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Structural and tectonic interpretation of deep seismic reflection data offshore Spain and Portugal: a tectonic rifting model

By

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ABSTRACT

Structural and tectonic interpretation of deep seismic reflection data offshore

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A portion of new seismic reflection/refraction data collected over the Iberia margin were processed and interpreted, resulting in the recognition of two previously unknown normal faults in the deep crust. These faults, named Q and R, formed during the Triassic-Cretaceous rifting of Pangea, and formation of the Atlantic basin. Previous studies have shown a third surface, the low-angle S reflector, in the area. S has previously been interpreted as a detachment surface or as a velocity interface. A model is proposed where extension was accommodated along Q, R and S during the rifting of Pangea. R and S acted as detachment surfaces as the hangingwall block moved westward. Late in the rifting process, the geometry of R and S was altered by motion of underlying serpentinized mantle material. This upwelling of the mantle controlled the shape of S in three dimensions. Total extension across the margin is estimated at 1.67.
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Chapter 1: Introduction

The study of passive margins is important for a number of scientific and economic reasons. Sedimentary systems on passive margins are significant sources of, and reservoirs for, petroleum (oil, natural gas, and gas hydrate) deposits. The current world economy relies on these deposits for fuel and raw materials to power and create the civilization we know. Passive margins also contain extensive deposits of carbonate rocks and other materials which sequester a significant portion of the carbon available in the carbon cycle.

Because both of these systems rest on passive margins, the study of the evolution of such margins is a fundamental step in understanding them. A large part of such study focuses on their evolution. All passive margins form during a rifting event. The distribution of extensional strain through time controls the location and architecture of such sedimentary systems. Therefore, an understanding of the structural geology of the crust underlying passive margins permits a more complete understanding of these surficial systems. At a global scale, study of passive margins leads to a greater understanding of the properties of continental crust. Studying why passive margins occur where they do leads to further understanding of the global stress field. Understanding the processes that drive extension and rifting also enables a better comprehension of how the forces of subduction are transmitted within the crust and mantle across continental distances. This knowledge furthers our insight into how the physical and mechanical properties of the crust and mantle change with depth and distance.
GEOLOGICAL & GEOPHYSICAL BACKGROUND OF THE IBERIA MARGIN

The Iberia margin (figure 1) was recognized by the JOIDES Passive Margin Advisory Panel as an attractive place to study a passive continental margin. The lures of Iberia include the absence of major offshore deltas and therefore a relatively thin blanket of clastic sediment, as well as a lack of volcanic features and a minimum of evaporites. These latter characteristics make the possibility of drilling into the crystalline crust underlying the margin more attainable in a location like Galicia than compared to African or South American margins (Shipboard Scientific Party, 1979). Also, this thin sedimentary blanket permits the recording of high-quality seismic reflection data to long two-way travel time (TWTT) (i.e. to deeper crustal levels), data that are key to understanding the regional and temporal evolution of a margin. When seismic data are combined with the detailed stratigraphic knowledge derived from individual borehole sites, a holistic picture of a margin and its life cycle can be developed.

This relatively thin (less than 3 seconds TWTT) sediment cover has encouraged a large number of scientific studies of Galicia over the past 30 years. One of the first was a visit by the *Glomar Challenger* during Leg 47 of the Deep Sea Drilling Project (DSDP) in 1976, when Site 398 was drilled on the southern edge of the Galicia Bank. This drilling study also collected a number of shallow seismic reflection lines within the study area. Other seismic reflection studies include the collection of 8350 kilometers of mainly single channel seismic data (Dupeuble et al., 1976; Dupeuble et al., 1977), discussed in Groupe Galice (1979).

These studies set the stage for Ocean Drilling Program (ODP) Leg 103 to investigate the Galicia Bank in April, May and June of 1985 (Boillot et al., 1987; Boillot et al., 1988c). Further to the south, ODP Legs 149 (Sawyer et al., 1994; Whitmarsh et al.,
1996) and 173 (Whitmarsh et al., 1998) were conducted across the Iberia Abyssal Plain (figure 1) in March, April and May of 1993. These experiments drilled and retrieved sediment and some basement cores, which greatly increased the understanding of the sedimentary units present along the margin. They also conducted underway geophysical studies which provided data number of conflicting hypothesis (c.f. Boillot, 1988; Winterer, 1988).

Earlier geophysical studies suggested that the basement of the Iberia margin under the Galicia Bank was faulted, and that N-S trending faults formed basins into which sediments accumulated (Montadert et al., 1974). Further work indicated that the basins to the east of the Galicia Bank in the Galicia Interior Basin (figure 1) typically formed full grabens, separated by horsts (Groupe Galice, 1979), while to the west, basins were half-grabens, with the block-bounding faults dipping to the west (de Charpal et al., 1978). De Charpal et al. (1978) also noted a sub-horizontal reflector at 8.0 to 10.0 seconds TWTT below the tilted fault blocks to the west of the Galicia Bank and on the northern margin of the Bay of Biscay (figure 1). De Charpal et al. (1978) named this reflector S. Above the tilted fault blocks, the sediments within the basins are categorized into pre-rift packages, syn-rift sediments, and post-rift sediments.

The DSDP, ODP and other studies provide a solid framework of data to the Iberia Seismic Experiment (ISE '97) which was conducted in June through August of 1997. The ISE '97 data represent a fundamental improvement in the signal quality and penetration depth compared to previous seismic studies on the Iberia margin. When combined with a concurrently collected detailed refraction study, which provides wide-angle velocity data along and across the margin, researchers now possess a dataset from
Figure 1. Map of the locations of ISE 97 lines, ODP sites, and physigraphic features of the Iberia Margin. An index map of the location of the study area is shown in the lower right corner.
which many of the questions raised by previous workers may be answered. This thesis presents a portion of the data collected during ISE '97, discusses a number of seismic processing techniques applied to these data, advances a structural and stratigraphic interpretation of the data, and finally, presents a possible hypothesis for the creation of the observed structures and features found along the Iberia Margin, and specifically, the Galicia Bank (see figure 1 for location of ISE '97 and prior studies). crustal structure prior to extension and rifting.

In order to understand the current state of the Iberia Margin, it is important to examine the pre-rift Variscan-Hercynian structures that formed a fabric which controlled the location and trends of breakup (Tankard and Welsink, 1988). These structures are related to the closing of the Iapetus Ocean in the Alleghanian during the Appalachian Orogeny and the formation of the Pangea supercontinent, as well as earlier structures created during the closure of Cambrian ocean basins, and the Proterozoic Grenville orogenic event (Rast, 1989). The dominant tectonic features of these orogenic episodes are thrust fronts created by contraction, zones of transverse motion which accommodated differential movement between thrust sheets, and foreland basins into which syn- and post-orogenic sediments were shed (figures 2 and 3).

Amalgamation of Pangea

In the Cambrian, Laurentia (composed of Archean and Precambrian Canadian Shield material), Baltica (composed of northern European continental blocks) and Gondwana (composed of the African, Antarctic, Australian and Indian continental blocks) interacted in a triple-point type system, separated from each other by the Iapetus Ocean (Rast, 1988). In the north, a number of exotic terranes laid between Laurentia and Baltica (c.f. Olszewski (1982). Shortening between Laurentia and Baltica in the
Paleozoic formed the northern Appalachian Mountains, while contraction between
Laurentia and Gondwana created the southern Appalachian mountain range.

Subduction of the Laurentian plate to the east in the Paleozoic (Strong et al.,
1974) caused two main contractional events in the northern Appalachians: the Taconic
Orogeny and Acadian Orogeny (Colman-Sadd, 1982) (figure 2). The Taconic Orogeny
occurred in the Ordovician as the Iapetus Ocean closed along an eastward dipping
subduction zone. During this event, a mid-ocean island arc collided with the Laurentian
margin, and accreted allochthons ranging from distal passive margin rocks through
ophiolites onto the former passive margin deposits.

During a significant pause of approximately 60 Ma (Colman-Sadd, 1982) erosion
deposited clastics and carbonates in a shallow sea east of the Taconic orogen and west of
the Grenville aged Avalon superterrane (Olszewski and Gaudette, 1982) Rast, 1989). In
the Late Silurian, renewed shortening caused the Acadian Orogeny as the Avalonian
block converged on Laurentia (Keen et al., 1990). This orogeny caused polyphase
deformation of the Central Mobile Belt of the Avalon terrane (Rast, 1989), and intrusion
of large, non arc-related, granitic plutons into the crust (Strong, 1980). Finally, the entire
ocean closed as Baltica collided with the east side of Avalon (Rast, 1988).

This series of accretionary events, separated by hiatuses, created the material that
was rifted during the formation of the North Atlantic. This shortening created orthogonal
trending lines of weakness, with a N-S trend related to the shortening structures and a E-
W trending fabric arising from the transfer domains. The structures along which these
various terranes were juxtaposed appear to have controlled the localization of strain
during the later Mesozoic extension.
Both major and minor transfer domains segment along-strike features in the northern Appalachians. The major transfer domains include the multiply-named Minas geo-fracture zone/Cobequid-Chedabucto fracture zone/south-east Newfoundland transform fault (MGFZ) on the south side of the Grand Banks and the Dover fault/Charlie Gibbs fracture zone to the north (figures 2 and 3). These transfer domains influence the tectonic features of the modern North Atlantic region. For instance, the MGFZ connects to the South Atlas fault in North Africa via the Pico fracture zone in the western North Atlantic, and the East Azores fracture zone (Keen et al., 1990; Rast, 1988; Widmier et al., 1985). In much the same way, the boundary between the Avalon terrane and the Central mobile belt extends to the northeast, linking into the Charlie Gibbs fracture zone, which extends eastward across the Atlantic Ocean until it terminates at the Rockall Trough near Porcupine Bank (Widmier et al., 1985).

Deformation of the crust during extension seems to be localized by these transfer domains. For instance, the Scotian basin WSW of the Grand Banks (figure 3) is composed of down to the ocean simple normal faults in a horst and graben regime (Keen et al., 1990). Just to the north of the Scotian basin, across the MGFZ, the Grand Banks and opposing Galicia margin appear to have extended in a radically different way along a crustal-scale simple shear zone (c.f. Wernicke, 1985). Within the Grand Banks domain there are at least two, smaller transfer zones (figure 3) (Welsink et al., 1989). These serve to segment the Grand Banks block into even smaller domains.

This fabric, formed from a combination of shortening and transverse motion, mechanically set the stage for the extension that commenced in the Permo-Triassic, and
continued diachronously as rifting began in Central Atlantic and propagated northward (figures 4 (a)-(d)).

**Rifting and Opening of the Atlantic**

After the Iapetus ocean closed and the Appalachians formed through collision and accretion in the Paleozoic, the Pangea supercontinent composed of Laurentia, Baltica, and Gondwana endured for nearly 120 million years, until it began to breakup in the Triassic (e.g. Manspeizer et al., 1989; Wilson et al., 1989). During the earliest phase of extension, N-S trending basins developed to the east of the entire Appalachian range, and terrestrial sediments (i.e. red beds and evaporites) were deposited into these basins (Tankard and Welsink, 1988; Murillas et al., 1990).

In the southern Appalachians, extension was a complex diachronous process stretching from the Late-Triassic through the Middle Jurassic (Withjack et al., 1998). East-west oriented extension and rifting of the continental crust first changed to sea-floor spreading in the Blake Spur area (Dunbar and Sawyer, 1989) off the coast of Florida. Sea-floor spreading propagated northward as Africa to rotated out of the Pangean supercontinent counter-clockwise, away from Laurentia and Baltica (Srivastava and Tapscott, 1986; Tucholke and Jansa, 1986) (figure 4(a)). Associated with the rift/drift transition in the southern North Atlantic are volcanic deposits found in rift basins along the east coast of the United States and in seaward dipping reflectors typically found offshore in the continent/ocean transition zone.

In the northern Appalachians, east-west directed Triassic extension appears to have ceased shortly after its onset, and forms aborted rift systems such as the Triassic
Figure 2. Hercynian and other pre-Triassic structures of the North Atlantic from Rast, 1989. This is one of many reconstructions of the North Atlantic during Permian times, however most follow similar trends. Note similarity of the orientation of rifting axes to the older contractional structures.
Figure 3. Map showing the Appalachian tectonics preserved on the western side of the North Atlantic, as well as the Triassic–Cretaceous basins that formed during North Atlantic rifting. Rift basins are the Orphan Basin (OB), Jeanne d’Arc Basin (JdA), Horseshoe Basin (HB), Whale Basin (WB) and Scotia Basins (SB). Transfer zones are indicated by TZ. Also shown is the interpreted OCT, the magnetic anomalies M0 and 34 from Widmier (1985) and transform faults in the ocean. These tectonic features controlled the localization and style of breakup of Pangea during the Cretaceous.
Figure 4. Four stages of the opening of the Atlantic from the Jurassic to the Late Cretaceous, after Tuchokle and Jansa, 1986. In (a), seafloor spreading has initiated in the southern North Atlantic, and has propagated northward until encountering the Minas geo-fracture/Atlas fault system. To the north of the Minas geo-fracture/Atlas fault, distributed extension is occurring between the Grand Banks and Galicia Bank, creating epicontinental basins into which shallow marine and clastic rocks are deposited. Distributed extension in the northern epicontinental basins continues in the Late Valanginian while seafloor spreading progresses eastward across the Strait of Gibraltar (b). The diffuse extension in the north increases the basin depth and thins the continental crust. In the Tithonian, breakup between the Grand Banks and the Galicia Bank occurred, and seafloor spreading initiated (c). Shortly afterward, seafloor spreading began in the Bay of Biscay and spread northwestward up the Labrador Sea. By the Late Cretaceous, (d) the Grand Banks and Galicia Bank were completely separated and no longer interacting.
Galicia Interior Basin between the Grand Banks and the Iberia Massif (figure 4(a)). Later, extension occurred in the hinterland of the Appalachian orogeny, as Africa continued to rotate counter-clockwise out of the Pangean supercontinent along the MGFZ (figure 4(a)). As this occurred, diffuse distributed extension occurred between the Grand Banks and the Galicia Bank (Srivastava and Tapscott, 1986; Tucholke and Jansa, 1986) and formed N-S trending basins into which clastic rocks and carbonates were deposited as the epicontinental paleo-Atlantic ocean entered via the Tethys seaway, (Tucholke and Jansa, 1986).

Extension in the Grand Banks/Galicia Bank region then paused again as displacement was passed on the MGFZ to the east of the Iberia Massif and sea-floor spreading occurred in the Tethys (Tucholke and Jansa, 1986). Tethian sea-floor spreading ceased in Late-Jurassic time, and deformation again became concentrated in the North Atlantic between the Grand Banks and Galicia Bank. During this time, more shallow marine carbonates and associated clastics were deposited in the Grand Banks/Galicia Bank area (figure 4(b)).

Final rifting and the initiation of E-W seafloor spreading at Galicia began in the Tithonian (Jansa et al., 1988), and continued through the Lower Aptian (Shipboard Scientific Party, 1987e) (figure 4(c) and (d)). At least two wide-spread unconformities related to the breakup are found across the Grand Banks. These suggest that the Grand Banks area was uplifted at the end of rifting and during the rift/drift transition (Keen et al., 1990). Seafloor spreading in the North Atlantic continues today, separating North America from Europe at approximately 2.5 cm/yr. (Tamaki, 1994).
SEDIMENTOLOGY AND SEISMOLOGY OF THE IBERIA MARGIN

The rocks of the Iberia margin have been sampled in a number of ways over the course of the last 20 years. Leg 103 of the Ocean Drilling Program (ODP) drilled five sites on the Galicia Margin in 1985 (Boillot et al. (1987); Boillot et al. (1988c) (see figures 1 and 6 for locations of these sites). Deep Sea Drilling Project Site 398 was drilled to the south-east of the Galicia Bank, in the Galicia Interior Basin (Sibuet et al., 1979). ODP legs 149 and 173 drilled a transect across the Iberia Abyssal Plain, 140 kilometers to the south of the Galicia Bank (Sawyer (1994); Whitmarsh (1996); Whitmarsh (1998).

Pre-Rift Sediment and Acoustic Basement

The oldest crustal unit recovered from along the Galicia Bank is the uppermost acoustic basement. It was sampled in ODP holes 639 E and 639 F (Shipboard Scientific Party, 1987d) (figure 5). This site is in the hangingwall on the western side of a tilted fault block. Recovery of material from these holes was difficult, but rhyolite cobbles, quartz sandstones, graywackes, metasandstones, and quartz breccias were retrieved. The provenance of this material is equivocal: it may be from a basal conglomerate, it could be part of a submarine talus bed resting upon the fault surface, or it may in fact be derived from brecciated rhyolite flows that compose the continental basement (Evans, 1988). To the west, the basement is exposed at the seafloor along a normal fault hangingwall on the western side of a rotated fault block. This site was sampled in situ in 1986 by the French submersible Nautilus during the Galinaute cruise at dive site 11. At this site, samples of deformed granite and granodiorite were recovered (Boillot et al., 1988a) (figure 5).

Elsewhere, the basement in the Galicia Bank area has been sampled by dredging. These dredge hauls indicate that basement rocks on the northwest side of the Galicia
Bank are composed of peraluminous granite and micaschist. To the southeast of the Galicia Bank the basement is composed of greenschist facies metasediments and biotite-bearing granodiorites while towards the south south-east edge of the Galicia Bank it is made up of granulites. This basement structure suggests that the tectono-metamorphic units that compose the crust along the northern Galicia Margin shallow towards the west (Capdevila and Mougnot, 1988).

The rocks that compose the basement are thought to have formed during the Appalachian orogeny, as mudstones, arkosic sandstones, and minor amounts of carbonates were metamorphosed and deformed with crenulation to slaty cleavage or schistose foliation. There are at least two generations of structure. Intrusion of calc-alkaline and peraluminous granites occurred early during the main phase of deformation, followed by later intrusions of peraluminous granites, calc-alkaline granites, granodiorites and tonalites. A second, lower temperature, deformation imposed a greenschist to low amphibolite facies metamorphic overprint (Capdevila and Mougnot, 1988). These rocks were then uplifted and eroded in the hinterland, before being extended and rifted (Colman-Sadd, 1982; Rast, 1989).

During the extension of the Appalachian hinterland, which was followed by rifting and later opening of the North Atlantic, sedimentary sequences were deposited into rift basins and low points along the collapsing orogen above the pre-rift sediments. As these sediments lithified, the basins into which they were shed rotated, forming tilted fault blocks composed of pre- and syn-rift sediments underlain by crystalline basement material. Seaward of this series of tilted fault blocks that are is a N-S trending outcrop of serpentinitized peridotite (figure 1). This peridotite ridge extends for at least 130
kilometers north-south (Boillot et al., 1988b), and was drilled at site 637 on ODP leg 103 (Shipboard Scientific Party, 1987b). The Peridotite Ridge was also sampled by Boillot et al (1988a) using the French submersible Nautil. Studies of recovered peridotite show that it is highly altered by seawater, with more than 90% of the peridotite being altered to serpentine. It is relatively undepleted in magmaphile elements, indicating that it has experienced little partial melting. Foliation in the peridotite dips 15° to 60° to the east or the northeast, showing top to the east shear indicators.

**Syn-Rift Sediments**

Syn-rift sedimentary rocks were sampled at ODP sites 638, 639, and 640 across the Galicia Bank (Shipboard Scientific Party, 1987d, e), and in DSPS hole 398 (Shipboard Scientific Party, 1979). Sites 638 and 640 were drilled on footwall blocks, within 2 km of the normal faults, in order to sample the syn-rift stratigraphy. Site 639 straddled a normal fault, penetrating both the footwall and the hanging wall, supplying information on both the pre-rift and syn-rift stratigraphy (Shipboard Scientific Party, 1987a).

The oldest syn-rift unit sampled was in hole 639D, where Tithonian limestone unconformably overlies the basement. This limestone is interstratified with marlstone and lesser sandstone and claystone, and is at least 84 meters thick (Shipboard Scientific Party, 1987d). The sediments directly above this early Tithonian limestone were not sampled in any Leg 103 holes, however limited amounts of stratigraphically higher dolomites and interbedded muds (Jansa et al., 1988) were recovered in holes 639C and B. Due to the limited recovery, the provenance of this material is uncertain: it may be either slump deposited or in its original depositional location. No age or thickness determinations were possible on this unit.
Hole 639A found a 50 m thick layer of Middle Early Valanginian marlstone, interbedded with minor sandstone above the older dolomites (Jansa et al., 1988). The next unit recovered was a series of Hauterivian sandstone and claystone turbidites in holes 638B and C (Jansa et al., 1988; Shipboard Scientific Party, 1987c). Above this is a 300+ m thick unit composed mainly of claystone and marlstone.

The lowermost unit of this syn-rift succession is interpreted to have formed just offshore from a carbonate bank in the Late Tithonian, in 30 to 70 m of water (Jansa et al., 1988) as rifting first began to extend the area. After these rocks were deposited, on the high points of tilted fault blocks, they were exposed subaerially, leading to the dolomization of the limestone sequence (Loreau and Cros, 1988). Later, as extension progressed, the Tithonian carbonates resubmerged and shallow water sediments were deposited in the middle Valanginian. Another sea-level fall may have occurred before the initiation of the final transgression which deposited Hauterivian aged turbidites (Jansa et al., 1988). At this point the extension was concentrated in the Galicia Bank area, and the water remained deep as the last syn-rift unit of marlstones, claystone and turbidites was deposited in the Hauterivian to Late Aptian (Shipboard Scientific Party, 1987c).

These sedimentary units were assigned to Acoustic Units 4 and 5/5A by Mauffret and Montadert (1988), following after Groupe Galice (1979) (see figure 6). Acoustic Unit 5 of Groupe Galice (1979) was interpreted as pre-rift sedimentary rocks. Unit 5A is composed of syn-rift material (found in hole 639D), and typically shows sigmoid to oblique reflections. Acoustic Unit 4 onlaps unit 5A, and is bound at the top by a major unconformity. This upper unconformity is the breakup unconformity, and may be Late
Peridotite Ridge

ODP Site 637 (Party, 1987a):
1) Highly altered peridotite (90% serpentinite) exposed at the seafloor.
2) Relatively undepleted in magmaphilic elements, indicating little to no partial melting.
3) PT study shows stability at 9kbar 900\(^\circ\)C.
4) Harzburgites and Iherzolites cut by calcite veins.
5) Foliation dips 30\(^\circ\)-60\(^\circ\) to the east.

Nautilie site 4 (Boillot et al., 1988a):
1) Lower Cretaceous syn-rift sediments onlap.
2) Foliation dips 15\(^\circ\) to the northeast.

Other Nautilie Peridotite Ridge sites:
1) Show Peridotite Ridge exhumed at seafloor only since Eocene/Oligocene (Iberian-European plate convergence).

Acoustic Basement

ODP Site 639 E, F:
1) Rhyolite cobbles, quartz sandstones, graywackes, metasandstones and quartz breccias (Evans, 1988)

Nautilie site 11:
1) Basement composed of granite, granodiorite covered with ~300 m poorly sorted coarse to medium grained sandstone and conglomerate (pre-rift) (Boillot et al., 1988a)

Syn-Rift Sediments

ODP Site 638, 639, 641:
1) Base is Tithonian carbonates, marlstone, sandstone (Unit V). Deposited at 30 - 70 m on carbonate bank (Moullade et al., 1988). Dolomitization indicates later emersion and sub-arial exposure due to early extensional tectonics prior to block tilting (Hole 639, Loreau and Cros, 1988).
2) Unit IV is undated dolomite, 215 m thick.
3) Indication of deepening water depth in Units V-IV (Jansa et al., 1988)
4) Unit III is Mid-Early Valanginian - Hauterivian. Marls, marlstone, sandstone. Shallower water depth, meaning uplift of fault block (Jansa et al., 1988)
5) Unit II is Hauterivian - L. Aptian nanofossil marlstone, claystone, microturbite.

Figure 5. Location map and brief summary of previous studies in the S Reflector area, Deep Galicia Basin, as they relate to this study. Faults from Thommeret et al., (1988).
Figure 6. Mauffret and Montadert's (1988) preexisting division of seismic units along the Galicia Margin. Acoustic Units 1-3 are interpreted to be post rift deposits, units 4 and 5/5A are thought to be composed of syn rift sediments and 5B and below are considered to be pre-rift strata (including crystalline basement). GP-101 is an east-west trending line collected along 42.1°N. The locations of ODP sites 638, 639 and 641 are projected onto the lower line (GP 101-B2). Note the S reflector indicated by the arrow below Site 639.
Barremian to Late Aptian in age. Reflectors in Acoustic Unit 4 are typically well layered, and onlap onto the sides of the fault-bound basins. They are typically fan-shaped, diverging to the east in the deeper portion of the basins. In places the reflector geometry becomes chaotic. This is probably related to syn-rift slumping and talus slopes (Mauffret and Montadert, 1988).

**Post Rift Sedimentary Units**

The uppermost sedimentary units along the Galicia margin are spatially diverse. They range from pelagic and hemipelagic clayey carbonate sediments with minor bottom current deposits along the southernmost portions of the Galicia Bank (Shipboard Scientific Party, 1979) to turbidites, contourites and pelagic clays on the distal western deep portions of the margin at Leg 103 Site 637 (Comas and Maldonado, 1988).

These sedimentary rocks are interpreted to have been deposited after the cessation of rifting, in a deep water environment (generally below the Carbonate Compensation Depth (CCD) (Shipboard Scientific Party, 1979, 1987a). They are considered to be Albian to Late Cretaceous at the base, and Holocene at the seafloor (Shipboard Scientific Party, 1987a). This sedimentary package is divided into three seismic stratigraphic units, labeled Acoustic Units 1-3, from the top down, as seen in figure 6. These seismic reflectors are dominantly conformable, and onlap onto local highs, typically tilted fault blocks.

Unit 3 is acoustically transparent, composed of black shales and mudstones, of Early Albian to Middle Cenomanian age, and thought to represent organic-rich distal turbidite deposits. This unit fills sub-basins and other low points that remained after the completion of rifting. Acoustic Unit 2, is a series of strong, parallel reflectors, interpreted to be Late Cretaceous to early Cenozoic condensed sections resulting from continued
pelagic sedimentation. Acoustic Unit 1 is composed of upper Eocene to Holocene pelagic sediments and interbedded turbidites, and is composed of a series of conformable reflectors (Shipboard Scientific Party, 1979, 1987a).

**S Reflector Area**

To the west of the Galicia Bank, between 8 and 10 seconds TWTT, lies the S reflector. Also know simply as S, this feature is sub-horizontal to dipping, and flat to concave or convex, depending upon the processing sequences used to image it. S is seen on ISE '97 Lines 1, 2, 4, 5, 6 and 15. On Line 4, it is visible across nearly 30 km, between CMPs 1500 and 3900, while on Line 1 it stretches for 45 km (CMPs 2500 to 6000). On the time sections S is undulatory, due to the effects of velocity pull-up. In depth converted sections, S tends to be flat to slightly undulatory.

Modeling of the reflection character of S by Krawczyk and Reston (1995) shows that the waveform that is S has a shape very similar to the water sediment interface in the S reflector area. Because of the lack of evidence for multiple interfaces comprising S, Krawczyk and Reston suggest that S arises from a single, high impedance interface, perhaps the Moho, rather than multiple interfaces that would be expected if S was caused by intrusion of volcanic sills.

An understanding of the structure of S, and of the rocks above and below S are crucial to understanding the rifting history of the Iberia Margin. Any model for the formation of the margin is required to explain S and the structures associated with S.

**Structure of the Newfoundland Margin**

The previous studies along the Iberia margin have provided a wealth of information on the tectonostratigraphic history of the region. They are the initial
glimpses into the history of the tectonics of Iberia. However, in order to fully understand
the Iberia Margin, we need to examine not just the sedimentology at a series of distinct
sites, but we also need to investigate the evolution of the conjugate margin and the
structure that underlies the Iberia Margin. A brief description of the conjugate margin is
supplied next, followed by a series of models that have been advanced so far to explain
the tectonic evolution of the Iberia Margin.

Opposing the Galicia Bank, on the western side of the Atlantic Ocean, is the
Newfoundland margin (figure 3). The Triassic-Cretaceous linkage of Galicia Bank and
the Grand Banks of Newfoundland is shown in figure 4. The Newfoundland margin is
bound by two zones of transform displacement: on the south is the Newfoundland
fracture zone (a portion of the MGFZ) and on the north is the Charlie Gibbs fracture
zone. Between these two major transform zones are a series of basin trends, offset into
individual basins by smaller transform zones (figure 3). The basement of the
Newfoundland margin is composed exclusively of Avalon block material.

A number of authors have reconstructed the pre-Atlantic rifting positions of the
Galicia Bank, Iberia Massif, Grand Banks of Newfoundland, Flemish Cap (c.f. Srivastava
et al., 1990; Tucholke and Jansa, 1986; Ziegler, 1989). These reconstructions show that
the Flemish Cap and Galicia Bank are conjugate. All of the tectonic reconstructions for
this region are very similar.

Much seismic data have been collected on the Newfoundland margin because of
the nearly 2 billion barrels of petroleum reserves discovered in the Jeanne d’Arc basin
(Tankard et al., 1989). Unfortunately, none of these data are directly conjugate to the ISE
’97 data. Therefore, the National Science Foundation funded Holbrook in 1999 to
acquire data in the summer of 2000 on the Newfoundland margin directly conjugate to the ISE '97 data (Holbrook, 1998).

Despite this lack of conjugate data, there is an interesting LITHOPROBE seismic line that was collected over the southern Jeanne d'Arc basin, about 250 kilometers to the south of the conjugate location of Line 1. This line is shown in figure 7, and its location is highlighted along line 85-4 in figure 3. While not directly opposite to the Galicia Bank, this section can be thought to be representative of the structure found further to the north, because the western basin bounding fault (the Murre fault) extends northward towards the Orphan Basin. Behind Flemish Cap the Murre fault enters a transfer zone, where its displacement is transferred eastward to the synthetic Mercury fault. The Murre fault is shown to extend to 22-26 km depth, where it becomes a detachment surface. Interpretation of the seismic data suggest that the Murre fault is the reactivation of a pre-Mesozoic lineament (Tankard et al., 1989).

The general structure of the Murre fault and the Jeanne d'Arc basin is one of a northeast-southwest trending major basin-bounding extensional fault dipping to the east (figure 7), creating a basin into which Late Triassic through lower Aptian syn-rift sediments were shed (Tankard et al., 1989). These syn-rift sediments thicken to the west. One hundred seventy kilometers to the east, a smaller synthetic extensional fault bounds a second major basin. Above the syn-rift sediments is the wide-spread Avalon unconformity (Kay et al., 1990). Within the basin is a pronounced series of northwest-southeast trending, northeast dipping transfer faults. These features acted as strike-slip faults during the pre-Aptian extension, as Newfoundland and Iberia were separating, and
Figure 7. A trace of an interpreted seismic section collected across southern Jeanne d'Arc basin, eastern Canada, after Keen et al. (1987). Notice the eastward dip of the basin bounding normal faults, the relatively shallow position of the Moho, and the position of the Avalon unconformity which eroded upper syn-rift sediments. Location of this line and other 1985 Lithoprobe lines shown in inset. This line indicated in green. Labels in inset: Nf - Newfoundland, GB - Grand Banks, FC - Flemish Cap.
then changed to dip-slip displacement as the Orphan Basin opened to the north during separation of Newfoundland from Northern Europe (Tankard et al., 1989).

**Previous Structural Models and Analogs**

A number of models have been proposed to explain the formation of the $S$ reflector, the extensional basins above $S$, and the sediments deposited in the Galicia Bank area. These range from pure shear models, where the crust and underlying lithosphere extend together, leading to roughly symmetric distribution of rock types and structures, to simple shear models, where thinning in the crust is offset laterally from the thinning in the subcrustal lithosphere, and distant lithologies are exposed at the surface (c.f. Wernicke, 1985). Four different models for the tectonic evolution of Galicia are shown schematically in figure 8.

**Models**

Montadert et al. (1974) discussed the deposition of sedimentary units into basins along the Galicia margin, and linked the basin formation to opening of the North Atlantic during the Late Jurassic-Late Cretaceous. De Charpal et al. (1978) noted a deep, sub-horizontal reflector between 9 and 11 seconds TWTT under the Galicia Bank, and a similar event present beneath the northeastern Bay of Biscay. They called this reflector $S$, and postulated that it was either the boundary between lower velocity upper crustal material ($4.9 \text{ km s}^{-1}$) and faster, lower crustal material ($6.9 \text{ km s}^{-1}$), or it was the seismologic Moho.

De Charpal et al. (1978) were the first to describe a model for the formation of the structures seen at Galicia. They advanced a model for the formation of $S$ whereby the
upper crust thinned brittly while the lower crust thinned viscously. Due to the differences in maximum strain rate between these two different manners of deformation, they suggested that the lower crust thins more than the upper crust across the entire zone of rifting. This creates a system where the upper crust passively deforms on a ductile lower crust, also known as a 'pure-shear system' (figure 8(a)).

The pure shear model is a traditional model for how continental crust extends. In this model, normal faults create grabens in the upper crust (figure 9(b)). Below the brittle zone of the crust, ductile material is undergoing extension, and thinning. In this model of extension, vertical lines in the lithosphere remain vertical during deformation. Typically the failure of the upper crust takes place within the region of maximum thinning, and mantle material moves upward from below to fill the space left by the ductily thinned lower and middle crust. In the late 1970's and early 1980's this model was used to explain the structure of a wide variety of rift zones. However, there were mass balance problems in many areas when pure shear theory was applied, in particular the recognition that in some places in the Basin and Range Province a lower crustal footwall was exposed at the surface below a hanging wall of minimally extended upper crust. This mass balance problem caused Wernicke (1985) to advance the simple shear model.

The simple shear model uses a crustal scale detachment to thin the crust (figure 9(a)). This detachment surface cuts through the brittle layers of the crust and soles in either a ductile portion of the crust, or into the crust-mantle interface. Above the detachment the hanging wall is moderately deformed as it rides over the underlying footwall. In particular, slices of the hangingwall may be removed and be attached to the footwall in a manner akin to the rolling hinge models of Buck (1988) and Wernicke and
Axen (1988). The hangingwall in this scenario is typically less deformed than the footwall.

The simple shear model framework was applied to Galicia in two very different models using exactly the same data by Boillot et al. and Winterer et al. in 1988. Boillot et al. (1988b) postulated that S is a crustal scale detachment dipping to the east. This means that the half grabens preserved above S are remnants of the hangingwall, and that footwall structures should be seen on the conjugate margin. This model was advanced to explain the low amount of extension preserved in the upper crust compared to the extreme thinning of the crust as a whole. Winterer et al. (1988) also proposed a hypothesis for the formation of the Iberia margin based upon the simple shear model. Their model however speculates that the detachment surface dips to the west, and that Galicia is the footwall of the system, while the Newfoundland Margin is the hanging wall. Their main reason for suggesting a west dipping detachment involves the sediment supply to the deep Galicia margin during extension, and the source area for that sediment. Because the Galicia Interior Basin blocked European-derived continental sediment from entering the deep Galicia Basin to west of the Galicia Bank, and by making volume calculations of the sediment in the deep Galicia Basin, they suggested that the source area for the roughly 10,000 km³ of rift sediments is the Galicia Hills, which were isostatically uplifted as the footwall was unroofed during extension. In this hypothesis, the hangingwall would also be available as a source of sediment throughout the rifting sequence.

Other authors have weighed in on the debate regarding the formation of S and its implications. A number of papers have been written by Krawczyk and Reston
(Krawczyk and Reston, 1995; Krawczyk and Reston, 1996; Reston et al., 1995)
describing their analyses of data that existed prior to ISE '97. These papers typically
support a westward dipping simple shear model for the formation of S. Reston also
participated in the ISE '97 experiment, and one of his students, Marta Perez-Gussinye has
been working on processing portions of the data.

**Analogs**

At least one apparent exposed analog for the Galicia Margin has been found.
Studies by Froitzheim and Manatschal (1996) and Manatschal and Nievergelt (1997) and
Manatschal and Bernoulli (1999) indicate that the Austroalpine and Penninic units in the
southern Alps show a remarkable similarity to the model proposed by Winterer et al..
(1988) when the effects of Alpine contraction are removed. These nappes are the former
passive margins of the Piemont-Liguria ocean which trended northeast-southwest to the
southeast of the European continent and which began opening in the Liassic (Froitzheim
and Manatschal, 1996). Froitzheim and Manatschal (1996) show that the extensional
faults began forming proximal to the margin in late Hettangian-Sinemurian times. These
faults formed in the upper crust and Froitzheim and Manatschal assume that they flatten
with depth in the middle or lower crust. Distal to this, other normal faults are thought to
have been forming, gently thinning the crust over a distance of 300 km in a symmetric,
bulk pure shear system, as shown in figure 10(a). This thinning brought mantle material
to relatively shallow levels, where it cooled, halting the rheologic decoupling of the upper
crust and mantle at the lower crustal level which had enabled the bulk pure shear system.
This freezing of the system forced a throughputgoing lithospheric shear zone to form, which
accommodated a second phase of rifting along crustal scale faults (Froitzheim and
Manatschal, 1996).
This second phase of rifting occurred in the late Toarcian to Middle Jurassic. As
the mantle strengthened under the previously stretched region, a lithospheric scale fault
initiated along detachment surface 1 (figure 10(a)). Synthetic with this detachment were
two other surfaces, labeled 2 and 3 in figure 10(a). The mantle was exhumed along
detachments 2 and 3. Deformation in the opposing margin (Briançonnais domain) does
not appear to have been accommodated along similar large-scale detachment surfaces.
Instead, the Briançonnais area was isostatically uplifted as the footwall was removed from
beneath the hangingwall. At the same time, the narrow rift shoulder (Bernina) to the east
of detachment 1 was uplifted as the hangingwall moved off the footwall, and the footwall
was unloaded (Froitzheim and Manatschal, 1996).

An interpretation of the final geometry of the eastern Piemont-Liguria passive
margin shows blocks of upper crust are separated by low-angle faults (figure 10(c)). The
mantle lithosphere has been denuded, and serpentinite is now exposed at the surface.
Upper-crustal blocks have been stranded directly on this mantle lithosphere, and the
mantle is flexing in response to their load.

Manatschal and Bernoulli (1999) compared the Galicia and Adria Margins. They
noted the following similarities; 1) a wide zone of diffuse extension, 2) a narrow zone of
concentrated subsidence (and therefore, extension), 3) a low-angle feature within the
zone of concentrated extension which was interpreted to be a throughgoing lithospheric
scale extensional fault which evolved into a detachment surface, 4) a narrow shoulder on
the footwall adjacent to the detachment surface which experienced uplift during rifting,
5) a broader shoulder on the opposing margin which underwent gentle uplift, or lesser
subsidence, 6) slices of the upper crust in contact with the mantle across a detachment
surface (Manatschal and Bernoulli, 1999). The comparison of Galicia with the Austroalpine and Penninic units in the Southern Alps appears to be valid, and the understanding of the Austroalpine-Penninic area will be applied to a model presented below.
Figure 8. Four models advanced to explain the seismic features of the Iberia margin. In each of the models, S acts as a detachment surface. In (a), S overlies a zone of distributed pure shear, with the overriding fault blocks soling into S (de Charpal, et al, 1978). In (b), S is a detachment surface which exhumes deep crustal and mantle materials in the lower plate on the west side (Boillot, et al, 1988b). (c) is merely a mirror image of (b) (Winterer, et al, 1988). In (d), end-member models (a) and (b) have been combined (Sibuet, 1992). Figure after Reston et al., (1996), models after their respective authors.
Figure 9. Two end-member models for the rifting of continental crust and lithosphere. (A) shows the simple shear model, where a single through-going detachment surface serves to accommodate extension between the hanging-wall and the footwall. The pure shear model, where the crust and the lithosphere thin symmetrically is shown in (B). The yellow lines show the deformation of originally vertical lines in each model. (A) after Wernicke, 1985, (B) after McKenzie, 1978.
Figure 10. Reconstructed kinematic evolution of the Piemont-Liguria ocean and its passive margins based on field exposures in the Austroalpine and Upper Penninic nappes in eastern Switzerland and parts of Italy and Austria. In (A) early rifting was accommodated by pure-shear extension, forming horst and graben structures in the upper crust and necking of the lithosphere. In (B), extension is localized along detachment 1, allowing mantle to be exposed at the surface. At (C), extension is completed, and new oceanic crust is forming to the west of the ocean-continent transition. Abbreviations are: Br, Briançonnais; M, Margna nappe; S, Sella nappe; Be, Bernina rift shoulder. After Froitzheim and Manatschal (1996).
Chapter 2: The 1997 Iberia Seismic Experiment

The Iberia Seismic Experiment was conducted along North Atlantic margin, on and offshore of Spain and Portugal, in July and August of 1997. The main goals of the experiment were to determine (Sawyer, 1997):

1) The seismic velocity of the rocks that bound the S reflector where it is best expressed in the crust of the Galicia Bank.
2) Using structure and seismic velocity, the character of the eastward termination of S where it appears to split into three or more, lower amplitude reflectors.
3) Using structure and seismic velocity, the character of S where it terminates westward and identify its relationship, if any, with the Peridotite Ridge.
4) The seismic character of the Moho, if it is distinct from S, beneath the Galicia Bank.
5) The velocity depth function above S and compare it to that obtained by migration focusing analysis and use the result to determine the three-dimensional shape of S.
6) The structure and thickness of the crust under the Galicia Interior Basin.
7) How the structure and thickness of the crust changes along the Iberia margin from the Galicia Bank into the Iberia Abyssal Plain and how those changes relate to the location and nature of a transform segment boundary.
8) The nature and thickness of the unriifted continental crust landward of the Galicia Interior Basin.

Experiment Design

ISE '97 was conducted on the R/V Maurice Ewing, operated by the Lamont Doherty Earth Observatory, between 10 July and 15 August, 1997 on cruise Ewing 97-05 (EW97-05). The cruise began in Ponta Delgada, the Azores and finished in Lisbon, Portugal, with a mid cruise stop in Lisbon. Members of the science party on the cruise included personnel from Rice University, The University of Texas at Austin, GEOMAR
(Kiel, Germany), the Institute for Earth Sciences (Barcelona, Spain), the University of Aberdeen (Scotland, UK), the University of Aveiro (Portugal), and the University of Madrid (Spain). The experiment was funded by the U. S. National Science Foundation, the German National Science Foundation and the Spanish National Science Foundation. The Chief Scientist of the Iberia Seismic Experiment was Dale Sawyer, of Rice University.

Seismic, magnetic, and bathymetric data were collected during EW 97-05. The seismic source used for ISE '97 was the RV Ewing's 20 gun deep penetration airgun array, towed at a nominal depth of 6.5 meters. The total volume of this array was nominally 8385 cubic inches. Shooting was done at intervals of 20, 40 or 60 seconds during the experiment, depending on the primary purpose of the line. The 20 second shot repetitions were done when the MCS data were the primary goal, while 60 second shot repetitions were done while recording primarily by the OBS array. The MCS data were collected and digitized by the Ewing's 160 channel streamer, nominally towed at its full length of 4038 meters. At times the streamer was shortened; when inshore in shipping lanes it was judged too dangerous to have it extended to its full length. The receiver spacing on the streamer was 25 meters. Once acquired by the streamer, the seismic signals were digitized and recorded aboard the Ewing by a DSS-240 seismic recording system, which output data to 3480 cartridges tape in demultiplexed SEG-D format (Sawyer, 1997).

The data collected are of very high quality, with a relatively high signal to noise ratio. Due to the depth of the water much of the data were collected in, no multiple is present in the region of interest on some portions of the lines. On other lines, a very
strong water bottom multiple is present in the crust and upper mantle, exactly where we would like to see in order to build a coherent lithospheric structural model for the evolution of the margin.

**Thesis Goals**

The goals of this thesis are a subset of those for the entire Iberia Seismic Experiment. In particular, I initially set out to:

1) “determine, using structure and seismic velocity, the character of the eastward termination of S where it appears to split into three or more, lower amplitude reflectors,
2) “determine, using structure and seismic velocity, the character of S where it terminates westward and identify its relationship, if any, with the Peridotite Ridge,” (Sawyer, 1997)
3) Create a palinspastic restoration of the eastern rift, and a model of the rifting process,
4) Investigate the styles of deformation that occurred during rifting and how (if) the deformation processes changed through time and three dimensional (east-west, north-south, and vertical) space.

All of these goals were addressed in my research, and other tasks arose. For instance, after doing the initial interpretation and creating a deformational model, I realized that I needed to see below the water bottom multiple. This lead me to investigate processing techniques to remove water bottom multiples. In addition, as I worked toward creating a reconstructed cross-section, I realized that I needed to create a upper-lithosphere model of the extensional deformation across the Galicia Bank, rather than merely creating an upper-crustal palinspastic cross section.

There is one goal to this thesis — to advance a possible deformational history of the Galicia Bank, as it interrelates with the Galicia Margin and Flemish Cap, and to explain how this deformational history may integrate with the opening of the Atlantic and
the preexisting Hercynian structural trends. Such a model is presented in Chapter 6. The rest of this work explains the techniques and procedures that I followed in order to support and advance such a hypothesis.
Chapter 3: Seismic Data Processing Flows

A basic processing flow was designed by Dale Sawyer shortly after the end of the ISE '97 cruise, and nearly all of the seismic data collected aboard the R/V Ewing had been run through this preliminary set of processing flows by August, 1998. After this, the data were interpreted and the interpretations were used in further processing steps. One goal of this project was to work towards enhancing the processing sequence, using data from the interpretation to supply parameters to standard ProMax tools, and to see how the two procedures could be integrated.

Preliminary Data Flow

Original Processing Flow

Before the interpretation was started, the data were loaded into ProMax and experienced a basic processing sequence. This was done by Dale Sawyer and Brian Pietruszewski at Rice University.

To do this, they setup the line geometry, using the navigation and seismic metadata recorded aboard the R/V Ewing, edited (killed) channels to remove a majority of noisy traces from the data set, re-binned shot gathers into 12.5 meter CMP gathers, and performed a preliminary stacking velocity analysis on the CMP gathers. From this velocity analysis, Sawyer and Pietruszewski produced stacked sections and post-stack, f-k, constant velocity time migrations for most of the 30 lines collected. These f-k time migrations at 110% water velocity are the sections that were used for the interpretation portion of this project.

Problems with Original Flow
While this processing sequence does an excellent job of displaying the data to the first order, there are a number of problems with seismic sections produced using it. These problems include water bottom multiples in the shallow portions of the sections, migration "smiles" of overmigrated events, poor resolution of the Peridotite Ridge, and a uncertainty of the depth to various events such as R. Example of the water bottom multiple and Peridotite Ridge issues are shown in figure x011.

Along the shallower water portions of the Line 1, the water bottom multiple is present at around 9 to 10 seconds TWTT, (figure x011(a)). Unfortunately, this is near where the Moho is expected to be on this line, making it difficult to resolve any features along the crust-mantle interface on a stacked or migrated section. Also, the landward detachments Q and R discovered during the interpretation are obscured at depth by the water bottom multiple, making their geometries difficult to discern. Both of these issues with the multiple are further exacerbated by the water depth in which the data was collected. Because of the nearly 4 kilometers of water overlying the seafloor, the water bottom multiple arrives late in the section, where the reflected primary energy is already at a minimum. Because the impulse which creates the water bottom multiple has traveled between two very reflective surfaces (the seafloor and the sea surface, both surfaces of very high impedance), it naturally has a very strong amplitude compared to the primary data. This means that the water bottom multiple signal swamps much of the primary energy in all the gathers, across all the channels of the gather. This makes attenuating the multiple, and recovering the primary energy within it, a difficult task.

On the eastern ends the dip parallel seismic lines, the contact between the Peridotite Ridge and the overlying sediments on the west side of the Peridotite Ridge is
difficult to pick (figure x011(b)). However, the geometry of this contact is important in deducing the formation of the Peridotite Ridge. Similarly, the internal structure of the Peridotite Ridge is poorly imaged. It is difficult to determine the orientation of reflection events within the Peridotite Ridge. In some places, these events appear as continuous reflectors, bowed upwards under the apex of the Peridotite Ridge. Elsewhere, reflectors appear to initiate at the Peridotite Ridge apex and fan downwards. A supportable hypothesis for the formation of the Peridotite Ridge would be greatly strengthened by resolution of the internal reflector geometry.

Because of these issues found in the preliminarily processed data, a number of more advance processing operations were applied to the data. The primary purpose of these processing steps was to remove the water bottom multiple from the data, though data improvement in other problem areas was hoped for.

**WATER BOTTOM MULTIPLE REDUCTION PROCESSING**

**Methodology of Application**

A number of processing methodologies were applied to these data in an attempt to remove the water bottom multiple from the data. In order to test the parameters required for each of these methods, short sections of Line 1 were processed with different parameter values for each run. The results were compared between runs, and parameters were re-chosen to maximize the improvement of the data.

Typically, particularly problematic areas were selected for this testing, in the hope that an improvement in such an area would provide the most gain. In general, this assumption worked well. However, in some case, different parameters were needed in
different areas due to differences in water depth, sedimentary thickness and crustal
structure.

**Water Bottom Multiple Suppression Techniques**

Four different water bottom multiple suppression techniques were attempted on
Line 1. These techniques were inside muting, f-k filtering, radon filtering, and stack
weighting.

**Inside Mute**

The inside mute technique involves removing the near-offset traces from the CMP
gather (figure x012). This relies upon the re-positioning of the traces during normal
moveout (NMO) to limit the water bottom multiple. The theoretical shape of a travel-time
curve in the CMP domain for a horizontal surface is a hyperbola, of the form,

\[ t_r^2(x) = t_r^2(0) + \frac{x^2}{v^2} \]  

(equation 4.1, (Yilmaz, 1988), p. 157)

where \( t(x) \) is the TWTT along a raypath for shot location to receiver at offset distance \( x \), \( v \)
is the velocity of the medium, and \( t(0) \) is twice the zero (or near) offset travel time.

Equation 4.1 describes normal moveout (NMO). NMO is the process whereby records at
\( t(x) \) are corrected for the velocity of the medium. By flattening the hyperbolic reflections
in the CMP gather, we can determine a stacking velocity \( v \).

A side effect of the NMO correction is that primary events should be flattened by
NMO, while multiples should arrive earlier at longer offset. Because of this, the
primaries should add constructively during stacking, while multiple arrivals ideally will
add deconstructively when stacked. However, because traces at near offsets are
minimally shifted by the NMO operation, the near offset multiples contribute strongly to
the stack. The purpose of the inside mute multiple reduction technique is to remove these strong signals from the gather.

The inside mute suppression technique worked moderately well in areas where the seafloor was horizontal, or nearly horizontal. This occurred because of the assumption in the NMO equation that reflectors are horizontal, and do not dip significantly. When the seafloor dipped steeply (as it does on the seaward side of the Galicia Bank, rising from the Deep Galicia Basin), this assumption was no longer valid, and significant multiple energy remained in the gather.

**f-k filtering**

F-k filtering also operates on CMP gathers, however, rather than operating in the $t$-$x$ domain, it operates in the frequency ($f$) - wavenumber ($k$) domain. Data are reversibly mapped between $t$-$x$ and $f$-$k$ domains via the 2-D Fourier transform.

To accomplish the f-k filtering, the data were moved out at a velocity lower than the stacking velocity (65% of stacking). This mapped the primaries into concave upward features in f-k space, while the multiples were convex. This simple separation of the primaries from the secondaries allowed easy f-k dip filtering. In ProMax, the f-k filtering was performed by defining regions (polygons) of data which should be rejected from the CMP gather. All values with negative $k$ were rejected during f-k filtering. This operation removed a significant portion of the water bottom multiple event from the data in many of the gathers (figure x013). However, portions of the data still showed a very strong water bottom multiple signal. This tends to happen where the multiple dips steeply on the flanks of the Galicia Bank, or where reflectors within the basins dip steeply.
Figure 11. Two examples of processing related issues in the data on Line 1. In a), a stacked section with AGC, the water bottom multiple is highlighted in the blue shaded area. Notice that it reaches 9 seconds TWTT in this area. Shallower events (indicated by the dashed purple lines) enter the multiple area and are lost. In b), a migrated section showing the Peridotite Ridge, the contact between the peridotite and the overlying sediments (indicated by the green line) along the eastern flank of the ridge is difficult to discern. Also, the internal structure of the ridge is unclear (shown by the dashed purple lines) - are the internal reflectors continuous, and bowed upwards under the Peridotite Ridge; or, are they discontinuous, and feather downwards from the apex of the ridge?
Alone, the $f-k$ technique did not improve the data enough to perform a clean migration. In places, the amplitude of the water bottom multiple could be filtered to below the amplitude of the signal. Elsewhere, the water bottom multiple amplitude could not be reduced below the signal level, thus migration created a large number of migration errors, contaminating the section, and making interpretation difficult.

**Radon Filter**

Another water bottom multiple reduction technique applied to Line 1 was a Radon Filter. This technique transforms gathers into time-moveout space, where events with the same moveout velocity condense to a point at their zero offset time. This allows discrimination of events based on velocity and time.

In order to perform Radon filtering on these data, they were moved out at the stacking velocity. This flattened primary events in the t-x domain. When transformed into the radon domain, the flattened primary events mapped at or near zero in residual moveout versus time space. However, when moved out at stacking velocities, multiples should have a positive residual moveout (i.e., they should be concave downward). By applying a radon filter that removes all events with positive residual moveout (figure x015), these slower events can be removed from the data.

A Radon filtered gather is shown in figure x014. Note the reduction in amplitude of the water bottom multiple between 9 and 10 seconds TWTT. Also note that more noise was removed from the far offset traces than the near offset traces. Radon filtering worked very well in some locations, and poorly in others. Typically it worked poorly in
areas where the strata were dipping, which caused the primaries and secondaries to have non-predicable residual moveout.

**Stack Weighting**

Weighted stacking works to remove water bottom multiple noise from CMP gathers during the stacking process. The strongest and most difficult to remove noise from the water bottom multiple is on the near offset traces. If, during stacking, the long offset traces are multiplied by a weighting factor, then they can contribute more to the amplitude of a stacked bin. Stack weighting works much the same way as the inside mute technique, but it works during the stacking operation, while inside muting works pre-stack.

This technique worked moderately well on these data, but once again, did not reduce the noise to below the signal level. Because of this, stack weighting was not useful for creating migrated sections.

**Combinations of Suppression Techniques**

A number of the techniques were tried together in various permutations. The best reduction was achieved using a combination of the $f$-$k$ filtering, radon filtering and weighted stacking. However, not enough noise was removed from these data to provide any more information than was already available.

**Evaluation of Water Bottom Multiple Suppression Techniques**

None of these water bottom multiple reduction techniques eliminated enough of the multiple to be useful in producing stacked sections for time migration. Attempts to perform Stolt or Kirchoff time migration using both a constant velocity and a percentage of stacking velocities performed poorly due to the high amplitude noise still remaining in the stacked sections after water bottom multiple reduction processing.
The difficulty in removing the water bottom multiple from these data is not surprising, in retrospect. The location of the multiple in the areas of interest on Line 1 was typically 3 to 6 seconds below the seafloor (6 - 12 TWTT). The low signal to noise ratio at these depths require reducing the amplitude of the multiple by a factor of nearly 20 in order to reach a signal to noise ratio of 1. It was not possible to reduce the multiple below this level across the entire section.

The structure of the area also created a number of difficulties in the water bottom multiple reduction processing. These techniques depend on an expected shape of a moveout curve and the discrimination of primary energy from secondary energy based on the shape of this curve. Because of this, dipping beds and structural complex regions are difficult to use these techniques on, since such areas do not fit the layer-flat earth model assumed by these techniques. Therefore, often the actual reflection gather does not fit the predicted gather, and the technique fails.

A final difficulty in these data were the stacking velocities used. Because of the short streamer available on the Ewing, accurate velocities were difficult to obtain. The difficulty increased with depth and velocity, as the primary events become nearly flat. This makes picking NMO velocities, which are essential for effective use of these techniques, very difficult.

Possibilities for Further Data Improvement

A number of improvements can be made to the data at this point. Re-picking stacking velocities on the water bottom multiple reduced sections would contribute to a stronger stack and cleaner migrations of these data. Also, a deconvolution operator should be applied to these data in future processing steps. This deconvolution should contribute to removing the remaining multiple, and also help to sharpen the section by
removing the reverberations under $S$ and other strong reflectors. Finally, these data could be greatly enhanced with the proper application of a depth migration processing sequence.
Figure 12. This is an example of how an inside mute is applied to traces in the CMP domain for water bottom multiple reduction. The inside mute (shown by the red line) is applied to the near offset traces, just above the water bottom multiple, before NMO. Once NMO is applied, the multiple curves upward in the CMP gather, and ideally, does not stack constructively. The purpose of the inside mute is to remove the traces which are not strongly modified by the NMO function, and which therefore contribute the majority of the energy to the stack. Both gathers have the same 1/dist spherical divergence correction applied.
Figure 13. An example of the effect of filtering in frequency-wavenumber space on the CMP gather. Notice that the water bottom multiple has been somewhat attenuated in the f-k filtered data, and the data appear to be cleaner. Both gathers have the same 1/dist spherical divergence correction applied.
Figure 14. Thie is an example of how a Radon mute has modified the data during water bottom multiple reduction. The Radon gate used for this processing step is show in figure 15. As can be seen from these data, the Radon transform drastically reduces the amplitude of the water bottom multiple present in the gather, though it does seem to create some artifacts. Despite these artifacts, the removal of the multiple drastically improves the usability of these data. Both gathers have the same 1/dist spherical divergence correction applied.
Figure 15. The red line is the bottom mute applied during the Radon muting to eliminate the water bottom multiple. This mute removes the majority of the energy which has a positive residual moveout. When energy has a positive residual moveout after NMO, it means that the events required slower NMO velocities to stack properly. This means that events that curve upward in the CMP domain are on the positive side of this diagram, while those events that curve downwards in NMO space are on the negative side. Since multiples are NMOed at velocities faster than their multiple stacking velocity, multiples typically curve upward in the CMP domain. The mute shown removes all energy in the t/p domain which has a residual moveout of more than 45.
Chapter 4: Seismic Data Interpretation

INTERPRETATION PROCESS

The data collected offshore Iberia in the summer of 1997 were run through an initial processing sequence which involved gathering the data, stacking it and Stolt migrating it at 1650 m/s. Once this was complete, the data were loaded into an electronic seismic interpretation system. This system was composed of a Dell Pentium II CPU, with 256 megabytes (MB) of random access memory (RAM), running Microsoft® Windows NT™ Server Version 4.0 (Service Packs 3-5) and was donated to Rice University by Intel Corporation.

Seismic Micro-Technology’s (SMT) KINGDOM Suite, Version 4 was used to interpret the data. Seismic Micro-Technology donated this software to the Rice University Department of Geology and Geophysics, and in particular to John Anderson’s research group. The KINGDOM suite acts much like any other digital interpretation system which affords an interpreter the speed and flexibility of automated horizon picking, x-z storage of horizon points and zoom of seismic data. All of these features were used in the interpretation of the data, however, none of the more advanced map making or horizon contouring were used in this project.

STRATIGRAPHIC PICKS

It is important to pick the same events as the same horizons throughout the seismic project area. This is aided by having a set of guidelines for picking each event
and a methodology of picking horizons. For the seismic lines in the S reflector area of ISE '97, the guidelines used are discussed here.

**Seafloor**

The seafloor was picked at the first positive excursion of the trace, continuous over at least 20 traces. In general, this was a very easy pick to make, and automated tools could pick very nearly the entire surface. There are reverberations present in the data above the seafloor, due either to the impulse response of the airgun array or artifacts caused early in the processing sequence, however, these are much lower amplitude (generally >20% of the water bottom), and are cut by diffraction parabolas in the migrated section that are rooted at the first true seafloor reflection. The seafloor picks vary very little in time trace-to-trace, and range in value from 3.18 to 7.06 seconds TWTT.

**Post-Rift sediments**

In general, just below the seafloor horizon are almost flat lying, nearly conformable reflectors, dipping gently to the west (into the North Atlantic ocean basin), if at all. These reflectors are generally finely bedded with little to no disruption in continuity. These reflectors typically are concordant with the seafloor, though in places the reflectors diverge from being sub parallel to the seafloor and show localized onlap onto basement highs (as indicated by (e) in figure 16 and 17), as well as internal onlap/toplap relationships. This package ranges from nearly 0 to more than 1.5 seconds thick in the thickest parts. The top of these sediments is the seafloor (3.17 to 7.06 sec TWTT), while the base ranges from 3.57 to 8.79 TWTT.

These reflectors are interpreted to represent sediments which were deposited after the onset of drifting, and are thickest within the sedimentary basin, and thinnest over
basement highs. Because the basin-bounding faults (discussed in Section III below) do not cut these packages, it is evident that the basin-bounding faults were not active at the time of deposition of these sediments. Since the basin bounding faults accommodated the upper-crustal extension (see Section III), these sedimentary packages are interpreted to be post-rift sediments. These packages are the same as Acoustic Units 1, 2 & 3 of Sibuet (1987) and Mauffret (1988) (figure 6).

**Syn-Rift Sediments**

Below the slightly dipping, sub-parallel concordant reflectors of the post-rift sediments lies a package of reflectors that are generally divergent in character. These are shown as (d) in figures 16 and 17. The thickness of this package varies greatly along the lines, ranging from 0 to nearly 1 second. The upper surface of this package is the lower surface of the post-rift sediments, and ranges from 3.57 to 8.79 seconds TWTT, while the base of the package ranges from 6.87 to 9.37 TWTT. The internal geometry of this package shows concordant to hummocky fanning reflectors, with downlap towards the east. The reflectors comprising this package are often tilted and terminate against the bounding faults of the half grabens.

The significant asymmetry shown by this package and the onlap of reflectors onto the top of pre-rift blocks and downlap onto pre-rift blocks is interpreted to be caused by its deposition into half-grabens during the active rifting cycle, hence the label of syn-rift sediments. As rifting progressed, basins formed, and then deepened, and the sediments deposited into those basins thickened toward the deeper extents. As further rifting occurred, the units in the basins rotated along the half graben bounding fault, and more sediment was deposited. This created fan shaped wedges in the basins (figures 16 and
17). The units classified as syn-rift sediments are the same as those named Acoustic Unit 4 by Sibuet (1987) and Mauffret (1988).

**Pre-Rift Sediments**

Directly below the syn-rift sediment package lies a set of reflectors that are generally sub-parallel to hummocky and non-contiguous. The reflectors in this package nearly always dip to the east, and are often marked by a high amplitude event at their top, which is nearly conformable with the base of the overlying syn-rift sediments. The upper surface of this package coincides with the base of the syn-rift package and ranges in TTWT from 3.60 to 6.87 seconds, while the base of the package is between 3.60 and 9.79 seconds of TWTT. This unit is shown as (c) in figures 16 and 17.

Because of the internal reflector geometry of this package, the discontinuities that bound it, and its covering sediments (i.e. the fanning, divergent reflectors in the overlying unit), it is interpreted to be the pre-existing sediment cover of the crystalline basement which was at the surface when rifting initiated, or from a time of the earlier extension. Evidence from some recovered ODP cores indicates that the high amplitude reflector at the top of the sequence is a Tithonian carbonate layer (Shipboard Scientific Party, 1987d). The hummocky to sub-parallel reflectors below the high amplitude event give the impression that this unit is composed of a thin layer of terrestrial sedimentary rocks (siliclastics, volcanioclastics and metamorphic sedimentary rocks), a supposition supported by drilling and sampling by manned submersible (Boillot et al., 1989; Shipboard Scientific Party, 1987a, d).

This sequence is composed of Acoustic Units 5A and 5B (Mauffret and Montadert, 1988). It is thought that these carbonate reefs formed in the earliest Jurassic, in shallow water during the initial onset of extension. As extension progressed towards
riding, the water depth in the basins increased, forcing the carbonate reefs to localize on
the top of the rift blocks. The extension thus segmented and dissected the carbonate
platform into isolated patch reefs and bioherms. Ultimately, as extension progressed
towards riding, water depths continued to increase, drowning these carbonate banks
(Boillot et al., 1988c).

**Basement**

The interface between the basement and the pre-rift sediments is often difficult to
discern. While the basement top is generally marked by a high-amplitude event on the
eastern ends of the analyzed lines, on the western ends there often is less amplitude
difference, and in places, the reflection section becomes “transparent” in this area, with
few to no continuous reflections visible. The upper surface of this unit occurs between
3.88 and 9.79 seconds TWTT. The reflections in this unit range from sub-parallel to
hummocky in character and are continuous over 20 to 200 traces, as shown by (b) in
figure 16. The reflections dip to the east and occasionally are erosional or sigmoidal at
the uppermost surface of the unit. Continuous reflector packages dipping to the east are
often cut by other reflectors dipping more steeply to the west (figure 16(a)). These west
dipping features are interpreted to be faults, and are discussed below.

This unit has a “wormier” appearance than the other units described. This is
probably due to the typically non-continuous nature of crystalline basement material,
which is thought to comprise this unit (Boillot et al., 1988a; Shipboard Scientific Party,
1987d). Or, this may be due to multiple stages of faulting during the contractional and
extensional history of the disrupting the coherence of previously parallel reflectors.
Figure 16. A typical seismic section along Line 4. (A) shows uninterpreted seismic data between CMPs 1400 and 2600 (B) shows primary features present in the data, indicated by a-e. These feature are recognized as follows: (a) a high amplitude, wavy event oblique to reflectors such as (b) and (c) which terminate against (a). Events (b) are hummocky, dip to the east and terminate against (a). Overlying (b) are sub-parallel reflectors as in (c), above which events (d) fan. Events indicated by (d) typically diverge towards the east, and terminate against (a), while being overlain by sub-horizontal to horizontal reflectors like (e).
Figure 17. Interpreted features of a typical seismic section along Line 1. (A) shows interpreted seismic data between CMPs 4350 and 5500. (B) shows primary features present in the data, indicated by a-e. These features are interpreted as follows: (a) basin bounding normal fault, which forms in the crystalline continental crust (b) and overlying sedimentary veneer (c). Motion on these faults during extension formed basins into which syn-rift sediments (d) were deposited. These syn-rift sediments often show a fanning, or basinward thickening. Above the syn-rift sediments are relatively flat-lying sediments (e), interpreted to have been deposited after the cessation of extension. Folding and faulting in these units is probably related to differential compaction.
The picks described above were generally done on dip lines, because the syn-rift package made an excellent marker for the interpretation of the pre- and post-rift packages. In the strike lines, the surfaces are more difficult to pick. However, when the horizons from the dip lines are plotted where they intersect the strike lines, the interpretation becomes simpler. In general, the units on the strike lines have similar seismic characteristics compared to the dip lines. However, the fanning geometry in the syn-rift package is much less pronounced, and the reflectors do not have the pronounced east-ward dip that they do on the dip lines (compare figures 18(a), 19(a), with 20(a)).

**Structural Picks**

**Block bounding faults**

The majority of the syn-rift reflectors terminate to the east at westward dipping events (indicated in figures 16 and 17 above, with (a)). Typically, these reflectors are composed of a negative-positive-negative amplitude response, and truncate eastward dipping sedimentary reflectors.

These reflectors extend from the seafloor horizon into the basement units, and are typically 800 to 4000 meters apart on the dip lines. They are generally listric on a time section (figures 18, 19, and 20), but only slightly listric on a depth converted seismic section (figures 21, 22, and 23). They occasionally bifurcate upward into two or more sub-parallel reflectors, while maintaining a similar dip. They always form the boundary towards which the syn-rift sediment packages thicken, and surfaces are generally interpreted to have normal displacement across them.
Because of the above features, these reflectors are interpreted as basin bounding normal faults, formed during extension and rifting of the Galicia margin (Sibuet et al., 1987). They were picked just below the first high amplitude positive-negative phase transition (black to white) in the negative trough (white).

These faults generally control the distribution and geometries of the overlying syn-rift sediment packages, and as such, are generally easy to localize in the upper portions of the section, as reflectors of drastically different dip or character are juxtaposed along a narrow interface (figures 16 and 17). In the deeper portions of the section (8 to 9 seconds and deeper) within the basement, they are much more difficult to pick. This is because of a lack of continuous, sub-parallel reflectors to judge offset and discontinuity against. Since the basement material generally has a broken, discordant, hummocky reflection character, any recognition of a fault needs to be based on impedance contrasts across the fault and the relative continuity of the faults, a much more difficult proposition than searching for sub-linear reflection discontinuities.

**Basal/Master Faults**

On the western ends of the dip lines, the basin-bounding faults meet a low angle surface within the basement, below 8 to 9 seconds. In the time section, this surface is gently mounded, with a concave downward trend (figures 18 and 19), and a slight dip towards the west. In depth sections, this surface is flat to concave downward, and dips both seaward and landward (figures 21 and 22). Typically picked just below the first high amplitude positive-negative phase transition (black to white) in the negative trough (white), this surface is synthetic with the overlying faults.

Similarly, to the east, there are two other reflectors into which some of the overlying basin-bounding faults sole. These are visible to 9 and 10 seconds TWTT, and
are some of the brightest reflectors in the deep portions of the section. Because they are linked with the overlying basin-bounding faults, and because of their gentle dip, the seaward most of these features is interpreted as a detachment surface by many authors (Boillot et al., 1988b; Krawczyk and Reston, 1995; Krawczyk and Reston, 1996; Reston et al., 1995; Sibuet, 1992; Winterer et al., 1988). This deep reflection which was interpreted to be a detachment surface by some authors is named the S reflector, or simply S (de Charpal et al., 1978). The two more landward deep reflectors have not been noted prior to this study. The labels Q and R, landward and seaward, respectively are suggested and used in this paper.
Chapter 5: Discussion

The Galicia Margin formed beginning in the Triassic and continuing through the Cretaceous, as the Atlantic Ocean opened. The structures resulting from this extension are preserved in the Galicia Interior Basin, while the structures that formed during the final rifting are recorded further to the west in the Deep Galicia Basin.

The final rifting of Pangea between the MGFZ zone to the south, and the Bay of Biscay to the north is preserved beneath the Deep Galicia Basin. The normal faults formed during this rifting are preserved as basin-bounding faults. These faults controlled the basin evolution and therefore the deposition of the sediments within the basins. These faults typically extend 2-4 seconds below the seafloor, and sole into deeper detachment surfaces. One of these surfaces is S, which was previously interpreted as an extensional detachment surface (Boillot et al., 1988b; de Charpal et al., 1978; Krawczyk and Reston, 1995; Sibuet, 1989).

De Charpal et al. (1978) first noted the existence of a sub-horizontal reflector between 9 to 11 seconds TWTT below the northern Bay of Biscay passive margin, and a similar feature below the extended crust to the west of Galicia Bank. They named the reflector beneath the Deep Galicia Basin S. As discussed by de Charpal et al. (1978), S beneath the Deep Galicia Basin was interpreted to be a detachment surface or an interface where a sharp change in velocity occurred (perhaps the Moho). In neither interpretation, did de Charpal et al. (1978) suggest that the basin bounding upper-crustal normal faults extend to S. Boillot et al. (1988b) interpreted S to be only a detachment surface, and to dip to the east. In their hypothesis, the material below S is the footwall, while the
material above S is the non-dismembered, coherent portions of the hangingwall, which moved to the east over the footwall. Winterer, et al. (1988) suggested that the sense of motion was opposite: that the material above S moved to the west, and the crustal material that remained above S was the dregs peeled off of the hangingwall as it moved to the west in a rolling hinge-type model (c.f. Wernicke, 1988; Buck, 1988). Krawczyk and Reston (1995) re-processed a number of the lines collected earlier by the French, and determined that the acoustic signature of S indicates that it arises from a single interface. This means that S is not a sub-horizontal volcanic sill. Instead, it strongly suggests that S is in fact a velocity discontinuity. Krawczyk and Reston (1995) also suggested that motion along S was top to the west, and that S acted as a detachment surface during deformation (figure 8 shows sketches of each of the models). All of these interpretations were limited by lack of data: collecting high-quality data was one of the main purposes of ISE '97.

**INTERPRETATION OF LINES 1 AND 4**

The reflection data collected during ISE '97 (figures 18, 19, and 20) begin to fill some of these data holes, enabling creation of a more complete model. In figures 18 and 19, it is clearly evident that the upper crustal normal faults penetrate to S, where they terminate. Long offset refraction data inversion (figure 25) shows that there is a sharp increase in velocity below S. These data suggest that material with mantle velocities does lie beneath S (Zelt et al., 1998; Zelt et al., 1999) and that S may in fact be the Moho. This indicates that S is a detachment surface where the upper crust has been detached from the lower crust and sits directly on a mantle footwall. S arrives closest to the surface near CMP 6060 on Line 1 and cross-cuts the pre-rift sediments near CMP 4690.
on Line 4. At least 7 secondary faults are seen above S on Line 4, while on Line 1 no more than 4 are obvious. These faults sole into S and bound supra-crustal sedimentary basins. In addition, the sub-horizontal geometry of the top of the pre-rift crust between CMPs 3300 and 3950 on Line 1 differs from the large east dipping rift basin geometry along the rest of the line. This may be suggests that that the top of the pre-rift unit was disrupted by a number of down-to-the-west normal faults with relatively small throw (below seismic resolution), effectively modifying the top of the pre-rift unit by structural simple shear.

This study discovered two additional detachment surfaces to the landward of S. These structures were named R and Q (Unger et al., 1999) (figures 18 and 19). Both Q and R show normal sense of motion, dipping to the west, with the footwall up to the east. Q dips at 19-27° on depth converted sections, decreasing with depth (figures 21 and 22). Dips on R range from 35° to the west near the surface to 4° to the east at 11 km depth where R flattens and becomes convex upward. Secondary normal faults sole into R. The fault bounded blocks above Q and R range in size (in dip section) between 3 and 10 km wide and have an unknown length north-south. Surface R forms a prominent fault scarp more than 1 second high at CMP 5000 on Line 4, and bounds a smaller hillock near CMP 7670 on Line 1.

Q and R act as master faults controlling the displacement and location of several minor faults within the Deep Galicia Basin. Q is characterized more by a series of interacting crustal scale normal faults, extending to 9 seconds TWTT, though it does flatten and appears to become listric between 7 and 9 seconds TWTT. R is a detachment
into which at least one normal fault soles (figure 19), although perhaps as many as four smaller normal faults terminate at R as well on Line 1.

The geometry of R is very different from that of Q. R can be seen to reach to the interpreted Moho in figure 19, and can be seen to penetrate through the Moho into the mantle in figures 18 and 20. Where Q is somewhat convex upward in dip time sections due to velocity pull-up, R is nearly concave upward, with minor velocity pull-up effects. The dip of R may have been modified after fault initiation by domino-tilting of crustal blocks and their mantle bases, or by deeper material moving upwards and/or laterally. This might have occurred as deep, mantle material moved to areas where overlying crust was thinner.

The geometry of S suggests a very different series of events in its formation. At S upper-crustal material of the hangingwall lies directly on mantle material comprising the footwall. On Line 1, S is convex upward, centered near CMP 4600 (figure 22). To the west, S is a structural contact between mantle and crustal material along a detachment fault, while to the east of CMP 4600 S transitions into a more diffuse eastward dipping boundary. This boundary may be structural feature that formed as mantle material moved westward and upward during and after rifting. The convex upward shape of S on Line 4 suggests that the mantle may have expanded during serpentinization by water penetrating along hanging wall faults. This expansion would cause the interface above the mantle to be deformed upward, as seen on Lines 1 and 4.

Above all of these deep penetrating faults (Q, R and S), the rocks of the hangingwall were modified during and after extension. Material was eroded from high points and deposited in low-lying areas. The distribution of this erosion and
sedimentation is well displayed in the seismic lines collected by ISE '97 and discussed in this paper (figures 18, 19, 20, 21, 22, and 23).

The thickness and geometry of the pre-rift sediments layer tells much about the temporal evolution of this area of the Deep Galicia Basin. By inspecting the thickness of the material interpreted as being pre-rift, and assuming that it was a constant thickness across the area, we can determine where erosion and deposition occurred (and therefore identify paleo-topographic highs and lows). By documenting how unit thickness changes throughout a region, it should also be possible to determine the relative ages of various erosion and deposition.

Measuring the thickness of the pre-rift materials on Line 4 (figure 21) shows that they are at most 1.5 km thick in the basin above Q near CMP 5600. To the west, this unit is a similar thickness in the basin at CMP 3800. However, in the hangingwall block above R this unit decreases in thickness to around 0 m. This indicates that the upper crustal portion of the S and Q faults formed before R, which caused the hanging wall of R to be a topographic high. Material on this high was eroded, thinning the unit by 1500 meters. This material was probably deposited in the nearby basins to the east and west of this paleotopographic high. Later, motion on fault R formed the basin above R, but because it formed later, less syn-rift material was deposited into his basin. Along Line 4 (figure 21), 750 m of syn-rift sedimentary rock exist above R (CMP 4300) compared to 1.75 km above Q (CMP 5800).

**ALONG STRIKE DEFORMATION**

Interpreting the data collected along strike lines (figure 20) is much more difficult than interpreting the data collected on the dip lines. Because Lines 5 and 6 run parallel to
strike of the margin, it is extremely difficult to distinguish the acoustic units based on seismic character alone. Where the strike lines cross the dip lines, the interpretation can be transferred from the dip lines, providing a pick to continue the interpretation on. In order to accurately interpret these strike lines, dip line picks need to be integrated with the velocity data to distinguish between crustal and mantle materials. Because of the short length of the seismic streamer used in collecting these data (4 km), and the depth of the water along strike line 5 (4500 meters), the moveout on the CMP gathers is insufficient to accurately pick velocities for stacking, migrating and interpretation. Despite this, these stacking velocities converted to interval velocity were used for the tentative interpretation of Line 5 (figure 20).

This interpretation shows the post-rift sediments in a mostly conformable contact with the underlying pre-rift sediments. The post-rift sediments do onlap older material on the southern end, near CMP 6600. The syn-rift sediments fill two wide depressions, each 30 –35 km in extent, one on the south end, the other on the north end of the line. Both the post-rift and the pre-rift units have a gentle (1°) slope to the south, and are slightly thicker on the southern end of the line (figure 23). Below the syn-rift unit is where the interpretation difficulties begin. The internal structure of the pre-rift and crust material is not distinct enough to permit infallible interpretation. Therefore, this line was interpreted using stacking velocities converted to interval velocities within ProMax as well as the structural and stratigraphic character of the data.

The structural interpretation of Line 5 is similarly difficult. The structural interpretation relies on the stratigraphic interpretation for validation, therefore difficulties in interpretation of the stratigraphic units lead to uncertainties in the structural
interpretation. The data from Line 5 (figures 20 and 23) appear to show a series of north
dipping shear zones that originate as sub-horizontal detachment surfaces. There is a
major interface between 8 and 9 seconds which serves to juxtapose pre-rift and crustal
hangingwall materials onto mantle footwall material. This is the detachment S. The
north dipping shear zones which arise from this interface accommodate normal-sense
displacement, with the hangingwall down to the north. The majority of the displacement
of the pre-rift and material was of course to the west, but these faults show that there was
north-south extension occurring as well.

Above the S detachment at 8 to 9 seconds (8 to 9 km) lies crustal and pre-rift
material. These units dip to the southeast, and are broken by north dipping secondary
normal faults. The units were displaced to the north along the detachment surface during
deformation of Galicia. The majority of displacement occurred along S, and the
overlying faults served to accommodate local strain. This caused the broken character of
the pre-rift unit, while the complete absence of crustal material from portions of the
section is due to movement along the detachment surface.

Line 5 suggests how the detachment surfaces recognized in the dip lines interact
in three dimensions. S is visible as the detachment surface along which the crust is
dismembered. Comparison of the intersections of the strike and dip lines shows that the
surface named R on Line 4 is a different surface from R on Line 1. Because of this, they
are indicated as R1 and R2 on Line 5 (figure 20(b) and 23(b)).

WIDE ANGLE DATA

Long offset data were also collected along a number of the seismic lines during
the ISE '97 experiment using ocean-bottom seismographs and hydrophones to collect
wide-angle data along the entire length of Lines 1, 12 and 9, as well as in a more tightly-spaced area over the S reflector (Sawyer, 1997) (figure 1). The seismic velocities determined from the wide-angle data (figure 25) confirm the position of the Moho, provide test of the deep structural picks of Q and R, and also supply a velocity model that aided in the depth conversion of these data. These data can then be transferred to the reflection sections to aid in interpretation of the reflection data. This is important because the velocity resolution that can be determined from the reflection gathers is severely limited by the short offset of the streamer used to collect the data.

These wide-angle data were presented by Zelt (1999). They show the crust to the east of the Galicia Interior Basin to be more than 20 km thick, and a thinning of the crust in the Galicia Interior Basin to 9 km. Westward of the Galicia Interior Basin, the crust thickens slightly, reaching nearly 17 km thick under the Galicia Bank, before thinning drastically above the S reflector. These data also show that S is an interface between material with crustal seismic velocities and material with mantle velocities, effectively making S the Moho under the Deep Galicia Basin.

**Beta/Gamma vaules from Wide-angle Data**

The models created using the wide-angle seismic data were used to calculate the stretching factors beta and γ (gamma) for the Galicia margin. The extension factor γ is a measure of the change in cross-sectional area of a region. It is useful for quantifying crustal-scale two dimensional strain. By determining a crust-mantle interface depth from the wide-angle model of Zelt et al. (1999), it is possible to plot gamma versus distance. Gamma was calculated in two different ways, the first assuming an initial crustal
thickness of 30 km, based on the presumed crustal thickness prior to extension of Rast (1989), while the second was derived from the total tectonic subsidence.

The first method is shown in equation 5.1. In this case, gamma is determined from the stretching value beta, which is the ratio of original crustal thickness ($Z_o$) divided by extended crustal thickness ($Z$) (equation 5.2). Beta will always be greater than 1, and gamma will always be less than 1. As crustal thickness goes to 0 and therefore extension increases, beta goes to infinity.

$$\gamma_{\text{crustal thickness}} = 1 - \frac{1}{B_{\text{crustal thickness}}} \quad \text{(eq. 5.1)}$$

$$B_{\text{crustal thickness}} = \frac{Z_o}{Z} \quad \text{(eq. 5.2)}$$

A second method of determining beta is using the subsidence of the margin. In this case, the total tectonic subsidence (TTS) is determined by using the water depth ($D_{\text{H2O}}$) and the depth to the sediment/crust interface ($Z_{\text{sediment}}$) determined by Zelt et al. (1999). Subtracting the seafloor from the depth to the sediment/crustal interface give the thickness of sediment in the basins. Assuming a subsidence of 1/3 of the thickness due to sedimentary loading, and adding the current water depth supplies the subsidence related to tectonic activity (equation 5.3). Assuming that average compensated seafloor ($D_{\text{ave}}$) sits 6 km below sea level, dividing $D_{\text{ave}}$ by (1-TTS) supplies a second form of gamma (equation 5.4).

$$TTS = D_{\text{H2O}} + \sqrt[3]{Z_{\text{sediment}}} \quad \text{(eq. 5.3)}$$

$$\gamma_{TTS} = \frac{D_{\text{ave}}}{1 - TTS} \quad \text{(eq. 5.4)}$$

These calculations show that crust in the Galicia Interior Basin thinned by more than 70% during the rifting of the Iberia margin (figure 24). Crustal thinning at Galicia Bank is 50%, while in the Deep Galicia Basin it is between 90% and 100%. One hundred
percent thinning indicates that no crustal material remains at the surface and that instead sub-crustal material should be exposed. The total crustal stretching is 200 kilometers across Line 1 between the shore of the Iberia peninsula and oceanic crust, a distance today of 320 km. This indicates that the Galicia margin underwent an extension \((e)\) of 1.67, by equation 5.5.

\[
e = \frac{l_d - l_o}{l_o} \quad \text{(eq. 5.5)}
\]

where \(l_d\) is the deformed length (320 km) and \(l_o\) is the original length (120 km).
Chapter 6: Summary

The deformation that formed the Galicia Bank, the Deep Galicia Basin and the Grand Banks of Newfoundland/Flemish Cap occurred over a significant period of time. Plate reconstruction studies have shown that there was more stretching before initiation of seafloor spreading between the Flemish Cap and Galicia Bank and that compared to extension to the south in the Central North Atlantic, stretching occurred over a longer period of time (Srivastava et al., 1990). This change in intensity and localization of strain was probably controlled by the location of the deformation just to the north of the MGFZ. Motion of Africa to the southwest during the Triassic rifting of Pangea was accommodated mainly on the MGFZ. While the majority of deformation between the Central North Atlantic and the Northern North Atlantic was localized along this feature, diffuse Triassic extension also occurred on the opposite side of the MGFZ within the Galicia Interior Basin to the east of the Deep Galicia Basin. Though extension began in the Triassic 300 km to the east, rifting at Galicia did not begin until the Tithonian, and final rifting may not have occurred until Albian or Late Cretaceous. This 50 to 60 million year period of gentle extension was caused by the tectonic setting of the extension — perpendicular to a major fracture zone.

The influence of this strong discontinuity in the crust may have allowed more diffuse spreading to occur. This diffuse and slow spreading controlled the style of structural deformation in the rift, and may also have strongly affected the petrogenetic and volcanic history during the rifting. Because extension occurred over such a long period of time, as the crust thinned under a combination of simple and pure shear, mantle
material moved upward slowly. But, because of the slow rate, adiabatic decompression of the mantle did not occur, and therefore, minimal partial melts were generated during rifting. This can explain why the Iberia Margin, as well as the margin along the Grand Banks do not have a large quantity of volcanic material, as is found along other portions of the Atlantic margin (c.f. Withjack et al., 1998), and also why the Peridotite Ridge is not depleted in magmaphile elements.

Although seismic reflection data from the Galicia Interior Basin were not studied in this project, the larger-scale wide-angle data collected within it along Line 1 (figure 1) were used to measure a number of stretching factors across the width of the Galicia Margin. These stretching factors supply a wider picture into which the reflection lines 1 and 4 can be fit. The thinning analysis shows that two crustal ‘neckings’ occurred along the Galicia margin; one within the Galicia Interior Basin, and another to the west of the Galicia Bank within the Deep Galicia Basin. This analysis also shows that the eastern two-thirds of the margin contain one-half of the stretching (figure 24). The final extension, leading to sundering of Pangea, is preserved in the outmost 100 km of the Iberia Margin. It is this extension that is seen in Lines 1 and 4 (figures 18 and 19). Both Lines 1 and 4 show listric faults which sole into the detachment surface S. Long-offset velocity modeling by Zelt (1999) shows that high velocity material directly underlies S, supporting the interpretation that the hanging wall blocks above S are upper crustal blocks over a mantle footwall. A reconstruction of a possible rifting scenario is shown in figure 26.

In this model, the locus of extension is seen to move westward through time. Initially, extension was accommodated along fault surface Q (figure 26 (a)), the landward
most of the large-scale normal faults in the S reflector area. Shortly afterward, some
extension occurred along surface S, lowering the pre-rift surface below the erosive base-
level, forcing the deposition of the syn-rift material eroded from the footwalls of S and Q
into the hangingwall basins. This is suspected because the thickness of the pre-rift unit is
thinner in the hangingwall above R than it is above Q or S, indicating that it was a
topographic high and was eroded. Later, the locus of extension moved from fault Q to
fault R (figure 26 (b)) probably after a pause in extension allowed the rheologic strength
of the crustal column near Q to strengthen. As this occurred, the mantle responded by
moving upward and pushing the lower crust aside as the upper crustal section thinned
through simple shear processes. The movement of the lower crust effectively thinned the
crust by pure shear means. After the majority of the lower crust was removed, extension
moved to surface S (figure 26 (c)), allowing the extension along S to exploit the
weakness at the interface of the crust and the mantle. As the majority of extension is
accommodated along S, portions of the hangingwall peel off and are stranded on the
footwall, as in the rolling hinge model of extension as in Buck (1988) or Wernicke and
Axen (1988). In the hangingwall, a single major down-to-the-east antithetic normal fault
may have formed (figure 26 (c)), as shown by Wernicke (1985). In fact, LITHOPROBE
line 85-4 on the Grand Banks of Newfoundland shows two crustal-scale faults dipping to
the east which appear to penetrate the crust to the mantle, but not offset the Moho.
Instead, the crust below the hangingwall of these faults is thinner than that in the
footwall. This geometry is exactly what would be expected for faults formed in the
hangingwall above a simple-shear detachment.
At this time, the crustal section has thinned significantly, and probably is what permits the rifting along S to ultimately succeed. Because of the synthetic normal faults above S which cut the entire upper crust (figure 26 (c) and (d)), seawater can penetrate to the mantle. This serpentinizes the uppermost mantle material, weakening it and causing it to expand due to hydration-related volume changes. At the same time, the density of the material is reduced, making it lighter. The combination of less dense and weaker material enables upward movement. The movement of the serpentinized mantle material upward (and perhaps laterally as well) deformed S into the pronounced convex upward shape observed on Line 1. This upward motion may have also affected the western end of R, distorting its geometry as well. The serpentinization and deformation enhances the seismic impedance of the crust-mantle boundary, creating the bright reflector S.

As the western hangingwall (today the Grand Banks of Newfoundland and Flemish Cap) pulled away from the eastern foot wall and the drift stage initiated, footwall mantle material was exposed seaward of S in a setting similar to a metamorphic core-complex. During and after this unroofing of mantle material, the upper mantle domed upward at the Peridotite Ridge, perhaps through buoyant processes, similar to salt diapirism (i.e. a salt pillow). Since the crust above the mantle had been thinned along north-south striking normal faults, the Peridotite Ridge exploited these weaknesses to form the N-S linear elongated nature of the Peridotite Ridge. As the mantle material moved upward, into the Peridotite Ridge, the deeper material would have moved further west than the shallower material, explaining the top to the east shear sense indicators found at ODP Holes 639 E and F.
This deformation at Galicia occurred in three dimensions. The majority of the extension was oriented east-west, but there was also a component of north-south extension during the rifting. As seen in figures 21 and 22, the mantle moved to within 3 km (1.5 seconds TWTT) of the current surface, and may have been exposed at the surface during rifting (see the mantle-pre-rift contact at CMP 4200 in figure 19). As extension progressed, the crustal material on the hangingwall moved to the west-northwest, creating down to the north normal sense shear zones on Line 5, and down to the west normal faults and detachments on Line 1 and 4. Keen et al. (1990) and Tankard and Welsink (1989) describe similar north dipping normal faults within the Jean d’Arc and Horseshoe Basins on the Grand Banks of Newfoundland, the hanging wall of Galicia.

Within the basins found on Line 5, the lowermost syn-rift sediments dip strongly to the south, while younger sediments dip weakly, if at all. This indicates that the more northerly oriented displacement occurred late in the rifting of Galicia from Newfoundland. This suggests a change in the orientation of extension, from generally east-west to more northwest-southeast directed. This may have occurred as rifting progress northward past the Galicia Bank/Grand Banks-Flemish Cap segment into what would become the Bay of Biscay and the Labrador Sea, the margins of which are oriented nearly perpendicular to the central North Atlantic rifted margins. This change in orientation of strain may have much to do with the formation of the north dipping extensional features seen in Line 5.

This extensional deformation occurred as seafloor spreading was occurring in the central Atlantic, and the MGFZ was accommodating transform motion between Africa and the Avalon terrane. The diffuse nature of the extension, combined with the
Figure 24. (a) Schematic diagram of the measurements used to calculate the values plotted in (b). (b) Plots of γ and total extension across the Iberia margin. Gamma (γ) is a measure of change in cross-sectional area, used for characterizing extension. Two methods of determining γ are used: Crustal Thickness (blue line) \[1-(1/B)\], where \[B=30/(\text{Moho depth - Basin depth})\] and Total Tectonic Subsidence (red line) \[(\text{Water depth} + 0.33*(\text{Basin depth - Water depth})/6)\]. The total extension (green line) is the sum, from east to west, of γ across the margin. This plot highlights the marked change in extension in the western 100 km of the margin. A wide-angle traveltime refraction model from Zelt et al., 1999 was used to pick the data on 1 km spacing, across all of Line 1.
Figure 25. Wide angle velocity model for the entire length of Line 1, from Zelt et al., (1999). (A) shows a non-exaggerated, 1:1 cross section, while (B) shows a 5:1 exaggeration. Note the location of the Moho (white line with black dots), and the thinness of the crust above the S reflector, where it approaches 5 km thick. Also note the necking of the crust within the Galicia Interior Basin. PR is the Peridotite Ridge, GB is Galicia Bank, GIB is the Galicia Interior Basin.
Figure 23. A tectonic model of the deformation seen in Line 1 at four time steps (A - D). At time A, extension is being accommodated along fault surface Q, the landward most of the low angle features seen along the Iberia Margin. Syn-rift sediments fill the half-graben bound by Q. In B, extension has progressed to surface R, and movement on Q probably ceases. As this is occurring, deeper (lower crust/upper mantle) material is moving upward as the crustal section thins through both pure and simple shear processes. At time C, extension accommodation moves to surface S, and the hangingwall is faulted (figures 7 and 9). By this time, the crustal section is substantially thinned, and S exploits the weakness at the interface of the crust and the mantle. The thinning of the crustal section to this point probably permits the rifting along S to ultimately be successful. Between times C and D, water penetrates along the upper crustal faults, serpentinizing the uppermost mantle material, weakening it and causing it to expand upward due to volume and buoyancy changes. This serpentinization and deformation enhances the seismic impedance of the crust-mantle boundary, creating the bright reflector S. As this mantle material continues to ascend beneath the foot wall, R and S are further deformed into convex upward geometries. As the western hanging wall is pulled away from the eastern foot wall and the drift stage initiates (after time C), mantle material is exposed seaward of S similar to a core-complex. During and after this process, the upper mantle domes upward at the Peridotite Ridge, perhaps through processes similar to salt diapirism (i.e. a salt pillow). Time D shows the current configuration of the Iberia Margin. No vertical exaggeration
sedimentary record of the Tithonian through Albian, suggests a long history of such
tension. When final rifting sundered Galicia from the Grand Banks, it occurred rapidly,
since little sedimentary record of the breakup was found in the ODP holes. The relative
subsidence of the eastern side (Galicia) of the rift versus the western side (Grand
Banks/Flemish Cap) as seen by the widespread Avalon unconformity at the top of the
syn-rift sediments further suggests that the footwall is on the eastern margin.

**FURTHER WORK**

This thesis is the first work to come out of the ISE '97 project, and as such, has a
number of gaps that could be filled by further work. Perhaps the most fruitful project
would be a depth migration of the seismic lines collected in the S reflector area. This
should greatly increase our understanding of the area by combining highly detailed
seismic cross sections with a detailed understanding of the velocities present. Some
depth migration was attempted in this project, however the tools available were too
primitive for success. Use of a true velocity/depth modeling package would greatly
increase the probability of success.

Once depth migrated, the seismic data and velocity/structural models could then
be combined into an intersecting three dimensional model over the S reflector region.
Combination of these velocity models based on migration velocities with the velocity
models derived from the long-offset data would provide further information for refining
the migration velocity models. With these data, a structural analysis would be possible,
and a number of structural models could be proposed. If the three dimensional velocity
modeling package could also perform structural restorations, then these larger-scale
models could then be validated.
A more detailed interpretation of the syn- and post-rift strata would provide important constraints on the relative timing of deformation within the S reflector area. The syn-rift strata would provide important timing relationships to determine the relative movement of the fault blocks (e.g. S, R, and Q), while analysis of the post-rift sediments may help to constrain the timing of any post-rift equilibration of the mantle under S and over the Peridotite Ridge. Once again, any such interpretation should be analyzed in three dimensions to highlight depositional systems, depocenters and topographic highs. These could then be used, along with the structural model, to determine the paleo-topography of the region, enabling the more complete understanding of both the structural geology and the sedimentology.

A final task in the S reflector area would be to determine subsidence curves (similar to figure 24) at as many places as possible across the margin using the existing seismic data, including the on-shore long offset data collected during ISE '97 by the Spanish and Portuguese participants. If combined with similar subsidence studies on the opposing margin, and the analysis of the syn- and post-rift sedimentary units mentioned above, a cohesive understanding of response of the mantle during extension and rifting may be achieved.
Chapter 7: Conclusions

Structures found along the Galicia Bank tell us much about how deformation along the Iberia margin occurred, both within the crust and the mantle. Extension began in the Triassic following the contractional mountainbuilding events that formed the Appalachian Chain in the Late Cambrian to Early Permian. The locus of extension between the Iberia Massif and the North American craton moved through time: initial extension occurred, creating basins such as the Galicia Interior Basin while later extension was concentrated in the Deep Galicia Basin. Total extension across the margin was 1.67 (more than 200 km) based on stretching determined from crustal thickness measurements.

Three main faults control the segmentation of the Deep Galicia Basin where studied along ISE '97 Lines 1 and 4. These faults dip to the west, and have normal sense, down to the west motion. They are named from the landward to seaward, Q, R and S, respectively. S was previously recognized by investigators as a detachment surface (Boillot et al., 1988b; de Charpal et al., 1978; Krawczyk and Reston, 1995; Reston et al., 1995). This study is the first to recognized that R acted as a detachment surface during extension as well.

A number of secondary faults accompany Q, R and S in the segmentation of the crust of the margin. All of these secondary faults occur in the upper crust and sole into or otherwise terminate at a master fault. These faults form half-graben basins into which syn- and post-rift sediments were deposited.
Above S the crustal blocks bound by the secondary faults have been domed upward by sub-detachment mantle flow. In this area, the mantle has become mobilized, perhaps via the addition of water to form serpentinite, which lowers both the strength and density of the upper mantle peridotite. This change in density and strength allowed the serpentinized upper mantle to flow under the gravity stress of the adjoining continent and the nearby upper crustal blocks.

Extension in the area of the Deep Galicia Basin occurred in at least three phases. In the first phase, motion along fault Q formed a basin above the footwall into which syn-rift sediments were shed. A second, perhaps concurrent phase occurred as fault S initiated and accommodated extension. Later, motion along fault R accommodated more extension, until finally, the hanging wall was detached from S. This caused the isostatic accommodation of the mantle under S and the formation of the Peridotite Ridge.
References


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Appendix A:

INTEGRATION OF SEISMIC INTERPRETATION WITH SEISMIC PROCESSING

One hope for the study of these data was that interpretation of the data could be integrated with the processing of the data. This involves the use of the stratigraphic and structural picks to modify the data and control parameters during processing within ProMax. It was hoped that this would enable refinement of the processing sequence, and provide enhanced output to enable refinement of the interpretation.

The horizons picked in SMT’s KINGDOM Suite were used for a variety of functions outside of KINGDOM Suite. They were mainly used as input into UNIX shell scripts and programs. The output from these applications was used in ProMax 6.2 and the Generic Mapping Tools (GMT) 3.2 and 3.3 (Wessel and Smith, 1991, 1998).

The picked horizons were processed by a number of FORTRAN programs to create data tables for import into Landmark Graphics ProMax 6.2 seismic processing application. The availability of such surfaces in an electronic format made certain parameters supplied to processing flows much more consistent than otherwise would have been possible, because functions could be computed mathematically from the interpreted horizons.

The horizons picked in the KINGDOM Suite were used to create ProMax horizons; to produce top, surgical and bottom mutes; and to create Miscellaneous Time Gates for advanced ProMax modules such as Time-Variant Filtering and Time-Offset Variant Gain. Automated time-gate production (in both source and CMP domains) via
external FORTRAN programs allowed for the mathematical determination of values, rather than using the interactive picking tools of ProMax. This allowed for more rapid production of such tables, and the reduction of human error in picking the values.

In addition, velocity fields were produced from the KINGDOM Suite horizons by an external FORTRAN program. This program, written by Dale Sawyer, created a file containing polygons to which a velocity could be assigned in ProMax. These polygons were then given a velocity function, and a t/x velocity field was created. This velocity field was then applied to Kirchoff time and depth migration, in an attempt to further refine the sections.

These horizon files were also used in GMT shell scripts to plot the surfaces on top of the seismic data for presentation of interpreted sections. GMT creates small PostScript files, which can then be imported into applications such as Adobe Illustrator for further refinement. All of the seismic figures in this thesis were created with GMT, and edited in Illustrator.

**DATA PRESENTATION**

**GMT**

Once interpreted and processed, these data were presented in a number of ways. Two posters were presented at American Geophysical Union meetings in 1999 showing these data (Unger and Sawyer, 1999; Unger et al., 1999). These data were exported to an IEEE Real SEG-Y formatted data file. These SEG-Y files were plotted using psssegy and the GMT application package (Wessel and Smith, 1991, 1998). Horizons and velocity fields were also plotted using GMT tools.

**IberiaWeb**
Information about ISE '97 was also presented on the World Wide Web at <http://terra.rice.edu/iberiaweb>. These pages were designed and created while I was waiting for the ProMax processing steps to complete. These pages were designed to supply information to researchers and lay people regarding the purpose of ISE '97, its location, the types of data collected during the experiment, how the data was collected, and what is being done to process these data. The rationale behind creating such a web site was to provide outreach to the academic and non-academic community; to educate; and to share information among ISE '97 researchers. Every visit begins at the start page, where the purpose of the experiment is discussed, and the site is described. From there, there are links to pages covering ISE '97 News, describing abstracts and papers being presented; People, the participants in ISE '97; Publications; Tools, the tools used to analyze and interpret the data; Data, the line and OBS latitude/longitude locations, processed lines, GMT scripts and maps, etc; and an About page, where there is a discussion of data collection methods and an overview of the project.

But not all the pages are for outreach. A number of the page were designed to facilitate information sharing within the Iberia research group at Rice University. These pages include listing of FORTRAN codes available to manipulate horizon files, GMT scripts to create various figures, and a listing of suggested processing flow names. This naming system suggests the naming of flows with a unique 3 number preface to order the flows into a coherent listing, starting with data loading and geometry set-up in the 001 – 009 range, noise checking, first velocity analysis and stack in the 010 – 019 range, finishing with data output flows in the 090-099 range. This use of a prefix forces the ordering of the flow in the ProMax Flow window, and permits easy location of a needed
flow. It brings order to a listing of 50+ flows in the Line 1 ProMax area, placing them in the order of use.

As of 21 June, 2000, these pages received 1,900 hits from 355 unique hosts. The site averaged 65 visits per week, with visitors from 35 different top level domains (TDL). More than 20% of the pages were served to .edu TDLs, suggesting that other researchers in U.S. educational institutions were accessing the pages. Hits from TDLs .de (Germany), .uk (the United Kingdom), .au (Australia), .ca (Canada) and .ie (Ireland) made up nearly 10% of the visitors. Fifty percent of the hits came from .com TDLs.
NOTE TO USERS

Oversize maps and charts are microfilmed in sections in the following manner:

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This reproduction is the best copy available.
Figure 18. Northernmost seismic line in the S reflector is the uninterpreted line in two-way travel time, while the structural interpretation in TWTT of these seismic sections most significant features visible on this line are the horst formed by the basin bounding normal faults which dip to the west. Secondary normal faults in the remnant horst between CMPs 1500 and 3800, formed as S domed up within the hanging wall of S may be lower angle extensional faults at below seismic resolution. These faults were accommodated earlier extension.

Faults R and Q break through the crust and form half-grabens. These faults do not appear to offset the E-W trending half-graben to the west of R (CMPs 4300-4700). The decreased thickness of this unit indicates that Q formed as the result of erosion to remove 50% - 80% of pre rift material (however, this material may not have had a constant thickness, so this determination may not be accurate). The thickness of the pre rift unit in the graben formed by motion on the upper portion of S may have occurred at the same time as well. At a later time, R formed, and accommodated 5 km of extension. The master fault Q accommodates 9 km of extension.

Compare to figure 10(c) and note the similarities in spacing and the final shape of the detachments, the distance the upper crustal blocks resting directly on mantle materials, the location of the Peridotite Ridge to the denuded mantle in the Malenco area. Other features of this line are discussed in the text.
Structural and tectonic interpretation of deep seismic reflection data offshore Spain and Portugal: a tectonic rifting model
Michael R. Unger, 2001
Figure 19. Seismic dip Line 1 in the S reflector area, 13 km to the south of Line 4 (figure 18). Line 1 is aligned with Line 4 at the crossing of Line 5 (vertical line). (A) shows the uninterpreted line while (B) shows an interpretation of these data. Note the fault-bound half-grabens, as on Line 4, the detachment surfaces R and S, as well as the convex upward, though slightly west dipping S surface. The material above the eastern end of the sub horizontal S surface is composed of coherent, cohesive blocks, while the material on the western end of S appears to be disaggregated by normal faults which are below seismic resolution. These faults and the updoming of S are related to the upwelling of the mantle beneath S at the end of rifting.

Compare to figure 10(c) and the similarities to the transition between passive margin (Bermia) and the Piemont-Liguria oceanic crust.
Figure 20. Strike parallel line, crossing Lines 4 and 1 at the indicated points. (A) is uninterpreted, (B) is interpreted. Notice the gentle slope and thickening of these units of the post and syn rift sediments to the south. Above the detachment surface S, the pre rift and crust units are severely dismembered, and in places, syn rift sediments directly overly mantle material. This indicates that the mantle was exposed at the surface during deformation. Note the segmentation of the crustal materials by north dipping faults. These faults show top to the north normal shear sense.

Below the S detachment, en echelon shear zones also plunge to the north. In places, lozenge- shaped structures in mantle may indicate shear around stronger boundin-type structures.
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Figure 21. Depth converted seismic Line 4, thin in the S reflector area. Line 4 trends east-west km to the north of Line 1 (figures 19 and 22). (a) the uninterpreted data while (b) shows an interpretative.

The crust is greatly attenuated across this > 17 km on the east side to a minimum of 0.5 km. Three major surfaces (Q, R and S) serve to the crustal material is the thickest. To the west, of somewhat less than 30° west, and cuts across the mantle. Further west is S, a convex upward at least 10 upper crustal faults terminate. Surface appear to offset the Moho, while S is the interface and mantle material.

The material above S is domed ~2 km upward 1500 and 4000. This suggests upward movement after the formation of S. This may be the cause spaced faults which terminate at the S reflector.

Compare to figure 10 (c).
Structural and tectonic interpretation of deep seismic reflection data offshore Spain and Portugal: a tectonic rifting model
Michael R. Unger, 1991

Plate 2
Figures 21, 22 and 23
Depth converted seismic sections

...
Figure 22. Depth converted seismic Line 1, 13 km to the south of Line 4 (figures 18 and 21). Line 1 is an east-west trending line, sub-parallel to the dip of the rift related structures (figure 1). Figure 22 (a) shows the uninterpreted line, while (b) shows an interpretation of these data.

Note the half-grabens bound by west dipping faults and containing a relatively thin (< 1.5 km) layer of syn-rift sedimentary units. The thickest syn-rift package is near CMP 4900, where the thickness of the crustal material reaches a minimum. This maximum thickness of syn-rift material suggests that the most subsidence occurred where the crust was thinnest.

The thickness of the pre-rift unit can be used to determine the relative ages of the major faults (Q, R and S), assuming a constant pre-rift thickness prior to the onset of deformation. This is because more erosion should occur on topographic highs than in low points. Therefore, thicker pre-rift packages should abut the oldest faults, while thinner pre-rift packages should be adjacent to younger faults. The thickest pre-rift package is in the hanging-wall of Q, suggesting that it formed first. The pre-rift packages within the hanging-walls of R and S are roughly similar in thickness, indicating that similar amount of erosion occurred above both R and S. This suggests that R and S are roughly contemporaneous.

Comparing the general thicknesses to those on Line 4 (figure 21) suggests that initial movement along Q occurred in the region of Line 1, and that motion along Q on Line 4 occur later. Similarly, thicknesses above R and S are less on Line 4, suggesting that motion occurred earlier closer to Line 1.

Note that S is generally west dipping here, and that the Moho has a convex upward shape similar to that seen in figure 21.
Figure 23. Strike parallel depth converted seismic Line 5, crossing Lines 4 (figures 18 and 21) and 1 (figure 19 and 22) as indicated. Figure 23 (a) is uninterpreted while (b) is interpreted. In (b), notice the gentle southerly slope and the southerly thickening of the post-rift units.

Here, S is sub-horizontal and serves as the interface between upper-crustal units and mantle material. Upper crustal faults, which serve to dismember pre-rift and crustal material, sole into S. Below S, Q and R on Lines 4 and 1 are seen as a north dipping series of shear zones within the mantle. These suggest that deformation occurred within the mantle during rifting. The geometry of these shear zones suggests that the overall movement of the mantle material during rifting had a southern directed component of motion.

Figures 21 through 23 suggests that mantle material moved to the west-southwest during rifting in this area. As this occurred, crustal material was domed upward, with maximum uplift occurring at CMP 3000 on Line 4, CMP 4500 on Line 1 and CMP 4300 on Line 5.