Extension of continental crust at the margin of the eastern
Grand Banks, Newfoundland

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A B S T R A C T

Seismic and gravity observations from the rifted margin of the eastern Grand Banks, Newfoundland, support a
new model for extension of the continental crust from the shelf edge to ODP Site 1277, where mantle rocks are
exhumed. We find that the largest decrease in crustal thickness, from about 28 km to 6 km, occurs beneath
the continental slope of the Grand Banks over a distance of just 20 km. This rapid decrease in crustal thickness
coincides with anomalously high seismic velocities (70–72 km s−1) in the lower crust of the shelf edge. The
thin crust of the continent–ocean transition (COT) in this area has a smooth basement surface, void of upper
crustal blocks and prerift sediments. We compare our geophysical results with a geodynamical model that
represents rifting of a relatively hot continental lithosphere and with another numerical model that represents
ripping of a cold lithosphere. Both geodynamic models suggest that crustal thinning beneath the continental
slope was achieved by extensional faulting in the upper crust and ductile shear zones in the middle crust. The
godynamic models provide an explanation for the formation of distinct continental slopes at rifted margins:
Beneath the continental shelf of the Grand Banks, the Moho and the strong lower crust rotated upwards toward
a 50° dip without visible internal deformation. The presence of these strong lower crustal rocks at shallow
depth in the rift flank subsequently helped to localize the extension farther seaward. With ongoing extension,
some high-angle normal faults may have rotated to a sub-horizontal orientation, which would explain the lack
of brittle deformation visible in the seismic reflection data. The two geodynamic models produce different
amounts of extension of continental crust in the distal margins. The hot rifting model localizes strain much
more rapidly, leaving narrow zones of extended continental crust, and it produces a relatively large amount of
melt (>30%) in the final stages of rifting. Continental breakup may occur rapidly in hot lithosphere (<5 Myr). On
the other hand, a cold extension model extends the continental crust to a thickness smaller than 10 km over a
width of 50 km in the distal margin, similar to what we inferred at the eastern Grand Banks. The cold
lithospheric model requires about 23 Myr of extension before continental breakup, and it predicts much less
melting in the mantle (13%). The long rift duration, wide zones of thinned continental crust, and small amount
of magmatism make the cold rifting model the most applicable to Newfoundland–Iberia rift.

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1. Introduction

Geophysical data gathered across various continental rifts show
that their evolution is closely related to the development of large
normal faults (e.g., Prodehl and Mechie, 1991). Widening of the rift
may be accompanied by an increasing amount of slip on these crustal
faults, rotation of active fault planes, or the development of new
normal faults. The style of faulting will depend to a large extent on
the coupling of brittle extension with ductile deformation at larger depth
(Buck, 1991; Scholz and Contreras, 1998). If extensional faulting leads
to complete rupture of the continental lithosphere, the crustal
structure of the rifted margins should show evidence for faulting in
areas where the crustal thickness decreases rapidly in the seaward
direction.

A regional gravity inversion for the Newfoundland Basin (Welford
and Hall, 2007) shows that most of the crustal thinning at the
Newfoundland margin occurs beneath the steep continental slopes of
the Grand Banks and Flemish Cap (Fig. 1A). Spatial variations in seafloor
topography and Moho depth are roughly isostatic, so the bimodal
distribution of bathymetry in this region (Fig. 1B) reflects a rapid
transition from thick to thin crust between 1000 and 2000 m water
depth. Seismic refraction data gathered in 2000 during the Studies of
Continental Rifting and Extension on the Eastern Canadian Shelf
(SCRECH) program confirm that the continental crust thins rapidly
beneath the continental slopes offshore Newfoundland. Whereas
continental crust of the continental shelf can be as thick as 30 km,
highly extended continental crust in the adjacent Newfoundland Basin
may be less than 6 km thick (Funck et al., 2003, 2006b; Van Avendonk et al., 2006). Multichannel seismic reflection (MCS) data show that large seaward-dipping normal faults accommodated some of this extension in the upper crust near the shelf edge of the southeastern Grand Banks (Austin et al., 1989; Lau et al., 2006a), but similar seismic reflection data on SCREECH Lines 1 and 2 show little faulting in the upper continental crust (Hopper et al., 2004; Shillington et al., 2006). The apparent discrepancy between extensional faulting in the basement and whole-sale crustal thinning has also been noticed at the conjugate Iberian margin, where Reston (2005) suggested that the amount of brittle extension was obscured by different generations of faults that cross-cut and offset each other. Alternatively, some of the basement surface at these nonvolcanic rifted margins may have formed when low-angle detachment faults exhumed deep-crustal and mantle rocks during the late stages of rifting (Lavier and Manatschal, 2006; Reston, 2007). The seismic reflection data from the SCREECH program did not reveal a crustal detachment in the basement (Hopper et al., 2006) such as the S reflection at the conjugate Iberian margin (Reston et al., 1996). However, in some areas of the Newfoundland margin, the SCREECH MCS data show smooth continental basement that does not appear to be covered by prerift sediments (Hopper et al., 2006; Shillington et al., 2006). We must consider the possibility that these sections of the basement surface are the exhumed footwall of a top-basement detachment fault that accommodated crustal thinning. In this paper we aim to investigate the role of both fault-controlled extension and ductile thinning of the crust in the development of a relatively abrupt transition from thick to thin crust that we observe at the continental slope of the eastern Grand Banks along SCREECH Line 2. The relative timing of strain localization near the continental slope of the eastern Grand Banks and the breakup of the Newfoundland–Iberia rift is another goal of our study. Both these aspects depend on the rheology of the lithosphere before and after the thinning of continental crust. We will explore the relationship between extensional deformation and lithospheric rheology using geodynamic models of lithospheric extension with two different mantle temperature depth profiles.

We compare results from numerical modeling of continental rifting with new, more detailed observations of the deep structure of the continental slope of the eastern Grand Banks. The SCREECH seismic refraction profiles provided us with a good understanding of the crustal–mantle boundary. In addition, wide-angle Moho reflections from beneath the continental slope give us a better definition of the crust–mantle boundary.

2. Crustal structure of the eastern Grand Banks

Due to the apparent lack of synrift volcanism and thin sediment cover, the rifted margins of Newfoundland and Iberia have been visited by numerous marine geophysical expeditions to study their crustal
structure. Many of these studies specifically targeted the transition from rifting to the onset of seafloor spreading (Tucholke et al., 1989; Pickup et al., 1996; Chian et al., 1999; Russell and Whitmarsh, 2003). To understand the early development of the rift and the localization of strain in the distal margin, it is necessary to gather data on transects that encompass a section of un-stretched continental crust. The margins of Newfoundland lend themselves better for this than the conjugate Iberian margins, because much basement of the Grand Banks and Flemish Cap (Fig. 1A) consists of continental crust of the Avalon terrane that did not undergo a significant amount of Mesozoic extension (Welsink et al., 1989). Unfortunately, the landward end of SCREECH Line 2 does not lie on thick continental crust either, since the area between Flemish Cap and Grand Banks was also stretched by v~20% during the early stages of the Newfoundland–Iberia rift (Keen and Barrett, 1981; Van Avendonk et al., 2006). The seismic refraction studies of SCREECH Lines 1 (Funck et al., 2003) and 3 (Lau et al., 2006b), as well as Lithoprobe line 91–2 near the Newfoundland coast (Marillier et al., 1994) show that the crust of the Grand Banks and Flemish Cap was originally about 30 to 35 km thick. The relatively dense instrument spacing of the three SCREECH transects gives us fairly good constraints on the variation of crustal thickness across the margin. If the 2 to 6 km-thick crust of the Cot is continental in origin (Van Avendonk et al., 2006), the total amount of crustal thinning may be as large as β=5 at the foot of the continental slope, increasing farther into the Newfoundland Basin, with final separation of continental crust near ODP Site 1276 (Fig. 2).

The crystalline crust of the Avalon terrane has a seismic velocity somewhat less than 6.0 km/s near the top of the basement, and it increases to about 7.0 km/s at the Moho (Marillier et al., 1994). The upper crustal velocities are consistent with the metamorphic and granitic rocks found in outcrops on the Grand Banks and Flemish Cap (Durling et al., 1987). A seismic velocity of 7.0 km·s⁻¹ in the lowermost crust is slightly less than what we would expect for a mafic composition (Rudnick and Fountain, 1995). This may be explained by a laterally homogeneous delamination of mafic lower crust during the amalgamation of the Paleozoic terranes surrounding the Newfoundland–Iberia rift (Hughes et al., 1994). The seismic velocities of 7.0 to 7.2 km·s⁻¹ that Van Avendonk et al. (2006) imaged in the lower crust beneath the continental slope are higher than those found elsewhere in the Avalon terrane offshore Newfoundland (Marillier et al., 1994; Funck et al., 2003; Lau et al., 2006b). Since this high-velocity anomaly lies outside the zone of extreme thinning, Van Avendonk et al. (2006) interpreted it as a mafic body that was emplaced in the lower crust long before rifting started. Similar gabbroic lenses in the lower crust have been observed in the Ivena zone in the Alps, where they were exhumed in the footwall of a major shear zone (e.g., the Pogallo shear zone) beneath thinned upper crust in the distal margin (Handy and Zingg, 1991; Manatschal, 2004) during the final stages of rifting. The localization of shear zones at the edges of strong, mafic lower crustal bodies may explain their current presence near the transition from thick to thin continental crust in rifted margins (Fig. 2).

3. Gravity analysis

The seismic velocity model of Van Avendonk et al. (2006) shows nearly complete coverage of the crustal structure along SCREECH Line 2. Using the 8 km·s⁻¹ velocity contour as our best estimate for the Moho, we find a ~27 km-thick continental crust beneath the eastern Grand Banks and crust thinner than 7 km beneath the flat basement of the Cot (Fig. 2). The Moho, defined by Van Avendonk et al. (2006) as the 8 km·s⁻¹ contour in their first-arrival topography model, is not constrained by the seismic data in this section of the rifted margin, because turning waves of seismic refractions did not reach into mantle with a seismic velocity of 8 km·s⁻¹. The nature of the crust of the shelf edge of the eastern Grand Banks is clearly important if we want to understand lithospheric necking, so other methods must be applied to determine its structure.

3.1. Velocity–density model

Surface gravity data cannot provide a unique solution for crustal structure, but they can complement seismic refraction data. If variations in seismic velocity and density are correlated, large-scale features in a seismic refraction model may be verified with gravity data. Many seismic refraction studies of rifted margins (Dean et al., 2000; Funck et al., 2003; Bullock and Minshull, 2005; Klingelhofer et al., 2005; Lau et al., 2006b) use an empirical relationship between velocity and density that is based on a field and laboratory data compilation from Nafe and Drake (Ludwig et al., 1970). The Nafe–Drake data compilation shows considerable scatter in seismic velocities and densities, as indicated by the shaded area in Fig. 3, so caution is necessary with the interpretation of the gravity anomalies (Barton, 1986). Nevertheless, density models obtained through this velocity–density conversion usually fit the gravity data within 50 mGal, and the inferred density structure along these seismic transects may be adjusted locally to obtain a better fit of the gravity data. For example, in areas where resolution of the seismic data is poor, crustal thickness estimates may need to be modified to explain gravity anomalies (Dean et al., 2000; Funck et al., 2003). In the case of SCREECH Line 2, the depth of the Moho beneath the continental slope cannot be constrained with seismic refraction data alone. We therefore use gravity data to constrain the thickness of the high-velocity lower crust in this region.

If the crustal structure of the Newfoundland margin does not vary strongly perpendicular to SCREECH Line 2, we can calculate the contribution of each infinitesimal 2-D element of this seismic velocity model to the predicted gravity signal of SCREECH Line 2 (Telford et al., 1976). To avoid modeling artifacts at the landward and seaward edge of our finite profile (Fig. 2), we must also assume that the crustal density is laterally homogeneous at the edges of our model. We approximate the gravity signal from the deeper mantle by adding the gravitational effect of a conductively cooling, homogeneous half space beneath our model profile of SCREECH Line 2. Since we know the approximate age of the basement in the Newfoundland Basin (Verhoef et al., 1996) and β factors (see Fig. 3).
previous section), we can estimate the gravity contribution of the lower lithosphere by assuming it underwent pure-shear extension (McKenzie, 1978). The long-wavelength thermal correction of the density structure of the deep mantle gives us a slightly smaller gravity signal (by $\sim 10$ mGal) towards the seaward end of our profile where the lithosphere is still slightly thinner and hotter, even $\sim 110$ Myr after all the crust on SCREECH Line 2 was formed.

After we accounted for the deep lithospheric structure, we first converted the seismic velocity structures imaged along SCREECH Line 2 to densities using the Nafe–Drake relationship (Ludwig et al., 1970). Compared to laboratory measurements on basement rocks (Christensen and Mooney, 1995), the Nafe–Drake relationship has very low seismic velocities for unconsolidated rocks and sediments. The sediments along SCREECH Line 2 also have very low seismic velocities, which is presumably due to high porosity (Shipboard Scientific Party, 2004). We found that for seismic velocities around 6 km/s, a density $\sim 50$ kg·m$^{-3}$ lower than that predicted by the Nafe–Drake curve gives us a good data fit. This adjustment may reflect a predominantly granitic basement beneath the eastern Grand Banks (Fig. 3). For higher seismic velocity we follow the Nafe–Drake curve. This may not be correct if serpentinite is the dominant rock type in the basement, since this hydrated mantle rock has relatively high density (Christensen, 2004).

In order to constrain crustal thickness with the gravity data, we assume that the density increases from $3.1 \cdot 10^3$ kg·m$^{-3}$ to $3.3 \cdot 10^3$ kg·m$^{-3}$ across the Moho. If this assumption is justified, we obtain fairly good resolution of the crustal thickness (Fig. 4). A model with constant crustal thickness of $\sim 28$ km, but increasing lower crustal density as predicted by tomography (Van Avendonk et al., 2006), gives us a data fit within 10 mGal along much of our profile. A 10 km shallower mantle gives us a predicted free-air gravity that is $\sim 35$ mGal too high (Fig. 4), whereas a 10 km deeper Moho underpredicts the gravity signal by $\sim 15$ mGal. Average values of 7.1 km·s$^{-1}$ and $3.05 \cdot 10^3$ kg·m$^{-3}$ for the seismic velocity and density of the lower crust

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**Fig. 4.** Density model for SCREECH Line 2 and free-air gravity fit. Using our choice of velocity–density relationship (Fig. 3), we varied the crustal thickness beneath the continental slope to get the best fit of the free-air gravity anomaly. (A) We tested three structural models for the eastern Grand Banks: The thickness of continental crust may be nearly constant at $\sim 28$ km (solid curve for Moho), it may form a keel beneath the continental slope that reaches 38 km depth (short dash), or it may taper gradually to a Moho depth of 10 km in the Newfoundland Basin (long dash). In each case we assume that densities increase rapidly from $3.1 \cdot 10^3$ kg·m$^{-3}$ to $3.3 \cdot 10^3$ kg·m$^{-3}$ across the Moho. (B) Free-air gravity predictions (in red) for the three models for Moho depth (A) along with the data (in black).

**Fig. 5.** Calculation of isostatic balance for three models of Fig. 4. For each of the three Moho curves of Fig. 4 we calculated the pressure at 40 km depth due to the weight of the density profiles. The corresponding pressure is scale is on the left. The dash–dotted curve is the bathymetry along SCREECH Line 2, with a corresponding scale on the left. The solid, short dashed and long dashed pressure curves correspond to the Moho curves of Fig. 4.
of the continental slope correspond well with the laboratory measurements of gabbro under $p\cdot T$ conditions that are representative of 20 km depth in continental crust (Christensen and Mooney, 1995) (Fig. 3).

3.2. Isostatic balance

A realistic density model for the rifted margin must be approximately in isostatic equilibrium (Barton, 1986), since large deviations from this equilibrium cannot be sustained over geologic time. We can therefore use the weight load of our crustal density model to provide an additional check on the choice of the Moho beneath the continental slope. The offshore deepening of the Atlantic Ocean creates a shallow mass deficit that is compensated by the shallower mantle in the Newfoundland Basin. We compare the pressure at 40 km depth in our density model between 70 km, where the water depth is only 500 m, and 170 km, where the water depth already exceeds 3500 m (Fig. 5). Both at the continental shelf (70 km) and in the Newfoundland Basin (170 km), the pressure at 40 km depth is between 1.18 and 1.19 MPa. Beneath the continental slope, at 130 km, we have a small mass deficit, as indicated by lower pressures at 40 km depth (Fig. 5). A constant Moho at 28 km depth beneath the shelf (solid curve) corresponds to a drop in pressure to 1.16 GPa. A Moho that is 10 km deeper beneath the continental slope would result in a pressure lower than 1.15 GPa (short dash), while a 10 km shallower Moho keeps the pressure at 1.17 GPa beneath the continental slope. Clearly, a thinner gabbroic lens beneath the continental slope better fits the requirement of isostatic equilibrium. However, the bathymetry of the shelf edge of the Grand Banks may be regionally compensated, so we think that our model with a constant crustal thickness of 28 km at the shelf edge (solid curve) is not unreasonable.

4. Finite-difference modeling of wide-angle data

The traveltime and gravity data presented in previous sections help us define the bulk properties and size of the high-velocity body in the lower crust of the continental shelf edge beneath SCREECH Line 2. The nature of the Moho discontinuity cannot be defined with these data, but wide-angle seismic PmP reflections recorded during the SCREECH study may provide some constraints on the crust–mantle boundary. In the tomographic inversion of SCREECH Line 2 traveltimes, Van Avendonk et al. (2006) used only first-arriving basement refractions because they did not observe enough Moho reflections to explicitly model the Moho interface along the entire profile. However, some excellent wide-angle PmP arrivals can be seen in the OBS records in the vicinity of the continental slope. For example, an interesting Moho reflection can be seen in the receiver gather of Orb6 (Fig. 6). Air-gun shots that were fired down the continental slope show a strong reflection between 14 and 18 km source–receiver offset. This observation is unusual, since the Pg/Pn crossover occurs only at an offset of 19 km (Fig. 6). Clearly, this seismic arrival pattern cannot be explained by model with a constant crustal thickness. We can therefore use this receiver gather to verify our density model (Fig. 4). In addition, we want to use the wide-angle reflection to

![Fig. 6. Seismic refraction data from instrument ORB6. Between 14 and 18 km offset, the record shows a strong PmP reflection arriving almost a second behind the Pg arrival. The PmP reflection is not observed at and beyond the Pg/Pn crossover at 19 km. The black rectangle indicates the portion of data that is modeled in Fig. 8.](image-url)

![Fig. 7. Acoustic finite-difference calculation in two models, using instrument Orb6 at 95.0 km in our model as a point source. Pressure wave field snapshots at 7.6 s are superimposed in blue (positive pressure) and red (negative). (A) The crust–mantle boundary beneath the continental slope is assumed to represent a first-order seismic velocity discontinuity, where the velocity increases from 7.0–7.2 km·s$^{-1}$ to 8.0 km·s$^{-1}$. (B) In our second model, the same increase in seismic velocity occurs over a distance of ~2 km. Modeled phases are a crustal refraction (Pg), Moho reflection (PmP), mantle refraction (Pn), and a headwave from the Moho beneath the continental slope (PmP2). Distance coordinates are the same as in Fig. 4.](image-url)
importantly, the amplitude of $P_{mP}$ and $P_{mP2}$ are much weaker than in (Fig. 8C) show a poorer traveltime to our Moho discontinuity beneath the Grand Banks that dips upward 4.1. Sharp Moho discontinuity

order) in the wave equation, we assume that the simulation does not derivatives of acoustic pressure in space (11th order) and time (5th order) velocity discontinuity at the Moho in the model of Fig. 7A is necessary to reproduce the wide-angle reflections (Fig. 7). Receiver gathers are easy to model with finite-difference methods, since the data record is identical to a hypothetical shot gather where all sources and receivers reversed. In Fig. 7 we show the implementation of reciprocity in a portion of a snapshot at 7.6 s traveltine from ocean-bottom seismometer (OBS) Orb6. The point source at Orb6 is not shown in this figure. The finite-difference calculations were performed on a 20 m by 20 m grid, stepping forward in time at 0.5 ms intervals. Since we use large finite-difference operators for the derivatives of acoustic pressure in space (11th order) and time (5th order) in the wave equation, we assume that the simulation does not suffer much from grid dispersion.

4.1. Sharp Moho discontinuity

For our first calculation, we adopted the velocity and density model of the previous section. This model has a relatively flat, sharp Moho discontinuity beneath the Grand Banks that dips upward toward the Newfoundland Basin (Fig. 7A). In this model with a sharp Moho (Fig. 7A), the snapshot at 7.6 s shows $P_g$ and $P_n$ arrivals with positive amplitude at $\sim$135 km. The amplitude of the wide-angle reflection from the flat Moho beneath the Grand Banks ($P_{mP}$) is also positive, since the impedance increases across the crust–mantle boundary. As the $P_{mP}$ arrival impinges on the dipping Moho, a head wave ($P_{mP2}$) traveling along this interface refracts back to the surface of our velocity model at a steep incidence angle (Fig. 7A). When we compare the data (Figs. 6 and 8A) and the finite-difference solution for our first model (Fig. 7A), we find an excellent fit of all the wide-angle arrivals (Fig. 8B). Both the OBS data and synthetics (Fig. 8A and B) show a strong, first-arriving crustal refraction $P_g$ from the instrument location to $\sim$18 km source–receiver offset, where it is overtaken by a very weak $P_n$ mantle refraction. The strong wide-angle $P_{mP}$ reflection between offsets 14 to 18 km is also reproduced well with our first seismic velocity model (Figs. 7A and 8B). Although the $P_{mP2}$ arrival is not very clear in the data (Fig. 8B), it appears that we have also modeled this phase which grazes the Moho beneath the continental slope (Fig. 8B).

4.2. Smooth Moho transition zone

Clearly, our seismic velocity model with the rapidly shallowing Moho beneath the continental slope is consistent with the wide-angle $P_{mP}$ reflection observed on instrument ORB6. In our second finite-difference model (Figs. 7B and 8C), we test whether the sharp first-order velocity discontinuity at the Moho in the model of Fig. 7A is necessary to reproduce the wide-angle reflections. We repeated the calculation for a model where the seismic velocities beneath the seafloor have been spatially averaged over $\sim$2 km (Fig. 7B). Compared to our first calculation, the resultant synthetic data of our second test (Fig. 8C) show a poorer traveltime fit of all wide-angle arrivals. More importantly, the amplitude of $P_{mP}$ and $P_{mP2}$ are much weaker than in the previous calculation.

Together with the tomographic inversion (Van Avendonk et al., 2006) and the gravity modeling (previous section), we think that the finite-difference calculation provides convincing confirmation of the deep structure of the continental slope at SCREECH Line 2. The complex Moho reflections recorded by instrument Orb6 are consistent with a seismic velocity of $\sim$7.1 km·s$^{-1}$ in the lower crust of the continental slope, and the data suggest that the Moho here dips $\sim$50° toward the Grand Banks. The finite-difference calculation shows that the crust–mantle boundary beneath the continental slope is a sharp contact.

5. Pre-rift sediments

The deep margin of Galicia Bank and adjacent Iberia Abyssal Plain comprise blocks of faulted and attenuated continental crust capped by prerift and synrift sediments (Mauffret and Montadert, 1988; Wilson et al., 2001; Péron-Pinvidic et al., 2007). In contrast, seismic data from the Newfoundland Basin show little evidence for faulted upper continental crust or prerift sediments in the deep margin (e.g., Shillington et al., 2006). The paucity of recognizable continental upper crust in the COT of the Newfoundland margin may in part be due to the masking effect of the U reflection (Tucholke et al., 1989). This prominent
seismic horizon in the Newfoundland Basin is, at least locally, caused by the intrusion of diabase sills 10–30 Myr after continental breakup between Newfoundland and Iberia (Karner and Shillington, 2005; Hart and Blusztajn, 2006; Shillington et al., 2007). A frequency analysis shows that these igneous sills severely limit the seismic depth penetration beneath the U reflection (D.J. Shillington et al., Seismic signal penetration beneath post-rift sills on the Newfoundland rifted margin, manuscript submitted to Geophysics, 2007). In this section we review evidence that shows that it is unlikely that the thin crust beneath the U reflection in the Newfoundland Basin (dashed line in Fig. 2) is covered with prerift sediments, because there is very little space, perhaps just 200 m, between the basement rocks and oldest postrift sediments cored at Site 1276 (Shipboard Scientific Party, 2004). However, there is evidence of prerift sediments on SCREECH Line 2 closer to the continental slope (Shillington et al., 2006). The distribution of prerift sediments offshore Newfoundland and Iberia are an important constraint for conceptual models of rifting and breakup of these conjugate margins. We assume that prerift sediments overlie continental upper crust that remained intact during rifting, and the absence of prerift sediments in parts of the deep margins suggests that basement surface here was created by exhumation of crustal or mantle rocks, or by incipient seafloor spreading.

The oldest sediments encountered in the 1740 m section recovered at ODP Site 1276 are no older than latest Aptian or earliest Albian age based on shipboard biostratigraphy (Shipboard Scientific Party, 2004). According to kinematic plate reconstructions (Srivastava et al., 2000) the basement beneath Site 1276 is older than anomaly M3 (Verhoef et al., 1996; Shillington et al., 2004), or ~18 Myr older than the oldest sediments cored at Site 1276. The large amplitude of the U reflection makes it difficult to establish a relationship between the postrift sediments and underlying basement. However, both MCS (Shillington et al., 2006) and wide-angle seismic reflections from OGS GSCA-1 on SCREECH Line 2 (Fig. 9) show that the basement lies only 100 to 200 m deeper than the mafic sills of the U reflection. The refraction data from GSCA-1 also show that the seismic velocity of the basement surface beneath ODP Site 1276 is higher than 5.5 km s⁻¹ near Site 1276 (Fig. 7). At these shallow depths, this seismic velocity could represent a granitic or gabbroic crustal composition (Christensen and Mooney, 1995), or the basement may consist of pervasively serpentinized mantle rock (Christensen, 2004). Given that there is not much space for Aptian and older sediments between the U reflection and the crystalline basement, we consider the possibility that the basement in the vicinity of ODP hole 1276 on SCREECH Line 2 (Fig. 2) may have formed by exhumation of deep-crustal or upper-mantle rocks relatively late in the evolution of the Newfoundland–Iberia rift. Late Aptian, Albian, and younger sediments may have been deposited directly on the new basement surface.

The most seaward block of unambiguous continental upper crust lies between the continental slope and the “flat” transitional basement on SCREECH Line 2 (between 155 and 165 km in Figs. 2 and 10A). Strike line SCREECH 301 indicates that this feature has a lateral dimension of approximately 10 km across SCREECH Line 2 as well (Fig. 10B). The block shows a clear Moho reflection ~7 km beneath the basement surface. The basement is topped by a ~1.5-km-thick stratified sequence at depths of 4–6 km below sea level (Fig. 10A), interpreted to represent prerift sediments by Shillington et al. (2006). This sequence appears to be unconformably overlain by young postrift sediments; the boundary between these units is marked by a bright reflection. Likewise, the contact between the layered sequence and the underlying crystalline basement is also marked by a ~0.5-km-thick package of bright reflections that appear to be offset by one or more faults.

A possible alternative interpretation for the layered sequence atop the most seaward continental block is that it corresponds to igneous stratigraphy. The occurrence of a magnetic high in this region (Verhoef et al., 1996) suggests that some magmatic addition could have occurred in this location. However, the velocities within the layered

![Fig. 9](image-url). Detail of wide-angle receiver gather from OBS GSCA-1 with predicted traveltimes in the velocity model of Van Avendonk et al. (2006) and corresponding ray paths. (A) Data fit. Just ~150 ms before basement reflection (in red) we see a strong reflection in the wide-angle data that may represent the U reflection (yellow dashed). (B) Velocity model and ray diagram. We infer that the U reflection lies just ~200 m above the basement surface, which locally has seismic velocities between 5.7 and 6.3 km s⁻¹.
sequence determined from prestack depth migration (2.7–3.7 km s$^{-1}$, Fig. 10) are most likely too low to correspond to extrusive volcanics. For example, PSTDV velocity analysis of a crustal block with interpreted igneous stratigraphy on SCREECH 1 yields velocities of $\sim$4.3 km s$^{-1}$ (Hopper et al., 2004). A compositional change in the crust between this continental block and the adjacent COT, together with an abrupt 1.5 km change in basement depth, can also explain the local magnetic anomaly (Verhoef et al., 1996).

6. Structural development of the rift

The main purpose of our study is to understand the development of variations in the thickness of continental crust that we observe across the margin of the eastern Grand Banks of Newfoundland. In the next section, we will attempt to explain some of our geophysical observations with a geodynamic model of lithospheric extension and continental breakup. We will parameterize this model using physical properties of the lithosphere in the Newfoundland–Iberia rift. First, we formulate five key issues regarding the development of the most important structural features of SCREECH Line 2 and the Newfoundland–Iberia rift that we seek to address:

1) Our new analysis of seismic and gravity data along SCREECH Line 2 confirms that the continental crust that lies just seaward of the continental slope is much thinner (~6 km) than the crust of the shelf edge (~25 km). The high-velocity and high-density anomaly in the lower crust beneath the shelf edge suggests that we must consider the role of a strong, possibly gabbroic, lower crust in the development of the rift. However, it is not yet clear whether the 50° dip of the Moho beneath the continental slope formed by rotation of a flat-lying continental Moho, or whether the base of the lower crust was offset by a normal fault that dipped beneath the western flank of the Newfoundland–Iberia rift.

2) We found good evidence for prerift sediments on SCREECH Line 2 lying on top of a 10-km-wide crustal block at the base of the continental slope, but the continental shelf edge, most of the continental slope, and the basement beneath the U reflection all appear void of prerift and synrift sediments. The absence of older sediments in shallow areas of the Grand Banks may be due to erosion during the Avalon uplift (Grant et al., 1988; Marsaglia et al., 2007). We will look into the possibility that the basement surface of the deeper continental slope, currently at 3000 m, was formed by exhumation of continental middle crust. The seismic structure of the basement beneath the U reflection between 165 km and 225 km in our model (Fig. 2) is also consistent with that of exhumed middle continental crust. At 225 km in our seismic velocity model (Fig. 9) a profound lateral change in the seismic velocity structure may mark the seaward exhumation of serpentinized mantle. These serpentinites were drilled in Site 1277 at 270 km (Fig. 2).

3) The presence of high mantle seismic velocities (~8 km s$^{-1}$) beneath the thin crust of the COT between 155 and 225 km in our model of SCREECH Line 2 is a fairly unique observation in the margins of Newfoundland and Iberia, suggesting that the mantle in this particular area was not serpentinized. The mantle emplaced here may have been too hot to serpentinize (~500 °C; Ulmer and Trommsdorff, 1995), or it may not have been in contact with percolating seawater during extension and exhumation. The high fluid fluxes required to serpentinize mantle rocks may only be sustained where fault zones extend into the mantle, such as those studied at the modern Mid-Atlantic Ridge (Boschi et al., 2006).

4) SCREECH Line 2 is approximately conjugate to the Lusigal 12 profile in the Iberia Abyssal Plain (Krawczyk et al., 1996; Chian et al., 1999). Upper-crustal allochthons with prerift and synrift sediments are abundant on the Iberian side, but they are absent from the smooth basement of the COT along SCREECH Line 2. This asymmetry between the two rift flanks may have been caused by movement of these upper crustal blocks along low-angle detachment faults during the final stages of rifting. The origin and evolution of such low-angle faults are important for our understanding of continental breakup (Froitzheim and Eberli, 1990; Manatschal et al., 2006).

5) Our geophysical constraints on the deep structure of the Newfoundland margin reveal the rifting architecture at the final stage of opening of the incipient ocean basin. These constraints are important, but we must also consider the duration of the rift from the onset of extension to continental breakup and the formation of a mid-ocean ridge. Evidence for widely distributed extension in the sedimentary basins of the continental shelves of Newfoundland and western Iberia dates back to the Late Triassic and Early Jurassic (Tankard and Welsink, 1989; Tucholke et al., 2007) with an increased extension commenced around Oxfordian time (155–160 Ma according to the time scale of Gradstein et al., 2004). Synrift sediments were initially distributed to marginal basins far from the later rift axis (Wilson et al., 1988), but around late Berriasian time (140 Ma) extension localized in the current deep-sea margins of Newfoundland and Iberia. This development of the rift is also evidenced by the subsidence history. Shallow-water carbonates formed off western Iberia in the Tithonian (145–150 Ma), and their succession by deep-water carbonates and clastic sediments in the Valanginian and Hauterivian (130–140 Ma) indicates rapid tectonic subsidence during this time (Wilson et al., 1996). In the late stages of
the rift, subcontinental mantle may have been exhumed to the seafloor of the Iberia Abyssal Plain as early as Berriasian time (142 Ma) according to 40Ar/39Ar dating (Jagoutz et al., 2007). If continental crust was as thick as 20 km in the Tithonian, most of the crustal thinning may have taken place in just 5 Myr, from the Tithonian (147 Ma) to Berriasian time (147 Ma) (Wilson et al., 2001). On the other hand, Manatschal et al. (2001) found evidence that the Tithonian shallow-water carbonates were emplaced on crust that was less than 10 km thick, in which case the crustal thinning episode that lead to crustal separation may have been longer, perhaps as long as 20 Myr.

7. Geodynamic modeling

7.1. Strategy

The structure of the rifted margins of Newfoundland and Iberia and the timing of the rifting events, however poorly constrained, may be explained with a geodynamic model for lithospheric extension. We seek to compare two geodynamic end member models for continental rifting that represent 1) rapid extension of weak continental lithosphere and 2) slow extension of strong lithosphere. In the case of both of these scenarios it is our goal to achieve a rift duration and structural features that are comparable to what we observe offshore Newfoundland and Iberia.

For a starting model we take a one-dimensional structure with a rheology that is consistent with the density and seismic velocity of the far landward edge of our model of SCREECH 2 (Fig. 4), since this area experienced little extension (Fig. 11A). The crustal thickness in this model is 30 km. The continental crust has a density of ∼2750 kg·m⁻³, consistent with a granitic composition, to a depth of 22 km. The lower crust consists of a 6-km-thick, possibly gabbroic layer with a density of ∼2950 kg·m⁻³, which is similar to what we imaged beneath the edge of the shelf at SCREECH Line 2. The upper mantle has a density of ∼3250 kg·m⁻³, which is lower than the density of the flanks at the rift zone (3300 kg·m⁻³). We introduced different mantle structure with a larger strength at the edges of the model space to make sure that extension and rifting will occur at sufficient distance from these edges.

Since the strength of the lithosphere is largely governed by the temperature, we choose a thermal age of 100 Myr for our weak model, and we choose a 250 Myr old continental lithosphere for our strong model (Fig. 12). Given that we assume that the final phase of extension in the Newfoundland–Iberia rift commenced in the Late Jurassic (∼160 Ma), these ages correspond to amalgamation of the continental lithosphere either at the end of the Alleghanian (∼260 Ma) or at the end of the Acadian (∼410 Ma) orogeny. For these two scenarios, we calculate the initial temperature profiles using a Fourier transform solution of the heat equation with radiogenic heat production in the top 10 km of the crust and top and bottom boundary temperature conditions of 10 °C and 1330 °C, respectively. Due to the longer duration of radiogenic heat production, the shallow crust of 250-My-old lithosphere is hotter in the upper crust than the 100 Myr old lithosphere (Steckler and ten Brink, 1986). However, the lower crust and upper mantle of the old lithosphere is significantly colder and stronger due to conductive heat loss. In both models, the upper crust will experience strain-dependent weakening which does not depend strongly on temperature. We calculate the melt production in our models during extension (Phipps Morgan, 2001) using the lithus plotted in Fig. 12. Clearly, our 100-Myr-old lithosphere model has a much greater potential for melt production than the model for a 250 Myr old continent. The viscosities of the two models (Fig. 11B and C) are based on the temperature profiles (Fig. 12) and a power-law rheology in the lower crust and upper mantle.

For our first geodynamic model we seek a relatively high extension half rate of 4.5 mm/yr in our cold lithosphere model. These two extension rates span the range of what may be considered realistic for the Newfoundland–Iberia rift. Extension rates in the Newfoundland–Iberia rift are poorly constrained since all magnetic anomalies older than C34 (84 Ma) have low amplitudes (Verhoef et al., 1996). A reconstruction of the margins requires an average extension or spreading half rate of about 6.7 mm/yr in the Late Jurassic, although correlation of these magnetic anomalies indicates that these extension rates may vary locally and with age (Srivastava et al.,...
Indeed, a deep-tow magnetic study of the earliest seafloor spreading anomalies of the Iberia Abyssal Plain inferred a spreading half rate of 10 mm/yr (Whitmarsh and Miles, 1995).

Our first geodynamical model, with high temperatures and a high extension rate, has the greatest potential for mantle melting during deformation, whereas the second case is least likely to experience melting (Pedersen and Ro, 1992; Bown and White, 1995; Minshull et al., 2001). We expect rapid strain localization, separation of continental crust, and continental breakup in our first model, whereas breakup should be much more delayed in our second model. It is therefore important to compare the duration of rifting in these models with Newfoundland and Iberia, where rifting proceeded to mantle exhumation after about 5 to 20 Myr (see previous section). For each of these two geodynamic scenarios we will also examine whether they can produce deformation patterns that are similar to our geophysical observations of the Newfoundland–Iberia rift.

Besides the temperature structure and extension rate, the outcome of geodynamic modeling will strongly depend on the assumed feedback between deformation and rheology. Recent numerical studies of continental rifting show that geologically realistic models of continental rifting can address the nature of strain softening in the Earth’s lithosphere (Huismans and Beaumont, 2002, 2007; Nagel and Buck, 2004; Lavie and Manatschal, 2006). These models sometimes predict different rift morphologies, but the localization of strain in shear zones always plays an important role in the thinning and breakup of continental margins. Two additional physical processes that may strongly influence the style of rifting are 1) the thermal state of the crust and mantle (Hirth and Kohlstedt, 1995) and 2) the hydration of the mantle (Pérez-Gussiény and Reston, 2001). We use the modeling code for extension developed by (Lavie and Manatschal, 2006) who parameterized the rheology to allow for the development of ductile shear zones in the brittle and semi-brittle crust. The yield criterion is based on Freudenthal’s critical plastic work criterion (Freudenthal, 1950; Goijaerts et al., 2000) that depends on both stress and strain. To account for the fact that localization is occurring in polymetallic rocks such as granite, we weaken the material strength from that of the strongest phase (plagioclase) to that of weakest phase (quartz) when the yield criterion is reached (Lavie and Manatschal, 2006). In areas of high strain in our geodynamic model, plastic deformation is governed by the Mohr–Coulomb yield criterion in a frictional and cohesive medium. For viscous deformation, we use power-law creep to calculate the effective viscosity for quartz, plagioclase, and olivine in a Maxwell visco-elastic constitutive relation (see Lavie and Manatschal, 2006 for details).

Fig. 13. Strain accumulated in our first geodynamic model, which represents rifting with a relatively hot geotherm. Time progresses from top to bottom as crust thins from 30 km to ~10 km after 2.5 million years. Left panels show viscous strain and right panels represent plastic strain accumulation during extension. (B) We trigger the development of a large normal fault by introducing a weak lineation at t=0.0 Myr. Black contours represent the 2.8·10^3 kg⋅m^−3, 3.0·10^3 kg⋅m^−3, and 3.2·10^3 kg⋅m^−3 density contours.
7.2. Results for hot and fast-rifting model

At the start of extension we introduce a weak fault zone (low friction, low cohesion) in the lithosphere that helps to focus the deformation in the central portion of our numerical model (Fig. 13B). At the onset of extension (Fig. 13C–D), both left-dipping and right-dipping high-angle normal faults develop throughout the lithosphere. These faults accumulate both viscous and plastic strain (Fig. 13C), although the lower portion of the granitic crustal layer mainly supports viscous strain since the formation of ductile shear zones develops a semi-brittle layer there (Fig. 13D). With increasing time, (Fig. 13E–F), plastic deformation dominates in the shear zones through the upper crust, whereas viscous strain dominates in the semi-brittle layer. Viscous deformation is also more pervasive in the gabbroic lower crust and upper mantle.

At 1.5 Myr in our first numerical model, the viscous shear zone in the deeper lithosphere has breached the lower crust in the center of the rift (Fig. 13E–F). Viscous shearing now focuses on the widening sub-horizontal boundary between the granitic upper crust and the underlying mantle. In the upper crust above, shear zones that dip towards the center of the rift grow larger than faults that dip away from it (Fig. 13E–F). With increasing amounts of extension (Fig. 13G–J), the crust becomes thinner, and the normal faults in the upper crust rotate to a shallower dip angle. Faults with long offsets start to develop a downward-concave shape, but the active portion of these faults remains steep. The seaward-dipping normal faults isolate a large triangular upper crustal block in the center of the rift (at 120 km) that is identified as “block H” in the model of Lavie and Manatschal (2006).

We find that the style of rifting is influenced over time by a decrease in the viscosity of the granitic middle crust. The coincidence of linear, steeply dipping low-viscosity bands and shear zones in our model shows that the deeper portion of the granitic layer is progressively weakened with accumulating strain (Fig. 14). At 2.0 Myr (Fig. 14C), a widespread low-viscosity channel has formed in the middle crust that allows extension of this weak layer towards the center of the rift by means of ductile shearing. The thinning of the crust near the rift flanks is therefore mostly due to brittle deformation in the upper crust and ductile shearing in the middle and lower crust. Pure-shear stretching of the uppermost mantle occurs mostly in the center of the rift, where the lower crust has been breached.

At 3.5 Myr, the mid-crustal low-viscosity channel starts to pinch out between the cool, brittle upper crust and the gabbroic lower crust (at the rift flank at 80 km in Fig. 14E). At the same time, the locus of crustal extension shifts from the left rift flank to the right rift flank. The high temperature of the rising asthenosphere does not appear to cause crustal weakening during the development of the rift flanks (Fig. 14D), but it may facilitate crustal deformation in the final stages of rifting (Fig. 14F).

After the continental crust has stretched to less than 10 km in a wide area, the rift is characterized by a rotation of both upper-crustal faults and lower crustal faults to a more sub-horizontal position, at the same time as these faults become inactive (Fig. 15). The offset on upper-crustal detachment faults exposes mid-crustal rocks at the basement surface of the rift flanks over a distance of 20–40 km. One of the upper crustal detachments and the underlying mantle shear zone merge together at 4.0 Myr (Fig. 15E), which leads to separation of the continental crust in this location (Fig. 15G). The central block H of continental crust is left stranded near the left rift flank where the lower crust was exhumed to its shallowest level (at a distance of 80 km in Fig. 15E). The presence of strong lower crust at relatively shallow levels may have caused the rift to fail farther to the right.

Our first numerical model with a hot geotherm predicts exhumation of mantle rocks at both margins during the final stages of rifting (Figs. 15I–L). In our model parameterization, mantle exposed to the surface is serpenitized, which makes it brittle and lowers its density to between $2.8 \cdot 10^{3} \text{ kg m}^{-3}$ and $3.0 \cdot 10^{3} \text{ kg m}^{-3}$. Adiabatic decompression of the rising asthenospheric mantle leads to a very substantial amount of melting (>30% in Fig. 15K–L). The geodynamic modeling code (Lavier and Manatschal, 2006) does not account for melt extraction from the sublithospheric mantle. The melt percentages given in Fig. 15 therefore reflect the amount of melt that is locally produced instead of the amount of melt that is retained deep in the mantle. The results from our geodynamic model with a hot geotherm indicate that mantle material that is exhumed to the seafloor after 5.0 Myr (Fig. 15I–J) is infiltrated with melts. Assuming that these melts migrate to shallower levels, they should form extrusive volcanics and gabbroic plutons in the incipient ocean basin.

7.3. Results for cold and slow rifting model

In our second numerical model we apply the lower mantle temperature profile (Fig. 12) to see if we can produce realistic rifted margin structure after a much longer episode of continental extension.

![Fig. 14. Viscosity (A,C,E) and temperature (B,D,F) of the three-layer geodynamic rift model that represents hot rifting at three different times in the evolution. Density contours (black) are as in Fig. 11. (E,F) White contour represents a melt fraction of 10%](image)
under cool conditions. For this simulation we extend the model of Fig. 11C at a constant half rate of 4.5 mm/yr. In Fig. 16 we show the development of plastic strain in 5 Myr intervals between 0 and 20 Myr. Similar to our hot rifting model (Figs. 13 and 15), widely distributed strain initially localizes in high-angle normal faults in the upper crust and also in the lower crust and upper mantle (Fig. 16A and B). The granitic middle crust at ~15 km depth is therefore again a weak zone that prevents the early formation of whole-lithospheric faults, just as we observed in our first numerical model (Fig. 14C).

After 10 Myr (Fig. 16C) the cold rift model starts to produce two rift zones at the left and right side of the model, 250 km apart. Initially, the largest amount plastic strain is accumulated in the right rift arm, but by 20 Myr the portion of the model between 200 and 300 km no longer accommodates extension. The 3.2·10^3 kg·m\(^{-3}\) density contour in the mantle outlines a growing body of hot, asthenospheric mantle at ~0 km (Fig. 16D), while the volume of low-density mantle at the failed rift at ~230 km is already shrinking due to conductive heat loss.

The plastic strain pattern of the rift zone between ~80 km and 100 km in our model at 20 Myr (Fig. 16D) shares some similarities with the rift zone of the hot extension model (Fig. 15). In both rift models, the steep flanks of the rift valley (at ~60 km and 80 km in Fig. 16E) are positioned above the upturned edges of gabbroic lower crust. A large amount of plastic strain is accommodated near the surface and at the crust–mantle boundary just seaward of the continental slopes (at ~50 km and at 40 km in Fig. 16E). Though the strain distribution in the hot and cold rifting models is similar, the cold rift model has much wider zones of continental crust that is stretched to less than 10 km thickness (Fig. 16E).

The cold rifting model extends for ~20 Myr before the crust thins to less than 10 km, but subsequently the rift proceeds rapidly towards breakup. We therefore show the last few million years of rifting at 1 Myr time slices (Fig. 17). At 22.0 Myr the continental crust pinches out in the center of rift zone, which so far has not yet experienced much plastic strain (Fig. 17B). On either side of the rift axis we now observe areas of high plastic strain just seaward of the continental slopes (Fig. 17C–E). As mantle is exhumed to the seafloor after 23 Myr of extension at 4.5 mm/yr (Fig. 17D and E), asthenospheric melts start to form at depths larger than 30 km. In contrast, in the hot rift model (extending at 12 mm/yr) melts formed well before the continental crust was breached by rifting (Fig. 15).

8. Discussion

8.1. Formation of continental slopes

We presented seismic and gravity data that help to better constrain the crustal structure of the eastern Grand Banks, Newfoundland along
Besides the drastic change in crustal thickness across the continental slope, we have fairly good evidence for a high-velocity (7.0–7.2 km/s), high-density (3.0·10^3 kg m^{-3}) anomaly beneath the shelf edge of the eastern Grand Banks. Most Avalon crust has velocities ranging between 6.0 and 7.0 km/s, which is consistent with a more granitic or perhaps andesitic composition (Christensen and Mooney, 1995; Behn and Kelemen, 2003). The high-velocity anomaly beneath the shelf edge may represent a more mafic, possibly gabbroic composition. High-velocity anomalies at rifted margins are typically interpreted as magmatic underplating or mantle serpentinization. Decompression melting of unusually fertile or hot mantle can lead to voluminous magmatic additions to the crust (White et al., 1987). Most of these melt products intrude and modify the crust of the COT, creating a layer whose density and seismic velocity typically lies between those of continental crust and mantle. The seismic velocity and density structure of the COT of SCREECH Line 2 (Fig. 4) are not compatible with a large amount of magmatic underplating in the COT during rifting. The well-defined crust–mantle boundary beneath the flat basement of the COT on SCREECH Line 2 (Fig. 4) suggests that mantle melting and magmatic underplating were limited, despite the fact that the distal margin must have experienced a large amount of extension. From the analysis of peridotite samples at ODP Site 1277, Müntener and Manatschal (2006) concluded that not much melt was generated during Mesozoic rifting at the Newfoundland margin because the continental mantle remained relatively cool during rifting, and it had already been depleted during a Paleozoic subduction event. We therefore consider it unlikely that the 7.0–7.2 km/s anomaly beneath the shelf edge of the Grand Banks represents magmatic intrusives that were emplaced during Mesozoic extension. At a depth of ~15 km, the crustal high-density anomaly of SCREECH Line 2 was probably not caused by serpentinization either, since seawater most likely did not percolate through the thick continental crust at this early stage of rifting (Pérez-Gussinyé and Reston, 2001). We therefore assume that the gabbroic lower crust is of older, probably Hercynian origin (Matte, 2001). Lower crustal rocks from the Iberian continent have also been found by ODP drilling in the Iberia Abyssal Plain margin (Manatschal et al., 2001), and high seismic velocity anomalies in the Voring margin have similarly been interpreted as mobilized Caledonian lower crust (Gernigon et al., 2004).

Assuming that the crust of the Avalon terrane was approximately of constant thickness prior to rifting, the 50° landward dip of the Moho beneath the continental slope requires an abrupt seaward increase in the amount of overall crustal thinning. Fault zones play a critical role in thinning the crust in our numerical rifting model, but neither of our two geodynamic models suggests that the Moho beneath the continental slope is a tectonic boundary. For example, in the hot rifting model (Fig. 13), the Moho beneath both rift flanks undergoes rigid rotation, rather than internal deformation, to achieve
a similar ~50° dip. Thinning of the crust beneath the continental slope is achieved primarily by 1) brittle, extensional faulting in the upper crust along a major seaward-dipping fault, and 2) ductile shear zones in the deeper portion of granitic crust at depths of 10 km to 20 km. Extension by ductile shearing in the middle crust is interrupted when the low-viscosity channel in the granitic crustal layer pinches out between the brittle upper crust and gabbroic lower crust (Fig. 14). Our cold rifting model (Figs. 16 and 17) shows a similar development of the continental slopes. However, the distal margins of the cold and hot continental rift look quite different in our models. The zone of thin continental crust is much wider in the cold rifting model (~80 km in Fig. 17) than in the case of the hot rifting model (Fig. 15). The 80-km-wide zone of highly thinned continental crust seaward of the continental slope along SCREECH Line 2 (Fig. 4) is clearly more consistent with the cold rifting model (Fig. 17) than with the hot rifting model (Fig. 15). The failed rift on the right side of our cold rifting model (Fig. 17) resembles the Galicia Interior Basin in size and shape (Pérez-Gussinyé et al., 2003).

Seismic reflection data from the continental shelf edge of SCREECH Line 2 do not yield clear evidence for large, seaward-dipping normal faults around a major seaward-dipping fault, and ductile shear zones in the deeper portion of granitic crust at depths of 10 km to 20 km. Extension by ductile shearing in the middle crust is interrupted when the low-viscosity channel in the granitic crustal layer pinches out between the brittle upper crust and gabbroic lower crust (Fig. 14). Our cold rifting model (Figs. 16 and 17) shows a similar development of the continental slopes. However, the distal margins of the cold and hot continental rift look quite different in our models. The zone of thin continental crust is much wider in the cold rifting model (~80 km in Fig. 17) than in the case of the hot rifting model (Fig. 15). The 80-km-wide zone of highly thinned continental crust seaward of the continental slope along SCREECH Line 2 (Fig. 4) is clearly more consistent with the cold rifting model (Fig. 17) than with the hot rifting model (Fig. 15). The failed rift on the right side of our cold rifting model (Fig. 17) resembles the Galicia Interior Basin in size and shape (Pérez-Gussinyé et al., 2003).

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8.2. Isolated block of continental crust with prerift sediments

The MCS data from SCREECH Line 2 (Fig. 10) show that despite the large amount of extension ($\beta = 5$) at the base of the continental slope, a block of continental crust with prerift sediments and a distinct Moho remained intact between 150 and 165 km in our model (Shillington et al., 2006). In Fig. 18A we have marked the Moho imaged in the MCS data with diagonal crosses. Although the fragment of continental crust is fairly small, and perhaps of limited extent along the strike of the margin (Fig. 10B), we do not discount it in our interpretation of the geophysical results. In the kinematic model that we propose for the eastern Grand Banks (Fig. 18), we assume that the prerift sediments and the continental Moho at the base of the continental slope are preserved during extension of the margin.

During the thinning of continental crust, the development of seaward-dipping normal faults at both rift flanks can lead to the isolation of a relatively undeformed wedge of upper-crustal rocks with prerift sediments, which is block H in the model of Lavier and Manatschal (2006) and in our two geodynamic models. As suggested by Scholz and Contreras (1998), the ultimate location of this triangular wedge of crustal material will depend on which normal fault develops into a lithospheric shear zone. The crustal block with prerift sediments at the base of the continental slope of SCREECH Line 2 (Fig. 10) may be a small remnant of a larger piece of continental crust that is similar to block H in the two numerical models (Figs. 15 and 17).

In our geodynamic model for rapid extension (12 mm·yr$^{-1}$ half rate) of relatively hot lithosphere we find that continental breakup will leave this crustal block stranded at either one of the two rifted margins (Fig. 15). The hot extension model predicts relatively little stretching of the continental crust in the final stages of rifting (Fig. 15), but block H may be reduced in size by tectonic erosion if one or both of its bounding normal faults develop into a rolling-hinge fault (Lavier and Manatschal, 2006). This type of low-angle fault can break crustal allochthons from the head wall of block H and transport them farther into the rift basin (Lavier and Manatschal, 2006). In our cold geodynamic rifting model (Fig. 17) detachment faults can isolate a piece of continental upper crust in a rift zone in a similar fashion, although the final phase of extension in this model is accompanied with a larger amount of pure-shear extension than in our hot, fast-rifting model.

The Moho that is preserved in the crustal block at 155–160 km at the base of the continental slope (Fig. 10) would be difficult to reconcile with the presence of a rolling-hinge fault that is rooted in the mantle. The absence of prerift sediments on both sides of the crustal block can still be explained by rolling-hinge detachment faults on both sides, but we propose that the detachment faults rooted in the middle of the continental crust. With ongoing extension, the two shear zones extruded middle crustal rocks both in a landward (Fig. 18C) and seaward (Fig. 18D) direction, until the block of continental crust was diminished in size, with a thickness of just 7 km from the top of the basement to the Moho (Fig. 18A).

8.3. Evolution of the distal margins

The nature of the basement between the most seaward portion of prerift sediments at 165 km in our model (Fig. 10) and the serpentinitized mantle with gabbroic intrusives of Site 1277 at 270 km (Robertson, 2007), has been discussed extensively in the recent literature. As in many other studies of rifted margins, Shillington et al. (2006) weighed the evidence for 1) a continental affinity of the basement of the COT, 2) basement composed of exhumed mantle, and 3) the possibility that the basement beneath the U reflection may be formed by ultraslow seafloor spreading. The
results from Site 1277 clearly indicate mantle exhumation at the seaward end of the COT of SCREECH Line 2 (Robertson, 2007; Tucholke and Sibuet, 2007), but the seismic velocity image on SCREECH Line 2 (Van Avendonk et al., 2006) is consistent with continental crust on the landward side of the COT (165–230 km in Fig. 2 and in Fig. 18).

Although Van Avendonk et al. (2006) and Shillington et al. (2006) inferred continental crust beneath the U reflection in the COT of SCREECH Line 2, it appears that the basement here is not covered with prerift sediments. By modeling the seismic refractions from the flat basement of the COT along SCREECH Line 2 (Fig. 9), we were able to determine the depth of the crystalline basement beneath the U reflection. After comparison with multichannel seismic reflection images (Shillington et al., 2006) and the sedimentary stratigraphy at ODP Site 1276 (Shillington et al., 2007), we found that the U reflection lies at a depth no more than 200 m above crystalline basement. It is therefore unlikely that the basement beneath the U reflection is covered with a significant amount of prerift or synrift sediments. To explain the lack of older sediments, we interpret the smooth basement of the COT as an 80-km-long, flat-lying detachment fault (Fig. 18A) that developed in the late stages of rifting. Since we have imaged the Moho beneath the upper-crustal block (Fig. 8), we assume that the detachment fault initiated as a rolling-hinge fault that rooted in the granitic middle crust of the Newfoundland margin (Fig. 18D).

If the detachment fault that unroofed the basement beneath the U reflection is rooted in the middle crust, as we suggest in Fig. 18A, we possibly have a scenario where seawater was not able to penetrate the underlying mantle. This is important, because our geophysical evidence clearly suggests that the mantle beneath the COT between 150 km and 230 km was not serpentinized: Van Avendonk et al. (2006) imaged seismic velocities higher than 8.0 km/s beneath the smooth basement of SCREECH Line 2. Our density model (Fig. 4) is consistent with this result, since the velocity–density relationship (Fig. 3) suggests that high-velocity mantle would have to be dense, which explains a slight rise (25 mGal) in the gravity profile between 130 km and 170 km (Fig. 4).

The apparent absence of prerift sediments in the deep margin of SCREECH Line 2 highlights a major difference in structure with the conjugate Iberian margin (Krawczyk et al., 1996). Similar observations in other rift settings have lead researchers to propose large-scale simple-shear mechanisms that would encompass the entire rift zone (Wernicke, 1985; Marton and Buffler, 1993). These models have received mixed support over the years, because straightforward mechanical considerations (Anderson, 1942) argue against low-angle extensional faults. However, normal faults that initiate at high angles may rotate to a shallower dip with ongoing extension (Spencer, 1984; Buck, 1988). The temporal development of our numerical models (Figs. 15 and 17) also predicts the development of low-angle detachment faults that initiated as high-angle normal faults.

8.4. Mantle exhumation

If the basement beneath the U reflection in SCREECH Line 2 is exhumed along a landward-dipping detachment fault that rooted in the middle crust (Fig. 18D), we expect to encounter rocks at the basement surface that were exhumed from progressively shallower crustal depths, as we look in a seaward direction, until we reach the breakaway point of the low-angle fault. This model may only hold for the basement beneath the U reflection, between 165 km and 230 km in our model (Fig. 18A), because seaward of 230 km the nature of the basement changes significantly. The increased basement relief and seismic velocities of ~7 km·s−1 in the basement seaward of 230 km appear much more consistent with exhumed continental mantle, such as the seafloor exhumed in the Iberia Abyssal Plain (e.g., Dean et al., 2000). Farther seaward at 270 km, results from Site 1277 (Robertson, 2007) confirm that the continental mantle of the Avalon terrane is exhumed here.

To explain these observations, we invoke a second, deeper detachment fault that cut off the mid-crustal detachment fault in the vicinity if Site 1276, at 230 km (Fig. 18A and E). The deeper detachment accommodated mantle exhumation by ductile shearing in the center of rift zone. During exhumation, the continental mantle was bent in the rolling-hinge detachment fault, such that extensional stresses allowed water to penetrate and form serpentinites deep beneath the basement surface (Sibuet et al., 2007). In our hot geodynamic extension model, mantle unroofing starts to occur on the right side of block H at 4.5 Myr (Fig. 15G and H). In our cold geodynamic model, it appears that mantle exhumation occurred much farther seaward (Fig. 17), just as we observed in the data of SCREECH Line 2 (Fig. 9).

8.5. Timing of separation of continental crust

The two geodynamic models that we presented in this paper show different predictions for the time that is required between thinning of the continental crust to mantle exhumation. The hot lithospheric model, with a constant extension half rate of 12 mm·yr−1, leads to separation of continental crust in 5 Myr (Fig. 15). On the other hand, the cold lithospheric model, with an extension half rate of 4.5 mm·yr−1, predicts that rifting takes about 23 Myr, with much of the localization of deformation in the last 5 Myr. These two models span the range of rift durations that have been suggested for the Newfoundland and Iberia margins (Wilson et al., 2001), if we consider only the final Late Jurassic to Early Cretaceous phase of extension between Newfoundland and Iberia. Our hot geodynamic model suggests that if the rift duration is indeed as short as 5 Myr, we should expect abundant mantle melting (>30%) in the final stages of rifting. Since the amount of rift-related igneous rocks at ODP Site 1277 appears small (Robertson, 2007), the cold extension model (Fig. 17) may be a better fit for the Newfoundland–Iberia rift.

The results of our geodynamic modeling (Figs. 11–17) shows a strain development in the rift model that predicts a distribution of rift basins and large blocks of continental crust in rifted margins that are reminiscent of images from geophysical data in the Newfoundland–Iberia rift and other rifted margins. In a similar geodynamic study, Huismans and Beaumont (2007) achieved similar success in reproducing first-order features of the margins of Newfoundland and Iberia. These studies show the importance of the feedback between deformation and rheology in the evolution of rifted margins. However, the extension rates in the geodynamic models of Huismans and Beaumont (2007) are much lower (~2.5 mm·yr−1 half rate), and their rift duration is twice as long (~46 Myr) as our cold rifting model.

Huismans and Beaumont (2007) justify the long rift duration in their model by assuming that the Late Triassic to Early Jurassic extension in the exterior margins of Newfoundland and Iberia (Tankard and Welsink, 1989) are part and parcel of the history of the Newfoundland–Iberia rift. This assumption seems valid, given that these geodynamic models correctly predict diffuse, widespread formation of sedimentary basins during early extension, and localization of strain afterwards. However, if we compare it to some local geophysical data, the 2.5 mm·yr−1 average extension rate of Huismans and Beaumont (2007) is much slower than the 10 mm·yr−1 extension half rate that marine geophysicists have estimated in the Iberia Abyssal Plain around the time of continental breakup (Minshull et al., 2001; Russell and Whitmarsh, 2003), and at a slow extension half rate of 2.5 mm·yr−1 it would be difficult to thin most of the continental crust in just 5 Myr, as was suggested by Wilson et al. (2001) for the Newfoundland and Iberia margins.

model, stress is a function of viscosity and strain rates, which entails that failure is reached much sooner at high strain rates than at low strain rates. As a consequence, higher extension rates will lead to more distributed deformation in a visco-plastic solid, while low extension rates are more likely to lead to strain localization and exhumation of the deeper lithosphere, of which we have ample evidence in the Newfoundland–Iberia rift. While it is clear that the extension rate is an important factor in geodynamical models of continental rifting, it may not be sufficient to run a series of models with a different constant extension rate. Future numerical experiments may have to address whether temporal variations in extension rate at rifted margins are caused by variability in far-field stresses, or whether they reflect an evolution in the rheology of the extending continental lithosphere.

9. Conclusions

1. We obtained new geophysical constraints on the structure of the continental slope along SCREECH Line 2 on the eastern Grand Banks, Newfoundland. We found fairly abrupt thinning of the crust from ∼28 km to ∼6 km beneath the continental slope. Moreover, the lower crust at the shelf edge has anomalously high velocity (7.0–7.2 km·s\(^{-1}\)) and density (3.0–3.2 g cm\(^{-3}\)). The basement surface of the rifted margin is overall smooth overall, and prerift sediments are found only at the base of the continental slope. Mantle seismic velocities are high beneath the COT (7.0–7.6 km·s\(^{-1}\)), but they are much lower (7.3–7.5 km·s\(^{-1}\)) further into the Newfoundland Basin near ODP Sites 1276 and 1277.

2. In order to better understand these results we present two geodynamic rifting models that we use to interpret some features in our data. The first model applies a 12 mm·yr\(^{-1}\) half rate to hot continental lithosphere, the second model applies a 4.5 mm·yr\(^{-1}\) extension half rate to cold lithosphere. A strain softening parameterization in the crust and upper mantle allows us to see the development of high-angle normal faults that either become inactive or flatten out to become low-angle detachment faults as extension continues. In both models, lithospheric shear zones eventually lead to continental breakup. However, the cold rifting model requires much more extension before lithosphere is breached, and it is accompanied by a much larger amount of thinning of the continental crust in the distal margins. The large amount of melt produced in the hot extension model (∼30%) may also be difficult to reconcile with the results of ODP Leg 210 in the Newfoundland Basin (Robertson, 2007).

3. A layer or a lens of strong, gabbroic lower crust beneath the eastern Grand Banks may have aided in the localization of the strain in the distal margin. With ongoing extension, the gabbroic lower crust rotated upwards towards the rift center until the Moho achieved a 50° dip. The presence of strong lower crust at relatively shallow levels beneath the continental slope may have caused the locus of extension to shift eastward towards the Iberian margin.

4. The apparent lack of prerift sediments in the distal margin of the eastern Grand Banks is consistent with the development of landward-dipping, low-angle detachment faults during the later stages of rifting. In our model for SCREECH Line 2 we invoke a crustal detachment that exhumed middle crust beneath the smooth base-"


