PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

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THE ROYAL SOCIETY

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The deep structure of non-volcanic rifted continental margins

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Delineating the nature of crustal variations across passive continental margins is fundamental to our understanding of how the uppermost lithosphere deforms under extension. This information is best obtained across margins where the extensional fabric within the crust has not been significantly modified by large-scale syn-rift volcanism. Previous studies of such margins have shown a range of extensional styles, which have been interpreted by various combinations of pure and simple shearing of continental crust and lithosphere. This paper reanalyses several key transects in the North Atlantic, including the conjugate margin pairs of Newfoundland–Iberia, Flemish Cap–Goban Spur and Labrador–W. Greenland. Analysis of both deep multichannel seismic reflection and wide-angle reflection/refraction profiles allows us to produce new joint constructions of velocity and reflectivity depth sections. These sections indicate major asymmetries in the width and faulting style of continental rifting between margin conjugates. They also show a general occurrence of a complex transition zone between extended continental and oceanic crust that is dominated by exposed and serpentinized upper mantle. Because serpentinite has a much lower strength than is typical of continental or oceanic crustal rocks, its existence within the transition zone may have a profound influence on how these margins develop. The possible effects that such time and spatially variable rheologies have on the development of these margins needs to be considered in future geodynamical models.

Keywords: seismic reflection; seismic refraction; ocean–continent transition; crustal structure; serpentinized peridotite

1. Background

During the past 20 years, two archetypal models have dominated our concepts of lithospheric extension across continental margins: (1) the pure-shear model of McKenzie (1978) and its numerous variants (e.g. Royden & Keen 1980); and (2) the simple-shear model of Wernicke (1985) and its application to continental margins (Lister *et al*. 1986). In pure-shear models, extension is produced by a series of faults in the brittle upper layer (primarily upper crust) and ductile deformation in the lower layer (lower crust and mantle), resulting in a symmetric pattern of crustal and lithospheric thinning across pairs of conjugate margins. In simple-shear models, extension occurs along low-angle detachment faults, which penetrate through the crust and possibly into the lithosphere and offset the thinning of the upper layer from that of the lower layer. This results in an asymmetric pair of upper-plate (hanging wall) and lower plate (footwall) margins.

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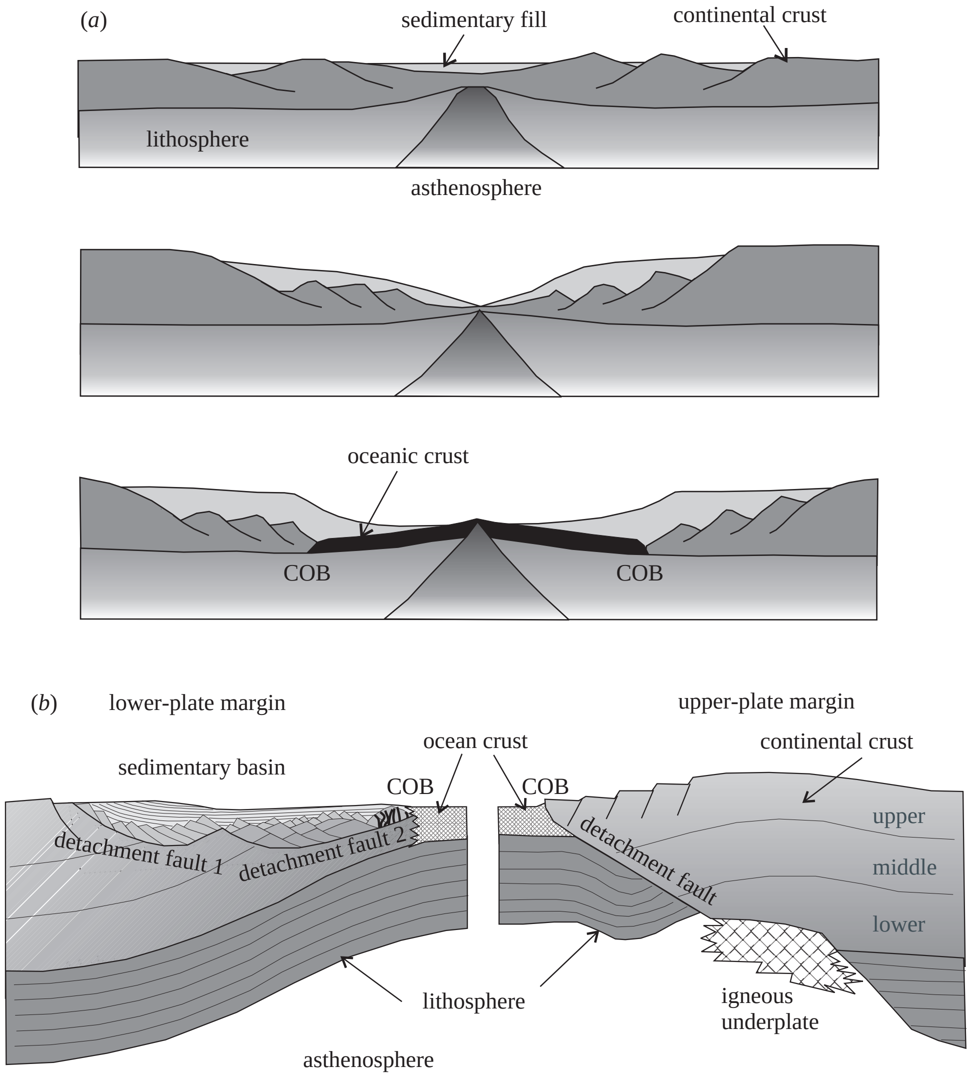

Figure 1. Conceptual models for pure and simple shear, which show predicted structures in the rifted continental crust: symmetric about the locus of asthenospheric upwelling for pure shear, and asymmetric about the crustal or lithospheric detachment fault for simple shear. In both cases, the contact between extended continental crust and oceanic crust (COB) is abrupt. After Open University (1998) and Lister *et al*. (1986).

Cartoon sketches of these two models (figure 1†) show predicted patterns of symmetric (pure-shear) and asymmetric (simple shear) extension within the continental crust. A number of deep seismic reflection profiles have been interpreted in terms of these predicted basement structures: a symmetric series of seaward-dipping rotated fault blocks indicative of pure-shear and flat-lying or landward-dipping detachment faults indicative of simple shear. The width of the zone of extended continental crust

† Full-size versions of all figures are available from K. Louden.

over which these extensional fabrics occur is typically 50–150 km, although sometimes as large as 400–500 km (e.g. Orphan Basin (Keen *et al*. 1987)). Both types of basement-fault geometries have been observed, primarily on margins where the overlying sediment layers are thin and not greatly reflective, which facilitates the seismic imaging of the underlying basement.

The idealized 'contact' between the thinned continental crust and the formation of oceanic crust, generally referred to as the continent–ocean boundary (COB), has typically been proposed in these models as an abrupt transition (figure 1). Numerous earlier attempts have been made to determine the nature and location of the COB primarily from the character of basement reflections (e.g. Keen & de Voogd 1988), but it has often remained an elusive target. More recently, seismic profiles and deep boreholes have indicated that this contact is not necessarily abrupt, but is characterized instead by a wide zone between rifted continental crust and plutonic oceanic crust, referred to as the ocean–continent transition (OCT). This zone has a rather uncertain crustal affinity, without a clear explanation in the above-mentioned idealized models, though it is generally thought to have formed during prolonged periods of extension when little or no melt was generated from the upper mantle. Several suggested possibilities for the origin of the basement in the OCT are (a) thinned continental crust intruded by melt from the mantle (e.g. Whitmarsh & Miles 1995); (b) thin and tectonized oceanic crust produced by ultra-slow sea-floor spreading (e.g. Srivastava & Roest 1995); and (c) serpentinized upper mantle (e.g. Boillot *et al*. 1987*a*).

It is the purpose of this paper to analyse characteristic images of crustal structure across continental margins in order to critically assess the nature of its extension. In particular, we want to consider not only the patterns of extension within the continental crust but also the nature of the transition between thinned continental and oceanic crust. We will use results from joint analyses of faulting patterns observed on deep seismic reflection profiles and of crustal velocity models determined from wide-angle reflection/refraction profiles. Profiles from conjugate transects will be used to look for predicted patterns of symmetry or asymmetry.

2. Crustal transects

In order to study crustal structures created by processes of continental rifting, we need to avoid margins where the extensional fabric may have been masked by large volumes of syn- or post-rift volcanism. This is a severe constraint because evidence for rift-related volcanism (i.e. seaward dipping reflectors associated with volcanic extrusives overlying an underplated zone of lower crust with high velocity) is increasingly evident from a majority of margins, even those (e.g. US East Coast) not necessarily associated with surface expressions of a mantle plume (Holbrook & Kelemen 1993). We also need to locate regions where deep-imaging crustal transects exist on both sides of respective conjugate margin pairs. The North Atlantic offers one of the few regions where such conditions are met. The plate reconstructions shown in figure 2 (Coffin *et al*. 1992) indicate a period of continental rifting between North America and Europe–Greenland from approximately 130 to 60 Ma, which is not associated with major amounts of volcanism. This period post-dates a previous episode of volcanic rifting beginning at *ca*. 180 Ma between North America, and Africa to the south;

and it pre-dates a later period of volcanic rifting associated with the Iceland plume beginning at *ca.* 60 Ma between Greenland and Europe to the northeast.

Rifting in this part of the North Atlantic may have started as early as the Late Triassic to Early Jurassic, as evidenced by rift sag successions encountered in marginal basins (e.g. Hiscott *et al*. 1990). Rifting continued in the Late Jurassic to Early Cretaceous, as evidenced by coast parallel dike swarms in SW Greenland (Watt 1969; Larsen *et al*. 1999) and shallow water carbonates on the Biscay and Galicia Bank margins (e.g. Moullade *et al*. 1988). During the Cretaceous period, there were three final episodes of continental rifting (figure 2), progressing from south to north. These episodes resulted in separations of (1) the Southern Grand Banks and Iberia; (2) Flemish Cap and Goban Spur; and (3) Labrador and SW Greenland. Additional separation between the Biscay and N. Iberia margins during episode 2 is less useful for purposes of conjugate reconstructions due to subsequent partial subduction of the North Iberia margin.

In this paper, we will constrain the amount and style of extension across the three pairs of non-volcanic conjugate margins shown in figure 3. We will use deep seismic reflection profiles to image the tectonic fabric (i.e. faults) and crustal velocity models, determined from wide-angle reflection/refraction profiles, to define variations in crustal thickness and composition. When possible, we will combine the reflectivity and velocity models to produce seismic depth sections. These displays will image the true crustal geometry better than is possible from time sections alone, due to the large offsets that changes in the water and sediment layer thicknesses and velocities across the margin have on underlying crustal vertical incidence travel-times.

(*a*) *Goban Spur–Flemish Cap*

The Goban Spur–Flemish Cap conjugate margins were formed during the Early Cretaceous by separation of the European and North American continents. Immediately to the south, a triple junction (figure 2) was active until Chron 33 (80 Ma)†, during rotation of the Iberian peninsula to form the Bay of Biscay (Sibuet & Collette 1991). The absence of magnetic chron M0 in the offshore oceanic crust (Verhoef *et al*. 1986) and the dating of syn-rift sedimentary sequences from drillholes on the Goban Spur and Biscay margins (Montadert *et al*. 1979; Roberts *et al*. 1981; Masson *et al*. 1985) indicate the beginning of rifting in the early Barremian (*ca.* 127 Ma) and the start of oceanic crustal accretion in the late Albian (*ca.* 110 Ma). Using these dates, the duration of rifting is 15–20 Ma.

Figure 4 shows the location of Goban Spur. It is a relatively wide and shallow margin compared with the Biscay margins to the southeast, with an extensive region less than 2000 m deep (shaded area). Locations are given of the BIRPS WAM reflection profile (Peddy *et al*. 1989) and a coincident crustal refraction profile (Horsefield *et al*. 1993). These profiles have previously been interpreted as evidence for pure-shear rifting, in which the upper crust thins by fault-block rotation and the lower crust thins by ductile extension in the same proportions. A sharp contact with oceanic crust is thought to exist at the base of the seawardmost tilted fault block, as evidenced by recovery of tholeiitic basalt typical of oceanic crust from DSDP Sites 551 and 550 (de Graciansky *et al*. 1984).

† The geological time-scale used for this and subsequent dates is that of Gradstein *et al*. (1994).

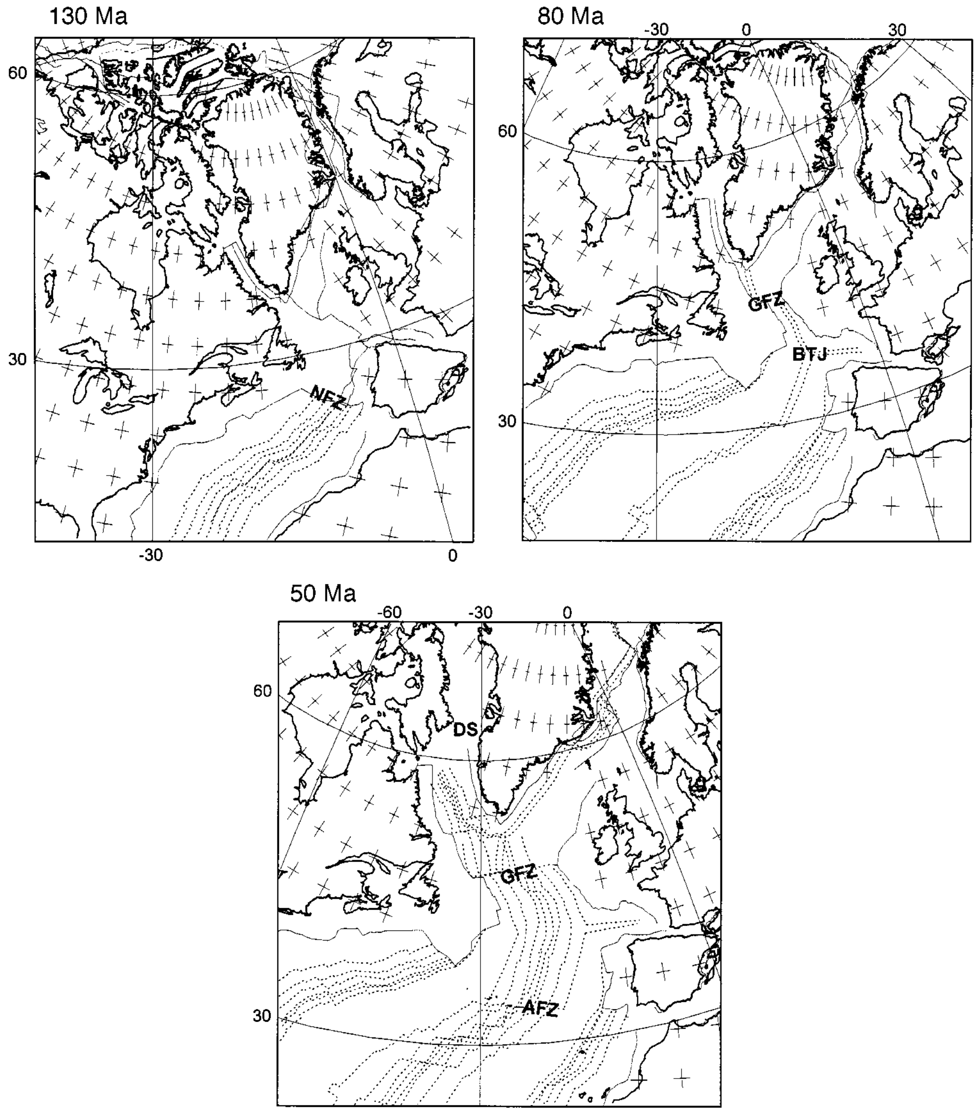

Figure 2. Plate reconstructions for the North Atlantic at 130, 80 and 50 Ma showing the break-up and separation of Europe and North America. Selected magnetic isochrons are given by broken lines. NFZ, Newfoundland Fracture Zone; GFZ, Charlie–Gibbs Fracture Zone; BTZ, Biscay Triple Junction; AFZ, Azores Fracture Zone; DS, Davis Strait. From Coffin *et al*. (1992).

In figure 5, we use the velocity model (shown in colour) determined from the refraction profiles to depth migrate the seismic reflection data (superimposed in black). This depth section reveals some interesting features not apparent on the original time section. The reflective lower crust is clearly imaged within the thicker continental crust (−10 to 50 km distance), the base of which corresponds reasonably well to Moho depths in the velocity model (R1). At about 60 km, the lower crustal reflectivity appears to terminate abruptly at a prominent horizontal reflection (R2),

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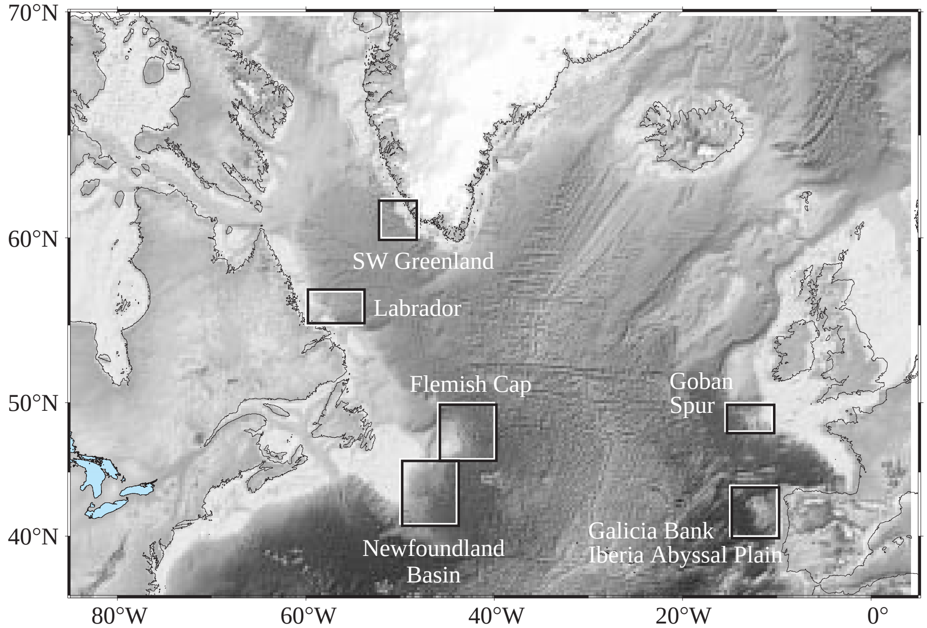

Figure 3. Location of the three conjugate margin study areas: Goban Spur–Flemish Cap; SW Greenland–Labrador; and Galicia Bank/Southern Iberia Abyssal Plain–Newfoundland Basin. Shaded bathymetry from Smith & Sandwell (1997).

which appears to cut across the simple shape of the Moho in the wide-angle velocity model. This suggests that the Moho may have a more complex shape in this region beneath the basement fault blocks. Note, however, that potential interference of crustal arrivals with water multiples across the continental slope complicates the interpretation of changes in reflectivity within the crust over this region (30–70 km distance).

The three large-scale upper crustal fault blocks (B1) between distances of 30–90 km are associated with a Moho thinning from 25 to 12 km in the velocity model. Seaward of 90 km there is no clear Moho reflection, but a number of dipping reflections, though weak, are observed within the crust (R3) and mantle (R4). Other reflections that dip more gently landward are observed to depths of 15–18 km within the mantle (R5, R6). The basement is initially flat lying from 90 to 150 km distances (B2) and then is marked by a series of highs from 150 to 215 km (B3). A reflection from oceanic Moho does not appear until a distance of *ca.* 230 km (R7) and it is shallower (at *ca.* 5 km below the top of basement) than predicted by the velocity model; though in fact, the velocity model is only constrained between distances of 30–140 km, since there are no receivers or crossing profiles seaward of 95 km. A sharp Moho in the oceanic crust as proposed by Horsefield *et al.* (1993) is also not consistent with the lack of wide angle PmP reflections observed on the surface at this station between distances of 100–150 km. Thus, the thinning of upper and lower continental crust across Goban Spur may not be as uniform as previously proposed. There also exists a 120 km wide zone between the seaward-most continental fault block and the first appearance of a reflection from oceanic Moho that has a different pattern of reflec-

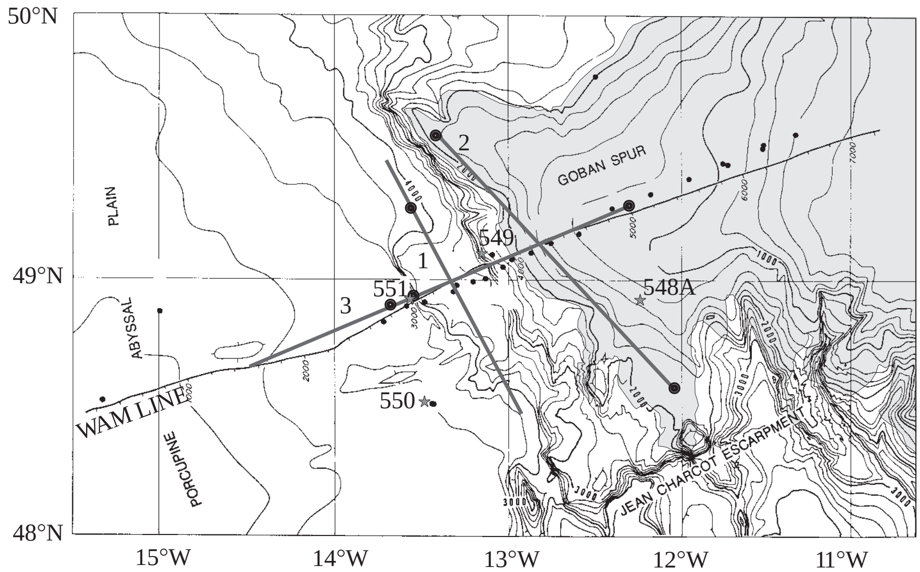

Figure 4. Bathymetry of Goban Spur (shaded for depths less than 2000 m) (Sibuet *et al*. 1985) and locations of the WAM deep reflection profile (thin line with shot numbers) (Peddy *et al*. 1989); coincident and crossing refraction profiles 1, 2 and 3 (thick lines) (Horsefield *et al*. 1993); and DSDP leg 80 borehole sites (stars) (de Graciansky *et al*. 1984). Location of refraction receivers shown by larger filled circles and heat flow stations (Louden *et al*. 1991) by small filled circles.

tions, though existing wide-angle seismic data are not sufficiently detailed in this region to characterize its crustal affinity.

We can compare these features with profile 85-3 (Keen & de Voogd 1988), which is located across the conjugate margin of Flemish Cap (figures 6 and 7). Unfortunately we are not able to convert this time section to depth because of the lack of wide-angle data except over a limited offshore region (100–150 km) (Reid & Keen 1990). Therefore, in figure 7 both the unmigrated seismic data and velocity model are plotted in two-way travel times. Clear Moho reflections are observed from *ca.* 11 s beneath Flemish Cap at distances of 25–50 km (R1). Seaward of this and in contrast to Goban Spur, there is no clear connection between tilted fault blocks and crustal thinning. A prominent fault block is observed at 65–70 km distance (B1) and is underlain by a horizontal reflection (R2) that may terminate at the top of basement (75 km distance). Within the area of the refraction profile, a number of both landward (R3) and seaward (R4) dipping events appear to border a layer of high-velocity (7.4 km s^{-1}) lower crust (HVLC). A possible Moho reflection (R5) appears at *ca.* 140 km distance consistent with the velocity model. The reflection from the top of basement (B2), associated with a velocity transition of 3.0–4.5 km s^{-1} in the refraction model, is flat and deepest over distances of 80–150 km, seaward of which lies a rougher elevated basement (B3) over distances of 160–185 km.

Therefore, in a manner quite similar to Goban Spur, a transitional region 70–100 km wide occurs between the last continental fault block and the first appearance of a reflection from oceanic Moho, with a similar pattern of basement morphology

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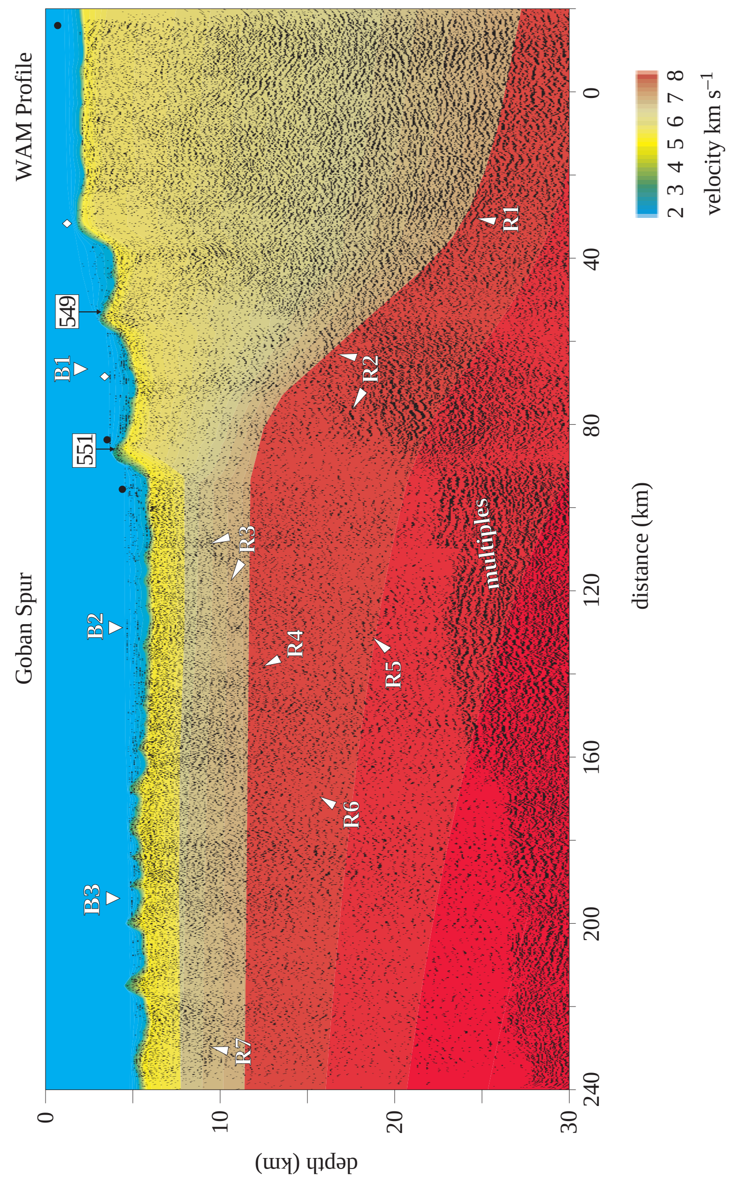

Figure 5. Seismic depth section across the Goban Spur margin, showing reflectivity along the WAM profile (Peddy *et al*. 1989) that is Kirchhoff depth-migrated using the velocity model of Horsefield *et al*. (1993) (shown in colour). Locations are given for ODP borehole sites (numbered boxes), refraction receivers (filled circles), and crossing refraction profiles (white diamonds). Note lack of velocity control and simple one-dimensional model seaward of *ca*. 120 km distance. Basement types (B1–B3) and reflections (R1–R7) are discussed in the text.

and crustal reflectivity. However, in contrast to the similarities in the transitional crust, the manner of continental faulting and crustal thinning appears quite dissimilar between the two margin conjugates, with a much broader region across Goban

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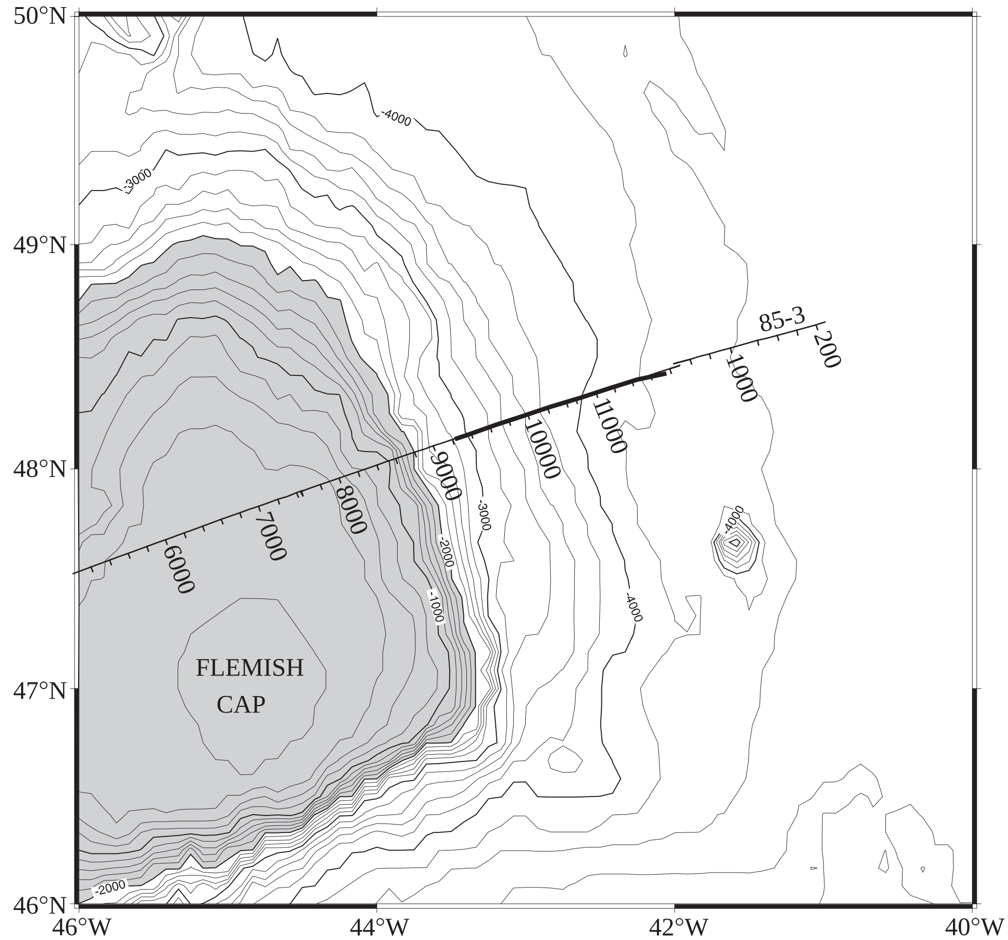

Figure 6. Bathymetry of Flemish Cap (ETOP05) and location of deep reflection line 85-3 (thin line with shot numbers) (Keen & de Voogd 1988) and refraction profile (thick line) (Reid & Keen 1990). Water depths shallower than 2000 m are shaded.

Spur compared to Flemish Cap. This was noted by Keen *et al*. (1989) in their initial reconstructions of the two time sections. They suggested, however, that this asymmetry might have occurred if final break-up was offset towards Flemish Cap. In this case, initial thinning of the continental crust would have occurred as pure shear and final separation as either pure or simple shear. When we include the additional interpretation of the velocity models, our interpretations suggest that the continental crust of Flemish Cap was thinned over a much shorter horizontal distance (20–30 km) than Goban Spur (60–80 km). Another asymmetry between the margins is indicated by the presence of the sub-horizontal mid-crustal reflection (R2) beneath the tilted fault block off Flemish Cap, which is not evident off Goban Spur. These asymmetries in addition to the asymmetry between upper and lower crustal thinning previously noted indicate that evidence for pure shear extension during final break-up is not as clear as previously suggested.

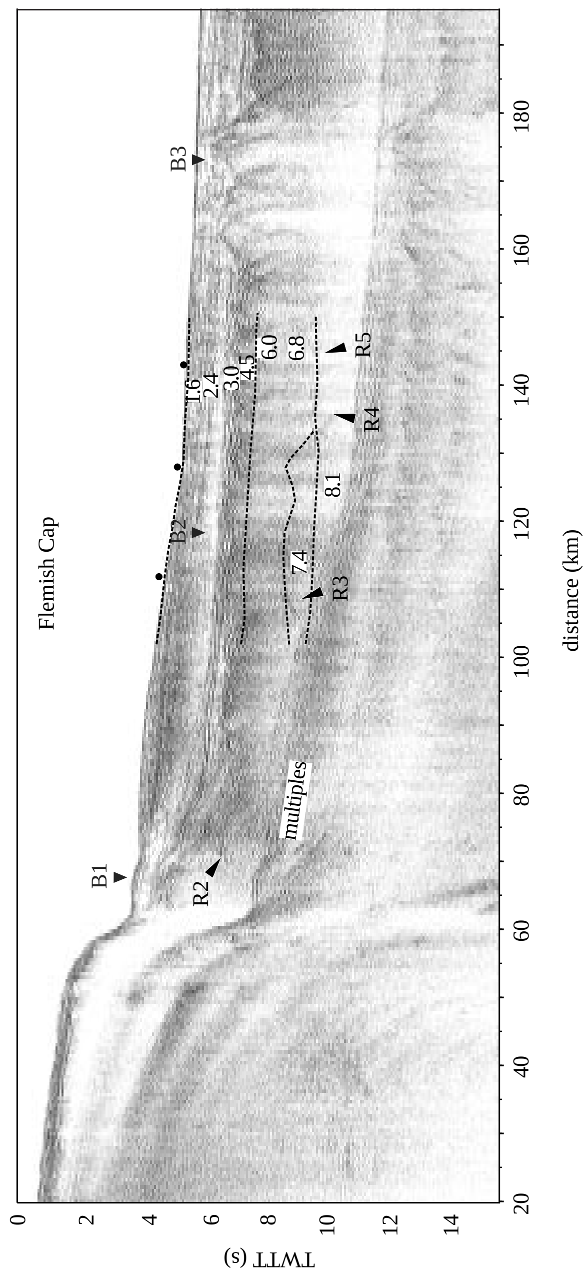

Figure 7. Time migrated reflection profile 85-3 (Keen & de Voogd 1988) and velocity model (Reid & Keen 1990) across the Flemish Cap margin. Location of refraction receivers (small filled circles), selected velocity layer boundaries (dotted lines) and layer velocities (in km s^{-1}) of the refraction model as shown. Reflections R1–R5 and basement types B1–B3 are discussed in the text.

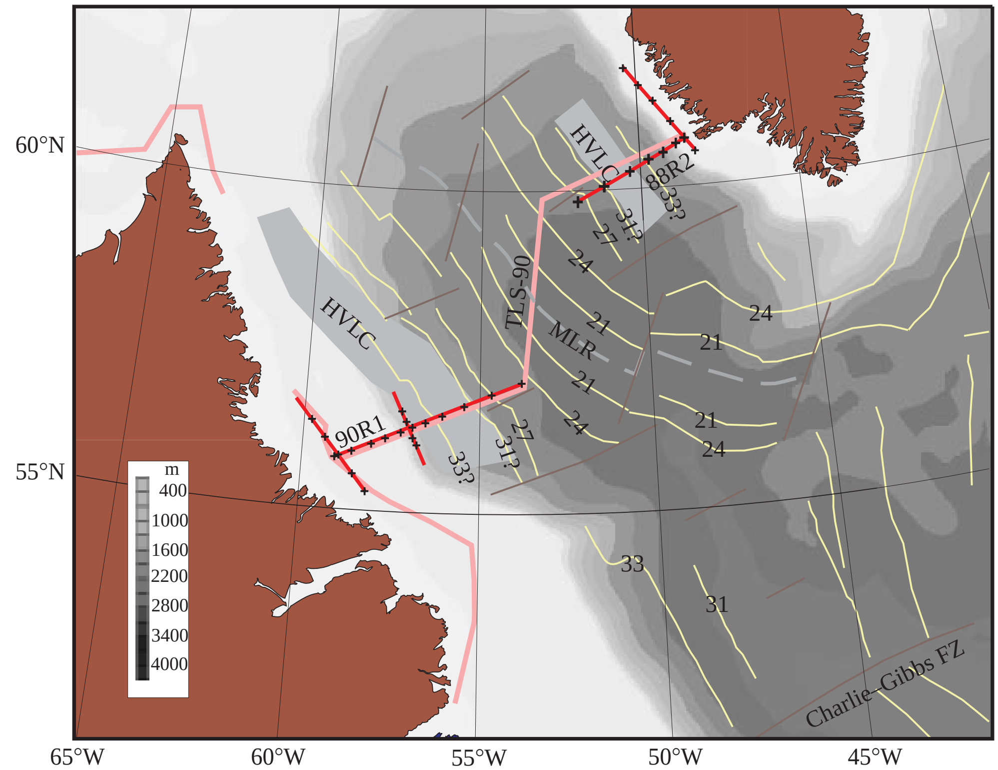

Figure 8. Bathymetry of Labrador Sea (ETOP05) and location of deep seismic reflection profile TLS90 (Keen *et al*. 1994) and refraction profiles 88R2 (Chian & Louden 1994) and 90R1 (Chian *et al*. 1995*b*). Magnetic anomaly lineations (white lines) and identifications in numbers from Roest & Srivastava (1989). Location of high velocity lower crustal layers (HVLC) from Chian *et al*. (1995*a*). MLR, Mid-Labrador rift.

(*b*) *Western Greenland–Labrador*

The Labrador Sea is a northwestward extension of the North Atlantic Ocean, from the Charlie–Gibbs fracture zone in the south to Davis Strait in the north, that separates southern Greenland from Labrador (figure 8). The continental shelf and slope south of *ca.* 67° N are much narrower on the Western Greenland margin than on the Labrador margin. Since the Cretaceous, a large volume of clastic sediments has been deposited into the Labrador Sea (Tucholke 1988), covering its basement morphology. The crust is deeply subsided under the Labrador shelf, partly due to sediment loading, whereas little or no subsidence has occurred under the Western Greenland shelf south of 63° N (Rolle 1985), except within some restricted half-grabens (Chalmers *et al*. 1993). Final rifting and break-up of these margins began during the Barremian (*ca.* 127 Ma) and ended in the Coniacian (*ca.* 88 Ma) (Balkwill 1987) or Palaeocene (Chalmers 1997). As Davis Strait is approached, the amount of syn-rift volcanism increases as evidenced by extrusives from Disko Island (Gill *et al*. 1992), seaward dipping reflectors on reflection seismic line BGR 77-6 (Chalmers *et al*. 1993; Chalmers 1997) and intrusives indicated by seismic velocity models beneath

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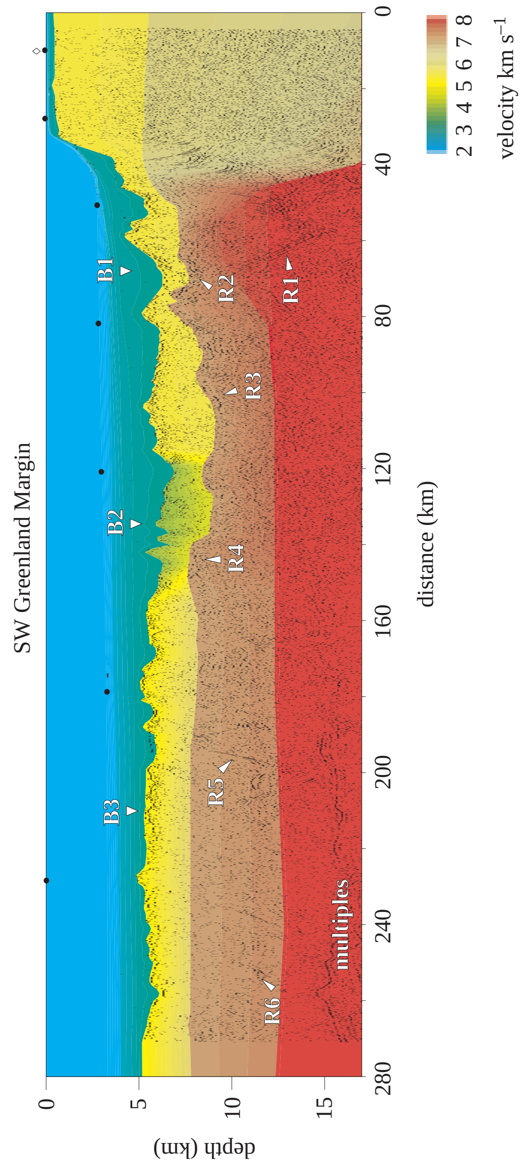

Figure 9. Seismic depth section across the SW Greenland margin, showing reflectivity along the time migrated TLS90-3 profile (Keen *et al*. 1994), which is converted to depth using the velocity model of Chian & Louden (1994) (shown in colour). Locations are given of refraction receivers (filled circles), and crossing refraction profile (white diamond). Note lack of velocity control and simple one-dimensional model seaward of *ca.* 240 km distance. Basement types (B1–B3) and reflections (R1–R6) are discussed in text.

the Greenland shelf at 64° N (Gohl & Smithson 1993). The subsequent sea-floor spreading history is documented by magnetic lineations, with well-defined anomalies observed between Chrons 21 and 27 (49–63 Ma) (figure 8) (Roest & Srivastava 1989).

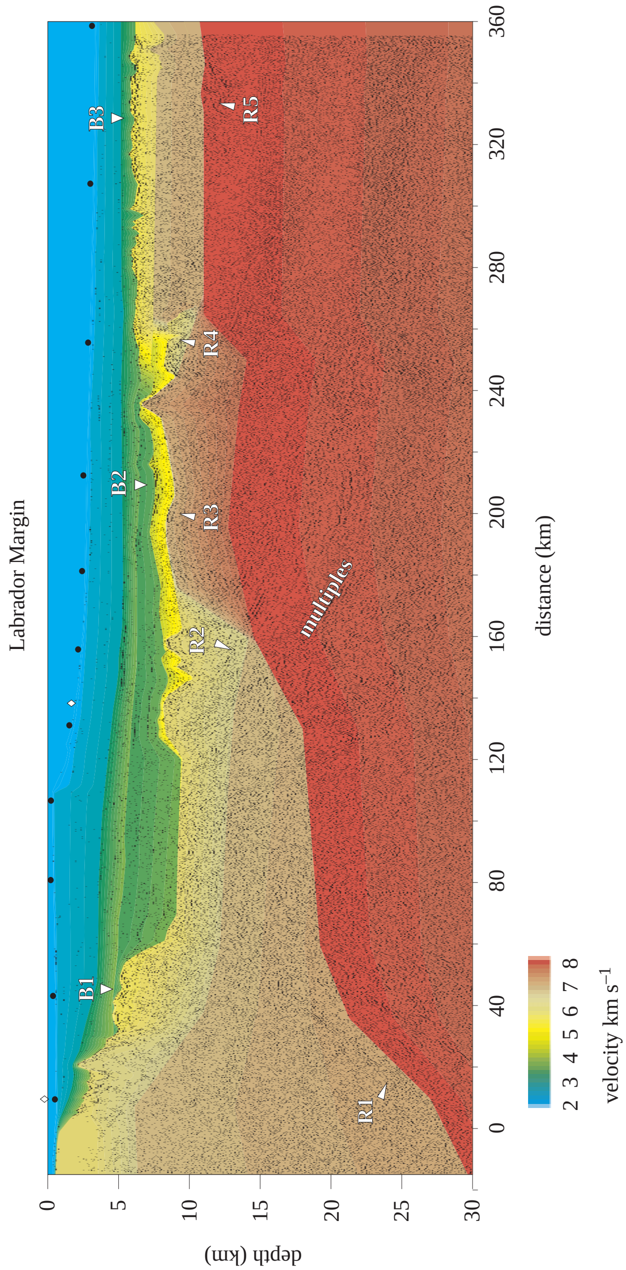

Figure 10. Seismic depth section across the Labrador margin, showing reflectivity along the time migrated TLS90-1 profile (Keen *et al.* 1994), which is converted to depth using the velocity model of Chian *et al.* (1995*b*) (shown in colour). Locations are given of refraction receivers (filled circles), and crossing refraction profiles (white diamonds). Note good velocity control along entire profile. Basement types (B1–B3) and reflections (R1–R5) are discussed in text.

The less clearly defined, low-amplitude anomalies older than Chron 27 have been interpreted by Roest & Srivastava (1989) and Srivastava & Roest (1995) as Chrons 31–33 (68–80 Ma). Alternatively, it has also been suggested that true oceanic crust did not form until either Chron 27 (63 Ma) (Chalmers 1991; Chalmers & Laursen 1995) or some time between Chrons 27–31 (63–68 Ma) (Chian & Louden 1994; Chian *et al*. 1995*a*, *b*). Using the full range suggested by these various dates, the period of rifting lasted about 40–65 Ma.

Figures 9 and 10 show seismic depth sections across the W. Greenland and Labrador margins. These images are produced from conversion of the time-migrated reflection profiles TLS90-1/3 (Keen *et al*. 1994) into depth, using the seismic velocity structure superimposed in colour as derived from the proximate wide-angle profile 88-R2 for W. Greenland (Chian & Louden 1994) and coincident profile 90-R1 for Labrador (Chian *et al*. 1995*b*).

The Greenland profile TLS 90-3 (figure 9) shows an abrupt crustal thinning across the narrow continental slope with weak evidence for some landward dipping reflections within the mantle (R1). As previously described for travel-time sections (Keen *et al*. 1994; Chian *et al*. 1995*a*), there are three zones defined by variations in basement morphology, which closely relate to changes in crustal velocity structure. Zone 1 (40–100 km distance) lies beneath the tilted fault blocks (B1), beneath which lies a well-developed horizontal mid-crustal reflection (R2–R3). This reflection is coincident with an abrupt velocity transition between upper crustal and HVLC layers. Zone 2 (110–160 km distance) is indicated by the seaward termination of large-scale basement fault blocks into a more fragmented boundary (B2), in which upper basement velocities are low (4.4–4.6 km s^{-1}) and lower crustal velocities are high (7.2–7.6 km s^{-1}). The mid-crustal reflection (R4) terminates towards the seaward end of this zone (*ca*. 150 km distance). Further seaward, basement becomes elevated and is characterized by a more coherent reflection boundary (B3) within Zone 3 (distances of more than 200 km). A prominent seaward dipping crustal reflection (R5) marks the landward edge of this zone (195 km distance) and the first appearance of possible Moho reflections. Lower crustal velocities reduce to 7.0–7.2 km s^{-1}. The Moho boundary in the velocity model is somewhat deeper than suggested by the reflection (R6), though the velocity model in this region is not as well constrained by the single-ended sonobuoy data due to failure of the seaward-most OBS receivers.

The Labrador profile (figure 10) shows a much broader zone of crustal thinning across the continental shelf (−15 to 160 km distance), with its very thick sedimentary basin (maximum sediment thickness of *ca*. 9 km). Most of the thinning occurs in the lower crust (V_p = 6.6–6.9 km s^{-1}), where a rising Moho in the velocity model corresponds to a series of deep reflections (R1) that span a range of depths within the lower crust and upper mantle (Chian *et al*. 1995*b*; Louden & Fan 1998). Faulted upper continental crust appears to terminate at a single large fault block (B1) at a distance of 40–60 km. The character of basement and crust is not well-imaged beneath the thick sedimentary basin, possibly due to the lack of strong velocity contrasts. Further seaward, an HVLC zone (180–240 km distances) with a smooth basement surface (B2) and prominent mid-crustal reflection (R3) is bounded by a series of seaward dipping reflections (R2) in the crust and mantle on its landward side. On its seaward side, the top of basement is shallower and the mid-crustal reflection (R4) terminates (at *ca*. 260 km distance). Further seaward, oceanic basement (B3) and reflections from the Moho (R5) are finally observed at 300–340 km distance. In

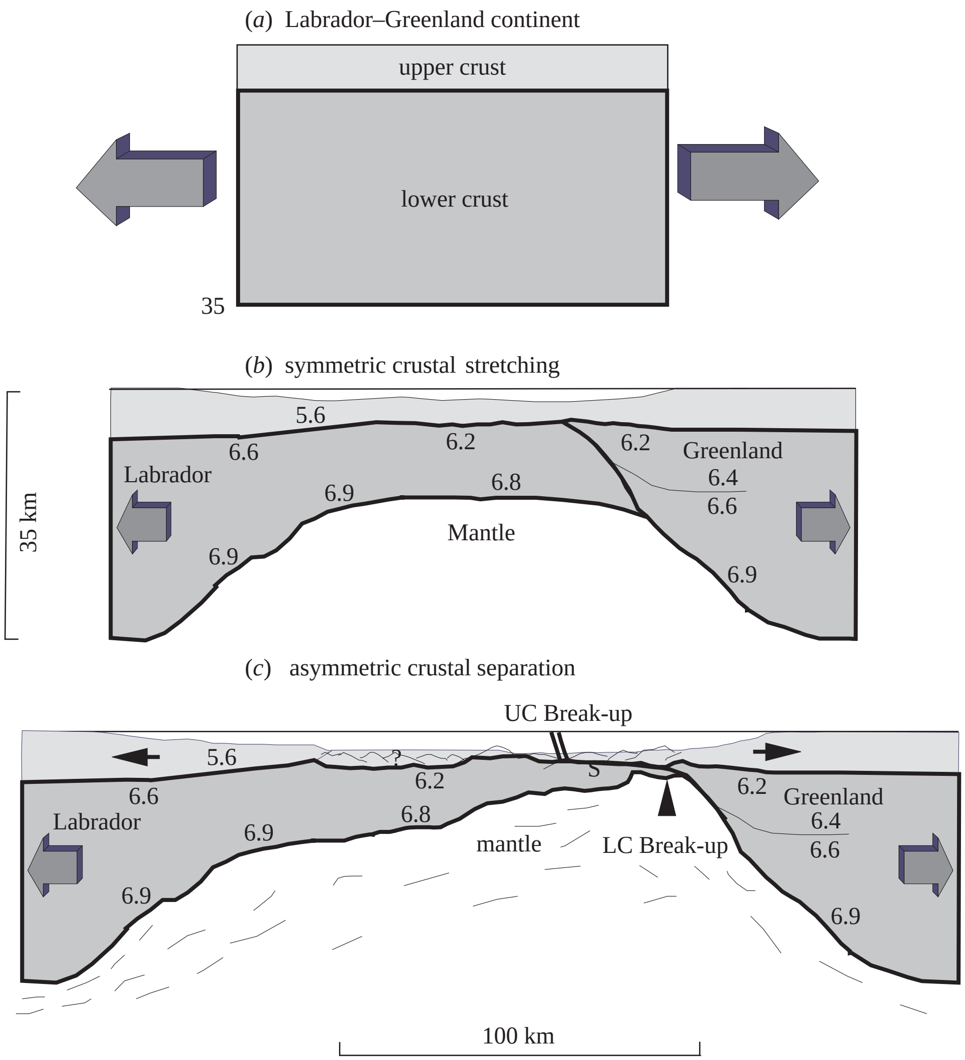

Figure 11. Possible scenario for continental rifting based on balanced cross-sections, using results of refraction models across the conjugate Labrador and SW Greenland margins (velocities given in km s^{-1}) (Chian *et al*. 1995*a*) and assuming that lower crust is allowed to move plastically beneath upper crust. (*a*) Initial configuration of Labrador–Greenland block. (*b*) Crustal thinning starts as symmetric pure shear until stretching factor $\beta \approx 2$. (*c*) Locus of stretching migrates toward the Greenland margin at which point break-up occurs with offset between lower crust (LC) and upper crust (UC).

this case, the well-controlled velocity model indicates a thin oceanic crust (*ca.* 5 km thick) in agreement with the Moho reflection here as well as on the seaward end of the Greenland profile (figure 9).

The fundamental asymmetry of the conjugate margins after removal of the similar oceanic and HVLC zones is shown in figure 11*c* (Chian *et al*. 1995*b*). Thinning of the lower crust is particularly asymmetric, with location of break-up offset towards

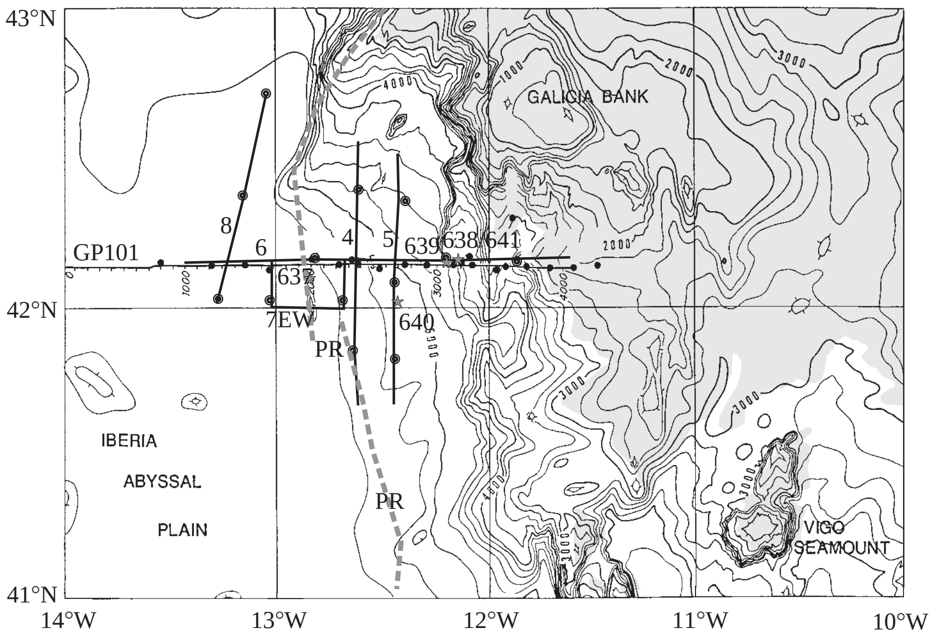

Figure 12. Bathymetry of Galicia Bank (shaded for depths less than 3000 m) (Sibuet *et al*. 1987) and locations of the GP101 deep reflection profile (thin line with shot numbers) (Mauffret & Montadert 1987), coincident and crossing refraction profiles (thick lines labelled 4, 5, 6, 7EW and 8) (Whitmarsh *et al*. 1996*b*), and ODP leg 103 borehole sites (stars) (Boillot *et al*. 1987*b*). Location of refraction receivers shown by larger bulls eyes and heat flow stations (Louden *et al*. 1991) by small filled circles. Thick dashed lines labelled PR are locations of the peridotite ridges (Beslier *et al*. 1993).

the Greenland margin and a broad zone of extended lower crust beneath the outer Labrador margin. It is clear that there is some discontinuity between the location of lower and upper crustal break-up, though the nature of the upper crust on the seaward sections of both the Greenland (90–120 km distance in figure 9) and Labrador (60–160 km distance in figure 10) Margins is uncertain. In figure 11, we close the two margins further by arbitrarily allowing the lower crust to deform plastically beneath Greenland. This scenario maintains balanced continental crustal sections, but does not include potential shortening of the upper crust by movement on small-scale faults. This result indicates that the initial thinning of the Labrador–Greenland continent may have occurred symmetrically until a factor $\beta \sim 2$ (i.e. 50% thinning), at which point the locus of thinning migrated toward the Greenland margin where final separation occurred once the crust had been thinned by a β-factor of 7–8.

(*c*) *Galicia Bank/Iberia Abyssal Plain–Newfoundland Basin*

The margins west of the north-central Iberia peninsula and east of the southern Grand Banks of Newfoundland formed as a result of several stages of Mesozoic rifting in a segment south of the Biscay triple junction and north of the Newfoundland–Azores fracture zone (figure 2) (e.g. Pinheiro *et al*. 1996; Tucholke *et al*. 1989). Initial continental stretching started in the Late Triassic to Early Jurassic, continued in

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the late Oxfordian to early Kimmeridgian, and culminated in the early Valanginian (137 Ma) to late Aptian (112 Ma). The first unequivocal sea-floor spreading anomaly in the Newfoundland and Iberia Abyssal Plains is the J anomaly, which lies between magnetic chrons M0 and M2 (*ca.* 121 Ma) (Tucholke & Ludwig 1982; Whitmarsh & Miles 1995). However, Whitmarsh & Miles (1995) and Whitmarsh *et al.* (1996*a*) have suggested that sea-floor spreading in the southern Iberia Abyssal Plain began during Chron M3 (125 Ma) just west of a prominent ridge (PR) where serpentinized peridotite was sampled at ODP Site 897 (Sawyer *et al.* 1994). Plate reconstructions to these isochrons (Srivastava & Verhoef 1992) join Galicia Bank to Flemish Cap, but leave a deep water ocean–continent transition (OCT) zone over 200 km wide between Chron M0 and the base of the continental rises off Iberia and the Grand Banks. Off Galicia Bank and Flemish Cap, sea-floor spreading began somewhat later in the Aptian (*ca.* 115 Ma) as evidenced by the absence of Chron M0 (Verhoef *et al.* 1986). Thus, rifting lasted *ca.* 25 Ma off Galicia Bank and Flemish Cap and *ca.* 15 Ma in the southern Iberia Abyssal Plain and Newfoundland basin. Alternatively, Wilson *et al.* (1996) have suggested from the lack of syn-rift sequences on reflection profiles and a radiometric date of 136 Ma for metagabbros from ODP Site 900 (Féraud *et al.* 1996) that rifting may have lasted for a much shorter period (*ca.* 5 Ma) in the southern Iberia Abyssal Plain.

Figure 12 shows the location of Galicia Bank. Reflection profile GP101 (Mauffret & Montadert 1987) and a coincident refraction profile 6 (Whitmarsh *et al.* 1996*b*) across the margin are plotted separately in figure 13. Due to difficulties in obtaining the original reflection data, we were not able to produce depth sections of the reflection data using the refraction velocity model. Improved time and depth migrations have been reported for GP101 and adjacent profiles (Krawczyk & Reston 1995; Reston 1996; Reston *et al.* 1996) but only for rather limited sections of the profiles landward of the peridotite ridge (PR). Profile GP-101 and Line 6 show that the tilted fault blocks are associated with thinning of the continental crust from a thickness of 17 km beneath Galicia Bank (0 km distance) to *ca.* 5 km at 70 km distance. From 70 to 115 km distance, an HVLC layer ($V_{\mathrm{p}} = 7.2$–$7.6\ \mathrm{km\ s^{-1}}$) is associated with the occurrence of a mid-crustal 'S-reflector' (Whitmarsh *et al.* 1996*b*) and peridotite ridge (PR). The crustal velocity models, defined by the refraction profile and by pre-stack migration of the reflection profile, are consistent with a sharp velocity boundary at the top of the HVLC layer when it is coincident with 'S'. This argues against the earlier interpretations of 'S' as a boundary between the brittle and ductile continental crust (Mauffret & Montadert 1987), based on analysis of reflection profiles across this margin and across the Biscay margin to the north where it was first identified (de Charpal *et al.* 1978; Le Pichon & Barbier 1987). However, note in figure 13 that the 'S' reflector only coincides with the top boundary of the HVLC between shot point 2200 (90 km distance) and the peridotite ridge. East of shot point 2200, 'S' shows a complex geometry in the pre-stack migration and shallows to a depth of *ca.* 8 km, which is located in the mid-crust *ca.* 2 km above the HVLC in the refraction model. Further seaward, the boundary between the HVLC and oceanic crust is not well defined along line 6 due to the lack of seaward receivers, but is consistent with the crossing line 8, although Moho reflections are nowhere evident along GP-101.

To the south of Galicia Bank within the southern Iberia Abyssal Plain, a large amount of geophysical data has been collected across the OCT, including reflection,

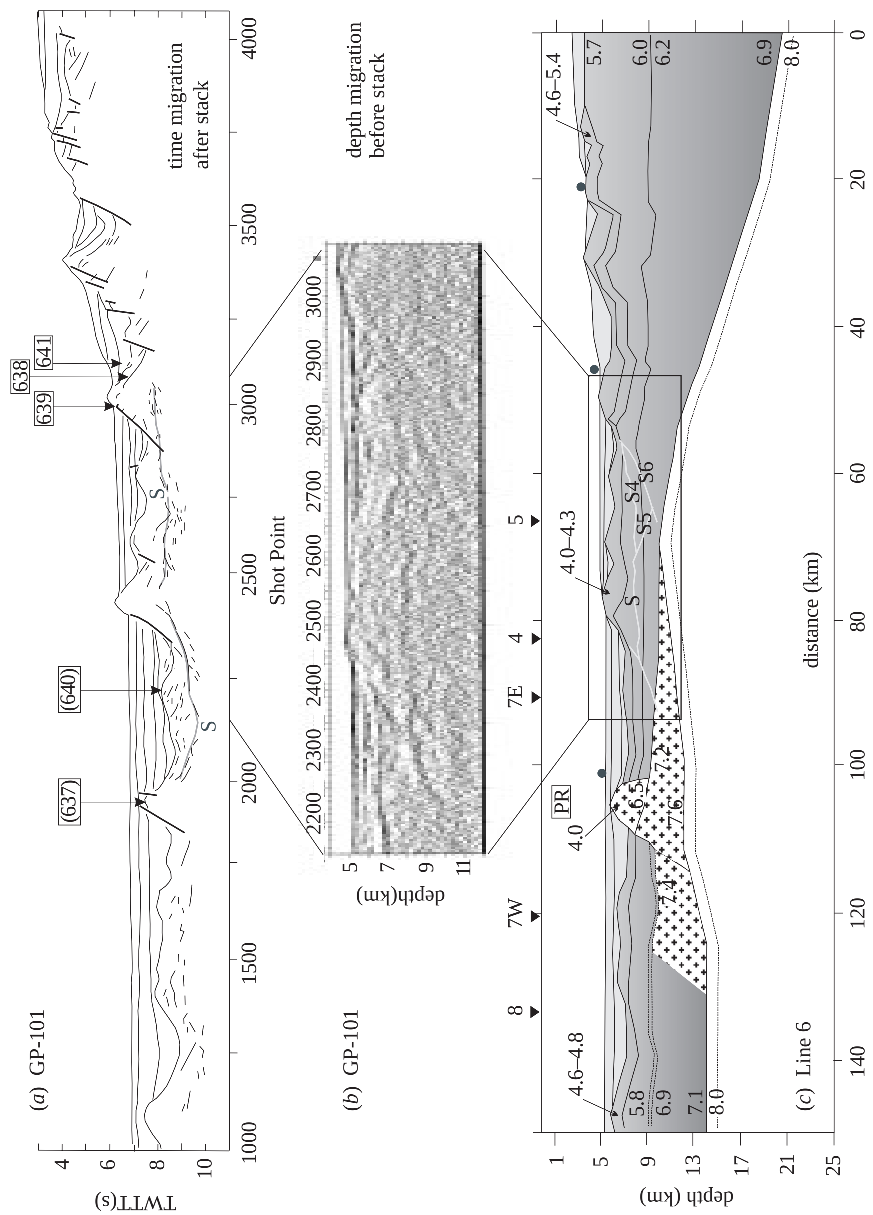

Figure 13. (*a*) Line drawing of reflection travel time profile GP101 (time migration) following interpretations of Mauffret & Montadert (1987) and Sibuet *et al*. (1995). Grey lines give location of 'S' reflector and boxes with arrows give locations of ODP borehole sites as identified (projected locations indicated by parentheses). (*b*) Depth-migrated section of GP-101 (Reston *et al*. 1996). (*c*) Velocity–depth model for line 6 (Whitmarsh *et al*. 1996*b*). Receivers locations (small filled circles), crossing profiles (triangles), layer boundaries (solid lines) and layer velocities (in km s^{-1}) are shown along the refraction model. Area highlighted with pattern indicates HVLC layer interpreted as a zone of serpentinized peridotite. Rectangle shows position from (*b*) of depth migrated section with 'S' reflector segments (grey lines).

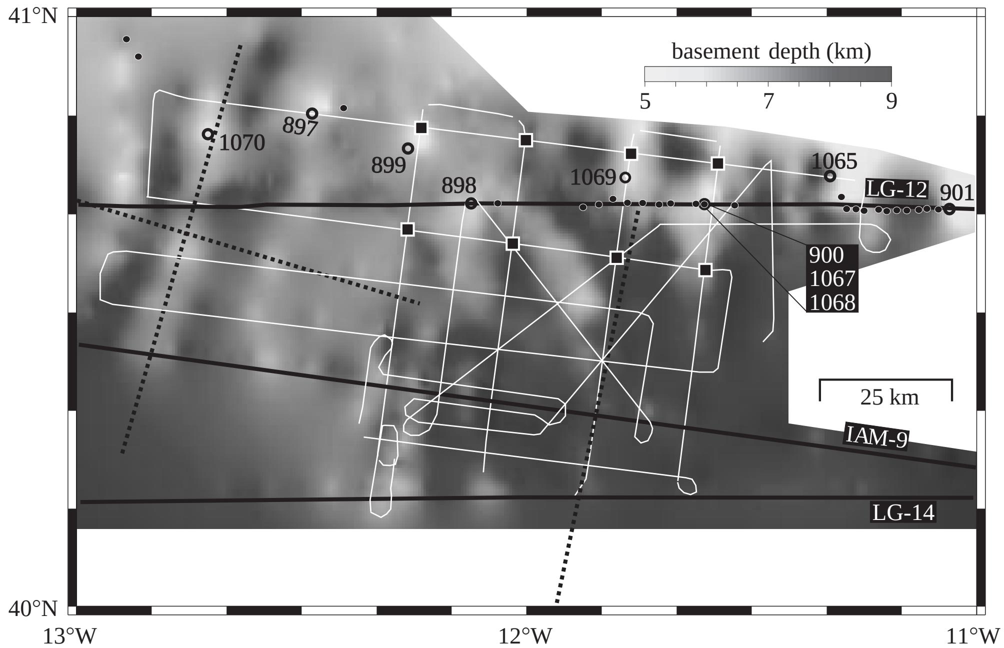

Figure 14. Detailed map of ODP drill sites (Sawyer *et al*. 1994; ODP Leg 173 Shipboard Scientific Party 1998) and seismic reflection and refraction lines in the southern Iberia abyssal plain. Thick black lines are deep reflection profiles LG-12, LG-14 and IAM-9. Thick dotted lines are previous refraction lines of Whitmarsh *et al*. (1990). The background image is depth to basement (in km; after Discovery 215 Working Group (1998)) and white lines are recent refraction/reflection profiles with eight northernmost receivers shown in filled black squares (Chian *et al*. 1999). Small black dots show locations of heat flow stations (Louden *et al*. 1997).

refraction and magnetic profiles along and adjacent to an E–W transect of ODP boreholes (Sawyer *et al*. 1994; ODP Leg 173 Shipboard Scientific Party 1998). Figure 14 shows the ODP drillsites and some of the reflection/refraction profiles located with respect to the basement topography (Discovery 215 Working Group 1998). In this paper, we analyse an older reflection profile LG-12 initially reported by Beslier *et al*. (1993), which we convert to depth using velocity models recently determined from a nearby set of crossing refraction profiles (Chian *et al*. 1999). A pre-stack depth migration of LG-12 between ODP Sites 901 and 898 has previously been presented by Krawczyk *et al*. (1996).

Features similar to those of GP101 are observed on LG-12, except that in this case the section of the profile shown does not extend onto the thicker Palaeozoic continental crust of the Iberian massif. From 5 to 70 km distance, a series of tilted basement fault blocks are observed (B1). Dipping reflections (R1–R2) associated with this basement topography appear to terminate at a sub-horizontal reflection (R3, referred to as the 'H-reflector' by Krawczyk *et al*. (1996), associated with a sharp transition from 6.6 to 7.3 km s^{-1} in the velocity model (Chian *et al*. 1999). Some reflections (R4–R5) penetrate this boundary and extend as deeply as 15–18 km (R6). The horizontal reflection R3 is missing beneath a basement high of gabbroic to serpentinized peridotite rocks sampled by ODP Sites 900, 1067 and 1068; but it may extend further west (R7) beneath continental basement sampled at Site 1069. An abrupt transition in basement morphology at 85 km distance is coincident with the seaward termination of the crustal horizontal reflection (R7) and deep cross-dipping lower crustal or upper mantle reflections (R8–R9). An abrupt westward reduction in crustal velocities of both upper and lower layers occurs at a similar position (Chian *et al*. 1999), east of the elevated basement where highly serpentinized peridotite was sampled at ODP Sites 898 and 897 (figure 14). This zone (B2) extends as far west as the basement ridge where Site 1070 sampled serpentinized peridotite and pegmatite gabbro. Whitmarsh *et al*. (1996*a*) consider this ridge to have an oceanic character based on magnetic modelling and a velocity model across the same structure to the south along IAM-9 has an oceanic structure (S. Dean, personal communication). Alternatively, the seismic velocity models and lack of clear Moho reflections on reflection profiles near Site 1070 (not shown) suggest that oceanic basement might not occur until somewhat further seaward towards the J-anomaly ridge (Chian *et al*. 1999).

Thus, the majority of the region shown in figure 14 may be underlain predominantly by serpentinized peridotite. The region of thinned continental crust in the upper basement layer is restricted to the N–S ridges in the northeastern sector of figure 14. Along LG-12, the upper basement changes at *ca*. 80 km distance from thinned continental crust towards the east to highly serpentinized peridotite towards the west beneath the complex series of basement highs. To the south of LG-12, the width of probable serpentinized basement within the OCT becomes even greater (Pickup *et al*. 1996).

We finally consider deep structures of the eastern Grand Banks and Newfoundland basin, which lies in conjugate position to Galicia Bank and the Iberia Abyssal Plain. Locations of deep reflection and refraction profiles across the margin are given in figure 16. Coincident reflection and refraction profiles exist along line 85-2, as originally reported by Keen & de Voogd (1988) and Reid (1994), although they are somewhat south of a conjugate position with respect to profile LG12 on the Iberia margin. Unfortunately, we were unable to produce a new depth section by combining both datasets, due to a critical missing section in the archived digital reflection data file. The original results are shown in figure 17. We note that the thinning of the crust across the continental slope is fairly rapid over an interval of *ca*. 40 km (60–100 km distance) and is primarily represented by thinning of the lower continental crust (reflections R and Mc). Only one or two tilted fault blocks (B1) can be observed in the upper basement. An HVLC layer ($V_\mathrm{p} = 7.2\ \mathrm{km\,s^{-1}}$) in the velocity model exists seaward of the thinned continental crust and is associated with landward dipping reflector 'L' and deeper basement (B2). A possible oceanic Moho reflection (Mo) does not occur until the extreme seaward-end of the profile, associated with rougher, elevated basement (B3) near the location of the J-anomaly (Chron M1 in figure 16 (Tucholke *et al*. 1989)). However, the velocity structure of the presumed oceanic crust is very unusual, with low basement velocity ($V_\mathrm{p} = 4.5\ \mathrm{km\,s^{-1}}$) lying above either a very high velocity lower crust or low velocity upper mantle ($V_\mathrm{p} = 7.7\ \mathrm{km\,s^{-1}}$). Unfortunately, this structure is not well-constrained due to the absence of receivers on the eastern end of the reflection profile. Similar velocities in the northern Newfoundland basin, adjacent to the southern boundary of Flemish Cap (figure 16), have been reported by Todd & Reid (1989).

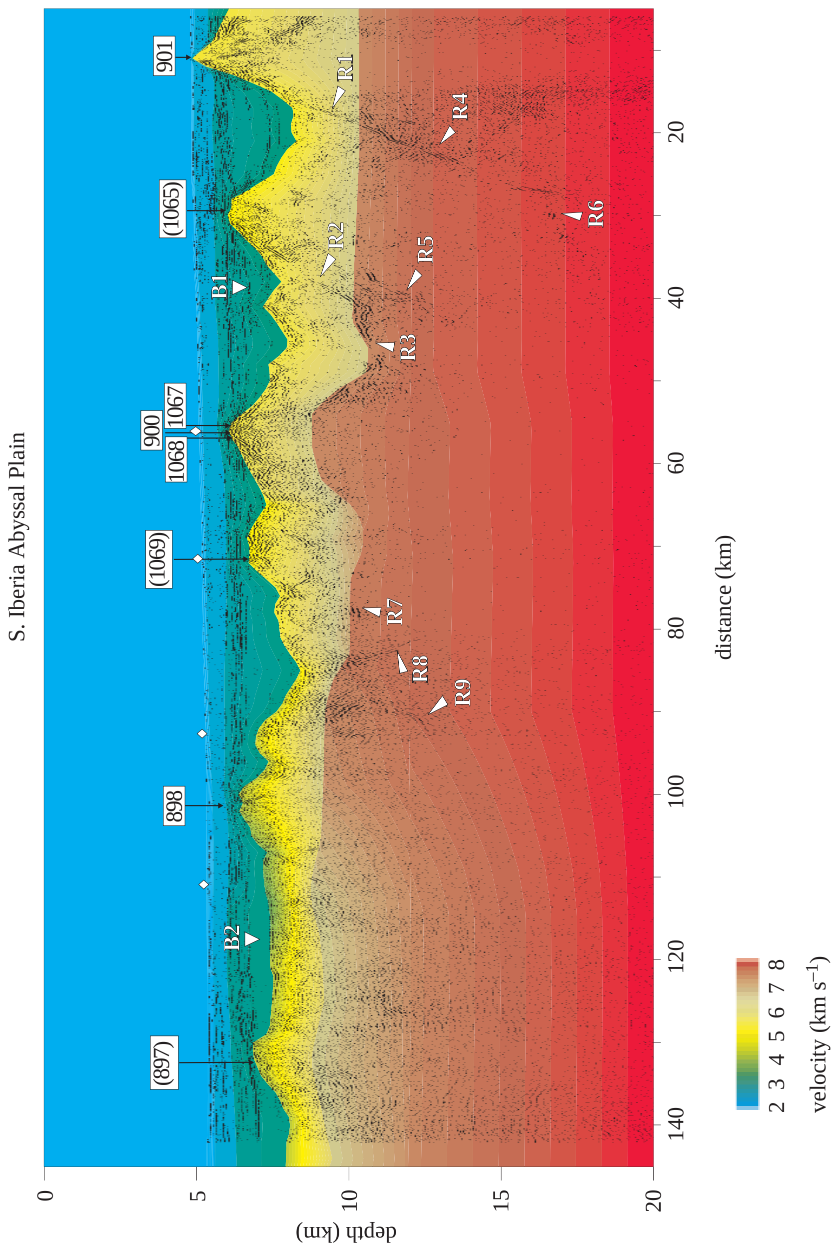

Figure 15. Seismic depth section across the southern Iberia Abyssal Plain, showing reflectivity along the time migrated LG-12 profile (Beslier *et al.* 1993) that is converted to depth using the velocity model of Chian *et al.* (1999) projected from adjacent profiles (shown in colour). Locations are given of ODP borehole sites (projected locations indicated by parentheses) and of crossing refraction profiles (white diamonds). Basement types (B1, B2) and reflections (R1–R9) are discussed in text.

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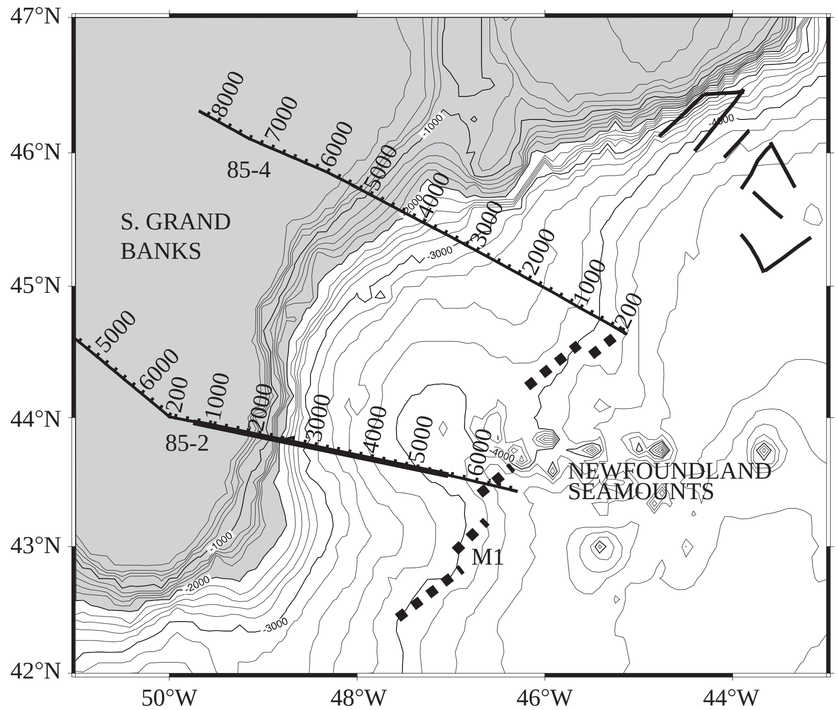


Figure 16. Bathymetry of the southern Grand Banks and Newfoundland basin (shaded for depths less than 2000 m) (ETOP05) with locations of the 85-2 and 85-4 deep reflection profiles (thin lines with shot numbers) (Keen & de Voogd 1988) and refraction profiles (thick lines) (Todd & Reid 1989; Reid 1994). Thick dashed lines show location of anomaly M1 (Tucholke *et al.* 1989).

3. Discussion

(*a*) *Interpretation of crustal sections*

The seven seismic sections across the three conjugate transects (figures 5, 7, 9, 10, 13, 15 and 17) show some remarkable similarities, which we group into the two end-member archetype crustal sections shown in figure 18. In one case, derived from the southern Iberia Abyssal Plain, Galicia Bank and the W. Greenland margins, rifting of the continent produced a zone of extremely thinned continental crust, characterized by tilted fault blocks that are primarily bounded at their base by a horizontal ('S-type') reflection. This reflection marks the top of the HVLC layer where velocities increase gradually with depth and approach a normal mantle velocity at depths of 15–20 km. Seaward of the thinned continental crust, a transitional region exists with low basement velocity and steep velocity gradient, but little reflectivity. Further seaward, the basement is characterized by a complex series of basement highs or ridges, which contain sea-floor-spreading magnetic anomalies and which eventually become sub-parallel to the strike of the oceanic spreading centre.

Thus, basement can consist of three types: rifted continental fault blocks, a smooth transitional region and elevated highs or ridges. Moho reflections are present only at the extreme ends of the section, beneath the thick continental crust on the landward side and beneath the thinner than normal oceanic crust on the seaward side. Within

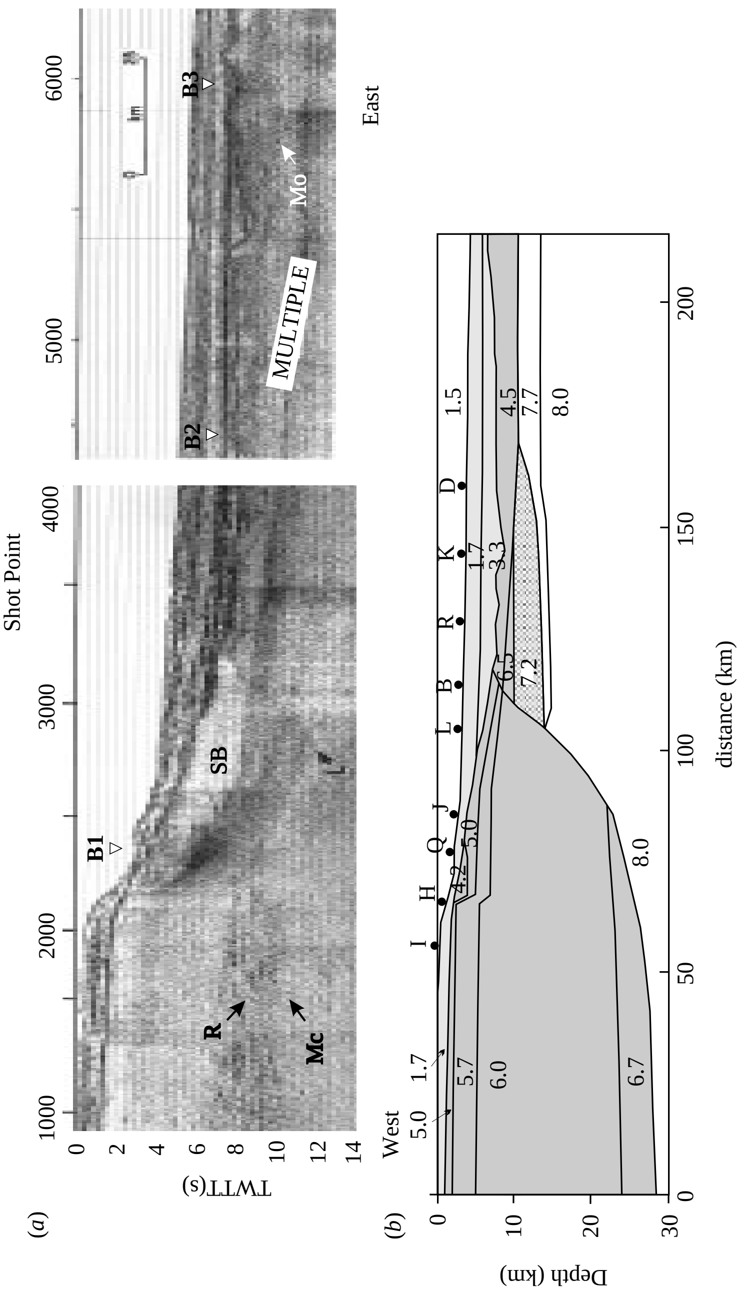
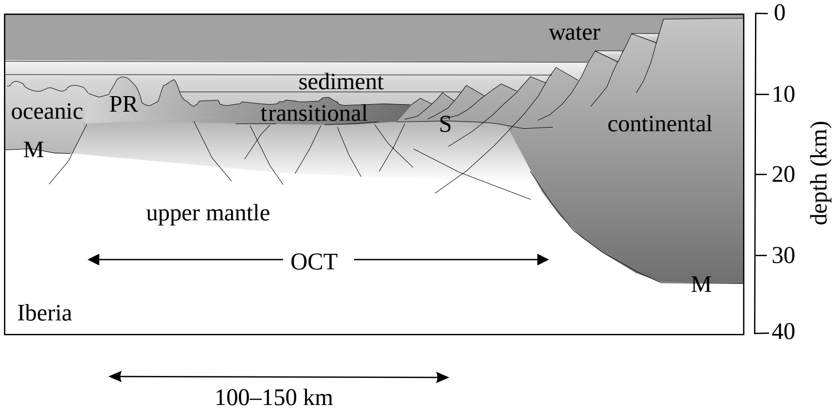


Figure 17. (*a*) Reflection travel time profile 85-2 (Keen & de Voogd 1988); and (*b*) velocity–depth model for coincident refraction profile (Reid 1994) across the southern Grand Banks–Newfoundland basin margin. Locations of refraction receivers (small filled circles), layer boundaries (solid lines) and layer velocities (in km s^{-1}) along the refraction model as shown. Area highlighted with pattern indicates HVLC layer. Reflections and basement types identified by labels are discussed in the text.



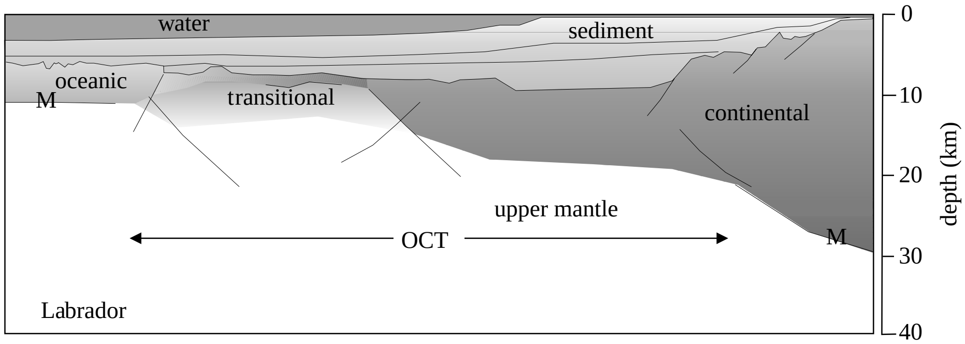


Figure 18. Cartoons of archetypal crustal sections across two end-member types of non-volcanic continental margins considered in this paper. The sediment-starved margin with zone of extended upper continental crust (upper panel) is based on profiles across the southern Iberia Abyssal Plain (Pickup *et al*. 1996; Chian *et al*. 1999), Galicia Bank (Whitmarsh *et al*. 1996) and SW Greenland margins (Chian *et al*. 1995*a*). The sediment-filled margin with zone of extended lower continental crust (lower panel) is based on the Labrador margin (after Chian *et al*. 1995*a*).

the OCT, Moho reflections are absent; instead, we observe a complex and variable reflectivity including both landward and seaward dipping events that can extend to depths of 15–20 km. The overall width of the OCT as well as the character of the various components varies from margin to margin. The southern part of the Iberia margin is characterized by a very wide transition region and complex series of seaward basement highs, while off Galicia Bank the transition region is very narrow or absent. On the W. Greenland margin, the zone of elevated basement highs between the transitional and oceanic basement types may be very restricted or absent, and the zone of initial thinning of the continental crust is extremely narrow.

Our other archetype section in figure 18 is based primarily on the Labrador margin. It differs from the other archetype primarily in the manner by which the continental crust thins. In this case, only one or two tilted fault blocks of upper continental crust are observed and the ‘S-type’ horizontal reflection is absent. A zone of thinned midto-lower continental crust extends seaward beneath a thick sediment basin. Further seaward, a transitional region with low basement velocity and HVLC layers occurs in a similar manner to the other archetypal crustal section. Also in similar manner, Moho reflections are missing throughout the OCT except at the extreme ends. In this case, however, dipping reflections within the upper mantle are less prevalent, except at the boundaries of the HVLC layer. For Labrador, the region of extended lower continental crust is very wide with a thick sedimentary basin; while for Flemish Cap and the Newfoundland basin, the width of extended lower continental crust is narrow or absent.

Characterization of the basement composition across the OCT has primarily focused on the transition across the 'S-type' reflector boundary and on the nature of the HVLC layer within the transition region. Previous interpretations of the 'S-reflector' suggested that it marks (a) the brittle–ductile transition within continental crust (de Charpal *et al*. 1978; Le Pichon & Barbier 1987); (b) a detachment fault that penetrates the entire lithosphere (Boillot *et al*. 1989*a*); (c) an intracrustal detachment fault (Sibuet 1992); (d) a detachment between upper continental crust and serpentinized upper mantle (Boillot *et al*. 1989*b*; Chian *et al*. 1995*b*; Reston 1996; Reston *et al*. 1996); or (e) a mixed boundary representing both (c) and (d) (Whitmarsh *et al*. 1996*b*).

Profiles considered in this paper indicate that the 'S-type' reflector primarily corresponds to an abrupt transition between the faulted upper continental crust and the top of the HVLC layer. We note, however, that some faults penetrate this boundary (e.g. R1/R4 and R5 on LG-12 profile), suggesting that faulting continued after its development. The high velocity of the HVLC ($V_p = 7.2$–$7.6\ \mathrm{km\,s^{-1}}$) is not present within the unrifted continental crust nor within the plutonic oceanic crust. This indicates that it consists either of an underplated melt zone, as previously suggested for the Newfoundland basin and Flemish Cap margins (Keen & de Voogd 1988; Reid & Keen 1990) or serpentinized peridotite, as suggested for the W. Iberian margin (Whitmarsh *et al*. 1993) and for the Labrador and W. Greenland margins (Chian & Louden 1994; Chian *et al*. 1995*a*, *b*; Chalmers 1997) and Newfoundland basin (Reid 1994). We favour the serpentinized peridotite interpretation for the HVLC because (a) no corresponding sequence of seaward-dipping reflections indicative of volcanic extrusives is observed in the overlying basement; and (b) melting models predict too great a thickness for the layer, in order to explain its high velocity by increased Mg content with increased melting temperature (Kelemen & Holbrook 1995). Similar arguments were used to argue in favour of a serpentinized upper mantle beneath Rockall Trough (O'Reilly *et al*. 1996).

In similar manner to our previous models for the Labrador and Western Greenland Margins (Chian & Louden 1992; Chian *et al*. 1995*a*, *b*), we also favour interpretation of the transitional 'crust' as primarily representing serpentinized peridotite. The characteristically thin, low velocity (4.0–$4.5\ \mathrm{km\,s^{-1}}$) of the basement layer and presence of an HVLC layer beneath are compatible with differing degrees of hydration, as previously suggested by Pickup *et al*. (1996) for the southern Iberia Abyssal Plain. Also, the velocity variations with depth within this zone in the Southern Iberia Abyssal Plain (Discovery 215 Working Group 1998; Chian *et al*. 1999) are not consistent with formation as oceanic crust at extremely low spreading rates, as previously discussed by Whitmarsh & Sawyer (1996).

(*b*) *Implications for geodynamic models*

The two primary implications from our seismic models for the OCT are that (a) late stages of continental rifting often appears asymmetric between margin conjugates, with one side having a broader region of highly faulted upper crust, typically underlain by a horizontal detachment between crust and serpentinized upper mantle; and (b) very little melt is produced over a zone 100–200 km wide where the continental crust has been thinned by very large (possibly infinite) stretching factors, exposing the upper mantle and leading to extensive serpentinization. We now briefly consider the implications that these results have for models of lithospheric extension.

One-dimensional models of decompressional melting (Pedersen & Ro 1992; Bown & White 1995) indicate that vertical conductive cooling during a finite period of rifting can decrease melt production at large stretching factors for extensional rates less than 15 km Ma^{-1}. Tett & Sawyer (1996) have shown that this result is also consistent with a two-dimensional finite-element model. This lack of melt production explains the existence of basins which are deeper than predicted by instantaneous rifting models with melt (Foucher *et al*. 1982). For Bown & White's (1995) one-dimensional model, rift durations of 15–25 Ma for the Goban Spur–Flemish Cap and Galicia Bank/Iberia–Newfoundland margins would result in less than 1 km of melt for stretching factors $\beta = 9$–15 (i.e. crustal thicknesses of 2–3 km remaining from an initial 30 km thick continental crust); and essentially no melt would be expected for the 40–65 Ma duration of rifting on the Labrador–Greenland margin (Chalmers 1997). Melt production would be even further inhibited by partial extension during earlier episodes of rifting (e.g. Galicia Bank), by two-dimensional conductive cooling, by advective cooling due to penetration of water into the upper mantle, or by simple shear extension of the lithosphere (Latin & White 1990). The seismic models are, however, not able to distinguish the presence of small amounts of melt within layers of primarily serpentinized peridotite because they can have similar velocities. Samples of basalt from ODP Sites 550 and 551 (Goban Spur) and gabbro at Sites 900 and 1067 (southern Iberia Abyssal Plain) suggest that some melting does take place. However, sampling of the OCT and constraints from radiometric dating on the relationship between rifting and melt production are still very limited. Even for the southern Iberia Abyssal Plain drilling transect (figures 14 and 15), which represents the best sampling of the OCT to date, basement samples primarily of highly serpentinized peridotite with essentially no basalt, are limited to basement highs. Extensive regions of deeper basement, as well as the thicker sections of syn-rift sedimentary sequences, remain essentially unsampled.

The asymmetry in the pattern of continental faulting may also be related to slow extension rate through the temperature dependence of rheological strength. The basic idea is that slow extension rates allow the upper mantle to cool and become more resistive to further extension than the equivalent thickness of quartz–feldspar continental crust. This may limit the amount of extension (England 1983) or cause the deformation to move laterally to an adjacent, undeformed area, therefore widening the rift zone (Kusznir & Park 1987). This effect may also be influenced by the initial conditions of continental crustal thickness and thermal conditions within the lithosphere (Buck 1991). Recently, Bassi (1995) has included the additional effects of mechanical instability, which under certain conditions may render the effect of syn-rift cooling less predictable through changes between creep and plastic deformation mechanisms. Bassi's (1995) results indicate that diffusive cooling of the mantle will modify the geometry of the rift primarily when the upper mantle is initially weak and viscous. In this case, the locus of maximum strain-rate moves laterally once the initial rifted region starts to cool until concentrating at the transition between deformed and undeformed lithosphere. The predicted pattern of asymmetric break-up, with juxtaposition of wide, uniformly thinned and very narrow margin conjugates closely resembles the crustal geometry which we have determined for the Labrador–Greenland margins. This situation may, however, be somewhat unique to this margin where rifting was extremely slow and where the lithospheric thermal structure may have been warmed and weakened by the beginning of the Iceland plume preceding eruption of basalts in the Palaeogene. Other peculiarities of individual rifting histories might also help to generate asymmetric break-up. For instance, Tett & Sawyer (1996) have shown that asymmetric rifting would be produced by multiple periods of extension relevant to the situation for Galicia Bank.

Because of numerical difficulties with including the effects of large-scale faults, most geodynamical models break down during the final phase of crustal rupture at large amounts of extension. Results such as those of Harry & Sawyer (1992) and Boutilier & Keen (1994) indicate that the presence of large-scale faults can influence the form of final break-up. Laboratory models, such as those of Brun & Beslier (1996), suggest that boudinage structures in the brittle mantle may be detached and offset from the faulted brittle crust by a lower-crustal dutile layer. This effect can produce an asymmetric crustal break-up, including exposure of the upper mantle, even though the underlying mantle necking is symmetric. A difficulty in applying the laboratory results to real margins is that they cannot include the effects of temperature on rheology, such as those included in the computer simulations, nor the formation of melt. A further complication is the likely existence of serpentinized mantle to depths of 15–20 km. Recent laboratory studies (Escartín *et al.* 1997) indicate that serpentinization can reduce the integrated strength of the lithosphere by up to 30%. This weakening of the mantle may help to explain the existence of the occasionally wide zones of serpentinized upper mantle within the OCT, as suggested by our integrated seismic reflectivity and velocity profiles.

4. Summary

Joint analyses of seismic reflection lines and velocity models determined from wide-angle reflection/refraction profiles indicate common features across the three North Atlantic conjugate non-volcanic margins studied.

(i) The late stages of continental rifting and break-up appear asymmetric with one margin having a broader zone of highly thinned crust, which is often underlain by a sub-horizontal reflector. This reflection is most probably associated with the contact between the base of the tilted fault blocks, within the seaward end of the stretched continental basement, and a high velocity lower layer of serpentinized mantle.

(ii) A zone of transitional basement exists seaward of the stretched continental crust and landward of the first oceanic crust. This zone is typically characterized by a low basement velocity and a lower layer with high velocity. It is associated with characteristic changes in basement morphology and depth

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across the transition zone, with the deepest, flat-lying basement on the landward side and elevated basement highs on the seaward side. Although this zone is most likely composed primarily of serpentinized mantle, it could also include minor amounts of melt. A Moho reflection is not present within this zone and does not appear until further seaward beneath the first true oceanic (plutonic) crust. The oceanic crust is often thinner (*ca.* 5 km thick) than typical of average oceanic crustal thicknesses (*ca.* 7 km thick).

Therefore, a very broad transition region (OCT) 150–250 km wide and composed primarily of serpentinized mantle forms the boundary between the initial stages of continental rifting and the final stage of oceanic crustal melt. The existence of this transition zone, as well as the asymmetry of late stage continental break-up, is most likely a consequence of very slow rates or multiple periods of extension. Laboratory (sandbox) models of continental rifting can simulate the occurrence of this transitional region, but computer models that include the effects of melt do not. The probable effect of serpentinization in lowering the effective strength of the mantle during late stages of rifting has not as yet been included in any model for lithospheric extension. Although the specific structures that we observe within the OCT of non-volcanic rifted margins differ from those created by oceanic rifts at very slow rates of spreading (Mutter & Karson 1992; Louden *et al*. 1996), some aspects of the rifting process may be similar. In each case, what we observe is a balance between tectonic thinning by crustal extension on the one hand and volcanic production on the other, which can vary both in space and time. At very slow rates of extension, tectonic extension dominates to produce large-scale faulting that can increase the penetration of water to lower crustal and upper mantle layers. Important influences on rifting style due to enhanced cooling by advection and to rheological weakening by serpentinization are not as yet understood. Through these processes, however, the structures within the transition zone may have important parallels to the existence of serpentinized mantle within oceanic fracture zones and sections of exposed lower crust and mantle at intersections of fracture zones and mid-ocean spreading centres.

At present, our understanding of these important processes is limited by a lack of data. Even after some considerable efforts over the past 20 years, no single conjugate margin transect contains a coherent suite of both seismic reflection and refraction profiles and basement samples from boreholes. The seismic profiles across the Labrador and SW Greenland margins represent the only crustal sections across the complete continent–ocean transitions of conjugate non-volcanic margins that include complete profiles of both deep multichannel seismic reflection and wide-angle refraction datasets. The best set of coincident basement samples and seismic profiles exist within the transition region of the southern Iberia Abyssal Plain, but the borehole sites are limited to basement highs and only general links can be made at present to the margin conjugate in the Newfoundland Basin. Profiles presented in this paper are limited to two-dimensional transects and we have only begun to look at three-dimensional variations within a single basin (Discovery 215 Working Group 1998; Chian *et al*. 1999). Finally, almost all data relevant to the study of non-volcanic margins come from a particular region of the North Atlantic where break-up occurred between North America and Europe through a rather complex pattern. It would be very instructive to compare these results with another quite different non-volcanic margin, but at present such data are lacking. For instance, recent velocity models of unreversed wide-angle profiles across the non-volcanic South Australia margin do not

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require a high-velocity lower basement layer within the OCT (Finlayson *et al*. 1998). However, the absence of ocean bottom receivers along these profiles severely limits the resolution of the crustal models. Much more detailed data need to be collected on suites of coincident refraction and reflection profiles, similar to or better than those shown in this paper. The resources required for collecting such images in both two and three dimensions on a larger number of conjugate margins are immense. However, as drilling for hydrocarbon resources progresses from shallow continental shelves to offshore regions in deeper water, the study of these deep basins may benefit in the future from the far greater resources available to commercial industry.

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Discussion

M. Osmaston (*Woking, UK*). As Dr Louden has emphasized, a strip of 'transitional' crust commonly intervenes between true continental and oceanic crust along passive margins. One of the possible origins I think he should consider is that this crust was separation-generated by a mid-ocean ridge process that would have been drastically modified by the concurrent heavy sedimentation that is to be expected at this stage. This is a matter to which, I feel, much more attention needs to be given. The northern Gulf of California and the Salton Sea offer the only natural laboratory for the study of this as an ongoing process. Sediment blocking of hydrothermal chilling channels and a transition to a sill mode of magmatism would ensure the absence of magnetic anomalies. The Salton Sea Drilling Project found syn-rift, inter-sill sediments metamorphosed to a high seismic velocity (laboratory-determined, pressure-corrected V_p just under $6\ \mathrm{km\,s^{-1}}$ at 2.5 km depth (Tarif *et al*. 1988), thus rendering them basement-like from a seismic point of view. Further work surely needs to be done in the northern Gulf of California (or at least a reconsideration of earlier work) with this perspective in mind, though the proximity of transform offsets there must, to some extent, prejudice the thermal/magmatic correspondence with situations having more orthogonal separation. On established passive margins, it seems

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highly desirable to drill and core the seismic basement of the 'transitional' crust at issue.

K. E. Louden. I certainly agree that it would be very useful to drill additional boreholes into the transitional crust of the southern Iberia Abyssal Plain, particularly at sites in the deeper parts of the basin. However, I do not agree with the parallels you draw between the regions of transitional crust that I have described and active, sediment-filled rifts such as the Gulf of California. For the regions I have described, the syn-rift sediment fill is not very thick. Of these margins, only the Labrador Margin contains a thick sedimentary basin over the transitional crust, and even in that case the larger part of the sediment fill post-dated rifting by a considerable period. Also, the seismic velocities in the upper basement of the transitional crust, for all cases where we have good data, are lower than expected when compared to the oceanic crust further seaward, rather than higher. This is opposite to the situation that, as you have correctly indicated, would be expected from oceanic basement formed beneath a large sediment fill.

R. S. White (*Bullard Laboratories, University of Cambridge, UK*). We have seen in the cross-sections of non-volcanic continental margins a zone about 250 km wide which has been identified either as 'exposed mantle' or 'serpentinite', or as mixed fragments of continental rock, solidified melt and serpentinite. Whenever mantle decompresses beneath thin lithosphere it has a propensity to melt, and indeed it is hard to stop it melting. Can Dr Louden explain what circumstances might allow a 250 km wide rift zone to develop with only extremely small amounts of melt being produced?

K. E. Louden. The consistent pattern across these margins that we observe in seismic profiles is one in which neither characteristically continental material nor characteristically oceanic material is observed within a wide transitional zone in the middle. Overall, results from drilling and seismic models are most consistent with a gross composition of this zone as exposed mantle, which has been serpentinized to varying degrees. This is not to say that no melt is ever present within this zone. Certainly for the Goban Spur margin there has to have been some melt because DSDP Leg 80 drilled 60 m of tholeiitic basalt. That's not a lot compared to seismic models, but it still looks exactly like mid-ocean basalt. So there has to have been some melt generated in that particular location. In the southern Iberia Abyssal Plain, where we have the greatest number of basement samples, virtually no basalt and only a small amount of gabbro have been recovered. So I think that the amounts of melt generated within these transition zones must be relatively small compared to their horizontal scale.

How this situation can be explained in terms of current theoretical models of extension is not clear, possibly because none of the models contains the full story. A simple answer is that these margins are formed by very slow rates of extension over periods greater than 15–20 Ma. This would inhibit melt production due to conductive cooling, as Professor White and Jonathan Bown have shown. The two-dimensional model of Gianna Bassi that I showed suggests that slow extension could also explain the asymmetry that is observed in conjugate sections, particularly in the Labrador–Greenland profiles. However, these models break down when the extension becomes very large, so they cannot tell us much about what happens as the plates finally break apart. To do this, one must use the type of sandbox models that Jean-Pierre

Brun showed (this issue), which explain some of the patterns that we observe in the seismic data quite well. In addition, if you weaken the mantle material by production of serpentinite and cool the plate by circulation of water, perhaps these additional processes will also inhibit melt production during extension. But we don't know because such effects have not been included in the models yet.

R. B. Whitmarsh (*Southampton Oceanography Centre, UK*). I would respond by saying that where we drilled on the ODP Site 900 high there is not very good evidence there at present for syn-rift melt, only for syn-rift tectonism of the gabbro. But at Sites 897 and 899, drilled further west in the ocean-continent transition, there are breccias and debris flows that contain small clasts of basalt which might be syn-rift, but we don't know. The other point is that about 40 km south of the drill sites there is a deep seismic reflection profile (IAM-9). The magnetic anomalies along this profile, which Simon Russell (SOC) has inverted, suggest that there might be magnetic source bodies, whose tops are buried 4 km or so within the basement, which could represent syn-rift melt which did not ascend to the top of basement. These bodies might be evidence of a limited amount of melting. The last point I would like to make is that you talked about 250 km of stretching. There was certainly 130 km of stretching at the latitude of the IAM-9 line, in the southern Iberia Abyssal Plain, but we do not really know what happened on the Newfoundland margin. Therefore, we cannot necessarily assume stretching was symmetric and double that number. Thus there was a minimum 130 km of stretching.

D. McKenzie (*Bullard Laboratories, University of Cambridge, UK*). It is extremely difficult to stretch the lithosphere by the amount Dr Louden proposes without generating the melt. It is not a question of whether faulting is involved, or where you put the faults. The problem is caused simply by the change in pressure: however you move material upwards into the space you create by separating the plates, it will melt unless you can cool it. A small amount of melt produced by this process will have a distinctive composition. Because melting will extract all incompatible elements, the melt composition will not be in the least like MORB. If you have any basalt at all, it should be diagnostic of the process you propose. The chemistry is easy to interpret, because the incompatible element concentrations are approximately proportional to the thickness of the crust that is formed. The peridotite compositions should also be less depleted than they are on ridges, but are harder to interpret, because they are spatially variable.

R. B. Whitmarsh. The ODP boreholes in the southern Iberia Abyssal Plain encountered serpentinized peridotite at four sites (Sites 897, 899, 1068 and 1070) and there is abundant core material. However, basalt was found only at Site 899 and then in very small quantities as small clasts in a debris flow deposit.

D. McKenzie. If you've got peridotite, you should have some clinopyroxene, whose composition can be measured using the using an ion probe.

K. E. Louden. I just wanted to make a comment on that, in the sense that my conclusions are not based on results from the computer models or the sandbox models. I'm looking at seismic information of crustal properties. You may say that it's impossible to create that kind of zone without melting, but all I'm saying is that whatever is there is very unusual compared to what we see at very slow spreading ridges, normal spreading ridges, or stretched continental blocks. It doesn't look like any of these. We have very few images even to look at, but all the images that I can see give a systematic pattern. The stretching in these areas is probably very slow, and it certainly would be nice to look at some other areas where the extension rates were higher but large amounts of volcanism still weren't produced. One such example might be south of Australia, where there probably was a non-volcanic margin and where we see similar kinds of features from a region of the mid-ocean ridge system which is totally different. The North Atlantic is really the only region of non-volcanic margins which have been well studied and maybe things were odd there for some reason.

R. C. L. Wilson (*The Open University, Milton Keynes, UK*). I have a comment and a question related to the duration of the rifting episodes in North Atlantic non-volcanic margins. Interpretation of seismic reflection profiles in the Galicia Bank and Iberia Abyssal Plain area using information from ODP Leg 149 and DSDP Site 398 suggests that rifting occurred over a very short time interval, perhaps only about 5 Ma (Wilson *et al*. 1996). Results from Hole 1069A favour this interpretation (ODP Leg 173 Shipboard Scientific Party 1998). This is much shorter than the 20 Ma long rifting phase identified to the west of Galicia Bank by Boillot & Winterer (1987). If correct, this conclusion and $\beta \sim 5$–7 calculated from total subsidence plots by Wilson *et al*. (1996) make the lack of evidence of significant mantle melting even more puzzling. How well constrained is the duration of the rifting episode in the other areas described in Dr Louden's paper?

K. E. Louden. The duration of rifting on the margins that I have discussed is typically considered to fall within the range of 15–25 Ma and possibly as large as 35–50 Ma for the Labrador–Greenland margins. These values are consistent with a reduced melt production in the one-dimensional model of Bown & White (1995). In addition, all these margins have evidence for extended periods of rifting which pre-date the actual break-up. However, I do not believe that the duration of rifting is very well constrained because cores into the base of the deeper rifted basins and into the earliest oceanic basement from which primary dates can be determined are not generally available. Instead, we rely on a collage of both primary and secondary information (e.g. wells but generally on topographic highs, interpretation of seismic stratigraphy and identification of magnetic anomalies), which are open to interpretation (e.g. Roberts *et al*. 1993). If the short duration suggested for the southern Iberia Abyssal Plain is correct, this certainly poses a major problem in explaining the apparent lack of magmatism. However, in this case direct constraints from drilling the deep faulted rift basins as well as the earliest oceanic crust are still lacking.

K. McClay (*Royal Holloway College, University of London, UK*). I just want to make the point that in rift systems where the extensional fault systems control the magmatism it will not be surprising that little evidence of magmatism will be found where only the highs are drilled. In addition, in regions such as the southern Red Sea and the adjacent Danakil area, there is a wide zone of extension with significant intervening horst blocks (cf. the Danakil Horst). If in this region extension continued to develop to an extent similar to the case Dr Louden has described, the result would be a very wide zone of extended continental crust similar to that on the west Iberia continental margin.

K. E. Louden. It would be very nice to be able to look at the more active centres of this process. Ocean Drilling Program Leg 180 will take place soon (June–August

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1998), to study extension of the Eastern Woodlark Basin into the continental crust of Papua New Guinea and the nature of the active low-angle faulting. In addition, other drilling has been proposed for the Gulf of Aden to look at some of the young margins there. Unfortunately, the politics are not very conducive to working in that region.

N. Kusznir (*University of Liverpool, UK*). Dr Louden interprets the difference between pairs of conjugate margins on either side of the ocean in terms of upper plate/lower plate. Is it not possible to explain the differences, or at least to understand the differences, in terms of just natural variability or natural heterogeneity in the continental material?

K. E. Louden. In our Labrador–Greenland transect, we show a very large asymmetry in the kind of continental extension that took place between the two conjugate pairs. This can be interpreted grossly in terms of upper crustal–lower crustal pairs, which were produced as the centre of extension migrated toward the Greenland side. Of course, this scenario is very simplistic and what happens as you finally pull the thing apart is undoubtedly a very complex process. For these profiles, we paid particular attention to measuring the seismic properties of the original unextended continental crust and they are quite similar. So I don't think that these differences are caused by heterogeneity within the continental crust on the opposite sides.

However, we have just the one transect so we cannot address the very interesting problem of whether heterogeneities in the continental crust parallel to the margin somehow influence the style of extension. Dale Sawyer looked at this problem previously using gravity profiles and extension histories. Labrador is a particularly interesting location because changes occur as you go up the basin in terms of the age and structure of the pre-existing continental crust. If extension is influenced by pre-existing conditions within the crust and lithosphere, as some models suggest, these changes in the original crustal structure might influence the style of extension that occurs. But at present we can't really say because we just have this one long transect.

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