A deep seismic investigation of the Flemish Cap margin: implications for the origin of deep reflectivity and evidence for asymmetric break-up between Newfoundland and Iberia

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SUMMARY

Seismic reflection and refraction data were acquired along the southeast margin of Flemish Cap at a position conjugate to drilling and geophysical surveys across the Galicia Bank margin. The data document first-order asymmetry during final break-up between Newfoundland and Iberia. An abrupt necking profile of continental crust observed off Flemish Cap contrasts strongly with gradual tapering on the conjugate margin. There is no evidence beneath Flemish Cap for a final phase of continental extension that resulted in thin continental crust underlain by a strong 'S'-like reflection, which indicates that this mode of extension occurred only on the Galicia Bank margin. Compelling evidence for a broad zone of exhumed mantle or for peridotite ridges is also lacking along the Flemish Cap margin. Instead, anomalously thin, 3–4-km-thick oceanic crust is observed. This crust is highly tectonized and broken up by high-angle normal faulting. The thin crust and rift structures that resemble the abandoned spreading centre in the Labrador sea suggest that initial seafloor spreading was affected by processes observed in present-day ultra-slow spreading environments. Landwards, Flemish Cap is underlain by a highly reflective lower crust. The reflectivity most likely originates from older Palaeozoic orogenic structures that are unrelated to extension and break-up tectonics.

Key words: continental break-up, Newfoundland margin, rifting.

1 INTRODUCTION

Investigations of non-volcanic margins have proven essential for gaining insight into the mechanical response of crust and lithosphere to extensional stresses. In these environments, the resulting structures are not modified by magmatic processes, thus permitting a better understanding of their kinematic evolution. More importantly, such a crust is more easily imaged by seismic methods, which are crucial for constraining the geometries and basin structures associated with rifting and the eventual creation of an oceanic basin.

Comprehensive surveys and sampling along the Iberian margin (Fig. 1) have been instrumental in establishing some of the key aspects of non-volcanic margin evolution. Complex structural patterns have been revealed, including the 'S' reflection off Galicia Bank. 'S' is hypothesized to represent a major detachment surface at the crust–mantle boundary and has been used to infer large-scale asymmetry of the entire rift system during the final stages of break-up (Reston *et al.* 1996; Whitmarsh *et al.* 2001). In addition, it is now recognized that a simple continent–ocean boundary may not exist in these magma-starved settings. Instead, a broad transition zone of mechanically unroofed upper mantle with little evidence for decompression melting is observed. This so-called zone of exhumed continental mantle (ZECM) is heavily serpentinized by reaction with sea water to form a 2- to 4-km-thick layer with crust-like seismic velocities (Pickup *et al.* 1996; Dean *et al.* 2000). Mechanical unroofing may occur along concave-downward faults with footwalls that rotate as they are unroofed, thus allowing for continued slip along a single fault and for exhumation of rocks from great depths (e.g. Buck 1988; Lavier *et al.* 1999; Manatschal 2004).

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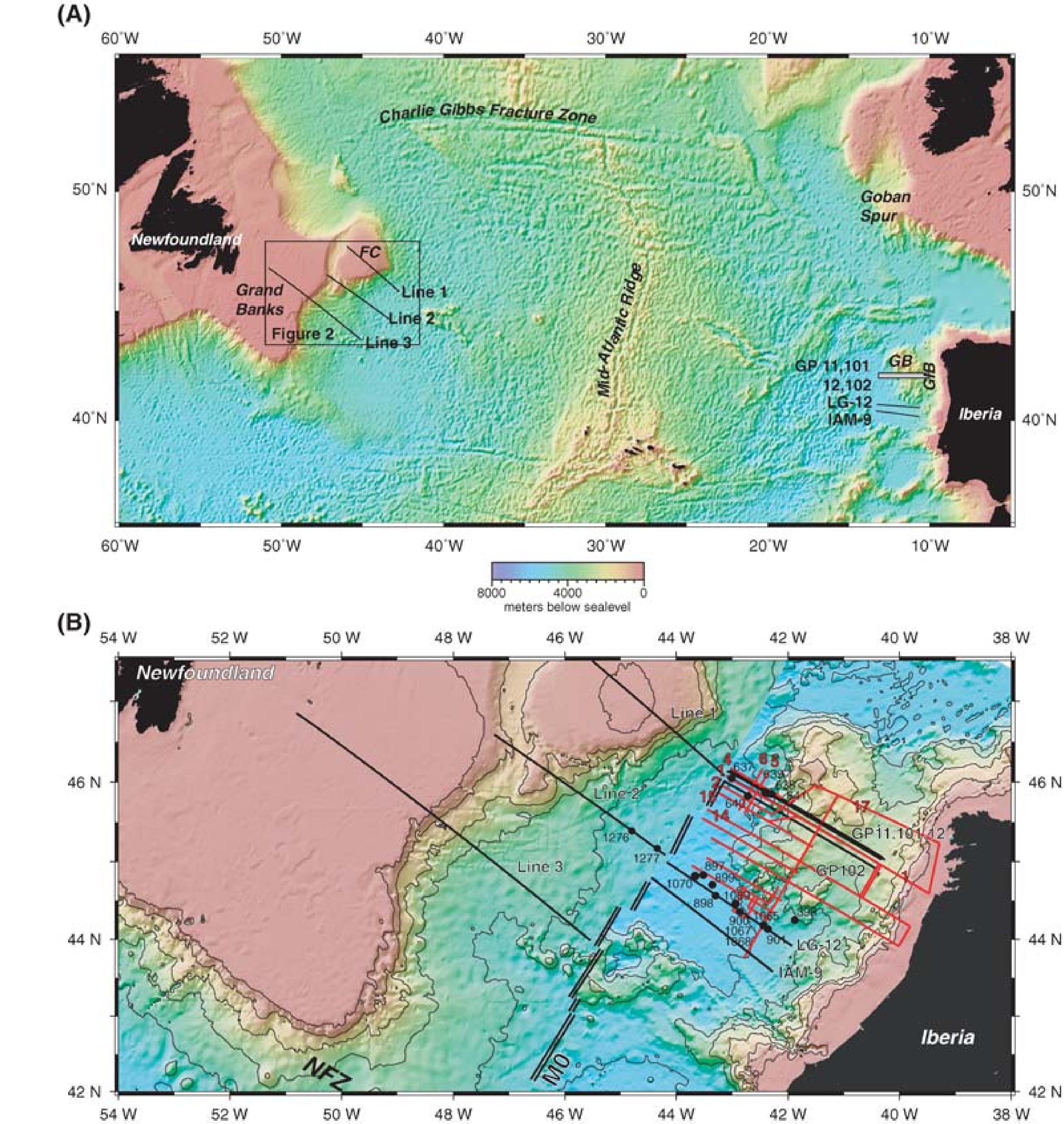

Figure 1. (a) Regional map of the North Atlantic Ocean showing the conjugate margins of Newfoundland and Iberia. Bathymetry from Smith & Sandwell (1994); depths are colour-coded as in Figs 1(b) and 2. The location of Fig. 2 is indicated. Lines 1–3 show the long, primary dip lines from three SCREECH transects across the Newfoundland margin. Seismic profiles described here are from SCREECH Transect 1 across Flemish Cap. Galicia Bank profiles GP11, GP101, GP12, GP102 are approximately conjugate to SCREECH Line 1. Also shown are profiles IAM-9 (Reston *et al.* 1996; Whitmarsh *et al.* 1996) and LG-12 (Beslier 1996), the latter of which is approximately conjugate to SCREECH Line 2. FC: Flemish Cap; GB: Galicia Bank; GIB: Galicia Interior Basin. (b) Reconstruction of the Newfoundland–Iberia rift at anomaly M0 time using poles of rotation from Srivastava *et al.* (1990) as modified by Srivastava *et al.* (2000). The bathymetric contours are shown at 200, 1000, 2000, 3000, 4000 and 5000 m. SCREECH Lines 1–3 are labelled. ODP drill sites are marked with small black numbers and black dots. Labelled black lines indicate seismic profiles discussed in the text. Red lines and numbers are the ISE97 seismic survey (Henning *et al.* 2004). NFZ: Newfoundland fracture zone.

Because of its position conjugate to Iberia, the Newfoundland rifted margin is a key area where models developed to explain the Iberia observations can be tested and examined more thoroughly. Although there has been much previous work on this margin (e.g. Keen *et al.* 1987; Tucholke *et al.* 1989; Reid 1994; Hall *et al.* 1998), there has been a lack of coincident wide-angle refraction and deep seismic reflection data at key locations relative to the major transects on the Iberian margin. In addition, until recently there was no drilling and sampling of the Newfoundland margin except in commercial exploration wells on the Grand Banks. To address these issues, a

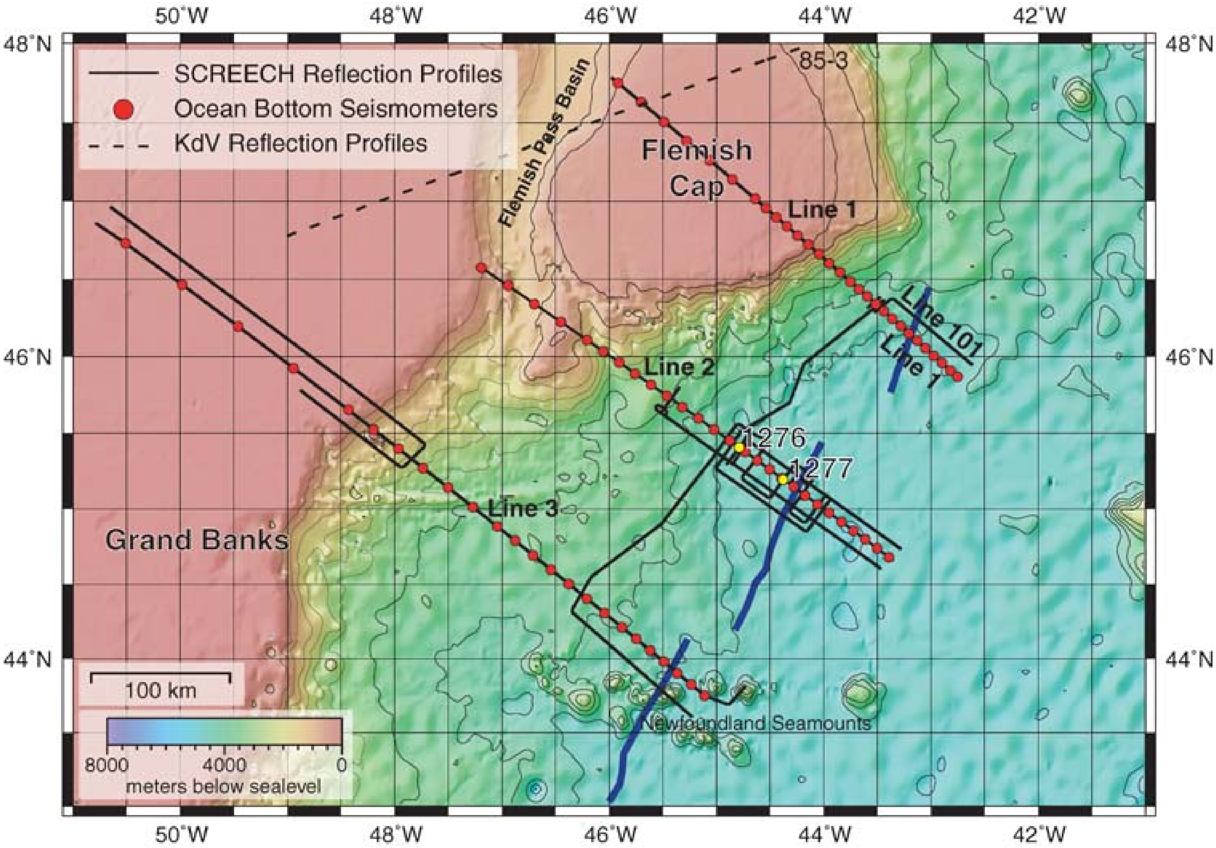

Figure 2. Detailed map showing the SCREECH survey on the Newfoundland margin. Bathymetry from Smith & Sandwell (1994) with a contour interval of 500 m. KdV: Keen & de Voogd (1988). Profile 85–3 is described in detail by Keen *et al.* (1989). OBS's along Line 1 are numbered from 1 to 29 beginning at the NW end of the profile. Thick blue line is magnetic anomaly M0 from Srivastava *et al.* (2000). Yellow dots are ODP Leg 210 Sites (Tucholke *et al.* 2004).

major seismic survey, called SCREECH (Studies of Continental Rifting and Extension on the Eastern Canadian sHelf), was carried out in 2000 July and August, wherein both reflection and refraction data were collected along three major transects (Figs 1 and 2). The seismic data were collected in support of Ocean Drilling Program (ODP) objectives to sample a conjugate margin pair. A first phase of drilling was carried out in 2003 during ODP Leg 210 (Fig. 2; Shipboard Scientific Party 2004b).

The northernmost seismic lines (SCREECH Transect 1) are conjugate to Galicia Bank where the 'S-reflection' has been well imaged and where previous ODP drilling was conducted on Leg 103 (Boillot *et al.* 1987). Fig. 1(b) shows the bathymetry and survey locations off Iberia, reconstructed to the Newfoundland margin at the time of magnetic anomaly M0 using the poles of rotation of Srivastava *et al.* (1990) as modified by Srivastava *et al.* (2000). In the vicinity of the Newfoundland fracture zone, a large-amplitude magnetic anomaly known as the J-anomaly is observed between anomalies M0 and M1 (Rabinowitz *et al.* 1978; Srivastava *et al.* 2000). Northwards from the Newfoundland fracture zone, the magnitude of the J-anomaly decreases significantly, but there is general agreement on the identification of M0 up to the Flemish Cap. Overall, the fit of the plate reconstruction is very good at M0 (see Fig. 2 of Srivastava *et al.* 2000), though inherent uncertainties could probably move the conjugate transects by several tens of kilometres along strike.

SCREECH Transect 1 is comprised of two dip lines, two short strike lines, and a third strike line that connects it to Transect 2 to the south. The velocity structure on the primary dip line (Line 1) is documented by wide-angle data reported by Funck *et al.* (2003), and pre-stack depth-migrated images of the continent–ocean boundary region are described by Hopper *et al.* (2004). Those data provide evidence for asymmetry in rift structure at a variety of scales when considered with the Iberia margin. The Newfoundland data further suggest that, in contrast to interpretations of the Iberian margin, final break-up was not entirely amagmatic even though it was clearly magma starved. Anomalously thin, apparently oceanic crust lies close to thin continental crust and is highly tectonized. Hopper *et al.* (2004) proposed that initial seafloor spreading occurred in an environment similar to modern-day ultra-slow seafloor-spreading regions such as Gakkel Ridge (Coakley & Cochran 1998; Jokat *et al.* 2003) or the Southwest Indian Ridge (Dick *et al.* 2003). To explain some unusual features in the data, a model was proposed for early seafloor spreading that implies large fluctuations in the available melt supply.

In the present paper, additional seismic reflection data along and parallel to SCREECH Line 1 are presented. The data:

(1) provide important insights into the origin of lower-crustal reflectivity along the Flemish Cap margin and suggest that it has little relation to rift-related extension,

(2) document the nature of asymmetry between Flemish Cap and Iberia on a variety of scales and

(3) provide further evidence in support of anomalously thin oceanic crust and ultra-slow seafloor spreading along the margin.

2 REGIONAL GEOLOGIC SETTING AND PREVIOUS DATA

Newfoundland is bounded offshore by the Grand Banks and the northeast Newfoundland shelf, which are broad, 300–400-km-wide platforms underlain by continental crust (Keen & de Voogd, 1988; Figs 1 and 2). Flemish Cap is a subcircular block of 30-km-thick continental crust with a three-layer velocity structure identical to Appalachian crust measured elsewhere (Funck *et al.* 2003). It lies northeast of the Grand Banks and is separated from the main platform by Flemish Pass Basin. Geologically, it is part of the Avalon terrane found onshore Newfoundland (e.g. King *et al.* 1985; Tankard & Welsink 1987). Throughout the Mesozoic, it appears to have behaved as a small microplate (Srivastava *et al.* 2000).

Two significant phases of extension associated with Atlantic Ocean opening occurred between Newfoundland and Iberia, while a third phase affected the northeastern Newfoundland margin and opened the adjacent Labrador Sea. Early rifting in Late Triassic to Early Jurassic time formed major rift basins within the Grand Banks and on the Iberia margin (Tankard & Welsink 1987; Wilson 1988; Rasmussen *et al.* 1998). This was followed by a 40–50 Myr period of relative quiescence, and the sedimentary basins record a gradual and uniform thermal subsidence (Tankard & Welsink 1987).

A second major phase of extension began in the Late Jurassic and culminated in seafloor spreading no later than Barremian to Aptian time (126–112 Ma on the timescale of Channell *et al.* 1995; Gradstein *et al.* 1995, used here). This second phase of extension resulted in separation of the southwest Flemish Cap margin from Galicia Bank. In the southern Iberia Abyssal Plain (IAP) conjugate to SCREECH Transect 2 (Figs 1 and 2), anomaly M3 is the oldest magnetic anomaly attributed to seafloor spreading on which there is general agreement (∼125 Ma; Whitmarsh & Miles 1995), although older anomalies have been suggested (Srivastava *et al.* 2000). Along SCREECH Transect 1 reported here, separation of continental crust was later (Funck *et al.* 2003). It has been proposed that anomaly M0 (earliest Aptian) is present (Srivastava *et al.* 2000), but it remains unclear whether there are older seafloor-spreading anomalies; for example, anomaly M3 proposed by Srivastava *et al.* (2000) is clearly over continental crust (Funck *et al.* 2003). Although some form of seafloor spreading appears to have occurred by the time of anomaly M0 (121 Ma), the 'break-up unconformity' on the conjugate Galicia Bank drilling transect has been dated to a much later time (late Aptian, ∼112 Ma; Boillot *et al.* 1987). Tucholke *et al.* (2006) have suggested that this unconformity developed when plate-wide extensional stress was relieved during the transition from extensional exhumation of mantle in the rift (i.e., near-amagmatic 'seafloor spreading') to a system of more normal seafloor spreading.

Srivastava *et al.* (2000) estimated the half spreading rate during early opening to be 7 mm yr^{-1} and proposed that ultra-slow seafloor spreading may have been important in margin development. Off Iberia, Russell & Whitmarsh (2003) estimated slightly faster rates of ∼10–14 mm yr^{-1}, but noted that drilling results suggest slower rates during the formation of the ZECM. There is a general consensus that the initial opening rates were slow to ultra-slow and comparable in magnitude to those of the Gakkel Ridge and Southwest Indian Ridge, which are the slowest-spreading segments of the modern mid-ocean ridge system (e.g. Coakley & Cochran 1998; Chu & Gordon 1999; Muller *et al.* 2000).

The third and final phase of extension opened the Labrador basin north and northeast of Newfoundland beginning about 80 Ma, separating the northeast margin of Flemish Cap from the Goban Spur margin (Keen *et al.* 1989). Seafloor spreading in the Newfoundland–Iberia rift along the southwest Flemish Cap margin was well underway by this time.

The nature of the transition from continental to oceanic lithosphere is well studied off Iberia. South of Galicia Bank, a broad, >130-km-wide zone of mechanically unroofed continental mantle separates continental crust from more normal oceanic crust (Dean *et al.* 2000). This zone narrows to the north. At Galicia Bank it appears to wedge out, and on the west side of Galicia Bank continental crust is separated from apparent oceanic crust by a single, <10-km-wide peridotite ridge (Whitmarsh *et al.* 1996b). Continental crust along this part of Galicia Bank is extremely thin, only 3–4 km thick immediately landwards of the peridotite ridge, and it is underlain by the so-called 'S' reflection, which is interpreted as a major detachment surface that played a key role in final break-up (Reston *et al.* 1996; Pérez-Gussinyé. & Reston 2001).

3 DATA ACQUISITION AND PROCESSING

SCREECH data across the Newfoundland margin were acquired during two-ship operations using *R/V Maurice Ewing* and *R/V Oceanus*. Ocean bottom seismometers (OBS) were deployed by *Oceanus* to record refraction and wide-angle reflection data along three primary dip lines, one within each of the main transects (Fig. 2). Shooting operations and multichannel seismic (MCS) reflection data acquisition were carried out on the *Ewing*. In addition to the long, primary dip lines, supplementary MCS data were collected along several lines parallel and perpendicular to the primary lines. The tuned source array consisted of 20 airguns with a total volume of 140 L (8540 cu. in.). Data were shot on distance and the primary dip line in each transect was covered twice, once with a shot spacing of 200 m for recording by the OBS's, and again with a smaller shot spacing for the MCS data. The shot spacing for most of the MCS data acquisition was 50 m, but strong currents occasionally caused the ship's speed over the bottom to be too fast for the source to repressurize fully. When this happened, the shot spacing was increased to either 62.5 or 75 m depending on the ground speed. MCS data were recorded on *Ewing*'s 6 km streamer with 480 channels (12.5 m group spacing), and data were binned into common depth point (CDP) gathers spaced every 6.25 m. Thus, the data fold is 60 for the 50 m shot spacing, but it drops to as low as 40 when the shot distance could not be maintained. The northernmost profiles, lines 1 and 101 in Transect 1, are shown here. On Line 1, the data is 60 fold from CDP 21 000–68 488 and lower for the rest of the profile. One Line 101, the fold is 60 from CDP 75 500–83 488 and lower for the rest of the profile.

The data processing sequence depended on water depth. The deep-water data had minimal problems with multiples and other coherent noise and was pre-stack depth migrated. In shallow water and along the slope, however, coherent noise problems made prestack migration impossible. The processing details for each case are described below, but in both cases, data were pre-processed by assigning geometry from the navigation files, applying spherical divergence amplitude corrections using estimates of the expected velocities, filtering using a minimum phase Butterworth bandpass from 4–80 Hz, and finally suppressing the bubble pulse of the source with a minimum phase predictive deconvolution filter.

3.1 Data processing: Shallow water and slope

Flemish Cap is nearly devoid of sediment and hard-rock basement occurs only a few milliseconds below the seafloor. Guided waves and energy trapped in the upper basement layer are a severe problem and significantly hamper imaging the deeper structure. Most of this energy has a linear slope of <5.4 km s^{-1} on shot gathers and was suppressed by f–k filtering in the shot domain, eliminating all low-velocity energy beginning at the first seafloor multiple. Hyperbolic semblance analysis was run and stacking velocities were picked at a minimum spacing of 625 m, or less where lateral variations were clearly indicated. Data were then sorted into common midpoint (CMP) gathers and additional predictive deconvolution was applied to reduce the ringing character of the data. Next, CMP gathers were combined into super-gathers and f–k filtered to suppress multiple energy. The f–k filtering was achieved by applying a moveout correction at 80 per cent of the picked stacking velocities, transforming the data to the f–k domain, and then discarding the positive wave numbers before transforming back to the t–x domain (e.g. Yilmaz 1987). Semblance analysis was computed on these filtered gathers and the velocities were repicked prior to final moveout correction and stacking. Last, a post-stack Stolt constant velocity migration was computed using a velocity of 1500 m s^{-1}.

3.2 Data processing: deep water

The coherent noise problems are absent in the deep-water data. Prestack depth migration using a Kirchhoff algorithm was done on all the data where the water depth is >3500 m. This included all of Line 101 and 115 km of the seaward-most data from Line 1.

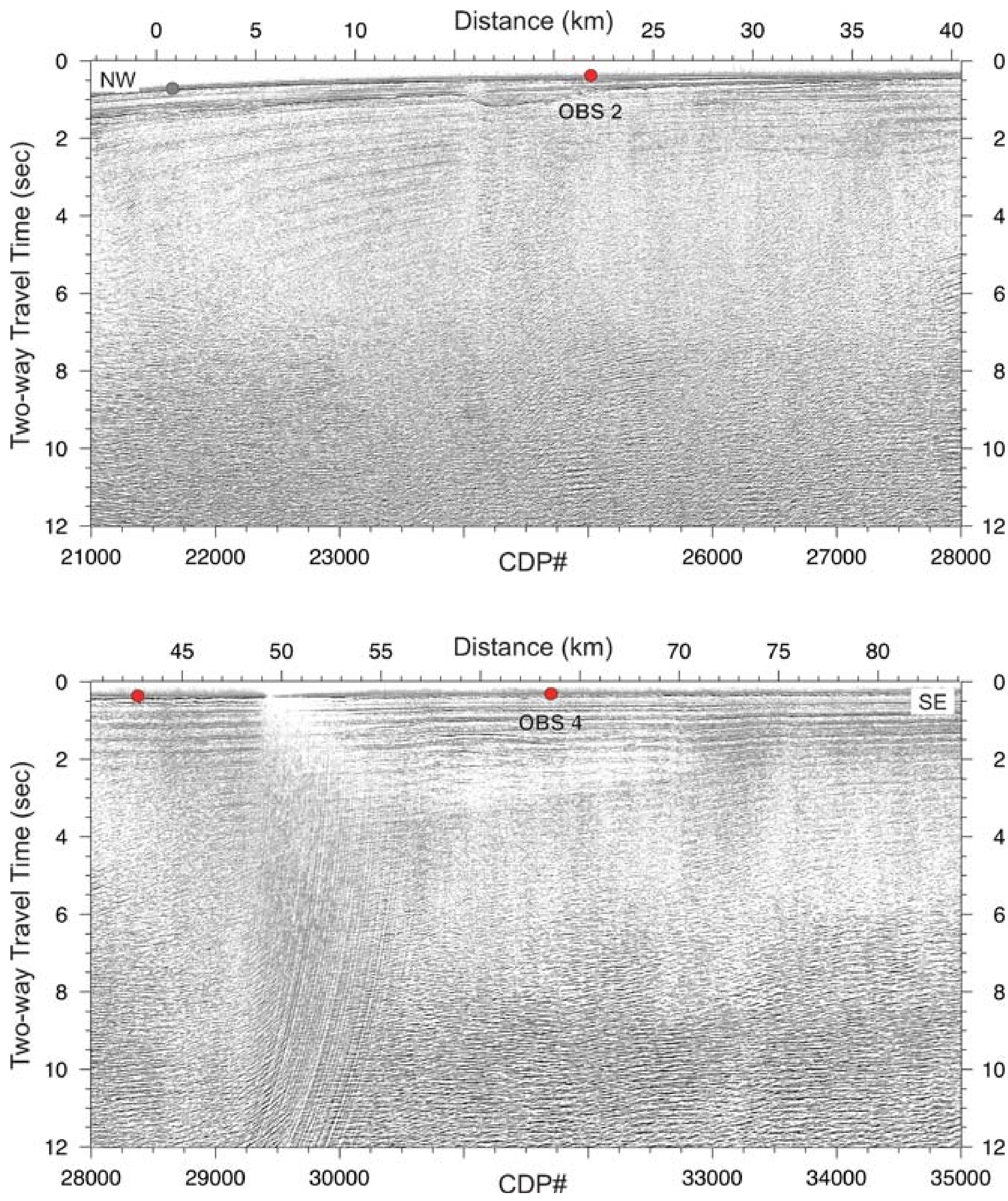

Figure 3. Segment of multichannel seismic reflection profile from SCREECH Line 1 over Flemish Cap. Dots are OBS locations. Grey dots indicate instruments that did not record good quality data. Red dots indicate instruments with data used in constructing the velocity model. CDP = Common depth point. See Funck *et al.* (2003) for full details on the refraction data. Moho reflection is marked where it is clearly defined. See text for discussion and information on data processing.

Because the main velocity contrast in the data is between sediment and basement, it is critical to the success of the migration that the basement be accurately mapped in depth. The migration velocity model was built up iteratively by first working through the sediments and under-migrating everything below. Once the sedimentary layers were satisfactorily migrated, the basement was picked and the velocities were increased slowly for several iterations until the basement surface was fully migrated. The deeper section presents two challenges for migration. First, deeper, high-velocity arrivals are difficult to constrain even with a 6 km streamer. Second, the deeper arrivals are lower amplitude and more poorly defined. Where lack of reflectivity or poor focusing analyses made building the velocity model difficult, we relied on the wide-angle results as a guide. The velocities were increased until the deeper section appeared fully migrated. In total, eight iterations of velocity picking and model refinement were necessary to migrate Line 1 and six iterations were required to complete Line 101.

4 IMPLICATIONS OF NEW SEISMIC DATA

The new data reveal several important features relating to the evolution of the region beginning with orogenesis during the Palaeozoic and ending with extensional and break-up processes during the Mesozoic. Here, we describe the key features seen in the seismic images and, where appropriate, compare to previous data collected nearby.

4.1 Lower-crustal reflectivity

Although the data across Flemish Cap is highly reverberatory and the imaging quality is generally poor, there is nonetheless coherent reflectivity visible in the section from 5 to 6 s two-way traveltime (TWTT) and deeper (Figs 3 and 4). The white line in Fig. 4 represents

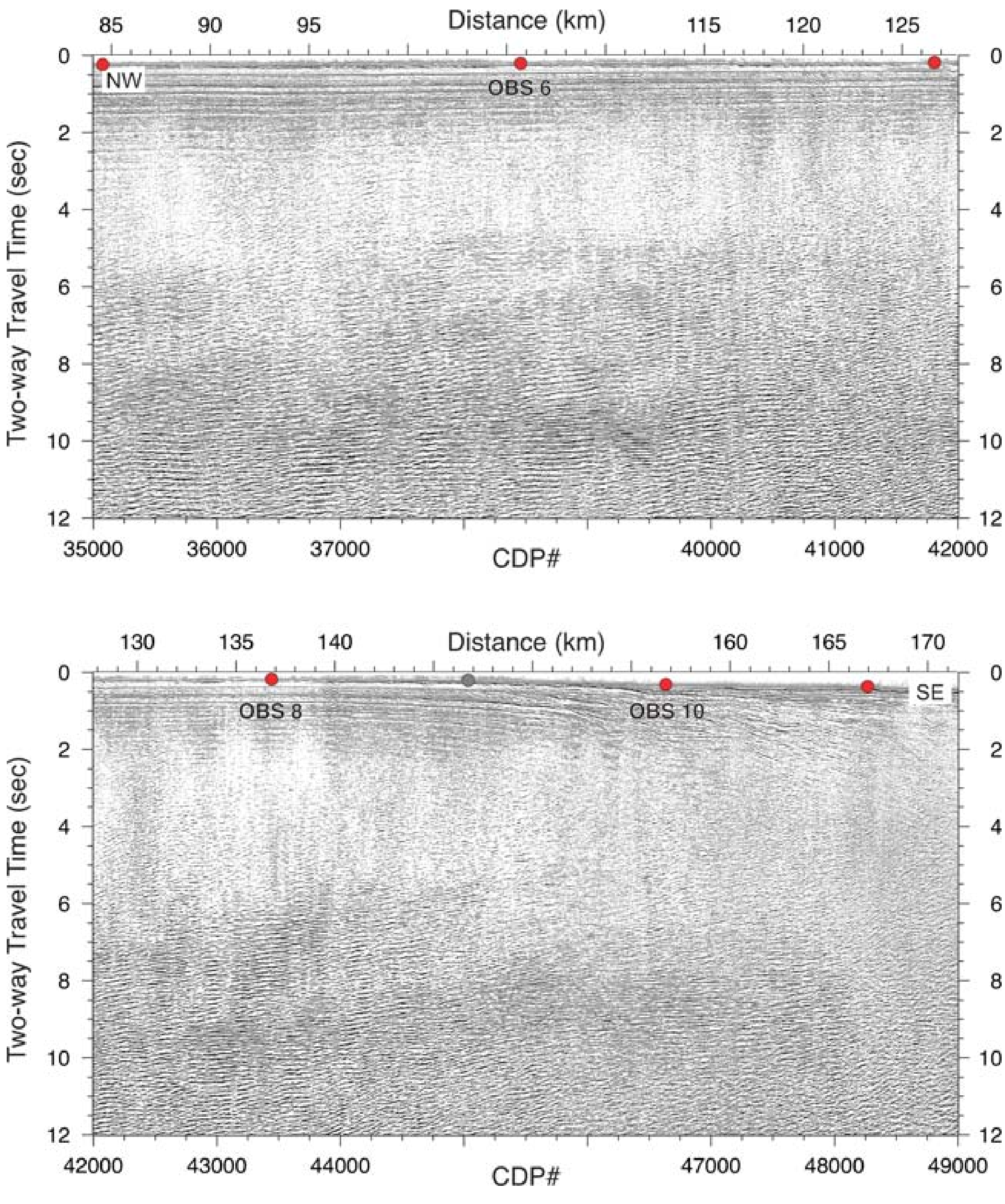

Figure 3. (*Continued.*)

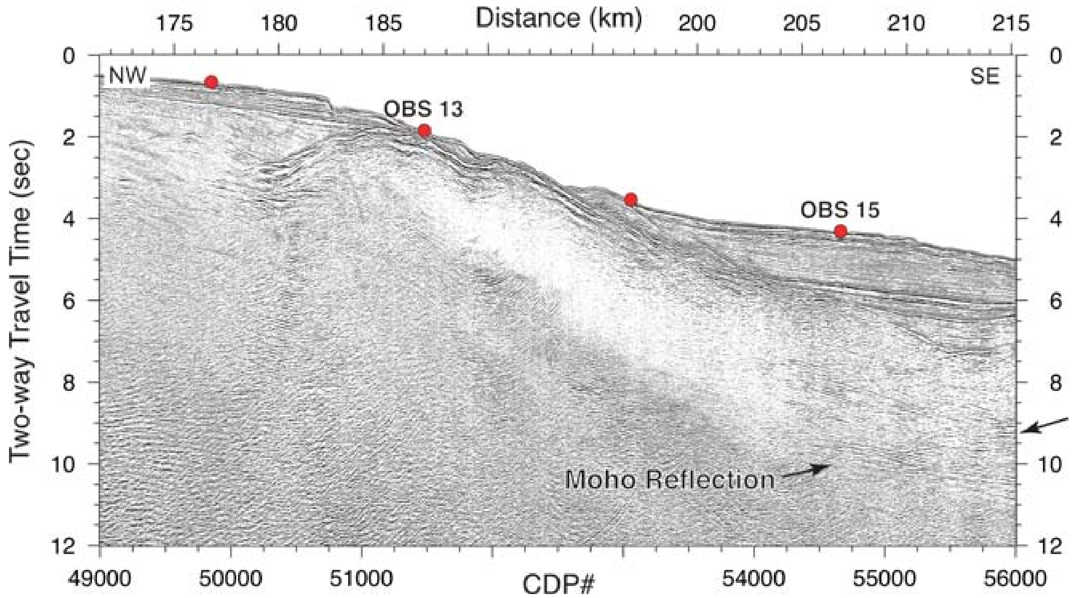

Figure 3. (*Continued.*)

the top of the lower crust determined independently by modelling the wide-angle data (Funck *et al.* 2003). The observed reflectivity is primarily confined to the lower crust and is subhorizontal to slightly landward dipping, although reflections reach into the middle crust at km 85–105 and km 35–45, with very coherent landward-dipping reflections at the latter location.

A key question is the extent to which the reflections interpreted on Fig. 4 are truly representative of real structure or are instead coherent noise artefacts that cannot be removed from the data. Generally, the interpreted structures do not follow the overlying structure in predictable ways as would be expected from internal multiples or reverberations from shallow-water features. For example the mid-crustal (5–5.5 s TWTT) reflections at km 40 are too steep to be from structure above. Similarly, packages of apparent NW dipping reflections from km 130–160, at 6–10 s TWTT, dip in the direction opposite that of the overlying structure and seafloor. Thus, while individual reflections interpreted on Fig. 4 should be treated with caution, the overall pattern of reflectivity seems representative of real structure.

Beneath the continental slope, the imaging quality is exceptionally poor and it is not possible to determine reflection characteristics. Farther seawards, however, distinct reflectivity in the middle and lower crust is again observed. A strong and coherent pattern of deep seaward-dipping reflectivity is seen from km 205 to 215 (Fig. 3). Weaker subhorizontal events are also observed in the middle crust. Both sets of reflections are clear in the time section (Figs 3 and 4) and in the depth section (Fig. 5). The deep, seaward-dipping reflections clearly continue from the lower crust into the mantle (Figs 3 and 5). A well-defined reflection Moho shallows from 10 s TWTT at km 207 to 9.3 s TWTT at km 215. Over this interval, the seaward-dipping reflections are observed as deep as 10.5 s TWTT. The possibility that these reflections are peg-leg multiple energy or out-of-plane energy (e.g. side swipe) has been considered. In both cases, such energy will be low velocity. In the first case, only horizons within the sedimentary sequence are candidates for producing peg-leg multiples. In the second case, out-of-plane energy would most likely result from a rough basement surface, and the reflected energy would have velocities appropriate for the overlying sediments. Such low-velocity energy will significantly degrade pre-stack migration, however. In the pre-stack depth migration focusing analysis, the bulk of the reflections did not generate low-velocity coherence peaks. In the final depth migration (Fig. 5, and Hopper *et al.* 2004), velocities of 6.4–6.8 km s^{-1} were used in this region. While some events appear to be over-migrated and thus may be either multiple or out-of-plane energy, the bulk of the reflections imaged in the depth section appear well migrated, which gives us confidence that most of these are real, deep reflections.

The highly reflective nature of continental crust in this region is well known (e.g. Keen *et al.* 1987; Keen & de Voogd 1988; van der Velden *et al.* 2004). Early hypotheses that reflectivity in the lower crust resulted from magmatic intrusions and sills emplaced during late stages of lithosphere extension are generally no longer favoured because of subsequent work demonstrating that break-up along Newfoundland was largely magma starved (e.g. Reid 1994). For Flemish Cap, the lack of evidence for significant extensional structures also is at odds with the large amounts of extension and decompression melting that would be required to create such a widespread and pervasive lower-crustal fabric by magmatic intrusion (e.g. Bown & White 1995). In addition, mafic underplating and intrusions would increase the average seismic velocity above that for typical Appalachian lower crust. Such a velocity increase is not observed in the wide-angle data (Funck *et al.* 2003).

A second explanation for the reflectivity is that it results from shear fabric that could be created by ductile flow in the lower crust (Keen *et al.* 1987; Reid & Jackson 1997). Lower-crustal flow during extension leading to break-up, however, may be difficult to reconcile with the magma-starved nature of extension. For example, Hopper & Buck (1998) showed that for a wide range of possible lower-crustal mineralogies, significant lower-crustal flow is predicted only for elevated lower-crustal temperatures. This would imply a warm mantle that would be expected to melt. More importantly, lower-crustal flow is likely to be restricted to a few km just above Moho and the penetrative shear fabric should be subparallel to the Moho (Bird 1991; Buck 1991; Hopper & Buck 1996). In contrast, the reflective pattern observed here begins at and is most pronounced at the top of the lower crust, approximately 10–15 km above Moho (Funck *et al.* 2003).

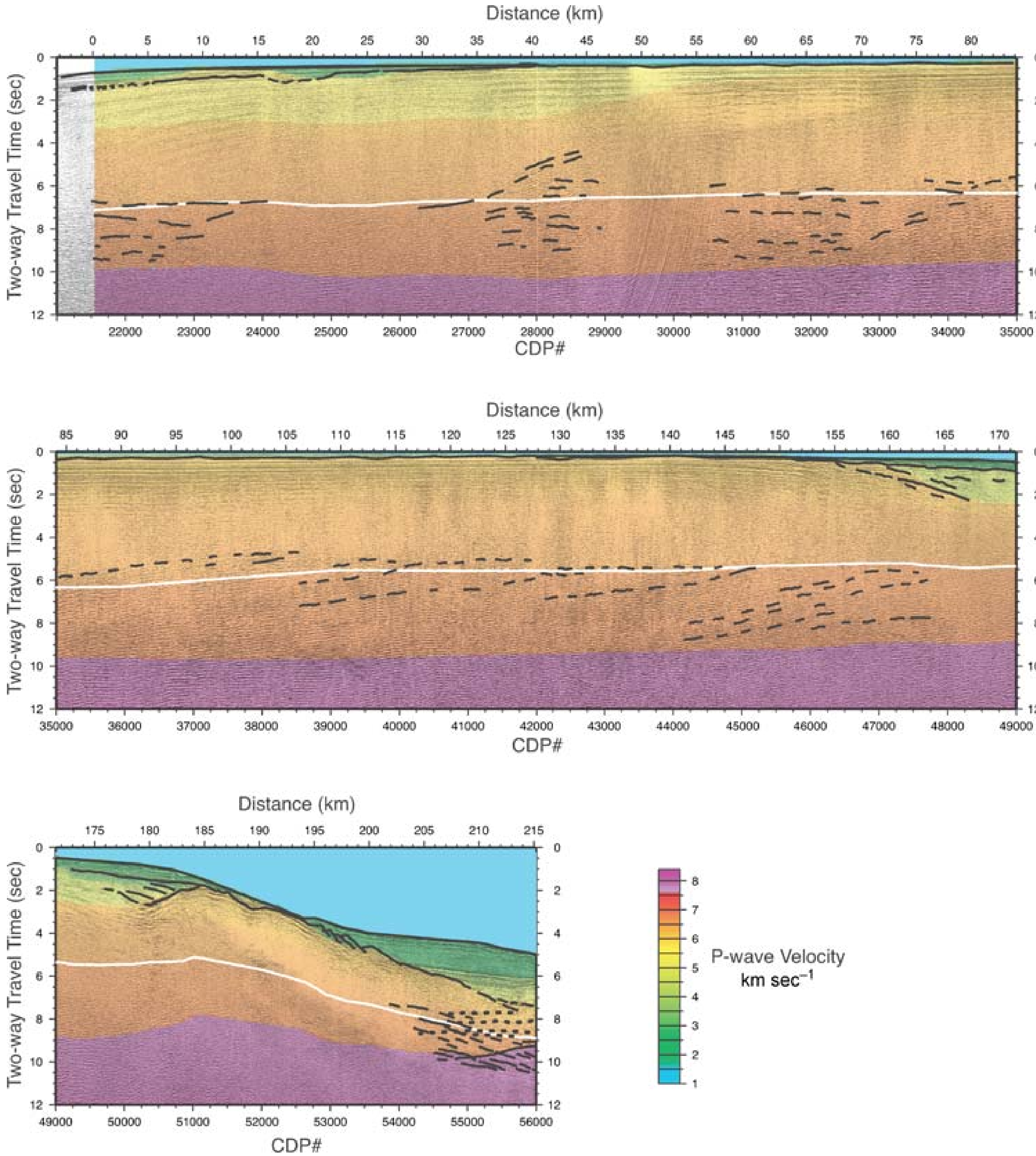

Figure 4. Segment of multichannel seismic reflection profile from SCREECH Line 1 over Flemish Cap (Fig. 3) with interpretation. The colour overlay shows seismic velocities derived from wide-angle data, with purple indicating mantle velocities (Funck *et al.* 2003). The white line indicates the top of the lower crust based on the velocity data. CDP = common depth point. See Fig. 5 for the seaward continuation of this profile over the interpreted continent–ocean boundary region

Lower-crustal flow and shearing associated with orogenic collapse that occurred long before extension finally led to break-up is also a possibility (Hall *et al.* 1998), although the depth range over which this fabric will develop is also likely to be restricted to the lowermost, warmest crust. While this may explain the lower-crustal reflectivity in the section from km 60–160, it cannot explain the seaward-most reflectivity. The reflections there are at a high angle to the Moho, varying from 40° to 60°, and they clearly continue into the mantle. Hall *et al.* (1998) noted several instances where deep reflectivity cuts the Moho, and they argued that such events are most likely the result of collisional orogenic processes rather than collapse processes. In addition, a deep reflection extending into the mantle is observed along the Goban Spur conjugate to the northeast Flemish Cap (Louden & Chian 1999). This too is suggested to be an older collision related structure. Such sutures could be preserved across the Moho, and they seem to offer the most viable explanation for the observed reflectivity.

From these considerations, reflectivity in the lower crust throughout the section appears to be pre-existing structure largely unrelated to extension and opening of the North Atlantic. Beneath Flemish Cap, it might be related to collapse structures, but it could equally well be caused by older orogenic structures associated with the

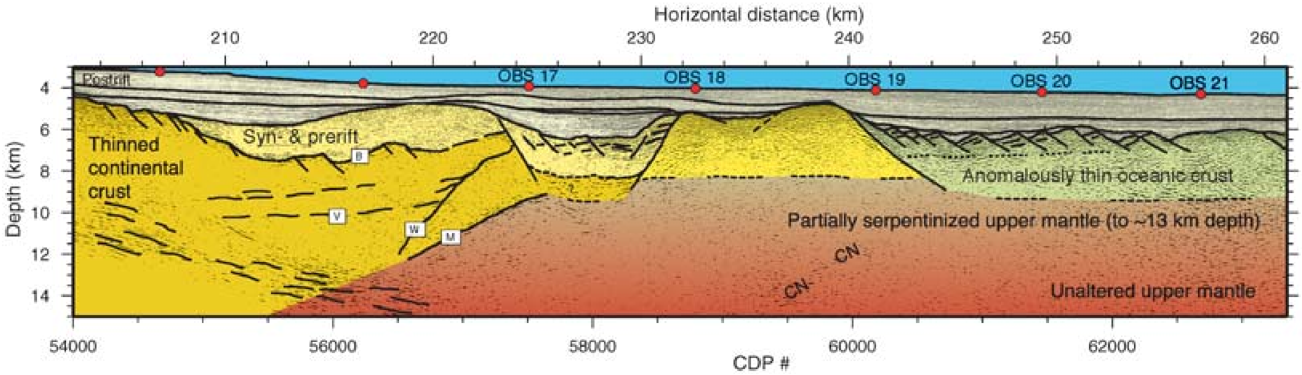

Figure 5. Pre-stack depth migrated reflection data and interpretation of a part of SCREECH Line 1 across the continent–ocean boundary region at the seaward edge of Flemish Cap (from Hopper *et al.* 2004). CN: coherent noise. B: top of continental basement or high-velocity sediments. V and W: prominent reflections discussed in Hopper *et al.* (2004); it is unclear whether V is within continental crust or high-velocity sediments. M: Moho reflection. CDP = common depth point. The colours do not indicate velocity and have no relation to the scale in Fig. 4. Yellow colours denote sediments and interpreted continental crust; green denotes interpreted oceanic crust; and brown denotes upper mantle. See Fig. 9 for the seaward continuation of this profile over interpreted oceanic crust.

construction of the Avalon terrain, which are known from onshore work to be reflective (e.g. van der Velden *et al.* 2004). The seaward-most reflections that cross Moho are almost certainly older sutures. The data thus have important implications for understanding Appalachian tectonics, reconstructing the Palaeozoic geography of the region, and interpreting the possible significance of pre-existing structure for later extension and break-up tectonics.

4.2 First-order asymmetry

One of the most important results from the Line 1 profile is that it firmly documents first-order asymmetry of the conjugate margins at a variety of scales. Although such asymmetry has been hypothesized previously based on extensive work on the Iberian margin and more limited work on the Newfoundland margin, this is the first data set with coincident wide-angle refraction and reflection data at positions conjugate to the major transects off Iberia. This allows us to quantify the nature of large-scale asymmetry in greater detail and to begin assessing possible causes and consequences of the asymmetry.

Fig. 6 shows cross-section reconstructions of data sets from the Flemish Cap and Galicia Bank margins. Fig. 6(a) represents a composite of several data sets and is intended to show the large-scale, first-order structure of the conjugate profiles; thus it is relatively insensitive to uncertainties of the plate reconstruction (Fig. 1b). Figs 6(b) and (c) show in detail the final, thinned continental crust on both margins immediately prior to mantle exhumation and/or magmatic seafloor spreading. Although conclusions based on this comparison are more sensitive to possible mismatch in conjugacy of the profiles, the Galicia Bank margin is covered by numerous seismic reflection profiles (Fig. 1b), which show that its main features within continental crust are consistent along strike.

The Iberian margin data begins onshore and ends seawards just short of the first peridotite ridge off Galicia Bank (based on profiles from Reston *et al.* 1996; Whitmarsh *et al.* 1996b; González *et al.* 1999; Pérez-Gussinyé *et al.* 2003). The SCREECH data are shown from the middle of Flemish Cap seawards to one of two places. In Figs 6(a) and (b), the splice point is where Funck *et al.* (2003) and Hopper *et al.* (2004) placed the continent ocean boundary. In Fig. 6(c), the splice point is at the westernmost location where alternate interpretations could possibly place the seaward limit of continental crust (see later discussion).

At a broad scale, the abrupt necking profile seawards of Flemish Cap (Fig. 6a) contrasts strongly with the Galicia Bank margin, which shows a wide zone of attenuated continental crust that thins gently westwards to very small thickness. The Galicia Bank thins from 16 to 3 km over a distance of nearly 100 km with an average Moho dip of $<8^\circ$. In contrast, the southeast Flemish Cap margin thins from 30 to 3 km over a distance of 80 km with an average Moho dip of 16°. Where it approaches the transition to oceanic crust, the Moho reflection at Flemish Cap is exceptionally strong and dips 30° in the depth images (km 220, Fig. 5). The transect probably does not follow a true dip line with respect to Moho, so the actual dip may be greater.

The contrasting abrupt and gradual necking profiles of the conjugate margins appear to be a general feature of this part of the rift. Keen *et al.* (1989) estimated an average Moho dip of $\sim 30^\circ$ along Line 85–3 (Fig. 2) across the northeastern margin of Flemish Cap, which rifted somewhat later than the southeast margin (e.g. Louden & Chian 1999). The Goban Spur conjugate, on the other hand, shows a much broader and gentler tapering of continental crust towards the oceanic basin (Keen *et al.* 1989). The contrasting necking profiles call to mind extension models of wide-rift mode; these models show that rifting is likely to localize and eventually break along one edge of the rift and leave a narrow/wide conjugate pair (e.g. Dunbar & Sawyer 1989; Hopper & Buck 1996).

Flemish Cap itself appears to be unaffected by the prolonged history of extension in the region. Our reflection data show little evidence for extensional structures or basins, and high-velocity basement forms the Cap beneath a thin veneer of sediment. The only exception is a 4-km-deep basin at the inboard end of the necking profile (km 165–180, Fig. 4). Somewhat reduced velocities in the upper part of basement at the western end of Line 1 (km 0–50, Fig. 4) probably represent Palaeozoic metasedimentary and metavolcanic rocks rather than Mesozoic rift-related deposits (Funck *et al.* 2003). Flemish Cap Basin, which is about 3 km deep, is located just beyond the western end of Line 1 and is distinct from the Flemish Pass Basin farther west. It most likely is filled with Cretaceous sediments (Grant & McAlpine 1990) although it has never been sampled.

The refraction data show that Flemish Cap is 30 km thick (Funck *et al.* 2003), only slightly thinner than the typical 35 km values measured for Newfoundland Appalachian crust (e.g. Marillier *et al.* 1994) and comparable to full-thickness Iberian crust, which is 32 km thick (Cordoba *et al.* 1987). Thus, Flemish Cap must be a relatively

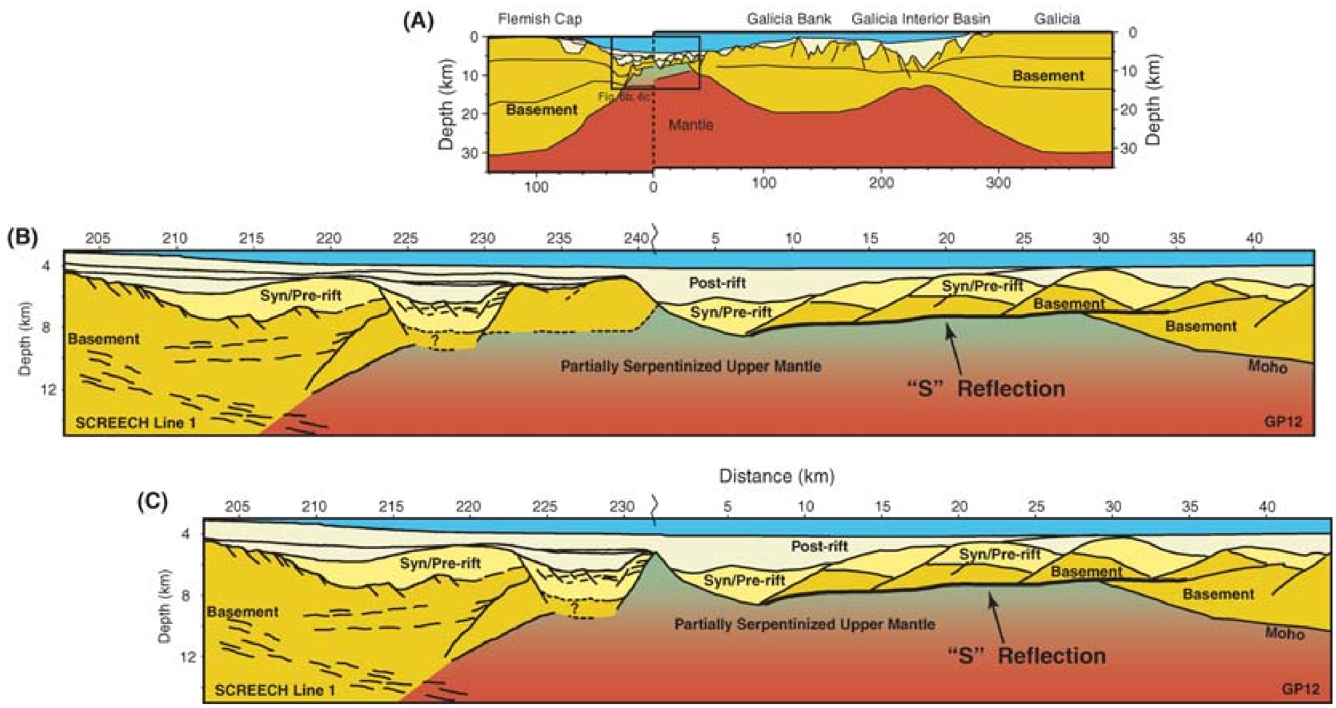

Figure 6. Interpretations of the eastern part of SCREECH Line 1 and representative data from conjugate profiles across the Iberia margin, spliced together at their seaward limits of continental crust. Data from the Iberia margin have been shifted upwards by ~1 km so that seafloor depth matches that on the Newfoundland margin. Yellow colours denote sediments or continental crust, brown denotes upper mantle, and greenish-grey grading to red denotes serpentinized upper mantle. (a) Broad-scale reconstruction vertically exaggerated by 2.5. The Flemish Cap section is from Funck *et al.* (2003) and the conjugate sections are from Pérez-Gussinyé *et al.* (2003), Reston (1996), Cordoba *et al.* (1999) and Whitmarsh (1996b). (b) Detail of the reconstruction immediately before exposure of the peridotite ridge off Galicia Bank. The Newfoundland data is SCREECH Line 1 (here and Hopper *et al.* 2004) and the conjugate data are from profile GP12 of Reston (1996); both the profiles from which interpretations were made were pre-stack depth migrated. The continent–ocean boundary position of Funck *et al.* (2003) and Hopper *et al.* (2004) is assumed, with the crustal block at km 232–241 interpreted as continental. Bold line is the 'S' reflection. (c) An alternate reconstruction that assumes the block of crust from km 232–241 is serpentinized peridotite rather than continental crust.

strong block of continental crust and lithosphere. It seems likely that rifting propagated around this strong block, consistent with its behaviour as a microplate (Srivastava *et al.* 2000).

Aside from the broad-scale character of the conjugate necking profiles, important asymmetry is also seen at smaller scales in the final stages of rift structuring immediately prior to break-up. Figs 6(b) and (c) show two, alternate cross-section reconstructions of the continental rift immediately prior to exposure of the peridotite ridge observed off Galicia Bank. Funck *et al.* (2003) and Hopper *et al.* (2004) placed the continent–ocean boundary at km 241 off Flemish Cap (Fig. 6b). This interpretation relies heavily on the seismic velocity information. In particular, OBS 19 (Fig. 5) provides clear evidence for a sharp lateral velocity boundary at this location. A distinct set of arrivals corresponding to oceanic layer 3 is observed to the east (Funck *et al.* 2003), and the crust there is best modelled assuming a typical oceanic crustal layering that is anomalously thin, only 3–4 km thick. This is consistent with observations of oceanic crust from ultra-slow seafloor spreading centres (Coakley & Cochran 1998; Muller *et al.* 2000; Jokat *et al.* 2003).

The crustal block from km 232–241, however, cannot be interpreted unambiguously as continental crust. An alternative would be to place the seaward termination of continental crust at km 232 (Fig. 6c). In this scenario, the block of crust from km 232–241 is serpentinized peridotite and Flemish Cap thus has a narrow transition zone consisting of exhumed upper mantle (ZECM) rather than a distinct continent–ocean boundary. The best-studied area of such exhumed mantle is along the Lusigal 12 and IAM-9 profiles off Iberia (Fig. 7). These profiles show ridges with irregular surfaces, which are known from drilling to be serpentinized peridotite (e.g. Whitmarsh *et al.* 1998). In addition, these profiles often show unreflective upper basement and increasing reflectivity with depth into basement (Pickup *et al.* 1996). On the IAM-9 profile in particular, the basement reflection is very weak and is primarily identifiable as the base of clear sedimentary layering (Fig. 7b). In addition, the velocity structure of this type of crust is distinct from both continental and oceanic crust. Dean *et al.* (2000) and Chian *et al.* (1999) show that the serpentinized 'crust' consists of a layer with extremely high-velocity gradients, from velocities of 4.5 km s^{-1} to more than 7 km s^{-1}, consistent with rapidly decreasing serpentinization with depth.

Unfortunately, it is unclear to what extent these observations are truly diagnostic of exhumed mantle. Hopper *et al.* (2004) argued that the distinct and smooth basement reflection from the block at km 232–241 off Flemish Cap is unlike that observed off Iberia, and thus they suggested that it may be continental (Fig. 7). However, recent study of peridotite ridges off Iberia shows that many have smooth surfaces, possibly from erosion of the weak serpentinites (Tucholke *et al.* 2006). More importantly, the wide-angle data off Flemish Cap do not show evidence for the high-velocity gradients in the upper crust (Funck *et al.* 2003), although it should be emphasized that this is not well resolved because the block is a small feature. Thus the origin of this block remains uncertain, and the alternative presented in Fig. 6(c) cannot be ruled out by the data currently available.

In either case, the crustal structure at the time of continental break-up is characterized by marked asymmetry. A key result from surveys on the western Galicia Bank margin has been the identification and

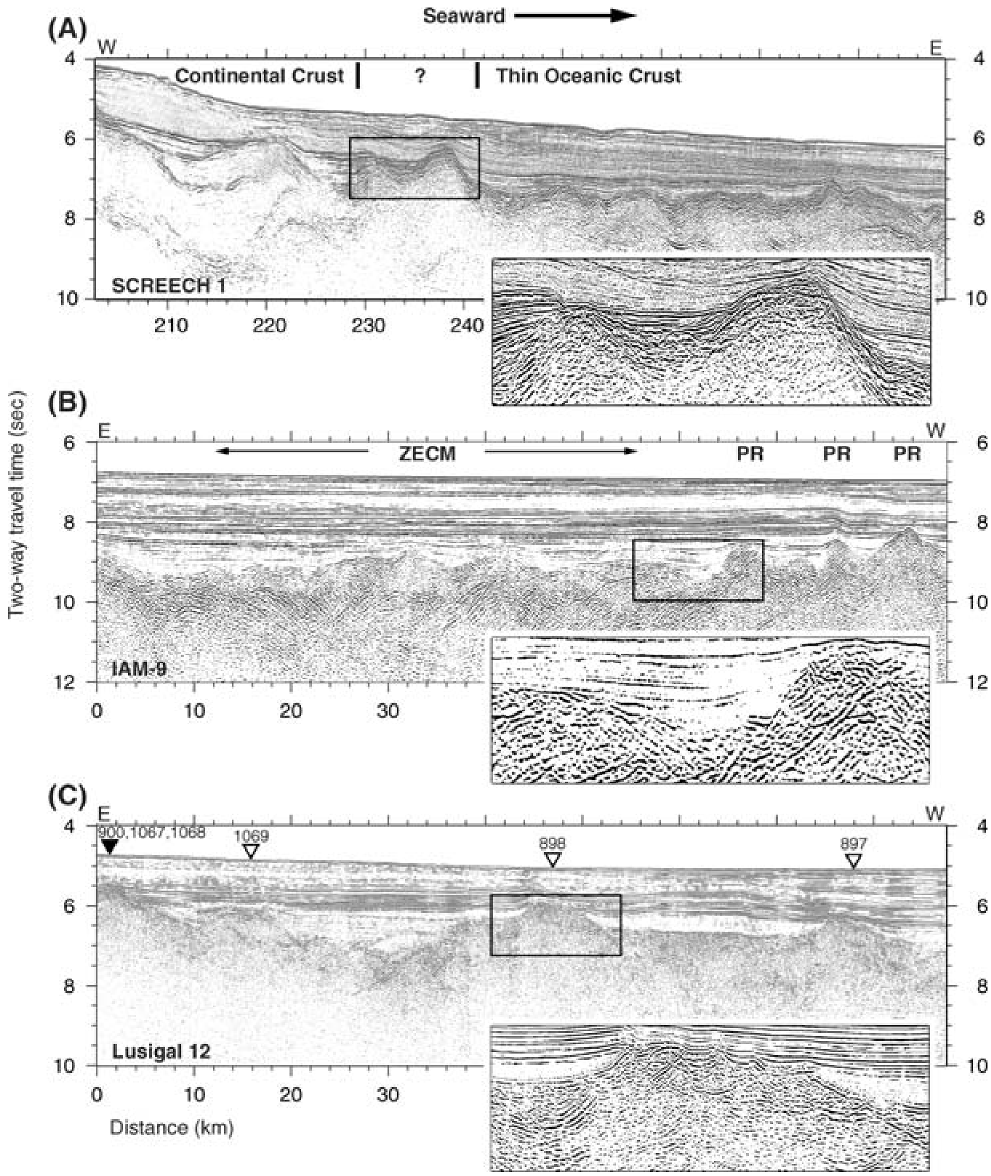

Figure 7. Segment of multichannel reflection profile SCREECH Line 1 across the continent–ocean boundary region (a), compared to segments of multichannel profiles across the southern Iberia Abyssal Plain where basement consists of exhumed, serpentinized mantle (b and c). The IAM-9 profile (b) is from Pickup *et al.* (1996) and the Lusigal 12 profile (c) is from Beslier (1996); locations of the full profiles are shown in Fig. 2. Insets show enlarged sections of the profiles as indicated. Note that all profiles are oriented with the seaward direction to the right side of the figure. The reflection character of the basement block at km 232–241 in SCREECH Line 1 differs from that of the serpentinized basement in these two Iberia profiles; see text for discussion. Along IAM-9, 'PR' labels indicate locations of interpreted peridotite ridges (e.g. Russell & Whitmarsh 2003). ZECM = zone of exhumed continental mantle. Along Lusigal 12, triangles locate ODP Legs 149 and 173 drill sites (Whitmarsh *et al.* 1996a, 1998).

mapping of a strong reflection, termed 'S'. It is observed on conjugate profiles GP-101, 102, 11, and 12 as well as ISE-1, 2, 4, 5, 6 and possibly 15, but it disappears southwards in the area of ISE-14 (see Fig. 1b for locations; Reston *et al.* 1996; Henning *et al.* 2004). The reflection is particularly well imaged on the depth sections along profile GP12, which is depicted in Figs 6(b) and (c). 'S' is described as a top-to-the-west detachment surface that formed at the crust–mantle interface during the final stages of break-up (e.g. Reston *et al.* 1996; Pérez-Gussinyé & Reston 2001). It thus represents a major compositional, and more importantly, rheological boundary.

Within unequivocal continental crust of Flemish Cap west of km 232, we see no evidence for an 'S'-like surface. Given that the reflection is so clear over a wide area on the conjugate Galicia profiles, we find it unlikely that this lack of an 'S' reflection results from a mismatch in profile conjugacy caused by uncertainty in plate reconstruction. The steeply dipping Moho reflection under Flemish Cap, which might be interpreted as a continuation of 'S', is crossed by the lower-crustal to upper-mantle reflections described earlier (Figs 4–6). The apparent lack of offset in these reflections indicates that the Moho reflection is not a detachment surface like that interpreted at Galicia Bank. The Moho reflection at Flemish Cap, however, is discontinuous, and defining a more precise relationship between it and the crossing reflections will require better imaging than is presently available.

4.3 Oceanic crust and ultra-slow seafloor spreading

An important component of understanding the final break-up of continents is to determine when and where mantle melting begins and at what point there is sufficient melt productivity to generate an oceanic crust. As described earlier, there is clear evidence suggesting that thin oceanic crust was produced at an ultra-slow seafloor spreading system beginning at km 241 on Line 1. 10 km to the north, a second MCS profile was shot (Line 101) that provides additional insight into early production of oceanic crust off Newfoundland.

Fig. 8 shows a pre-stack depth migrated image of Line 101. Several key features can be noted. A short section of crust where the top-basement reflection is very weak is found in the west at km 256. Basement there is marked only by the base of clearly stratified post-rift sediments. This is similar to the top of basement seen in the ZECM along the IAM-9 profiles (Pickup *et al.* 1996; Fig. 7). Unlike the Iberian margin, however, there is almost no basement relief and there is no increase in reflectivity with depth. Thus, there is some doubt about how to interpret this section based on reflection data alone. This area is bounded seawards by a deep basin filled with horizontal sediments. Segments of strong, low-frequency reflections within this fill could be sills. Much of the remainder of the section consists of domino-style fault blocks similar to those observed along the oceanic section of Line 1 from km 240–260. Unfortunately, we lack wide-angle data along this profile, so velocity information is available only from focusing analysis that was done for the depth migration. This means that there are essentially no constraints on crustal velocities below the basement surface.

The general character of the reflection images is similar along Lines 1 and 101, suggesting that the sections share a common origin (Fig. 9). In detail, however, there are key differences. In particular, it is difficult to correlate single features on Line 1 with features on Line 101. Thus, there is significant along-strike variation, indicating either that individual structures are small scale (<10 km) or, perhaps less likely, that a segment boundary separates the profiles. Notably, the strong 'Z' reflection observed on Line 1 (Figs 5 and 9) is not seen on Line 101. This implies that the 'Z' reflection has limited extent, which has important implications when interpreting the origin of extremely thin, 1–1.5-km-thick crust observed from km 280–290 along Line 1 (Fig. 9). This thin crust appears to be oceanic despite the fact that oceanic layer 3 is missing (Hopper *et al.* 2004). It exhibits tilted and rotated fault blocks more commonly associated with continental crust, but to explain this section as continental crust would require a ridge jump with an abandoned rift valley at some poorly defined location to the west. In this scenario, the crust would be a slice of continental Galicia Bank that was stranded on the Newfoundland margin, and the strong 'Z' reflection could be a segment of the Galicia 'S' reflection. However, because 'S' is a strong regional reflection on the Galicia margin and 'Z' is a much smaller-scale and apparently local feature, it is unlikely to be a conjugate segment of 'S'. This removes one argument that the overlying crust might be continental.

Stratified and rotated fault blocks such as those observed at km 280–300 (Fig. 9) seem not to be uncommon in oceanic crust. In particular, Srivastava & Keen (1995) showed that crust from the abandoned, ultra-slow-spreading centre in the Labrador Sea contains large blocks of oceanic crust that are similar to those observed here (Fig. 10). Salisbury & Keen (1993) also argued that rotated and stratified blocks observed off Nova Scotia are oceanic, although recent velocity modelling suggests that this may not be the case (Wu *et al.* 2006). In magma-starved settings, such stratification in oceanic crust may be created by basalt flows interlayered with

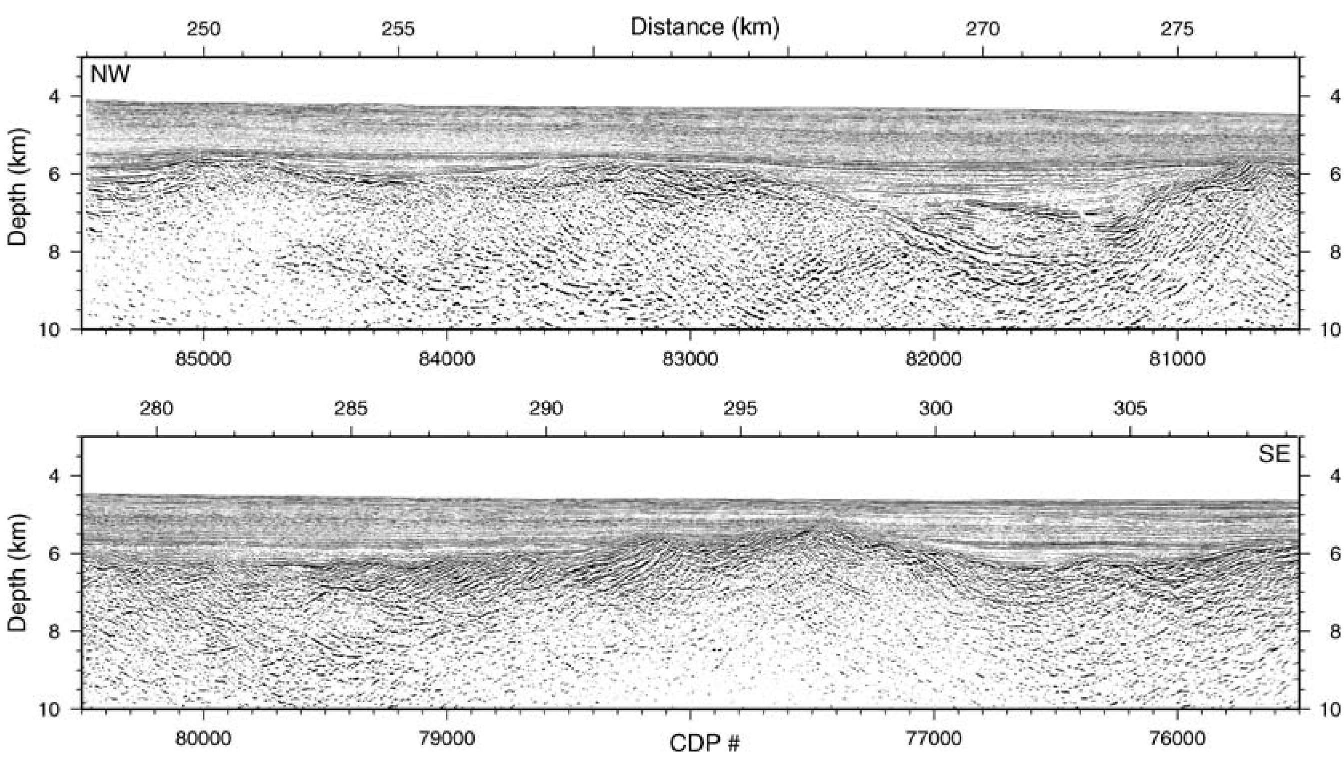

Figure 8. Pre-stack depth migrated reflection profile along SCREECH Line 101, located 10 km northeast of Line 1. Data are plotted with no vertical exaggeration. CDP = common depth point. The distance scale at top is marked such that numbers reflect approximate along-strike correlations to SCREECH Line 1 to the southwest (see Fig. 9 for a direct comparison between lines). See text for processing and description.

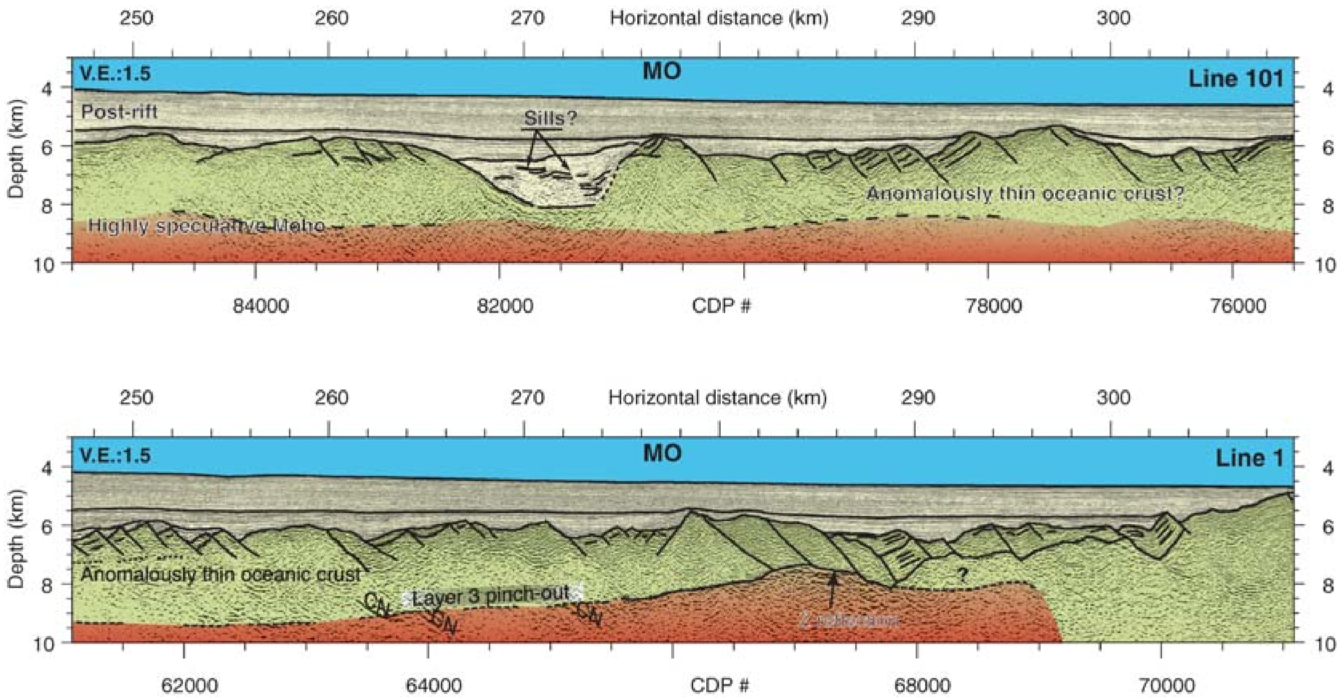

Figure 9. Pre-stack depth migrated profiles of SCREECH Line 101 and the corresponding, along-strike part of Line 1, with interpretations superimposed. CDP: common depth point. The profiles are aligned to correlate approximately along strike and are plotted with a vertical exaggeration of 1.5:1. Note that with the exception of the layered, domino-faulted basement centred at ~km 290, there is little correlation of structural features between profiles. See text for discussion. CN: coherent noise. Green colours denote interpreted oceanic crust and brown colours denote upper mantle.

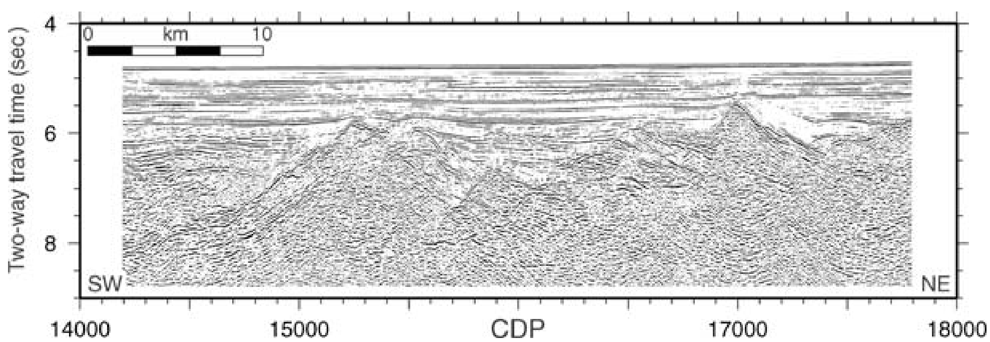

Figure 10. Section of Line 90–2 from Srivastava & Keen (1995) showing ultra-slow spreading crust. CDP: common depth point. The centre of the abandoned spreading is at ~CDP 14 000 and corresponds to magnetic anomaly 13. Magnetic anomaly 21 is at ~CDP 18 000. The half spreading over this interval is 3–3.5 mm yr^{-1}. Note in particular the layered nature of the crust and the rotated tilted fault blocks.

pelagic sediments. Subsequent tectonic extension of these deposits could produce fault blocks similar to those found in continental extensional basins. However, Osler & Louden (1995) and Louden *et al.* (1996) showed that in the case of the Labrador Sea ultra-slow spreading crust, seismic phases corresponding to oceanic layer 3 are observed in wide-angle data near the Srivastava & Keen (1995) profile. Thus the process noted above does not explain the absence of oceanic layer 3 on Line 1.

To explain the missing layer 3, Hopper *et al.* (2004) proposed a conceptual model where the layer is mechanically removed during a period of extreme melt starvation that resulted in creation of an oceanic core complex. It is also possible that much of the gabbroic melt might not reach crustal levels in very-slow-spreading environments, as suggested by recent studies in old North Atlantic crust and along the Mid-Atlantic ridge (Lizarralde *et al.* 2004; Shipboard Scientific Party 2004a). In extreme cases, gabbro could be sequestered in discrete bodies in the mantle while only a thin layer-2 crust accretes at the sea floor. This may be the case for part of the Gakkel Ridge today (Jokat *et al.* 2003). A last possibility that cannot be excluded by data presently available is that Line 1 runs across an oceanic fracture zone, where thin crust and mantle exhumation is commonly observed in slow-spreading environments (e.g. Tucholke *et al.* 1998; Canales *et al.* 2004).

Overall, our data suggest that crust immediately east of Flemish Cap has highly variable and strongly 3-D structure at small scales (±10 km) and that it probably was produced in an ultra-slow spreading environment. While Hopper *et al.* (2004) argued for large temporal variations in the melt supply, it is clear from the data here that large spatial variations in melt supply are also important. For example, along Line 101 the deep basin at km 272 (Fig. 9) probably indicates a period of extreme magma starvation and tectonic extension, while along strike on Line 1 (and thus presumably at about the same time) a higher level of magmatism was generating 4-km-thick oceanic crust. Large variability in melt production in both space and time is likely a general feature of ultra-slow spreading environments, and it merits further investigation.

5 CONCLUSIONS

Seismic reflection and refraction data from the southeastern Flemish Cap margin show abrupt thinning of continental crust from 30 to 3 km thick over a distance of 80 km. Over this interval, reflection Moho is defined by both wide-angle and MCS data, and it locally dips as steeply as 30°. Extremely thin continental crust is observed only in a very narrow zone, and evidence for a conjugate counterpart to the 'S' reflection observed beneath Galicia Bank is lacking. Anomalously thin oceanic crust either immediately abuts the thin continental crust or is separated from it by only a narrow (10 km) zone of exhumed mantle. There is no convincing evidence for a broad zone of exhumed mantle off Flemish Cap. These observations document major, first-order asymmetry in the final development of continental rifting that led to seafloor spreading between Flemish Cap and Galicia Bank. The data also shed new light on the nature of lower-crustal reflectivity observed beneath Flemish Cap. The reflections are most likely older, pre-existing structures that are unrelated to extension and break-up.

Finally, structural differences between parallel dip lines across the outer margin show that tectonic extension and melt production were highly variable both spatially and temporally in ultra-slow spreading ocean crust that was produced immediately following rifting. The data also suggest that initial seafloor spreading off Newfoundland occurred in a magma starved, but not completely amagmatic, setting. The structure and layer thicknesses of crust produced in such an environment show dramatic departures from those expected in 'normal' ocean crust.

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REFERENCES

Beslier, M.-O., 1996. Data report: Seismic line LG12 in the Iberia Abyssal Plain, in *Proceedings of the Ocean Drilling Program, Scientific Results,* Vol. 149, pp. 737–739, eds Whitmarsh, R.B., Sawyer, D.S., Klaus, A. & Masson, D.G, College Station, Texas, Ocean Drilling Program.

Bird, P., 1991. Lateral extrusion of lower crust from under high topography, in the isostatic limit, *J. geophys. Res.,* **96,** 10 275–10 286.

Boillot, G., Winterer, E.L., Meyer, A.W. & Shipboard Scientific Party, 1987. Introduction, objectives, and principal results: Ocean Drilling Program Leg 103, West Galicia Margin, in *Proceedings of the Ocean Drilling Program, Initial reports* Vol. 103, pp. 3–17, eds Boillot, G., Winterer, E.L. & Meyer, A.W., College Station, Texas, Ocean Drilling Program.

Bown, J.W. & White, R.S., 1995. Effect of finite extension rate on melt generation at rifted continental margins, *J. geophys. Res.,* **100,** 18 011–18 030, doi: 10.1029/94JB01478.

Buck, W.R., 1988. Flexural rotation of normal faults, *Tectonics,* **7,** 959–973.

Buck, W.R., 1991. Modes of continental lithospheric extension, *J. geophys. Res.,* **96,** 20 161–20 178.

Canales, J.P., Tucholke, B.E. & Collins, J.A., 2004. Seismic reflection imaging of an oceanic detachment fault: Atlantis megamullion (Mid-Atlantic Ridge, 30°10′N), *Earth planet. Sci. Lett.,* **222,** 543–560.

Channell, J.E.T., Erba, E., Nakanishi, M. & Tamaki, K., 1995. Late Jurassic–Early Cretaceous time scales and oceanic magnetic anomaly block models, in *Geochronology, Time Scales and Global Stratigraphic Correlation*, Society of Economic Paleontologists and Mineralogists (Soc. Sediment. Geol.) Special Publication 54, pp. 51–63, eds Berggren, W.A., Kent, D.V., Aubry, M.-P. & Hardenbol, J., Society of Economic Paleontologists and Mineralogists, Tulsa, Oklahoma.

Chian, D., Louden, K.E., Minshull, T.A. & Whitmarsh, R.B., 1999. Deep structure of the ocean-continent transition in the southern Iberia Abyssal Plain from seismic refraction profiles: Ocean Drilling Program (Legs 149 and 173) transect, *J. geophys. Res.,* **104,** 7443–7452.

Chu, D. & Gordon, R.G., 1999. Evidence for motion between Nubia and Somalia along the Southwest Indian ridge, *Nature,* **398,** 64–68.

Coakley, B.J. & Cochran, J.C., 1998. Gravity evidence of very thin crust at the Gakkel Ridge (Arctic Ocean), *Earth planet. Sci. Lett.,* **162,** 81–95.

Cordoba, D., Banda, E. & Ansorge, J., 1987. The Hercynian crust in northwestern Spain: a seismic survey, *Tectonophysics,* **132,** 321–333.

Dean, S.M., Minshull, T.A., Whitmarsh, R.B. & Louden, K.E., 2000. Deep structure of the ocean-continent transition in the southern Iberia Abyssal Plain from seismic refraction profiles: the IAM-9 transect at 40°20°N, *J. geophys. Res.,* **105,** 5859–5885.

Dick, H.J.B., Lin, J. & Schouten, H., 2003. An ultraslow-spreading class of ocean ridge, *Nature,* **426,** 405–412.

Dunbar, J.A. & Sawyer, D.S., 1989. Patterns of continental extension along the conjugate margins of the central and North Atlantic Oceans and Labrador Sea, *Tectonics,* **8,** 1059–1077.

Funck, T., Hopper, J.R., Larsen, H.C., Louden, K.E., Tucholke, B.E. & Holbrook, W.S., 2003. Crustal structure of the ocean-continent transition at Flemish Cap: seismic refraction results, *J. geophys. Res.,* **108,** 2531, doi: 10.1029/2003JB002434.

González, A., Córdoba, D. & Vales, D., 1999. Seismic crustal structure of Galicia continental margin, NW Iberian Peninsula, *Geophys. Res. Lett.,* **26,** 1061–1064.

Gradstein, F.M., Agterberg, F.P., Ogg, J.G., Hardenbol, J., van Veen, P., Thierry, J. & Huang Z., 1995. A Triassic, Jurassic and Cretaceous time scale, in *Geochronology, Time Scales and Global Stratigraphic Correlation*, Society of Economic Paleontologists and Mineralogists (Soc. Sediment. Geol.) Special Publication, Vol. 54, pp. 95–126, eds Berggren, W.A., Kent, D.V., Aubry, M.-P. & Hardenbol, J., Society of Economic Paleontologists and Mineralogists, Tulsa, Oklahoma.

Grant, A.C. & McAlpine, K.D., 1990. The continental margin around Newfoundland, in *Geology of the continental margin of eastern Canada,* Vol. 2, pp. 239–292, Keen, M.J. & Williams, G.J., Geological Survey of Canada, Geology of Canada.

Hall, J., Marillier, F. & Dehler, S., 1998. Geophysical studies of the structure of the Appalachian orogen in the Atlantic borderlands of Canada, *Can. J. Earth Sci.,* **35,** 1205–1221.

Henning, A.T., Sawyer, D.S. & Templeton, D.C., 2004. Exhumed upper mantle within the ocean-continent transition on the northern West Iberia margin: Evidence from prestack depth migration and total tectonic subsidence analyses, *J. geophys. Res.,* **109,** B05103, doi: 10.1029/2003JB002526.

Hopper, J.R. & Buck, W.R., 1996. The effect of lower crustal flow on continental extension and passive margin formation, *J. geophys. Res.,* **101,** 20 175–20 194.

Hopper, J.R. & Buck, W.R., 1998. Styles of extensional decoupling, *Geology,* **26,** 699–702.

Hopper, J.R., Funck, T., Tucholke, B.E., Larsen, H.C., Holbrook, W.S., Louden, K.E., Shillington, D. & Lau, H., 2004. Continental breakup and the onset of ultraslow seafloor spreading off Flemish Cap on the Newfoundland rifted margin, *Geology,* **32,** 93–96, doi: 10.1130/G19694.1.

Jokat, W., Ritzmann, O., Schmidt-Aursch, M.C., Drachev, S., Gauger, S. & Snow, J., 2003. Geophysical evidence for reduced melt production on the Arctic ultraslow Gakkel mid-ocean ridge, *Nature,* **423,** 962–965.

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

Keen, C.E., Boutilier, R., de Voogd, B., Mudford, B. & Enachescu, M.E., 1987. Crustal geometry and extensional models for the Grand Banks, eastern Canada: constraints from deep seismic reflection data, in *Sedimentary Basins and Basin-forming Mechanisms,* Vol. 12, pp. 101–115, eds Beaumont, C. & Tankard, A.J. Can. Soc. Pet. Geol., Memoir.

Keen, C.E., Peddy, C., de Voogd, B. & Matthews, D., 1989. Conjugate margins of Canada and Europe: results from deep reflection profiling, *Geology,* **17,** 173–176.

King, L.H., Fader, G.B., Poole, W.H. & Wanless, R.K., 1985. Geologic setting and age of the Flemsih Cap granodiorite, east of the Grand Banks, of Newfoundland, *Can. J. Earth Sci.,* **22,** 1286–1298.

Lavier, L.L., Buck, W.R. & Poliakov, A.N.B., 1999. Self-consistent rolling-hinge model for the evolution of large-offset low-angle normal faults, *Geology,* 1127–1130.

Louden, K.E. & Chian, D., 1999. The deep structure of non-volcanic rifted continental margins, *Phil. Trans. R. Soc. Lond. A,* **357,** 767–804.

Louden, K.E., Osler, J.C., Srivastava, S.P. & Keen, C.E., 1996. Formation of oceanic crust at slow spreading rates: new constraints from an extinct spreading center in the Labrador Sea, *Geology,* **24,** 771–774.

Lizarralde, D., Gaherty, J.B., Collins, J.A., Hirth, G. & Kim, S.D., 2004. Spreading rate dependence of melt extraction at mid-ocean ridges from mantle seismic refraction data, *Nature,* **432,** 744–747, doi:10.1038/nature03140.

Manatschal, G., 2004. New models for evolution of magma-poor rifted margins based on a review of data and concepts from West Iberia and the Alps, *Int. J. Earth Sci.,* **93,** 432–466.

Marillier, F. *et al.*, 1994. Lithoprobe East onshore-offshore seismic refraction survey; Constraints on interpretation of reflection data in the Newfoundland Appalachians, *Tectonophysics,* **232,** 43–58.

Muller, M.R., Minshull, T.A. & White, R.S., 2000. Crustal structure of the Southwest Indian Ridge at the Atlantis II Fracture Zone, *J. geophys. Res.,* **105,** 25 809–25 828.

Osler, J.C. & Louden, K.E., 1995. Extinct spreading center in the Labrador Sea: crustal structure from a two-dimensional seismic refraction velocity model, *J. geophys. Res.,* **100,** 2261–2278.

Pérez-Gussinyé, M. & Reston, T.J., 2001. Rheological evolution during extension at nonvolcanic rifted margins: onset of serpentinization and development of detachments leading to continental breakup, *J. geophys. Res.,* **106,** 3961–3975.

Pérez-Gussinyé, M., Ranero, S.R., Reston, T.J. & Sawyer, D., 2003. Structure and mechanisms of extension at the Galicia Interior Basin, west of Iberia, *J. geophys. Res.,* **108,** 2245, doi:10.1029/2001JB000901.

Pickup, S.L.B., Whitmarsh, R.B., Fowler, C.M.R. & Reston, T.J., 1996. Insight into the nature of the ocean-continent transition off west Iberia from a deep multichannel seismic reflection profile, *Geology,* **24,** 1079–1082.

Rabinowitz, P.D., Cande, S.C. & Hayes, D.E., 1978. Grand Banks and J-Anomaly Ridge, *Science,* **202,** 71–73.

Rasmussen, E.S., Lomholt, S., Andersen, C. & Vejbæk, O.V., 1998. Aspects of the structural evolution of the Lusitanian Basin in Portugal and the shelf and slope area offshore Portugal, *Tectonophysics,* **300,** 199–225.

Reid, I.D., 1994. Crustal structure of a nonvolcanic rifted margin east of Newfoundland, *J. geophys. Res.,* **99,** 15 616–15 180.

Reid, I.D. & Jackson, H.R., 1997. A review of three transform margins off eastern Canada, *Geo-Marine Lett.,* **17,** 87–93.

Reston, T.J., Krawczyk, C.M. & Klaeschen, D., 1996. The S reflector west of Galicia (Spain): evidence from prestack depth migration for detachment faulting during continental breakup, *J. geophys. Res.,* **101,** 8075–8091.

Russell, S.M. & Whitmarsh, R.B., 2003. Magmatism at the west Iberia non-volcanic rifted continental margin: evidence from analyses of magnetic anomalies, *Geophys. J. Int.,* **154,** 706–730.

Salisbury, M.H. & Keen, C.E., 1993. Listric faults imaged in oceanic crust, *Geology,* **21,** 117–120.

Shipboard Scientific Party, 2004a. Leg 209 Summary, in *Proceedings of the Ocean Drilling Program, Initial Reports*, Vol. 209, pp. 1–139, eds Kelemen, P.B., Kikawa, E. & Miller, D.J., College Station, Texas, Ocean Drilling Program.

Shipboard Scientific Party, 2004b. Leg 210 Summary, in *Proceedings of the Ocean Drilling Program, Initial Reports*, Vol. 210, pp. 1–78, eds Tucholke, B.E., Sibuet, J-C. & Klaus, A., College Station, Texas, Ocean Drilling Program.

Smith, W.H.F. & Sandwell, D.T., 1994. Bathymetric prediction from dense satellite altimetry and sparse shipboard bathymetry, *J. geophys. Res.,* **99,** 21 803–21 824.

Srivastava, S.P. & Keen, C.E., 1995. A deep seismic reflection profile across the extinct Mid-Labrador Sea spreading center, *Tectonics,* **14,** 372–389.

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

Srivastava, S.P., Sibuet, J.-C., Cande, S., Roest, W.R. & Reid, I.D., 2000. Magnetic evidence for slow seafloor spreading during the formation of the Newfoundland and Iberian margins, *Earth planet. Sci. Lett.,* **182,** 61–76.

Tankard, A.J. & Welsink, H.J., 1987. Extensional tectonics and stratigraphy of Hibernia oil field, Grand Banks, Newfoundland, *American Association of Petroleum Geologists Bulletin,* **71,** 1210–1232.

Tucholke, B.E., Austin, J.A. Jr. & Uchupi, E., 1989. Crustal structure and rift-drift evolution of the Newfoundland Basin, in *Extensional tectonics and stratigraphy of the North Atlantic margins,* Vol. 46, pp. 247–263, eds Tankard, A.J. & Balkwill, H.R., AAPG Mem.

Tucholke, B.E., Lin, J. & Kleinrock, M.C., 1998. Megamullions and mullion structure defining oceanic metamorphic core complexes on the Mid-Atlantic Ridge, *J. geophys. Res.,* **103,** 9857–9866.

Tucholke, B.E., Sawyer, D.S. & Sibuet, J.-C., 2006. Breakup of the Newfoundland-Iberia rift, in *Extensional Deformation of the Lithosphere*, eds Karner, G., Manatschal, G. & Pinheiro, L., Columbia University Press, New York.

van der Velden, A.J., van Staal, C.R. & Cook, F.A., 2004. Crustal structure, fossil subduction, and the tectonic evolution of the Newfoundland Appalachians: evidence from a reprocessed seismic reflection survey, *Geol. Soc. Amer. Bull.,* **116,** 1485–1498.

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

Whitmarsh, R.B., Sawyer, D.S., Klaus, A., Masson, D.G. & Shipboard Scientific Party, 1996a. *Proceedings of the Ocean Drilling Program, Scientific Results,* Vol. 149, pp. 1–754, College Station, Texas, Ocean Drilling Program.

Whitmarsh, R.B., White, R.S., Horsefield, S.J., Sibuet, J.-C., Recq, M. & Louvel, 1996b, The ocean-continent boundary off the western continental margin of Iberia: crustal structure west of Galicia Bank, *J. geophys. Res.,* **101,** 28 291–28 314.

Whitmarsh, R.B., Beslier, M.-O., Wallace, P.J. & Shipboard Scientific Party, 1998. *Proceedings of the Ocean Drilling Program, Initial Results*, Vol. 173, pp. 1–294, College Station, Texas, Ocean Drilling Program.

Whitmarsh, R.B., Manatschal, G. & Minshull, T.A., 2001. Evolution of magma-poor continental margins from rifting to seafloor spreading, *Nature,* **413,** 150–154.

Wilson, R.C.L., 1988. Mesozoic development of the Lusitanian Basin, Portugal, *Revista de la Sociedad Geológica de España,* **1,** 393–407.

Wu, Y., Louden, K.E., Funck, T., Jackson, H.R. & Dehler, S.A., 2006. Crustal structure of the central Nova Scotia margin off Eastern Canada, *Geophys. J. Int.,* in press.

Yilmaz, O., 1987. *Seismic Data Processing*, Society of Exploration Geophysicists, Tulsa, Oklahoma.