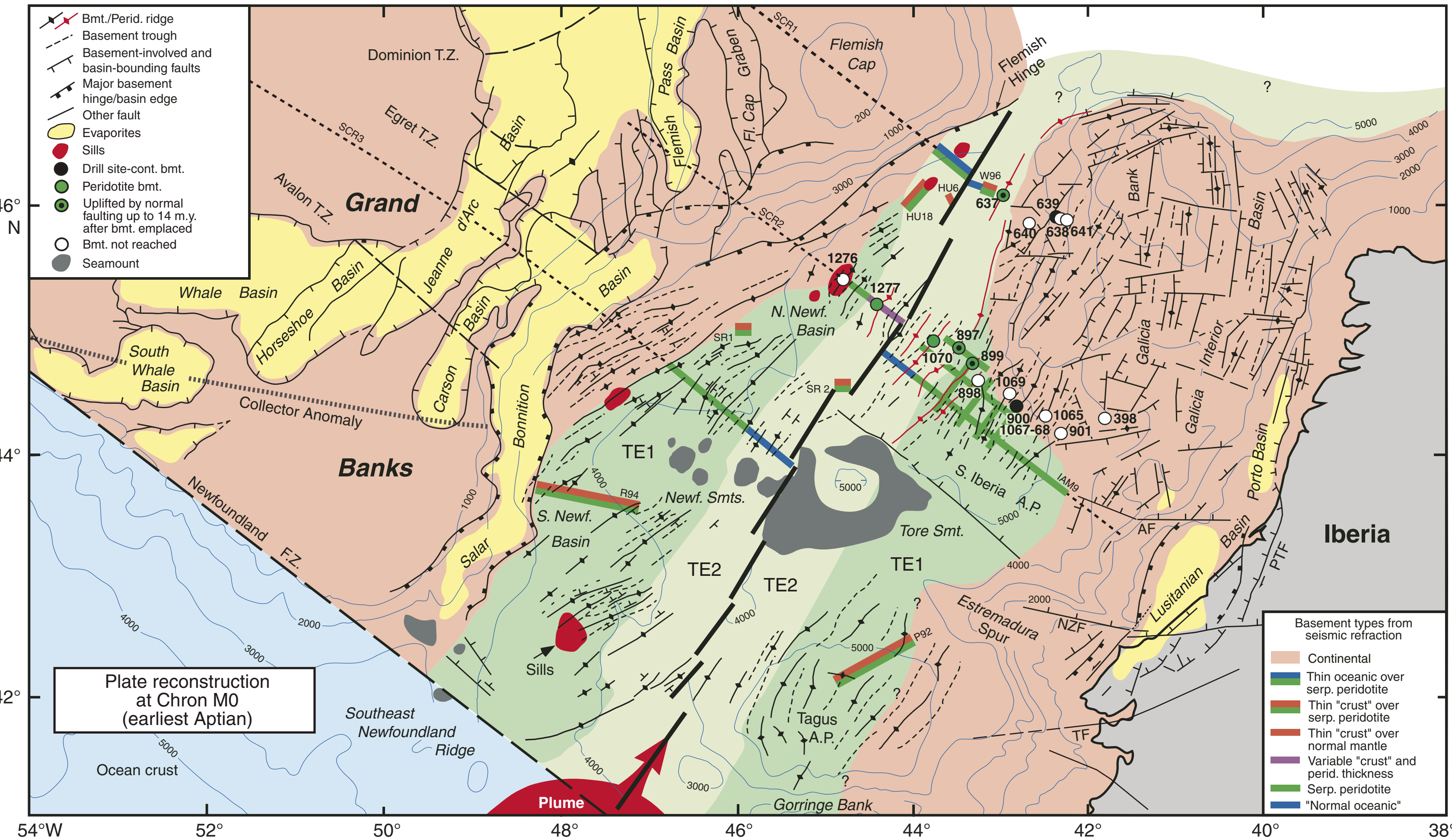

**Figure F2.** Bathymetry of the Newfoundland Basin, contoured at 200-m intervals, with the locations of Sites 1276 and 1277, ODP Leg 210. Magnetic anomalies M3 and M0 are shown, together with the approximate location of the Aptian/Albian boundary in crust farther seaward. The bold line at M0 shows the extent of the high-amplitude J Anomaly. Locations of seismic profiles in Figures F3, p. 48, and F9, p. 54, are indicated.

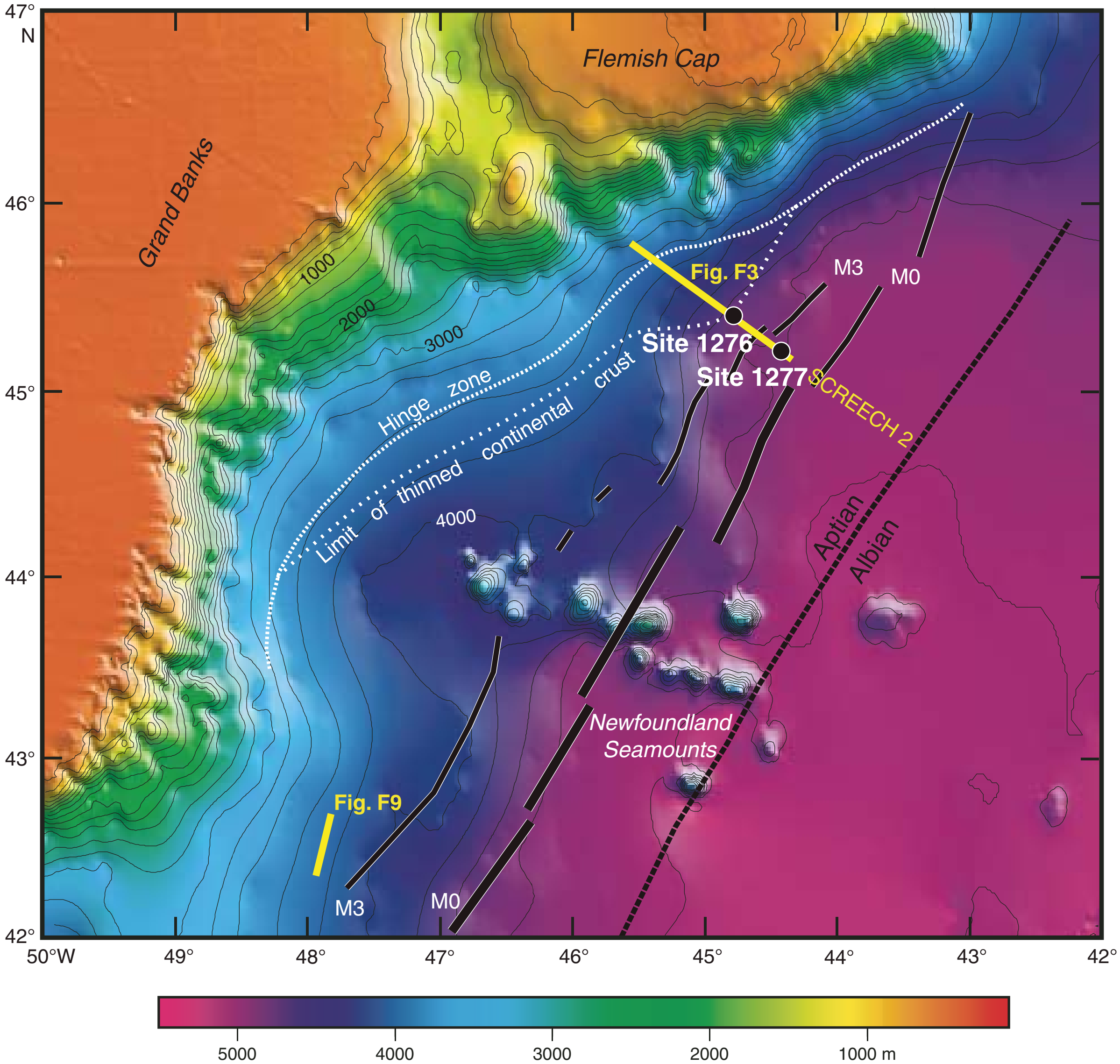

**Figure F3.** Parts of the SCREECH Line 2 multichannel seismic profile east (top) and west (bottom) of Site 1276. Note that the two panels are scaled differently. Locations are shown in Figure F2, p. 47. Magnetic anomalies M3 and M1 are indicated. In the top panel, three arrows at left point to changes in the U reflection (= Aptian event) that suggest disruptions by igneous sills. In the lower panel, two reflections that are candidates for correlation to Horizon $A^U$ in the western North Atlantic are traced by colored lines (see text for discussion). The upper reflection is interpreted to be equivalent to Horizon $A^U$, and it is shown as such in the top panel.

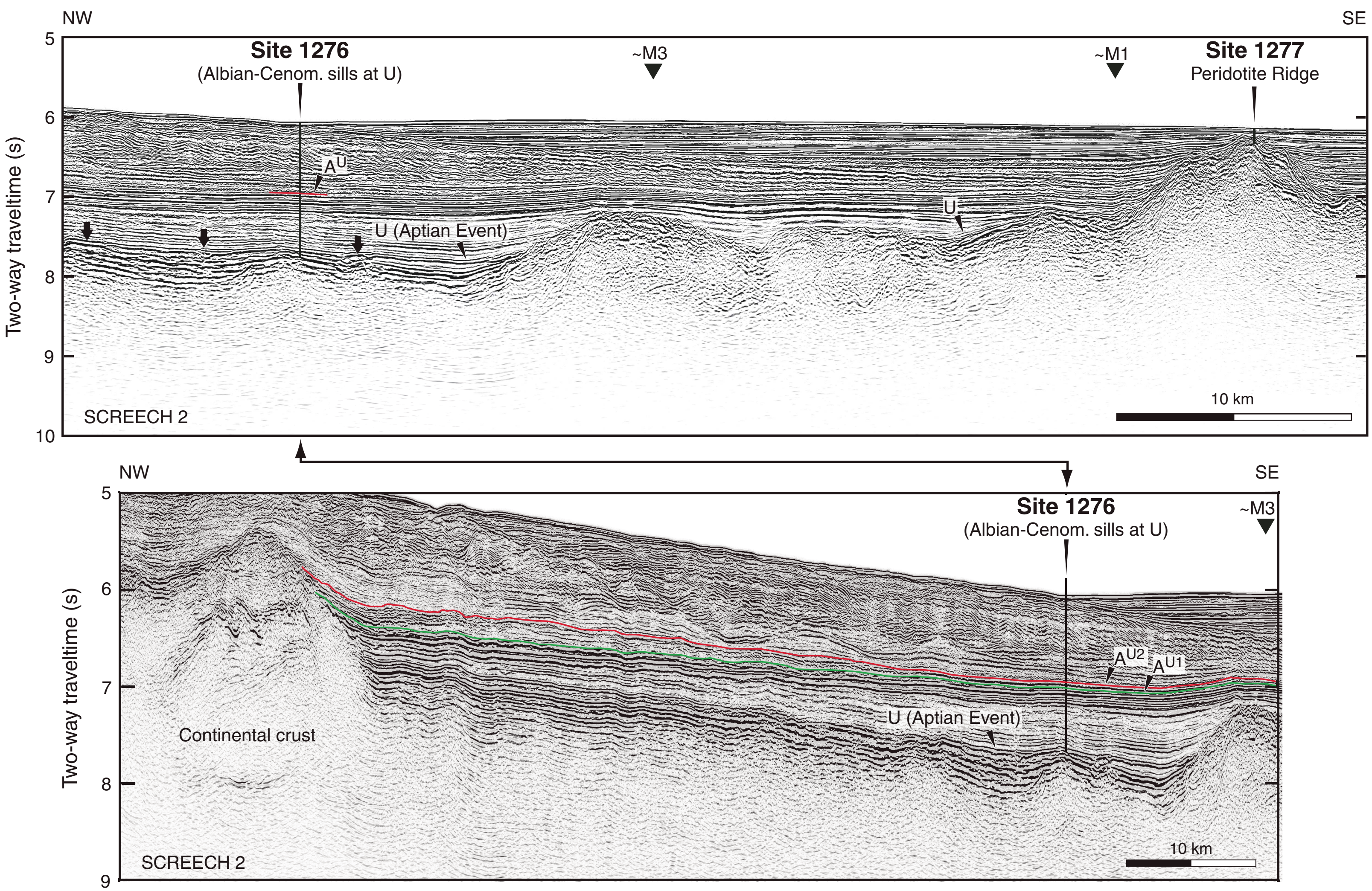

**Figure F4.** Summary of tectonic events, magmatic events, sedimentary history, and deep-basin paleoenvironment in the Newfoundland-Iberia rift from latest Jurassic to Paleogene time. The lithology of Site 1276 is summarized at the center of the figure. Formally defined formations of the main North Atlantic Basin (Jansa et al., 1979) are shown at the right, together with mid-Cretaceous Oceanic Anoxic Events (Leckie et al., 2002, as modified by Gradstein et al., 2004). CCD = calcite compensation depth. This figure is also available in an **oversized format.**

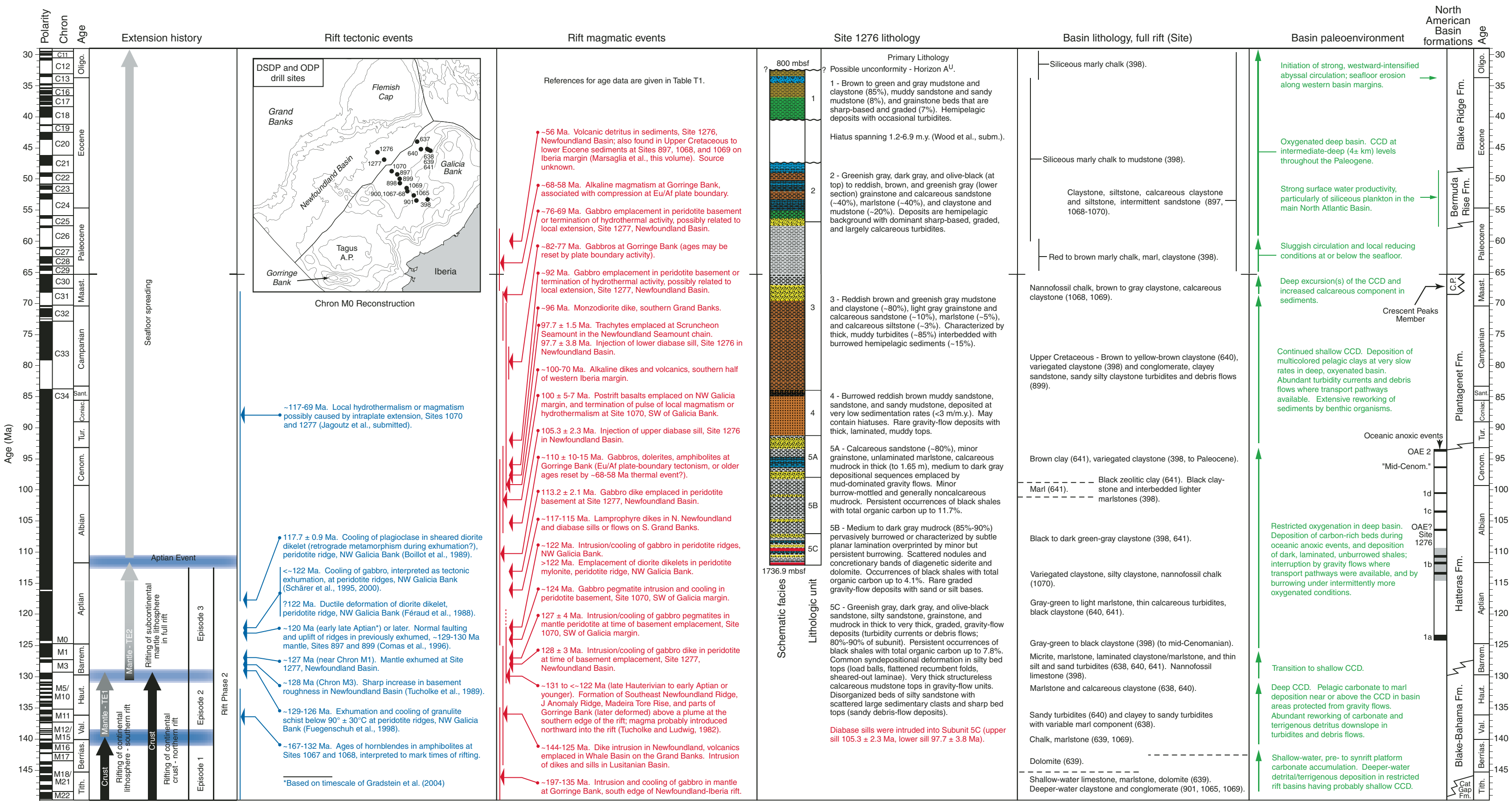

**Figure F5.** Multichannel seismic reflection profiles across thinned continental crust of the western Galicia margin, adapted from Tucholke et al. (2007). Profile locations are shown on the inset map together with bathymetric contours in meters. Seismic sequences are as follows: 6 = Berriasian, Tithonian, and older sediments and basement; 5 = Valanginian to Hauterivian; 4 = Barremian to Aptian; 3 = Albian and younger. This numbering follows the conventions of Mauffret and Montadert (1988). The Aptian event is approximately at the Aptian/Albian boundary. In panels A and B, the S reflection is interpreted to be a detachment surface between continental crust and underlying mantle (e.g., Reston, 1996). Bold arrows show fault slip, and light arrows highlight sedimentary onlap.

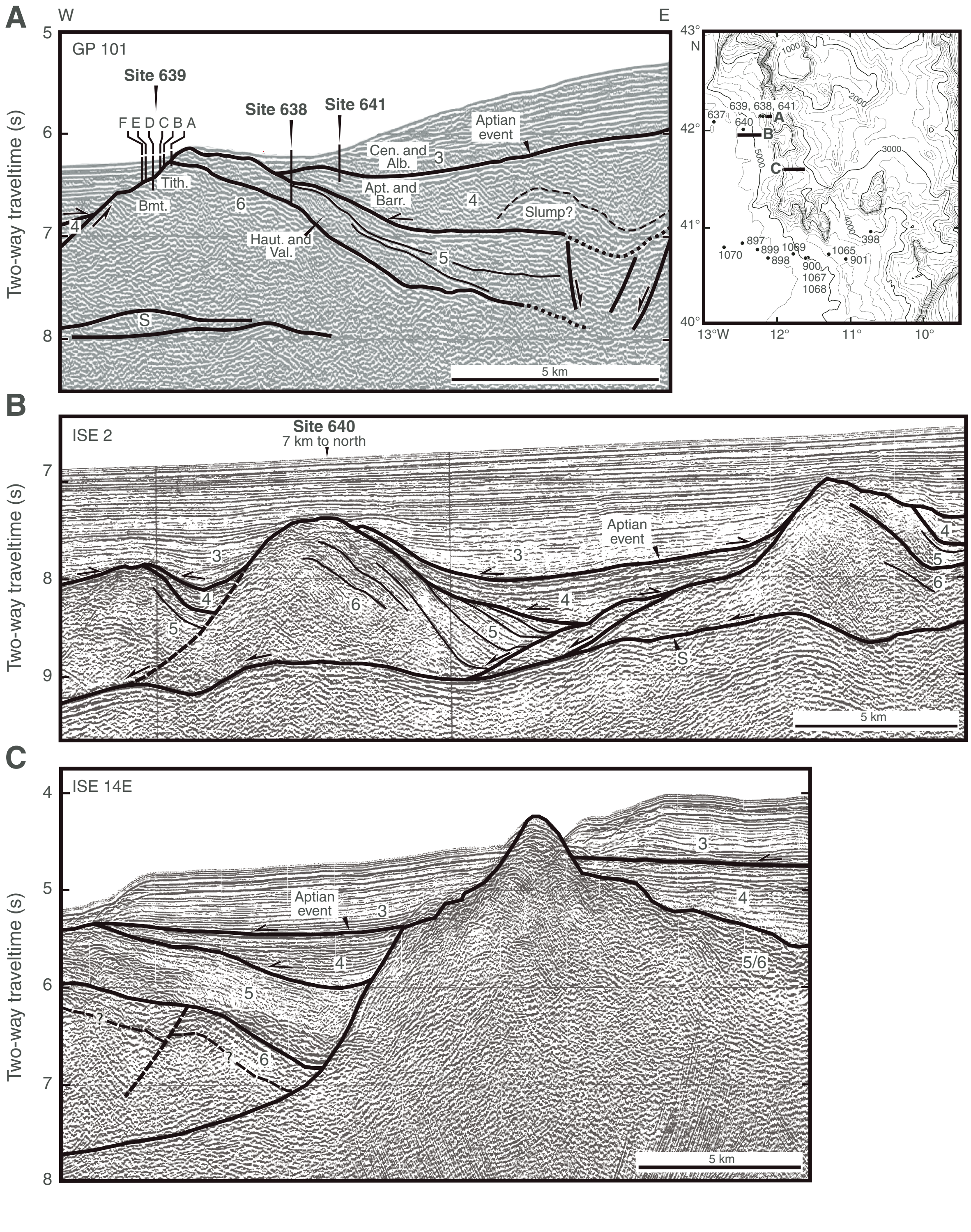

**Figure F6.** Multichannel seismic reflection profile showing strong Aptian event reflection (arrows) that merges with the surface of a peridotite ridge on the western Galicia margin (adapted from Tucholke et al., 2007). Map shows location, with bathymetric contours in meters. Note that the strong Aptian event reflection masks underlying basement structure near the peridotite ridge. The reflection amplitude decreases with distance from the ridge, suggesting that weak serpentinites on the ridge provide a source of coarse debris.

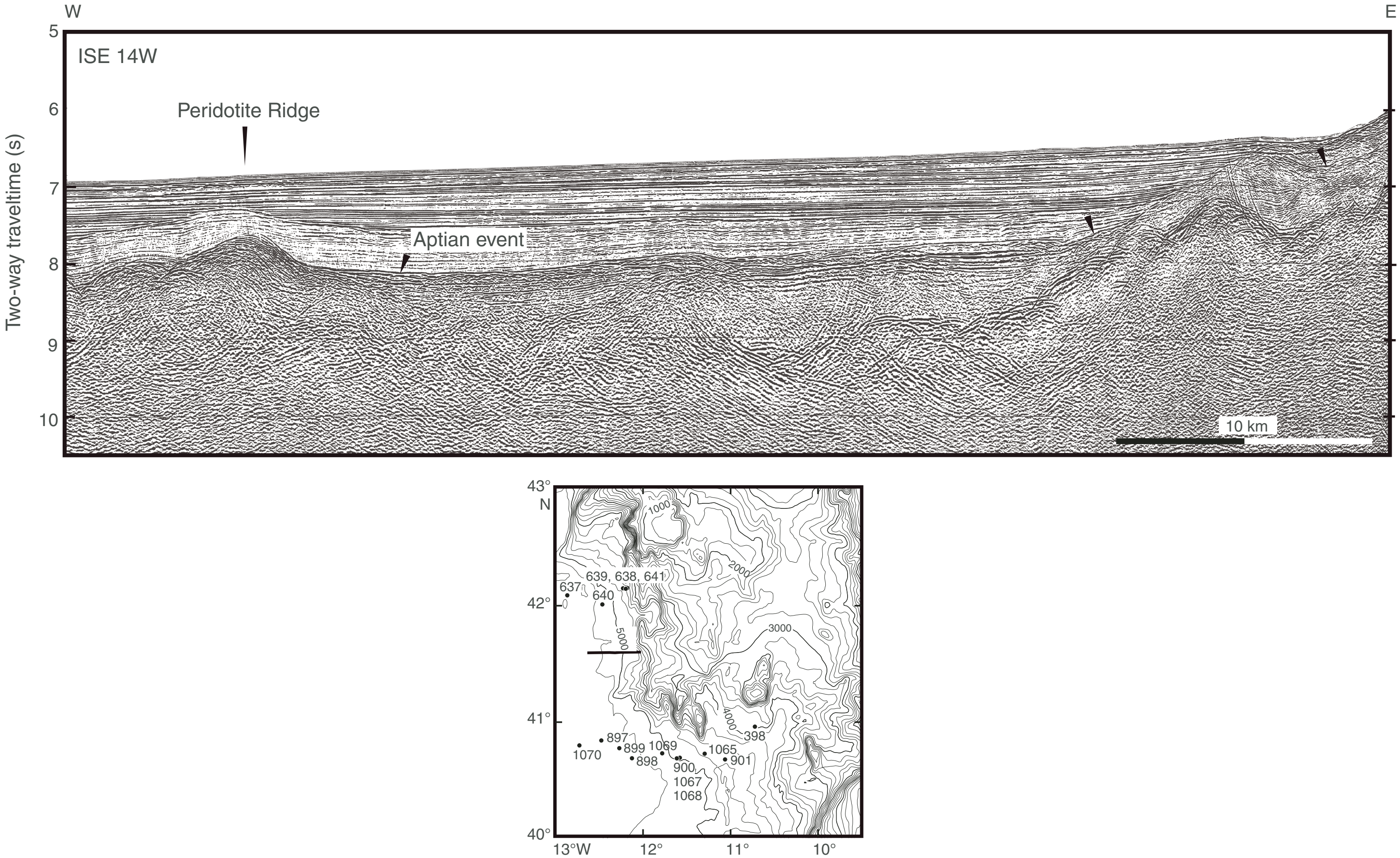

**Figure F7.** Correlation between observed magnetic anomalies and calculated M-sequence anomalies for conjugate magnetic profiles in the northern Newfoundland-Iberia rift, adapted from Srivastava et al. (2000). Profile locations are shown on the inset map, and ODP drill sites along the profiles are noted in blue. **A, B.** Profiles ER-56 and LG-12 are conjugate surface profiles; the ER-56 profile from the Newfoundland margin has been flipped end-for-end to facilitate comparison with the LG-12 profile. **C.** Modeled surface magnetic anomaly, based on the reversal model in G. **D, E.** SAR-93 and TOBI-91 deep-tow magnetic profiles from the Iberia margin along the western part of the Leg 149/173 drilling transect (Whitmarsh et al., 1996a; Sibuet et al., 1996). **F.** Corresponding modeled magnetic anomaly at the sea bottom. The TOBI-91 deep-tow profile is along the same line as the LG-12 surface profile. **G.** Reversal model shows thickness of the magnetic layer, extension rates (computed by using the Kent and Gradstein, 1986, timescale), and magnetization. **H, I.** In I at the bottom of the figure, a model with basement topography and constant magnetization (5 A/m) is shown to evaluate the effect of the basement topography. Magnetization direction (Dr = 0° and Ir = 46°) is from Van der Voo (1990) for the mid-Cretaceous. Corresponding magnetic anomalies in H are calculated at the sea surface (gray line) and sea bottom (red line). They show that variations of the basement topography explain only ~10% of the observed magnetic anomaly amplitude (note that H has an amplitude scale different from the scales in A–F).

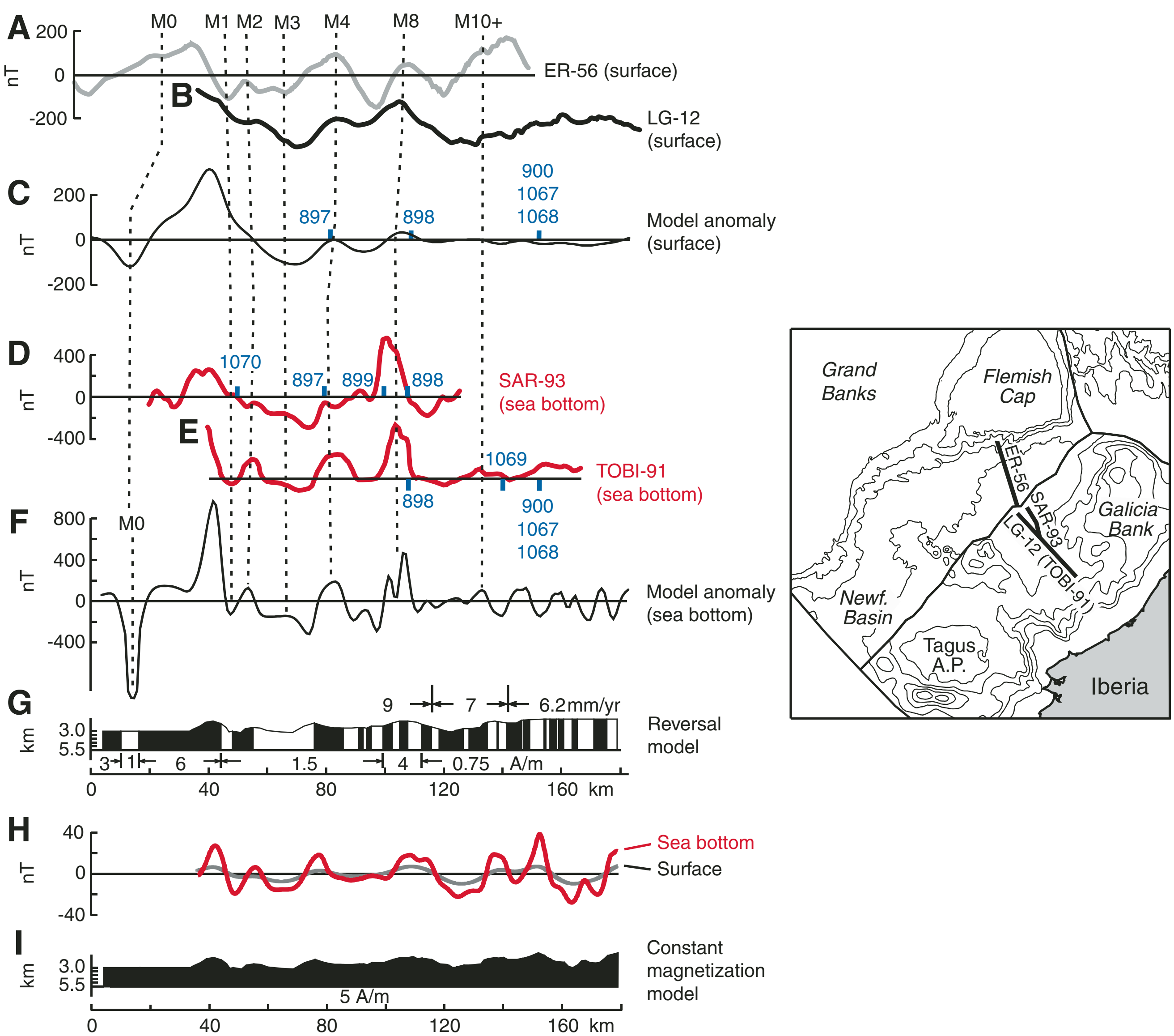

**Figure F8.** Half-rates for extension/spreading in the Newfoundland-Iberia rift (lower part of figure) calculated from magnetic anomaly identifications of Srivastava et al. (2000) using three different timescales (upper part of figure). The recent Fiet et al. (2006) Cretaceous timescale is shown for comparison (note the very large discrepancy between the G04 and Fiet et al. timescales). In the lower part of the figure the solid horizontal lines (color-coded according to timescale used) show rates calculated on the Newfoundland (NF) and Iberia (IB) plates at the indicated time intervals, or on only one plate where picks were not made on both sides of the rift. Thick color-coded lines show averages and trends of these rates through time, based on the different timescales. Major discrepancies among the three timescales are in the Early Cretaceous, especially for the G04 timescale (see text for discussion). Because of poorly resolved anomalies older than ~M8 (Fig. F7, p. 52), calculated rates older than this may not be significant. For comparison, two orange dotted lines labeled with a W show extension rates proposed by Whitmarsh et al. (2001), based on geometrical reconstruction of normal faults along the southern margin of the Galicia block. TE1 and TE2 = Transitional Extension periods 1 and 2. Plume magmatism at the southern edge of the rift is represented by the SENR = Southeast Newfoundland Ridge; JAR = J Anomaly Ridge; and MTR = Madeira-Tore Rise. Times of plume magmatism, rifting, mantle exhumation, and seafloor spreading are keyed to the Gradstein et al. (2004) timescale. Drill sites and selected radiometric ages (see Table T1, p. 55) are indicated.

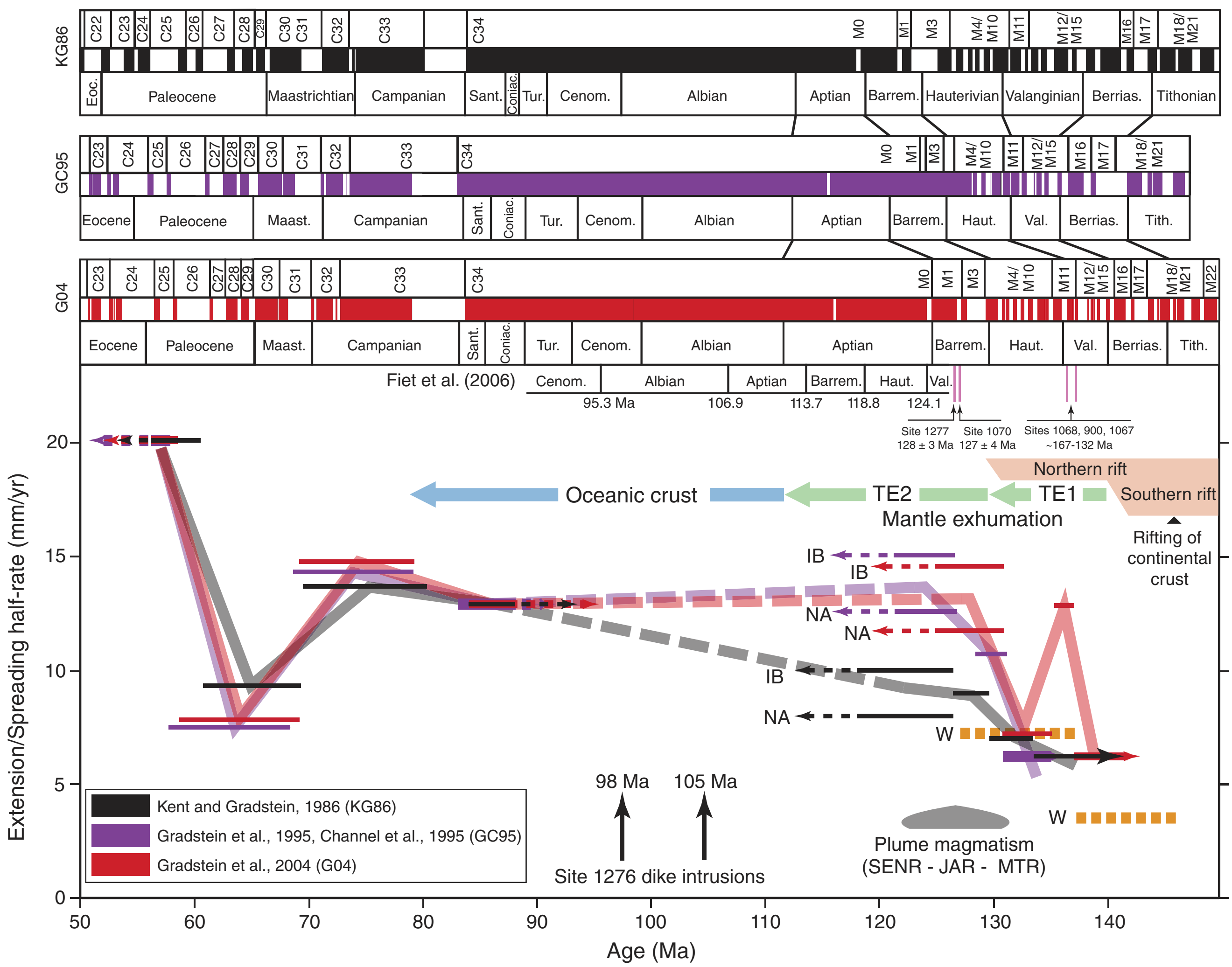

**Figure F9.** Multichannel seismic reflection profile showing sills at the level of the U reflection (Aptian event) in the southern Newfoundland Basin (adapted from Tucholke et al., 2007). Location is shown in Figure F2, p. 47. Note the strong reflectivity and sharp terminations of the sills, which appear to extend over distances of a few kilometers in this area.

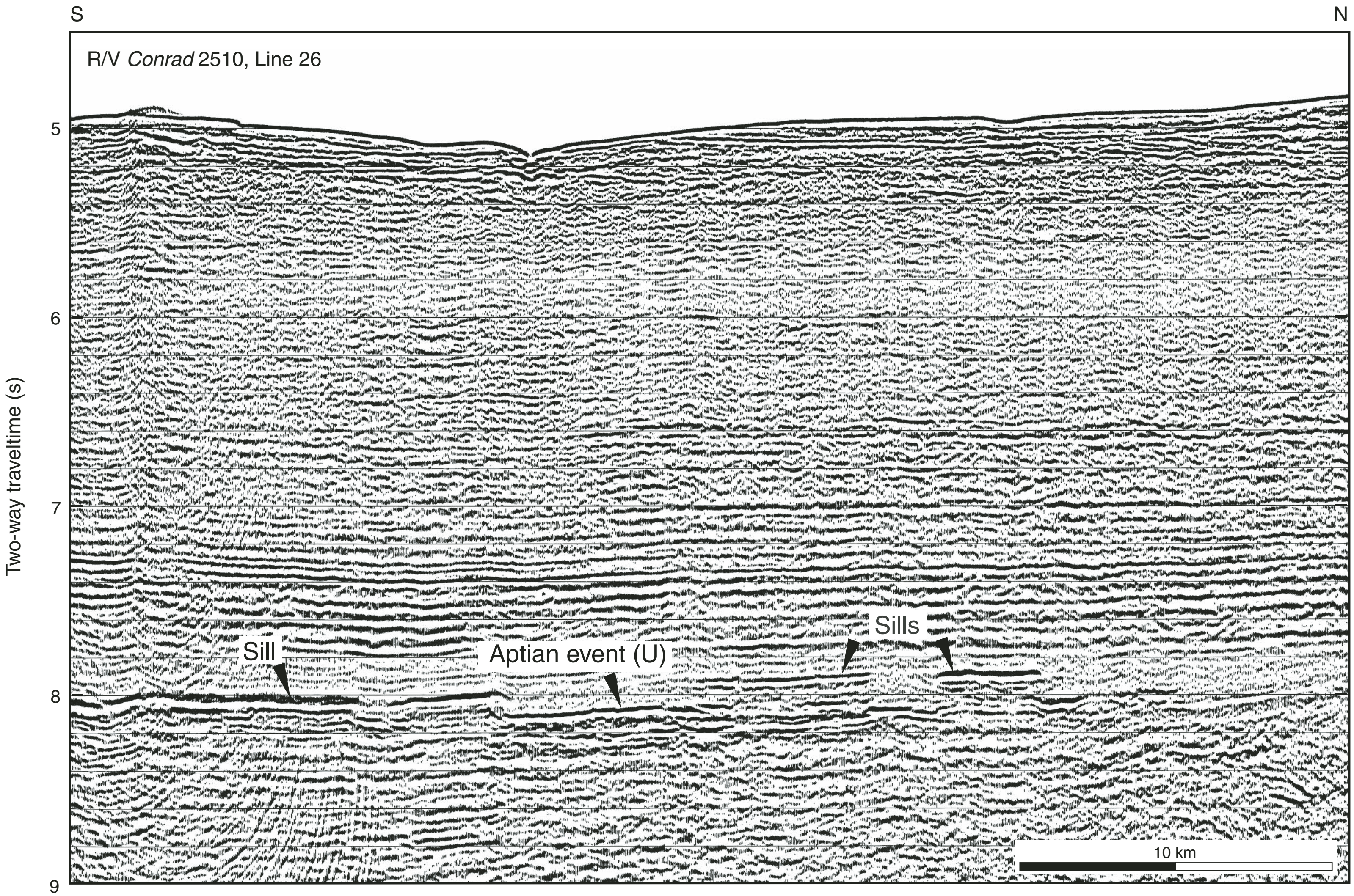

**Table T1.** Ages and characteristics of igneous and metamorphic rocks in the Newfoundland-Iberia rift. (This table is available in an **oversized format.**)

# CHAPTER NOTES*

**N1.** Jagoutz, O., Müntener, O., Manatschal, G., Rubatto, D., Péron-Pinvidic, G., Turrin, B.D., and Villa, I.M., submitted. The rift-to-drift transition in the southern North Atlantic: a stuttering start of the MORB engine? *Geology.*

**N2.** Georgescu, M.D., Leckie, R.M., and Hiscott, R.N., submitted. Latest Paleocene and early Eocene large-sized reworked benthic foraminifers in the Newfoundland Basin (Ocean Drilling Program Leg 210, Site 1276) and their paleoclimatic significance. *Rev. Micropaleontol.*

**N3.** Wood, A., Gardin, S., and Wise, S.W., Jr., submitted. Age constraint and interpretations of a disconformity in the Newfoundland-Iberia rift basin, Ocean Drilling Program Leg 210. *Rev. Micropaleontol.*

*Dates reflect file corrections or revisions.