Seismic velocity structure of the rifted margin of the eastern Grand Banks of Newfoundland, Canada

Harm J. A. Van Avendonk,[1] W. Steven Holbrook,[2] Gregory T. Nunes,[3] Donna J. Shillington,[4] Brian E. Tucholke,[5] Keith E. Louden,[6] Hans Christian Larsen,[7] and John R. Hopper[8]

[1]   We present a compressional seismic velocity profile of the crust of the eastern margin of the Grand Banks of Newfoundland, Canada. This velocity model was obtained by a tomographic inversion of wide-angle data recorded on a linear array of 24 ocean bottom seismometers (OBSs). At the landward side, we imaged a crustal thickness of 27 km in Flemish Pass and beneath Beothuk Knoll, which is thinner than the 35-km-thick crust of the central Grand Banks. We therefore assume that the eastern rim of the Grand Banks stretched uniformly by 25%. Farther seaward, the continental crust tapers rapidly beneath the continental slope to ~6 km thickness. In the distal margin we find a 60-km-wide zone with seismic velocities between 5.0 and 6.5 km s^{-1} that thins to the southeast from 6 to 2 km, which we interpret as highly extended continental crust. Contrary to other seismic studies of the margins of the Grand Banks, we find seismic velocities of 8 km s^{-1} and higher beneath this thin crustal layer in the continent-ocean transition. We conclude that mantle was locally emplaced at shallow levels without significant hydration from seawater or serpentinized mantle was removed along a décollement in the final stages of continental rifting. The outer edge of highly extended continental crust borders a 25-km-wide zone where seismic velocities increase gradually from 6.3 km s^{-1} just below the top of acoustic basement to 7.7 km s^{-1} at 5 km below basement. We interpret this area as a relatively narrow zone of exhumed and serpentinized continental mantle. Seaward, we imaged a thin and laterally heterogeneous layer with a seismic velocity that increases sharply from 5.0 km s^{-1} in basement ridges to 7.0 km s^{-1} at its base, overlying mantle velocities between 7.8 and 8.2 km s^{-1}. We interpret this area as unroofed mantle and very thin oceanic crust that formed at an incipient, magma-starved, ultraslow spreading ridge. A comparison of the conjugate rifted margins of the eastern Grand Banks and the Iberia Abyssal Plain show that they exhibit a similar seaward progression from continental crust to mantle to oceanic crust. This indicates that before continental breakup, rifting exhumed progressively deeper sections of the continental lithosphere on both conjugate margins. A comparison between the continent-ocean transition of the Grand Banks and Flemish Cap shows that the final phase of continental rifting and the formation of the first oceanic crust required more time at the Grand Banks margin than at the southeastern margin of Flemish Cap.

[1]University of Texas Institute for Geophysics, Jackson School of Geosciences, Austin, Texas, USA.
[2]Department of Geology and Geophysics, University of Wyoming, Laramie, Wyoming, USA.
[3]ExxonMobil, Houston, Texas, USA.
[4]National Oceanography Centre, Southampton, UK.
[5]Department of Geology and Geophysics, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts, USA.
[6]Department of Oceanography, Dalhousie University, Halifax, Nova Scotia, Canada.
[7]Integrated Ocean Drilling Program, Sapporo, Japan.
[8]Department of Geology and Geophysics, Texas A&M University, College Station, Texas, USA.

1. Introduction

[2]   Studies of continental rifting have shown the response of the Earth's lithosphere to extensional forces in a large number of settings [*Ruppel*, 1995]. The apparent relationship between the state of the lithosphere during extension and the morphology of continental rift basins gives insight into the geological processes that control deformation at different depths [*Buck*, 1991]. Understanding the evolution of a continental rift toward breakup is particularly challenging because there can be important feedbacks between the rheology of the thinning lithosphere and the spatial distribution of deformation. For example, during breakup the
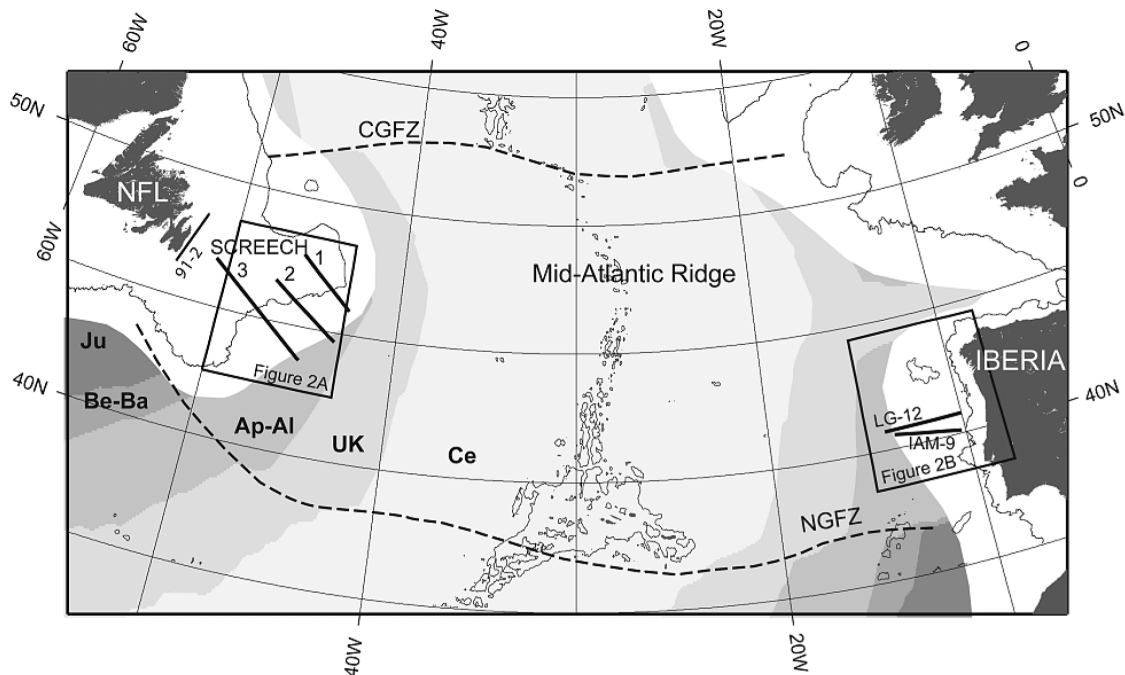
Figure 1. Regional map of the North Atlantic Ocean, including the margins of the Newfoundland-Iberia rift. The rift is bounded by Newfoundland (NFL) and Iberian margins and the Newfoundland-Gibraltar (NGFZ) and Charlie-Gibbs (CGFZ) fracture zones. The Mid-Atlantic Ridge, Flemish Cap, and Galicia Bank are marked with the 2000 m bathymetry contour. Gray shaded ages of the seafloor [*Müller et al.*, 1997] are Jurassic (Ju), Berriasian to Barremian (together comprising the Neocomian, Be-Ba), Aptian and Albian (Ap-Al), Upper Cretaceous (UK), and Cenozoic (Ce). Seismic transects discussed in this paper (thick black lines) are the three SCREECH transects (see text), LG-12 [*Krawczyk et al.*, 1996], and IAM-9 [*Dean et al.*, 2000]. FC, Flemish Cap; GB, Galicia Bank. Boxes locate the maps of Figure 2.

final separation of the continental lithosphere may be facilitated by local weakening of the relatively strong upper mantle by serpentinization [*Pérez-Gussinyé and Reston*, 2001], or it may be caused by the presence of melt generated by adiabatic decompression of the rising mantle [*Schärer et al.*, 1995; *Desmurs et al.*, 2002]. The interaction of these shallow and mantle processes during the late stages of continental breakup can be investigated at passive margins by probing the enigmatic zone of "transitional crust", the seafloor that lies between the thinned continental crust and the crust formed at a mid-ocean spreading ridge [*Louden and Chian*, 1999; *Whitmarsh et al.*, 2001a].

[3] Over the last few decades, many studies of continental rifting have focused on the passive margins of the Grand Banks of Newfoundland and Iberia (Figure 1). The Newfoundland-Iberia region experienced extension in the Triassic, but the final opening of the North Atlantic Ocean started in the Late Jurassic. The paleomagnetic and stratigraphic record shows that rifting of these margins generally progressed from south to north [*Masson and Miles*, 1984]. First, the Iberia Abyssal Plain and southern Newfoundland Basin formed by extensional deformation that localized in the distal margins of the new ocean basin [*Tucholke et al.*, 1989; *Whitmarsh and Sawyer*, 1996]. As rifting moved northward to the current Flemish Cap and Galicia Bank, extension was initially accommodated in the Galicia Interior Basin on the Iberian margin [*Pérez-Gussinyé et al.*, 2003] and in the Orphan Basin and Flemish Pass offshore eastern

Canada [*Sibuet et al.*, 2006] (Figure 2). Rifting in the Flemish Pass and Galicia Interior Basin did not lead to continental breakup there, but the continental lithosphere eventually ruptured seaward of these basins between the Galicia Bank and Flemish Cap. As a result, Galicia Bank and Flemish Cap form two promontories of continental crust on opposite sides of the Atlantic Ocean.

[4] To understand the south-to-north change in rift kinematics in the central North Atlantic Ocean we must have good spatial constraints on the deep crustal structure. In the 1980s and 1990s, excellent data were gathered on the Iberian margin during marine geophysical surveys and by the Ocean Drilling Program (ODP). These studies have shown that the basement of the West Iberian continent-ocean transition zone (TZ) mostly consists of exhumed mantle, but this area is significantly wider in the southern Iberia Abyssal Plain than on the Galician margin [*Henning et al.*, 2004]. Peridotites that were recovered from basement highs on the Iberia margin appear to have a subcontinental mantle origin [*Charpentier et al.*, 1998; *Hébert et al.*, 2001]. Studies using ocean bottom seismograph (OBS) refraction data found seismic velocities in the TZ that increase rapidly in the first few kilometers below basement to $\sim$7.6 km s^{-1} [*Chian et al.*, 1999; *Dean et al.*, 2000]. This velocity is too high to represent igneous underplating at such shallow depths. It is more likely that penetrating seawater caused partial serpentinization of exhumed mantle peridotites [*Christensen*, 2004]. The absence of Moho reflections in
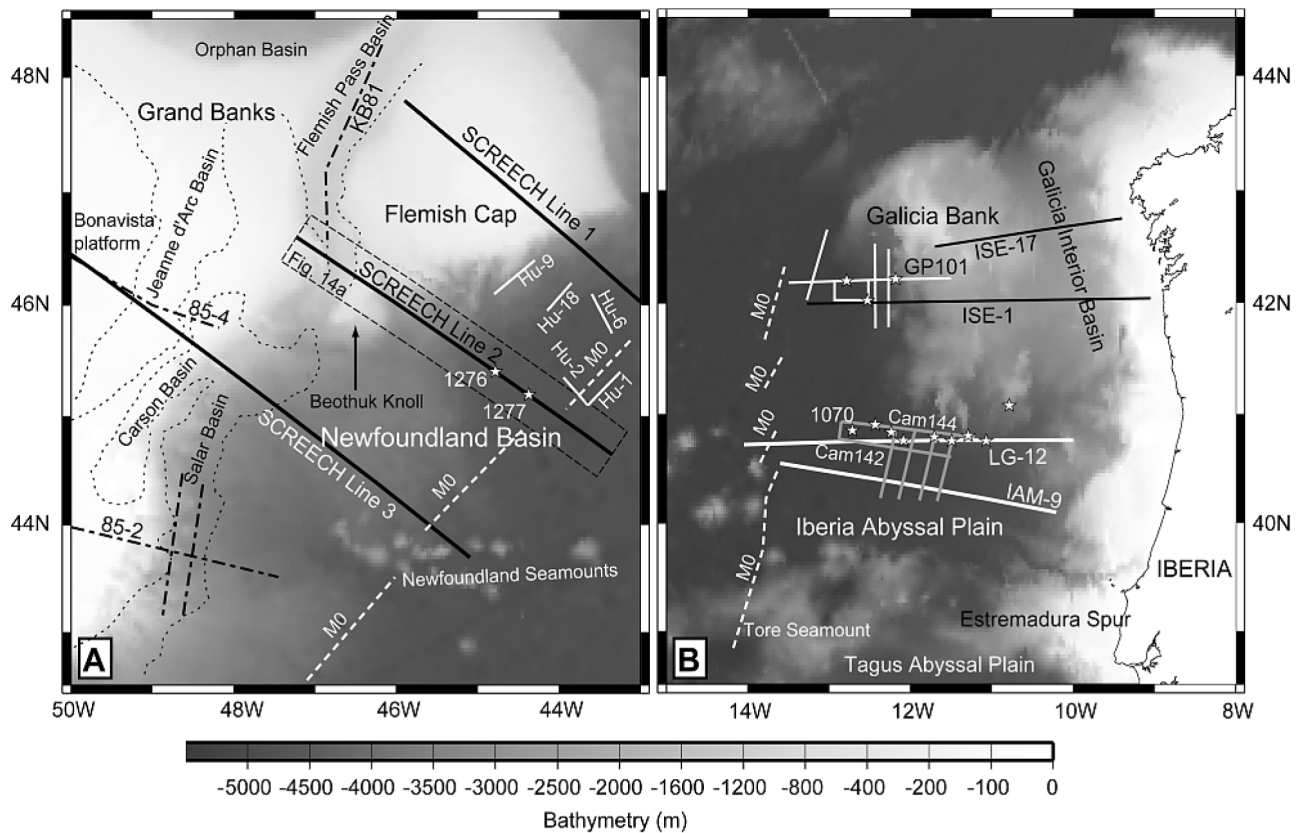
Figure 2. Two location maps for the Newfoundland and conjugate Iberia margins. Magnetic anomaly M0 picks [*Srivastava et al.*, 2000] are indicated with a white dashed line, and ODP drill sites are marked with white stars. (a) Location map of the Newfoundland margin. The SCREECH transects (thick black lines) are shown together with the locations of past seismic refraction studies. Hu profiles (white lines) between SCREECH lines 1 and 2 are from *Todd and Reid* [1989]. KB81 is the seismic refraction of *Keen and Barrett* [1981]. Other seismic refraction lines are 85-2 [*Reid*, 1994] and 85-4 [*Keen and de Voogd*, 1988]. Sedimentary basins are outlined with a black, short dashed line. (b) Location map of the conjugate Iberian margin. Other seismic reflection and refraction lines shown are IAM-9 [*Dean et al.*, 2000], LG-12 [*Krawczyk et al.*, 1996], Cam142 and Cam144 [*Chian et al.*, 1999], GP101 [*Whitmarsh et al.*, 1996], ISE-1 [*Zelt et al.*, 2003], and ISE-17 [*Pérez-Gussinyé et al.*, 2003].

the TZ offshore Iberia [*Pickup et al.*, 1996] further suggests that the serpentinized mantle is not covered by thinned continental crust or oceanic crust. On the landward side of the Iberian margin, the zone of continental crust that is thinned to less than 10 km (hereinafter referred to as thin continental crust) is just 40 km wide in the southern Iberia Abyssal Plain [*Dean et al.*, 2000]. In contrast, the combined width of thin continental crust in the Galicia Interior Basin [*Pérez-Gussinyé et al.*, 2003] and on the deep western Galicia margin [*Whitmarsh et al.*, 1996] is 110 km.

[5] Owing to hydrocarbon exploration, the stratigraphic history of deep sedimentary basins on the shelf of the Grand Banks is well known from the Triassic to the Cenozoic [*Tankard and Welsink*, 1987; *Enachescu*, 1992; *Driscoll et al.*, 1995], but the deep crustal structure of the distal margin in the Newfoundland Basin has been much debated. *Tucholke and Ludwig* [1982] postulated that magmatism that caused the J anomaly, a high-amplitude M0−M1 magnetic lineation extending north and south from the Newfoundland Fracture Zone a the southern edge of the Newfoundland-Iberia rift [*Rabinowitz et al.*, 1978], corre-

lated with the onset of seafloor spreading in the Newfoundland Basin to the north. This would imply that the ∼150-km-wide zone of smooth basement that lies between the continental slope of the southeastern Grand Banks and the northward extension of the J anomaly is highly extended continental lithosphere [*Sullivan*, 1983; *Tucholke et al.*, 1989]. Alternatively, *Keen and de Voogd* [1988] interpreted seismic reflection profiles across the edge of the Grand Banks and Flemish Cap as consistent with just ∼50 km of thinned continental crust abutting oceanic crust of normal thickness near the continental slope. In a seismic refraction study at the southeastern margin of the Grand Banks, *Reid* [1994] found that the crust seaward of the continental rise is ∼4 km thick over a distance of about 100 km, with a velocity of 4.3 km s^{-1}, overlying a layer with velocities between 7.2 and 7.7 km s^{-1}. He interpreted this result as thin oceanic crust overlying serpentinized mantle that was created at a magma-starved mid-ocean ridge. *Todd and Reid* [1989] obtained similar results from seismic refraction data obtained south of Flemish Cap. Unfortunately, these short seismic refraction lines were not reversed, and
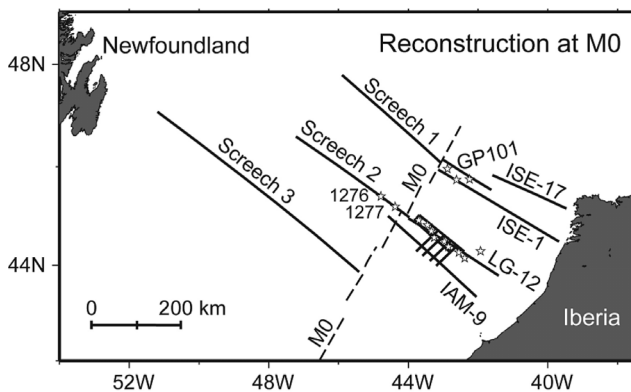
Figure 3. Reconstruction of the Newfoundland-Iberia rift at M0 time [*Srivastava et al.*, 2000]. The geographical coordinates are fixed with respect to North America.

they did not resolve well the nature of the TZ in the Newfoundland Basin.

[6] In the summer of 2000, we carried out a seismic data acquisition program termed "Studies of Continental Rifting and Extension on the Eastern Canadian shelf" (SCREECH). During this cruise, we collected three long profiles with both seismic reflection and refraction data across the margins of Grand Banks and Flemish Cap (Figure 2). All three profiles run in the presumed direction of extension, from the continental platform onto the abyssal plain well beyond magnetic anomaly M0 [*Srivastava et al.*, 2000]. In the north, SCREECH line 1 traverses Flemish Cap and its southeastern margin into the Newfoundland Basin. Line 1 yielded results that were rather different from the conjugate Galicia margin. The crust of Flemish Cap thins abruptly in the seaward direction, from a maximum thickness of 30 km to just 1.5 km over a distance of 90 km, with 40 km of continental crust thinner than 10 km [*Funck et al.*, 2003]; the seismic reflection data from this profile give no evidence for a large crust-penetrating fault [*Hopper et al.*, 2004] such as the S reflection beneath Galicia Bank [*Reston et al.*, 1996]; continental rocks may be juxtaposed against very thin oceanic crust that was formed by ultraslow spreading [*Hopper et al.*, 2004]. SCREECH line 3 lies across the southeastern margin of the Grand Banks in the vicinity of a few older seismic lines, such as 85-4 (Figure 2) [*Keen and de Voogd*, 1988; *Tucholke et al.*, 1989]. SCREECH line 3 therefore helps to assess contrasting views on the structure of the southern Newfoundland Basin. The seismic velocity structure along SCREECH line 3 shows that thin continental crust underlies the Carson and Salar Basins over a width of 120 km [*Lau et al.*, 2006b]. Farther seaward an 80-km-wide zone is interpreted to be exhumed continental mantle [*Lau et al.*, 2006b]. Contrasts between the results from SCREECH line 1 and 3 show that there are large differences in architecture between the southern and northern half of the Newfoundland-Iberia rift.

[7] Imaging the crust along SCREECH line 2 helps to understand the structure of an important section of the Newfoundland-Iberia rift. In a reconstruction of the rift at M0 time [*Srivastava et al.*, 2000], SCREECH line 2 and LG-12 [*Krawczyk et al.*, 1996] form a continuous transect between the eastern Grand Banks and the Iberia Abyssal

Plain just south of Galicia Bank (Figure 3). The basement along LG-12 was well sampled during ODP Legs 149 and 173. In 2003, the Newfoundland margin was also drilled at two sites (1276 and 1277) along SCREECH line 2 during ODP Leg 210. The seismic velocity model of *Chian et al.* [1999] covers the area of the Iberia ODP sites, and seismic reflection and refraction profile IAM9 [*Dean et al.*, 2000] provides a wider cross section of the Iberian margin 30 km south of LG-12.

[8] In this paper we present a tomographic analysis of seismic refraction data along SCREECH line 2. *Shillington et al.* [2006] conducted a study of coincident multichannel seismic (MCS) data on this line. This combined Newfoundland-Iberia transect represents the best sampled profile of a rifted-margin pair in the North Atlantic Ocean. Comparison of SCREECH line 2 seismic refraction data with the two other SCREECH profiles also provides more insight into the along-strike variability and timing of rifting processes in the Newfoundland-Iberia rift. Our results show that the TZ at the eastern Grand Banks near Flemish Pass is not as wide as along SCREECH line 3. However, the similarities between SCREECH lines 2 and 3 suggest that the mechanism of breakup did not change much along strike south of Flemish Cap. On the other hand, final rifting of the Flemish Cap and Galicia Bank led to relatively rapid onset of seafloor spreading.

2. Geological Setting

[9] To infer the amount of extension of the Newfoundland margin along SCREECH line 2, we must establish the crustal structure of the Grand Banks before Mesozoic rifting. Drilling and seismic reflection surveys show that basement of the northern Grand Banks largely consists of Proterozoic bedrock of the Avalon terrane [*O'Brien et al.*, 1983; *King et al.*, 1986; *Durling et al.*, 1987]. As this continental block docked during the Appalachian orogeny, it acquired a 4- to 8-km-thick sequence of Ordovician to Devonian sediments on much of the Grand Banks [*Durling et al.*, 1987; *O'Brien et al.*, 1983]. These relatively undisturbed strata have seismic velocities of just 3.5 to 4.7 km s^{-1} [*Sheridan and Drake*, 1968; *Reid and Keen*, 1988], which is significantly lower than the seismic velocity of the Proterozoic basement (>5.5 km s^{-1}). Outcrops of Precambrian granodiorites on the eastern Grand Banks and Flemish Cap [*King et al.*, 1985] suggest that Paleozoic sediments were eroded or never deposited there. A marine seismic refraction study near the Avalon peninsula (line 91-2 in Figure 1) determined a velocity of 5.7 km s^{-1} just below basement to 7.0 km s^{-1} near the Moho at 40 km depth [*Marillier et al.*, 1994]. The thickness of the Proterozoic crust of the Grand Banks therefore may have been about 35 km. Similar thicknesses of 30 to 35 km depth have been estimated elsewhere on the continental platform of the Grand Banks [*Keen and de Voogd*, 1988; *Lau et al.*, 2006b]. The seismic velocity structure of the Grand Banks is similar to that of the Avalon peninsula of Newfoundland [*Hughes et al.*, 1994; *Hall et al.*, 1998]. The general absence of seismic velocities >7.0 km s^{-1} in the Avalon terrane suggests that the lower crust of the Appalachian orogen of northeastern Newfoundland is more intermediate than mafic in composition [*Hughes et al.*, 1994].

[10] North-south trending magnetic anomalies on the Grand Banks show continuity of basement structures with those of the Iberian peninsula, which reveals their common Paleozoic history [*Haworth and Lefort*, 1979]. It is believed that Mesozoic rifting in the Atlantic was facilitated by these Paleozoic structures [*Tankard et al.*, 1989; *de Voogd et al.*, 1990]. The same tectonic fabric is visible in a series of extensional ridges and basins along the eastern rim of the Grand Banks, termed the Outer Ridge Complex [*Grant et al.*, 1988]. The most prominent of these sedimentary basins are the Carson, Jeanne d'Arc, and Flemish Pass basins [*Enachescu*, 1987; *Grant et al.*, 1988; *Welsink et al.*, 1989] (Figure 2a). Growth of these basins during extension may often have been controlled by east dipping listric faults [*Tankard and Welsink*, 1987; *Enachescu*, 1993]. In some areas, the crust beneath the Jeanne d'Arc and Flemish Pass basins may be less than half the thickness of unstretched crust of the Grand Banks [*Keen and Barrett*, 1981; *Reid and Keen*, 1990]. The deeper sections of the Carson and Jeanne d'Arc Basin contain evaporites that were deposited during the Triassic phase of extension when North America was rifting from Africa [*Enachescu*, 1988; *Austin et al.*, 1989]. These fault-bounded basins were also active during the final phase of rifting in the Late Jurassic through Early Cretaceous [*Jansa and Wade*, 1975; *Tankard and Welsink*, 1987].

[11] Perhaps the most prominent tectonic event in the Jurassic-Cretaceous rifting episode is represented by the widespread Aptian age Avalon unconformity, which is thought to represent uplift of the Grand Banks in response to continental breakup [*Grant et al.*, 1988]. *Meador et al.* [1988] correlated this unconformity across the continental slope into the Newfoundland Basin, where it appears to coincide with the U reflection, a high-amplitude seismic horizon that blankets and commonly masks smooth basement from the continental slope out to crust at magnetic anomaly M3 age (early Barremian, 125–126 Ma) [*Tucholke and Ludwig*, 1982; *Tucholke et al.*, 1989]. In this paper we use the timescale of *Gradstein et al.* [1994] to reference geological ages and magnetic anomalies, although there are some significant discrepancies with a newer timescale [*Gradstein et al.*, 2004]. If the U reflection is truly a breakup unconformity, a wide zone of the Newfoundland Basin seafloor must represent thinned and subsided continental crust. A similar and apparently correlative unconformity may also mark the cessation of rifting on the conjugate Iberian margin [*Groupe Galice*, 1979; *Meador and Austin*, 1988; *Alves et al.*, 2006]. In contrast, seafloor spreading as early as anomaly M5 (late Hauterivian, ∼128 Ma) has been proposed [*Whitmarsh et al.*, 2001b; *Russell and Whitmarsh*, 2003], although drilling results at ODP sites 897, 899, and 1070 indicate that apparently subcontinental mantle lithosphere in the TZ was in extension at least 14 Ma beyond that time [*Comas et al.*, 1996; *Whitmarsh and Wallace*, 2001]. The nature of the crust in the TZ of both margins therefore must be established independently.

[12] It is noteworthy that SCREECH line 2 appears to lie near the boundary between the northern (Flemish Cap–Galicia) and southern (Grand Banks–southern Iberia) segments of the Newfoundland-Iberia rift (Figure 2a). Interpreted magnetic anomalies of *Srivastava et al.* [2000] suggest that M15 (137 Ma) is the oldest magnetic anomaly along SCREECH line 2 at the base of the continental slope

(Figure 4). However, magnetic anomalies older than M3 are very weak and difficult to correlate in this area [*Shillington et al.*, 2004]. The basement thus could be extended continental crust, although denuded mantle with few igneous intrusions could also explain the weak magnetic anomalies (J.-C. Sibuet et al., Exhumed mantle forming transitional crust in the Newfoundland-Iberia rift and associated magnetic anomalies, submitted to *Journal of Geophysical Research*, 2006, hereinafter referred to as Sibuet et al., submitted manuscript, 2006). To the north of SCREECH line 2, Flemish Cap appears to have been attached to Galicia Bank until near chron M0 [*Sibuet et al.*, 2006]. Thus the older magnetic lineations along SCREECH line 2 must terminate northward against a transfer zone. This boundary, indicated in Figure 4, may be an extension of the Dominion transfer zone [*Tankard et al.*, 1989], a large fault zone that truncates ridges and basins on eastern Grand Banks.

3. SCREECH Seismic Data

[13] The SCREECH seismic study is a collaboration of scientists from the United States, Canada, and Denmark. The marine seismic experiment, which took place in summer 2000, used the R/V *Maurice Ewing* for seismic shooting and recording and the R/V *Oceanus* for the deployment and recovery of ocean bottom seismometers and hydrophones [*Funck et al.*, 2003; *Shillington et al.*, 2006]. We shot each of the three main lines across the Newfoundland margin twice, once to acquire seismic reflection data and once to acquire refraction data. We recorded the reflection data on the 6-km-long 480-channel seismic streamer of the R/V *Maurice Ewing*. During the MCS profiling, the shot spacing was 50 m, but the spacing was increased to 200 m for the refraction measurements in order to limit noise from previous shots at long offsets [*Nakamura et al.*, 1987]. Both data sets were shot with the 20-gun, 140-L array of the R/V *Maurice Ewing*.

[14] SCREECH line 2 crosses the south end of Flemish Pass and Beothuk Knoll between the eastern Grand Banks and the Newfoundland Basin. On this 368-km-long line, we deployed thirteen four-component ocean bottom seismographs (OBS) from Dalhousie University and the Geological Survey of Canada and 14 ocean bottom hydrophones (OBH) from Woods Hole Oceanographic Institution. We lost two instruments at sea and one instrument did not record useful data. We determined the location of the other 24 instruments on the seafloor using the direct wave from the air guns to the instruments. Table 1 and Figure 4 show the location of the instruments on this refraction profile.

[15] We designed the spacing of instruments on line 2 to capture the seismic velocity structure of the crust and upper mantle. Seismic refractions from the Mesozoic and Cenozoic sediments can be seen on many of the seismic refraction records, but only at short source-receiver distances. The instrument density (Table 1) was therefore not sufficient to constrain either the detailed basement topography or lateral variations in the seismic velocity of these sediments [*Nunes*, 2002]. The 6-km-long streamer of the R/V *Maurice Ewing*, however, provides good control on the seismic structure of these sediments. We therefore adopted the results from prestack depth migration of MCS data on line 2 (Figure 5) [*Shillington et al.*, 2006] in our velocity
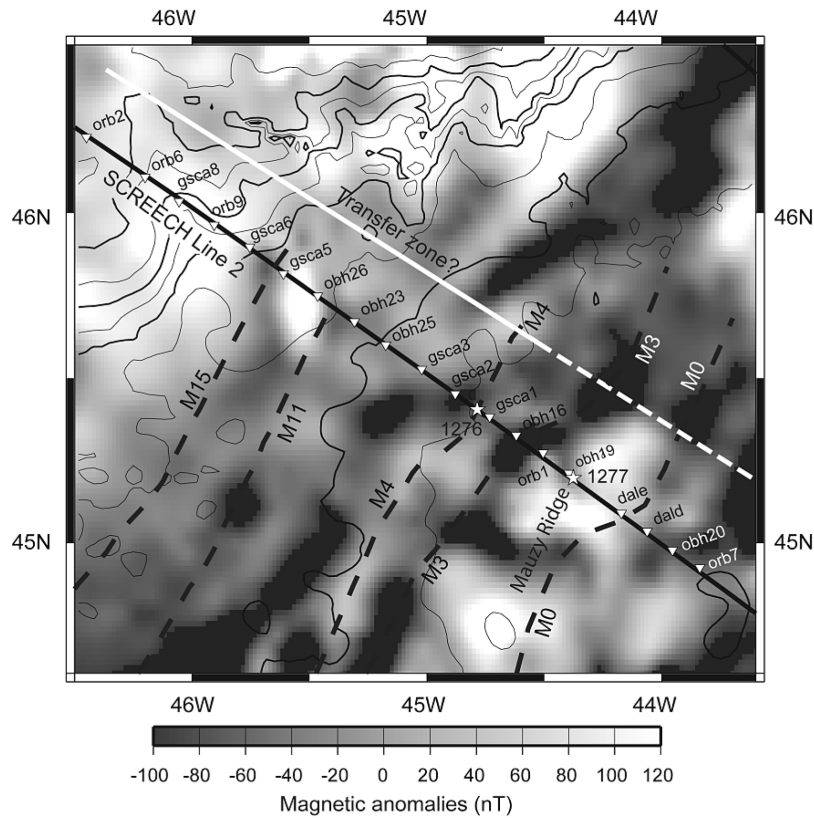
Figure 4. Gridded compilation of magnetic data in the Newfoundland Basin [*Verhoef et al.*, 1996] with magnetic picks [*Srivastava et al.*, 2000]. Bathymetry is contoured every 500 m (thin and thicker black lines). Magnetic picks older than M0 terminate against a transfer zone (white line) postulated by *Sibuet et al.* [2006] to account for motion between North America and Flemish Cap. Ocean bottom instrument locations are marked with triangles, and ODP drill sites 1276 and 1277 are marked by stars.

Table 1. List of 24 Instruments That Recorded Data on SCREECH Line 2[a]

OBS	Latitude	Longitude	X, km	Z, m	N	Δt, ms	σ, ms	χ^2
ORB8	46° 34.50′N	47° 11.50′W	3.02	502	45	32	50	0.63
ORB3	46° 27.46′N	46° 56.46′W	26.02	720	118	47	69	0.77
GSC4	46° 20.60′N	46° 41.65′W	48.88	1152	341	58	71	0.91
ORB2	46° 13.49′N	46° 26.98′W	71.91	455	394	46	72	0.91
ORB6	46° 6.31′N	46° 12.04′W	94.95	1580	490	48	56	1.28
GSCA8	46° 1.94′N	46° 03.35′W	108.87	2148	436	76	71	1.48
ORB9	45° 57.73′N	45° 54.35′W	122.97	2371	448	61	69	1.20
GSCA6	45° 53.76′N	45° 45.26′W	136.71	3178	561	55	64	0.98
GSCA5	45° 48.91′N	45° 36.74′W	150.87	3432	611	61	94	0.77
OBH26	45° 44.98′N	45° 27.90′W	164.76	3538	343	78	85	1.07
OBH23	45° 40.27′N	45° 18.47′W	178.44	3774	414	70	90	0.87
OBH25	45° 36.08′N	45° 10.51′W	192.63	4002	165	83	59	1.61
GSCA3	45° 31.61′N	45° 1.38′W	206.79	4242	164	63	76	0.89
GSCA2	45° 27.14′N	44° 52.72′W	220.80	4423	235	48	83	0.67
GSCA1	45° 22.97′N	44° 43.96′W	234.60	4538	335	82	97	1.03
OBH16	45° 19.62′N	44° 37.22′W	245.39	4539	239	58	73	0.96
ORB1	45° 16.46′N	44° 30.28′W	256.43	4562	221	47	83	0.62
OBH19	45° 12.70′N	44° 23.70′W	267.22	4590	333	47	88	0.72
DAL-E	45° 5.46′N	44° 10.36′W	289.11	4685	327	51	72	0.87
DAL-D	45° 2.15′N	44° 3.77′W	299.72	4710	324	76	92	0.96
OBH20	44° 58.63′N	43° 57.22′W	310.66	4732	376	61	96	0.76
ORB7	44° 55.46′N	43° 50.29′W	321.34	4742	136	53	56	1.06
OBH27	44° 51.26′N	43° 43.73′W	332.58	4960	180	102	95	1.78
DAL-C	44° 48.26′N	43° 37.52′W	342.76	5011	127	106	79	1.93

[a]The distance X is measured along a great circle through the first and last seismic refraction shot fired on this profile. The number of first arriving traveltime picks N is relatively low for instruments in shallow water, but other record sections provide good data coverage. The average data misfit for each station following the tomographic inversion is given by δt, and the average of the errors assigned to each traveltime pick is given by σ. The inversion procedure achieved an overall weighted data misfit χ^2 of 1.0, which means that on average δt compares well with σ. The χ^2 for individual record sections shows that each of these records has an acceptable data fit.
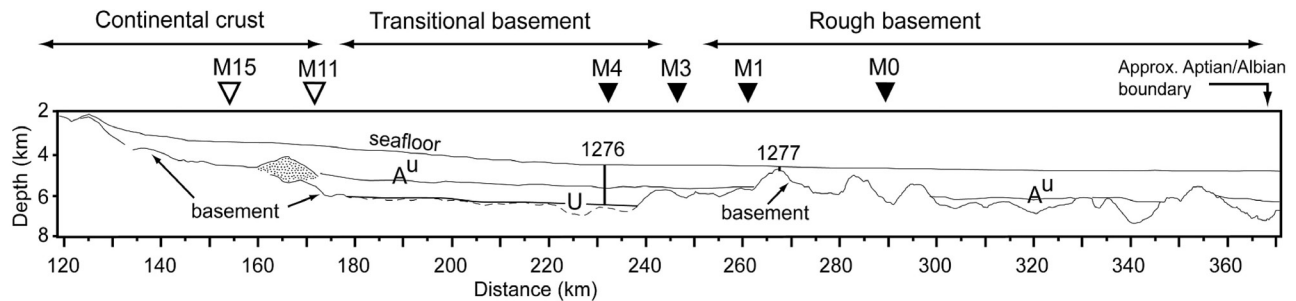
Figure 5. Profile of SCREECH line 2 adapted from the prestack depth migration of *Shillington et al.* [2006]. Magnetic anomalies M15 to M0 are as interpreted by *Srivastava et al.* [2000]. A package of apparently prerift sediments (stippled area) on top of the crustal block at 160–170 km suggests that the picks M15 and M11 (open triangles) are wrongly interpreted. Deep-lying basement at 175 to 240 km is covered by the U reflection. This strong seismic horizon, which locally coincides with postrift diabase sills drilled at ODP Site 1276 [*Shipboard Scientific Party*, 2004a], masks basement relief below it (dashed line) [*Shillington et al.*, 2006], and it onlaps rough basement at anomaly M3. The two main sediment packages of postrift sediments are separated by the A^U reflection [*Shipboard Scientific Party*, 2004a; *Stant and Wise*, 2004].

model. The sedimentary section can be divided broadly in two layers that are divided by the A^U reflection [*Shillington et al.*, 2006] (Figure 5), which is thought to mark the initiation of strong abyssal circulation in the Atlantic Ocean in late Eocene to early Oligocene time [*Tucholke and Mountain*, 1979; *Davies et al.*, 2001]. Seismic velocities typically vary from 1.8 km s^{-1} in the youngest sediments to 2.8 km s^{-1} just above basement.

[16] The basement topography changes character along the profile (Figure 5). Landward, faulted blocks of what appears to be continental crust step down beneath the continental slope. The most seaward block (160 to 175 km) appears to be covered by rotated, prerift sediments [*Shillington et al.*, 2006]. Because magnetic anomaly picks of chron M15 of *Srivastava et al.* [2000] lie landward of this continental block, their interpretation as seafloor spreading anomalies is not warranted. Seaward of 175 km, between magnetic anomaly picks M11 and M4, lies 65 km of deep and relatively smooth basement of the TZ, and seaward of M4 the basement topography is rough.

[17] Because the shallow structure of line 2 is well constrained by the reflection data, we analyzed the seismic refraction data primarily to image the crust and upper mantle. We enhanced the quality of the refraction data using a few simple procedures. First, we band-pass filtered the data mostly between 4 and 14 Hz. Autocorrelation of the seismic traces shows reverberation in the waveforms at 50 to 100 ms. We applied predictive deconvolution to remove the reverberations and sharpen the seismic pulse. These steps helped to identify seismic refractions at source-receiver offsets larger than 100 km on many instruments. The data recorded over the continental shelf (Figure 6) were nevertheless relatively noisy, because they contained seafloor multiples.

[18] Compared to SCREECH line 1 and line 3, SCREECH line 2 does not capture a very long section in the shallow water setting of the Newfoundland margin. Nevertheless, instruments located on the eastern Grand Banks (Figure 6) recorded seismic refractions over more than 100 km of the continental shelf. First arriving crustal

refractions can easily be traced in most receiver gathers, and wide-angle seismic reflections of high amplitude and other late seismic arrivals can be seen intermittently over distances of 10 to 20 km in the landward portion of line 2. For example, instrument ORB9 recorded a wide-angle reflection from the base of the continental slope at 165 km in our transect (Figure 6). On the continental slope, instrument GSCA5 shows similar wide-angle reflections at 110 km and 180 km (Figure 7). Farther seaward, rapid thinning of the continental crust is evident in the data. At the base of the continental slope, wide-angle *PmP* reflections appear at very short source-receiver offsets (Figure 8). *Funck et al.* [2003] and *Lau et al.* [2006b] used both wide-angle refractions and reflections to constrain a relatively simple seismic velocity model with three crustal layers for the shelf of Flemish Cap and Grand Banks. These crustal boundaries are important because they may represent rheological contrasts in the continental crust [*Hughes et al.*, 1994]. Wide-angle reflections were also observed in our SCREECH line 2 beneath Beothuk Knoll and Flemish Pass, but they are not consistent with a uniform layered velocity structure.

[19] Instruments located in the Newfoundland Basin show seismic refractions with a higher apparent velocity, even at short source-receiver offsets [*Nunes*, 2002]. The record section of OBH16, located at the seaward end of the TZ (Figure 4), shows that almost all seismic refractions emerge with an apparent velocity larger than 7 km s^{-1}, even at short source-receiver offsets (Figure 9). If any kind of crust exists in this section of the model it must be very thin. Farther to the southeast, instrument OBH19 (Figure 10) overlies a basement ridge that was drilled at ODP site 1277 and is termed Mauzy Ridge [*Shipboard Scientific Party*, 2004b]. In contrast to the near-offset arrivals recorded on OBH16, the OBH19 seismic refractions have a relatively low apparent velocity. Wide-angle reflections are very weak in the southeastern half of line 2 (Figures 9 to 11), which may be explained by a gradual increase of seismic velocity with depth from the crust into mantle. There appears to be a wide-angle reflection 10 km seaward of OBH19, which could be a shallow Moho reflection. Farther seaward, the
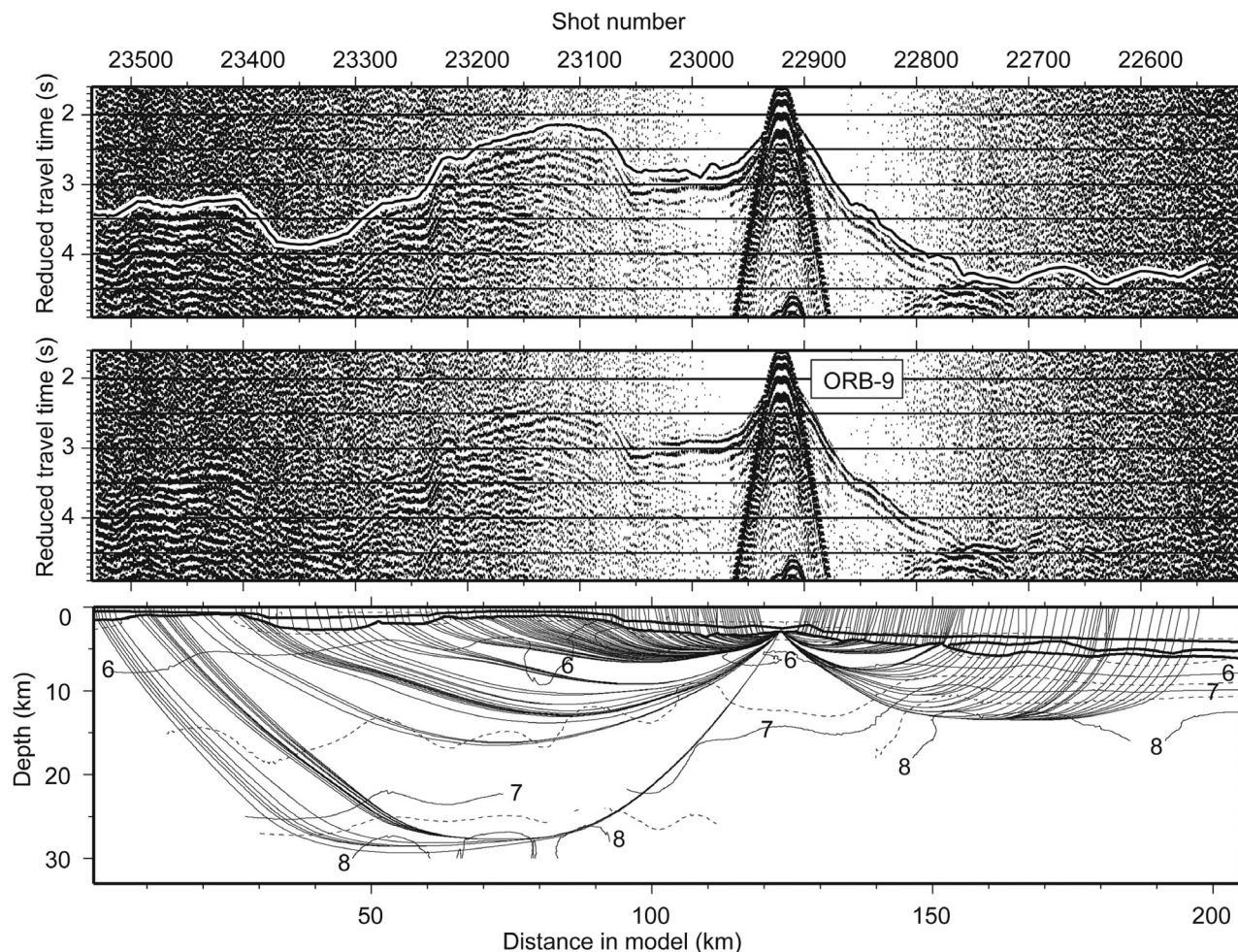
Figure 6. Seismic refraction data and ray diagram for instrument ORB9, situated on the seaward edge of what can be confidently characterized as continental crust. The OBS recorded air gun shots over a distance of 200 km. (top and middle) Refraction data are shown with a reduction velocity of 7.0 km s^{-1}. The traveltime picks are highlighted by a black line. (bottom) Ray paths show that the seismic refractions from shots 23480 to 23530 appear to turn in the mantle. Seismic velocities are contoured every 0.5 km s^{-1}.

apparent velocity of seismic refractions increases more gradually with source-receiver offset from about 6 to 8 km s^{-1}, which is suggestive of a thick Moho transition zone (Figure 11).

[20] We picked first arriving phases for all 24 record sections, which amounted to a total of 7430 traveltimes. We assigned errors to these traveltimes based on the quality of the data, varying from 30 to 150 ms (Figure 7). The average assigned error σ is given for each instrument in Table 1. These first arriving traveltimes can be used in an analysis of the crustal and upper mantle structure without making assumptions regarding the nature of the Moho or other crustal boundaries. If the model space is sampled by sufficient first arriving waves it may therefore be preferable to use just these arrivals to constrain the seismic velocity structure [*Zelt et al.*, 2003].

4. Tomography

[21] We used the 7430 traveltime picks in an iterative tomographic inversion to find the seismic velocity model with the least amount of complexity that fits the data within a tolerance level [*Van Avendonk et al.*, 1998, 2004]. With these data we created a seismic velocity model for the crust and uppermost mantle beneath the basement surface. The seismic structure of the overlying sediments, determined by prestack depth migration of MCS data [*Shillington et al.*, 2006], was not modified in our tomographic inversion. We chose a starting model with seismic velocities varying from 5.0 km s^{-1} near the top of the basement to 8.1 km s^{-1} at 33 km depth for an iterative tomographic inversion. In the iterations we traced rays between the sources and receivers in the current seismic velocity model. Delays between arrival time picks and computed traveltimes were then used in a linearized least squares inversion to obtain a new seismic velocity model. The traveltime misfits were scaled with assigned data uncertainties to give appropriate weight to every traveltime pick in the inversion. The linearized inversions were designed to produce the smoothest and flattest seismic velocity model with a scaled data misfit χ^2 of 1.0. For the inversion we discretized the model with a grid spacing of 1.7 km in the horizontal direction and a vertical
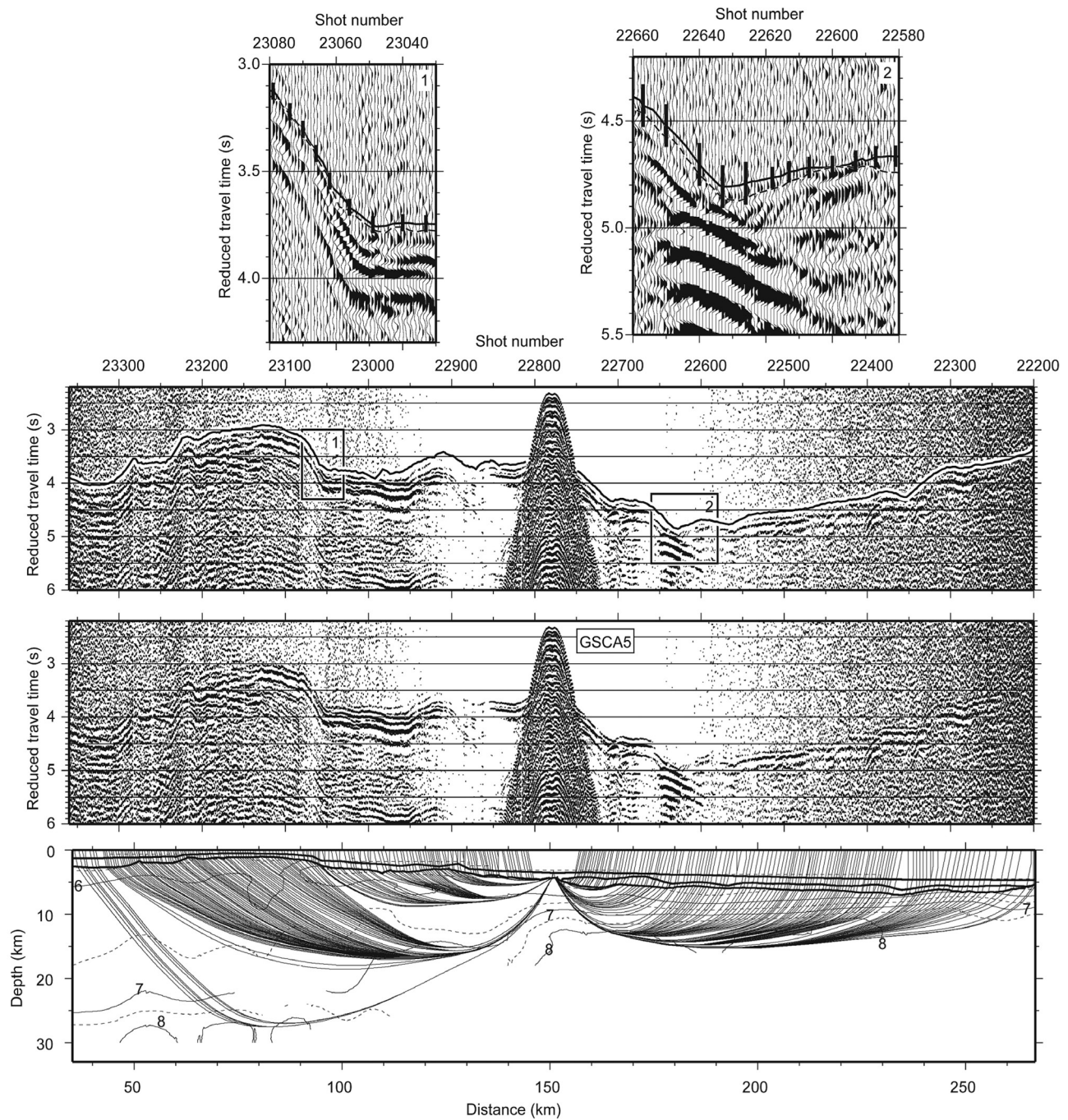
Figure 7. Seismic refraction data and ray-tracing diagram for instrument GSCA5. Seismic refraction data are shown as in Figure 6. The two insets show assigned errors (with bars) and data fit (thin dashed line) of our seismic velocity model. The ray paths show that mantle refractions arrive at much shorter source-receiver offsets in the Newfoundland Basin than on Beothuk Knoll.

spacing that increased linearly from 0.6 km in the top of the model to 1.1 km at the bottom. Of all 9495 grid nodes, 5278 points are situated deeper than the basement surface and lie close enough to one of the ray paths to be included in the vector of model parameters. In a regularized inversion we estimated updates for these parameters using the 7430 traveltime data and additional constraints that minimized the first and second derivatives of the resulting model [*Van Avendonk et al.*, 2004]. We used scaling lengths for the

spatial derivatives that were 8 times longer in the horizontal direction than in the vertical direction. This modeling choice represents our assumption that the Earth's seismic structure is predominantly stratified.

[22] After twelve iterations we obtained a root-mean-square traveltime misfit of 64 ms. Comparison of the distribution of traveltime delays and scaled data misfits shows that the latter has a more Gaussian shape with shorter tails (Figure 12). This suggests that, on average, large data
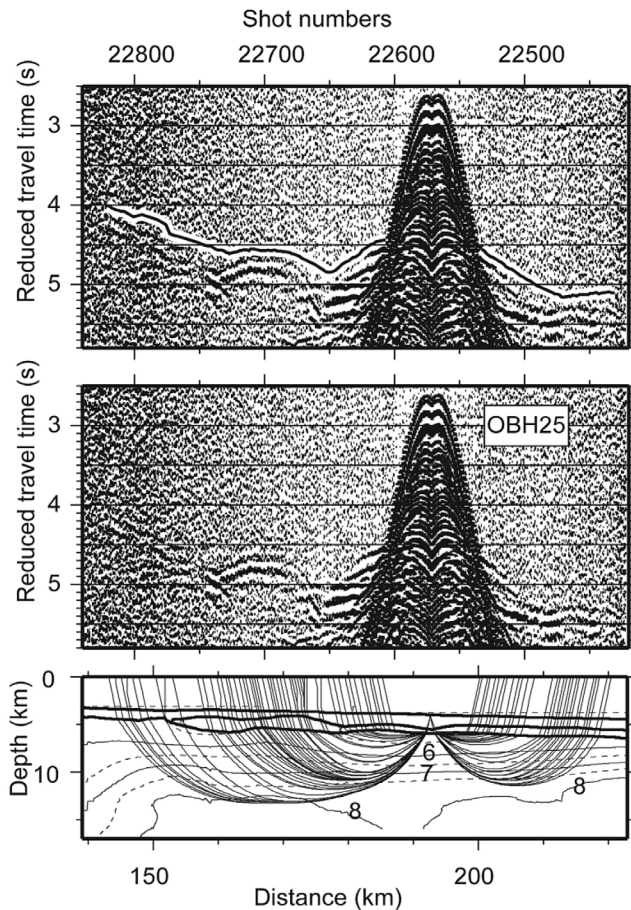
Figure 8. Seismic refraction data and ray paths for instrument OBH25 displayed as in Figure 6. This short seismic record of OBH25 offers a contrast to the instruments on the continental shelf. The seismic refractions move out with a high velocity at short source-receiver distances. The ray diagram shows that the data are consistent with a shallow mantle.

misfits are caused by noisy data to which we assigned large uncertainties. The result gives us confidence that poor traveltime data did not dominate the least squares inversion. Although the distribution of delay times is wider, most of the delays are smaller than 100 ms. Traveltime curves in space and time are drawn for all instruments in Figure 13. As the histograms showed before, the calculated curves lie near picked traveltimes. Moreover, the discrepancies between the seismic refraction data and model predictions are systematic. Traveltime residuals of ~50 ms vary slowly with offset over a few tens of kilometers. The static time shifts can be explained by small errors in the shallow seismic velocity structure near the instruments. For example, local variations in the basement depth or seismic velocity of overlying sediments could lead to systematic data misfits because they affect the arrival times of most or all refractions recorded by the instrument. The apparent velocities in the seismic refraction data are well predicted by the tomographic inversion. We therefore believe that the seismic velocities in the deep crust were modeled accurately.

[23] The inversion of first arriving traveltimes produced the smoothest and flattest seismic velocity model with a χ^2 misfit of 1.0 (Figure 14b). If imaged seismic velocities higher than 8.0 km s^{-1} represent the upper mantle, we find a Moho depth of ~28 km beneath the continental shelf, and very thin crust (<7 km) beneath the abyssal plain of the Newfoundland Basin. The ray paths of first arriving refractions (Figure 14c) show that the continental shelf is not sampled as densely as the deep margin. Some long-offset seismic refractions dive into the mantle beneath the continental shelf, but the thick lower crust is not well covered. The ray paths otherwise indicate very dense sampling of the crust and upper mantle along SCREECH line 2.

[24] The seismic velocity model shown in Figure 14b represents the best fit to our data and smoothness constraints, but it is not a unique solution for the traveltime data. We constructed a resolution matrix for our inverse problem using the method described by *Van Avendonk et al.* [2004]. This matrix shows how seismic velocity structures were imaged in our inversion. We applied the resolution matrix to elliptical seismic velocity anomalies of three different sizes that can occur anywhere in the model (Figure 15). We correlated the input seismic velocity model with its image in order to measure how much of the seismic velocity anomaly is reproduced. We first used elliptical bodies of 10 km by 4 km (Figure 15a) to show that small seismic velocity anomalies are well resolved (>50%) in the upper mantle beneath the abyssal plain. This result is not a surprise, because we recorded seismic refractions that turn in the upper mantle between source-receiver offsets of 10−80 km (Figures 9 to 11). In contrast, we only find ray paths that turn in the thin crust of the Newfoundland Basin at source-receiver offsets smaller than 15 km. Thus, with our instrument spacing of 10−15 km (Table 1), the fine-scale velocity structure of the thin crust is not as well constrained, and the resolution test of Figure 15a shows that seismic velocity anomalies of 10 km by 4 km are only ~20% recovered. Some portions of the thick crust beneath the continental shelf have enough crossing ray paths to resolve these small seismic velocity anomalies. Resolution tests for larger structures of 20 km by 8 km (Figure 15b) and 40 km by 16 km (Figure 15c) show that these features are reliably recovered in most of our velocity model.

[25] Although the tomographic inversion that we used is an effective method to explore models that fit large traveltime data sets, the RAYINVR program of *Zelt and Smith* [1992] has most often been used to image seismic velocity structure of rifted margins [*Chian et al.*, 1999; *Dean et al.*, 2000; *Funck et al.*, 2003; *Lau et al.*, 2006b]. This method carries out a damped traveltime inversion to determine seismic velocities and layer boundaries in a coarsely parameterized model. The variable spacing of the grid nodes allows the user to test geologically relevant models. As such, the motivation for the underparameterized inversion differs from the philosophy behind a tomographic inversion, which uses a fine and evenly spaced grid for its model space [*Zelt et al.*, 2003]. The smoothness constraints in the overparameterized tomographic inversion are designed to select the simplest model that fits the traveltime data. Seismic velocity anomalies imaged in a tomographic inversion are therefore less influenced by the parameterization of the model space. Although layer interfaces have sometimes
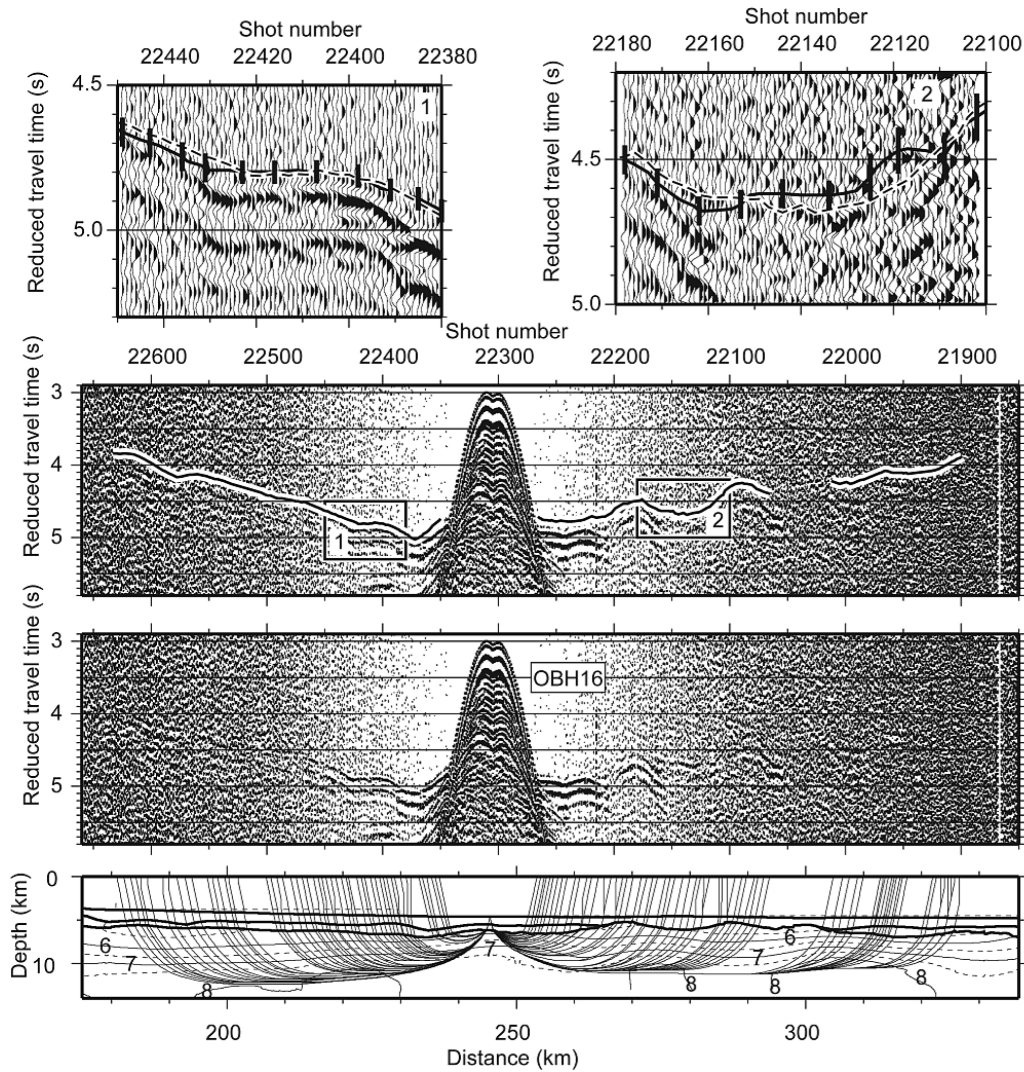
Figure 9. Refraction data and ray paths for instrument OBH16 displayed as in Figure 6. Instrument OBH16 is located at the seaward end of the flat-lying basement. Seismic refractions recorded by this OBS arrive with high apparent velocity, particularly those from the seaward side of SCREECH line 2.

been parameterized as first-order seismic velocity discontinuities introduced in tomographic inversions in order to incorporate wide-angle seismic reflections and later refractions [*McCaughey and Singh*, 1997; *Van Avendonk et al.*, 2004], we did not assume that these discontinuities are present in our model.

5. Results

5.1. Seismic Velocity Structure

[26] Our seismic velocity model (Figure 14b) shows a dramatic change in structure across the continental slope, but there also appear to be variations in the nature of the crust far into the Newfoundland Basin. We here describe characteristics of the model that are significant according to the resolution tests (Figure 15).

[27] Beneath the continental shelf the crust has a nearly constant crustal thickness of 27 km and is apparently covered with a thin layer of sediments. The seismic velocity increases from 5.5 to 6.5 km s^{-1} in the upper two thirds of the crust. The lower third of the crust has velocities higher than 6.5 km s^{-1}, but there is no evidence for a significant amount of crust with seismic velocities higher than 7.0 km s^{-1}. Beneath the continental slope, from 100 to 140 km in our model (Figure 14b), a crustal layer of ~12 km thickness with velocities up to 6.5 km s^{-1} appears to be underlain by velocities between 7.0 and 7.5 km s^{-1} over a depth range of 10 km. The resolution tests (Figure 15) suggest that these seismic velocities, which are high for continental crust but low for mantle rocks, may be accurately modeled.

[28] At the base of the continental slope (150 km in our model), the crust thins rapidly to just 4 km. Further to the southeast, between 160 and 170 km, an isolated crustal block has a thickness of 7 km. Prestack depth migrations of MCS data along SCREECH line 2 show a very similar shape and depth of the Moho in this part of the model [*Shillington et al.*, 2006]. Because this piece of crust is
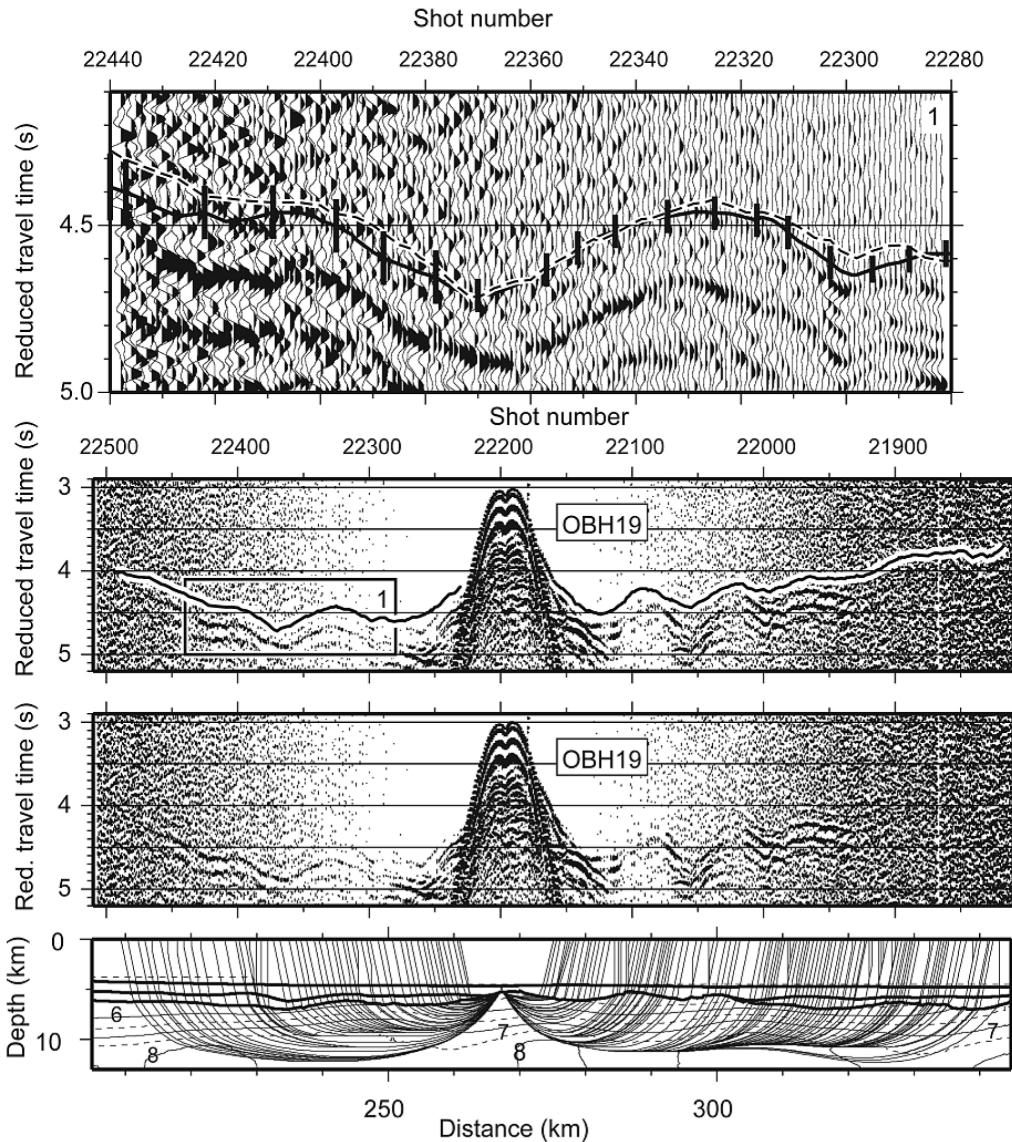
Figure 10. Refraction data and ray paths for instrument OBH19 displayed as in Figure 6. Instrument OBH19 was located on Mauzy Ridge. The bathymetric effect on traveltime curves is large, but the apparent velocity appears low at small source-receiver offsets.

buried by rotated layers of sediments with relatively high velocity (3.5 km s^{-1}), *Shillington et al.* [2006] considered it to be the seaward edge of upper crust of the Avalon terrane (Figure 5). Deeper in the Newfoundland Basin, between distances 180 km and 235 km, the top of the basement is relatively smooth but it is mostly obscured by the high-amplitude U reflection [*Tucholke et al.*, 1989; *Shillington et al.*, 2006]. We imaged seismic velocities lower than 6.5 km s^{-1} several kilometers into the basement of the TZ. At larger depth, seismic velocities between $7.0-7.6$ km s^{-1} are nearly absent here, in contrast to other deep margin areas sampled of the Newfoundland-Iberia rift. If we take the 7.5 km s^{-1} velocity contour as the Moho, we find that the crust thins steadily seaward from 6.0 to 2.0 km beneath the apparently flat basement (Figure 14b).

[29] Between ODP sites 1276 and 1277 we imaged seismic velocities greater than 6.0 km s^{-1} in the upper-most crust, increasing gradually to 7.7 km s^{-1} at depths up to 6 km below basement. Presumably, deeper rocks in this region reach mantle seismic velocities of 8.0 km s^{-1}, but they were not sampled by our data. This seismic structure is reminiscent of the wide TZ on the conjugate Iberia Abyssal Plain, although it covers only a 25-km-wide zone on our profile. It coincides with the onset of rough basement topography near chron M3 at the seaward edge of the low-lying basement covered by the U reflection [*Shillington et al.*, 2006]. It also contrasts with the outer part of the deep-water section of SCREECH line 2, where mantle seismic velocities higher than 8.0 km s^{-1} are observed (Figure 14b). We performed a ray-tracing test for instrument Dal-E (Figure 16) to investigate whether our data are consistent with mantle velocities higher than shown in our model between 240 km and 260 km. We tested four seismic velocity profiles with seismic velocity gra-
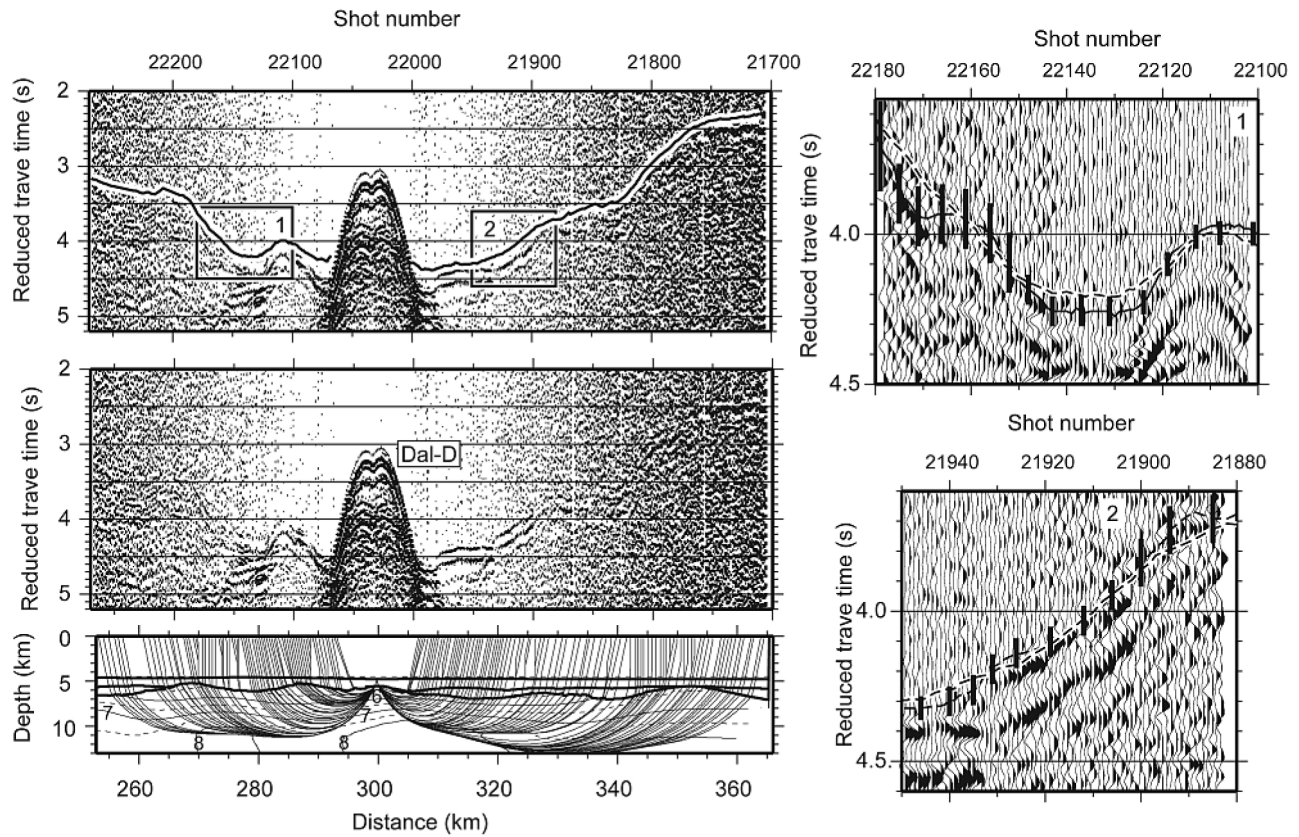
Figure 11. Refraction data and ray paths for instrument Dal-D displayed as in Figure 6. Instrument Dal-D was located far into the Newfoundland Basin. The refraction data recorded by this instrument do not show wide-angle reflections that would be expected if a sharp Moho contrast were present.

dients ranging from 0.2 to 0.6 s^{-1} from the basement into the mantle, where the seismic velocity reaches 8.0 km s^{-1} (Figure 16a). The tests show that gradients of 0.4 and 0.6 s^{-1} would give us seismic refractions that arrive about 200 ms earlier than what we see in our data (Figure 16b). A seismic velocity gradient of 0.2 s^{-1} leads to a traveltime advance of less than 50 ms, which is more difficult to refute with our data. The ray trace test nevertheless shows that the mantle
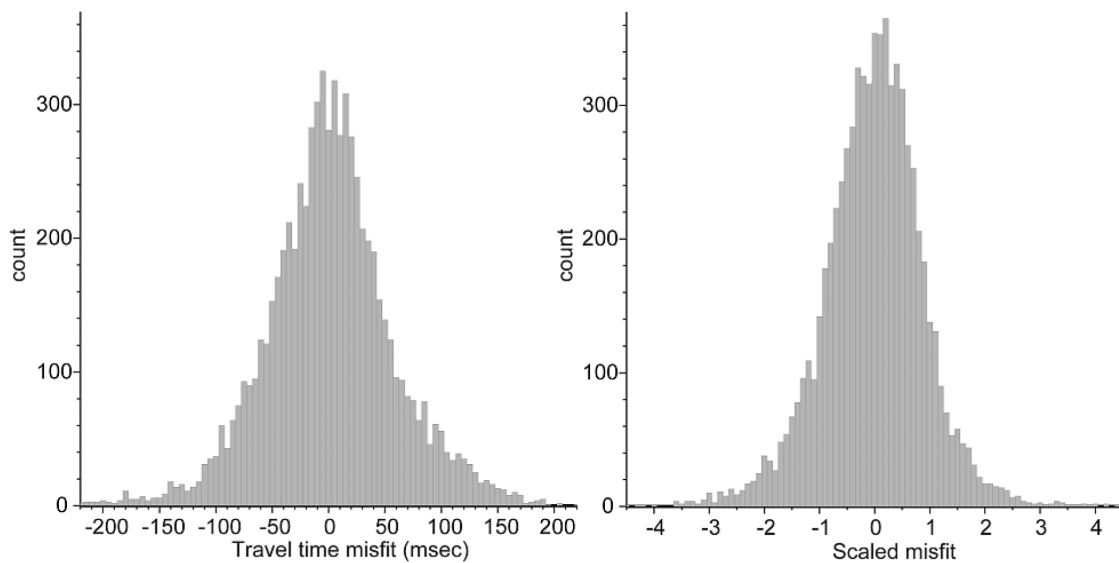

Figure 12. Two histograms show the distribution of (left) unscaled and (right) scaled distributions of the traveltime delays after tomographic inversion. The bin sizes are 5 ms and 0.1, respectively. The tomographic inversion is designed to minimize both the scaled misfit and the roughness of the seismic velocity.
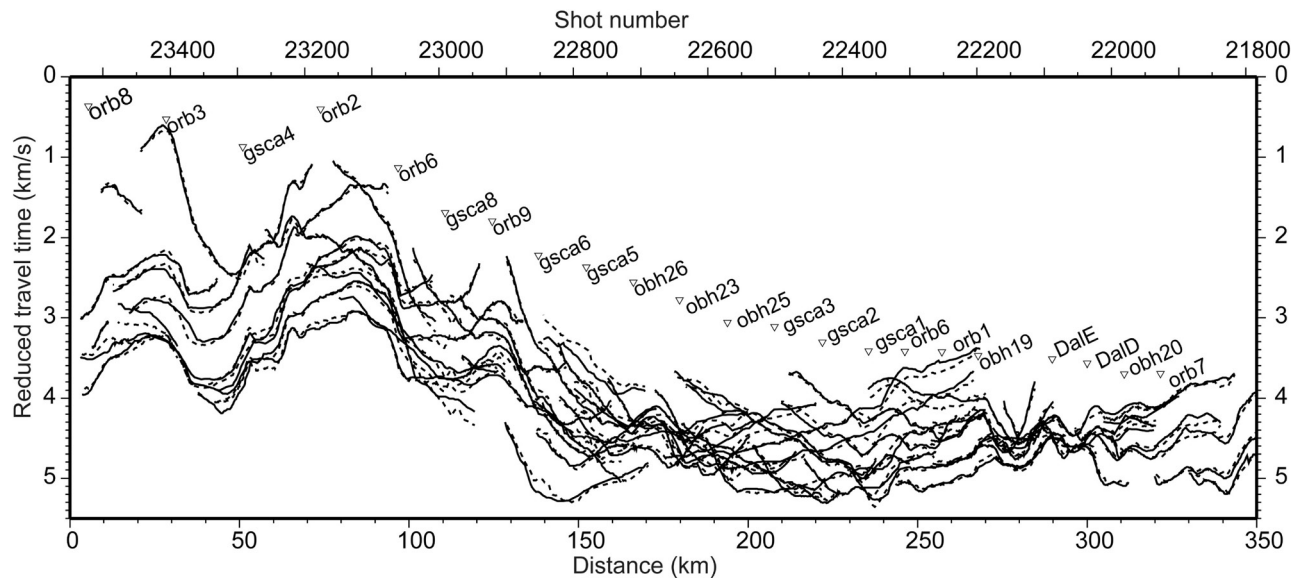
Figure 13. Traveltime curves of all seismic refraction data plotted in one frame. Traveltime picks are shown as solid curves; predicted traveltimes are dashed. The horizontal axis is the distance along SCREECH line 2. On the vertical axis, the source-receiver traveltimes are reduced by 7 km s^{-1}. Depths of instruments (triangles) are converted to traveltime using the speed of sound in water (1.5 km s^{-1}).

seismic velocities are significantly lowered over a depth range of at least 5 km.

[30] The Newfoundland Basin east of magnetic anomaly M1 is characterized by high basement relief with an average amplitude of about 1 km [*Shillington et al.*, 2006]. Our results indicate that the basement highs have seismic velocities of ~5.0 km s^{-1} near the top, increasing sharply with depth (Figure 14b). The 7.5 km s^{-1} velocity contour appears mostly to anticorrelate with basement relief, so the crust varies in thickness from 3 to 6 km with a typical wavelength of 30 km. There is no large-scale trend in crustal structure superimposed on this irregularity.

5.2. Moho

[31] In numerous seismic studies of continental and oceanic crust the Moho has been variably imaged as a sharp seismic velocity discontinuity, as a band of reflective layers, or as a transition zone between the crust and upper mantle [*Braile and Chiang*, 1986; *Collier et al.*, 1994; *Cook*, 2002; *Nedimovic et al.*, 2005]. Where mantle peridotite is exhumed to shallow levels, a serpentinization front may be evidenced by weak wide-angle reflections in seismic refraction data [*Muller et al.*, 2000]. In general, the nature of the seismic Moho likely varies laterally across rifted margins as it is formed by different tectonic and magmatic processes. Given that wide-angle reflections are weak or absent along SCREECH line 2 (Figures 9 to 11), the crust-mantle boundary appears to be expressed as a velocity gradient between 7 and 8 km s^{-1} over one to several kilometers depth, if it exists at all. Our minimum structure seismic velocity model (Figure 14b) may underestimate strong seismic velocity gradients, but the very tight traveltime fit (64 ms root-mean-square error) gives us

confidence that large-scale features of our model are well constrained.

6. Discussion

[32] In the discussion below, we compare seismic velocity profiles from our SCREECH line 2 analysis with velocity models for other parts of the Newfoundland-Iberia rift, even though these models were obtained using different methods. We find that the largest problem in making these comparisons is not difference in methodology, but differences in data quality. At the time that they were collected and analyzed, refraction data contributed by *Todd and Reid* [1989] and *Srivastava et al.* [2000] filled a large gap in constraints on the crustal structure of the Newfoundland Basin. Unfortunately, the range in velocities among these profiles is probably more indicative of uncertainty in modeling results than of spatial variability in crustal structure. We therefore limit our discussion of the seismic structure of basement in the Newfoundland-Iberia rift to results from recent and densely sampled seismic refraction profiles.

6.1. Continental Crust

[33] Results from a seismic refraction study of line 91-2 offshore from the Avalon Peninsula (Figure 1), indicate that the 30 to 35 km thick Proterozoic crustal bedrock is overlain by 4 to 9 km of Paleozoic sediments [*Marillier et al.*, 1994]. Our image of SCREECH line 2 shows that seismic velocities are mostly higher than 5.5 km s^{-1} on the eastern Grand Banks. We therefore assume that the Paleozoic sediments are absent along our profile, which is consistent with nearby presence of Precambrian rock dredges [*King et al.*, 1985]. The 27-km-thick crust that we imaged at the landward limit of SCREECH line 2 (Figures 14 and 17) is thinner than the
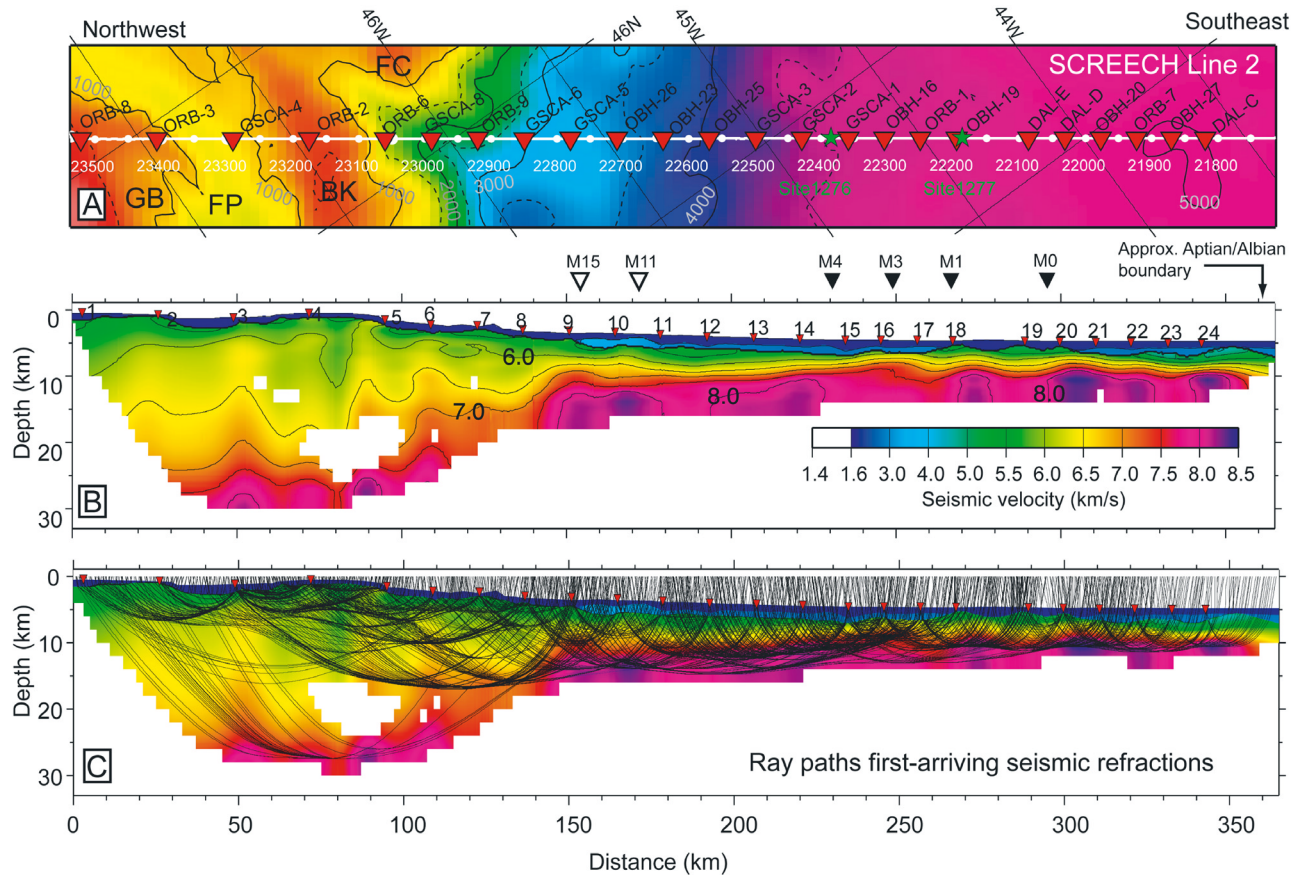
Figure 14. Results from tomographic imaging of first arriving traveltimes along SCREECH line 2. (a) Contoured color map of the bathymetry (m) in the vicinity of SCREECH line 2. The seismic refraction line crosses the eastern edge of the Grand Banks (GB), Flemish Pass (FP), Beothuk Knoll (BK), and south of Flemish Cap (FC). Bathymetry map is contoured every 500 m. Seismic shot numbers are labeled in white. Instrument locations are indicated with inverted red triangles. (b) Seismic velocity model along SCREECH line 2, contoured every 0.5 km s^{-1}. Magnetic anomaly picks are from *Srivastava et al.* [2000]. (c) Ray paths for first arriving seismic waves in the tomography model. Only every fifth of the ray path in our inversion is shown.

crystalline crust offshore Avalon peninsula, so continental crust beneath the eastern Grand Banks and Beothuk Knoll probably thinned by 5 to 10 km during the opening of the Atlantic Ocean. Farther north in Flemish Pass, *Keen and Barrett* [1981] determined that crustal thickness is just 22 km. A larger amount of crustal stretching and thinning in northern Flemish Pass is consistent with the clockwise pivotal motion of Flemish Cap that is thought to have started in the Late Jurassic [*Sibuet et al.*, 2006].

[34] The seismic structure of the Avalon terrane was well constrained on SCREECH line 1 and line 3. These two transects each covered 180 km of the continental shelf of Flemish Cap and the southeastern Bonavista Platform, respectively (Figure 2). Crustal thicknesses of 30 km observed on line 1 [*Funck et al.*, 2003] and 36-km-thick crust on line 3 [*Lau et al.*, 2006b] indicate that extension may have been negligible in these two platforms. In addition, on both transects the seismic velocity models consist of three continuous and relatively uniform crustal layers [*Funck et al.*, 2003; *Lau et al.*, 2006b]. Seaward dipping normal faults that bound the Jeanne d'Arc, Carson, and Flemish Pass basins probably cut deep into the crust of

the eastern Grand Banks [*Enachescu*, 1993; *Tankard et al.*, 1989]. We therefore expect the crustal structure at the northwest end of SCREECH line 2 to be more heterogeneous than other areas of the Grand Banks. Unfortunately, the northwestern section of our SCREECH line 2 velocity model lacks the resolution to see detailed variations in structure (Figure 15).

[35] SCREECH line 2 crosses from the eastern tip of the Grand Banks across Beothuk Knoll and into the Newfoundland Basin (Figures 2 and 14a). In this region, our tomographic model shows that the crust thins from ∼25 km at 100 km to ∼7 km at 150 km. Seismic velocities between 6.5 and 7.0 km s^{-1} make up a 10-km-thick layer beneath the continental shelf, but they are nearly absent beneath the continental slope. Between 110 km and 140 km, we imaged a 14-km-thick crustal layer beneath the continental slope with velocities of 5.5 to 6.0 km s^{-1}, overlying a large body with seismic velocities of 7.0 and 7.5 km s^{-1} (Figure 14b). The resolution tests (Figure 15) suggest that this structure may be significant, so we can associate velocities of 7.0 to 7.5 km s^{-1} beneath the continental slope with rapid thinning of the crust. Velocities of 7.0 to 7.5 km s^{-1} are too high for
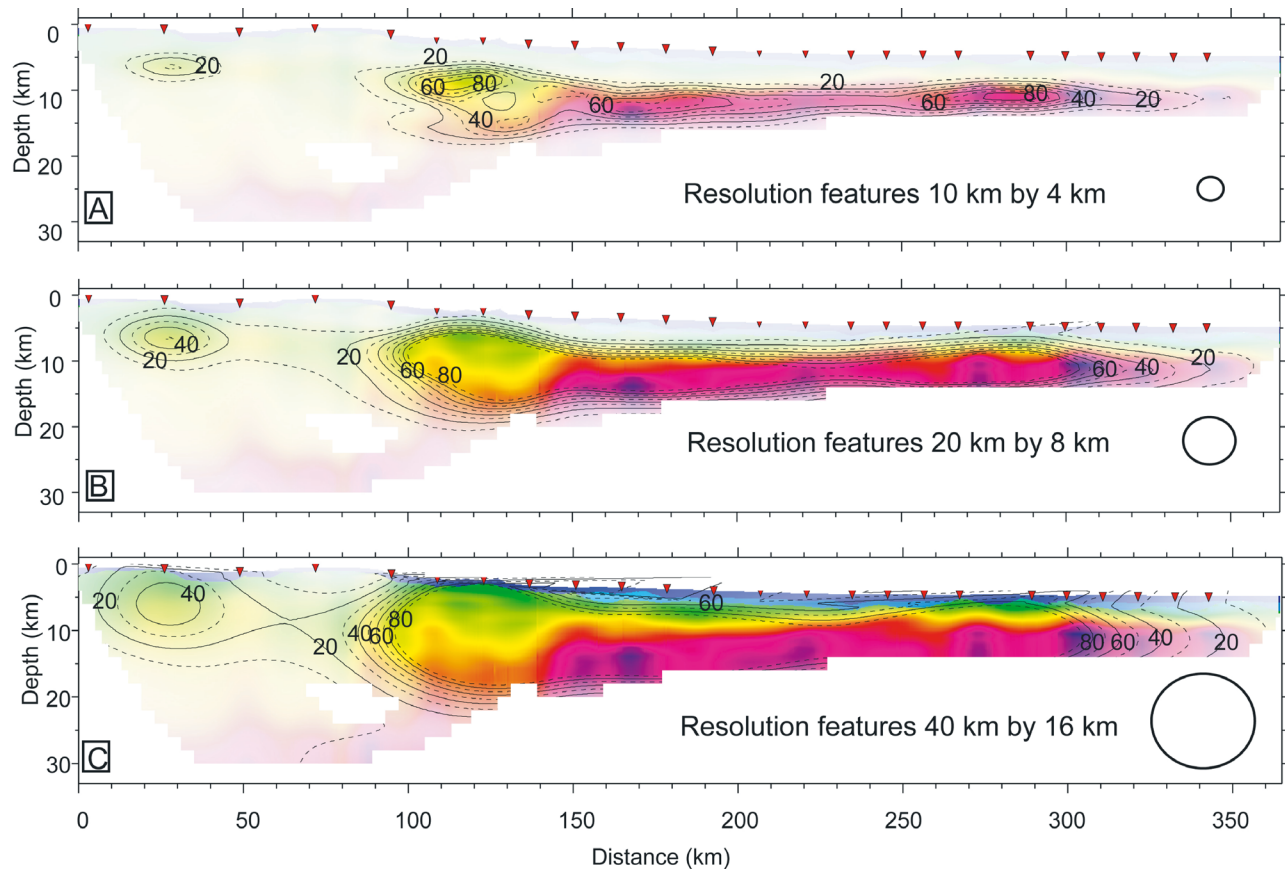
Figure 15. Resolution of the tomographic inversion. (a) Resolution of elliptical structures of 10 km by 4 km, contoured at every 10% recovery. The seismic velocity model is faded in areas where the resolution is low. (b) Resolution of 20 km by 8 km structure. (c) Resolution of 40 km by 16 km structure.

normal continental crust, but they are similar to those of serpentinized peridotite [*Horen et al.*, 1996]. However, serpentinization requires that water penetrated in faults in the brittle crust and uppermost mantle to hydrate the mantle. Numerical modeling suggests that these conditions do not arise until the crust has thinned to about 10 km [*Pérez-Gussinyé and Reston*, 2001], a depth somewhat less than the upper limit of our layer with velocities exceeding 7.0 km s^{-1}. If this layer beneath the continental slope with velocities between 7.0 and 7.5 km s^{-1} is an effect of serpentinization during rifting, we would also expect to find a similar but wider region farther seaward where the crust is thin. These factors suggest that this region beneath the continental slope, where seismic velocities lie between values that are typical for the crust and mantle, may not be serpentinized peridotite.

[36] Alternatively, the intermediate seismic velocities may indicate magmatic underplating [*Holbrook et al.*, 1992]. Numerical models of lithospheric necking and mantle decompression often predict melt extraction even at low strain rates [*Bown and White*, 1995], and the locus of melting does not need to coincide with the position of the crustal breakaway point [*Harry and Sawyer*, 1992]. The presence of a large magma body beneath this rift flank is nevertheless hard to reconcile with the small amounts of igneous intrusions interpreted elsewhere along highly stretched margins of the Newfoundland-Iberia rift [e.g.,

Whitmarsh et al., 2001b]. Another possibility is that mafic melt may have intruded the lower continental crust well before rifting started. Recent numerical models show that strong, mafic bodies in the lower crust can then restrict ductile flow during the early phase of rifting, such that faults penetrate the entire crust and localize the deformation [*Lavier and Manatschal*, 2006]. *Manatschal* [2004] suggested that gabbros of the Malenco area [*Müntener and Hermann*, 2001] played such a role in the formation of Tethyan extensional margins that are now exposed in the European Alps. The scenario might be invoked for SCREECH line 2, because the 7.x km s^{-1} anomaly lies beneath rapidly thinning continental crust (Figure 18). We have no constraints to confirm that rocks forming the seismic velocity anomaly on the Newfoundland margin are older than Mesozoic rifting, although this scenario is supported. by exhumed Variscan gabbros on the conjugate Iberian margin [*Capdevila and Mougenot*, 1988; *Pinheiro et al.*, 1996; *Manatschal et al.*, 2001].

[37] The crustal structure of the continental shelf and slope of the Iberian margin conjugate to SCREECH line 2 is not well known, because seismic reflection line LG-12 [*Krawczyk et al.*, 1996] and the seismic study of *Chian et al.* [1999] do not cover the proximal margin offshore Iberia (Figure 2b). Seismic line IAM-9, which crosses the Iberian continental slope ~40 km south of LG-12 (Figure 3), shows the crust thinning from 28 km to less than 10 km thickness
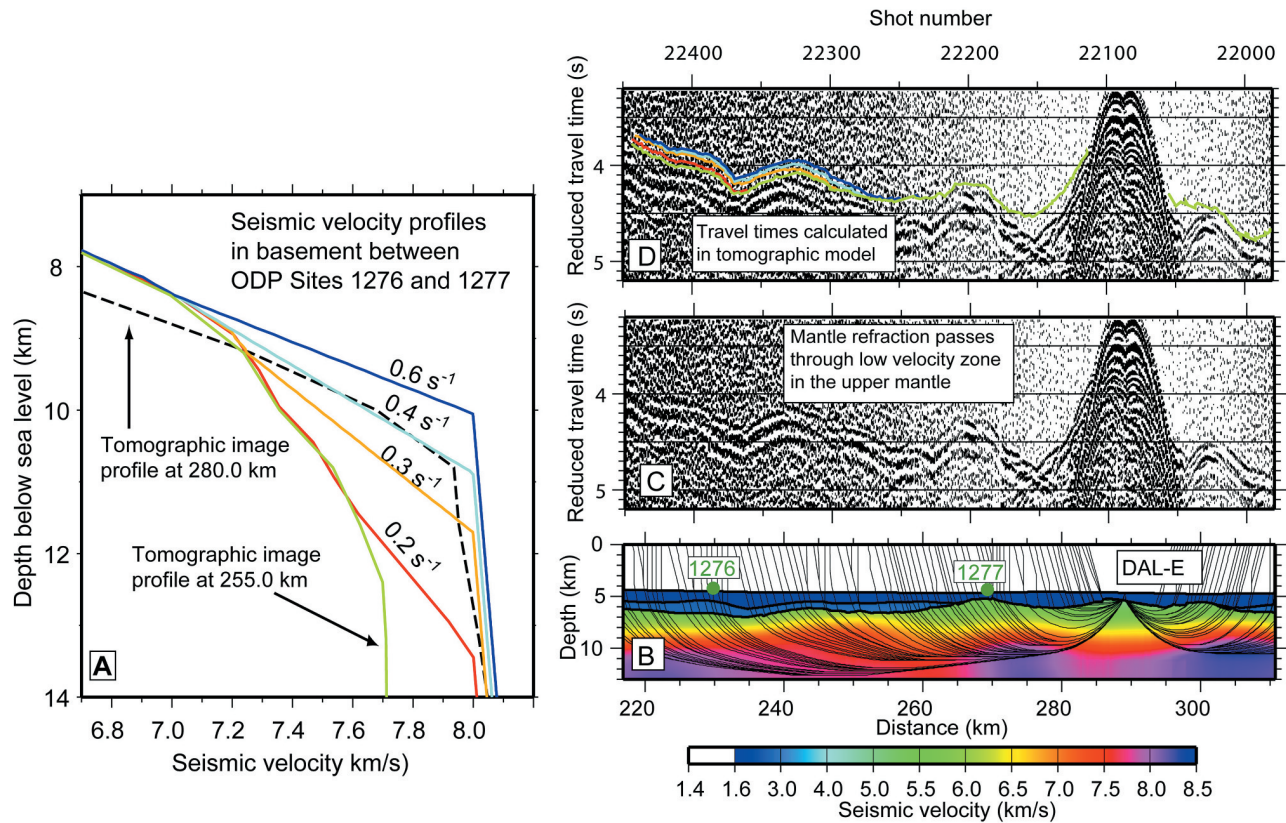
Figure 16. Ray trace test for receiver gather Dal-E. (a) Tomographic inversion, resulting in a seismic velocity profile at 255 km (green line) that differs strongly from adjacent areas. A velocity profile at 280 km is shown for comparison (black dash). We calculated traveltimes for four different velocity gradients in the upper lithosphere between 240 km and 260 km in our model (red, orange, cyan, and blue lines) to show that a steep velocity gradient is not warranted by the data. (b) Many ray paths for the seismic refractions recorded by instrument Dal-E from the northwest pass through the apparent low-velocity anomaly between ODP sites 1276 and 1277. (c) Seismic refraction data shown with a reduction velocity of 6.0 km s^{-1}. (d) Traveltime curves with colors corresponding to the velocity profiles in Figure 16a. The velocity profile with a gradient of 0.2 s^{-1} (red) results in traveltimes that are only slightly faster (50 ms) than the data, but steeper velocity gradients cause a larger advance.

over a distance of 80 km [*Dean et al.*, 2000]. The two other SCREECH profiles documented the same abrupt crustal thinning across the Newfoundland margin as on line 2 [*Funck et al.*, 2003; *Lau et al.*, 2006b]. Therefore a rapid transition from thick to thin continental crust appears to be a characteristic feature of the Newfoundland margins.

6.2. Transitional Basement

[38] The seafloor that lies between the continental slope and the beginning of rough basement at magnetic anomaly M3 (Figure 5) has been referred to as the transition zone (TZ). Most studies of the Newfoundland-Iberia rift focus on three simple hypotheses for the origin of the TZ, although a model that fits all geological observations would certainly have to be more complex:

[39] 1. The TZ may be highly extended continental crust that did not separate until about the time of chron M3. This interpretation would be favored if the U reflection were indeed a breakup unconformity associated with separation of continental crust [*Tucholke et al.*, 1989]. However, coring of postrift sediments at ODP Site 1276 shows a deep-water

setting [*Shipboard Scientific Party*, 2004a], so the U reflection did not form by subaerial or shallow marine erosion.

[40] 2. The relatively smooth basement of the TZ may be the surface of a rolling-hinge detachment fault [*Lavier et al.*, 1999] along which subcontinental mantle was exhumed to the seafloor. However, SCREECH MCS data show no clear evidence for allochthons of continental crust on the distal margins in the Newfoundland Basin [*Shillington et al.*, 2006]. Thus if mantle unroofing occurred off Newfoundland, it was not accompanied by fine-scale dissection of thinned continental crust such as that observed on the conjugate Iberian transect LG-12 [*Krawczyk et al.*, 1996].

[41] 3. Basement seaward of the last continental block at ~170 km (Figure 5) may have accreted by very slow seafloor spreading. Deep-tow magnetic results from the conjugate Iberian Abyssal Plain suggest that seafloor spreading did not initiate here before chron M3 (125–126 Ma) or M5 (128 Ma) [*Whitmarsh and Miles*, 1995]. Therefore an oceanic origin for the TZ basement below the U reflection would require that the oceanic crust was isolated on the Newfoundland margin by extremely asymmetric spreading or by a ridge jump at about the time of chron M3 or M5.
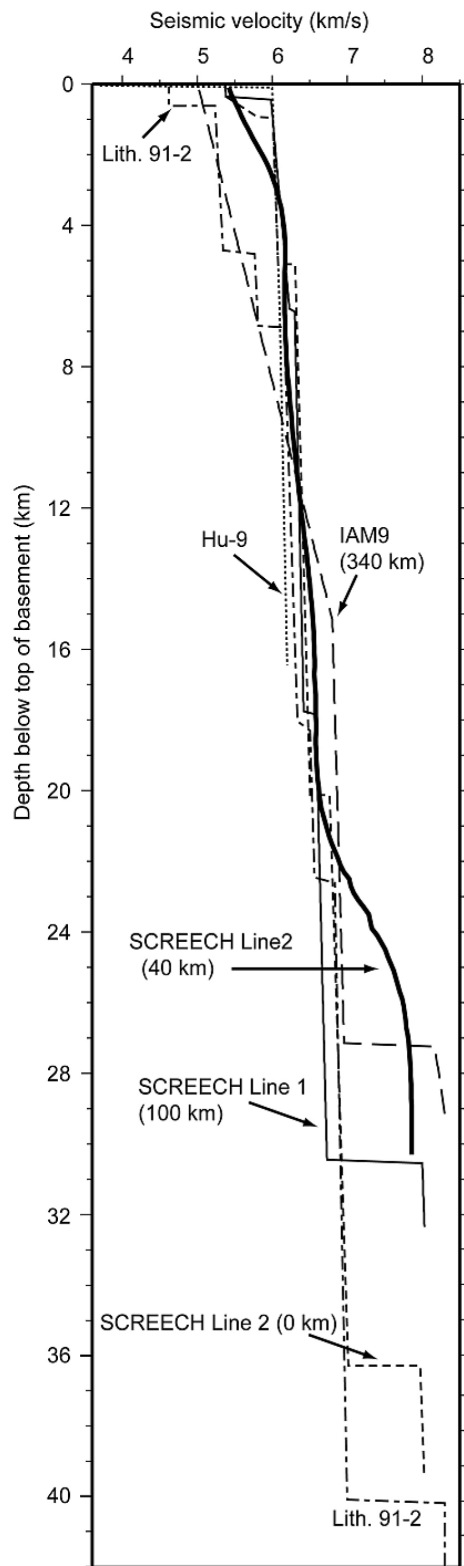
Figure 17. Seismic velocity profiles of continental crust. Profile Hu-9 is the most landward refraction line from *Todd and Reid* [1989]. Generally, the velocity profiles are similar, although the shallow layers with velocities between 4.5 and 5.5 km s^{-1} on line 91-2 offshore of the Avalon Peninsula [*Marillier et al.*, 1994] and IAM-9 near the continental slope of southern Iberia [*Dean et al.*, 2000] are absent on the three SCREECH transects.

However, the absence of clear magnetic anomalies predating chron M3 in the Newfoundland Basin [*Verhoef et al.*, 1996; Sibuet et al., submitted manuscript, 2006] argues against the accretion of oceanic crust with a significant magmatic component over this time period.

[42] The tomographic image of SCREECH line 2 (Figure 14b) shows a remarkably regular eastward thinning of the crust in the northern Newfoundland Basin from 6 km at 165 km distance to just 2 km at ODP Site 1276 (230 km). At 200 km in our model, we imaged ~4.5-km-thick crust of which the upper 3.0 km has a velocity lower than 6.5 km s^{-1} (Figure 19). Deeper in the crust the velocity gradient steepens until the seismic velocity exceeds 8 km s^{-1}. *Lau et al.* [2006b] found velocity structure on SCREECH line 3 in the southern Newfoundland Basin in the first ~60 km seaward of the continental slope. They interpreted this zone as highly extended continental crust, and we find this explanation equally plausible for the relatively smooth basement between the continental slope and ODP site 1276 on SCREECH line 2 (Figure 18). A 60- to 65-km-wide zone of highly thinned continental crust may therefore extend farther seaward than observed prerift sediments at 160 to 170 km (Figure 5) [*Shillington et al.*, 2006]. Coring at Site 1276 showed that the basement was in deep water when lower Albian and younger sediments were deposited [*Shipboard Scientific Party*, 2004a]. It is therefore unlikely that prerift sediments were eroded within the TZ during extension. New basement may have formed by exhumation of continental middle or lower crust shortly before crustal separation between Grand Banks and Iberia, which would explain some of the subsidence of the passive margin [*Driscoll and Karner*, 1998]. A thin layer of continental crust was also found on the conjugate profile in the Iberian Abyssal Plain [*Chian et al.*, 1999], but just 40 km to the south, this thin layer of continental crust appears absent [*Dean et al.*, 2000].

[43] The ray-tracing test (Figure 16) shows that the crystalline basement between 235 km and 270 km in our model of SCREECH line 2 probably consists of unroofed mantle (Figure 14b). The seismic velocities of 6.0 to 7.7 km s^{-1} in the top 5−6 km (Figure 18) are consistent with 60% to 15% serpentinization [*Christensen*, 2004]. A similar but 80-km-wide zone of serpentinized mantle was inferred seaward of the thinned edge of continental crust on SCREECH line 3 [*Lau et al.*, 2006b]. We assume that continental mantle was exhumed along the entire eastern margin of the Grand Banks much as it was in the area of the southern Iberia Abyssal Plain [*Dean et al.*, 2000]. On both margins, progressively deeper stratigraphic units of the continental lithosphere are exposed farther seaward in the TZ, which suggests that the Newfoundland-Iberia rift is symmetric on a large scale.

[44] A surprising outcome of our analysis of the SCREECH line 2 refraction data is that we obtained mantle velocities of at least 8.0 km s^{-1} beneath the lower continental slope and rise in the northern Newfoundland Basin. In contrast, SCREECH lines 1 and 3, and the data from *Reid* [1994] in the southern Newfoundland Basin, show upper mantle seismic velocities between 7.6 and 7.9 km s^{-1} over several kilometers depth in the TZ. We cannot attribute this difference to the inversion methods applied to the data sets, because an inversion with RAYINVR [*Zelt and Smith*,
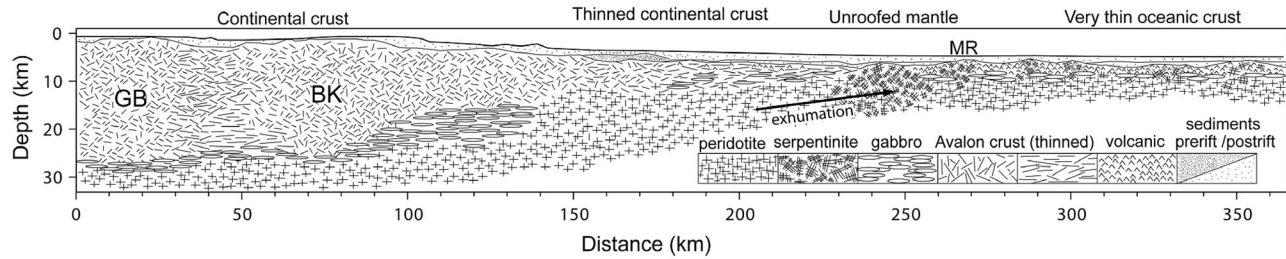
Figure 18. Interpretive sketch of SCREECH line 2 based on our seismic velocity model. Prerift sediments overlie crust at the base of the continental slope, but they have not been detected over the thinned continental crust in the distal margin. We therefore propose that middle and lower continental crust of the Grand Banks are exhumed in this area. A relatively narrow zone (~25 km) of exhumed continental mantle lies in between the seaward edge of thinned continental crust and the oldest basaltic crust. MR, Mauzy Ridge. Farther seaward the depth of mantle serpentinization appears to be much shallower, and the basement may incorporate volcanics produced by an incipient slow spreading mid-ocean ridge.

1992] of the line 2 refraction data also produced high mantle velocities [*Nunes*, 2002]. Therefore the upper mantle in the TZ on SCREECH lines 1 a 3 is probably partially serpentinized, whereas the mantle underlying the thin TZ crust on line 2 between the continental rise and ODP Site 1276 is probably relatively pristine. *Pérez-Gussinyé et al.* [2001] presented a numerical model of serpentinization and melt production for various amounts of extension and rift duration. Their results show that if the mantle remains too hot (>500°C) for serpentine to be stable [*Ulmer and Trommsdorff*, 1995; *Evans*, 2004], several kilometers of melt would be produced during pure shear extension, unless the amount of extension β is small. Such pure shear extension seems to be inconsistent with our results which show highly thinned crust (2–6 km) with relatively low seismic velocities (5.5–6.5 km s^{-1}) that preclude large volumes of synrift igneous intrusions. On the other hand, if the final stage of rifting involved a significant amount of simple shear, relatively hot mantle rock might be emplaced in the TZ without generating much melt [*Latin and White*, 1990]. Thus the distributions of apparently serpentinized and unserpentinized mantle in the TZ of SCREECH line 2 may reflect the temporal and thermal evolution of one or more major detachment faults (Figure 18).

6.3. Oceanic Basement

[45] The seismic velocity structure seaward of Mauzy Ridge at 260 to 270 km in our model (Figure 14b) is laterally more heterogeneous than the transitional basement discussed in the previous section. Basement ridges with ~1 km elevation coincide with crustal blocks that appear to be up to 6 km thick, each spanning only ~15 km along our profile. The depressions in between these basement highs have a crustal thickness of perhaps just 2 or 3 km, as can be seen by the undulating 7.5 km s^{-1} velocity contour. The seismic velocity increases steadily with depth inside the crests from 5.0–5.5 km s^{-1} to 7.5 or 8.0 km s^{-1}. Velocity gradients in the mantle are comparatively weak (Figure 20). *Shillington et al.* [2006] interpreted the reflectivity in their MCS data as igneous layering in the swales and faulting throughout the crust. Rough basement, restricted volcanism, thin crust and diffuse Moho are common characteristics of seafloor spreading at very slow rates [*Cannat*, 1996; *White*

et al., 2001]. In order to estimate the spreading rate along SCREECH line 2 during the Cretaceous, we measured the distance along a flow line between magnetic anomalies M0 and C33. Using the timescale of *Gradstein et al.* [1994] we obtain a spreading half rate of about 9 mm yr^{-1}. It must be noted that *Gradstein et al.* [2004] adjusted their estimate for the age of M0 from 120 Ma to 125 Ma, which results in a slower half spreading half rate of 7.5 mm yr^{-1}.

[46] The 7.5 or 9.0 mm yr^{-1} half spreading rate on our seismic transect in the Newfoundland Basin is smaller than estimates for most of the modern Mid-Atlantic Ridge, but it
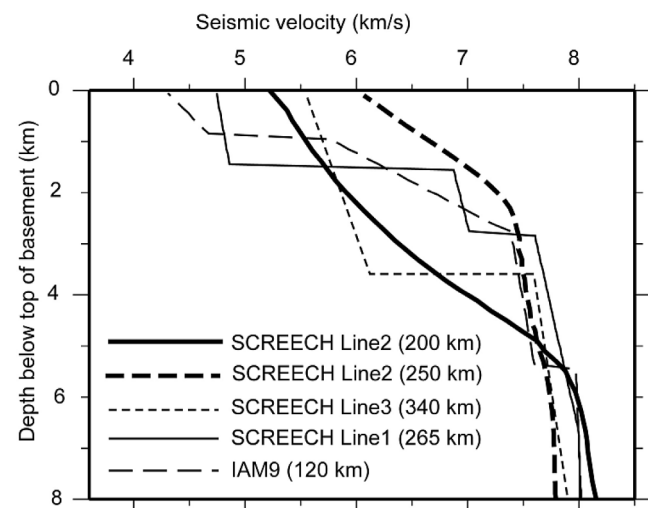

Figure 19. Seismic velocity profiles of the TZ. On both SCREECH line 2 and SCREECH line 3, there appears to be a wide zone of thin crust with relatively low seismic velocities beneath the continental rise, interpreted as extended continental crust. On both SCREECH line 1 and IAM-9, seismic velocities increase more rapidly with depth in the TZ, which has been interpreted as thin oceanic crust near Flemish Cap [*Funck et al.*, 2003] and unroofed continental mantle on the Iberian margin [*Dean et al.*, 2000]. At the outer limit (250 km) of the TZ on line 2 (bold dash) we find higher velocities immediately below basement that also indicate mantle exhumation.
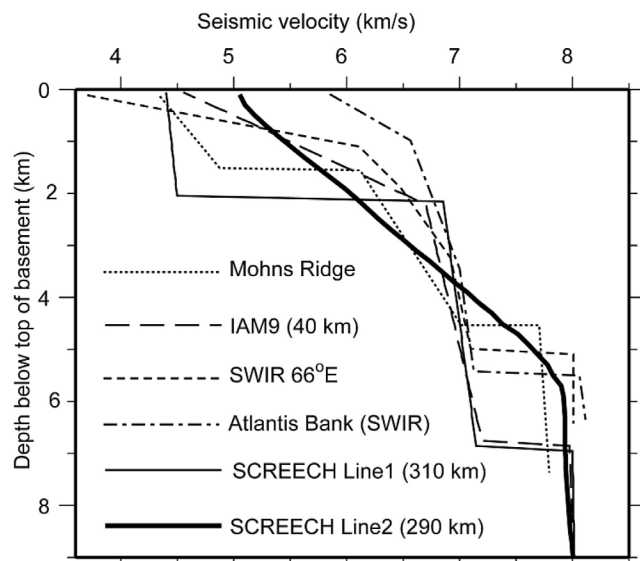
Figure 20. Seismic velocity profiles from the most seaward sections of SCREECH line 2 in this study, SCREECH line 1 [*Funck et al.*, 2003], and IAM-9 on the Iberia margin [*Dean et al.*, 2000]. For comparison we also show the seismic structure of the Atlantis Bank oceanic core complex [*Muller et al.*, 2000], from oceanic crust at the eastern limb of the Southwest Indian Ridge (SWIR) [*Muller et al.*, 1999], and oceanic crust at Mohns Ridge [*Klingelhöfer et al.*, 2000]. The profile from SCREECH line 2 shows the seismic structure of a basement ridge at 290 km (solid thick line).

is comparable to the rate observed at both the Mohns Ridge in the Norwegian Sea [*Géli et al.*, 1994] and at the Southwest Indian Ridge [*Patriat and Segoufin*, 1988; *Chu and Gordon*, 1999]. Some sections of these ultraslow spreading ridges consist of short magmatic segments bounded by transform faults, but other volcanic centers are instead linked by oblique nonmagmatic segments [*Grindlay et al.*, 1998; *Dick et al.*, 2003]. The seismic refraction studies of *Muller et al.* [1999] on the Southwest Indian Ridge and *Klingelhöfer et al.* [2000] on the Mohns Ridge show that oceanic crust formed in the ultraslow spreading regime is very similar in thickness and structure to the crust at the seaward end of SCREECH line 2 and other sections of the Newfoundland-Iberia rift (Figure 20). The reduced mantle upwelling and conductive heat loss at mid-ocean ridges spreading at less than 10 mm yr^{-1} half rate can explain the much thinner (<4 km) crust than normal oceanic crust (6 to 7 km) [*White et al.*, 2001]. Gabbros appear to make up most of the lower oceanic crust at spreading rates higher than 10 mm yr^{-1} half rate, but at lower spreading rates these intrusives may be retained deeper in the mantle lithosphere [*Lizarralde et al.*, 2004]. The very thin oceanic crust produced at ultraslow spreading rates may therefore consist of a thin volcanic layer on top of serpentinized mantle rock. Gabbros and highly serpentinized mantle rock have similar seismic velocities [*Cannat*, 1996], so we must be cautious to equate our seismic

estimates of crustal thickness with igneous production at a spreading center.

[47] At the seaward limit of the TZ (260 km), the onset of rough basement topography in the Newfoundland Basin coincides with a lateral change in seismic velocity (Figure 14b), from weak to strong seismic velocity gradients below basement. We compare the ~1 km basement relief seaward of chron M3 on SCREECH line 2 with Atlantis Bank, a well-studied, young oceanic core complex at the Southwest Indian Ridge [*Dick et al.*, 2000; *Baines et al.*, 2003]. Seismic velocities of 5.5 to 6.0 km s^{-1} near the surface of this highstanding platform [*Muller et al.*, 2000] are consistent with uplift and exhumation of lower crustal and upper mantle rocks. In comparison to Atlantis Bank, the seismic velocities of abyssal hills along SCREECH line 2 are significantly lower to a depth of 5 km below the basement (Figure 20). Perhaps the basement ridges in the seaward portion of SCREECH line 2 consist of exhumed lower crustal rocks and mantle peridotites that were almost fully serpentinized. At larger depth beneath the rough basement, our inversion showed little evidence for seismic velocities between 7.0 and 7.5 km s^{-1}, which would correspond to mantle peridotites with roughly 20% or 10% serpentinization [*Christensen*, 2004]. The in situ replacement of ultramafic rocks by serpentinite requires a volume expansion [*Coleman*, 1971] that may constrict the deep percolation of seawater, and hence result in a serpentinization front that separates rocks that underwent massive hydrothermal alteration from relatively fresh mantle rocks [*Minshull et al.*, 1998]. However, the most likely occurrence of serpentinization deep into the oceanic lithosphere is where active faults cut through the entire crust [*Francis*, 1981]. In the SCREECH line 2 MCS data the best example of extensional faults penetrating the crust are found between 310 and 320 km in our model [*Shillington et al.*, 2006]. Deep serpentinization of mantle rocks can be ruled out here, because crust of 2–4 km thickness overlies mantle with seismic velocities larger than 8.0 km s^{-1} (Figure 14b).

[48] If some bathymetric highs on SCREECH line 2 were not formed by exhumation of mantle peridotites, they must have a volcanic origin instead. Large volcanic constructions are thought to form in the axial valley of the ultraslow spreading Southwest Indian Ridge, but they are widely spaced in the spreading direction [*Cannat et al.*, 1999]. *Mendel et al.* [2003] found that slow spreading centers with greater melt supply produce abyssal hills of larger width and height. Their segment 27 of the Southwest Indian Ridge at 50°30'E formed off-axis ridges that are similar in size to those imaged seaward of Mauzy Ridge on SCREECH line 2 [*Mendel et al.*, 2003; *Shillington et al.*, 2006]. Small amounts of gabbro and basalt were recovered from ODP Site 1277 [*Shipboard Scientific Party*, 2004a], but these igneous rocks did not derive from the peridotites that were cored deeper in Mauzy Ridge [*Müntener and Manatschal*, 2006]. Ophiolite studies in the Alps similarly show that the oldest igneous crust formed in the Tethyan Ocean did not derive from decompression melting of the mantle peridotites that were accreted to the new lithosphere [*Piccardo et al.*, 2004; *Rampone*, 2004]. More plausibly, asthenospheric melts percolated through relatively thick lithosphere to form the first mid-ocean ridge basalts in both ocean basins [*Müntener et al.*, 2004; *Piccardo et al.*, 2004].

[49] The tomographic image of SCREECH line 2 (Figure 14b) shows a general increase in mantle seismic velocity from Mauzy Ridge to the seaward end of the seismic profile. This trend may imply that hotter peridotites were exhumed farther in the Newfoundland Basin, leading to shallower serpentinization since chron M1 (Late Barremian) on SCREECH line 2. The magnetic anomalies older than M1 are weak near SCREECH line 2 (Sibuet et al., submitted manuscript, 2006), but M0 is a more prominent positive magnetic anomaly that correlates with the larger J anomaly [*Tucholke and Ludwig*, 1982] south of the Newfoundland Fracture Zone (Figure 1). The stronger anomaly M0 in the vicinity of SCREECH line 2 (Figure 4) may represent the first significant magmatic event in this area, perhaps 3 to 4 Ma after the start of exhumation of hotter mantle lithosphere. Thin, heterogeneous crust lies farther seaward on SCREECH line 2, which indicates that early seafloor spreading in the Newfoundland Basin was magma-starved at least until late Aptian time (Figure 14b).

[50] The suggested onset of hotter mantle exhumation in the Newfoundland Basin after Late Barremian roughly coincides with a phase of extension in the Jeanne d'Arc Basin [*Driscoll et al.*, 1995]. It is nevertheless difficult to correlate the evolution of crustal structure in the Newfoundland Basin with the stratigraphic record of the adjacent margins, because different sedimentary basins of the Newfoundland-Iberia rift experienced extension during different episodes [*Alves et al.*, 2002, 2006]. The Newfoundland proximal and distal margin both accommodated extension until late Aptian or earliest Albian [*Driscoll et al.*, 1995; *Tucholke et al.*, 2006], which indicates that the lithosphere at the rift axis was under extensional stress when new crust was formed seaward of chron M0 (Figure 14b).

6.4. Along-Strike Variability

[51] The three SCREECH transects show evidence for crustal thinning, variable amounts of mantle exhumation, and very thin oceanic crust at the seaward end of each of these profiles. Together with seismic data from the conjugate Iberian margin, the SCREECH results give an impression of the along-strike variability in the geometry of the Newfoundland-Iberia rift. We compiled published seismic refraction studies to illustrate the evolution of the rifted-margin pair. In Figure 21 we show the interpreted distribution of crustal and mantle rocks prior to breakup along three profiles that extend in the landward direction as far as we have constraints from seismic refraction studies. The profiles represent a cross section of the rift during the time that volcanic upper crust is interpreted on top of exhumed mantle.

[52] The reconstruction of Figure 3 shows that SCREECH line 3 and IAM-9 are not precisely conjugate, but they together represent the best summary transect across the southeastern Grand Banks and southern Iberia Abyssal Plain (Figure 21c). Compared to the gentler slope of the Moho at southern Iberia margin [*Dean et al.*, 2000] the southeastern Grand Banks thins very rapidly from 35 km on the shelf to a thickness of just a few kilometers to the southeast [*Lau et al.*, 2006b]. In the central Newfoundland Basin, continental crust was stretched to less than 10 km thickness over a distance of 110 km [*Lau et al.*, 2006b], while almost no thin continental crust is found at the

conjugate side [*Dean et al.*, 2000]. This asymmetry indicates that the continental crust separated southeast of the rift axis. The zone of exhumed continental mantle is widest along IAM-9 [*Dean et al.*, 2000], but continental mantle was also exhumed in the southern Newfoundland Basin before oceanic spreading started in late Hauterivian time (~129 Ma) [*Whitmarsh and Miles*, 1995; *Lau et al.*, 2006a].

[53] Farther north, seismic profiles of SCREECH 2 (this study) and Cam144 [*Chian et al.*, 1999] together define the deep crustal structure of the rift near the drill sites of ODP leg 149, 173 and 210. Unfortunately, refraction data on this transect do not reach thick continental crust on either side of the rift. Both rifted margins have thin continental crust in the TZ, but they are nevertheless different in nature. As we noted earlier in the Discussion, the apparent absence of prerift sediments on the smooth basement of the TZ of the Newfoundland Basin suggests that it was formed by exhumation of midcrustal or lower crustal rocks. In contrast, in the Iberia Abyssal Plain, blocks of continental upper crust appear to be cut by seaward dipping normal faults along Cam144. As a result, the basement relief along Cam144 is higher and the continental crust is thinner [*Chian et al.*, 1999] than in the TZ of SCREECH line 2 (Figure 21b). Continental mantle was exhumed on both margins over a distance of ~50 km before the first basaltic crust formed in late Barremian time (~124 Ma).

[54] Crustal structure is more complex in the northern part of the Newfoundland-Iberia rift between Flemish Cap and Galicia (Figure 21c). Here Triassic to Jurassic extension initially resulted in the formation of the Flemish Pass Basin [*Keen and Barrett*, 1981] and Galicia Interior Basin [*Pérez-Gussinyé et al.*, 2003] (Figure 21a). The relatively thin crust of the Galicia Bank (20 km) suggests that this continental block also stretched during late Jurassic rifting. The sum of thin (<10 km thickness) continental crust on the conjugate margins of Galicia Bank (110 km) and Flemish Cap (40 km) is approximately the same as the combined width of thin crust on the Grand Banks (120 km) and southern Iberia Abyssal Plain (40 km) to the south. Contrary to the southern margin pair, more thin continental crust lies to the southeast (offshore Galicia) [*Whitmarsh et al.*, 1996; *Zelt et al.*, 2003] than to the northwest (near Flemish Cap) [*Funck et al.*, 2003]. The location where the continental crust was breached may therefore have shifted westward within the rift zone when the rift tip reached the Flemish Cap and Galicia Bank. Breakup between Flemish Cap and Galicia Bank led to the formation of Peridotite Ridge on the Iberian side [*Boillot et al.*, 1980], but the separation of continental crust was soon followed by the formation of the first oceanic crust near Flemish Cap in Late Barremian time (~123.5 Ma). It appears from our seismic results that on both SCREECH lines 1 and 2 the first oceanic basement accreted soon after magnetic anomaly M0 was formed. However, on SCREECH line 2 the continental crust was breached before anomaly M3 time (~126 Ma), while on SCREECH line 1 thinned continental crust extends beyond the anomaly M3 pick of *Srivastava et al.* [2000] at the foot of Flemish Cap [*Funck et al.*, 2003].

[55] The large-scale comparison between the three transects of Figure 21 shows that all have significant asymmetries. In addition, the central (Figure 21b) and southern transect (Figure 21c) share some important similarities. On
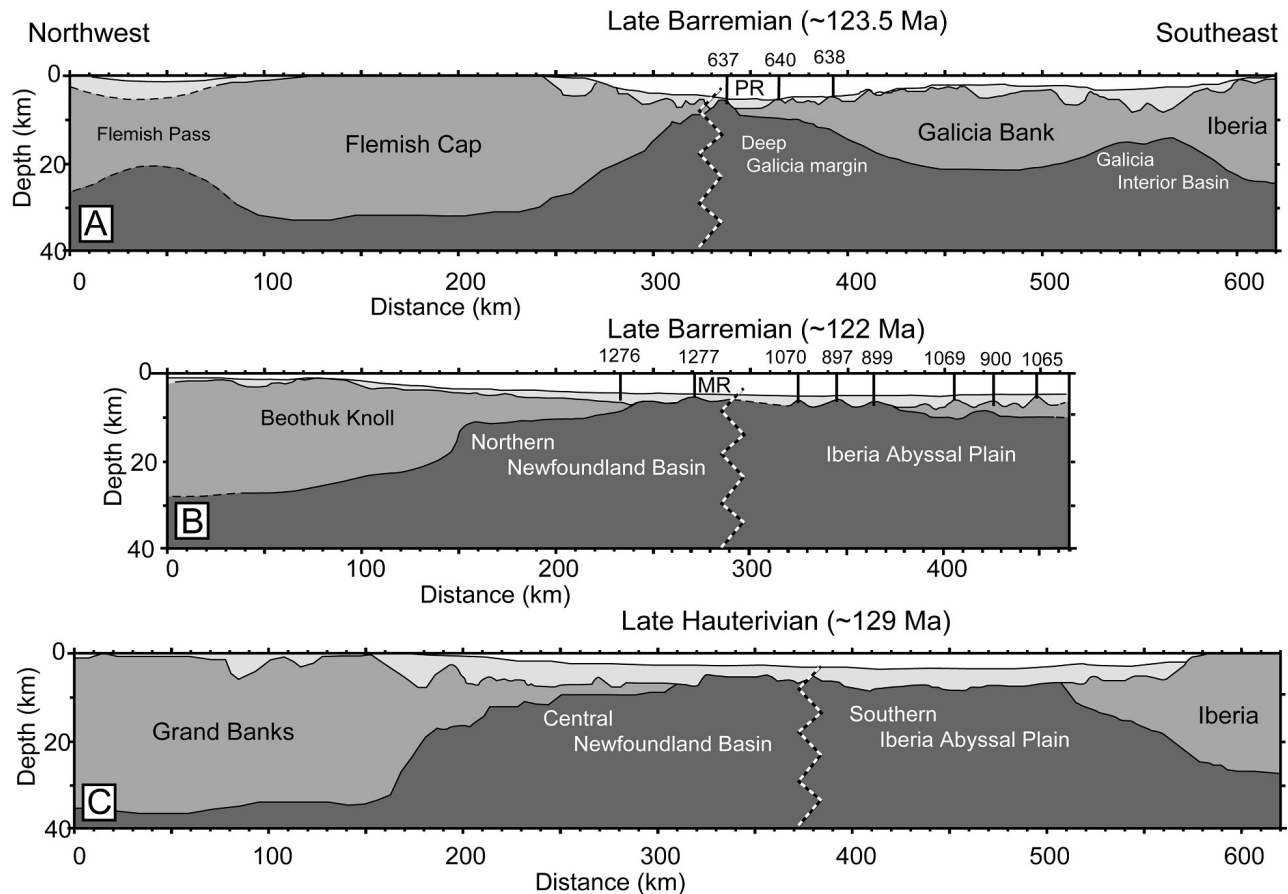
Figure 21. Three composite transects across the Newfoundland-Iberia rift. Each profile shows the structure of a cross section of the rift just prior to the accretion of the first basaltic upper crust, which occurred progressively later as the rift propagated to the north. Synrift and postrift sediments are shown in light gray, continental crust in middle gray, and continental mantle in dark gray. Labeled black vertical lines are ODP drill sites. The dashed jagged lines are placed where the first basaltic upper crust is formed at the end of continental rifting. (a) Northern section of the rift. Thinner crust is found in the Flemish Pass [*Keen and Barrett*, 1981] and the Galicia Interior Basin [*Pérez-Gussinyé et al.*, 2003]. The structure of Flemish Cap and the deep Galicia margin are adapted from SCREECH line 3 [*Funck et al.*, 2003] and *Whitmarsh et al.* [1996]. PR is Peridotite Ridge. (b) Similar transect along SCREECH line 2 (this study) and the drilling transect of the Iberia Abyssal Plain [*Chian et al.*, 1999]. MR is Mauzy Ridge. (c) Southern transect. The geometry of the southeastern Grand Banks [*Lau et al.*, 2006b] and southern Iberia Abyssal Plain [*Dean et al.*, 2000] is shown.

both SCREECH line 2 and SCREECH line 3, continental crust thins rapidly in the seaward direction beneath the continental slope of the Grand Banks. A wide zone of very thin continental crust extends from the continental slope farther into the Newfoundland Basin. A zone of exhumed continental mantle of variable width lies between the thin continental crust and oldest ocean crust of both the Newfoundland Basin and Iberia Abyssal Plain. The zone of exhumed continental mantle (Figures 16 and 18) on SCREECH line 2 is relatively narrow compared to SCREECH line 3 and the conjugate Iberia Abyssal Plain. Nevertheless, for a few million years continental mantle appears to have been exhumed between ODP Sites 1276 and 1277 without significant melt production (Figure 14b). In contrast to the two transects across the Grand Banks (Figures 21b and 21c), the rifted margin southeast of

Flemish Cap (Figure 21a) is very narrow, and continental breakup here happened relatively fast.

7. Conclusions

[56] A total of 24 ocean bottom seismometers and hydrophones deployed on SCREECH line 2 recorded seismic refractions over distances greater than 100 km on the continental shelf of the Grand Banks and more than 70 km in the Newfoundland Basin. These offsets were sufficient to sample full thickness crustal structure along this seismic transect. A tomographic inversion of first arriving travel-times gives a well-constrained image of seismic velocities that, together with other data from the Newfoundland-Iberia margins, provide new insight in the development of the

Newfoundland-Iberia rift. We summarize our findings as follows:

[57] 1. The continental crust of the eastern rim of the Grand Banks has seismic velocities varying from ~5 km s^{-1} near the basement to ~7 km s^{-1} near the Moho. This compares well with other estimates of the deep crustal structure of the Avalon terrane to the west. Although a crustal thickness of ~35 km is found elsewhere on the Grand Banks, we find only ~27-km-thick crust at the landward side of SCREECH line 2. Thinner crust beneath Flemish Pass and Beothuk Knoll indicates that the eastern Grand Banks underwent ~25% stretching in Late Triassic and Early Jurassic before the extension localized in the distal margin. Extension beginning in Late Triassic is also manifested by the development of the adjacent Jeanne d'Arc Basin.

[58] 2. At the base of the continental slope, the continental crust thins rapidly to a thickness of just 6 km. All three SCREECH transects [Funck et al., 2003; Lau et al., 2006b] show a transition from thick crust beneath the continental shelf to very thin crust in the distal margin over ~50 km, which is more abrupt than on the conjugate Iberian margins. The S reflection on the Galicia margin may be the expression of a west dipping detachment fault that produced this asymmetry in crustal structure between Flemish Cap [Hopper et al., 2006] and the conjugate deep Galicia margin [Reston et al., 1996]. There is no evidence for a similar rift-wide lithospheric detachment fault between the Grand Banks and Iberia Abyssal Plain.

[59] 3. We found a layer thicker than 5 km at a depth of 15 to 20 km with seismic velocities between 7.0 and 7.5 km s^{-1} beneath the continental slope. If the seismic results are sufficiently constrained, this layer may correspond to strong, possibly Variscan mafic lower crust that facilitated the localization of strain in the Newfoundland-Iberia rift. The seismic results of SCREECH line 2 give no other evidence for magma production during continental rifting.

[60] 4. Farther seaward, a ~60-km-wide zone of deeply-lying, relatively smooth basement has seismic velocities similar to continental crust. The crustal thickness thins from 6 km near the base of the continental slope to just 2 km in the Newfoundland Basin. There is no evidence for prerift sediments on top of the basement [Shillington et al., 2006]. It is therefore possible that middle or lower continental crust was exhumed to surface in the TZ of SCREECH line 2.

[61] 5. The zone of thinned continental crust appears much wider in the southern Newfoundland Basin [Lau et al., 2006b] than in the conjugate southern Iberia Abyssal Plain [Dean et al., 2000]. This difference in the distribution of thin continental crust is less pronounced between SCREECH line 2 and the conjugate Iberian margin [Chian et al., 1999]. To the north, the results from SCREECH line 3 [Funck et al., 2003] and the deep Galicia margin [Whitmarsh et al., 1996; Zelt et al., 2003] show that most crustal thinning occurred in the Iberia side of the rift, and that final extension leading to breakup was localized on the west side of the rift near thick crust of Flemish Cap (Figure 21).

[62] 6. Beyond the edge of thinned continental crust along SCREECH line 2 lies a ~25 km wide zone between ODP sites 1276 and 1277 with velocities >6 km s^{-1} just below the basement that increase to 7.7 km s^{-1} to a depth of at least 5 km (Figure 16). This area appears to be exhumed

mantle that was serpentinized to large depth. Similar zones of exhumed continental mantle are much wider (at least 60 to 160 km) on SCREECH line 3 [Lau et al., 2006b] and in the Iberia Abyssal Plain [Chian et al., 1999; Dean et al., 2000].

[63] 7. Between ODP Site 1277 to the seaward end of SCREECH line 2, the vertical seismic velocity gradient in the first 5 km into basement is much higher than between ODP Sites 1276 and 1277. The steeper seismic velocity gradient suggests that exhumed peridotites were locally too hot to be serpentinized. The transition from deep to shallow serpentinization in the crystalline basement roughly coincides with the timing of the J anomaly, a strong linear magnetic anomaly that terminates south of SCREECH line 2.

[64] 8. The crust that lies at the seaward end of SCREECH line 2, which formed in Aptian and perhaps earliest Albian time, shows variations in thickness between 3 and 6 km. The seismic velocity increases rapidly with depth from ~5.5 to 8.0 km s^{-1}, which suggests that both gabbros and serpentinized mantle rock can be present in volumetrically limited amounts. We therefore interpret this area as the oldest, ultraslow spreading oceanic crust in the rift basin.

[65] 9. Along the Grand Banks (SCREECH line 1 and line 3) the first oceanic crust formed after a few million years of exhumation of continental mantle. In contrast, along SCREECH line 1 oceanic crust formed soon after the continental crust separated between Flemish Cap and Galicia Bank [Funck et al., 2003; Hopper et al., 2006]. The final rifting phase between crustal separation and the onset of oceanic spreading was therefore completed much faster between Flemish Cap and Galicia than between the Grand Banks and southern Iberia.

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W. S. Holbrook, Department of Geology and Geophysics, University of Wyoming, 1000 E. University Ave., Dept. 3006, Laramie, WY 82071-3006, USA.

J. R. Hopper, Department of Geology and Geophysics, Texas A&M University, 3115 TAMU, College Station, TX 77843-0000, USA.

H. C. Larsen, Integrated Ocean Drilling Program, Hokkaido University, N21W10 Kita-ku, Sapporo, 001-0021, Japan.

K. E. Louden, Department of Oceanography, Dalhousie University, Halifax, Nova Scotia B3H 4J1, Canada.

G. T. Nunes, ExxonMobil Exploration, P.O. Box 4778, Houston, TX 77210, USA.

D. J. Shillington, National Oceanography Centre, School of Ocean and Earth Science, University of Southampton, Waterfront Campus, European Way, Southampton SO14 3ZH, UK.

B. E. Tucholke, Department of Geology and Geophysics, Woods Hole Oceanographic Institution, Clark 241, MS 22, Woods Hole, MA 02543, USA.

H. J. A. Van Avendonk, University of Texas at Austin Institute for Geophysics, 4412 Spicewood Springs Road, Austin, TX 78759, USA. (harm@ig.utexas.edu)