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The rifting history of the Newfoundland-Iberia conjugate margins: A geodynamic analysis

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THE RIFTING HISTORY OF THE NEWFOUNDLAND-IBERIA CONJUGATE MARGINS: A GEODYNAMIC ANALYSIS

by

DAVID L. TETT

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE MASTER OF ARTS

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ABSTRACT

The Rifting History of the Newfoundland-Iberia Conjugate Margins: a Geodynamic Analysis

by

David L. Tett

Rifting between Newfoundland and Iberia occurred in two distinct phases—the first late Triassic to early Jurassic, the second late Jurassic to early Cretaceous—culminating in the creation of the North Atlantic Ocean. A dynamic modelling method was used to examine the implications of multiple phases of rifting on the development of the Newfoundland-Iberia conjugate margins.

The models predicted a lack of magmatism on these margins, and suggested that extension was significantly greater in the second rifting phase than in the first; these predictions agree with geological observations. The models could not predict the existence of highly thinned continental crust on both conjugate margins, however. In addition, a set of generic models roughly based on the Newfoundland and Iberia margins suggested that, where two rift phases occur, the site of the original rift usually will not be favored for extension when stretching resumes.
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Chapter 1. Introduction.

In past studies of the processes of continental breakup and ocean basin formation, geoscientists have usually characterized rifting as a single, continuous event. The "classical" rifting scenario begins with horst-and-graben normal faulting, followed by a concentration of extension in a narrow zone and finally a pulse of volcanism indicating the initiation of sea-floor spreading.

The opening of the North Atlantic Ocean, where Newfoundland was once adjacent to the Iberian peninsula (Figure 1), is not this simple, however. On these margins, there is evidence for two phases of rifting, with an intervening tectonically-quiet phase; ocean-floor creation did not begin until after the second phase of rifting successfully broke North America and Iberia apart. In other words, the "driving forces" of rifting appear to have stopped for a period of time on these margins. What implications did such a tectonic "resting phase" have for the rifting history as well as the features observed today on the Newfoundland and Iberia margins?

To attempt to answer this question, I have employed a dynamic modelling method used most recently by Harry and Sawyer (1992a,b). I have devised a set of models which incorporate the resting phase. This contrasts with the models of previous modellers, who have included only one rifting "event" (e.g., Dunbar and Sawyer, 1989a; Chéry et al., 1990; Keen and Boutilier, 1990; Bassi, 1991; Harry and Sawyer, 1992a,b; Bassi et al., 1993).

I will begin by reviewing the geological and geophysical observations from the Newfoundland and Iberia margins, and will summarize the history of both Paleozoic mountain-building and Mesozoic rifting and continental breakup. I
Figure 1. Location of relevant physiographic features on the anomaly 34 reconstruction for the central North Atlantic. Contours indicate water depth. Lines indicate locations of Figures 16 and 17. (after Masson and Miles, 1984)
then discuss the implications of the subsidence amounts on both margins on
the rifting style. I follow with an examination of the consequences of multi-
phase rifting in a set of generic models roughly based in the Newfoundland and
Iberia margins. Finally, I attempt to create a model which approximates the
crustal thickness and subsidence profile and the patterns of extension of the
Newfoundland and Iberia margins.
Chapter 2. Geologic Setting.

DESCRIPTION OF THE NEWFOUNDLAND-IBERIA CONJUGATE MARGINS

The Newfoundland and Iberian margins are, to the first order, grossly similar. Each margin contains two sets of basins: one landward of the continental shelf break, and the other seaward of the shelf break. Thus, four basin sets comprise the conjugate margins. They are, from west to east, the basins of the Grand Banks of Newfoundland (GBN), the Newfoundland basin (NB), the Iberian Abyssal Plain (IAP), and the Lusitanian basin (LB). I will describe the margins within this framework (Figures 2 and 3).

The Grand Banks of Newfoundland. The Grand Banks is a 450-kilometer-wide continental shelf off Newfoundland (Tankard and Welsink, 1987) whose water depth averages less than 200 m (Welsink et al., 1989). The Grand Banks is underlain by continental crust between 32 and 38 km thick (Keen and de Voogd, 1988; Tankard and Welsink, 1987, 1989; Tankard, 1990; Reid and Keen, 1990). A 7.2 km/sec layer lies at the base of the crust, below crust of 5.8 to 6.1 km/sec; this high-velocity layer may represent magmatic underplating or intrusion that occurred during extension (Reid and Keen, 1990). In addition, the crust beneath Newfoundland itself ranges between 40 and 45 km in thickness (Keen et al., 1990).

The surface of the Grand Banks consists of many half-graben basins which were created during two diachronous orientations of rifting (Figure 4) (Tankard and Welsink, 1987). The first orientation separated North America from Iberia, and created northeast-southwest trending basins primarily on the southeast flank of the Grand Banks (for example, the Jeanne d'Arc, Carson,
Figure 2. Seismic structure map of the Grand Banks compiled at 1:1,000,000 scale, and simplified to emphasize major faults. Note positions of Jeanne d'Arc basin, Carson basin, Flemish Cap, Newfoundland Basin, and Southeast Newfoundland Ridge. The Southeast Newfoundland Ridge is a major transfer zone marking the southern edge of the Newfoundland basin. (Welsink et al., 1989)
Figure 3. Western Iberia and the adjacent Atlantic Ocean. Contours represent water depth in meters. Surface geology is shown on the continent. Dashed lines are magnetic anomalies 33, 34, and M0. P = Porto Seamount, V = Vigo Seamount. (Wilson et al., 1989)
Figure 4. Principal structural elements of the Grand Banks. Contours represent water depth in meters. Note the two margin (northeast and southeast) where the Grand Banks experienced its two major phases of extension, the first southeast-directed (south of Flemish Cap), the second northeast-directed (north of Flemish Cap). (Tankard et al., 1989)
and Horseshoe basins) (Tankard and Welsink, 1987). The second orientation separated Britain from North America, and caused the formation of basins trending roughly northwest-southeast on the northeast side (primarily the Orphan basin) (Tankard and Balkwill, 1989). We are concerned here with the basins formed in the first orientation, since these were the basins created by the separation of Newfoundland from Iberia.

Located on the southeast flank of the Grand Banks, the Jeanne d'Arc basin is one of the more thoroughly studied basins of the Grand Banks. This basin is about 14 km deep and 50 km wide (Tankard and Welsink, 1987). Although there is much structural interaction between the two phases of rifting in the northern part of the basin (Tankard and Welsink, 1987), the northwest-southeast rifting phase is the only phase evident in the southern half. The basin is floored by the down-to-the-east Murre fault--which is listric and becomes horizontal at a depth of about 22 km--and by an unconformity at the base of the Mesozoic section (Tankard and Welsink, 1989; Keen et al., 1987).

Two other notable basins on the southeast Grand Banks are the Carson basin and the Salar basin (Figure 5). The Carson basin lies slightly seaward of the Jeanne d'Arc basin (Figure 4), and formed along listric faults similar to the Murre fault (Tankard and Welsink, 1989). It contains up to 10 km of Mesozoic and Cenozoic sedimentary rocks (Grant et al., 1988). The Salar basin lies east of the Carson basin, underneath the continental slope and rise (Figures 4 and 5). Both basins are floored by extensive halite-rich evaporite deposits (Austin et al., 1989; Holser et al., 1988).

The Newfoundland basin. The Newfoundland basin extends about 200 km to the southeast of the Grand Banks continental slope (Figure 5) (Tucholke et al., 1989). Its bathymetry ranges from 2000 to 5000 m, but averages about 4000 m (Figure 5). This basin contains between 2 and 7 km of sedimentary
Figure 5. Simplified tectonic map of basement across the eastern Grand Banks and the Newfoundland basin. The 5.5 and 6.5 km contours are basement depth. Crosshatched area shows location of interpreted igneous sills. The "L" and dashed "L" symbols represent evaporites and presumed evaporites, respectively. The 1000 m and 3000 m bathymetric contours are shown for reference. (Tucholke et al., 1989)
rocks (Tucholke et al., 1982, 1989). The Newfoundland basin is bounded to
the south by the Southeast Newfoundland Ridge, which is probably volcanic
in origin (Sullivan, 1983) (Figure 5). The Flemish Cap limits the NB to the
north.

Many workers (Tucholke et al., 1989; Austin et al., 1989; Sullivan, 1983;
Masson and Miles, 1984; Meador and Austin, 1988) have interpreted most of
the crust under the Newfoundland basin to be highly extended continental
crust with a thickness of 4 to 8 km (Tucholke et al., 1989); they located the
ocean-continent boundary (OCB) at the Fontinalis and Trutta Ridges (Figure
5). These ridges are the site of the "J-anomaly" (Pitman and Taiwani, 1972), a
magnetic anomaly which approximately parallels the M0 to M1 anomalies on
both sides of the Atlantic (Tucholke and Ludwig, 1982; Malod and Mauffret,
1990; Srivastava et al., 1990). The origin of the J-anomaly (as well as the
Fontinalis and Trutta Ridges) is still a mystery, in that it does not fit with
recognized sea-floor spreading anomalies; it probably resulted from the
emplacement of anomalously magnetized or anomalously thick oceanic crust
(Austin et al., 1990; Tucholke et al., 1989; Tucholke and Ludwig, 1982).

Several reasons have been given for locating the OCB at the J-anomaly
ridges. Tucholke et al. (1989) noted that fault blocks landward of the J-
anomaly exhibit structures seen elsewhere in extended continental crust on
the north Biscay and west Galicia margins; these same authors also pointed
out that a single unconformity, the "U" or "Avalon" unconformity, can be
correlated on seismic data from the Grand Banks across the Newfoundland
basin. In addition, seismic data indicate that the fault blocks of the
Newfoundland basin are eroded, suggesting subaerial exposure during rifting
and thus continental affinity (Tucholke et al., 1989). Austin et al. (1989)
argued that the location of the Salar basin evaporites within the
Newfoundland basin means that the NB is probably floored by continental crust, since other major evaporite basins around the Atlantic rest on continental crust (Holser et al., 1988). Magnetic data are also used to argue for a J-anomaly location for the OCB: landward of the J-anomaly ridges, the residual magnetic anomaly field is relatively smooth, but seaward of it, a zone of high-amplitude anomalies is present (Tucholke et al., 1989; Meador and Austin, 1988; Sullivan, 1983).

The choice of this site for the OCB is by no means universally agreed upon (Parson et al., 1985). Keen and de Voogd (1988) placed the OCB closer to the shelf break on the basis of landward-dipping seismic reflectors under the continental slope (Figure 6). Malod and Mauffret (1990) commented that the "smooth magnetic signature" of the Newfoundland basin may be a manifestation of peridotite or serpentinite sea-floor. In addition, Salisbury and Keen (1993) imaged listric normal faults in oceanic crust off the coast of Nova Scotia; thus, contrary to Tucholke et al. (1989), the extensional structures seen in the Newfoundland basin need not imply that its crust is continental. There are problems with this location for the OCB: not only might the landward-dipping reflectors result from something other than magmatic material associated with emplacement of oceanic crust, but the proponents of a continental-slope OCB also do not explain the eroded fault blocks of the Newfoundland basin. Thus, although the issue has yet to be clearly resolved, I favor a continental affinity for the NB crust.

**Iberian Abyssal Plain.** The Iberian Abyssal Plain lies seaward of the narrow Iberian continental shelf, and is about 350 km wide (Figure 7). Water depth ranges from 3000 to over 5000 m, crustal thickness appears to be 5 to 8 km, and sediment thicknesses are 1 to 3.5 km (Whitmarsh et al., 1990). The IAP is bounded to the north and south by topographic highs (Figure 7); as will
Figure 6. Small black rectangles along the Ocean-Continent Boundary (cross-hatched and labelled "COB") mark location where a landward-dipping reflector could be identified from released industry data. Location of the deep multichannel seismic reflection profile 85-4 is also shown. Major basins are stippled. (Keen and de Voogd, 1988)
Figure 7. Bathymetry of the Iberian Abyssal Plain. 1, 2, 3, and 4 are locations of refraction seismic lines. S1, S2, and S3 denote disposable sonobuoy lines. 398 denotes DSDP site 398. Contour interval is 200 m. J indicates the J-anomaly. Continent appears in southeast corner of map. (Whitmarsh et al., 1990)
be discussed below, these represent separate structural domains from the IAP.

As in the Newfoundland basin, the location of the OCB is uncertain. Whitmarsh et al. (1990; in press) proposed that much of the IAP is underlain by thinned continental crust. They distinguished oceanic crust from continental using seismic velocity structure, and showed a magnetic model which placed the OCB between 12°10'W and 12°30'W longitude (Figure 7). This site is somewhat landward of the J-anomaly on the Iberian margin. The IAP is underlain by half-graben fault blocks tilted landward (Whitmarsh et al., 1990); the seaward 150 km are thought to be penetrated by dikes and lava flows (Whitmarsh et al., in press).

Parallel to the continental margin in the vicinity of the OCB is a basement high composed of serpentinized peridotite. This feature was recently drilled on Leg 149 of the Ocean Drilling Program (ODP); it is the southward, subsurface continuation of a 10-kilometer-wide peridotite ridge found off the Galicia margin (Beslier et al., 1993). That ridge was originally discovered by dredge sampling (Boillot et al., 1980) and confirmed by coring during ODP leg 103 (Boillot et al., 1987, 1988, 1989a,b). Beslier et al. (1993) considered that the ultrathin continental crust of the IAP may be underlain by similar serpentinized peridotite.

The seismic stratigraphy on the IAP can be correlated with that of the Newfoundland basin. On lines GP-19 (northern IAP) and NB-4 (central Newfoundland basin), similar changes in seismic character can be seen across what has been interpreted as the breakup unconformity (Meador and Austin, 1988). The interpretation of the breakup unconformity on both deep margins also substantiates Tucholke et al.'s (1989) location of the OCB on the Canadian margin and Whitmarsh et al.'s (1990) on the Iberian margin.
The Lusitanian basin. The Lusitanian basin is completely underlain by continental crust with a thickness of about 30 km (Banda, 1988; Cordoba et al., 1988; Whitmarsh et al., 1990). It lies parallel to the Portuguese coast, and is over 250 km long and 50 to 100 km wide (Figure 8). Except for the basin's northern end, it lies onshore; it was uplifted primarily because of compression related to Pyrenean and Alpine mountain-building (Wilson et al., 1989). Sediment thicknesses vary from 0 to 4 km (Wilson et al., 1989). Although Wilson et al. (1989) found differences in structure and stratigraphy between the Lusitanian basin and the Galicia Interior basin (GIB) to the north, Murillas et al. (1990) asserted that the GIB is simply an offshore continuation of the Lusitanian basin (Figure 8).

PALEOZOIC STRUCTURAL INHERITANCE ON THE NEWFOUNDLAND AND IBERIA MARGINS

Mountain-building episodes lasting from the Ordovician to the Carboniferous left the areas around Iberia and Newfoundland with anomalously thick crust and weak lithosphere (Lallemand and Sibuet, 1986). Four orogenic episodes produced weaknesses which guided the rifting that was to tear these margins apart during the Mesozoic; the role of these Precambrian and Paleozoic episodes in Mesozoic rifting are now discussed; a more thorough account of the orogenies that occurred throughout the Paleozoic on both sides of the Atlantic is given by Lefort (1989).

At the opening of the Ordovician, the continental crust that now composes the passive margins of Iberia and Newfoundland was part of three different terranes (Figure 9). The first, Laurentia, had developed a passive margin on its east coast. To the east, across the Iapetus Ocean, was the Avalon prong (or Avalonia), a wedge of continental material that protruded from the vicinity
Figure 8. Sketch of the western Iberian marginal basins; bounded to the west from Galicia Bank to Berlenga Ridge by the "Western Banks" and to the east by the Porto-Coimbra-Tomar fault (P-C-T). Note the general continuity of the Lusitanian and Galicia Interior Basins. (Murillas et al., 1990)
Figure 9. Relative positions of Laurentia, Gondwana, Iberia, the Avalon Prong, and the Iapetus and Theic oceans during Early Silurian time (Lefort, 1983).
of northern Europe. Avalon basement today floors the eastern part of Newfoundland, most of the Grand Banks, and most of Iberia (Haworth and Lefort, 1979; Capdevila and Mougenot, 1988; Lefort, 1989). The remaining continental crust, comprising the southern parts of the Grand Banks and of Iberia, belongs to the Meguma terrane. In early Paleozoic time, this terrane was a part of the northern Gondwana margin (Lefort, 1989; Capdevila and Mougenot, 1988; Rast, 1988). The Theic Ocean separated the Avalon prong from Gondwana.

Between 700 and 570 Ma, part of the Avalon prong underwent the Avalonian (North American name), or Cadomian (European name), orogeny (Lefort, 1989). Remnants of the Avalonian can be seen in the arcuate patterns of gravity and magnetic anomalies on the Grand Banks; when the Paleozoic plate positions are restored, these anomalies are concentric with the thrusts of the Ibero-Armorian Arc (IAA) in Iberia and France. (Figure 10) (Lefort and Haworth, 1979; Haworth and Lefort, 1979; Lefort, 1989). These anomalies are the manifestation of volcanic ridges dating to the Avalonian orogeny (Lefort and Haworth, 1979). On the southeastern Grand Banks, rift basins localized on crustal weaknesses produced by the Avalonian orogeny (Lefort and Haworth, 1979; De Chassy et al., 1990). The trends of both the Jeanne d'Arc basin and the Carson basin are the same as the Avalonian ridges. In addition, seismic data reveal that the Murre fault is seated on an older planar structure which cuts to the base of the reflective crust (Tankard and Welsink, 1989).

Around the middle Ordovician (450 Ma), the Iapetus Ocean closed, and Avalonia impinged upon the east coast of Laurentia. (Please note that the time scale of Harland et al. (1990) is used throughout this thesis in figuring absolute ages of events.) This mountain-building episode, named the
Figure 10. Geophysical expression of Canadian basement ridges and European Variscan zoning arranged in their pre-Mesozoic configuration. F = France; UK = United Kingdom; GB = Galicia Banks; FC = Flemish Cap; OK = Orphan Knoll; LT = Lizard Thrust; GBN = Grand Banks of Newfoundland; P = Porcupine Bank. (Lefort et al., 1993)
Taconic (North America) or Caledonian (Europe) orogeny, led to the suturing of Avalonia to Laurentia. A change in the character of seismic reflection data underneath the Dunnage terrane suggests that the suture between Laurentia and Avalonia lies there (Keen et al., 1986, 1990).

The Avalon terrane, which lies east of the Taconic suture and is composed of Avalonian continental crust (Figure 11), shows scarce evidence of the Taconic orogeny (Williams and Hatcher, 1983). The boundary between the Gander and Avalon terranes (Figure 11) is the subvertical Dover Fault (Keen et al., 1986). The Avalon terrane probably accreted to North America towards the end of the Taconic, primarily via transcurrent movements (Williams and Hatcher, 1983). Unfaulted plutons straddling the Dover Fault indicate that it underwent no further displacement after the Devonian--363 Ma at the latest (Lefort, 1989; Williams and Hatcher, 1983). The Dover Fault appears to be the landwardmost strike-slip fault on the North American margin at this latitude; the importance of the Dover Fault in Mesozoic rifting will be discussed in a later chapter.

After Avalonia attached itself to Laurentia, the Theic Ocean was subducted in a northward direction under both terranes, as Gondwana crept northward. Gondwana finally collided with Laurentia and Avalonia to create the Acadian (North American name) or Ligerian (European) orogeny, which lasted roughly from 420 Ma to 375 Ma (Lefort, 1989). The suture between the Gondwanan Meguma terrane and Avalonia/Laurentia lies under the Collector anomaly on the Grand Banks (marked "II" in Figure 10) (Haworth and Lefort, 1979; Keen et al., 1990), and along the ophiolitic "Beja" complex in southern Portugal (which probably represents the remnants of the floor of a back-arc basin) (Figure 12) (Lefort, 1989; Dallmeyer et al., 1993; Piqué and Skehan, 1992). This boundary continued to act as a subvertical transform during the subsequent
Figure 11. Map showing terranes of Newfoundland Appalachians and location of deep seismic reflection profiles. Black regions are ophiolites emplaced on miogeoclone during Taconian orogeny. (Keen et al., 1986)
Figure 12 (continued). Legend for schematic map (see preceding page) showing features of the Acadian-Ligerian orogeny (Lefort, 1989)

1a zones with Acadian southerly tectonic transport
1b zones with probable southerly Acadian tectonic transport
2 zones with Acadian northerly tectonic transport
3a Meguma zone facies
3b possible Meguma facies
4 passive Iberian margin
5 eastern limit of Acadian tectonic effects
6 Moroccan Coastal Horst
7 mylonitic zone and orthogneisses
8 Siluro-Devonian ultramafics
9 gravity and magnetic highs
10 thrust faults
11 transcurrent faults
12 island arc volcanics
13 fault
14 post-Upper Devonian faults
15 back-arc or intra-plate volcanics
16 drill holes
17 blueschists
18 riebeckite--granites
(4: Arronches; 5: Monforte)
19 drill-core sampling
20 geophysical trends (gravity and magnetics)
21 offshore limit of Moroccan Coastal Horst

CAPITALS:
A Avalon Peninsula
AL Almaden syncline
CIZ Central Iberian zone
GAB Guadalquivir basin
GB Galicia Bank
GBNF Grand Banks of Nfld.
L Lisbon
MW Murre well
OM Ossa Morena zone
PBCZ Porto-Badajoz-Cordoba shear zone
PSB Petite Sole Bank
RI Rif
SPZ South Portuguese Zone
Hercynian orogeny (Lefort, 1989).

The Porto-Badajoz-Cordoba shear zone (PBCZ) in Iberia began sinistral movement during the Ligerian orogeny which continued through the Hercynian (Figure 12; labelled on Figure 8 as "P-C-T FAULT") (Burg et al., 1981). Gravimetry indicates that, like the Dover Fault in Newfoundland, this fault traverses the entire crust (Lefort, 1989; Dallmeyer et al., 1993). During Mesozoic rifting, the Lusitanian basin formed on the western side of the PBCZ, which was presumably reactivated as a basin-bounding fault. Moreover, no Mesozoic extension took place landward of the PBCZ.

A change in the relative motions of Gondwana and Laurentia-Avalonia—from simple north-south convergence to dextral oblique convergence—led to the Alleghanian (North America) or Hercynian (Europe) orogeny (ca. 360 to 330 Ma). Actually, the transition from "Ligerian" to "Hercynian" deformation was gradual, so a distinction between the two "events" in somewhat artificial. Hercynian shearing caused the southern tips of the north-south trending Avalonian ridges—and their accompanying gravity and magnetic anomalies—to be deflected to the west (Figure 10) (Lefort, 1989; Lefort and Haworth, 1979; Haworth and Lefort, 1979).

Inherited Hercynian structures served as transfer faults throughout the Grand Banks (northwest-southeast-trending faults of Figure 4) (Tankard and Welsink, 1989; Verhoef and Srivastava, 1989). On the southern Grand Banks, the Avalon-Meguma terrane boundary was reactivated in a left-lateral sense during Mesozoic extension (Tankard and Welsink, 1989). Keen et al. (1987) indicated that the interaction of extension direction and pre-existing structural grain may dictate the rifting style; they suggest that the difference between the southeastern Grand Banks and the Orphan basin on the northeastern Grand Banks as an example of this rule at work.
The Iberian margins basins exhibit similar Paleozoic inheritances. As on the Grand Banks, weaknesses created in the Hercynian orogeny act as transform faults in the LB (the east-west faults of Figure 8) (Wilson et al., 1989; Groupe Galice, 1979). Boillot et al. (1989a) indicated that the structures on the Galician margin also follow similar Hercynian patterns.

The Newfoundland-Iberia conjugate margins remained relatively tectonically inactive from the Permian through the middle Triassic. A lack of deposition during the Permian suggests that the entire area was above sea level (Keen et al., 1990; Rast, 1988). Rifting did occur to the north in the vicinity of the North Sea and to the east in the Tethys area during this time; this extension, however, is thought not to have affected the Newfoundland-Iberia margins.

The Triassic phase of rifting that affected all of the North Atlantic may have come about in response to buoyancy forces related to the late Paleozoic thickening of the continental crust (Tankard and Welsink, 1988; Dewey, 1988). There is no evidence, however, that the Moho was heated to above 700°C, a condition given by Sonder et al. (1987) for spontaneous orogenic collapse.

RIFTING HISTORY

The first rifting on the Newfoundland-Iberia margins dates to the Late Triassic, about 235 Ma. Rifting between Newfoundland and Iberia progressed from south to north (Murillas et al., 1990).

The Grand Banks of Newfoundland. The Jeanne d'Arc is the deepest and best studied basin on this margin. Tankard and Welsink (1987) interpreted the stratigraphy of the Jeanne d'Arc basin in terms of the extensional history. The first deposits, from the Late Triassic (235 to 210 Ma), include 2 km of
terrestrial course-grained red beds; these were laid down as the surrounding continental highs shed their sediment into the newly-formed basin. A thick mat of evaporites followed as the area began to subside slightly. This stratigraphy suggests that the first phase of crustal extension had begun, creating a depocenter next to a fault in the upper crust (the Murre fault).

For the next 50 million years (ca. 210 to 161 Ma), the basin was filled with a monotonous 7-km succession of Lower to Middle Jurassic carbonates and calcareous mudstones (Tankard and Welsink, 1987). Because these facies show little evidence of synsedimentary faulting or unconformities, a period of tectonic quiescence is postulated for this period.

Late Callovian to Aptian (161 to 124 Ma) alluvial and shallow marine sandstones and mudstones predominate the next 4 km, with numerous unconformities (Tankard et al., 1989). The coarsening of sediments as well as the higher sedimentation rate suggests renewed basin subsidence. These two developments are interpreted as resulting from an intense, second phase of rifting on the southeast Grand Banks. Extension was accommodated by listric normal faults inherited from the first rifting episode (Manspeizer, 1988). The most intense rifting occurred during the Tithonian (152-146 Ma), and extension gradually decreased through the Barremian (Tankard and Welsink, 1987). Grant et al. (1988) speculated that this second phase of rifting began slightly later, in the Kimmeridgian (ca. 154 Ma), when sedimentation rates in the Jeanne d'Arc basin almost doubled. Doming of the southern Grand Banks, called the "Avalon uplift," created the Avalon unconformity during this second phase of rifting; the unconformity surface indicates uplift and erosion from the late Kimmeridgian (153 Ma) until at least the Albian (110 Ma) (Grant et al., 1988). Meanwhile, pronounced subsidence continued in the basins (Keen et al., 1990).
After this phase, continental rifting is thought to have ended and sea-floor spreading begun. Within the basins of the Grand Banks, the end of deposition of "syn-rift" sediments and the beginning of "post-rift" sediments (the breakup unconformity) occurs at the Avalon unconformity; the temporal extent of this hiatus is much less in the basins than on the interbasin highs, allowing estimation of the time of continental breakup (Grant et al., 1988). Tankard and Welsink (1987) and Tankard et al. (1989) placed the breakup unconformity (and by inference the onset of sea-floor spreading) at the Barremian-Aptian boundary (124 Ma). A single unconformity may not serve as the "breakup unconformity," however: a set of unconformities in the Avalon sequence (Hauterivian-Barremian) of the Jeanne d'Arc basin blur the transition from syn-rift to post-rift sedimentation (Tankard and Welsink, 1987, 1988).

The remaining post-rift deposits are mainly shales and mudstones deposited in shallow-shelf seas, ranging in age from Upper Cretaceous through Quaternary (Tankard and Welsink, 1987). Slow, thermal subsidence dominated as the North America and Iberia drifted apart. Thermal subsidence downwarped the Bonavista platform westward of the Murre fault (Tankard and Welsink, 1988).

The Newfoundland basin. Tucholke et al. (1989) documented the evolution of the Newfoundland basin. Rifting began in the Salar basin (as well as in the Carson basin on the Grand Banks) during the Carnian and Norian (235 to 210 Ma); this is evidenced by the occurrence of widespread evaporites over 4 km thick. The presence of the salt basins suggests that a restricted seaway existed between Newfoundland and Iberia at this time (Austin et al., 1990). I was not able to find documentation of extension or salt deposition at this time on most of the Newfoundland basin, although this does
not mean that no extension occurred; it could have been obliterated by the late Jurassic to early Cretaceous extension that occurred on this margin. The formation of the Carson and Salar evaporite basins coincided with the first phase of rifting in the Jeanne d'Arc basin.

The second rift phase (between the Callovian and the Aptian, 161 to 124 Ma) probably caused most of the extension of the Newfoundland basin (Tucholke et al., 1989). Seismic data indicate that fault blocks were eroded and covered with syn-rift sediments in some places, and numerous unconformities were created (Tucholke et al., 1989). The thinness of the syn-rift sedimentary section and the lack of erosion of fault blocks within 50 km of the J-anomaly suggests that the latest extension was concentrated there; thus, the rift zone appears to have narrowed with time. The J-anomaly is thought to have been emplaced subaerially (Tucholke and Ludwig, 1982).

When ocean crust formation finally began, a breakup unconformity was created across the Newfoundland basin; Tucholke et al. (1989) claimed that it can be correlated with the Avalon unconformity on the Grand Banks. The observation that the breakup unconformity cuts across fault blocks and syn-rift sediments suggests that the crust probably remained at shallow water depths, elevated perhaps by the same forces that caused the Avalon uplift. As discussed above, the OCB probably lies just landward of the J-anomaly. Since this anomaly is roughly coeval with anomaly M1, the date of first sea-floor spreading would be about 125 Ma. The crust under the basin rapidly subsided after sea-floor spreading began; deep-water sediments make up the rest of the sedimentological sequence (Tucholke et al., 1989).

The Iberian Abyssal Plain. Little work has been done to date the extension of the IAP margin. Although Whitmarsh et al. (1990) attempted to pinpoint the OCB, they did not describe the extended continental crust landward of it.
Montenat et al. (1988) suggested that extension on the IAP was not initiated until the early Cretaceous (after 145 Ma). Also, Wilson et al. (1989), referring to the results of ODP Leg 103, remarked that the onset of extension on the deep Galicia margin to the north occurred during the early Valanginian (140 Ma); since the west Iberian margin rifted progressively from south to north, rifting in the IAP probably had started before 140 Ma.

The Galicia margin breakup unconformity is interpreted as late Aptian (Murillas et al., 1990)—later than the apparent breakup on the IAP—but this difference probably reflects the progressive northward opening of the margin. An unconformity found at DSDP site 398 (Figure 13) was used to place continental breakup at the Aptian-Albian boundary (112 Ma; Groupe Galice, 1979); this unconformity, however, may reflect the breakup on the nearby Galician margin to the north and not on the IAP. As in the Newfoundland basin, the location of the OCB near the J-anomaly suggests a 125 Ma date for the first sea-floor spreading (Whitmarsh et al., 1990). This date contrasts with a final rifting date of 133 Ma for the Tagus Abyssal Plain to the south (Pinheiro et al., 1992) and of 115 Ma for the deep Galicia margin to the north (Boillot et al., 1987). Murillas et al. (1990) illustrated the evolution of the processes of rifting and oceanic spreading on the western Iberian margin (Figure 14). Note that the primary extensional episode on the IAP is Valanginian in age (141 to 135 Ma).

The Lusitanian basin. Wilson et al. (1989) provided a thorough sedimentary history of the Lusitanian basin. Rifting was initiated in the late Triassic (235 Ma), as both grabens and half-grabens were filled with red fluvial sediments and thick evaporites (Leinfelder and Wilson, 1989). Lower and middle Jurassic sediments show no major variations in thickness, and rates of subsidence were low (Wilson et al., 1989). The subsidence rate
Figure 13. Site 398 location map (Leg 47), showing Galicia Bank and Iberian Abyssal Plain bathymetry. (Shipboard Scientific Party, 1979)
(1) __________ Basin-bounding transfer and transform faults
(2) __________ Ocean-continent boundary
(3) __________ Abandoned Spreading Center
(4) __________ "J" magnetic anomaly
(5) __________ Western boundary of the first oceanic spreading event in the TAP

Continental areas where the main extensional episode is:

(6) \[\square\square\square\square\] Oxfordian-Kimmeridgian
(7) \[\square\square\square\square\] Valanginian
(8) \[\square\square\square\square\square\] Hauterivian-Aptian

(9) \[\square\] High areas

Figure 14. Map (see following page) of the western Iberian margins to illustrate the evolution of the process of rifting and oceanic spreading during Late Jurassic-Early Cretaceous times. (Murillas et al., 1990)

<table>
<thead>
<tr>
<th>AB</th>
<th>Alentejo basin</th>
<th>GB</th>
<th>Galicia Bank</th>
</tr>
</thead>
<tbody>
<tr>
<td>AF</td>
<td>Aveiro Fault</td>
<td>GIB</td>
<td>Galicia Interior Basin</td>
</tr>
<tr>
<td>BAP</td>
<td>Biscay Abyssal Plain</td>
<td>GOB</td>
<td>Gorringe Bank</td>
</tr>
<tr>
<td>BT</td>
<td>Beira Trough</td>
<td>IAP</td>
<td>Iberian Abyssal Plain</td>
</tr>
<tr>
<td>ES</td>
<td>Estremadura Spur</td>
<td>PS</td>
<td>Porto Seamount</td>
</tr>
<tr>
<td>ET</td>
<td>Estremadura Trough</td>
<td>TAP</td>
<td>Tagus Abyssal Plain</td>
</tr>
<tr>
<td>FFZ</td>
<td>Figueira Fault Zone</td>
<td>VS</td>
<td>Vigo Seamount</td>
</tr>
</tbody>
</table>
Figure 14. (continued).
apparently decreased exponentially during this time (Wilson et al., 1989), suggesting thermal relaxation following the late Triassic extensional episode (Montenat et al., 1988). Murillas et al. (1990) also indicated a lack of tectonic activity during this interval.

A second phase of extension, starting in the Oxfordian (ca. 157 Ma), is evidenced by high subsidence rates in the eastern part of the basin, while the western part was uplifted along with a horst to the west (Wilson, 1975; Montenat et al., 1988; Leinfelder and Wilson, 1989); this uplift may have been related to the inception of the Avalon uplift on the Canadian margin. Red beds were again present at the beginning of this phase, and salt diapirism contributed to the formation of sub-basins (Wilson et al., 1989; Montenat et al., 1988). Oxfordian and Kimmeridgian stretching was responsible for most of the extension in the Lusitanian basin (157 to 152 Ma) (Murillas et al., 1990; Hiscott et al., 1990; Leinfelder and Wilson, 1989; Mauffret and Montadert, 1988). This period closed with the deposition of fluvial sandstones as well as shallow-marine carbonates and shales.

Extension in the Lusitanian basin ended in the earliest Cretaceous. By the Valanginian (140 Ma), subsidence had decreased dramatically, and the remaining sediments, though similar in lithology, were much thinner (Wilson et al., 1989). This decrease in subsidence—plus the occurrence of the greatest extension on the IAP during the Valanginian (Murillas et al., 1990)—suggests that extension had largely ceased in the Lusitanian basin, and that the bulk of the extensional strain was occurring in the IAP. In fact, the Lusitanian basin was almost totally emergent throughout much of the early Cretaceous (Montenat et al., 1988); the basin simply filled up with sediment. (Extension may have stopped even earlier: Montenat et al. (1988) suggested that extension ended in the late Kimmeridgian (152 Ma), and Leinfelder and
Wilson (1989) claimed that parts of the Lusitanian basin were emergent by the Tithonian (146 Ma).

Because of the thinness of the beds near the time of final rifting, a breakup unconformity is difficult to interpret. Wilson et al. (1989), nonetheless, suggested that an Aptian unconformity (between 124 and 115 Ma) is coincident with the beginning of sea-floor spreading. Shallow-marine limestones and shales dominate the post-rift sequence from late Aptian through Turonian (115 to 89 Ma) (Wilson et al., 1989).

The idea of a northward-propagating rift is substantiated by observations in the Lusitanian basin and the Galicia Interior basin to its north. Murillas et al. (1990) noted that the two basins have roughly the same tectonic and sedimentary histories, but also pointed out that the main extensional episode in the GIB occurred in the early Cretaceous (around 140 Ma), as opposed to an Oxfordian-Kimmeridgian date for the LB.

**Summary.** A common history for the rifting of these four basins can be assembled from the observations from each individual basin. Figure 15 gives an overview of the tectonic history and stratigraphy of the basins on both sides. The evidence indicates that rifting on the Iberian margin progressed from south to north; such a diachronous opening is also postulated for the Newfoundland margin (Srivastava et al., 1988). The rift structures were guided by Avalonian and Hercynian crustal weaknesses.

Rifting began in the Carnian (235 Ma), when an extensional phase took place in the Lusitanian basin and in the Grand Banks basins, and red beds and evaporites were deposited; this phase ended at about 210 Ma. The geochemistry of the halite-dominated evaporites suggests a continental source early in the rifting, later succeeded by "normal marine" sources (Holser et al., 1988). Incidentally, when the North American, Iberian and European
Figure 15. Comparison of tectonic and stratigraphic history of the Jeanne d'Arc basin, Newfoundland basin, Iberian Abyssal Plain, and Lusitanian basin. Left half of each column shows (highly generalized) stratigraphy using common lithological symbols. Heavy diagonal stripes in right half of each column indicate intense rifting; lighter stripes indicate gentler rifting. Data for Jeanne d'Arc basin from Tankard and Welsink (1987, 1989); data for Lusitanian basin from Wilson et al. (1989). Numerals indicate other sources of data: 1 = Austin et al. (1989); 2 = Groupe Galice (1979); 3 = Meador and Austin (1988); 4 = Tucholke et al. (1989).
plates are restored to their late Triassic positions, the Triassic basins of the southwest British Isles follow the same trend as the late Triassic basins on the Newfoundland-Iberia margins (Masson and Miles, 1986). A common tectonic origin is also suggested by transfer faults in the Porcupine basin south of Ireland, whose trend is similar to the transform faults on the Grand Banks (Verhoef and Srivastava, 1989).

A period of slow thermal subsidence followed on both margins. At about 160 Ma, extension was renewed on the Grand Banks and in the Lusitanian basin, and was most intense through about 150 Ma. Extension did not resume, however, in the British basins at this time (Masson and Miles, 1986). Deposition of shallow marine sediments and some carbonates occurred throughout the rifting areas. The Avalon uplift was created at this time, as well as a horst on the west side of the Lusitanian basin; these phenomena may have had a similar cause.

Extension became more concentrated in the zone of the future OCB through the early Cretaceous. Extension began in the Newfoundland basin and on the IAP by 140 Ma at the latest. During the final 20 m.y. or so of rifting, there was little additional extension in the shelf basins of both margins. At about the Barremian-Aptian boundary (124 Ma), creation of oceanic crust began, and rapid subsidence in the seaward basins occurred; a breakup unconformity (the Avalon unconformity in the basins of the Grand Banks) marked this event. Serpentinized peridotite was emplaced on the Iberian margin at the ocean-continent transition, and perhaps on the Newfoundland margin as well.

Masson and Miles (1986) provided a possible explanation for this pattern of rifting. During the late Triassic phase of rifting, Iberia moved with Africa, so that extension took place simultaneously between the eastern U.S. and Africa.
and between Newfoundland and Iberia. After 195 Ma, a transform boundary
developed between Iberia and Africa and along the southern margin of the
Grand Banks, allowing Africa and North America to continue rifting and finally
to separate (around 175 Ma) while little or no extension occurred between
Newfoundland and Iberia (Masson and Miles, 1986; Tankard and Welsink,
1988). Srivastava and Verhoef (1992) suggested that, during this period, the
southern Grand Banks high was dragged away from the rest of the Grand
Banks by Africa. After 160 Ma, extension resumed again between these
margins, and progressed to the point of continental breakup and sea-floor
spreading. Srivastava et al. (1990) indicated that Iberia was moving as a
separate plate when spreading began.

Little rift-related magmatism is evidenced on either side of the Atlantic. The
central Newfoundland basin is dotted by the mid-Cretaceous Newfoundland
Seamounts, but these are unrelated to the rifting of these margins. Also,
aside from some dike intrusion in southeastern Newfoundland during the
early Jurassic and some minor volcanics drilled in wells, rift-related igneous
activity is scarce on the Grand Banks (Keen et al., 1990; Tankard and
Welsink, 1988). Mesozoic magmatism in the Lusitanian basin is insignificant
as well (Montenat et al., 1988).

The Newfoundland and Iberia margins show gross symmetry. During both
phases of extension, rifting was distributed symmetrically through time on
both margins, first on the shelf basins, and later in the more seaward basins.
Common lithologies, and even common microfauna (Exton and Gradstein,
1984), are found on both sides of the Atlantic throughout the rifting history. In
addition, the two phases of extension were centered in the same location, so
that the resulting conjugate margins are roughly mirror images of one
another.
The conjugate margins display some mild asymmetries, however. The IAP appears to be slightly wider than the Newfoundland basin (350 km versus 200 km); this probably indicates that the site of sea-floor spreading was somewhat off-center in a broad area of highly extended crust. In addition—as discussed in detail in the next chapter—the subsidence amounts for the Iberian Abyssal Plain and the Newfoundland basin are somewhat different. These asymmetries are limited to the highly-thinned areas adjacent to the OCB, however, and are minor when the margins are considered as a whole.

One last apparent asymmetry deserves mention. The continental shelf on the Newfoundland margin is an order of magnitude wider than that on the Iberian side; at first glance, this variance seems to be a significant asymmetry. Much of the Grand Banks, however, is the unextended Bonavista platform (Srivastava and Verhoef, 1992), and the Iberian shelf has become emergent largely due to Cenozoic compression. Thus, while the basins on the Grand Banks are still wider than the Lusitanian basin, the asymmetry of the shelf widths is not as extreme as the bathymetries would suggest.
Chapter 3. Sense of Shear on the Newfoundland and Iberia Margins.

A number of investigations have explained the rifting on the Newfoundland and Iberian margins in terms of a crustal scale or even lithospheric-scale detachment zone in the style of Wernicke (1985). Tankard and Welsink (1987, 1988) interpreted a down-to-the-west detachment based on seismic profiles from the Grand Banks and Galicia margins. Winterer et al. (1988) also portrayed simple shear on these margins as down-to-the-west.

Others invoking a simple-shear model to explain these margins favored a down-to-the-east detachment. Boillot et al. (1987, 1988) speculated that the peridotite ridge observed on the Iberian margin may be mantle rock unroofed by rifting. In their interpretation, the mantle was exposed by an eastward-dipping detachment zone, after which sea-floor spreading began to the west of the peridotite ridge, and left the ridge on the Iberian margin. Hiscott et al. (1990) noted that broad doming of Iberia during the latest rifting would be easiest to explain if Iberia were on the upper plate of a crustal-scale detachment, where "theory predicts broad thermal uplift relatively late in the extensional history." Etheridge et al. (1989) distinguished the foot and hanging walls of simple-shear detachments by the extensional style exhibited on the margins. By their standards—which ascribe a narrow continental shelf to the hanging wall and a broad area of fault-controlled basins to the foot wall—Newfoundland would again be placed on the foot wall of this detachment.

As part of my investigation of these margins, I asked the question, "Can subsidence be used to indicate the sense of shear of a detachment zone
between Newfoundland and Iberia?" To answer this question, I analyzed the Total Tectonic Subsidence (TTS) across the Newfoundland basin and the Iberian Abyssal Plain, where, as I mentioned at the end of Chapter 2, subsidence values are somewhat different. TTS is simply the sediment-unloaded depth of basement below sea level (Sawyer, 1985b). It is computed by mathematically removing the weight of the sediments and allowing the basement to rebound isostatically. TTS has a value somewhere between the original basement depth and the water depth. TTS is interpreted to be the amount that the basement has subsided since rifting began (Sawyer, 1985b), and is the sum of two components of subsidence: syn-rift subsidence, resulting from the thinning of continental crust, and post-rift subsidence, which occurs as the lithosphere cools and thus gets more dense (McKenzie, 1978). TTS has been used to locate the OCB (Sawyer, 1985b), as well as to determine the amount of extension on a continental margin (Le Pichon and Sibuet, 1981; Dunbar and Sawyer, 1987, 1988, 1989b).

Subsidence patterns on both margins should also be an indicator of whether a pure- or simple-shear mechanism is called for to explain rifting on these margins. A pure-shear model would predict a maximum TTS of about 5.9 to 6.2 km (Sawyer, 1985b; Le Pichon and Sibuet, 1981; Le Pichon et al., 1982). A maximum limit exists because, in theory, there is a limit to the extent to which continental crust can thin before production of oceanic crust begins. Le Pichon and Sibuet (1981) assumed that at a syn-rift subsidence of 2.5 km—the height of the asthenosphere geoid—magma would be able to break through to the surface and initiate creation of oceanic crust. The syn-rift subsidence amount of 2.5 km would then sink to about 6 km as post-rift thermal subsidence occurred. Given Le Pichon and Sibuet's (1981) assumption, a total subsidence of greater than about 6 km could never be reached in a pure-shear setting.
Figure 1 shows the location of the transects along which I estimated TTS. The margins where the transects are located were affected primarily by the opening between Newfoundland and Iberia. Plate reconstructions for these margins (Srivastava and Tapscott, 1986; Masson and Miles, 1984; Srivastava et al., 1988, 1990b; van der Voo, 1988; Verhoef and Srivastava, 1989; Malod and Mauffret, 1990; Sibuet, 1992; Srivastava and Verhoef, 1992) indicate that these transects were roughly conjugate before the onset of rifting.

The TTS data for the Newfoundland margin (Figure 16) were digitized and calculated from maps of depth to basement and sediment thickness in the North Atlantic (Tucholke et al., 1982; Tucholke and Fry, 1985). As the seaward limit of the digitized profile, I used Tucholke et al.'s (1989) location for the OCB; as the landward limit of my profile, I chose a point on the southeast coast of Newfoundland, which lies entirely on unextended continental crust. The TTS values on the Newfoundland basin seaward of the continental slope (which occurs at about 625 km, Figure 16) remain less than 6 kilometers, slightly less than would be predicted for a passive margin rifted by a pure-shear mechanism. TTS decreases almost linearly up the continental slope, and remains at less than 1 km for most of the continental shelf (with the notable exception of the Jeanne d'Arc basin; Figure 16).

For the Iberian margin, I took a cross-section of the crust from a gravity model introduced by Whitmarsh et al. (in press). This model (Figure 17) is constrained by four seismic refraction stations (Moho depth) and continuous reflection data (basement depth), and is in good agreement with the observed gravity anomaly profile. The eastern side of the profile is coincident with the Portuguese continental shelf. I took the ocean-continent boundary to lie approximately at the location of refraction line "L3" (Figure 17) (Whitmarsh et al, 1990); this is the seawardmost location that could be chosen for the OCB.
**Figure 16.** Total Tectonic Subsidence on the Newfoundland margin; location of transect in Figure 1. Landward side of profile is near southeast coast of Newfoundland; OCB coincides approximately with J-anomaly. Ocean floor and basement depth digitized from Tucholke et al. (1982) and Tucholke and Fry (1985). Location of Jeanne d'Arc basin is also shown.
Figure 17. Gravity model for Iberian Abyssal Plain margin; location in Figure 1. East side of figure coincides with continental shelf; OCB is approximately at "L3." (Whitmarsh et al., in press)
The pre-Mesozoic basement is taken to be the top of the layers with density 2.5 g cm\(^{-3}\) and 2.55 g cm\(^{-3}\).

The TTS on the IAP transect (Figure 18) increases quickly off the continental shelf to values of 4 to 5 km, and from there grows gradually deeper. For much of the profile, TTS is greater than 6 km and even exceeds 7 km in some places, significantly more than a pure-shear model of rifting would predict. Foucher et al. (1982) speculated that similar overdeep continental basement on the Armorican margin may have resulted from extension of a pre-existing basin, but there is no direct evidence to support such an hypothesis for the IAP.

Thus, one margin (Newfoundland) features subsidence of less than 6 km, the other (Iberia) greater than 6 km; these values are, respectively, less than and greater than values predicted by a pure-shear mechanism. This difference in the TTS values pointed to a simple-shear detachment model to explain the anomalous values. Hypothetically, a model in which the lower crust is denser than the upper crust should yield different subsidence values on the conjugate margins.

My model of subsidence resulting from simple-shear lithosphere-scale extension depends on (1) the crustal thickness before rifting, (2) the lithospheric thickness before rifting, (3) the angle of the simple-shear detachment, and (4) a density function for the crust that varies with depth. The foot and hanging walls are sheared by a constant distance, and the elevation of the lithosphere is reequilibrated using local isostasy. The model (Figures 19 and 20) ignores the flexural strength of the crust and assumes that heat flow is only vertical. The crustal thickness and the detachment angle determine the width of the zone of extension, and the crustal density function determines the magnitude of the TTS difference on the foot and hanging walls.
Figure 18. Total Tectonic Subsidence on the IAP margin. Ocean floor and basement depth digitized from Figure 17.
Figure 19. Simple shear model immediately after instantaneous extension. Thin line within crust indicates separates pre-rift "upper crust" from pre-rift "lower crust." See text for details.
Figure 20. Simple shear model at thermal equilibrium. Thin line again represent original "upper/lower crust" boundary. See text for details.
For this work, a crustal thickness of 40 km, a lithospheric thickness of 125 km, and a detachment angle of 10 degrees were used. The crustal density function increases linearly from 2700 kg/m$^3$ at the surface to 3100 kg/m$^3$ at its base. The model was stopped at the point where any further extension would unroof lithospheric mantle rock.

Figure 19 shows the model immediately after instantaneous extension. Note that the hanging wall has been uplifted over the area of subcrustal thinning. In addition, the ocean basin at the upper left is deeper over the foot-wall crust than over the hanging-wall crust. The original boundary between "upper" and "lower" crust has been indicated to illustrate the bowing up of the foot wall at the rift zone.

Figure 20 shows the model at thermal equilibrium; in effect, the lithosphere has been allowed to cool for an infinite time and to return to a constant thickness of 125 km. The uplift over the hanging wall has disappeared. The ocean basin has gotten deeper everywhere, but it is still deeper above the foot wall, since the vertically-averaged foot-wall density is greater than that of the hanging wall. The crust does not undergo any further deformation, but merely moves vertically in response to isostatic changes. Also, although lateral heat flow is ignored in the rifting step, it has no effect on subsidence at great amounts of time after rifting: in Figure 20, the model has returned to thermal equilibrium, yet the subsidence difference persists, since it is the result of a difference in crustal densities, which do not vary through time.

Figure 21 plots predicted TTS for the model at thermal equilibrium. In fact, this is merely a subset of Figure 20, showing only the surface of the crust. The site of sea-floor spreading is exactly in the middle of the area of subsidence. The thin lines indicate a constant TTS gradient. The foot wall subsidence is
Figure 21. Predicted Total Tectonic Subsidence for simple shear model. This plot is actually a subset of Figure 20. Thin lines indicate a constant TTS gradient.
everywhere greater than this constant gradient, and the hanging wall subsidence is everywhere less. Furthermore, the difference is by no means undetectable: the amount of deviation from a constant gradient can be as much as 0.5 km.

Compare Figure 21 to the figures showing the observed TTS values on the Newfoundland-Iberia margins (Figures 16 and 18). The foot wall of the model may correspond to the Iberian margin (with its anomalously high subsidence) and the hanging wall to the Newfoundland margin (with its relatively low subsidence). Thus, a detachment zone between these margins would dip down-to-the-west.

The asymmetry exhibited by the subsidence values—and the use of a simple-shear detachment model—seems to contradict the "gross symmetry" discussed at the end of Chapter 2. This symmetry could not have resulted if the margin rifted in a simple-shear manner from beginning to end. It is more likely that, for most of the margins' rifting history, a pure-shear mechanism dominated, leading to gross symmetry observed in the less-thinned parts of the margins. It was only late in the second phase of rifting, when extension was concentrated in the highly-thinned crust of the Newfoundland basin and Iberian Abyssal Plain, that a simple-shear mechanism took over, resulting in differently-subsided crust in the deep-water basins. This possible solution will be discussed further in Chapter 8.

I do not intend to argue for or against the feasibility of a simple-shear detachment as the primary mechanism for continental rifting. However, if a rift indeed involves a uniform-sense simple shear of the crust, one would expect an increase of TTS values on the foot wall and a decrease on the hanging wall. The anomalous TTS values on the Newfoundland-Iberia margins could have been produced by such simple-shear rifting. Any future interpretation of the
sense of orientation of such a detachment should include an analysis of the TTS data from either side.
Chapter 4. Estimating Extension on the Conjugate Margins.

At this point, I return to the main investigation of this thesis, the impact of multiple phases of rifting on the rifting style. The dynamic rifting models described later in this thesis require as input the width of the initial model and the amount of extension across the entire model. So that realistic values are used in these models, I chose to estimate the initial width of and amount of extension that the Newfoundland and Iberian margins underwent. I used two different methods on the Newfoundland and Iberia margins.

METHODS

The First Method: Using TTS. Le Pichon and Sibuet (1981) established a relation linking TTS to $\beta$, where $\beta$ is the ratio of the pre-extension crustal thickness to current (thinned) crustal thickness. This relation is

$$Z(t) = c_3 \left( 1 - \frac{1}{\beta} \right)$$

where $Z(t)$ is Total Tectonic Subsidence and $c_3$ is a constant defined as

$$c_3 = \frac{\{\rho_a \rho_c - (\rho_c - \rho_c(\rho_c - \rho_c)) \exp(-t/\tau_c)\}}{\rho_a - \rho_w} + \frac{\{(\rho_m - \rho_c)[1 - (\alpha/2)T_a(h_c/h_l)] h_c[1 - \exp(-t/\tau_c)]\}}{\rho_a - \rho_w}$$

and
\( h_l \) = initial thickness of lithosphere,
\( h_c \) = initial thickness of crust,
\( \rho_a \) = density of asthenosphere,
\( \rho_c \) = density of crust,
\( \rho_l \) = density of lithosphere,
\( \rho_w \) = density of water,
\( \rho_m \) = density of mantle at 0°C,
\( \rho_{c0} \) = density of crust at 0°C,
\( \alpha \) = coefficient of thermal expansion,
\( T_a \) = temperature of the asthenosphere,
\( t \) = time since the instantaneous rifting event, and
\( \tau_c \) = time constant associated with lithospheric cooling.

In addition,

\[
\begin{align*}
\rho_a &= \rho_m(1 - \alpha T_a), \\
\rho_c &= \rho_{c0}(1 - (\alpha/2) T_a (h_c/h_l)), \text{ and} \\
\rho_l &= \rho_m(1 - (\alpha/2) T_a - (\alpha/2) T_a (h_c/h_l)).
\end{align*}
\]

This relation assumes instantaneous, one-dimensional extension of the lithosphere (McKenzie, 1978). For a margin as old as the southeast Grand Banks, this assumption creates a negligible error, since the margin has experienced at least 85 percent of its thermal subsidence already (Jarvis and McKenzie, 1980; Cochran, 1983; Alvarez et al., 1984). For the present margins, the value of \( c_3 \) was calculated to be 9.75, such that

\[
Z(t) = 9.75 \left( 1 - 1/\beta \right).
\]

For a discussion of the relation of TTS to \( \beta \) and the parameters used in finding the value for \( c_3 \), see the Appendix.

Once \( \beta \) is found for each point on a profile, the value of \( \beta \) for the entire profile can be found by integrating \( 1/\beta \) over the length of the profile (Dunbar
and Sawyer, 1987). The current width of a margin is considered to be the length of the profile; since the value of $\beta$ for the profile is equal to the ratio of the current width of the margin to its original width, the original width can now be calculated. The amount of extension (in kilometers) is simply the difference between the current width of the margin and its original width.

It is worth noting that the assumption of instantaneous yields a maximum value of $\beta$ for a given subsidence. During an actual rifting event (which takes tens of millions of years, and is not instantaneous), some of the thermal subsidence takes place during rifting, since the lithosphere is cooling as it stretches. When instantaneous rifting is assumed, it is assumed that no thermal subsidence takes place during rifting. The thermal subsidence that actually does occur during rifting is included as part of the "syn-rift" subsidence in an instantaneous rifting model; in this way, syn-rift subsidence is overestimated. Since syn-rift subsidence is interpreted to be linearly related to amount of extension $\beta$ (Le Pichon and Sibuet, 1981), the value of $\beta$ is consequently overestimated as well.

The Second Method: Direct Measurement of Crustal Thickness. When estimates of basement and Moho depth are constrained by seismic data, the crustal thickness along a profile can be directly computed from these two values. The crustal thicknesses of an entire profile can be used to compute the two-dimensional cross-sectional area of crust underneath the profile. This area is assumed to have been conserved during rifting. If the pre-rift crustal thickness is taken to be a uniform 40 km, the original horizontal extent (that is, the original width) of the margin can be calculated from the conserved area; in effect, the profile is restored to its original width. The amount of extension "is
again the difference between the current width of the margin and its original width.

APPLICATION OF METHODS TO NEWFOUNDLAND AND IBERIA MARGINS

The Newfoundland Margin. The TTS transect discussed in Chapter 3 was used as input for the first method to estimate crustal thickness. The cross-sectional area was then restored to 40 km thickness to yield the original width of the margin. Again, Tucholke et al.'s (1989) location for the OCB was used as the seaward limit of the profile, and the same point in southeast Newfoundland as the landward limit. In this case, assuming a pre-rift crustal thickness of 40 km suggests that the pre-rift width of the crust was about 500 km. Since the current width of the margin—the distance from the point in the unextended crust to the OCB—is roughly 670 km, that means that the margin has been stretched by 170 km (Table 1). Since the choice of OCB in this calculation is a seaward extreme, the width will also be a maximum; hence, the amount of extension calculated is a maximum value.

For the second method, I employed the crustal thickness estimates of Reid and Keen (1990) and Tucholke et al. (1989) as input. The crustal thickness profile included a 360-km zone of 38-km-thick crust (the Grand Banks), a 150-km transitional zone whose thickness varied linearly from 38 km to 8 km (the continental slope), and a 160-km zone of 8-km-thick crust (the Newfoundland basin). Restoring this crustal volume to 40 km thickness again yielded an original margin width of 500 km; since the current width of the margin is 670 km—the same as used in the first method—the amount of extension is again computed to be 170 km (Table 1).

The Iberian margin. Using the TTS method for the Iberian margin suggests that the pre-rift width of the crust was about 120 km (Table 1). Since the
<table>
<thead>
<tr>
<th></th>
<th>Total Length of Profile</th>
<th>Method 1 Original Width of Margin</th>
<th>Amount of Extension</th>
<th>Method 2 Original Width of Margin</th>
<th>Amount of Extension</th>
</tr>
</thead>
<tbody>
<tr>
<td>IBERIAN MARGIN</td>
<td>280 km</td>
<td>120 km</td>
<td>160 km</td>
<td>70 km</td>
<td>210 km</td>
</tr>
<tr>
<td>NEWFOUNDLAND MARGIN</td>
<td>670 km</td>
<td>500 km</td>
<td>170 km</td>
<td>500 km</td>
<td>170 km</td>
</tr>
<tr>
<td>BOTH MARGINS</td>
<td>950 km</td>
<td>620 km</td>
<td>330 km</td>
<td>570 km</td>
<td>380 km</td>
</tr>
</tbody>
</table>

**Table 1.** Estimated amount of extension across the Newfoundland and Iberia margin transects. The "Method 2" values were based on better estimates of current crustal thickness, and were thus the values used in subsequent models. See text for a discussion on the methods used.
current width of the margin is roughly 280 km, that means that the margin has been stretched by an additional 160 km. As in the Newfoundland basin, the choice of OCB in this calculation is a seaward extreme; hence, the width and the amount of extension calculated will be maximums.

For the second method, crustal thickness was taken from the gravity model introduced by Whitmarsh et al. (in press) (Figure 17). The crust-mantle boundary was taken to be the boundary between rocks with densities greater than 3.0 g cm$^{-3}$ and those with densities less than or equal to 3.0 g cm$^{-3}$; the landward and seaward limits of extended continental crust and the basement depth were again identical to the ones used in TTS computation. Using a pre-rift crustal thickness of 40 km yielded values of 70 km for the margin's original width and 210 km for the amount of extension (Table 1). As on the Newfoundland margin, the choice of Whitmarsh et al.'s (1990) OCB is a seaward extreme, and thus the computed value of extension will be a maximum for this margin.

The amount of extension estimated by the first method for the Iberian margin is substantially less than the value arrived at by the second method. The values calculated in the second method are probably more accurate, however: that method uses crustal thicknesses constrained by seismic data. In contrast, in the first method, not only does the estimation of extension from TTS involve the assumption of many parameters, but the TTS on this margin is anomalously high. Hence, the amount of extension estimated by the second method is used in subsequent models.

SUMMARY

The total extension preceding the initiation of sea-floor spreading is obtained by adding the extension amounts on the two margins (Table 1). The
Iberian profile does not extend up onto the continental shelf and into the Lusitanian basin; given the relatively thin sequence of syn-rift sediments (Wilson et al., 1989) and the narrowness of the basin, it probably did not account for more than 25 km of additional extension across these margins. The omission of the Lusitanian basin extension is mitigated in importance when it is considered that the estimates given in Table 1 are already maximum values. It is important to note that the amounts of extension on both sides are roughly the same; in this sense, the rifting of the Newfoundland and Iberia margins was grossly symmetrical.
Chapter 5. The Dynamic Modelling Method.

To model the extension of continental lithosphere, I used the finite element method (FEM) originally employed by Dunbar and Sawyer (1988, 1989a) and also by Harry and Sawyer (1992a,b). (A more complete treatment of the modelling method can be found in Dunbar and Sawyer (1989a) and in Harry and Sawyer (1992a).) With the FEM, a vertical slice of the lithosphere is represented as a two-dimensional thermo-mechanical continuum. The continuum is approximated as a "mesh" of rheological elements.

Each quadrilateral element has eight nodes, one on each side and one on each corner. Thus, a side-node is shared by two elements, and a corner-node by four. A time-stepping algorithm determines the nodal velocities and steps the nodal positions forward to the next time step. Then, the algorithm computes the temperature change of each node over the last time interval using a finite-element, transient heat conduction algorithm. The FEM used here ignores adiabatic changes in temperature, since these changes are smaller in magnitude than the uncertainty of the initial geotherm (Dunbar and Sawyer, 1989a).

The deformation of each element is governed by two different empirically-determined rheological laws. The yield stress of each element is computed using Byerlee's law for frictional sliding and also using a power-law viscous creep. The value for yield stress actually used for each element at each time step will be the lower of the values predicted by these two laws, since deformation is likely to occur by the mechanism that requires the least energy.
The temperature, pressure, strain rate, and lithology all influence the deformation mechanism for which the yield stress with be least.

Brittle faulting in the upper crust is approximated by ideal plasticity, and no deformation by discrete faulting occurs in the model. Thus, while the formation of individual fault-controlled basins is not computed, the overall extension in the upper crust is estimated. The FEM method does not include the thermal or flexural effects of sedimentation. In addition, neither partial melting nor any other magmatic process is approximated by the FEM. I will discuss the melt generation of particular extensional models in a later section.

**Boundary Conditions.** The "driving forces" of plate tectonics, and continental lithospheric extension in particular, are approximated by constant velocity boundary conditions on the sides of the model; in other words, we assume a "passive" rifting mechanism, in which extension is driven by far-field tensile stresses on the continental lithosphere, and not by the separating force of upwelling asthenosphere (Keen, 1985; Kusznir and Ziegler, 1992; Ziegler, 1992). When a model is symmetrical, it is halved, with the center (which becomes a side) held fixed, and the other side moved at half the total extension rate of the full model. This modification significantly reduces the computing time for each model.

Buoyancy forces arising from lateral density differences contribute to the deformation of these models. In addition, the model is supported from within by its own flexural strength, and by a buoyancy force across its base. This latter force, which is the difference between the weight of the unextended model lithosphere and an equivalent thickness of zero-age oceanic lithosphere, stays constant through time in each model. If the top of the model is below sea level, the weight of overlying sea water is also included.
A flexural correction can be applied after each time step. The correction is approximated as the response of a thin elastic plate whose thickness is given by the depth to the 550°C isotherm, and whose load is given by the difference between the weight of a vertical slice of the model (at the selected time) and its weight prior to extension. This modifies the topography, but is not included in the model run since the correction does not affect the model's gross structural features.

The FEM includes heat generation in the crust; the amount of heat generated per unit volume decreases exponentially with depth. The top of the model is kept at 0°C throughout the model run. At the bottom of the model lithosphere, a constant heat flow is maintained. I have chosen this boundary condition since I am interested in modelling the effect of a resting phase on the pattern of extension; the conductive cooling of extended lithosphere during this phase would occur too slowly, if at all, if a constant temperature bottom boundary condition were chosen.

In the modelling routine used here, two thermal boundary conditions for the base of the model were available: one which imposed a constant temperature, the other which maintained a constant heat flow into the model. Most investigators have chosen constant-temperature bottom boundary conditions (Beaumont et al., 1982; Braun and Beaumont, 1987; Issler et al., 1989; Dunbar & Sawyer, 1989a; Chéry et al., 1990; Harry and Sawyer, 1992a,b). Such a boundary condition assumes that the underlying asthenosphere is inviscid and well-mixed, and thus maintains a constant temperature. Keen and Boutilier (1990), for purposes of simplification, did not include any thermal effects in their models. Like Braun and Beaumont (1989a,b) and Bassi (1991), however, I chose a constant heat flow as the bottom boundary condition in my models. I was interested in modelling the thermal effects of a "resting phase;" the choice
of a constant-temperature boundary condition would have maintained the base of the model at an artificially-high temperature, and would have precluded observation of the manner in which the lithosphere cools during a resting phase.

Recently, Bassi et al. (1993) argued that the constant heat flow boundary condition allowed the lithosphere to cool too fast, and chose to implement a more complicated boundary condition. In their scheme, a constant temperature is imposed at a given depth, but the depth of this boundary condition is far below the base of the mechanical model. This scheme allows for upwelling asthenospheric material to supply heat from below (overcoming the deficiency of the uniform heat flow boundary condition), but prevents the unrealistically-high temperatures of the constant-temperature boundary condition.

While Bassi et al.'s (1993) method does indeed better approximate the lithospheric temperature regime, modifying the FEM code to accommodate this scheme is outside the scope of this thesis. Hence, my use of the overly-cool constant heat flow boundary condition may overestimate the strength of the lithosphere and underestimate the amount of magma generated (see Chapter 8). Nevertheless, simple one-dimensional calculations I have done suggest that the temperature difference between Bassi et al.'s (1993) regime and a constant heat flow regime are minor in all but the deepest parts of the models, and that difference is smaller than the uncertainty in the initial geotherm. Thus, the error resulting from the selection of a constant heat flow bottom thermal boundary condition is negligible.

In the models run here, the bottom boundary condition is determined in the following manner. The unextended initial model is allowed to equilibrate thermally with a temperature of 0°C at the top and 1333°C at the bottom of the model. The vertical temperature gradient across the bottom row of elements is
multiplied by the mantle conductivity to arrive at the vertical heat flow, which varies horizontally across the model. These heat flow values are used as the bottom boundary condition for the rest of the model run.

Rheologies. In these following models, the lithosphere is represented by three different rheologies, described in Table 2. The mantle is approximated by the Wet Aheim Dunite of Chopra and Paterson (1981). The crust is characterized by two different rheologies: Wet Quartz Diorite (Hansen and Carter, 1982) represents typical continental crust, and Wet Granite (Hansen and Carter, 1983) is used to model weakened continental crust. The thermal properties listed in Table 3 are also used in all models.
Table 2. Rheological parameters used in the models presented here.

<table>
<thead>
<tr>
<th>Material</th>
<th>A (Pa⁻ⁿ s⁻¹)</th>
<th>Qc (kJ mole⁻¹)</th>
<th>n</th>
<th>ρ  (kg m⁻³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wet Ahein Dunite</td>
<td>4 x 10⁻²⁵</td>
<td>498</td>
<td>4.5</td>
<td>3300</td>
</tr>
<tr>
<td>Wet Quartz Diorite</td>
<td>5 x 10⁻¹⁸</td>
<td>219</td>
<td>2.4</td>
<td>2850</td>
</tr>
<tr>
<td>Wet Granite</td>
<td>8 x 10⁻¹⁶</td>
<td>137</td>
<td>1.9</td>
<td>2670</td>
</tr>
</tbody>
</table>

Table 3. Thermal parameters used in the models presented here.

- **Crustal Heat Production**
  - Surface Heat Production: 3 x 10⁻⁶ W m⁻³
  - Exponential Decay Depth: 10 km

- **Conductivity**
  - Crust: 2.5 W m⁻¹ °K⁻¹
  - Mantle: 3.4 W m⁻¹ °K⁻¹

- **Specific Heat**
  - Crust: 875 J kg⁻¹ °K⁻¹
  - Mantle: 1250 J kg⁻¹ °K⁻¹

- **Coefficient of Thermal Expansion**: 3.1 x 10⁻⁵ °K⁻¹

- **Asthenosphere Density**: 3189 kg m⁻³

DESCRIPTION OF THE MODELS

The intent of this lithospheric modelling experiment was to test the response of initial models with different lithospheric weaknesses to extension via different multi-phase extension "paths."

Each initial model is 800 km across, and undergoes 500 km of extension over the lifetime of the model run. These values were preliminary estimates for the Newfoundland and Iberia margins, which were superseded by the values calculated in Chapter 4 for the "specific" model (discussed in Chapter 7, below). These preliminary values have similar amounts of extension, and hence similar strain rates, to the ones used in the specific model. Each model is run for 120 m.y., which is approximately the length of time from the first rifting (about 230 Ma) to sea-floor spreading (about 110 Ma) on the Newfoundland-Iberia conjugate margins. By choosing these values, I did not intend any of these models to approximate the rifting of these margins; I simply wanted to use extension amounts and rates that are similar in scale to the ones observed on the Newfoundland and Iberian margins. In addition, by choosing symmetrical models, I could concentrate on the effects of multiple phases of rifting on rifting style without the complicating factor of asymmetry.

Rifting "paths." While all models have the same total amount of extension over the same total duration of time, they have different instantaneous extension rates during that interval. Over its 120-m.y. span, each model has one rifting rate for the first 25 m.y., a different rate for the next 45 m.y., and yet another rate for the final 50 m.y.. These durations were chosen since they
approximate the durations of the first (late Triassic) rifting phase, the resting phase, and the second (late Jurassic-early Cretaceous) rifting phase on the Newfoundland-Iberia margins. Each sequence of extension rates for the three phases is called a rifting "path."

The rifting paths (1 through 5) used in this set of models are described in Table 4, and shown graphically in Figure 22. The extension rate (and hence the amount of extension) for the middle phase is zero in all models (except those following the constant-rate path 4); this represents the period in which there is thought to have been no divergent motion between North America and Iberia (about 215 to 160 Ma). The paths are intended to simulate different distributions of the total extension between the first and second rifting phases. Path 4 simply describes a constant extension rate for the entire model duration; models following path 4 were not divided into three phases like the others.

Initial Models. Three different initial models are tested, each of which contains a different type of lithospheric weakness. The weakness types are taken from Dunbar and Sawyer (1989a). Because all the models are symmetric, they were halved along their axes of symmetry and stretched by half the extension rates listed in Table 4.

The first model, "MW," contains a mantle weakness, formed by a section of crust which is 5 km thicker than the flanking "normal-thickness" crust (Figure 23). The crust in this model is represented entirely by a quartz diorite rheology. In the models considered here, the mantle and lower crust lie mainly in the ductile field, where yield strength varies inversely with temperature. At identical temperatures, the quartz diorite rheology is much weaker than the dunite rheology. Hence, where a crustal welt replaces mantle dunite with crustal quartz diorite in a model, the vertically-integrated strength of a column of lithosphere is lowered.
<table>
<thead>
<tr>
<th>PATH</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Phase 1: 0-25 Ma</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Extension rate (km Ma(^{-1}))</td>
<td>0</td>
<td>2</td>
<td>6</td>
<td>4.17</td>
<td>10</td>
</tr>
<tr>
<td>Percent of total extension during this phase</td>
<td>0</td>
<td>10</td>
<td>30</td>
<td>20.8</td>
<td>50</td>
</tr>
<tr>
<td>Amount of extension during this phase (km)</td>
<td>0</td>
<td>50</td>
<td>150</td>
<td>104</td>
<td>250</td>
</tr>
<tr>
<td><strong>Phase 2: 25-70 Ma</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Extension rate (km Ma(^{-1}))</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>4.17</td>
<td>0</td>
</tr>
<tr>
<td>Percent of total extension during this phase</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>37.5</td>
<td>0</td>
</tr>
<tr>
<td>Amount of extension during this phase (km)</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>188</td>
<td>0</td>
</tr>
<tr>
<td><strong>Phase 3: 70-120 Ma</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Extension rate (km Ma(^{-1}))</td>
<td>10</td>
<td>9</td>
<td>7</td>
<td>4.17</td>
<td>5</td>
</tr>
<tr>
<td>Percent of total extension during this phase</td>
<td>100</td>
<td>90</td>
<td>70</td>
<td>41.7</td>
<td>50</td>
</tr>
<tr>
<td>Amount of extension during this phase (km)</td>
<td>500</td>
<td>450</td>
<td>350</td>
<td>208</td>
<td>250</td>
</tr>
</tbody>
</table>

Table 4. Description of the rifting paths used in the models that follow.
Figure 22. Rifting paths (1 through 5) used in generic set of models. Total extension is 500 km in all generic models. Slope of each line indicates rate of extension.
Figure 23. Initial model "MW" (mantle weakness only). See text for details.
The second model, "CW," contains a crustal weakness (Figure 24). Here, a section of quartz diorite crust is replaced with the wet granite rheology. The thickness of this weakness ranges from 10 to 20 km.

The third model, "BS," contains both a mantle and a crustal weakness, symmetrically superimposed on one another (Figure 25). (The abbreviation "BS" stands for "Both Symmetric.") The sizes of the weaknesses are the same as in the individual MW and CW models, although in BS, the crustal weakness makes up 25 to 50 percent of the crust, independently of the crustal thickness.

Executing the three different starting models over five different extension paths produced 15 models in total. This is the "generic model suite."

RESULTS OF THE GENERIC MODEL SUITE

The mechanical meshes of each of the fifteen models are shown in Figures 26 through 40. For models following path 1, I show the start, middle, and end of the only rifting phase for that path (the 70 to 120 m.y. phase). For path 4, the steps shown are roughly equally spaced (in the time domain) from beginning to end. For paths 2, 3, and 5, I show the beginning, middle and end of both the first and the second rifting phases. Note that, for these paths, the mesh at the end of the first rifting phase (25 m.y.) is roughly the same as the mesh at the beginning of the second (70 m.y.), since between these two times, no extension is applied to the models. Keep in mind that these models represent half of a symmetrical rifting margin.

I will describe the individual models in the following paragraphs, then summarize the conclusions from these generic models. For clarity, after each model name, I have listed the percentages of total extension for each phase of that path; for example, MW2(10,0,90) indicates that 10 percent of the extension
Figure 24. Initial model "CW" (crustal weakness only). See text for details.
Figure 25. Initial model "BS" (both crustal and mantle weaknesses). See text for details.
Figure 26. Deformation of mechanical mesh for model MW1(0,0,100). See text for details.
Figure 27. Deformation of mechanical mesh for model MW4(constant). See text for details.
Figure 27 (continued).
Figure 28. Deformation of mechanical mesh for model MW2(10,0,90). See text for details.
Figure 28 (continued).
Figure 29. Deformation of mechanical mesh for model MW3(30,0,70). See text for details.
Figure 29 (continued).
Figure 30. Deformation of mechanical mesh for model MW5(50,0,50). See text for details.
Figure 30 (continued).
Figure 31. Deformation of mechanical mesh for model CW1(0,0,100). See text for details.
Figure 32. Deformation of mechanical mesh for model CW4(constant). See text for details.
Figure 32 (continued).
Figure 33. Deformation of mechanical mesh for model CW2(10,0,90). See text for details.
Figure 34. Deformation of mechanical mesh for model CW3(30,0,70). See text for details.
Figure 34 (continued).
Figure 35. Deformation of mechanical mesh for model CW5(50,0,50). See text for details.
Figure 35 (continued).
Figure 36. Deformation of mechanical mesh for model BS1(0,0,100). See text for details.
**Figure 37.** Deformation of mechanical mesh for model BS4(constant). See text for details.
Figure 37 (continued).
Figure 38  Deformation of mechanical mesh for model BS2(10,0,90). See text for details.
Figure 38 (continued).
Figure 39. Deformation of mechanical mesh for model BS3(30,0,70). See text for details.
Figure 39 (continued).
Figure 40. Deformation of mechanical mesh for model BS5(50,0,50). See text for details.
Figure 40 (continued).
occurred in the first phase (0-25 m.y.), 0 percent in the second, "resting" phase (25-70 m.y.), and 90 percent in the third phase (70-120 m.y.). Model 4 has a constant rifting rate, so it is denoted, for example, "MW4(constant)."

MANTLE WEAKNESS MODELS

MW1(0.0.100). Necking begins in the center of this model, but the locus of necking moves closer to the edges of the model as time progresses (Figure 26).

MW4(constant). When rifting proceeds even slower than in model MW1, the lithosphere is necked even less in the center of the model, and extension is more evenly distributed across the original mantle weakness (Figure 27). This mesh looks very similar to the "runaway thinning" model presented by Bassi et al. (1993).

MW2(10.0.90). When stretching in the first rifting phase is even 10 percent of the total extension, the second phase of rifting shifts to the outer flank of original wide weakness (Figure 28). The center of the model stretches a bit more in the second phase, but by the end of the second phase, necking on the flanks of the original mantle weakness has taken over.

As can be seen on the gray-scale plots of strain rate in Figure 41, the strain rate diminishes slowly in the center of the model, and concentrates at the flank of the original crustal welt; the transition from broad rifting to concentrated rifting at the flanks occurs between 95 and 120 m.y., relatively late in the phase. Note also that the extension in the crust is more diffuse: at 120 m.y., for example, the crust on the left edge of the plot has a higher strain rate than the mantle below it. This difference in strain rates is accommodated by shearing in the lower crust.
Figure 41. Gray-scale strain rate plot for model MW2(10,0,90). Grays range from $\log_{10}$ (strain rate) of -14.2 (black) to -11.6 (white). Light background delineates maximum width of model and original thickness of model.
**MW3(30.0.70).** Again, when stretching resumes in the second rifting phase, it concentrates at the flanks of the original crustal welt (Figure 29).

**MW5(50.0.50).** Stretching the mesh 50 percent in the first rift phase shows clearly the "shift of necking" phenomenon discussed in the previous four models (Figure 30). Strain in the second rift phase localizes immediately on the original welt flank, and concentrates even more there with time. Again, the gray-scale plot of strain rate (Figure 42) shows the concentration of strain at the location of the hottest Moho. Compare these plots to the those in Figure 41; the localization of rifting at the edge of the welt occurs earlier than in MW2(10,0.90). Again, note the differential crust and mantle strain rates, and the lower-crust shearing.

**CRUSTAL WEAKNESS MODELS**

**CW1(0.0.100).** This mesh necks in the center, and appears at the point of failure there, too (Figure 31). Compare this to the style of rifting in MW1, where the location of necking moves outward from the center.

**CW4(constant).** Strain rate diminishes slowly in the center of this model, and the location of highest strain rate moves slowly outward (Figures 32 and 43). In fact, the locus of extension remains in a position just to the left of the locus of greatest previous necking. At this rate of rifting, narrow necking does not occur; thus, a crustal weakness does not predispose the lithosphere to narrow rifting; compare this model to CW1(0,0,100) (Figure 31).

**CW2(10.0.90) and CW3(30.0.70).** A mantle strength forms under the crustal weakness after the first phase of rifting thins the lithosphere at this location (Figures 33 and 34). Again, necking in the second rifting phases localizes on the flank of the necked region produced in the first phase.
Figure 42. Gray-scale strain rate plot for model MW5(50,0,50). Grays range from $\log_{10}(\text{strain rate})$ of -14.9 (black) to -12.0 (white). Light background delineates maximum width of model and original thickness of model.
Figure 43. Gray-scale strain rate plot for model CW4(constant). Grays range from $\log_{10}(\text{strain rate})$ of -14.8 (black) to -11.9 (white). Light background delineates maximum width of model and original thickness of model.
**CW5(50,0,50).** This model begins to neck in the center, just as
CW1(0,0,100) did, since the most extension is assigned to the first phase in
this path (Figure 35). At 25 m.y., strain is concentrated entirely in middle of the
model. Figure 44 shows that the strain rate is extremely low in the center of
the model throughout the second phase of rifting; the area of highest strain rate
always stays about halfway between the model's center and its outer edge.
The strain rate is even fairly high at the outer edge of the model. As Figure 44
indicates, the necking shape produce in the first rifting phase has been
practically frozen into the model during the rifting phase.

**COMBINED WEAKNESS MODELS**

**BS1(0,0,100).** The location of necking moves from the center outward, but
thinning of the lithosphere increases as the distance from the model's center
increases (Figure 36). The crust in this mesh also has the appearance of
"runaway thinning" (Bassi et al., 1993). Crustal extension remains in the center
of the model for much of the time, while mantle extension moves outward
sooner; this produces a zone of shear in the lower crust.

**BS4(constant).** This mesh appears similar to BS1(0,0,100), except that
both crustal and mantle extension is more evenly distributed throughout the
model (Figure 37); this difference is probably due to the lower extension rate.
In the strain rate plot in Figure 45, the area of highest mantle strain rate moves
from center outwards; the strain rate decreases more slowly in the crust at the
center than in the mantle. Extension is always more diffuse in crust, and again
produces pervasive shear just above the Moho. (This lower-crust shear
provides a "reasonable" mechanism for decoupling of crust and mantle
stretching, which Rowley and Sahagian (1986) claimed was lacking in previous
non-uniform stretching models.) It is clear that, from 70 m.y. onwards, necking
Figure 44. Gray-scale strain rate plot for model CW5(50,0,50). Grays range from $\log_{10}$(strain rate) of -14.9 (black) to -12.4 (white). Light background delineates maximum width of model and original thickness of model.
Figure 45. Gray-scale strain rate plot for model BS4(constant). Grays range from $\log_{10}$ (strain rate) of -14.9 (black) to -11.4 (white). Light background delineates maximum width of model and original thickness of model.
Figure 45 (continued).
occurs in a location slightly outward from the previously-necked area (Figure 45); this mechanism allows the locus of necking to migrate outward.

**BS2(10.0.90).** As previous "path 2" models, necking in the second riftting phase in concentrated on the outer flank of the original crustal welt (Figure 38). There is some strain in the center of the model early in the second phase, but that quickly diminishes in favor of riftting on the flanks (Figure 46). Interestingly, by 120 m.y., the highest strain rate occurs from the middle of the area of greatest thinning outward; meanwhile, from that point inward, strain rate is at its minimum. Moho shear occurs throughout the second riftting phase, and crustal extension is again more diffuse.

**BS3(30.0.70).** It is evident from the mesh in BS3 that, in the first phase, the locus of necking has begun moving outward when the first phase comes to an end (Figure 39). Thus, when the second riftting phase begins, and riftting localizes at the flanks of the original weakness, a remnant of the shape of the first episode of riftting is left at the center of the model.

**BS5(50.0.50).** During the first riftting phase, the location of necking has begun to migrate outward by 25 m.y. (Figures 40 and 47). Then, in the second rift phase, strain begins immediately on the flanks of the original mantle weakness (Figure 47), and stays here throughout the model run. Even more so than in BS3(30,0,70), the shape of the first-phase necking is preserved in the base of the lithosphere. Again, Moho shear occurs because of differential distribution of extension in the crust and in the mantle.

The following conclusions can be drawn from the foregoing models:

1. At any point in time, the upper mantle extension will localize where the Moho is hottest. For the power-law creep rheology, strength is inversely related to temperature; cooling the mantle only a little will greatly strengthen it.
Figure 46. Gray-scale strain rate plot for model BS2(10,0,90). Grays range from \( \log_{10}(\text{strain rate}) \) of -14.9 (black) to -11.0 (white). Light background delineates maximum width of model and original thickness of model.
Figure 47. Gray-scale strain rate plot for model BS5(50,0,50). Grays range from $\log_{10}$(strain rate) of -14.9 (black) to -11.6 (white). Light background delineates maximum width of model and original thickness of model.
In the models presented where mantle extension migrates laterally, cooling is sufficiently fast relative to the rate of extension that the upper mantle becomes stronger than the neighboring hotter (and weaker) mantle. Because the upper mantle is usually the strongest portion of a vertical slice of lithosphere (Figure N), when the upper mantle strength decreases, the strength of the entire lithosphere decreases. Indeed, the mantle weakness becomes a mantle strength once it cools sufficiently, preventing further significant extension at that location. This increase in strength from cooling is the same phenomenon presented by England (1983) and discussed most recently by Bassi et al. (1993). The upper mantle may become shallow enough to enter the brittle-failure regime (Sawyer, 1985a), but it still remains stronger than neighboring hotter mantle. It is this strengthening of the zone of initial rifting that cause the locus of extension to move laterally outward from the center of the model.

This behavior is exhibited by many of the generic models, where the locus of extension migrates laterally. In models MW1(0,0,100) and MW4(constant), the rate of extension is slow enough to allow the mantle in the center to cool and strengthen and cause a shift in necking location (Figures 26 and 27). Since the extension rate in MW4(constant) is less than in MW1(0,0,100), the lithosphere in the center cools off before it can thin as much as that in MW1. Similarly, in models MW2(10,0,90), MW3(30,0,70), and BS2(10,0,90), when stretching resumes in the second rifting phase, it concentrates at the flanks of the original crustal welt, where the Moho is deepest and hottest (Figures 28, 29, 38, 41 and 46). At the beginning of the second rifting phase of model CW5(50,0,50), the upper mantle is extremely cold and strong at the center, and the necking shape created in the first phase is practically frozen in (Figures 35 and 44).
2. The location of crustal weaknesses will be the initial location of crustal extension. Once the mantle lithosphere becomes sufficiently thinned at a different location, however, crustal thinning will proceed there. Early during extension, the crustal weakness (with a granite lithology) is partly in the ductile-deformation regime, and is thus weaker than the surrounding quartz diorite crust. As the crustal weakness is necked and cools, it lies entirely in the brittle-deformation field. Since a brittle rheology equalizes all lithologies in terms of strength, the granite crustal weakness is no longer weaker than the neighboring "normal" quartz diorite crust. With the crust equally strong everywhere, crustal deformation is controlled by the location of upper mantle strain, and hence the entire lithosphere will begin to strain at the same location. Models BS4(constant) and BS2(10,0,90) exhibit such a migration of crustal extension (Figures 37, 38, 45, and 46). This is a similar result to that reached by Harry and Sawyer (1992a).

3. A crustal weakness requires a lower rate of extension before it exhibits a lateral shift in necking. This is seen by comparing models MW1(0,0,100) and CW1(0,0,100), both of which are extended at 10 km Ma\(^{-1}\) (Figures 26 and 31). Because the crustal weakness in CW1(0,0,100) does not involve lateral variations in the Moho, the initial Moho temperature is relatively constant. This prevents one part of the Moho from becoming sufficiently hotter at the flanks of the original necking to overcome the weakness presented by the crustal granite rheology. Thus, extension continues to localize in the center. In contrast, at the slower rate of 4.17 km Ma\(^{-1}\), both MW4(constant) and CW4(constant) exhibit a later shift in locus of extension.

4. The occurrence of a resting phase between two episodes of rifting greatly affects the morphology of a continental rift. In most cases, the site of the original rift will not be favored for extension when stretching begins anew.
If extension in the first phase is significant, the crust will be thinned and the Moho elevated. During the resting phase, the upper mantle will cool and strengthen. When rifting begins anew in the second phase, the original site of rifting will be a lithospheric strong zone, and second-phase rifting will occur in a different, weaker location. Furthermore, a resting phase (or other variation through time in the rate of rifting) could serve as a possible explanation for areas where the location of rifting is observed to migrate through time.
Chapter 7. The Specific Model.

DESCRIPTION AND PURPOSE OF THE SPECIFIC MODEL

Based on the conclusions gleaned from the generic models, I attempted to design a model that more closely approximated the rifting history of the Newfoundland-Iberia conjugate margins—the "specific" model. The generic model suite results suggested that, if first-phase extension had been significant, second-phase extension would have occurred at a different location. As mentioned in Chapter 2, the second rift phase on these margins occurred in roughly the same place as the first. Hence, first-phase extension must have been sufficiently minor not to create a strong zone in the lithosphere and shift second-phase rifting elsewhere.

I set out to approximate only the gross lithospheric features of the Newfoundland and Iberia margins. Primarily, I was interested in imitating the crustal thickness profile across the two margins. On the Canadian side, the crust is 35 to 38 km thick across the Grand Banks, and thins gradually seaward of the Jeanne d'Arc basin to an ultrathin 4 to 8 km beneath the Newfoundland basin. On the Iberian side (as reflected in Figure 17), the crust on the continent is about 30 km, and thins gradually to 5 to 8 km over a broad area. As well as the crustal profile, the distribution of upper-crust extension and subsidence would be targets for the specific model.

Several estimates were used to constrain all attempts at a specific model. The original width of the model and the amount of extension it underwent were based on figures given in Table 1; for simplicity, I rounded the original width to 600 km and the amount of extension to 400 km. Of that 400 km, I assigned 60
km to the first rifting phase and 340 km to the second. This choice was based on the value of $\beta = 1.1$ that Hiscott et al. (1990) cited for the Triassic sequences of the Bristol Channel area and in the Wessex basin in the British Isles. While I am uncertain of the method by which this figure was estimated, and even though it applies to the British basins and not the ones examined in this thesis, it is the best estimate available. In addition, paleomagnetic data cannot resolve the amount of extension between Iberia and North America (J. E. T. Channell, personal communication). The late Triassic syn-rift sediments of the Newfoundland and Iberia shelf basins are buried so deeply, and salt tectonics have deformed the sediments so extensively, that reliable estimates of extension for the Triassic simply are difficult to obtain.

I assigned similar durations (25 m.y. and 45 m.y., respectively) to the first rifting phase (which lasted roughly from 230 to 205 Ma) and the resting phase (which lasted from 205 to 160 Ma) as I did in the generic model suite. I chose a duration of 35 m.y. (instead of 50 m.y.) for the second rift phase, however. The second phase began at about 160 Ma; my preliminary choice of 110 Ma for the initiation of sea-floor spreading yielded a duration of 50 m.y. for the second phase. A date of 125 Ma is a more accurate date for the rift-drift transition on the transects discussed here; hence, for the specific model, I revised my estimate of second-phase duration to 35 m.y.. The choices for distribution and duration of extension during the two rifting phases results in the rifting path shown in Figure 48. The extension rate of 9.7 km/Ma for the second phase closely resembles the rate of initial sea-floor spreading on the southern half of the Newfoundland margin (12 to 13 km/Ma) given by Srivastava et al. (1990).

Several parameters for the initial model were less well constrained. The original crustal thickness is almost impossible to constrain. The crust beneath
Figure 48. Rifting path used for "specific" model. Total extension for this model is 400 km and duration 105 Ma. Slope of line indicates rate of extension.
Newfoundland itself, in the heart of unextended Appalachian orogen, ranges in thickness from 40 to 45 km. The late Paleozoic orogen was enormously complex, and the part of the orogen about which we are most concerned has been obliterated by extension. Lacking any independent constraints on the original crustal thickness profile, I chose original crustal thicknesses between 40 and 46 km.

The crustal weakness in the specific model was chosen with a similar level of speculation. It was designed to mimic the distribution of the rift basins on the continental shelves of the Newfoundland and Iberia margins. For a reason I was not able to determine, the shelf basins nucleated on weaknesses inherited from different orogenies on both sides. On the Iberian side, the PBCZ—the major Paleozoic transform fault in western Iberia—became the eastern boundary of the Lusitanian basin, and served as the landward limit of extension. In southeastern Newfoundland, a similar major strike-slip fault, the Dover fault, appears to be the PBCZ's counterpart on the opposite margin, and would seem to be a location favored for upper-crust extension. Basins on the Newfoundland margin, however, nucleated on Precambrian Avalonian weaknesses, hundreds of kilometers seaward of the Dover fault. Thus, it is impossible to attribute with any certainty a single factor that controlled the upper crustal basin formation, and hence that could be chosen as a "crustal weakness" in the specific model. The role of any crustal weakness is a major uncertainty in the specific model.

The width and location of the crustal welt (the mantle weakness) in the initial model is speculative as well. Without exception, attempts to design a specific model with a wide crustal welt led to "wide rifting"—a style that I do not think is manifested on these conjugate margins. To achieve the narrow rifting style that appears to have occurred here, I had to use a narrow mantle weakness.
The position of the crustal welt in the specific model reflects the estimates of original margin width of Table 1: the Newfoundland margin (to the left) is designed to be much wider than the Iberian (to the right). Also, while the TTS values on the deep margins suggest asymmetrical rifting, the location of shelf basins and the crustal profile indicate a predominantly symmetrical rifting style. Attempts to model these margins using a specific model with moderate to large asymmetry resulted in extremely poor fits to the observed crustal profile. Thus, the weaknesses in the initial model are roughly symmetrical.

The initial specific model appears in Figure 49. The width of the model, the crustal thickness in the model, and the location of the mantle weakness are all as described above.

RESULTS OF THE SPECIFIC MODEL

The specific model mesh is shown in Figure 50. The larger date to the right of each mesh indicates the geologic date that the mesh is supposed to represent for the Newfoundland-Iberia margins; the smaller date indicates time since the start of the model, and corresponds with the time scale of the generic models. The lithosphere stretches very little in the first phase, and the root of the original orogen is removed, but further necking does not occur at the site of the mantle weakness. The Moho beneath the original crustal welt started with a higher temperature than the adjacent Moho. Although the upper mantle cools during the resting phase, this temperature difference is preserved, and by the beginning of the second rift phase (160 Ma/+70 m.y., Figure 50), the Moho beneath the area of first-phase necking still has a higher temperature that the neighboring Moho. Thus, the lithosphere is weakest there, and second-phase necking concentrates in the same place as first-phase extension. Necking continues at the same place until 130 Ma/+100 Ma. After this time, the mesh
Figure 49. Initial "specific" model. See text for details.
Figure 50. Deformation of "specific" model mechanical mesh. See text for details.
Figure 50 (continued).
becomes so necked that the method breaks down mathematically, and the model does not execute its last time step (125 Ma). Nevertheless, the mesh is sufficiently necked there to assume that, in the final time step, necking would simply become more extreme there.

The gray-scale plot of strain rate (Figure 51) shows the final concentration of necking of the rift. In strain rate plots of times before 145 Ma, the strain is similarly distributed. Between 145 and 135 Ma, however, necking rapidly concentrates at the location of highest strain. This timing is consistent with the shift of extension from the Grand Banks and Lusitanian basin to the Newfoundland basin and Iberian Abyssal Plain at the opening of the Valanginian (140 Ma).

Figure 52 shows the elevation profile of the specific model. The elevation on the "Iberian" (right) side of the specific model does not match the TTS on the IAP margin (Figure 18). Nevertheless, the abrupt change in slope between 100 and 200 km on the profile resembles the shelf break at about 450 km on the plot of Newfoundland margin TTS (Figure 16), and the slope of the model's surface between 150 and 350 km is similar to the continental slope on the Newfoundland margin (400 to 600 km, Figure 16). Although the model elevation will drop further due to thermal subsidence--rendering the fit poorer--the shape of the Newfoundland margin is predicted by the specific model.

Similarly, the crustal profile of the specific model fits the actual crustal thickness reasonably well on the Canadian margin, but poorly on the Iberian side. None of my attempts at specific models--including the final one shown here--featured broad zones of highly thinned crust, as are observed on both conjugate margins. I will discuss why I think the specific model failed in this regard in the next chapter.
Figure 51. Gray-scale strain rate plot for "specific" model. Grays range from $\log_{10}(\text{strain rate})$ of -14.9 (black) to -9.8 (white). Light background delineates maximum width of model and original thickness of model.
Figure 52. Elevation of the surface of the "specific" model at time 130 Ma/+100 Ma, the last executed step. See text for discussion.
This last issue notwithstanding, of all the specific models I tried, this model provided the most accurate fit to the crustal profile and the subsidence. If it is indeed true that the driving forces of extension between North America and Iberia paused for almost 50 million years, the model here suggests that the original crustal welt must have been narrow, and the amount of the extension during the first phase must have been minimal. Both implications are difficult to verify: the original crust has been obliterated by the extension we are examining, and the majority of first-phase syn-rift sediments accessible only through seismic surveys. I do not intend to imply that this model is the unique solution to the rifting between Newfoundland and Iberia; it is simply one possible model which fits the observations better than any other model I considered in my research. Nevertheless, any future attempt to model spreading between Newfoundland and Iberia must consider the effects of the resting phase which separates the two.
Chapter 8. Discussion and Conclusions.

CATEGORIZATION OF RIFTING STYLES

I have attempted to analyze my models in terms of the categories set forth by Buck (1991), who classifies all rifts into the categories of "narrow," "wide," and "core complex." These categories are based on such parameters as crustal rheology, crustal thickness, Moho temperature, surface heat flow, and strain rate. In all cases, the models in my generic suite fell into the "narrow rift" category; most of my models appear to exhibit features characteristic of Buck's "wide-rift" category, however.

This discrepancy can be explained by a difference in the parameters used in the categorization. Buck considered primarily dry rheologies, while my models incorporated only wet ones. Even so, a comparison to his Figure 12b (in which he uses wet quartz) still places my generic models in the "narrow" category. Buck's use of high strain rates in his models is probably the cause of this additional discrepancy: even though he uses a rifting velocity similar to mine, in his model all strain is concentrated over only 40 km, while my models are allowed to deform throughout. I feel that, even though our models do not agree upon first examination with Buck's categorization, they probably would be in accord once the difference in rheologies and strain rate is accounted for.

MAGMATISM IN THE DYNAMIC RIFTING MODELS

A number of workers have discussed the issue of magmatism at rift zones (Foucher et al., 1982; McKenzie and Bickle, 1988; White and McKenzie, 1989; White, 1993). As pointed out earlier, the Newfoundland-Iberia margins display
little rift-related volcanism, and are thus labeled "non-volcanic." Other areas, such as the Norwegian margin, are considered "volcanic" margins, producing large volumes of magma. It was my desire to estimate the amount of magma generated by the foregoing models.

Like Harry and Sawyer (1992b), I considered only partial melt generated within the upwelling asthenosphere (that is, material that lies "underneath" the finite element mesh); the mantle rock within the mesh is probably too cold to produce any magma. I next needed to consider the temperature of the underlying asthenospheric material. As discussed earlier, Bassi et al. (1993) believed that the heat flow bottom boundary condition allows the lithosphere to cool too fast; therefore, I attempted to imitate their temperature boundary conditions in order to generate a more realistic magma-generation model. For each vertical slice across each model, and at every 5 million years throughout each model, I computed the geotherm down to the original model depth (about 125 kilometers in all cases). This was done by using a one-dimensional finite difference routine to simulate the conduction of heat through the lithosphere. (The temperatures in this model were computed independently of those used in the dynamic models.) I maintained the surface of the model at 0°C, and a constant 1333°C at a depth of 125 kilometers.

At each time step, as the lithosphere thinned, I added an amount of "fresh" asthenosphere material at 1333°C at the bottom, whose thickness was equal to the amount of thinning. This scheme approximates the upwelling of asthenosphere into the space left by thinning lithosphere, and the subsequent accretion of that material onto the base of the lithosphere. Beneath an actual rift, this material may continue to convect, and thus maintain elevated temperatures, instead of accreting to the lithosphere. This possibility suggests that the temperature scheme used here may still underestimate temperatures
in the space down to 125 km depth. However, the scheme used here
approximates the lower-lithosphere processes better than the schemes
available in the dynamic models.

Once I had calculated a geotherm for every location and time in each
model, I computed the pressure imposed by the overburden. Hence, having a
pressure and temperature at every point, I used McKenzie and Bickle's (1988)
relations to compute the percentage of partial melt generated at each point.
Integrating the result over each vertical slice yielded the thickness of magma
that would be erupted on the surface at the top of each slice at each point in
time (assuming the melt travelled directly and immediately upward).

I applied this analysis to each model. In each case, no magma was
generated at any time at any location throughout the model run. For the rates
of rifting modelled here (10 km/Ma and less), the mantle cools too quickly to
allow asthenospheric material to decompress sufficiently to create melt. Since
even our fastest model does not produce any melt using this method, it can be
safely concluded that the slower models--and the actual margins they may
represent--are not close to producing any partial melt. This observation applies
to the specific model as well as to all of the generic model suite.

This conclusion is consistent with the observations from the Newfoundland
and Iberian margins: both the models and the actual margins show no
significant magmatism at the surface. Furthermore, the rate of rifting has a
decided impact on whether or not, and how much, melt will be generated by
riifting margins. In addition to considering the initial temperature of the
asthenosphere and the presence or absence of a "mantle plume" beneath a
riifting margin, those interested in examining melt generation at rifting margins
should consider the rate of rifting as well. Instead of considering only "hot and
cold rifts" (White, 1993), we should also consider "fast" and "slow" rifts in the magma-forming process.

CONCLUSIONS

Through analysis of the Total Tectonic Subsidence on the Newfoundland and Iberia conjugate margins, and dynamic modelling of multi-phase rifting in general and of the conjugate margins in particular, the following have been determined:

1. If it is assumed that extension between Newfoundland and Iberia took place via Wernicke-style (1985) simple-shear lithospheric or even crustal scale detachment, then the TTS data suggest that the detachment dipped down to the west. Only the final stages of rifting occurred by a simple-shear mechanism, however, since only the highly-subsided, seaward portions of each margin display asymmetry.

2. Upper mantle extension localizes where the Moho is hottest. This observation explains lateral migration of the locus of rifting as well as the rifting style produced by multi-phase rifting. In most cases, the site of the original rift will not be favored for extension when stretching resumes, since the Moho will have been elevated and cooled by the first rifting phase.

3. Rifting between Newfoundland and Iberia took place in the same location during both rift phases. If the first rifting phase (late Triassic-early Jurassic) had elevated and cooled the Moho--thus strengthening the lithosphere--the second phase (late Jurassic-early Cretaceous) probably would have occurred elsewhere. Hence, the first phase of rifting was probably relatively minor. In addition, the estimated amount of extension (about 400 km over an original width of about 600 km) and the duration of extension (35 m.y. in the last rifting phase) appear to be reasonably close to the correct amounts. While the
dynamic model of the Newfoundland-Iberia conjugate margins is not a unique solution, it provides the best fit to the observations on those margins.

4. The present models cannot predict the existence of a 400-km-wide zone of highly thinned continental crust. As discussed in the next section, this shortcoming may be inherent in the modelling routine used here.

5. No magmatism is predicted by the specific model. This observation is consistent with the lack of rift-related igneous activity on the Newfoundland and Iberia margins. The lack of volcanism is probably due to both the low initial asthenosphere temperature and the slow rate of rifting.

UNRESOLVED ISSUES

There are a few pertinent questions about the geology of the Newfoundland and Iberia margins that has been left unanswered by the dynamic modelling applied to these margins. The site of continental breakup occurred between two major strike-slip systems--the Dover fault and the Porto-Badajoz-Cordoba shear zone. Such a site would be expected: horizontal tensile stresses are more likely to cause a nearly vertical transform fault plane to fail than a low-angle thrust, so strike-slip faults serve as substantial crustal weaknesses. On a more detailed level, however, the Lusitanian basin nucleated on the PBCZ, while the Dover Fault is undisturbed by Mesozoic extension. I believe that, while the crustal weakness created by transform faults plays a major part in determining the location of basin formation and upper-crust deformation, it is not the major factor in the eventual location of continental breakup. The location of mantle weaknesses, specifically those created by thickened crust, serve as the primary factor in determining the future site of sea-floor spreading. On the Newfoundland-Iberia transect, this means that the crust thickened by Acadian-Ligerian and Alleghanian-Hercynian orogeny ultimately localized the
continental breakup. That the basins on the Grand Banks formed on inherited 
Precambrian structures, and not on Appalachian ones, suggests that those 
crustal weaknesses were simply the most convenient to manifest the upper- 
crustal expression of lithosphere-scale deformation—which was controlled by 
where the lithospheric mantle was weakest.

Strangely, the crust under Newfoundland itself has remained thicker than 
the surrounding crust. There, the crust is over 45 km thick, while to both the 
est and west, thicknesses range from 34 to 39 km (Keen et al., 1990). This 
crustal welt lies at the site of the Taconic suture between Laurentia and the 
Avalon prong. Since crust was thickened here—and hence the lithosphere 
weakened—this location should have been a favored location for Mesozoic 
extension. This is not the case, however; there is no evidence of any extension 
on Newfoundland itself. In fact, Taconic structures do not appear to play a part 
in the localization of extension on these margins. It seems that the forces 
which separated North America from Iberia were applied seaward of the 
Taconic crustal root, since modelling suggests that a crustal root is the first 
feature to disappear upon application of tensional stresses. This issue appears 
to be one that only detailed analysis of mantle dynamics can resolve.

The most important unresolved issue is the cause of hundreds of kilometers 
of ultrathin continental crust adjacent to the OCB. I have not yet seen a model 
created by the finite element method and by rheologies used here that results 
in such a feature. It must be noted, however, that the dynamic modelling 
routine used here does not allow for the addition of material to the model. 
Since the base of the specific model cools to well below 200°C at its final time 
step, it is not unreasonable to assume that asthenospheric mantle that was 
one below the model (and too hot and weak to contribute to the lithosphere's 
strength) has cooled to temperatures at which it can play a substantial role in
the strength of the crust. The durations of rifting modelled here are certainly slow enough to allow sufficient cooling for this to take place. Hence, new lithosphere may be added to models and subsequently extended. This would allow the crust to thin to greater and greater amounts, beyond the point where one would normally expect initiation of sea-floor spreading. This hypothesis is supported by the lack of magmatism on the margins: the absence of a magma supply would allow lithospheric thinning to continue further and further before the creation of oceanic crust began. This hypothesis is also supported by the serpentinized peridotite that may underlie the highly thinned crust of the IAP, which would represent upper mantle rock elevated to near the surface. In addition, the serpentinite samples from the Galicia margin appear to have been fractured (Boillot et al., 1987); this agrees with the specific model, where temperatures at the base of the model are low enough that mantle material would be well into the brittle-failure regime.

The idea that the peridotite may have been accreted to the bottom of existing lithosphere has been presented before (Boillot et al., 1987, 1988, 1989b). Boillot and his colleagues, however, invoke a simple-shear detachment which cuts entirely across unthinned continental crust to explain the Iberian margin. While the lithosphere might have ultimately failed by such a simple-shear mechanism, it may have occurred only towards the end of a long rifting episode. Pure-shear deformation, as portrayed by the models here, may neck the lithosphere to the point where it is thin enough to allow a single through-going normal fault or shear zone to form, unroofing the upper mantle material. Indeed, the asymmetry of the TTS data, coupled with the apparent symmetry of the margins as a whole, suggest that simple-shear rifting did not develop until late in the rifting history. Thus, a crustal-scale detachment is not necessary to bring upper mantle rock to the surface.
FUTURE EFFORTS

Before further investigation of these margins is undertaken using dynamic modelling, I believe a revision of the modelling routine is called for. The current models clearly cannot explain the existence of the broad zone of highly thinned continental crust; this feature may due to the addition of material to the lithosphere. Thus, I would recommend that future modellers modify their method to include this addition of material; the solution to the ultrathin-crust problem may lie in such a modification. Similarly, I would adopt a thermal bottom boundary condition similar to that used by Bassi et al (1993); such a change would better approximate the model's thermal regime, and hence, improve the accuracy of the models that are executed.
Appendix. Calculation of $\beta$ from TTS.

Le Pichon and Sibuet (1981) provided the equations for subsidence after an instantaneous one-dimensional "riifting event." The equation relating initial subsidence to extension parameter $\beta$ is

(1) \[ Z_i = c_1 \left( 1 - \frac{1}{\beta} \right) \]

where

(2) \[ c_1 = \frac{h_i \rho_a - h_c \rho_c - (h_i - h_c) \rho_l}{\rho_a - \rho_w} \]

and

$Z_i =$ initial (syn-rift) subsidence,
$h_i =$ initial thickness of lithosphere,
$h_c =$ initial thickness of crust,
$\rho_a =$ density of asthenosphere,
$\rho_c =$ density of crust,
$\rho_l =$ density of lithosphere, and
$\rho_w =$ density of water.

In addition,

$\rho_a = \rho m(1 - \alpha T_a)$,
$\rho_c = \rho c_0(1 - (\alpha/2)T_a(h_c/h_i))$, and
$\rho_l = \rho m(1 - (\alpha/2)T_a - (\alpha/2)T_a(h_c/h_i))$, 
where

\[
\begin{align*}
\rho_m &= \text{density of mantle at } 0^\circ\text{C}, \\
\rho_c0 &= \text{density of crust at } 0^\circ\text{C} \\
\alpha &= \text{coefficient of thermal expansion, and} \\
T_A &= \text{temperature of the asthenosphere.}
\end{align*}
\]

Le Pichon and Sibuet (1981) also provided the equation for total subsidence, which includes both the initial (syn-rift) subsidence and the thermal (post-rift) subsidence; this equation is

\[(3) \quad Z_\infty = c_2 \left(1 - 1/\beta\right)\]

where

\[
Z_\infty = \text{total (syn-rift plus post-rift) subsidence, and}
\]

\[(4) \quad c_2 = \frac{(\rho_m - \rho_c) \left[1 - (\alpha/2)T_A(h_c/h_l)\right] h_c}{\rho_a - \rho_w}\]

From here, I assumed that thermal subsidence occurred in exponentially-decaying fashion, such that

\[(5) \quad Z(t) = Z_i \exp(-t/\tau_c) + Z_0[1 - \exp(-t/\tau_c)]\]

where

\[
\begin{align*}
Z(t) &= \text{subsidence at time } t, \\
t &= \text{time since the instantaneous rift event,} \\
\tau_c &= \text{time constant associated with lithospheric cooling, and} \\
Z_0 &= \text{thermal (post-rift) subsidence, } Z_\infty - Z_i.
\end{align*}
\]
Thus, the subsidence starts at $Z(t) = Z_i$ at $t = 0$ and approaches $Z(t) = Z_\infty$ asymptotically as $t \to \infty$. By algebraically combining equations 1 through 5, it is possible to write current subsidence $Z(t)$ in terms of $\beta$:

$$Z(t) = c_3 \left(1 - \frac{1}{\beta}\right)$$

where

$$c_3 = c_1 \exp(-t/\tau_C) + c_2[1 - \exp(-t/\tau_C)].$$

Since $c_3$ is kept the same across a transect of a margin, the extensional parameter $\beta$ can be found at each point on that transect as a function of the observed TTS, $Z(t)$.

For the margins considered here, the parameters in Table A1 were used. These parameters yielded

$$c_3 = 9.75 \text{ km.}$$
\( h_l \) Initial thickness of lithosphere 135 km
\( h_c \) Initial thickness of crust 40 km
\( \rho_w \) Density of water 1020 kg m\(^{-3}\)
\( \rho_m \) Density of mantle at 0\(^\circ\)C 3350 kg m\(^{-3}\)
\( \rho_c \) Density of crust at 0\(^\circ\)C 2780 kg m\(^{-3}\)
\( \alpha \) Coefficient of thermal expansion 3.28 \( \times \) 10\(^{-5}\) \(^{\circ}\)K\(^{-1}\)
\( T_a \) Temperature of the asthenosphere 1333 \(^\circ\)C
\( t \) Time since the instantaneous rifting event 124 Ma
\( \tau_c \) Time constant associated with lithospheric cooling 62.8 m. y.

**Table A1.** Parameters used in the calculation of \( c_3 \).
References.


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