

Structure of the Flemish Cap margin, Newfoundland: insights into mantle and crustal processes during continental breakup

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Abstract: Seismic reflection and refraction data from the Flemish Cap margin off Newfoundland reveal the large-scale structure of a magma-starved rifted margin. There is little evidence for significant extensional deformation of the Flemish Cap, consistent with the hypothesis that it behaved as a microplate throughout the Mesozoic. The seismic data highlight important asymmetries at a variety of scales that developed during the final stages of continental breakup and the onset of oceanic sea-floor spreading. In strong contrast to the conjugate Galicia Bank margin, Flemish Cap shows: (1) an abrupt necking profile in continental crust, thinning from 30 km thick to 3 km thick over a distance of 80 km, and a narrow, less than 20 km-wide, zone of extremely thin continental crust; (2) no clear evidence for horizontal detachment structures beneath continental crust similar to the 'S' reflection; and (3) evidence for at least a 60 km-wide zone of anomalously thin oceanic crust that began accreting to the margin shortly after continental crustal separation. The oceanic crust averages only 3–4 km thick and in places is as thin as 1.3 km thick, although seismic layer 3 is missing where this occurs. The data suggest that there are large spatial and temporal variations in the available melt supply following continental breakup as oceanic sea-floor spreading becomes established. In addition, wide-angle data show that anomalously slow mantle P-wave velocities appear approximately where continental crust has thinned to 6–8 km thick, indicating that low-degree serpentinization begins where the entire crust has become embrittled.

Continental extension leading to lithospheric rupture and the creation of new oceanic sea-floor spreading systems has long been a cornerstone of global plate tectonics. Because rifted margins contain the geological record of key processes involved with extension and breakup, investigations of these margins are central to advancing our understanding of lithospheric rupture.

Thinning of the crust and lithospheric mantle is the most obvious manifestation of continental extension. Two key consequences of thinning are brittle failure of the upper crust and decompression melting of asthenospheric mantle as it nears the surface. Despite the seeming simplicity of these two effects, considerable controversy over basic aspects of how they work has resulted from enigmatic observations from a number of areas. For the brittle behaviour of the crust, the possible role of low-angle normal faulting and sub-horizontal detachments during extension is still debated (e.g. Lavie *et al.* 1999; Lavie & Manatschal 2006). In the case of mantle melting, there is no firm consensus on the primary causes of the extreme range of magmatic productivity during the initial stages of

sea-floor spreading (e.g. Boutilier & Keen 1999; Nielsen & Hopper 2004).

The rifted passive margins around the North Atlantic are a unique natural laboratory where many of these questions can be studied. Magnetic anomalies permit the reconstruction of conjugate margin pairs with a reasonable degree of certainty (Srivastava & Tapscott 1986). This is critical as many rift systems exhibit asymmetry that can only be quantified by characterizing the entire rift system. In addition, the North Atlantic margins are themselves extremely diverse. Figure 1 shows the regional setting with measured oceanic crustal thicknesses marked along several of the margins. At one end are the margins along Greenland, Norway and the British Isles where 15–30 km-thick oceanic crust accreted to the rifted continental margin (Holbrook *et al.* 2001). At the other extreme are the margins of SW Greenland and Iberia, where 0–5 km-thick oceanic crust is observed (Chian & Louden 1994; Minshull *et al.* 2001).

The lack of volcanism along these latter margins has proven particularly useful for studying

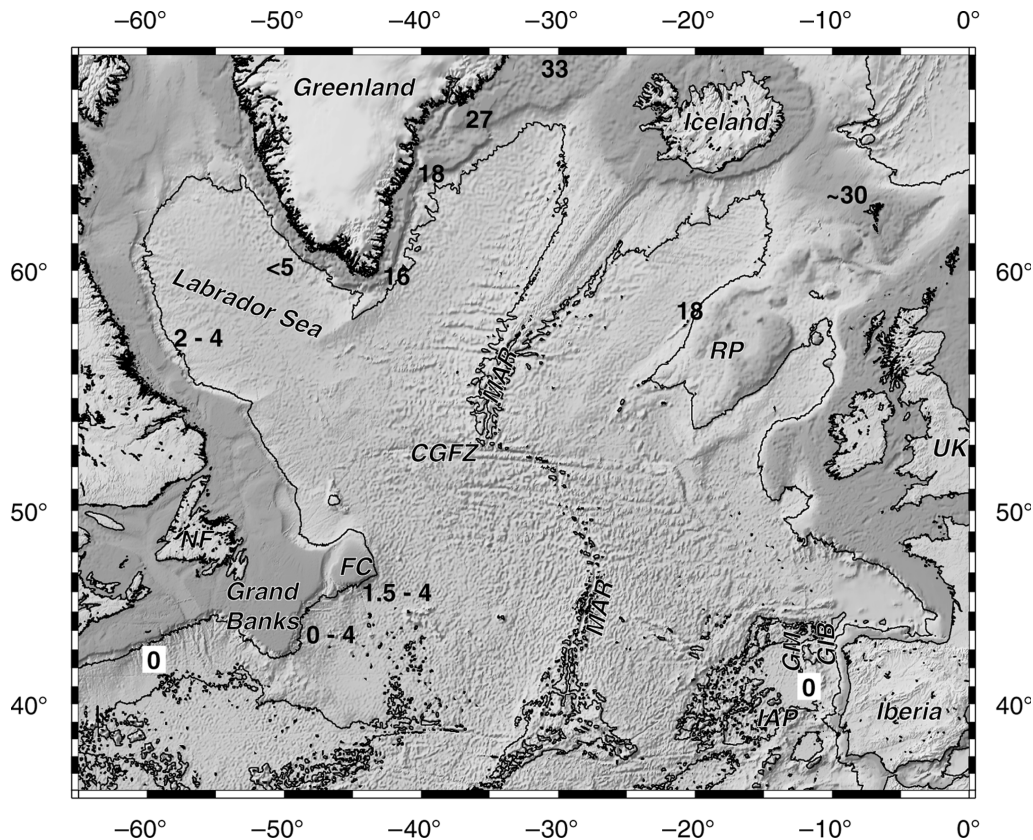


Fig. 1. Regional overview of the North Atlantic Ocean and rifted margins. The map is shaded relief of topography and bathymetry with shallow-water areas highlighted in darker shades. Bold numbers indicate the thickness of igneous crust produced immediately after continental crustal separation (values taken from Fowler *et al.* 1989; Chian & Loudon 1994; Chian *et al.* 1995; Smallwood *et al.* 1999; Holbrook *et al.* 2001; Whitmarsh *et al.* 2001; Funck *et al.* 2003, 2004; Hopper *et al.* 2004). CGFZ, Charlie Gibbs Fracture Zone; FC, Flemish Cap; GIB, Galicia Interior Basin; GM, western Galicia margin; IAP, Iberia Abyssal Plain; MAR, Mid-Atlantic Ridge; NF, Newfoundland; RP, Rockall Plateau; UK, United Kingdom.

mechanical deformation of the upper crust because the extensional structures are accessible to seismic imaging techniques. On the Galicia Bank margin off Iberia, the strong regional 'S' reflection is observed that is now believed to be a detachment surface at the crust-mantle boundary (Reston *et al.* 1996). The detachment is thus a major rheological boundary that probably formed as a heavily serpentinized ductile shear zone during the final stages of extension (e.g. Reston *et al.* 1996; Pérez-Gussinyé & Reston 2001). The uniform sense of shear and rotation indicated by fault blocks above 'S' has led to various hypotheses about possible asymmetry during extension and breakup.

Farther south along Iberia, a broad, up to 100 km-wide, zone of exhumed serpentinized mantle was exposed at the sea floor (e.g. Whitmarsh

et al. 2001), reminiscent of Hess' (1962) original hypothesis for the formation of oceanic basins. In this so-called transition zone, the mantle is mechanically unroofed by brittle extension and interacts with sea water to form a thin layer of serpentinite with crust-like seismic velocities (Dean *et al.* 2000). Initially, this exhumation is thought to be largely amagmatic, but as extension progresses more melt is eventually produced. In the Alps, where these melt products can be sampled and analysed, there is a progression from transitional mid-ocean ridge basalt (T-MORB) towards the continent to normal mid-ocean ridge basalt (N-MORB) towards the ocean (Desmurs *et al.* 2002; Schaltegger *et al.* 2002). Desmurs *et al.* (2002) suggest that this indicates progressive thinning of continental lithospheric mantle, which is eventually

replaced by asthenospheric 'oceanic' mantle as a magmatic sea-floor spreading system becomes established. It is becoming increasingly clear that resolving the nature and origin of 'transition zone' crust is central to understanding final lithospheric rupture along magma starved margins.

In this contribution, results from a recent seismic reflection and refraction survey conducted off the coast of Newfoundland are summarized. The data shown are published in Funck *et al.* (2003) and Hopper *et al.* (2004, 2006). We do not attempt to replicate all of those results here, but focus on key summary figures and portions of the data set that are particularly relevant for the numerical modelling exercises described in this volume.

Regional tectonic setting and survey description

The extensional history of the region between Iberia and Newfoundland spans nearly 100 million years, beginning with an early phase of rifting in the Late Triassic–Early Jurassic that formed the major basins within the Grand Banks and off Portugal (e.g. Tankard & Welsink 1987; Wilson 1988;

Tucholke *et al.* 1989; Rasmussen *et al.* 1998; Wilson *et al.* 2001). A 40–50 million year period of quiescence was followed by a second major phase of extension that ultimately led to complete lithospheric rupture and the establishment of magmatic sea-floor spreading in the North Atlantic some time between Barremian and Aptian time (127–112 Ma using the timescale of Gradstein *et al.* 1994). Details regarding the full history and development of the region are summarized in Tucholke *et al.* (2007) and Tucholke & Whitmarsh (2007).

The SCREECH survey (Studies of Continental Rifting and Extension off the Eastern Canadian sHelf) was conducted in 2000 to better understand the large-scale structure of the Newfoundland margin at positions conjugate to the major geophysical surveys and drilling transects off the Iberian margin (Funck *et al.* 2003; Hopper *et al.* 2004, 2006; Lau *et al.* 2006a, b; Shillington *et al.* 2004, 2006; Van Avendonk *et al.* 2006). Multichannel seismic reflection and wide-angle seismic refraction data were acquired along three primary transects. Each transect began on the shelf where continental crustal thinning should be minimal and continued seaward well past magnetic anomaly M0 identified over oceanic crust (Figs 2 & 3). Crustal thickness,

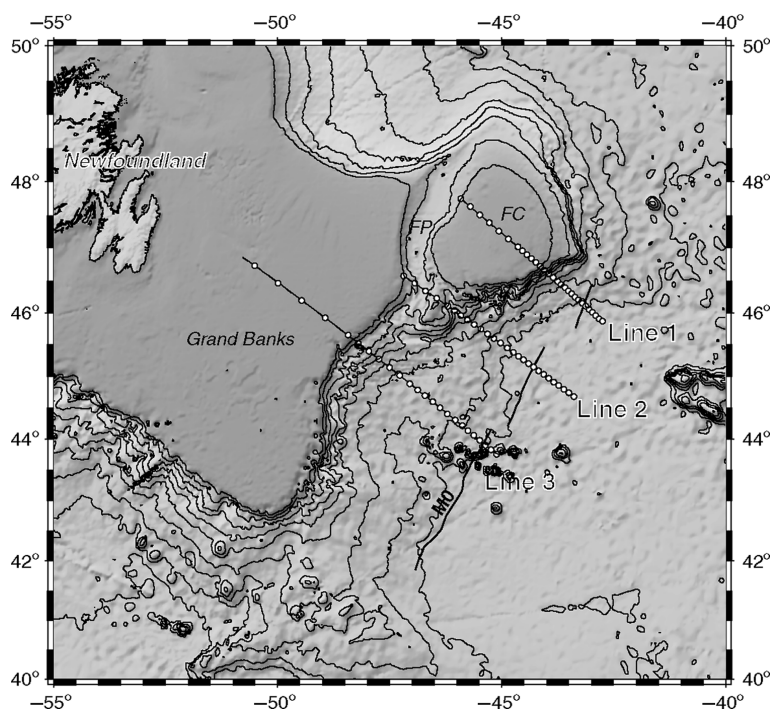


Fig. 2. Expanded view of the region offshore Newfoundland showing the SCREECH survey. Open circles are OBS locations along the three major transects, labelled Lines 1, 2 & 3. The thick black line is magnetic anomaly M0 as picked by Srivastava *et al.* (2000). FC, Flemish Cap. Bathymetric contours are spaced every 500 m.

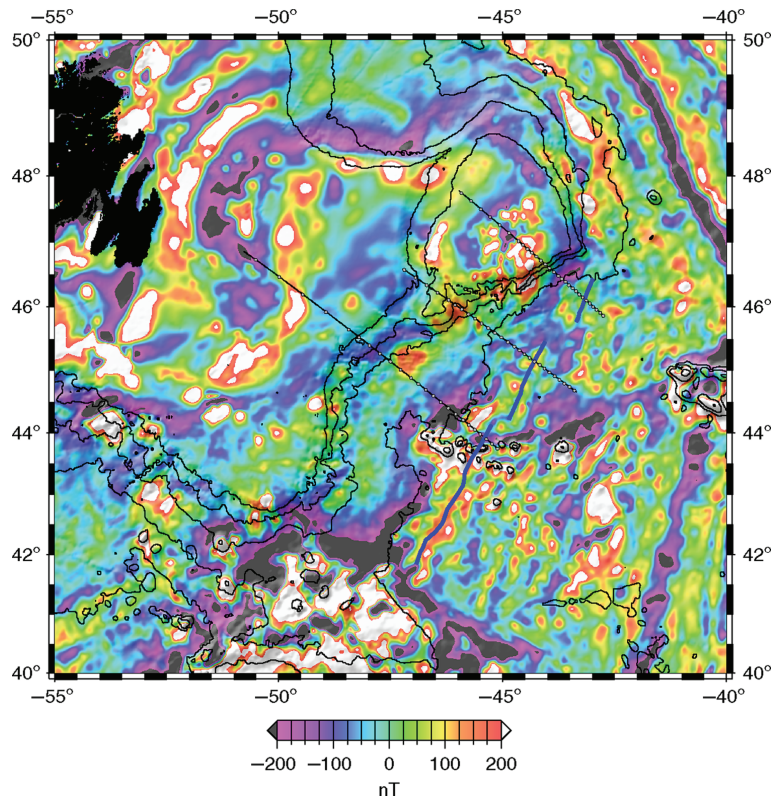


Fig. 3. Regional magnetic anomaly map based on the data of Verhoef *et al.* (1992). Seismic transects and the anomaly M0 pick from Figure 2 are also shown. Bathymetric contours are at 1000 m intervals, and the bathymetry is shaded with illumination from the NW.

seismic P-wave velocity and reflection imaging provide critical constraints on the large-scale structure of the full rift system between North America and Europe.

Here, the primary focus is on the northernmost transect across the Flemish Cap (Line 1 in Fig. 2), a block of Appalachian continental crust that is conjugate to the northernmost Iberian peninsula where the Galicia Interior Basin (GIB) and Galicia Bank margins are located. The profile is also conjugate to the Ocean Drilling Program (ODP) Leg 103 drilling transect as well as a number of seismic reflection and refraction profiles described in Reston *et al.* (1996), Zelt *et al.* (2003) and Henning *et al.* (2004). Data from SCREECH Line 1 are summarized in Figures 4 & 5.

The oldest rift-related sediments found by ODP Leg 103 are Tithonian in age (c. 142 Ma Boillot *et al.* 1987; Enachescu 1987) and, within the GIB, the oldest sediments that can be identified with confidence are also Tithonian (Murillas *et al.* 1990; Pérez-Gussinyé *et al.* 2003). Murillas *et al.* (1990)

described a seismic reflection sequence beneath the Tithonian unit that they proposed is either Late Triassic–Middle Jurassic or older Palaeozoic sediments. In addition, the Flemish Cap graben along the SW part of the cap contains evaporites that are presumably Triassic (Enachescu 1987). However, available data suggest that the Late Triassic phase of extension may have only minimally affected the region this far north. In considering the timing of rifting and the duration of extension along the conjugate profiles, however, it should be noted that drilling and sampling along the Flemish Cap side of the margin is essentially non-existent.

An additional point regarding the conjugate margin is that continental crust terminates just seaward of the ‘S’ reflection at a single peridotite ridge beyond which magmatically produced oceanic crust is thought to exist. This is unlike the area farther south along Iberia where a broad, approximately 100 km-wide, zone of exhumed continental mantle and multiple peridotite ridges were exposed at the sea floor prior to magmatic sea-floor

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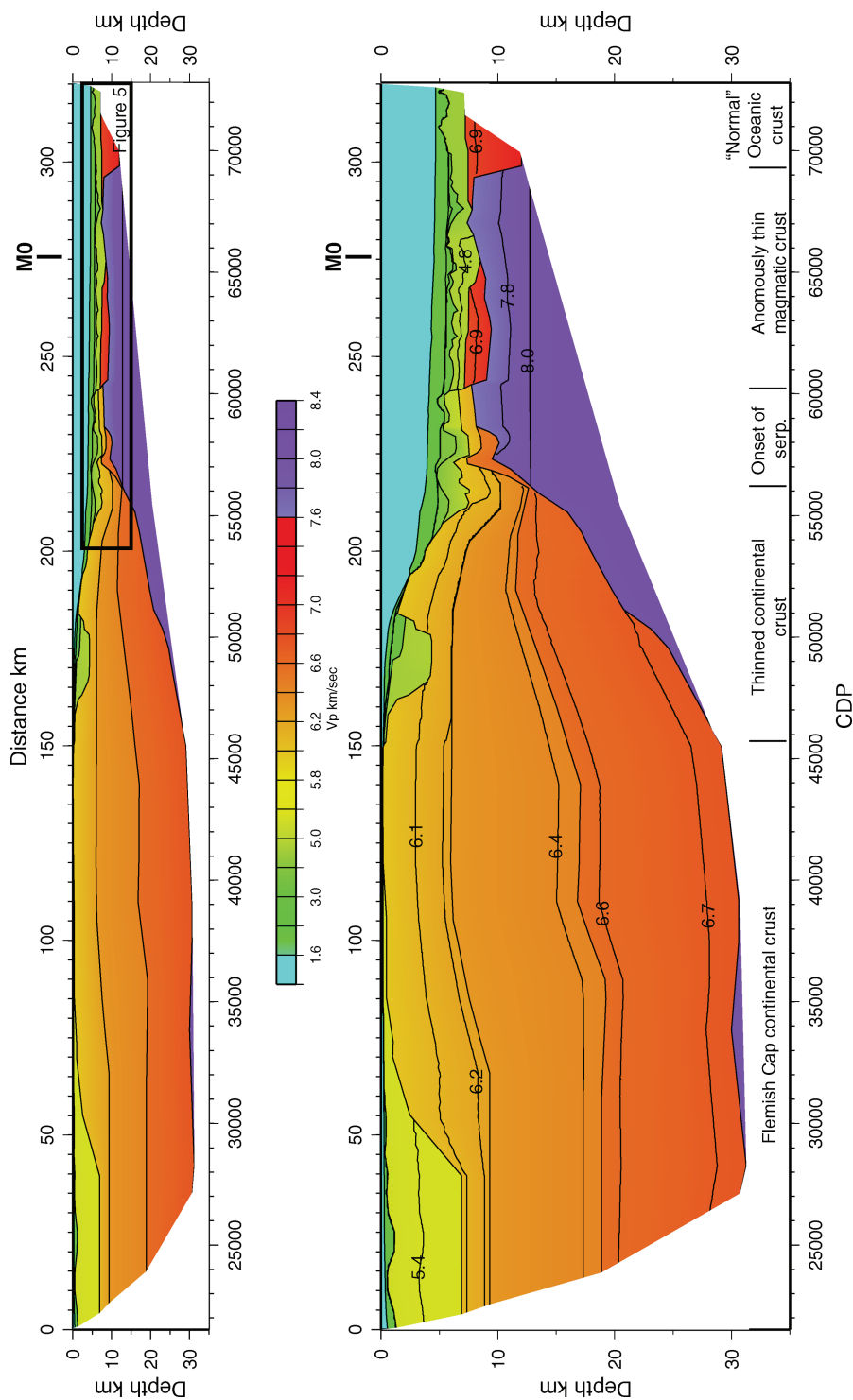


Fig. 4. Seismic velocity model along the SCREECH 1 transect derived from wide-angle data. See text and Funck *et al.* (2003) for details. Top panel is the model with no vertical exaggeration. Bottom panel is exaggerated 3:1 to show detail. Labelled contours are P-wave velocity.

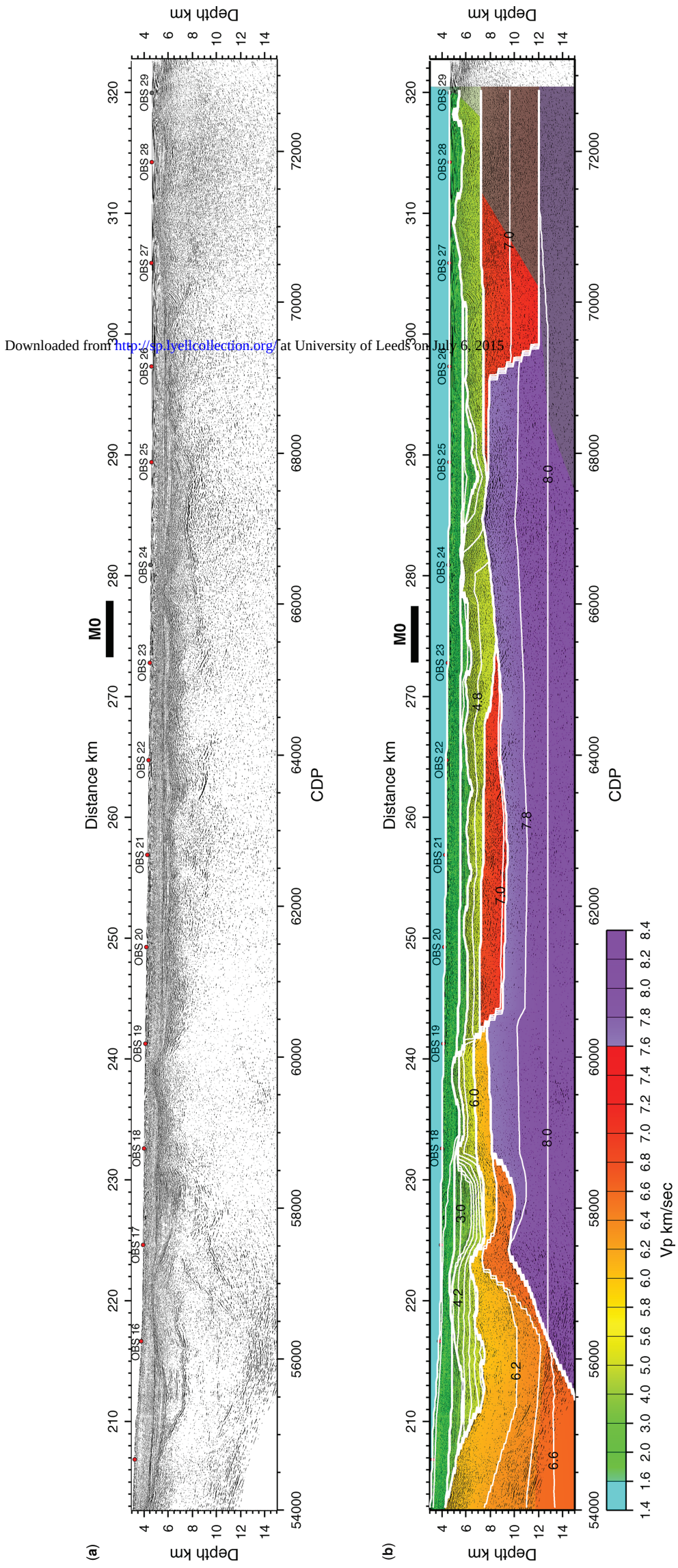


Fig. 5. (a) Pre-stack depth migrated data from Hopper *et al.* (2004) shown from 200 to 320 km along the SCREECH 1 transect. No vertical exaggeration. (b) same as (a) but with seismic velocity model from Figure 4 superimposed. White lines are velocity contours at 0.2 km s^{-1} interval. Greyed area towards the end of the profile indicates no ray coverage in the wide-angle data.

spreading (Pickup *et al.* 1996; Dean *et al.* 2000; Whitmarsh *et al.* 2001). However, seismic refraction surveys along Galicia Bank terminate less than 40 km seaward of the peridotite ridge (Whitmarsh *et al.* 1996), so precisely where magmatically produced oceanic crust begins is not well constrained.

Initial condition and crustal thinning

Flemish Cap itself is 30 km thick and appears to be unaffected by regional extension (Fig. 4). Large, sediment-filled rift basins are absent in the reflection images and the wide-angle data show no evidence for sediment cover (see Funck *et al.* 2003; Hopper *et al.* 2006). The basement has uniform P-wave velocities between 6.0 and 6.7 km s⁻¹, consistent with that typical of Appalachian continental crust. King *et al.* (1985) correlate Flemish Cap to the Avalon Terrane onshore Newfoundland, which has a seismic velocity structure very similar to that measured along the central part of the cap (Fig. 4) (Hall *et al.* 1998; Funck *et al.* 2003). Rock samples from the cap are primarily granodiorites and granites (King *et al.* 1985) and the crust onshore has whole-crust bulk Poisson's ratios of 0.23–0.24, indicating that quartzo-feldspathic rocks dominate the crustal composition (Hughes *et al.* 1994; Hall *et al.* 1998).

One important constraint for modelling exercises is to establish probable compositions for the lower continental crust. In the velocity model of Funck *et al.* (2003), there are no refracted arrivals from the lower-most layer in the crust. The velocity is thus constrained primarily by P_mP phases (Moho reflections), which are clear on most of the Ocean Bottom Seismometers (OBSs) across the Flemish Cap. Despite the inherent ambiguity in determining velocity–depth profiles based only on reflections, lower crustal velocities in excess of 6.75 km s⁻¹ are not supported by the data. This effectively rules out the possibility of significant gabbroic intrusions in the lower crust. At pressures appropriate for 25–30 km depth, gabbro has a P-wave velocity of more than 6.85 km s⁻¹ (Christensen & Mooney 1995). The Flemish Cap data are most consistent with measured velocities for intermediate composition rocks such as diabase (see the compilation in Christensen & Mooney 1995). More mafic rocks such as amphibolites, anorthositic granulites and mafic granulites are also possible, but require temperatures in excess of 650 °C. No heat-flow data exist on Flemish Cap, but such high temperatures seem unreasonable in this setting. Given the similarity of the velocity structure with onshore observations, where Poisson's ratio provides additional constraints, and the difficulty with reconciling mafic compositions with the available data, it

is very likely that Flemish Cap is dominated by relatively weak crustal mineralogies.

Towards the Flemish Pass basin at the NW end of the transect, reduced seismic velocities are observed in the upper crust. Although the shape of this unit is suggestive of a sedimentary basin, the lack of stratigraphic layering in the reflection section argues against its origin as a rift basin associated with Triassic or later rifting, which should be well preserved. In addition, the seismic velocities are much higher than observed in other Mesozoic basins around Newfoundland. Funck *et al.* (2003) suggested that this unit most probably corresponds to Precambrian and Palaeozoic sediments and volcanics associated with the Avalon Terrane of the Appalachians.

The lack of significant basin formation indicates that Flemish Cap is an intact block of continental lithosphere and thus can be considered as a microplate. This is consistent with previous work on reconstructions of the region (Srivastava *et al.* 2000; Sibuet *et al.* 2007). Behaviour as a microplate implies that Flemish Cap must be a strong block of lithosphere that did not deform internally in response to regional extensional stresses responsible for lithospheric rupture. Given the likelihood that the crust is dominated by weaker lithologies, it is inferred that the strong part of the lithosphere must rest primarily in the mantle. This places a qualitative constraint on the initial thermal condition of the lithosphere at this location. In particular, a strong upper mantle indicates reduced temperatures and thermal gradients and implies a cool initial condition (Brace & Kohlstedt 1980). This probably precludes the possibility of significant lower crustal flow in response to crustal buoyancy forces as this requires relatively warm lithosphere even for very weak crustal mineralogies (e.g. Hopper & Buck 1998). Lower crustal flow, it should be noted, is distinct from ductile stretching of the lower and middle crust during extension.

The seaward transition from relatively undeformed, 30 km-thick, continental crust to extremely thinned, less than 5 km-thick crust occurs over a narrow interval of 80 km. Moderately thinned continental crust is marked by a single sedimentary basin at 180 km (Figs 4 & 6). Although the basin is clearly observed in the both the wide-angle and normal incidence data, the reflection image quality is poor due to the incomplete removal of multiple energy. However, weak sedimentary reflections appear to fan seaward, indicating a half-graben bounded by a landward-dipping fault. The crust beneath this basin is 20 km thick, but it thins abruptly seaward to a thickness of less than 5 km over only 40–50 km.

The abruptly necking zone of crust from 150 to 240 km is particularly interesting. Figure 7 shows

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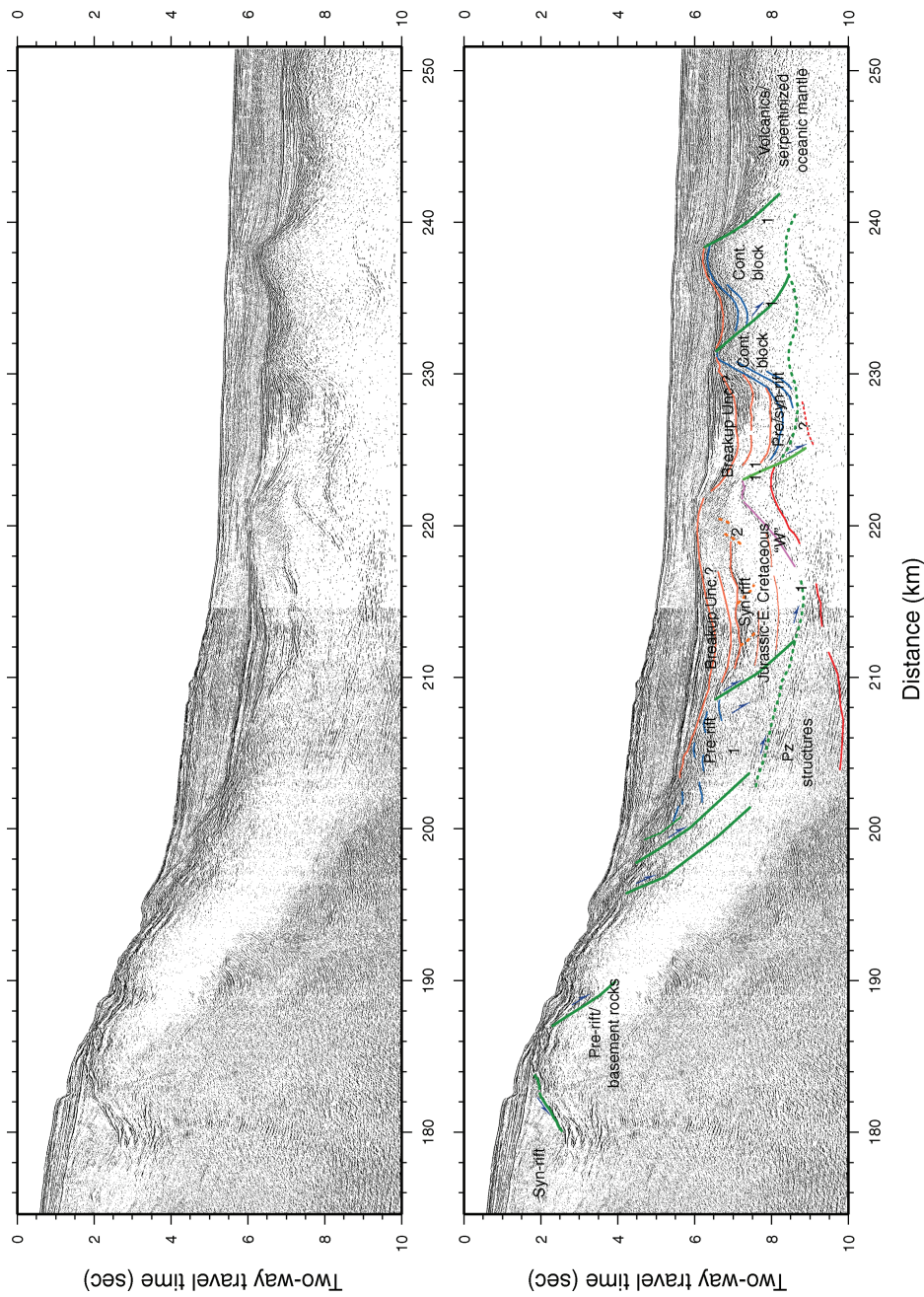


Fig. 6. Seismic reflection section in two-way travel time, where continental crust thins abruptly from depths 170 to 215 km. Top panel is uninterpreted data, bottom panel includes an interpretation as discussed in the text. Seaward of 215 km, the data are time-converted from the pre-stack depth migration from Hopper *et al.* (2004). The full depth migration is also shown in Figure 5.

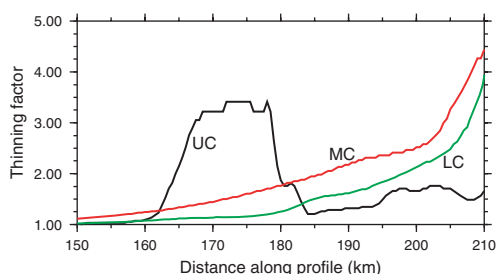


Fig. 7. Thinning factors of different seismically defined layers as a function of distance along the profile. The thinning factor is simply a reference thickness divided by the actual crustal thickness at a location. Reference thickness is taken from 110 km, which is approximately where each layer is thickest across the Flemish Cap on the SCREECH profile. UC, upper crust ($V_p = 6.0$ – 6.2 km s^{-1}); MC, middle crust ($V_p = 6.3$ – 6.4 km s^{-1}); LC, lower crust ($V_p = 6.6$ – 6.7 km s^{-1}). See text for further details.

the thinning factors of different crustal layers as defined by the seismic refraction data. The upper crust has velocities from 6.0 to 6.2 km s^{-1} , middle crust has velocities from 6.3 to 6.4 km s^{-1} , and lower crust has velocities from 6.6 to 6.7 km s^{-1} (Funck *et al.* 2003). From 150 to 210 km , the upper crust shows extensional thinning and strain concentrated in a local area defined by the rift basin at 160 – 180 km . Little thinning of the upper crust is observed elsewhere over this interval. In contrast, the middle and lower crust thin uniformly as is expected for layers that have undergone ductile stretching by bulk pure shear. At least locally, these observations seem to require that deformation of the upper crust was detached from deformation deeper in the crust. The overall amount of extension in the upper crust seems to balance with the amount of extension indicated in the middle and lower crust over this interval despite a small lateral offset in the location where strain was focused. The amount of stretching in the middle and lower crust increases rapidly from 200 km seaward and one or both these layers may pinch out. The details of this pinch-out, however, are difficult to resolve. In addition, the use of seismic velocities to define original upper, middle or lower crust under such extreme thinning is questionable because decompression, porosity changes and mineral alteration from circulating fluids will alter the seismic velocity as deeper material is brought closer to the surface. Thus, we cannot attach significance to layer thickness changes defined solely by seismic velocities beyond 210 km .

In the zone from 210 to 240 km , the middle and lower crust appear to be very thin, and the upper crust thins significantly. The presence of small rift

basins indicates brittle failure of the upper crust, but specific interpretation of the fault pattern is difficult because of the lack of reflections from fault surfaces. One possible interpretation of this section is shown in Figure 6. An important feature of the data is an interval of east-dipping reflections primarily in the lower crust that is interpreted to be Palaeozoic structure formed during the assembly of the Avalon terrane (Hopper *et al.* 2006). These reflections continue through and below reflection and refraction Moho, although the reflection Moho is discontinuous through the cross-over. It is possible that the Palaeozoic structure was reactivated by normal faulting and may have acted as a detachment surface at which overlying high-angle normal faults terminate.

The small basins centred at 216 and 227 km appear to enclose an interval with high velocities of 6.0 – 6.1 km s^{-1} . On the basis of velocity alone, this might be considered to represent upper continental crust. However, the sections show horizontal layering that is unconformable to the basin margins, suggesting a possible sedimentary origin. It is possible that they are limestones and dolomites, which can have velocities of 6 km s^{-1} or more (e.g. Press 1966). In addition, it is known from drilling at ODP site 639 that Upper Jurassic–Lower Cretaceous dolomite is present on the conjugate Galicia margin (Boillot *et al.* 1987). If the interpretation that much of the seismic upper crust is in fact sedimentary, then faulting probably formed the basins early during the second Newfoundland–Iberia rift phase. Faulting during the earlier, Late Triassic–Early Jurassic, rifting is unlikely because there is no indication of evaporites, which probably would be mobile and would distort the sedimentary fill. Deposition of evaporites was widespread in the basins formed on the Grand Banks and Iberia margins during the first rift phase (Tankard & Welsink 1987; Wilson *et al.* 1989).

The origin of the basement highs at 230 – 240 km is uncertain. Hopper *et al.* (2006) present arguments for and against various hypotheses of the section as either continental blocks or as exhumed continental mantle. Figure 6 shows one possible interpretation where the blocks are continental crust bounded by east-dipping normal faults that formed during the early phase of extension that led to the abrupt thinning along the margin. Erosion of the tops of the blocks would explain their rounded tops and apparent coarse debris shed into the adjacent basins. The fault labelled 1.1 would post-date the main faulting associated with these blocks, thus forming the basins centred at 216 and 222 km probably during the Jurassic (see above). The orange faults within the sedimentary fill are related to extension that post-dates the main crustal thinning phase of extension.

Although it is not possible to rule out the possibility that the blocks from 230 to 240 km could be exhumed continental mantle, major large-scale asymmetry between Flemish Cap and Galicia is clearly supported by the data. The abrupt necking profile and narrow region of basin formation is in strong contrast to the conjugate, which shows a broad, 300 km-wide region of attenuated continental crust. Figure 8 shows the profile described here reconstructed to a composite profile of the Galicia Bank and Galicia Interior Basin based on data from Pérez-Gussinyé *et al.* (2003), Reston *et al.* (1996), González *et al.* (1999) and Whitmarsh *et al.* (1996). Importantly, the asymmetry observed is independent of the precise placement of the seaward termination of continental crust along Flemish Cap profile. Moving the boundary 10 km landward and interpreting the blocks as exhumed upper mantle does not fundamentally alter the conclusion of asymmetric margin formation.

The question of symmetry v. asymmetry on conjugate rifted margins has received a great deal of attention (e.g. Lister *et al.* 1986; Dunbar & Sawyer 1989b; Sibuet 1992; Loudon & Chian 1999). Early models suggested that asymmetry requires strong crust–mantle decoupling, whereby crustal strain is displaced from mantle lithospheric strain, eventually resulting in asymmetric conjugate rifted margin pairs (e.g. Braun & Beaumont 1989; Dunbar & Sawyer 1989a). Such a mode of decoupling, however, may be difficult to achieve for realistic geological conditions (Hopper & Buck 1998). Hopper & Buck (1996) proposed instead that asymmetric margins form when wide-rift mode extension breaks along one edge. Both these possibilities, decoupling and wide-rift mode extension, probably require warm lithospheric conditions (Buck 1991), which seems at odds with the arguments presented above for cool, stronger lithosphere.

Recently, a new class of dynamical models of extension have shown that symmetry and asymmetry may instead be controlled by strain-dependent

rheologies (Huisman & Beaumont 2002, 2003; Lavier & Manatschal 2006). In particular, the results of Huisman & Beaumont (2003) suggest that strongly asymmetric rifting occurs when: (1) extension is slow; (2) there is significant strain softening of the dominant rheology; and (3) crust is coupled to the lithospheric mantle. Their conclusions are broadly consistent with data available from the Flemish Cap.

Mantle serpentinization

It has become increasingly clear that penetration of sea water into the upper mantle can result in large-scale serpentinization along divergent plate boundaries. This results in fundamental but still poorly understood changes in the rheology of the upper mantle that probably have a significant impact on mantle and crustal deformation during extension. In addition to having rheological effects, the serpentinization process is strongly exothermic and results in significant volume expansion and density reduction (e.g. Schroeder *et al.* 2002). Thus, it may play an important role in the thermal structure and evolution of divergent plate boundaries as well as having an impact on the dynamic evolution through buoyancy changes.

One of the most important results of research on the Iberia margin is the recognition of the important role that mantle serpentinization plays in the final stages of non-volcanic margin development (e.g. Manatschal 2004). In particular, crustal separation is not always coincident with final lithospheric rupture. Instead, a broad zone of exhumed continental mantle forms and there is an absence of extrusive magmatic rocks and oceanic crust. A key result of the SCREECH Line 1 is that, with the possible exception of the crustal block at 230–240 km, there is no convincing evidence of mantle exhumation occurring in a similar environment. In previous work, Funck *et al.* (2003) and Hopper *et al.* (2004,

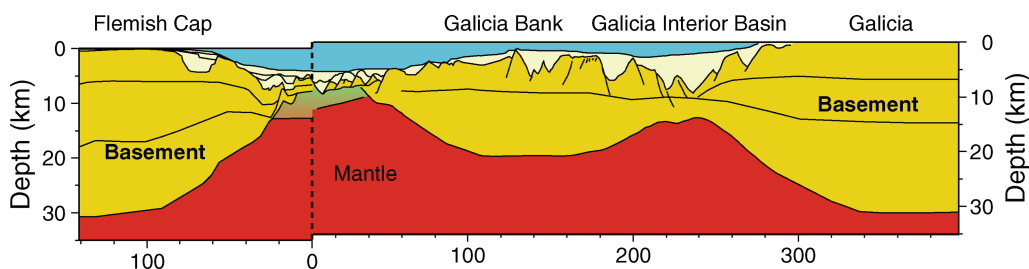


Fig. 8. Plot of Flemish Cap data summary from this work reconstructed to conjugate data sets. Data were spliced together at the seaward-most point where continental crust can be identified in each case and the data shifted so that the sea floor is at the same depth. Conjugate data are based on profiles published in Reston *et al.* (1996), Whitmarsh *et al.* (1996), González *et al.* (1999) and Pérez-Gussinyé *et al.* (2003).

2006) presented a number of arguments demonstrating that some type of oceanic crust exists seaward of 240 km. OBS 19, just on the eastern side of the basement blocks at 230–240 km, shows clear evidence for a rapid seaward velocity change that is best explained by anomalously thin oceanic crust overlying serpentinized mantle beginning at this location (Figs 4 & 5) (Funck *et al.* 2003). Thus, it appears that a combination of extrusive magmatism and mantle exhumation characterized extension shortly after continental crustal separation along this transect.

Nevertheless, mantle serpentinization is clearly indicated along the profile, and it must have played an important role in the late stages of margin development that led to full lithospheric rupture. Well-defined and easily identifiable upper mantle P_n phases from beneath the thin magmatic crust along the profile consistently show phase velocities of 7.6 km s^{-1} . This is significantly less than the 8 km s^{-1} measured below Flemish Cap (Funck *et al.* 2003), and it suggests a maximum of approximately 15–20% serpentinization of olivine peridotite immediately beneath the thin crust (100–150 Mpa; e.g. Christensen 2004). Amplitude modelling further constrains the velocity gradient in the mantle and shows that velocities of 8 km s^{-1} are reached by about 5 km below the base of the crust.

The velocity data provide convincing evidence that serpentinization becomes important when the crust thins to less than 6–8 km thick. Figure 9 shows the mantle P-wave velocity just below Moho as a function of crustal thickness along the SCREECH Line 1. Assuming that reduced mantle velocities indicate the onset of serpentinization, this process appears to initiate at a crustal thickness of approximately 8 km. Simple thermal models of lithospheric extension show that this is about the point at which the entire crust becomes brittle (Pérez-Gussinyé & Reston 2001). In the absence of significant melt production, extension can then only be accommodated by brittle failure, which

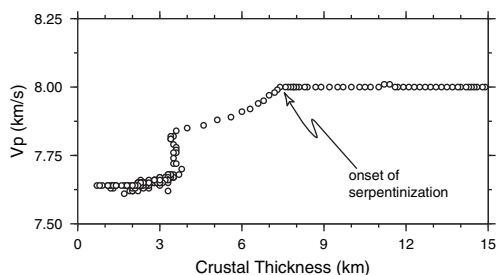


Fig. 9. Mantle P-wave velocity just below the Moho as a function of thickness of the overlying crust. See the text for explanation.

provides pathways for water to penetrate into the mantle and to facilitate serpentinization. This idea is generally consistent with the interpretation of the seismic reflection section of SCREECH transect, given that major faults bounding the basins at 210–230 km most probably cut deep into the crust and possibly into the upper mantle. These faults could provide the conduits necessary for fluids to circulate deep into the lithosphere.

Magma production and sea-floor spreading

The last important observation the SCREECH data provide is information on the onset of melting and the generation of magmatically produced oceanic crust, which marks final lithospheric rupture. Historically, amagmatic margins have been studied to understand brittle and mechanical deformation of the crust. The absence of melt in the system has important consequences. From an observational point of view, extensional structures that result from thinning and stretching are easier to image and sample as they are not covered up by volcanic rocks. From a modelling point of view, extensional stress cannot be relieved by magmatic intrusion and dyke injection so only mechanical deformation of the lithosphere needs to be considered up to the time that volumetrically significant melt is produced.

The SCREECH Line 1 results are fully consistent with an absence of significant decompression melting prior to breakup of the continental crust, despite extreme stretching and thinning. Across Flemish Cap and along the abruptly thinning crust farther seaward, seismic velocities measured in the lower crust are low, consistent with an absence of underplating or gabbroic intrusion into lower crust. Crustal velocities of more than 6.8 km s^{-1} are not observed anywhere in the section until 240 km, where thin magmatic crust is interpreted. In addition, the reflection data show little evidence for significant sill intrusion into the few sedimentary basins that are observed. To the extent that melts may have formed and infiltrated the upper mantle, there was insufficient production to generate extrusive flows or geophysically resolvable intrusive layers within the crust.

Melting and magmatism, when it did appear, produced anomalously thin, 3–4 km thick, oceanic crust, indicating a magma-starved crustal accretion system. Such crust is observed in modern settings along ultra-slow spreading ridges like the Gakkel Ridge where crust is only 4 km thick (Coakley & Cochran 1998; Jokat *et al.* 2003). This is further consistent with Srivastava's estimate that extension at the time of opening was approximately 14 mm

year⁻¹ (full rate), which is also comparable to Gakkel Ridge spreading rates. Crust with closer to average oceanic thickness did not appear until about 295 km on SCREECH Line 1, some 20–25 km seaward of anomaly M0 as picked by Srivastava *et al.* (2000) (Fig. 5). Unfortunately, velocities and crustal thickness on this part of the refraction line are not well constrained (Funck *et al.* 2003), and the line does not extend far enough to confirm that the crust continues seawards with normal oceanic thicknesses.

While it is tempting to conclude that extension rate was the controlling factor in suppressing melting in the Iberia–Newfoundland rift, it is not clear that this is the only important factor. Reston & Phipps-Morgan (2004) proposed that anomalously cool upper mantle may be important in limiting or suppressing melt production. A compilation of continental geotherms by Röhm *et al.* (2000) shows that the potential temperature of the upper mantle may have a natural variation of several hundred degrees, and the possibility that the Newfoundland Iberia rift was underlain by an asthenospheric cool spot is one that must be considered.

It is clear that large variations in melt supply are indicated once magmatism was established in the rift. At 273–290 km, oceanic layer 3 is absent, and the crust thins to as little as 1.5 km thick in part of this zone. Hopper *et al.* (2004) attributed this to a cessation of magmatism that resulted in the formation of an oceanic core complex in which the lower crust was mechanically removed. They further proposed that magmatism was episodically more robust, in order to explain apparent volcanic stratigraphy that is observed in the upper crust. Such volcanic stratigraphy is also observed on data 10 km to the north of Line 1 (Hopper *et al.* 2006). This kind of variation is not unique to the Flemish Cap–Galicia transect. There are significant disparities in the amount of melt introduced throughout the rift following breakup of the continental crust, although a common denominator appears to be that all areas exhibit significant exhumation and serpentinization of mantle (see Tucholke *et al.* 2007). Chemical heterogeneities may have also played a key role in the temporal and spatial variability of magmatism. In a magma-starved rift, small pockets of anomalously fertile mantle could have a large effect on local areas. For example, passively imbedded mantle inhomogeneities have been proposed to explain isolated magmatic centres along the ultra-slow Gakkel Ridge (Cochran *et al.* 2003).

The available data indicate that large spatial and temporal variations in melt supply may be important for understanding the detailed structure of non-volcanic margins during the extensional phase following complete separation of continental crust

and prior to the establishment of a robust sea-floor spreading system. A complex interplay of brittle faulting and mantle exhumation, melting, and intrusion and extrusion of melt controls the detailed structure of this crust. Geophysical data alone cannot constrain these processes and further progress in this area will require basement sampling and geochemical analyses.

Conclusions

The seismic reflection and refraction data from the Flemish Cap margin, together with seismic surveys over the conjugate Galicia Bank margin and Galicia Interior Basin, provide first-order constraints for numerical modelling of lithospheric rupture in a non-volcanic rift. In particular, the data show important differences in margin development and document asymmetries on a variety of scales.

Well-defined crustal and mantle arrivals provide detailed constraints on the crustal thickness across the entire transect. Flemish Cap itself has crustal seismic velocities consistent with velocities measured in Appalachian crust from the Avalon terrane throughout the Newfoundland region. Such crust is probably dominated by quartzo-feldspathic lithologies (Hughes *et al.* 1994). There is a conspicuous absence of significant rift basins in the continental block comprising Flemish Cap. Rift basins appear on the eastern margin of the block, but they are small compared to other rift basins on the western side of the block and within the Grand Banks. The continental crust thins rapidly on the eastern margin of Flemish Cap, from 30 to approximately 3 km thick, over a distance of about 80 km.

Mantle velocities beneath Flemish Cap are 8.0 km⁻¹ and thus do not indicate anything unusual about the composition of the upper mantle. Reduced seismic velocities are observed under thinned continental crust farther seaward, beginning where the crust is approximately 8 km thick. Reflection images suggest that the embrittlement of the entire crust may have occurred at this point, providing pathways for fluids to penetrate into the mantle and facilitate serpentinization. However, with the possible exception of one 10 km-wide basement block at the edge of the thinned continental crust, there is no evidence that mantle was extensively exhumed as is observed off the Iberia margin. Nevertheless, serpentinization of the upper 5 km of mantle is indicated by anomalously low seismic velocities. Serpentinization is an exothermic reaction that results in a significant volume increase and density reduction. In addition, the rheology of serpentinite is considerably different than peridotite. Thus, there clearly are important, but as yet unexplored, implications of this process

for understanding the dynamic evolution of rifted margins during the final stages of lithospheric rupture.

We do not observe high seismic velocities in the lower crust that would be expected from underplating or gabbroic intrusion as a result of decompression melting during rifting of the continental crust. There is no evidence for melt production until after the breakup of the continental crust, at which point thin (3–4 km) magmatic crust with oceanic velocity structure was produced. This crust shows a strong variation in thickness and it consistently overlies serpentinized mantle several kilometres thick, indicating a complex phase of mantle extension with variable magmatism before a more magmatically robust sea-floor spreading system was established.

The timing of rifting events on the Flemish Cap margin is uncertain because of a lack of drilling and sampling. It seems likely that the first phase of Newfoundland–Iberia extension (Late Triassic–Early Jurassic) did not significantly affect the crust between Flemish Cap and Galicia Bank. The oldest sediments drilled on Galicia Bank are shallow-water Tithonian carbonates and mudstones, which suggests that rifting of the crust there began during the second, Late Jurassic–Early Cretaceous, phase of extension. Possible high-velocity sediments in the small rift basins at the seaward edge of thinned Flemish Cap crust may be the counterparts of these Tithonian carbonates at the western edge of the rift. If it is assumed that rifting of the eastern Flemish Cap margin began in the Tithonian, then continental crustal extension followed by the formation of thin magmatic crust combined with mantle extension and exhumation occupied an interval of at least approximately 25 million years until apparently normal ocean crust was formed seaward of anomaly M0.

There are several parameters critical for numerical modelling that the presently available data do not constrain very well. There is no information on the lithospheric thickness of the region, and we can only speculate on the thermal structure of the lithosphere prior to rupture. However, the data are consistent with Flemish Cap behaving as a microplate, implying that is a relatively strong block of lithosphere. Thus, a relatively cool geotherm and thick lithospheric mantle is indicated.

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