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Structure and development of the southeast Newfoundland continental passive margin: derived from SCREECH Transect 3

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SUMMARY

New seismic reflection data from the Grand Banks of Newfoundland and the Newfoundland Basin add to the growing knowledge of the composition, structure and history of this non-volcanic margin. Geophysical imaging is now approaching the extent of that done previously on the conjugate margin along Iberia, providing a valuable database for the development of rifting models. Two parallel profiles over the shelf platform image deep crustal fabric representing Precambrian or possibly Appalachian deformation as well as Mesozoic extension. Progressively more intense extension of continental crust is imaged oceanwards without the highly reflective detachments frequently seen on profiles off Galicia. A landward-dipping event 'L' is imaged sporadically and appears to be analogous to a similar event on the Iberian IAM9 profile. The transition zone is probably exposed serpentinized mantle as interpreted off the Iberian margin although there appears to be a difference in the character of ridge development and reflectivity. The distinctive 'U' reflection identified previously at the base of the Newfoundland Basin deep water sedimentary section and recently identified as one or more thin basalt sills is imaged on newly presented profiles that connect previously published profiles SCR3 and SCR2 showing that 'U' is highly regular and continuous except where interrupted by basement highs. 'U' is also seen to have a major impact on the ability to image underlying basement. A full transect beginning over completely unextended continental crust through to oceanic crust has provided a data set from which estimates of extension and the pre-rifting location of the present continental edge can be made. Two estimates were obtained; 85 km based on faulting and 120 km based on crustal thickness.

Key words: Crustal structure; Continental margins: divergent; Submarine tectonics and Volcanism; Atlantic Ocean.

1 INTRODUCTION

The southeast Newfoundland continental margin, in the North Atlantic Ocean, is conjugate (Srivastava *et al.* 1990) to the Iberian margin (Fig. 1) with the Flemish Cap conjugate to the Galicia Bank and the northern Grand Banks conjugate to the southern Iberia Abyssal Plain where it approaches the coast of Iberia south of the Galicia Bank (Fig. 1). The Newfoundland-Iberia conjugate margin pair has been an attractive target for seismic imaging because it is almost devoid of extrusive volcanic rocks and was relatively starved of sediment during subsidence allowing for more effective seismic imaging of the deep crust. The Iberia margin has been studied in detail over many years now, but data need to be collected over both sides of a conjugate margin pair to develop a valid rifting model because there is evidence that the breakup processes may not happen symmetrically on both sides (Louden & Chian 1999). As part of the exploration prior to the 2003 drilling of Ocean Drilling Program (ODP) Leg 210 on the Newfoundland margin, regional deep seismic profiles (SCREECH: Study of Continental Rifting and Extension on the Eastern Canadian Shelf) were collected along and between three major transects across the margin, located to match along-margin variations on the more intensively studied Iberian margin (Boillot *et al.* 1989; Whitmarsh & Sawyer 1996). The central SCREECH profile (SCR2; Shillington *et al.* 2006; Van Avendonk *et al.* 2006) was positioned to be collinear, pre-sea floor spreading, with boreholes from Deep Sea Drilling Program (DSDP) Leg 47B and Ocean Drilling Program (ODP) Legs 149 and 173. Transects to north (SCR1; Funck *et al.* 2003; Hopper *et al.* 2004, 2006)

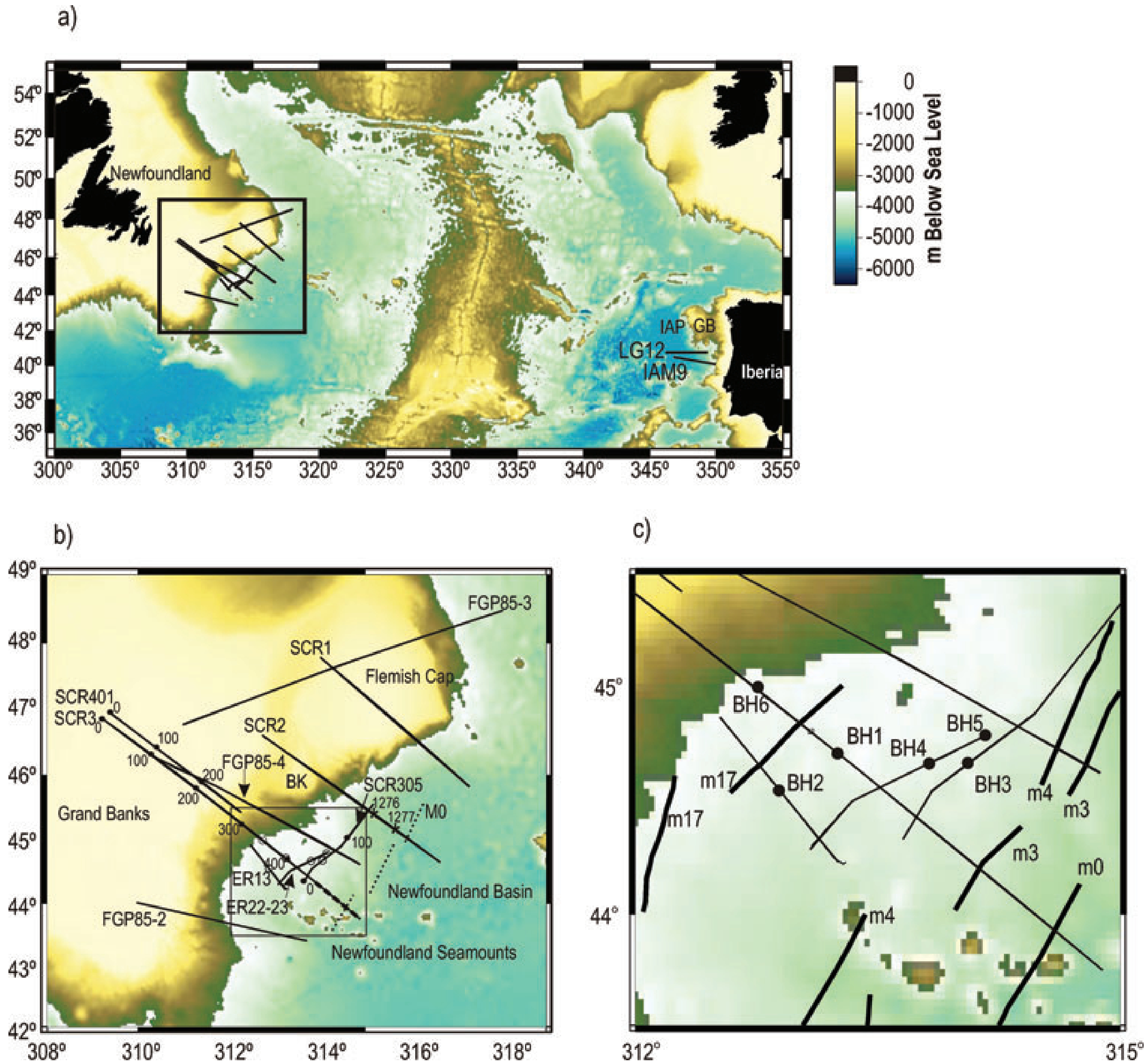

Figure 1. Location maps of seismic profiles referred to in the text with bathymetric data from Smith and Sandwell (1997). Elevations above sea level are indicated in black. (a) Bathymetry of the North Atlantic Ocean with profiles located over the Grand Banks and Newfoundland Basin shown relative to the conjugate IAM9 profile south of Galicia Bank. (b) Bathymetry of the Grand Banks and Newfoundland Basin (inset part a) with regional seismic profiles reproduced in Figs 2–5 or referred to in the text. The dashed end of SCR3 is not reproduced in this paper. Circles show the location of basement highs referred to in the text and labelled in part (c) and Figs 3 and 5. Stars mark the location of ODP Leg 210 borehole sites 1276 and 1277. The change from brown to green shading of bathymetry roughly marks the change in seabed dip from the steeper slope environment to the shallower dip of the rise. The dotted line marks the location of M0 (Srivastava *et al.* 1990). (c) Inset from (b) with basement highs (BH) labelled as described in the text and further m-series anomalies. SCR: SCREECH, FGP: Frontier Geoscience Project, ER: Erable, LG12: Lusigal 12 profile, GB: Galicia Bank, BK: Beothuk Knoll, IAP: Iberia Abyssal Plain, BH1–6: basement highs described in the text.

and to south (SCR3; Lau *et al.* 2006a,b) provide information about along-margin variations.

In this paper, we present new data on the shelf of the Grand Banks in the form of two parallel closely spaced (∼20 km) deep crustal seismic reflection profiles, one of which is continuous with SCR3 published by Lau *et al.* (2006b). In addition, the offshore portion of the profile has now been pre-stack depth migrated. New analysis of SCR3 includes estimates of extension throughout the extended continental crust domain using measurements of both faulting and crustal thickness from the seismic data. Previously unpublished images of the transition zone are included from the Erable surveys (described below) that show details of basement high structures and the ‘U’ horizon in the vicinity of SCR3. A previously unpublished SCREECH profile (SCR305) that ties the transition zone of SCR2 (Shillington *et al.* 2006) with SCR3 is also presented here to show the regional variation in character of ‘U’ parallel to the rift. SCR3 is compared here to IAM9, the most southerly regional seismic profile in the Iberian Abyssal Plain, because it is the closest to a conjugate location although it is displaced significantly to the north (∼100 km).

2 REGIONAL GEOLOGY OF THE NEWFOUNDLAND MARGIN

The pre-Mesozoic continental basement of the northern Newfoundland margin (Fig. 1) is composed of Neoproterozoic sedimentary

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and volcanic rocks, with occasional thin outliers of Palaeozoic sediments. It is part of the Avalon Zone of tectonostratigraphic associations spatially linked with the Cadomian basement in Iberia at the end of the Cadomian orogenic cycle (McNamara *et al.* 2001). The late Proterozoic tectonism was succeeded by the Appalachian/Caledonide events associated with the opening and closure of the Iapetus Ocean (Williams 1979). The Avalon Zone was peripheral to Appalachian orogenic activity, especially farther east on the continental shelf of the Grand Banks, but Precambrian–Cambrian transcurrent ductile shear zones are common in the Avalon Terrane and have had a major influence in Phanerozoic development of the area (Gibbons 1990). Silurian to Carboniferous successor basins, often in narrow zones associated with post-collisional strike-slip faulting, occur in Newfoundland and continue offshore.

The post-Appalachian geologic history has been controlled by the progressive rifting of Pangea and opening of the Atlantic Ocean (Tankard & Welsink 1988). Breakup in the north Atlantic first occurred between Nova Scotia and NW Africa at around 150 Ma (mid-Jurassic). It then progressed northwards: separating the Grand Banks and Iberia at about 115 Ma (lower Cretaceous) and the Flemish Cap and Galicia at about 100 Ma (mid-Cretaceous) (Tankard & Welsink 1987; Srivastava & Verhoef 1988).

3 GEOPHYSICAL INVESTIGATIONS ALONG THE MARGIN

3.1 Previous work

Four ODP Legs (47B, 103, 149, 173; Sibuet *et al.* 1979; Shipboard Scientific Party 1987, 1994, 1998), numerous other geophysical surveys (Krawczyk *et al.* 1996; Dean *et al.* 2000; Whitmarsh *et al.* 2000; Henning *et al.* 2004) and dredge samples (Boillot *et al.* 1980) have provided information to help understand the nature and evolution of the existing crust for hundreds of kilometres along the offshore Iberian coast. Much of the work offshore Iberia relevant to this paper has been designed to characterize two distinct zones: the northern part of the margin along Galicia Bank, and the central margin adjacent to the south Iberia Abyssal Plain.

Key crustal domains in each area include block-faulted, thinned, continental crust, a transitional crust of serpentinized mantle with a serpentinite ridge system and oceanic crust. Along the western margin of the Galicia Bank, block-faulted continental crust occurs above bright reflectors (Krawczyk *et al.* 1996; Whitmarsh *et al.* 2000; Henning *et al.* 2004) usually identified as detachment faults that separate crustal material and mantle (Reston 1996; Reston *et al.* 1996). In the Iberia Abyssal Plain, just south of Galicia Bank, seismic profile IAM9 does not show strong intracrustal reflectors that would represent extensive detachment faulting (Pickup *et al.* 1996; Dean *et al.* 2000). Exposed serpentinized mantle off the Iberian coast was first identified by collecting dredge samples from a seabottom ridge west of the Galicia Bank (Boillot *et al.* 1980). The topographic high was further mapped by seismic profiling as a north–south trending basement ridge along the Galicia Bank (GB) and as a segmented en echelon ridge to the south in the Iberia Abyssal Plain (IAP) broadly collinear with the ridge to the north and part of a broader (100 km) zone of low relief serpentinized mantle (Boillot & Winterer 1988; Beslier *et al.* 1990, 1993, 1996; Whitmarsh *et al.* 2001; Dean *et al.* 2000; Chian *et al.* 1999). Serpentinization is recognized through seismic refraction surveys (Whitmarsh & Sawyer 1996; Chian *et al.* 1999) to extend beneath thin oceanic crust to the west as well as, though to a lesser extent, beneath the thinned continental crust to the east. Drilling on topographic highs sampled serpentinized peridotite with continental geochemical characteristics (Whitmarsh & Wallace 2001) and subsequently this section has come to be known as the 'zone of exhumed continental mantle' (ZECM; Whitmarsh *et al.* 2001). Compared with the wide zone of crustal thinning across the Galicia Bank and narrow transition to oceanic crust seaward of Galicia Bank, the Iberian Abyssal Plain has a narrow zone of thinned continental crust, with a wide transition to oceanic crust (Pickup *et al.* 1996). The distance from the present-day Iberian shelf edge to oceanic crust is similar in both areas.

Whitmarsh *et al.* (2001) and Boillot *et al.* (1989) summarize present understanding of the evolution of the Iberian margin as follows. Rifting of continental crust produced a series of thinned block faults with thinning of the upper mantle prior to breakup, followed by a late-developing concave-downward detachment that accommodated continental crust breakup and exposure of mantle at the seafloor. Exhumation of mantle continued in the low-magma rifting environment until seafloor spreading and oceanic crust formation began. Low-amplitude magnetic lineations in the transition zone have developed as a result of gradual onset of magmatism (Whitmarsh *et al.* 2001; Russell & Whitmarsh 2003) leading to seafloor spreading or as part of the serpentinization process (Sibuet *et al.* 2007a,b).

3.2 Recent findings

Significant discoveries from the drilling of Leg 210 sites 1276 and 1277 on the Newfoundland margin include (i) the presence of post-rift sills at the level of the strong 'U' reflector without strong sedimentary acoustic impedance contrasts indicating that 'U' is a result of sills, and (ii) a lack of oceanic crust in the highly serpentinized basement high just seaward of M1 suggesting that oceanic crust developed at the time of M0 or later (Tucholke *et al.* 2004). Rifting is now being referenced in terms of 'crustal' and 'lithospheric' separation (Tucholke *et al.* 2007) to account for the broad zone of exhumed mantle now interpreted between continental and oceanic crust (Tucholke & Sibuet 2007). The formation of 'thin' oceanic crust over serpentinized mantle (Hopper *et al.* 2004) has recently been recognized as an intermediate stage between extension by exhumation of subcontinental mantle and fully developed oceanic crust.

3.3 The SCREECH survey

In the summer of 2000, three parallel transects orientated perpendicularly to the SE Newfoundland continental margin (Fig. 1) were recorded with multichannel and wide-aperture seismic techniques, magnetics and gravity. Other multichannel seismic profiles were also recorded to link the main transects and to locate drill sites for ODP Leg 210 (Shillington *et al.* 2004). The multichannel data were recorded on a 6000 m, 480-channel streamer with a group spacing of 12.5 m. The source was a 20-element, 140 l airgun array shot at a spacing of 50 m and yielding a common midpoint (cdp) spacing of 6.25 m with a nominal fold of 60. A 16 s recording time allowed for recording possible arrivals from the upper mantle. A 4 ms sample rate was sufficient for recording the signal that was dominantly below 90 Hz.

Each of the main reflection profiles (SCR 1, 2 and 3) extends from full thickness continental crust on the shelf to the west through to oceanic crust to the east. The continental portion of the northern profile, SCR1, crossed the Flemish Cap (Funck *et al.* 2003; Hopper *et al.* 2004, 2006). SCR2 crossed the Flemish Pass Basin, which separates Flemish Cap from the Grand Banks, and Beothuk Knoll at its western end before crossing the transitional zone and ending on oceanic crust to the east. SCR3 (Lau *et al.* 2006a,b) is presented in full, for the first time, in this paper. Over the Grand Banks, SCR 401 (Fig. 1; Solvason 2006) was recorded with the same experimental parameters just 20 km to the north of, and parallel to, SCR3. The previously unpublished connecting profile between SCR2 and SCR3 over transitional crust (SCR 305; Fig. 1; Eustace 2002) is also presented here.

3.4 The Erable survey

In 1992, a joint cruise of the Geological Survey of Canada (GSC) and l'Institut français de recherche pour l'exploitation de la mer (IFREMER) was carried out in the Newfoundland Basin and around the Flemish Cap to collect geophysical data in preparation of early proposals for ODP Leg 210. Two of the Erable profiles relevant to this discussion are presented in this paper, ER13 (Fig. 1; Quigley 2005) and ER22-23, (Fig. 1). The profiles were processed at Memorial University of Newfoundland from archived field tapes obtained from the Lithoprobe database. The field parameters for the seismic reflection profiles presented in this paper were a source array of 6 Sodera airguns, 96-channel streamer with a 50 m shot spacing, 25 m group spacing and a maximum offset of nearly 2675 m. The recording geometry produced a cdp spacing of 12.5 m and a nominal fold of 24 with a record length of 12 s.

3.5 The Frontier Geoscience Project surveys

The Geological Survey of Canada collected deep regional marine seismic reflection and refraction data, offshore Atlantic Canada as part of an initiative called the Frontier Geoscience Project (FGP, http://gsca.nrcan.gc.ca/pubprod/FGP/index_e.php, Keen *et al.* 1987) during 1984–1990. The seismic reflection data were recorded and processed by Geophysical Services Inc.. Reflection data were usually recorded on a 120-channel streamer with receiver group spacing at 25 m and shot spacing at 50 m resulting in a far offset of about 3200 m and 30-fold cdp gathers with a 12.5 m spacing. Maximum recording time was at least 15 s. Profile FGP85-4 (Fig. 1) is reproduced here for comparison purposes.

3.6 Generalized processing description

Processing sequences for the different data sets varied as a result of (i) environmental factors such as water depth, sediment thickness above basement and water bottom topography and (ii) the recording parameters, particularly the configuration of the airgun array. All data sets required a low-cut bandpass filter to remove typical streamer noise. The Erable data also benefitted from a spiking deconvolution early in the processing sequence. Handpicked mutes to remove direct waves in shallow water areas and residual far-trace normal moveout distortion, as well as near-trace mutes for multiple removal in deep water, were usually necessary.

In the shelf environment, short-gap deconvolution and amplitude control were crucial for removing persistent reverberations. The f-k removal of steep dipping linear noise from shallow sources was vital because the water-borne energy is of high amplitude relative to the basement reflections (Solvason 2006).

On data recorded in deeper water, other techniques such as radon filtering and f-k multiple removal were used to reduce high-amplitude water bottom multiples. The slope environment on SCR3 also had highly reflective sediment interfaces as well as a strongly reflecting basement surface that created pegleg reflections that were very difficult to remove. Water bottom multiples recorded in the slope environment were generally aliased at farther offsets and so required muting or removal by variable offset bandpass filtering if the f-k or radon filter did not remove the multiple events sufficiently.

SCR3 was pre-stack depth migrated in the slope and deep water parts of the line: this had not been attempted in the previous publication of these data (Lau *et al.* 2006b). All other profiles were migrated with Kirchhoff post-stack time migration or Stolt F-K migration (SCR305; Eustace 2002). After time migration, the data were depth converted for interpretation and display using a simplified velocity model taken from the pre-stack depth migration. The velocity model for all depth-converted profiles here in the Newfoundland Basin used 1480 m s^{-1} for the water layer, 2000 m s^{-1} for uppermost sediments down to 6.0 s, 2400 m s^{-1} from 6.0 to 7.0 s, 2900 m s^{-1} from 7.0 s to basement surface and 6000 m s^{-1} below the top of basement. The 6000 m s^{-1} is high for what is thought to be serpentinized mantle in most places but the pre-stack depth migration was unconstrained in basement because of a lack of reflectors. The lack of reflectors makes the basement velocity somewhat immaterial for our purposes here. The velocity structure obtained in the pre-stack depth migration is very near to the laboratory-measured horizontal velocities from the borehole core samples. It is not surprising that the migration velocities were higher than the measured vertical velocities because measurements in the laboratory at atmospheric pressure are expected to produce lower velocities than *in situ* measurements. The velocities resulting from the pre-stack depth migration are somewhat lower than the velocity model developed from wide-aperture measurements (Lau *et al.* 2006a).

The shelf portions of SCR3 and SCR401 required semblance filtering to produce a low signal-to-noise ratio and are plotted with a 1:1 vertical-to-horizontal scale. Offshore, the top basement events became very difficult to see, so data were plotted at a 2:1 vertical exaggeration and semblance filtering was not used because the signal-to-noise ratio was higher.

4 DESCRIPTION OF THE NEWFOUNDLAND MARGIN FROM SEISMIC DATA

4.1 Introduction

The seismic profiles are described here in terms of four geographic domains, each reflecting a history of distinctly different large-scale geologic processes. The naming conventions follow that established in Lau *et al.* (2006b) with the addition of the shelf section. The westernmost domain is the continental shelf of the Grand Banks. In this domain, the continental crust is of full thickness, or moderately thinned below major inboard basins such as the Jeanne d'Arc. Basement consists of Avalon Terrane assemblages and Triassic extension is confined to the east, leaving the Avalon Platform of the western Grand Banks unaffected by post-Appalachian extension. East of the shelf, the second domain consists of the slope and rise where water depths increase, continental crust decreases in thickness and Triassic extensional faulting is pervasive. The third domain, farther east again, is a section with a seismically non-descript basement with little internal structure or topographic relief imaged. Furthest to the east, the fourth domain is where basement relief is prominent and is characteristic of oceanic crust.

4.2 Domain 1: continental shelf

The continental shelf of the Grand Banks is imaged in parallel profiles SCR3 and SCR401 (Figs 1 and 2). Pre-Mesozoic shallow structures are not imaged in the seismic data, partly due to the overwhelming energy of water column reverberations relative to the weaker reflections from within the crystalline basement and the inherently low fold of shallow data (Hurich 1991). From middle to lower crustal levels, the seismic data image large-scale fabric.

As might be expected in lines so closely spaced, SCR3 and SCR401 are very similar and provide a further legitimacy to the identification of what are often indistinct and low-amplitude events as primary reflected energy. The seismic reflection fabric in basement (Fig. 2) is dominantly of two orientations, a subhorizontal fabric concentrated in parallel zones in the lower crust and a crosscutting dipping fabric. The subhorizontal events seem to occur at two distinct levels, a shallower set approximately 20–30 km deep ('A' events, Fig. 2) and a deeper set about 35–42 km deep ('B' events, Fig. 2). The base of the B events coincides with Moho as determined from the coincident wide-angle seismic profile (Lau *et al.* 2006a) as is often the case in continental crust (Klemperer *et al.* 1986). The crosscutting events are less confined in depth and occur as events that are spatially correlated with the extension creating the Jeanne d'Arc Basin ('D' events, Fig. 2, see below) and other packages that are not clearly related to any observed surface structure ('C' events, Fig. 2, see below). These other dipping reflections occur from the mid-crust down as deep as 50 km (well within the mantle), and though well imaged on SCR3, 'C' events are faint on SCR401. These mid- to lower crustal and mantle events are recorded throughout the Grand Banks up to the edge of the continental slope where deep crustal energy seems to be overwhelmed by noise problems associated with shallow geological complexities.

West of the Jeanne d'Arc Basin, crustal reflectivity is more clearly imaged in profile SCR3 than in SCR 401 but the patterns imaged are similar. The reflected signal occurs as packages or even more subtle 'patches' of short events rather than as continuous, laterally extensive individual reflections. Mid- and deep crustal reflective zones A1 and B1 are separated by a relatively transparent zone beginning west of the Jeanne d'Arc Basin and continuing to the west until cross-cut by west-dipping events, C1 and C2 (SCR3). West of C1, A2 and B2 are identified as being thinner zones of reflectivity than A1 and B1 and A2 occurs significantly deeper than A1. The base of B2 is similar in position to the base of B1 although B2 is weaker. Wide-aperture data (Lau *et al.* 2006a) also suggest that the base of the crust is flat throughout the area west of the Jeanne d'Arc Basin. At the extreme western end of both SCR3 and SCR401, a mid-crustal package of events, A3, is imaged although on SCR401, the internal dips are both east and west directed whereas on SCR3 the individual events within the package are horizontal. At the eastern end of the shelf section, before the shelf break, A4 and B4 appear as moderately west-dipping distinct packages, the dip of which appears to reflect the easterly thinning of the crust associated with Mesozoic extension.

Consistently dipping reflections in deep continental crust are usually interpreted as shear zones (Smithson & Johnson 1989). The lack of significant offset of the Moho across the C1, C2, C3 and C4 reflection packages indicates that the offset has relaxed isostatically over time, or been reversed by inversion. Profiles FGP85-3 and FGP85-4, also over the Grand Banks, have lower crustal events similar to those described for SCR3 and SCR401 (Keen *et al.* 1987). Other seismic profiles recorded nearby over the Appalachian orogenic belt in and around Newfoundland and Labrador also image complicated crosscutting events at deep crustal to Moho depths consistently dipping towards the continent and the axis of the Appalachian orogen (Hall *et al.* 1998, 2002). Hall *et al.* (1998) describe in detail the reflectivity relationships in the Newfoundland Appalachian crust and conclude that most of the deep crustal reflectivity (including dipping events crosscutting the Moho) was created during Silurian collision although some modification occurred during post-collisional relaxation. The C events show no correlation with Mesozoic structure. We interpret them as having an earlier origin, possibly related to the late-Precambrian assembly of the Avalon Terrane, or to Appalachian deformation though this markedly decreases eastwards in Newfoundland (Williams 1979) and so would be expected to be quite modest on the Grand Banks.

The Murre fault is the major eastward-dipping normal fault that bounds the Jeanne d'Arc half graben (Tankard & Welsink 1987) formed during Mesozoic extension. The 'D' events on Fig. 2 are approximately collinear with the Murre fault and so are interpreted to arise from intrabasement shear zones that represent the Murre fault at depth. On SCR401, D1 and D2 events are well defined and laterally continuous and, although they terminate at a depth near the mid-crustal reflectivity, they are strongly dipping suggesting that the fault may sole out into the uppermost lower crust as concluded by Dentith and Hall (1990) from a cross-sectional restoration based on FGP85-4 (Keen *et al.* 1987). On profile SCR3, D1 is a diffuse pattern of low-amplitude east-dipping reflectivity and D2 is a package of concave-upward events that suggests that extensional deformation soles out in the mid-crust. Events D3 and D4 on profile SCR3 do not appear on SCR401 and are not easy to interpret. They may be out-of-plane events from shallow structural complexities in the Jeanne d'Arc basin or they may come from another Murre-related fault that is very steep and offsets the Moho as inferred from the dip and possible offset of west-dipping events at the base of the crust labelled B3. Since these profiles are only 20 km apart, an interpretation of a steep bounding fault on SCR3 that offsets Moho under the Jeanne d'Arc Basin there, but not under SCR401, would necessitate a change in this regional structure over a short distance.

4.3 Domain 2: extended continental crust, continental slope and rise

The zone of pervasive extension on the slope is imaged over about 150 km on SCR3 (230–380 km; Figs 3 and 4) from the edge of the continental shelf in the west to a landward-dipping basement reflection 'B' with progressively thinner continental crust towards the seaward end (Fig. 4). The domain is characterized by rotated blocks identified through the geometry of reflections from the upper surfaces of the faulted segments. At the shelf edge, several steep faults (F2) produce wide steps in the basement surface but are followed to the east by a major fault (F1 at 240–250 km) preserving a thick sequence of pre-rift, probably earliest Triassic, sediments (Solvason 2006) and a prominent basement high (BH). SCR401 (Fig. 2) shows a very similar structure. Beyond BH, crustal thinning becomes extreme ($<$10 km thickness of basement) and is referred to later in this paper as the zone of highly extended continental crust. The fault blocks are spaced typically at 5–10 km, bounded by down-to-the-east normal faults, and in some cases, preserve pre-rift sediments forming parallel dipping reflection packages at the upper surface of the fault block. Towards the eastern end of the zone of highly extended continental crust, there is an area (330–350 km) that shows no evidence of block faulting but rather is a rounded dome-like feature projecting up into the post-rift sedimentary section (BH6). We interpret this high as the first occurrence of exposed serpentinized mantle normally indicative of the transition zone. Farther east, a short section (∼360–370 km) is interpreted to have further minor block-faulted crustal basement before the 'B' event. Further details about identification of the block-faulted section are presented in the structural reconstruction section (Figs. 4–6).

East of the upper slope basement high (BH), is a complicated west-dipping horizon that is not offset by the faulting of overlying basement (Figs. 3 and 4). This feature was labelled 'L' in Lau *et al.* (2006b) to indicate a possible connection to the landward-dipping events imaged in FGP85-2 to the south (Keen & de Voogd 1988) though there are no analogous events under the slope on SCR2 to the north (Shillington *et al.* 2006) or nearby profiles FGP85-4, NB-1 (Lau *et al.* 2006b) or ER13 (Fig. 3). Lau *et al.* (2006b) speculated that the event may have been lost in processing on nearby profiles but a careful review for this paper of ER13 shot gathers and velocity analyses found no significant reflection energy that could be attributed to an 'L'-type event, suggesting that 'L' is a sporadic though significant feature. Two nearby proprietary industry seismic lines between SCR3 and FGP85-4 illustrate further the local variability of 'L' as the line nearest SCR3 has a well-developed 'L'-type event (Fig. 3) but the line closer to FGP85-4 does not. It is first imaged deep within basement at about 280 km on SCR3 (Fig. 3) and terminates to the east close to the basement surface. Throughout the block-faulted section eastward of 'L' (Fig. 4), there are no more crustal events associated with possible large-scale detachments or the crust–mantle boundary until the west-dipping feature that defines the boundary to the thinned continental crust at about 380 km (B on SCR3; Fig. 3).

ER13 (Fig. 3) is subparallel to SCR3 from a point about halfway down the slope. A similar zone of block faulting ends at a distinctive basement high (BH2). The similar location of basement highs on both ER13 (BH2) and SCR3 (BH1) suggests that the feature is a local ridge imaged on both lines at the edge of the zone of extended continental crust although there is a significant difference in character between the two. On ER13, the basement high rises farther through the sedimentary section and is sharply peaked with no internal reflective character. On SCR3, the basement high is low and rounded with a highly reflective internal character. FGP85-4, to the north of SCR3, does not show a basement high in this part of the profile and has no other distinct feature by which to identify the limit of extended continental crust and onset of exposed mantle. A detailed examination of extension of continental crust on this margin, based on the new data, is presented later.

4.4 Domain 3: transition zone

4.4.1 Impact of 'U' reflector on imaging basement

Throughout the Newfoundland Basin a strongly reflective horizon historically referred to as 'U' has been recognized (Tucholke *et al.* 1989) near the base of the sediments that generally overlies a seismically non-descript basement with minor topography and unreflective upper surface. The 'U' reflector was previously interpreted to be the rift–drift unconformity (Tucholke *et al.* 1989) but drilling results from Leg 210 (Tucholke *et al.* 2004) identified thin (∼10 m) post-rift (Hart & Blusztajn 2006) basalt sills at the level of 'U' emplaced into unlithified Aptian-Albian age sediments (115–111 Ma; Hart & Blusztajn 2006). Low-velocity overpressured sediments discovered between the sills further enhance the impedance contrasts and create an extraordinarily reflective environment (Shillington *et al.* 2007). Weak basement reflectivity could result from (i) reverberatory secondary arrivals from the large amplitude and highly coherent 'U' reflection package masking slightly later events, or (ii) the highly reflective 'U' could be preventing energy transmission into deeper levels of the crust (Shillington *et al.* 2008) or (iii) the basement surface could have an acoustic impedance similar to the overlying sediments. The presence of occasional basement highs in the Newfoundland Basin that extend above the 'U' event well into the overlying sediments provide a window of illumination into reflective basement suggesting that the basement overall may not be featureless but simply masked by the well-known difficulty of seismic imaging through basalts. It is important to point out that 'U' may not be strictly related to the composition of basement. In other words, defining the transition zone according to 'U' may lead to erroneous assumptions about basement as clarified below.

4.4.2 Basement highs in the transition zone

Two northeast-trending profiles (SCR305, ER22-23, Fig. 7) image transitional crust perpendicular to SCR3, ER13 and FGP85-4 (Fig. 3) illustrating the character of the transition zone in the Newfoundland Basin. To discuss 'U', basement highs occurring on the various profiles will first be described. East of 'B' on SCR3 (Fig. 3) lies a structurally simple section about 70 km in extent with one basement high (BH1). Lau *et al.* (2006b) describe BH1 on SCR3 as having a very complicated internal reflection character, in contrast to the featureless BH2 shown here from ER13 (Fig. 3). These basement highs near the landward edge of the transition zone on SCR3 and ER13 also differ in the character of their relief. BH1 is low and rounded, whereas BH2 is sharply peaked. Significant basement highs were not imaged between clearly continental and near oceanic environments on SCR2 (Shillington *et al.* 2006) or FGP85-4 (Fig. 3). The crossline, SCR305 (Fig. 7), which connects the transition zone of SCR3 with that of SCR2 (Fig. 1), is remarkably uniform over the 160 km between the two profiles, except for one internally reflective basement high (BH3) about 40 km north of SCR3. Although internally reflective, the basement high appears more like that seen on ER13 due to its high relief and strongly peaked topography. Erable profiles 22 and 23 (Figs 1 and 7) are subparallel to, and 20 km landward of, SCR305. There are two basement highs (BH4 and BH5) not completely covered by the 'U' reflector that are low and rounded and internally reflective, similar to BH1. BH6 on SCR3 is interpreted here as a serpentinized mantle basement high present within the zone of extended continental crust and similar in character to BH1, BH4 and BH5. Continental crust is interpreted at the basement surface between BH6 and the dipping basement reflector B. Lau *et al.* (2006b) noted that the magnetic anomalies M17 and older (Figs 1 and 3) in the Newfoundland Basin occur under the seaward edge of where we have interpreted very thin continental crust basement. The interpretation here of exposed serpentinite within the zone of highly extended continental crust illustrates that mantle may be very shallow and a significant factor even within the zone of extended continental crust.

The basement highs within the transition zone do not consistently belong to any geographic location within the Newfoundland Basin. On profiles SCR3 and ER13, basement highs seem to mark the boundary between basement with highly extended continental crust and transitional crust. Basement highs imaged on ER22-23 are located well within the transition zone. The basement high on SCR305 is near Domain 4 (oceanic). There are a variety of possible explanations for the origin of basement highs such as the topographic expression of a major transfer fault, a volcanic edifice associated with the nearby Newfoundland seamounts or early rifting, or a serpentinite ridge.

4.4.3 The 'U' reflector

The extent of 'U' on each profile is indicated in Fig. 3. 'U' continues landward conformably with surrounding sediments over extended continental crust on profiles SCR3 and FGP85-4 where its amplitude diminishes gradually (Fig. 3) but 'U' appears to end abruptly on ER13 against basement high BH2 (Fig. 3). It must be noted that where 'U' overlies extended continental crust on profile SCR3 there are basement reflections visible, perhaps as a result of a change in reflective character of 'U'. 'U' changes along that profile from a strong two-cycle event over the transition zone to a higher frequency single event over extended continental crust. On lines FGP 85-4 and ER13 (Fig. 3), the location of the boundary between Domains 2 and 3 is unclear because of the difficulty of imaging basement. 'U' continues seaward approximately as far as M3. Fig. 7 presents a detailed view of 'U' along SCR305 that shows the variability in the fine structure despite the apparent overall uniformity. A close inspection of 'U' shows that it varies in number and phase of cycles so that individual peaks come and go within a multicyclic package.

Some observations of 'U' along SCR305 that have implications for basement history include numerous small discontinuities in the vicinity of the basement high, which are interpreted here as being either E-W oriented strike slip faults (since they are not apparent on the nearby E-W oriented SCR3) or 'vent' features created by overpressuring of unlithified sediments, or both. If the breaks in the 'U' event are a result of leaking from overpressured sediments, the observation that they are not imaged on the crossline nearby suggests that another mechanism such as a series of parallel minor strike-slip faults provide the initial pathway for the vents to form. Another intriguing observation of 'U' along SCR305 is that it occurs at different depths across the basement high (BH3, Fig. 7). 'U' is about 1 km deeper south of the basement high which suggests an association with a major structural discontinuity. The continuous, bright, reflection sequence at about 7 km depth throughout the Newfoundland Basin is capped by a mid-Eocene unconformity (A^u) and occurs at the same level both north and south of the basement high, effectively limiting vertical movement across this boundary to prior to mid-Eocene. This basement high lies on the extrapolation of a major transfer fault mapped on the slope (Solvason 2006) although there is no evidence of its existence on nearby parallel profile ER22-23. It also lies on the trend of an offset in the oceanic magnetic anomalies (Fig. 1).

The overall continuity in reflectivity of 'U' suggests a continuity in origin. Despite small-scale variability, SCR305 suggests that overall these sills are very regular considering how thin they are and their widespread occurrence throughout the Newfoundland Basin over hundreds of square kilometres (Tucholke *et al.* 2004, ch. 3). It is remarkable to imagine such a widespread distribution of thin sills, but similar observations have been made in the vicinity of the Jan Mayen Ridge microcontinent in the north Atlantic, north of Iceland (Gudlauggson *et al.* 1988; Kuvaas & Kodaira 1997). Although the eastern margin of Jan Mayen is dominantly volcanic in origin, the highly reflective, widespread, post-rift sills (possibly thin flows) occur on the western margin that appears, otherwise, to have a non-volcanic history.

4.4.4 Nature of the transition zone

Since the 'U' reflector generally obscures basement, the use of the term 'transition zone' off of Newfoundland reflects some uncertainty as to the nature of basement. The transition zone of SCR3 and SCR2 has been interpreted through wide-aperture data to represent different types of crust by different workers. The transition zone of SCR3 and ER13 (Fig. 3) is interpreted to be a zone of mantle exhumation and serpentinization of smooth topography with little or no influence from seafloor spreading processes (Lau *et al.* 2006b). Van Avendonk *et al.* (2006) have interpreted what they call transition zone on SCR2 to be highly extended continental crust that correlates with Domain 2 described here and in Lau *et al.* (2006b). It has already been observed here that the boundary between basement that includes extended continental crust and basement that is presumably entirely serpentinized mantle (with or without magmatic contributions from the onset of slow seafloor spreading) is not apparent on nearby line FGP85-4 due to the overlying 'U' reflector. SCR305 shows that there is no difference in reflection character between the transition zones of SCR3 and SCR2 along that transect so care must be taken in making assumptions about basement based on the presence of 'U'.

4.5 Domain 4: oceanic crust

The fourth domain begins at the eastern edge of the transition zone on SCR3 with two minor topographic basement highs (30 km wide) with intervening strong crustal events. Immediately to the east, strong west-dipping events mark the onset of increasing topography and the well-known 'J' magnetic anomaly that is widely accepted as the onset of oceanic crust. The rough topography associated with the surface of oceanic crust is imaged for the remaining 80 km of SCR3 (Lau *et al.* 2006b). Sporadic intracrustal events are imaged but nothing consistent to mark a crust–mantle boundary.

The onset of oceanic crust in both profiles SCR2 and SCR3 appears similar. The relief of basement topography increases and basement velocities are represented by gradients reflecting likely serpentinized upper mantle (Van Avendonk *et al.* 2006; Lau *et al.* 2006a). On SCR3, the onset of oceanic crust is very similar to the oceanic crust of SCR2 beyond 270 km (M1) whereafter Moho topography is anticorrelated to basement surface topography and upper basement velocities are on the order of 5.0 km s^{-1} (Van Avendonk *et al.* 2006; Lau *et al.* 2006a). The presence of serpentinite and seafloor spreading-related volcanics was confirmed on SCR2 when ODP Leg 210 site 1277 was drilled (Tucholke *et al.* 2004). The presence of serpentinite seaward of previously interpreted magnetic lineations defining oceanic crust indicates that crust is transitional in nature farther out to sea than previously thought or else serpentinite is more a part of the slow spreading oceanic environment than previously thought. The drilling results may be representative of the onset of oceanic crust on SCR3 as well.

5 STRUCTURAL RECONSTRUCTION OF THE EXTENDED ZONE

5.1 Introduction

Passive margins and their sedimentary basins are created by extension of continental crust (McKenzie 1978; Wernicke 1985; Lister *et al.* 1986; Lavier & Manatschal 2006) and its ultimate splitting, with new oceanic crust forming during seafloor spreading (Whitmarsh *et al.* 2001). The pre-rifting fit of the continental edges is still apparent after separation despite their extensional deformation (Bullard *et al.* 1965), but detailed reconstructions often fail to produce unambiguous configurations and lead to implications of significant microplate rotations (e.g. Sibuet *et al.* 2007b). In such a context, it is likely to be important to estimate the variability of extension along and across conjugate margins. Because SCR3 extends from unextended continental crust, across rift basins on the shelf and the continental edge, to end on undisputed oceanic crust, it offers an opportunity to measure extension to be compared with estimates from the conjugate margin and other transects across this margin.

Estimates of extension along SCR3 have been made in two ways. Displacement of top pre-rift rocks across major faults, especially those bounding half grabens, provides a minimum estimate. Total extension would also include the effects of minor faults that are inadequately imaged on SCREECH, and ductile stretching, which might well be responsible for cryptic strains in the sediment column and is rheologically likely in deep basement. Varying thickness of crustal basement also provides an indirect estimate of extension, assuming a pre-rifting thickness like that of the unextended crust today, assuming no out-of-plane motion (i.e. the stretching does not vary at right angles to the profile used for calculations), and that there have been no additions to the crust (e.g. by underplating) during the rifting process. Detailed mapping in a small area at the southern edge of the Galicia Bank (Péron-Pinvidic *et al.* 2007) illustrates the continuity of faulting in continental blocks that vary laterally but do not appear to have significant out-of-plane motion. We present results from these separate estimates, and then discuss the implications.

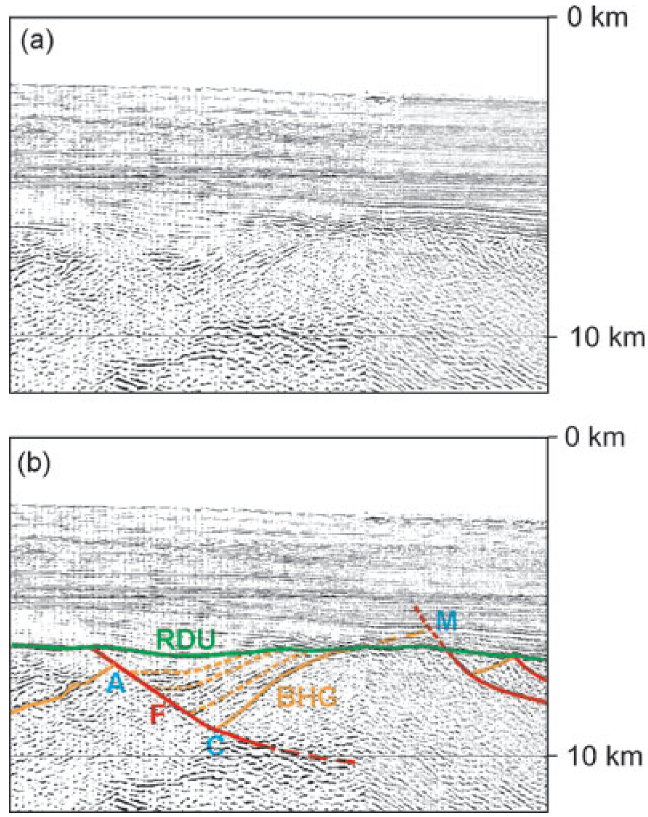

Figure 5. Blow-up of the rectangle at 290 km in Fig. 4, showing (a) uninterpreted and (b) interpreted sections illustrating the pick of a syn-rift sequence, below the rift-drift unconformity (RDU), and above the base of the half graben (BHG). The base of the syn-rift sequence is displaced from A to C across the fault, F, and then truncated by the unconformity, which has eroded the former cut-off, M, on the fault bounding the next half graben.

5.2 Estimates of extension from faulting

The SCR3 reflection profile has been converted to depth and interpreted to show the base of syn-rift rocks and the major faults bounding rift sequences (Fig. 4). The location of the base of the crust (Moho) is overlaid on the reflection image from the wide-angle seismic model of Lau *et al.* (2006a,b). The pick of base syn-rift rocks is determined subjectively as the base of reflective layering that shows evidence of originating in sedimentary sequences with diverging dips indicative of thickening sequences in wedge-shaped accommodation spaces produced by listric faulting (Fig. 5). Most often on SCR3, this horizon is also top seismic basement, but occasionally (especially below the Jeanne d'Arc Basin and below the Carson Basin at 250 km, Fig. 4) pre-rift sediments may be present and included above the horizon. The implication of error in the extension estimates from this is discussed below. The seismic velocity model has also been used to condition our selection, but again with some ambiguity. Where seismic *P*-wave velocities are below 3.5 km s^{-1} over depth intervals of 2 km or more, the material is likely to be sedimentary rock of tertiary or later Mesozoic (i.e. post-rift) age. Where the *P*-wave velocity is higher than 4.5 km s^{-1}, over such depth intervals the rock is likely to be crystalline or highly indurated sedimentary rock such as those typical of the Precambrian of the Avalon Zone, the Palaeozoic and earliest Mesozoic. In applying these loose velocity criteria, we recognize the horizontal smoothing of lateral velocity changes implicit in wide-angle modelling, such that the short-wavelength, saw-toothed features of the extensional half graben may not be accurately represented by the wide-angle model.

The major faults are evident from truncations, or offset, of reflections. They are readily picked out where the sedimentary response is strong, but their course deep within basins and within basement is somewhat conjectural. Where reasonable, we link well-determined fault loci within the sediments with deep reflections in basement that may be ductile shear zone equivalents of the shallow brittle faults. Those faults, that can be traced this way deep into the basement, appear to be listric, and so we assume that all the major bounding faults are listric. We then identify the corresponding cut-off points of base syn-rift rocks along each fault. There are some examples where the original upper cut-off point has been eroded by one of the several unconformities that mark the episodic rifting in this area. Where this appears to be the case, we have extrapolated the base of syn-rift rocks up from where it emerges downwards below the unconformity and extrapolate the fault upwards above the unconformity. The two lines of extrapolation intersect at the supposed cut-off point.

We measure the length of base syn-rift rocks between successive fault cut-offs and equate that with the unextended length for that segment. We also measure the horizontal offset of the cut-off points of top basement on either end of the segment and equate that with the extended length. The ratio of extended length to unextended length provides an estimate of the extension factor (β), and the difference of the two lengths gives the actual extension.

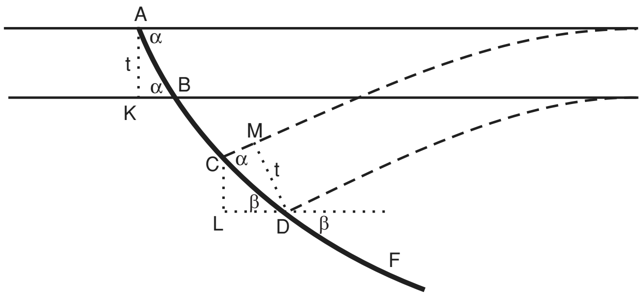

Figure 6. Fault, F, displaces a pre-rift sequence of thickness, t, from AB to CD. The fault changes dip from α to β from A to C. The rotation of the pre-rift sequence is assumed to conserve the angle, α, between layer and fault. Estimates of error based on including the pre-rift sequence with the syn-rift sequence are calculated from this figure and derived in the text.

An example from the inset rectangle of Fig. 4 (290 km) illustrates how extension is measured (Fig. 5). A weak fault reflection (F) truncates bedding reflections in a small half graben that shows internal growth above a lowermost strong reflector, assumed to be the base of syn-rift sediments. The half graben is syn-rift, with its base (BHG) displaced across the fault, F, from A to C. The heave across the fault (horizontal component of AC) is a close approximation to the extension associated with the half graben. However, the original horizontal bed length at top basement in strata within the half graben is given by the bed length along BHG from M to C. In this example, the rift–drift unconformity (RDU) has eroded the upper part of the half graben. The true extension is thus the heave minus the difference between the true bed length along MC and its horizontal component. This difference increases as the dip along MC increases. The β value associated with the half graben is the true extension divided by the original bed length MC.

Pre-rift sedimentary rocks are not observed over much of SCR3. They should lie below the base syn-rift picked. In rare cases (the Jeanne d'Arc Basin and the Carson Basin), there may be some pre-rift successions mistakenly included in the syn-rift fill. What effect might this error have on extension estimates? Suppose a listric normal fault, ABCDF (Fig. 6), with surface dip, α, causes a pre-rift formation of thickness, t, to be displaced from AK down to MD, where the fault dips at angle β, but the layer retains its thickness, t. Also suppose that the dip of the pre-rift formation is sufficiently modest that the heave across the fault is a good approximation to the extension. The extension calculated (correctly) at the top of the pre-rift sequence would be the horizontal component of AC, which equals the sum of the horizontal components of AB and BC.

The extension calculated if the pre-rift sequence were included mistakenly in the syn-rift sequence would be the horizontal component of BD, which equals the sum of the horizontal components of BC and CD. The error in extension by mistaking the pre-rift for syn-rift sediment would be the difference between these two lengths, that is, the horizontal components of BD – AC, or (BC + CD) – (AB + BC), that is, the horizontal component of CD – AB. From Fig. 6, this difference is LD – KB. From the various right-angled triangles, note that

$$\mathrm{KB} = t/(\tan\alpha.)$$

Also

$$\mathrm{CD} = t/(\sin\alpha),$$

and that

$$\mathrm{LD} = \mathrm{CD}(\cos\beta),$$

so that

$$\mathrm{LD} = t(\cos\beta)/(\sin\alpha).$$

Thus, the error in extension estimate, $\mathrm{LD} - \mathrm{KB} = t\ \{(\cos\beta)/(\sin\alpha) - 1/(\tan\alpha)\}$ or $t\{(\cos\beta) - (\cos\alpha\}/(\sin\alpha)$.

In the Carson Basin (Fig. 4, 250 km), the pick of the base syn-rift has a strongly reflective sequence immediately overlying it that shows only modest growth towards the bounding fault of the half graben. If it is supposed that this is actually pre-rift, then its thickness might be as much as 3 km. The bounding fault shows a shallowing in dip from about 35° (α) to 25° (β) over the slip zone affecting this supposed pre-rift sediment succession. Substituting these values in the error estimate, the amplitude of the error is $\sim$ 0.5 km. The heave used herein is actually 7 km, so that the error is about 7 per cent. It is concluded that the errors in extension estimates associated with mistakenly picking the base of pre-rift, instead of base syn-rift, are finite, but small, even in the most extreme case along this profile.

The resulting estimates of the lateral variation in β are given in Fig. 4. From fault displacements, the cumulative extension over the SCR3 profile is 85 km, 6 km over the shelf (as far as km 220) and 79 km over the slope.

5.3 Estimates of extension from crustal thinning

The SCR3 transect starts on the feather edge of post-rift sedimentary rocks. SCR3 wide-angle seismic results indicate a crustal thickness of around 36 km, where constrained, a few tens of kilometres east of the end of the line, within the cover of post-rift sediments but not within any notable rift structures (Lau *et al.* 2006a). A Lithoprobe East wide-angle seismic profile 120 km farther west yielded a crustal thickness of 40 km on the unextended Avalon Zone of the Appalachians (Marillier *et al.* 1994). This thickness characterizes unextended Avalon Zone and is corroborated further by earlier results (Keen *et al.* 1986, 1987). A recent seismic reflection transect on land in southwest Iberia images a very consistent Moho at about 10.5 s corresponding to a depth of 30–35 km (Simancas *et al.* 2003).

We calculate crustal extension over successive length segments of a few tens of kilometres of the SCR3 profile. Assuming conservation of cross-sectional area (no net out-of-plane movement), the ratio of extended to original segment length (β) is equal to the ratio of the unextended crustal thickness to the crustal thickness after extension (thickness today). The assumed thickness of unextended continental crust along SCR3 is 36 km. Had we assumed a 40 km unextended crustal thickness, then our results for estimates of the extensional β factor would have been 10 per cent higher. To calculate present crustal thickness, we use the location of Moho as determined by Lau *et al.* (2006a) from wide-angle data, and our pick of top pre-rift rocks, as discussed above.

The resulting estimates of β are shown in Fig. 4. The total extension from the crustal thinning estimates is given by summing the extensions estimated for the series of segments into which the crust may be divided (and indicated by a β value on Fig. 4), where the segment extension is given by its length multiplied by $(1 - 1/\beta)$. Thus, from crustal thinning estimates, we estimate extension over the shelf of around 6 km and over the slope of 114 km, to give a total extension of 120 km. Given that the most outboard occurrence of continental crust today occurs at around 380 km (Fig. 4), the edge of the original continental crust (pre-rifting) would have been at 260 km, about 25 km beyond the present shelf edge.

5.4 Comparison of extension estimates

Extension estimated from interpretable faulting (85 km) is rather less than that estimated from crustal thinning (120 km). This is probably because the faults interpreted do not account for all the extension within the upper crust: minor faults within the half graben may also be present and unaccounted, either because they are invisible on the seismic profile or identifiable but with an indeterminate offset. This kind of discrepancy has been pointed out by Gibbs (1984) and he explains why hanging wall deformation geometries often demand additional faults within the half graben, both synthetic and antithetic to the major bounding fault. Many examples have been observed, for example in the northern North Sea (e.g. Christiansson *et al.* 2000), from which the sum of heaves across the internal minor faults may readily achieve several tens of percent of that across the basin-bounding fault. In addition to the brittle response manifest in the faulting, there may also be a modest amount of ductile extension within the sedimentary section. It is also possible that some extension might have occurred without accommodation space for a preserved sedimentary record that would have otherwise contributed to offset across faults. This might happen if heat flow were particularly high, or extension relatively higher in the lower lithosphere. Whatever the cause, we do observe erosional unconformities along SCR3, which are a manifestation of uplift to base level during rifting.

Reston (2007a) has also addressed this issue, assessing the roles of depth-dependent stretching and early faulting not recognized in seismic sections. In the examples considered, the evidence suggests that failure to recognize all the faulting is probably the more likely cause. It is well known that as displacement increases across listric normal faults, they tend to flatten to the point where they become impotent to further stress release and new faults form cutting across the old ones (Proffett 1977). It is of interest in this context to note that on SCR3, beyond the basement high at 260 km (Fig. 4), the estimates of β increase from landward values of 1.0 to 1.3 to 3.3 and then decrease to values between 1.1 and 2.1, where crustal thinning indicates the values should be 3.5 or more. It is in this zone of high extension that early faults with significant displacements across them might remain unrecognized in our assessments.

The crustal thinning estimates assume a constant crustal thickness (of 36 km) prior to extension. We have noted above that variation from this within the Avalon Zone is expected to be modest (possibly as much as +10 per cent), and we expect the Avalon Zone to lie below all the crust here considered: potential field analyses show the Avalon anomaly pattern extending from land to Flemish Cap (Haworth & Lefort 1979). The unextended basement in NW Spain is Hercynian and is characterized by a lower crustal thickness, around 32 km (Cordoba *et al.* 1987). Somewhere between Newfoundland and NW Spain, Hercynian characteristics may be impressed on older features. There is no geological evidence that this occurs below the Grand Banks, though geophysical arguments have been presented to suggest that there may be Hercynian trends below the Grand Banks (Lefort *et al.* 1993). Tolerating all such possibilities indicates an uncertainty of +/−10 per cent in β estimates from crustal thinning.

Given all this background, the difference of 35 km between the two estimates of extension is not unreasonable. We suggest that the fault estimate is a minimum estimate, and the crustal thinning estimate, although subject to gross assumptions about original crustal thickness, may be closer to true. Accepting this, the conclusion that the original break point, assuming it was vertical, was 25 km beyond the present-day shelf edge can be compared with other estimates. Sibuet *et al.* (2007b) used Bouguer anomalies to estimate the position of the original break, by associating it with the inflection point in the present-day Moho (this assumes that continental crust extending now beyond the point can be packed exactly into the space between the thinned crust inboard of the inflection point and the original deeper Moho). Their estimate applied to SCR3 would have the original break point at around 245 km (Fig. 2), 15 km landward of our estimate and about 10 km oceanward of the present shelf break. Minshull *et al.* (2001) calculated an extension of 56 km on nearby FGP85-4 based on crustal thickness measurements and the limit of continental crust as defined by Keen and de Voogd (1988). As seen in Fig. 4, the limit of continental crust on FGP85-4 was picked about 40 km landward of our pick on SCR3. FGP85-4 is particularly difficult to pick because of the poor basement imaging under the 'U' reflector. Their (Minshull *et al.* 2001) calculation is also relevant to end-stage extension only and uses a total crustal thickness of 27 km and so is underestimated. Crosby *et al.* (2008) calculated 126 km of extension across imaged faults along SCR3, but they included faulting over 100 km of transitional crust in addition to the thinned continental crust used in our estimates. Though they took a pre-extension crustal thickness of only 30 km, the two results appear to be compatible.

It is noteworthy that these estimates of extension exceed those from the conjugate margin: 40 km from crustal thinning (Minshull *et al.* 2001) along IAM9 and 35 km from detachments (Manatschal *et al.* 2007) along Lusigal 12. Both these profiles are located significantly farther north than the conjugate location of SCR3 but are the closest data we have for comparison. The combined total extension across the conjugate margins is about 160 km from crustal thinning and at least 120 km from structurally defined shearing in the upper crust, with final separation leaving the majority of the thinning on the Newfoundland margin.

6 DISCUSSION

6.1 Comparison with the conjugate margin

The Tore seamount on the Iberian margin is directly conjugate to SCR3 and obscures the continent–ocean transition so profile SCR3 cannot be compared with any profile directly conjugate. Regional seismic profile IAM9 (Pickup *et al.* 1996; Dean *et al.* 2000) is located about 100 km to the north of the Tore seamount and possibly across a minor fracture zone (Sibuet *et al.* 2007a), but is the closest available crustal scale seismic reflection profile for conjugate comparison (Fig. 3). The bathymetry and basement topography along IAM9 is unaffected by Galicia Bank, unlike LG12 (Krawczyk *et al.* 1996) to the north, which, along with the location of the continental shelf, makes IAM9 more representative of the southern Iberia Abyssal Plain. It has significant similarities to, and differences from, SCR3. IAM9 images only the lower slope and rise out to oceanic crust so will be compared to the deeper water part of Domain 2 through Domain 4 of SCR3. Although the object of this paper does not include a detailed comparison with the Galicia Bank margin, a few key observations are referred to here to facilitate the following discussion.

6.1.1 Landward-dipping boundary event

The IAM9 profile has a high-amplitude, complicated and discontinuous event ('L', Fig. 3) that is very similar in character to the landward-dipping event 'L' (Lau *et al.* 2006b) described in Domain 2 of SCR3. It occurs in both cases near the base of the slope and when extrapolated to the surface marks the boundary between thick, moderately extended continental crust and thin (<10 km), highly extended continental crust. The velocity model on the Iberian side (Dean *et al.* 2000) and interpretations of refraction data on the Newfoundland side (Reid 1994; Lau *et al.* 2006a) suggest that the location of 'L' is a contact between lower continental crust on the continent side of the boundary and serpentinized mantle on the seaward side.

The 'L' reflections may be relicts from early stage localization of extension where stress concentrations first extend through the whole crust (Brun & Beslier 1996), allowing fluids into the mantle and subsequent serpentinization. This process of stress concentration would necessarily be heterogeneous as 'L'-type reflectors are sporadically imaged.

6.1.2 *Thinned continental crust*

The lateral extent of highly thinned continental crust is about 100 km (Lau *et al.* 2006b) in the Newfoundland Basin and is about 40 km on IAM9 (Pickup *et al.* 1996; Dean *et al.* 2000) totalling ∼140 km. The conjugate SCR2–LG12 transect to the north (Krawczyk *et al.* 1996; Shillington *et al.* 2006) has a total width of ∼110 km for the highly thinned continental crust if the transition zone of SCR2 is continental crust (Van Avendonk *et al.* 2006). If the SCR2 transition zone is actually serpentinized mantle, the zone of highly extended continental crust would be only 65 km wide and would all be located on the eastern margin. This interpretation infers a narrowing of the zone of intense crustal thinning northwards and a possible change in locus of crustal breakup. South of LG12, crustal separation seems to have occurred on the eastern end of the zone of highly thinned continental crust whereas to the north, crustal breakup occurred at the western edge of the extended continental crust.

A lack of reflective detachments is a general characteristic throughout the Newfoundland Basin. Lau *et al.* (2006a) pointed this out for nearby profiles NB1 and FGP85-4 and we have made further observations on ER13 and nearby proprietary industry profiles. The zone of highly extended continental crust imaged on SCR3 differs from that observed on the profiles along the western Galicia Bank due to the lack of extensive reflective detachment faults ('S' events, Reston 1996; Reston *et al.* 1996; Lau *et al.* 2006a). North of IAM9, along the southern edge of Galicia Bank, LG12 (Krawcsyk *et al.* 1996) has at least one detachment fault ('H') where the first peridotite ridge is exposed. Although not imaged elsewhere, the 'B' feature on SCR3 shown in Fig. 3 may be a well-defined, but localized, shear zone in brittle upper mantle that represents the point at which upper crust separated between the two sides.

6.1.3 *Ocean–continent transition*

On the Newfoundland Basin side, the transition zone is about 70 km wide, whereas on IAM9 the interpreted transition zone is about 100 km wide. A comparison of the transition zones is complicated by the pervasive presence of the strong 'U' reflector that occurs only on the Newfoundland side. At first glance, it appears that topography on the basement surface of the transition zone is rougher on IAM9 than SCR3 but the overlying flat 'U' event could have an apparent smoothing effect on the appearance of the basement surface in the Newfoundland Basin. Even on SCR2 where basement is often described as smooth, it is apparent that there is at least 1 km of relief under the 'U' reflector (Shillington *et al.* 2006). 'Windows' into the basement on the profiles in the Newfoundland Basin where 'U' is absent show that basement is clearly imaged at its upper surface and is often internally reflective. This is similar to basement on IAM9 except for the uppermost 1–2 km of basement which on IAM9 is much weaker in amplitude than the immediately underlying highly reflective basement. This observation has been interpreted as an effect of pervasive serpentinization in the uppermost basement with increasing reflectivity due to more heterogeneous serpentinization at depth (Pickup *et al.* 1996). The basement highs where 'U' is lacking in the transition zone of the Newfoundland Basin do not show a similar amplitude increase with depth (see also Shillington *et al.* 2006) suggesting that any apparent transparency (Lau *et al.* 2006b) results from a lack of reflections where 'U' is influencing signal. The heterogeneous dipping events in the transition zone basement of SCR3 and IAM9 indicate that faulting in the upper mantle is widespread.

The transition zone in the Newfoundland Basin has a variety of basement highs with different topographic profiles and internal reflective characteristics. The highs do not seem to form extensive ridges except perhaps in the vicinity of oceanic basement where overall topography is consistently greater. The apices of basement highs imaged in the Newfoundland Basin have been plotted in Fig. 1. Basement highs in the transition zone as imaged in ER22-23 and SCR305, though in close proximity to each other, are not collinear and so, apparently, are not part of a continuous feature. BH1, identified on SCR3, may have a counterpart on ER13, but not on FGP85-4. These highs also occur at the continental end of the transition zone instead of the oceanic end as observed on IAM9. There are minor basement highs that occur on the oceanic end of the transition zone of SCR3 (∼CDP 538000, Fig. 3) but these are not evident on FGP85-4 and so do not appear to be as extensive and persistent as the ridge complex on the Iberian side. Lau *et al.* (2006b) and Shillington *et al.* (2006) both noted that basement highs appeared as ridges over distances of tens of kilometres in the vicinity of the oceanic environment and we note a similar possibility for correlation between a more landward basement high on SCR3 and ER13. Determining whether the near-oceanic ridges are as continuous as similarly located ridges on the Iberia side would require more seismic profiles.

Inspection of current satellite data in the Newfoundland Basin shows only a limited association of basement highs with the complicated gravity field. The largest, most well-defined basement highs on ER13 and SCR305 do not have associated gravity highs which supports an expectation that there is a great deal of serpentinization in basement producing a density similar to that of the overlying sediments. The linear basement high off the Galicia Bank was also observed to lack an associated gravity anomaly.

6.1.4 *Water depth*

The depth to seabed is significantly different between the two conjugate margins (Louden *et al.* 1991; Shipboard Scientific Party 2004). On the Newfoundland side, the seafloor depth below the slope is ∼3.5 km versus ∼5 km off of Iberia (see bathymetric contour colouring in Fig. 1). This difference in depth is apparent over oceanic crust as well (Fig. 1). The sediments on the Newfoundland side appear to be thicker than off Iberia, but not so much as to offer an isostatic explanation. Müntener & Manatschal (2006) have compared chemical analyses indicating that the mantle composition is distinctly different between the margins along the SCR2–LG12 transect with the Newfoundland mantle being more depleted. This suggests that the breakup of continental mantle might have been initiated at a compositional discontinuity inherited from a previous orogenic event and that a difference in overall topography may be the result. Crosby *et al.* (2008) attribute the difference to asthenospheric convective flow.

6.1.5 *Oceanic crust*

The boundary between transitional and fully developed oceanic crust is under question still along the Newfoundland margin. In the slow-spreading environment, there may be a gradual onset of oceanic volcanics without an identifiable boundary (Whitmarsh *et al.* 2001; Jagoutz *et al.* 2007). Lau *et al.* (2006) interpreted oceanic crust to begin with the increase in topography seaward of the transition zone in the vicinity of magnetic anomaly M4/M3 (∼130 Ma). Shillington *et al.* (2006) also favoured the M4/M3 area as the onset of oceanic crustal development but pointed out that drilling results of sites 1277 and 1070 (which were both located slightly seaward of M3) did not recover fully developed oceanic crust, but rather some units of oceanic affinity overlying serpentinized mantle. The geochemical analysis of serpentinites from site 1277 (Müntener & Manatschal 2006) further supports the interpretation of mantle of continental affinity. These results indicate that incipient seafloor spreading at non-volcanic margins develops gradually and there may be a significant amount of exposed mantle amongst the intrusives of the early oceanic crust environment. Van Avendonk *et al.* (2006) suggest that basaltic volcanics and intrusions accumulated between exposed mantle basement highs in the oceanic environment of SCR2.

6.2 Rifting model

Rifting models that could explain the exposure of serpentinized mantle without seafloor spreading volcanism have evolved from simple end-member one-stage processes of pure shear or simple shear mechanisms. Workers now often invoke a multistage process involving early overall pure shear to account for early symmetry followed by late stage detachment faulting after the crust has been highly thinned (summarized in Péron-Pinvidic 2008). The nature of deformation within the mantle throughout these processes is less often discussed but may be key to understanding the process of mantle exposure at the seafloor.

A cartoon illustrating the observations made along SCR3 is presented in Fig. 8 for reference during the following discussion. The evolution of the margin at the SCR3 rift segment began with a period of widespread symmetrical faulting, as the shelf and upper slope basins of both Newfoundland and Iberia show, with faults dipping seawards and crustal thinning modest to moderate. This is consistent with a relatively weak mid-crust and perhaps lowermost lower crust where the large listric crustal faults were able to sole (Nagel & Buck 2004; Lavier & Manatschal 2006; Huismans & Beaumont 2007). The rheological distribution during early rifting is assumed to have included rigid upper crust/ductile mid-crust above a rigid lower crust/ductile lowermost crust (Pérez-Gussinyé *et al.* 2001). The upper mantle and lower crust brittle zone accommodated additional extension through faulting and possible boudinage (Brun & Beslier 1996; Huismans & Beaumont 2007; Reston 2007b).

As rifting progressed, continental crust continued thinning by listric faulting until weak zones no longer existed and the whole crust was in the brittle state (Pérez-Gussinyé *et al.* 2001) and mechanically coupled to the brittle upper mantle. Eventually, under

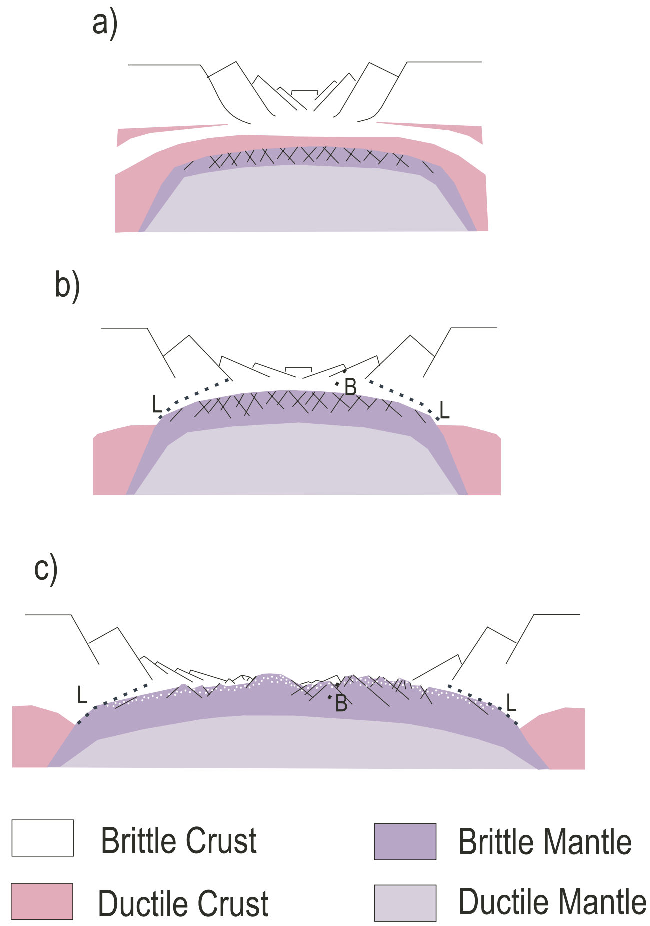

Figure 8. Schematic summarizing history based on observations made in the various data sets highlighted in this paper. A simplified starting rheology of brittle upper crust and lower crust with ductile zones in the mid and lower crust is portrayed before significant thinning occurred. (a) Pure shear extension dominates with brittle upper crust and lower crust and upper mantle and thinning crust, shallower mantle. (b) Pure shear extension dominates; early stress concentrations (L) define limits of zone of symmetric extension and occur approximately at the point where the whole crust becomes brittle. Rupture of crust occurs on the eastern side (B) leaving a larger extent of continental crust to the west (SCR3) and a more abrupt termination of continental crust to the east (IAM9). (c) Overall pure shear extension continues to dominate, segmenting upper crust and producing extensive fracturing and pervasive serpentinization (indicated by white dots) of brittle upper mantle until a future point when asthenosphere rises to the surface and melt is produced leading to the formation of new upper mantle and eventually oceanic crust.

influence of the extensional stress field and whole crust brittle conditions, the upper mantle began to be serpentinized by the infiltration of seawater. On both SCR3 and IAM9, major landward-dipping shear zones (reflector 'L', Figs 3 and 4) formed under the slope where crustal thickness is associated with the onset of serpentinization (Pérez-Gussinyé *et al.* 2001). This point also marks the change between moderate and extreme crustal thinning. Velocity models also support this interpretation (Dean *et al.* 2000; Lau *et al.* 2006a), showing a juxtaposition of serpentinized mantle with lower continental crust in the vicinity of 'L'. Landward dipping shear zones could be interpreted as a mantle shear zone similar to that predicted from Brun and Beslier's (1996) physical modelling results. Apparent seismic images of mantle shear zones have been reported in margin environments and have been described as a zone of decoupling moderately thinned from highly thinned continental crust (Reston 1993).

When serpentinization had progressed enough to produce a continuous weakened layer at the top of the mantle, the continental crust became decoupled from the mantle. According to Pérez-Gussinyé *et al.* (2001), thinning and faulting of the continental crust can continue along faults soling into serpentinite décollement zones until crust separates. Such detachments have been imaged along the Galicia Bank ('S' reflector, Reston *et al.* 1996). It is important at this point to note that detachments are not imaged at the SCR3–IAM9 segment except perhaps over a small distance at the point of crustal separation (event 'B' in Figs 3 and 4). Since the inspection of other regional profiles shows that the lack of seismically imageable detachment faults is a general observation in this part of offshore Newfoundland, it is likely that there was a difference between the SCR3–IAM9 segment of the margin and that farther north along the Galicia Bank. It may be that continental crust and upper mantle remained mechanically coupled until a late stage of extension when crust was extremely thinned. Another possibility may be that deformation was more vertically distributed in the zone of serpentinization in this area rather than occurring along well-defined deformation zones at the upper surface of the serpentinized mantle. Significant mobilization and diapirism of serpentinite from under adjacent crust resulting in a serpentinite mound (BH6, Figs 3 and 4) may remove regular reflecting surfaces. Serpentine that is able to flow would come only from the uppermost mantle where alteration has been extreme leaving less altered serpentinite in place. A similar process has been proposed in the Porcupine Basin where a diapiric mound of serpentinite appears to have flowed out over adjacent block faulted crust (Reston *et al.* 2004). In the Porcupine Basin images, the detachment is not very reflective in areas of interpreted serpentinite mobilization.

The point of crustal breakup occurred closer to the Iberian side of the highly thinned continental crust, producing an asymmetry between the two margins in final geometry at least in terms of continental crust distribution. A much larger extent of the highly thinned crust is on the Newfoundland side and most of the faults are seaward dipping. Asymmetrical breakup is consistent with model predictions in slow rifting environments (Bassi 1995; Huisman & Beaumont 2007; Van Avendonk *et al.* 2008).

The mechanism(s) accommodating ongoing exposure of subcontinental mantle after the rupture of continental crust is a topic of much debate at the moment. Simple shear explanations have proposed an asymmetrical process of large-scale detachment faulting where deeper levels of mantle are exposed upon extension from a single fault zone (Wernicke 1985; Lister *et al.* 1986, 1991). We image nothing along the SCR3–IAM9 segment that represents a major detachment fault and the seismic observations of exposed serpentinite zones across SCR3–IAM9 are quite symmetrical. The zones are similar in size, are of fairly low relief at the landward end, increase in relief near oceanic crust and have abundant heterogeneous reflectivity with a wide variety in dips.

Recent refinements to mantle exhumation models include explanations for the development of low-amplitude magnetic anomalies and differ largely in tectonic concept between large-scale concave-downward individual detachments (Lavier & Manatschal 2006), an environment of widespread normal faulting (Whitmarsh *et al.* 2001; Russell & Whitmarsh 2003) and centralized slow seafloor spreading that is asymmetrical at any moment in time but broadly symmetric over large periods of time (Sibuet *et al.* 2007a). Another mechanism similar to slow seafloor spreading suggests dominantly landward dipping mantle shear zones symmetric about the axis of extension (Nagel & Buck 2004; Reston 2007b). The variety in dip of basement events alluded to previously in the transition zones of both the Iberia and Newfoundland Basin data suggests that extension processes were more heterogeneous, perhaps more distributed, during rifting of the transition zone. At least some of the mantle reflectivity could have developed prior to crustal separation as brittle processes can happen both above and below the serpentinized mantle detachment zone, and even prior to serpentinization when crust–mantle deformation processes were linked. Highly stretched mantle could also produce large offset and rotated normal faults accounting for the tectonized basement surface drilled on basement highs.

An explanation suggested here for the generally low relief throughout the area of exposed mantle is the relatively high degree of surface serpentinization that may not support strong relief. Submarine erosional processes as well as the tendency to flow may have a smoothing effect. In contrast, features of the near-oceanic crustal environment may help to preserve higher relief such as basalt flow caps, gabbroic intrusions and more intense tectonic faulting due to narrowing of the zone of extension.

The processes operating throughout rifting in this non-volcanic environment may be related ultimately to the processes that are observed at presently active slow spreading centres (Nagel & Buck 2004; Cannat *et al.* 2006; Sibuet *et al.* 2007a; Reston 2008). The presence of weak magnetization, subdued topography and reflective basement leads us to prefer an overall pure shear mechanism that gradually localizes and develops into slow but stable seafloor spreading similar to Reston's (2007b) description.

It is important to be careful extrapolating interpretations from one part of the margin to another. There are significant differences between interpretations based on observations along the SCR2–LG12 transect (Van Avendonk 2008; Péron-Pinvidic 2008) to the north and the SCR3–IAM9 transect described in this paper. Some of these differences include bright reflectors from possible detachments seen on LG12, lack of highly thinned normal faulted crust on SCR2, lower crust anomalous velocity at the shelf edge on SCR2 not interpreted anywhere else and anomalous velocity structure in the landward section of the transition zone on SCR2 under 'U'. Much progress has been made in developing creative and more complex models describing margin evolution through time based on increasingly more complex data and it may well prove to be the case that complexity and chain of events differ along the margin as well.

7 CONCLUSIONS

Refined processing of the SCR3 profile, imaging of its entire shelf section, analysis of crustal extension, regional interpretation with a number of adjacent deep profiles and comparison with the conjugate margin allow us to extend and adapt the earlier interpretation of Lau *et al.* (2006b) as follows:

(1) *Continental shelf of the Grand Banks*: the shelf platform is underlain by Avalon crust residual from Appalachian and earlier tectonic events. The crust below the unextended platform has a consistent thickness of around 36 km. It contains two ~5 km thick zones of laterally intermittent subhorizontal reflectivity, one at 20–30 km depth and the other just above the Moho at 32–40 km depth. These are crossed by zones of west-dipping reflectivity, extending from the lower crust into the mantle, but without obvious offset of the Moho. These features extend across the entire shelf section, without spatial correlation with basins resulting from Mesozoic extension.

We interpret these fabrics as pre-Mesozoic, possibly Appalachian, but more likely of Precambrian age. The east-dipping listric Murre fault, which bounds the Jeanne d'Arc basin, soles into the lower crust, confirming earlier observations (Keen *et al.* 1987).

(2) *Extended continental crust of the SE Newfoundland slope and rise*: rifting produced a series of half grabens with dominantly east-dipping bounding listric normal faults. The zone of extension reaches 170 km beyond the shelf edge. Estimates of absolute extension from fault heaves and basement thinning are 85 km and 120 km, respectively. Reconstruction of the margin suggests that the 36-km-thick continental crust that has been extended into the thinned crust on this margin had its pre-rifting edge some 25 km seaward of the present-day shelf edge.

(3) *Transition zone*: the transitional zone between continental and oceanic crust is characterized by serpentinized mantle. Our interpretation suggests that there may be a window into exhumed mantle within the zone of extended continental crust, so that it first appears below rift sediments 50 km farther east than in the earlier interpretation of Lau *et al.* The seismic velocities in the window are low so that the earlier interpretation focused on continental crust, but extreme serpentinization could also produce low velocities.

The transition zone is dominated by, and to a great extent its basement is hidden by, the strong 'U' reflector, now known to be produced by basalt sills intruded into sediments just above the basement. The 'U' event is widely present throughout, although not necessarily restricted to, the transition zone along the SE Newfoundland rifted margin. Basement under 'U' appears to be of low relief but additional profiles reveal that basement highs are not uncommon throughout the transition zone. These basement highs also reveal that basement is often highly reflective with steeply dipping events such as those found on IAM9. The basement highs do not appear to have strong linear trends subparallel to the margin, as observed off Iberia, especially along Galicia Bank, but appear to have a patchy distribution except in the near-oceanic environment where topography increases. Disruptions in 'U' occur in limited areas and appear to be aligned perpendicular to the margin. These may be related to transfer faults associated with rifting in the shelf-edge basins.

(4) *The L reflector*: a sporadically imaged landward-dipping reflector, 'L', observed on several profiles below highly extended continental crust of the SE Newfoundland slope also appears on IAM9. Wide-angle data (Lau *et al.* 2006a) were interpreted to suggest that this reflector separates continental crust from underlying serpentinized mantle.

(5) *The rifting model*: observations of seismic data presented here suggest that if IAM9 is representative of the Iberian margin south of that profile, rifting processes have been largely symmetric along this part of the margin. Differences between these profiles and others farther north indicate that the relative contribution of asymmetric and symmetric mechanisms of rifting may vary between different parts of the margin.

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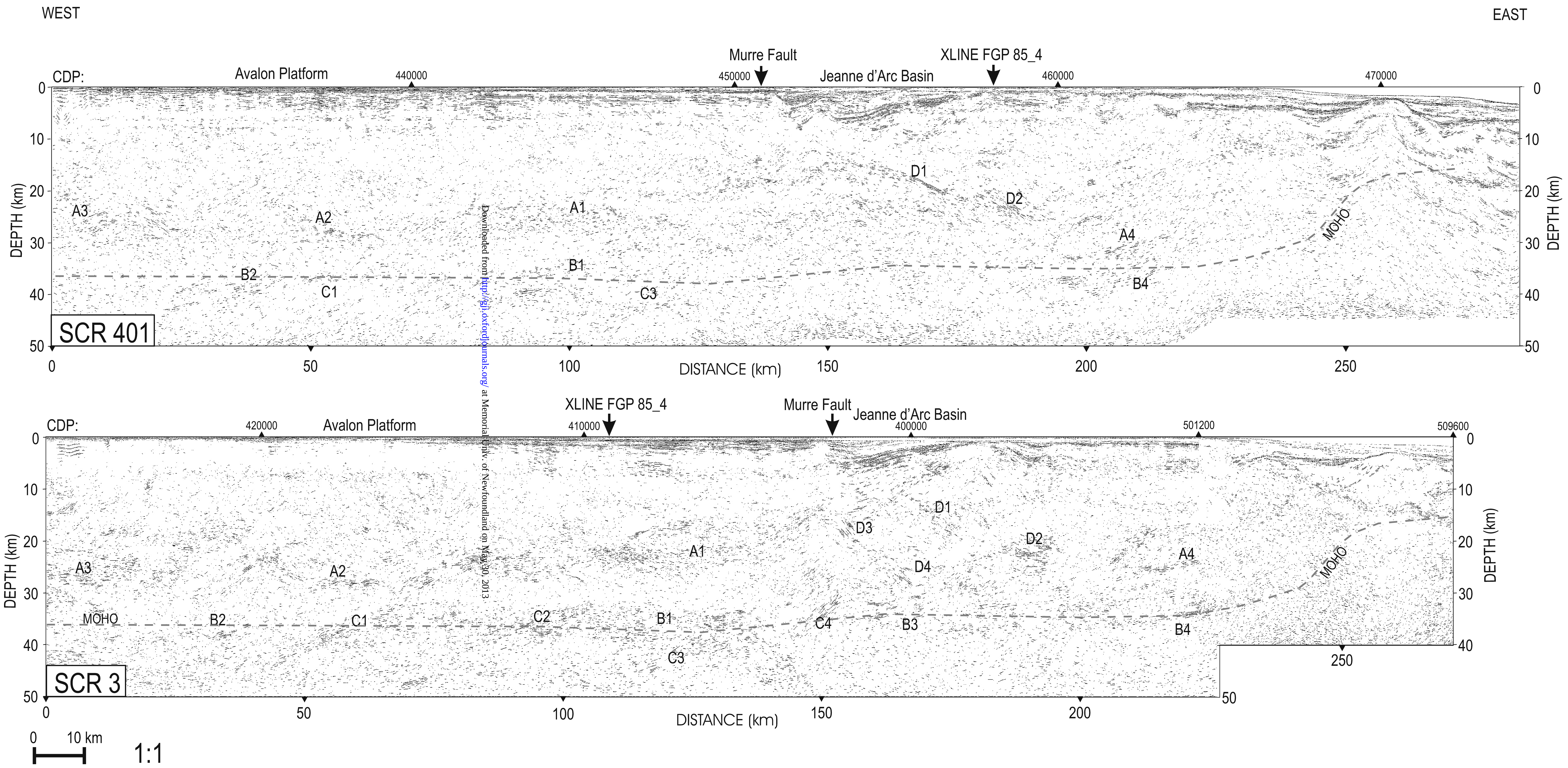

Figure 2. Crustal reflection images of continental shelf from profiles SCR3 and SCR401. Post-stack time migrations are converted to depth using the velocity structure in Lau *et al.* (2006a) and plotted here with no vertical exaggeration. Moho indicated in grey according to Lau *et al.* (2006a,b) is based on wide-angle data collected along SCR3 and reproduced on SCR401 for comparison purposes. Reflection character labelled according to: A (subhorizontal mid-crustal fabric), B (subhorizontal lower crustal fabric), C (dipping lower crustal or upper-mantle fabric), D (events associated with the Murre fault). Scale 1:1.

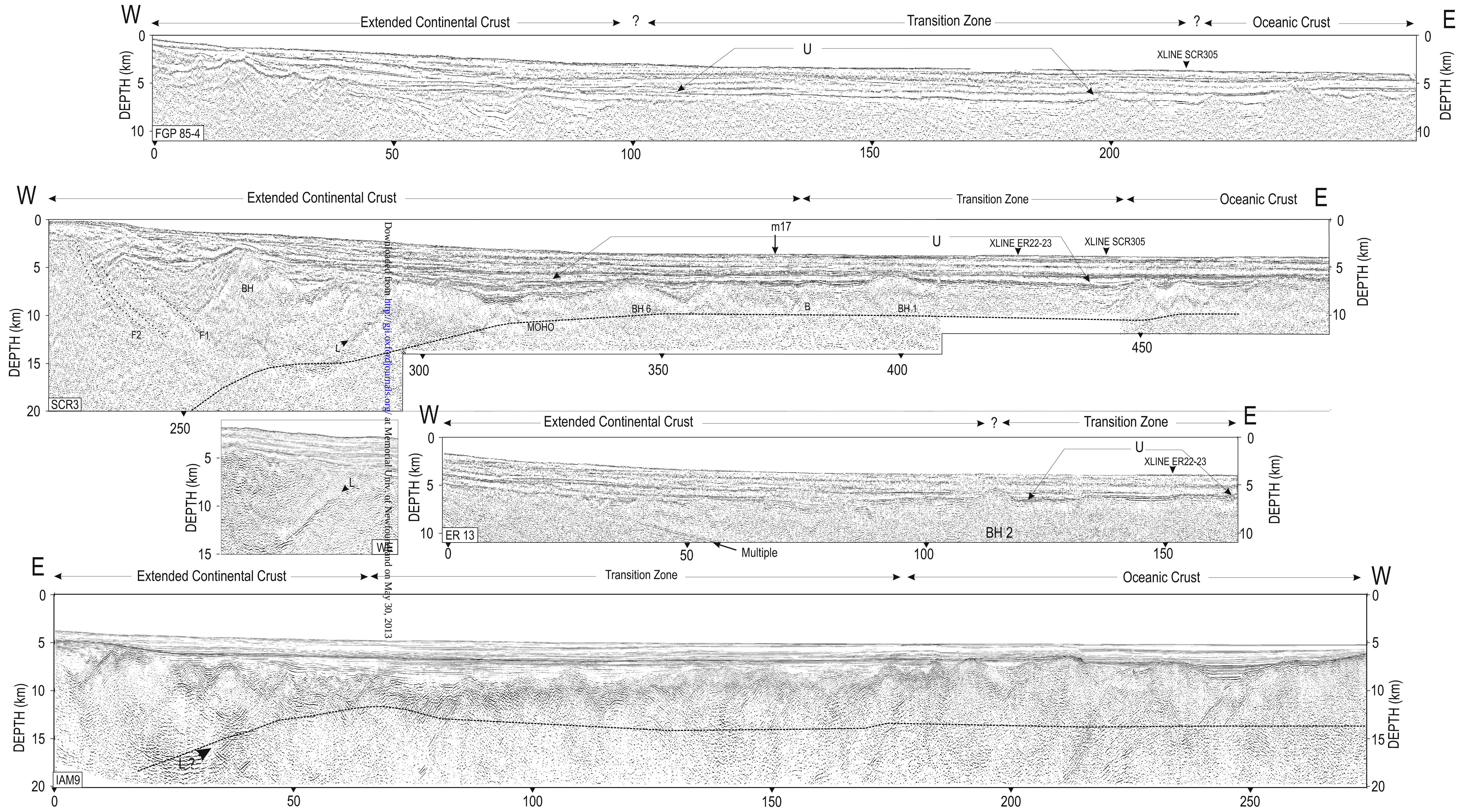

Figure 3. Seismic reflection profiles across the continental slope and rise over the Newfoundland basin and conjugate profile IAM9 across the Iberia Abyssal Plain. IAM9 is plotted in reverse orientation for easier comparison with the Newfoundland Basin profiles. WE is a proprietary industry profile not located in Fig. 1 that occurs north of SCR3. SCR3 is time migrated from 170–295 km and pre-stack depth migrated from 295–470 km. FGP 85-4, ER13, WE and IAM9 are post-stack time migrated. Moho is indicated as a grey dashed line on SCR3 (from Lau *et al.* 2006b) and IAM9 (from Dean *et al.* 2000; Pickup *et al.* 1996). Depth conversions were made with velocity models from coincident wide-angle surveys where present (SCR3 and IAM9) and by extrapolating the SCR3 velocity model for other Newfoundland Basin profiles. F: large faults at the shelf break, BH: basement high as referred to in the text, U: widespread strong reflector found throughout the Newfoundland Basin, L: landward dipping reflection as described in the text. Vertical exaggeration 2:1.

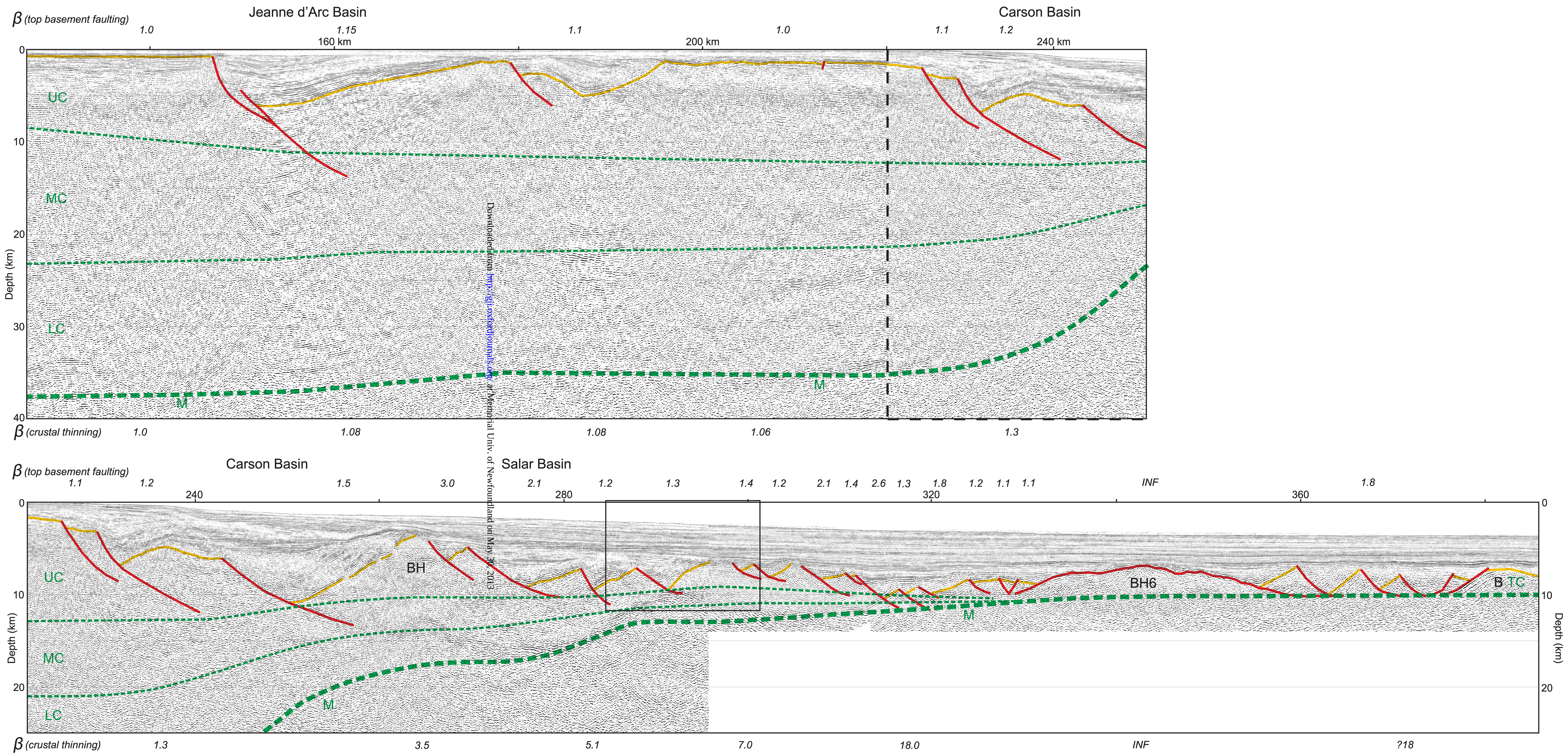

Figure 4. Interpretation of extension along SCR3. Crustal units defined according to the velocity structure in Lau *et al.* (2006a). Calculations of β according to top basement faulting and crustal thinning are located above and below the seismic plot, respectively. The depth-converted data are plotted at a vertical-to-horizontal scale of 1:1. The rectangle at 290 km highlights the area expanded in Fig. 5. β labels 'INF' refer to infinity.

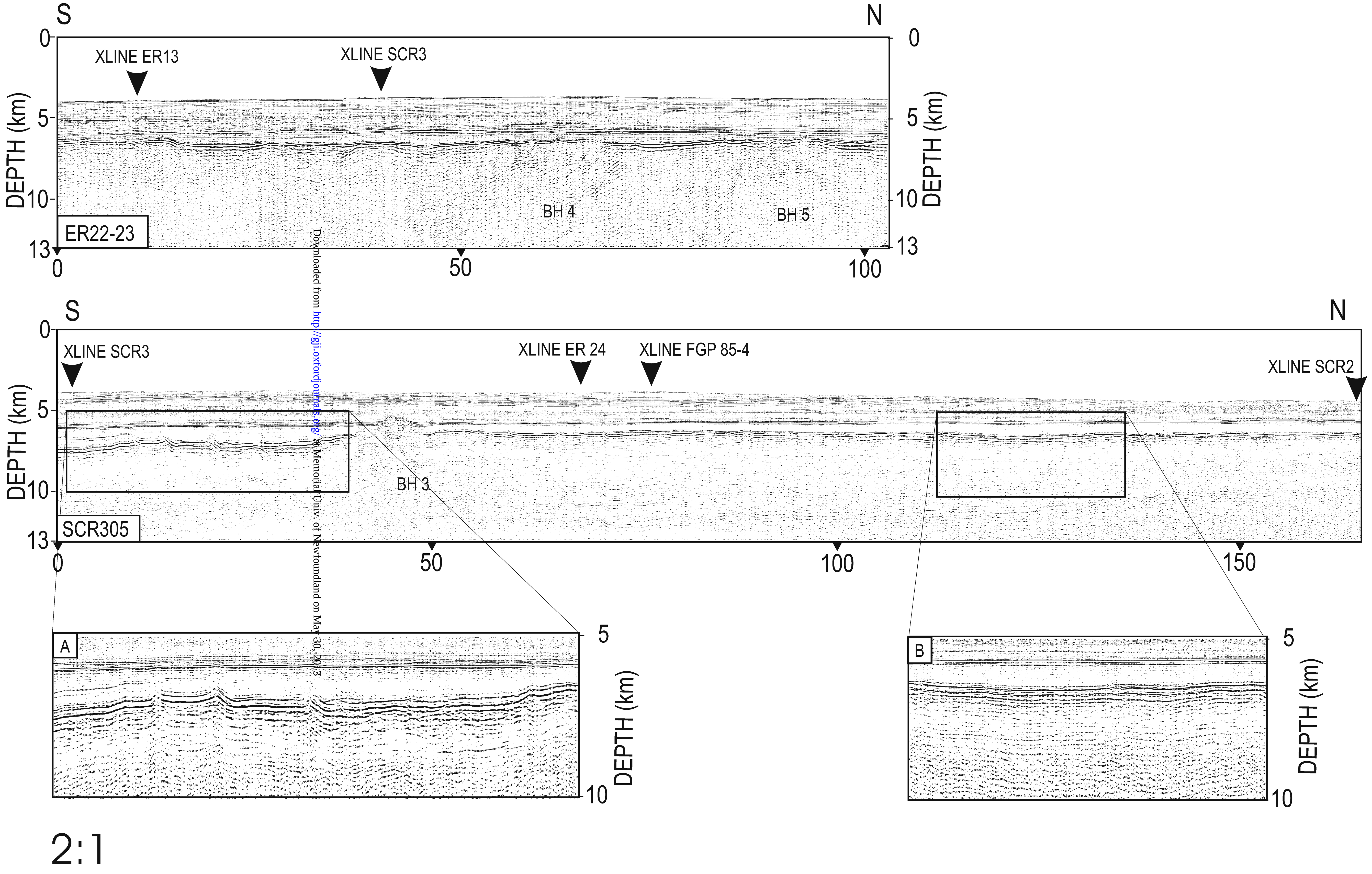

Figure 7. The transition zone in the Newfoundland Basin as imaged in profiles ER22-23 and SCR305 that are oriented perpendicular to the profiles in Fig. 3. BH: internally reflective basement highs. SCR305 closeup A illustrates the disruptions in 'U' that may originate from venting of overpressured sediments. These features are also present north of the basement high. Closeup B illustrates the complexity of the fine structure of 'U'. Vertical exaggeration 2:1.