CRUSTAL STRUCTURE OF NORTH ATLANTIC FRACTURE ZONES

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Abstract. Seismic studies have established that large-offset transforms along the slow spreading Mid-Atlantic Ridge exhibit anomalous crustal structures that fall well outside the range typically associated with oceanic crust. Seismically, fracture zone crust in the North Atlantic is extremely heterogeneous in both thickness and internal structure. It is frequently quite thin (<1–2 km thick) and is characterized by low compressional wave velocities and the absence of a normal seismic layer 3. A more gradual crustal thinning can extend up to several tens of kilometers from these fracture zones. Anomalously thin crust has also been inferred from both seismic and gravity studies at smaller ridge axis discontinuities along the Mid-Atlantic Ridge. The geological nature of the seismically anomalous crust found within Atlantic fracture zones, and how this crust forms, are still controversial. One interpretation consistent with available seismic observations is that the crust within North Atlantic fracture zones consists of a thin, intensely fractured, and hydrothermally altered basaltic section overlying ultramafics that are extensively serpentinized in places. Variations in apparent seismic crustal thickness along fracture zones may reflect different degrees of serpentinization of the upper mantle section or changes in the thickness of the igneous crust. The existence of a thinner crustal section in fracture zones can be explained by a reduced magma supply within a broad region near ridge offsets due to the three-dimensional nature of upwelling beneath a segmented spreading center and by tectonic dismemberment of the crust by large-scale detachment faults that form preferentially in the cold, brittle lithosphere near the ends of segments.

INTRODUCTION

The global mid-ocean ridge system is segmented into discrete spreading ridge segments by a variety of ridge axis discontinuities (for a review, see *Macdonald et al.* [1991]). The largest of these are transform faults which link ridge segments along a narrow strike-slip fault zone and offset opposing ridge segments by distances of tens to hundreds of kilometers. These boundaries are relatively stable for millions of years, and their aseismic extensions can, in some cases, be traced as small circles about the pole of relative plate motion for thousands of kilometers onto the ridge flanks [*Menard and Chase*, 1970]. Numerous smaller, nontransform discontinuities also have been identified along ridge crests. They include a range of offset geometries (e.g., overlapping spreading centers along the fast spreading East Pacific Rise [*Macdonald and Fox*, 1983; *Lonsdale*, 1985] and oblique shear zones on the slower spreading Mid-Atlantic Ridge [*Schouten and White*, 1980; *Sempere et al.*, 1993]). They typically are associated with small (less than a few tens of kilometers to less than a kilometer) ridge crest offsets, and their off-axis traces show that they can migrate along the ridge axis. The larger nontransform offsets can persist for several million of years, while the smaller offsets are not associated with recognizable off-axis traces and appear to be relatively short-lived [*Macdonald et al.*, 1991]. In the North Atlantic, where transform faults and smaller nontransform discontinuities are typically found every 10–100 km along the Mid-Atlantic Ridge, a significant fraction (10–20%) of the oceanic crust is formed in or near these features.

The rugged topography and geologically complex tectonic setting of transform faults and other ridge axis discontinuities make them far from ideal sites for seismic crustal studies. The extreme topographic relief and geological heterogeneity of the transform domain make them difficult areas in which to carry out sesimic studies and introduce significant uncertainties in the associated crustal velocity solutions. Consequently, of the hundreds of seismic experiments that have been

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carried out in the ocean basins over the past 40 years, only a few have attempted to investigate the structure of oceanic fracture zones, and most of these have been carried out during the past decade.

A major motivation for attacking the difficult problem of determining fracture zone crustal structure was the realization that transform faults constitute a major source of crustal heterogeneity in the ocean basins [*Fox*, 1978]. It had been recognized for many years that the walls bordering many large Atlantic fracture zones (e.g., Oceanographer, Vema, Romanche, Kane) expose a diverse suite of gabbroic and ultramafic rocks that are not commonly found elsewhere in the ocean basins [*Miyashiro et al.*, 1969; *van Andel et al.*, 1973; *Fox et al.*, 1972, 1976; *Bonatti*, 1976, 1978]. Submersible studies of several ridge-transform intersections [*Ballard et al.*, 1979; *Stroup and Fox*, 1981; *Karson and Dick*, 1983; *OTTER Scientific Team* 1984, hereinafter referred to as OTTER] and of the walls of the Oceanographer [*OTTER*, 1985] and Vema [*Auzende et al.*, 1989] transforms have confirmed the ubiquitous exposure of basaltic, gabbroic, and ultramafic rocks in transform valley walls. *Francheteau et al.* [1976] argued that these rocks, which are generally associated with the lower crust and upper mantle, can be exposed on the small-throw faults that form fracture zone valley walls only if the basaltic layer (layer 2) is unusually thin in fracture zones or if large-scale faulting exposes the lower crust. In either case it was clear that an understanding of fracture zone crustal structure, and its relationship to "normal" oceanic crust, would be important in order to use rocks recovered from fracture zone walls to reconstruct the geologic structure of oceanic crust.

Another motivation for determining the variation in seismic crustal thickness near segment boundaries is to constrain the integrated flux of magma along a spreading center. This in turn will be controlled by the pattern of upwelling and melt migration in the underlying mantle. A systematic variation in crustal thickness from segment midpoints to segment offsets would provide evidence for a focusing of upwelling in the underlying mantle while a relatively uniform crustal thickness along-axis would suggest a more two-dimensional upwelling pattern [*Lin and Phipps Morgan*, 1992]. Crustal thickness variations along and across a spreading axis may also reflect tectonic stretching and thinning of the oceanic crust by faulting. An asymmetry in crustal thickness across ridge segments, especially near segment boundaries, would provide important evidence for the kind of large-scale detachment faulting often associated with crustal extension in continental rifts [*Karson*, 1991].

In this paper we review the results from 15 seismic investigations of fracture zone crustal structure in the North Atlantic (Table 1). Comparatively little information is available on the crustal structure of fast-slipping transforms, and we will thus focus in this paper on the crustal structure of transforms and ridge axis discontinuities at the slow spreading Mid-Atlantic Ridge. Most of these studies have relied on conventional seismic refraction techniques, often using ocean bottom receivers, and have concentrated primarily on determining the crustal structure at several major Atlantic fracture zones. Recently, however, several experiments have been carried out at smaller Atlantic fracture zones, and normal incidence and two-ship multichannel seismic techniques have also been employed. This paper is divided into three sections. The first two sections review seismic results from large- and small-offset North Atlantic fracture zones. In the final section these observations are integrated into a model of fracture zone crustal structure, and the implications of this model for crustal accretion along the slow spreading Mid-Atlantic Ridge are discussed.

GLOSSARY

For the nonspecialist we include a short glossary of terms used in this paper:

Crust-mantle triplication: seismic energy turning in the steep seismic velocity gradient at the base of the oceanic crust, which results in anomalously high amplitudes as three different phases interfere. The shot-receiver range at which this caustic occurs is diagnostic of the thickness of the oceanic crust.

Fracture zone: strictly speaking, the inactive trace of a transform fault which extends flankward of the offset ridge crests, although the term is often used synonymously with transform fault in many circumstances.

Mantle Bouguer anomaly (MBA): a gravity anomaly calculated from the free-air anomaly by subtracting the gravitational attraction of the water-crust and crust-mantle boundaries assuming a constant-thickness, constant-density crust. The MBA highs associated with parts of some Atlantic fracture zones, and the traces of smaller ridge axis discontinuities, indicate the presence of anomalously thin crust and/or higher than normal crustal or upper mantle densities.

Moho: an abbreviation for the Mohorovicic discontinuity, the seismically defined boundary between the crust and upper mantle. In "normal" oceanic crust the Moho is marked by a relatively abrupt increase in seismic compressional wave velocity to values of $V_p >$ 7.6 km s^{-1} and typically occurs about 6–7 km below the seafloor. The "crust," by definition, has lower compressional wave velocities and its thickness is determined by the depth to Moho. The geological interpretation of this seismically defined crust is somewhat problematic, especially since the crust formed proximal to oceanic transforms may differ structurally and compositionally from that found elsewhere in the

TABLE 1. Atlantic Fracture Zone Seismic Experiments

Fracture Zone	*Offset*	*Reference*	*Summary of Results*
Charlie Gibbs	340 km	*Whitmarsh and Calvert* [1986]	Two OBS refraction lines across and along the southern trough of the active transform. Two-dimensional ray tracing interpretation shows thin crust (about 4 km thick) with low crustal velocities (<6 km s^{-1}) and the complete absence of layer 3 beneath the transform valley. Transverse ridge underlain by relatively normal crustal thicknesses and velocities.
Vema	320 km	*Ludwig and Rabinowitz* [1980]	Unreversed sonobuoy refraction line interpreted using slope-intercept method indicates a thicker layer 2 and a thin or absent layer 3 beneath the transform valley with a total crustal thickness of about 5 km.
Vema	320 km	*Detrick et al.* [1982]	A 75-km-long OBS refraction line in the Vema transform valley. Results similar to those of *Ludwig and Rabinowitz* [1980] with near-normal crustal thickness (~5 km), but low velocities throughout the entire crustal section.
Vema	320 km	*Potts et al.* [1986*a*] *Louden, et al.* [1986]	OBS refraction experiment. Region of anomalous crust extends 20 km to either side of the fracture zone. Thinnest crust is found beneath the edges of the fracture zone valley. Seismic layer 3 absent in the fracture zone, and crustal velocities are anomalously low. Line down adjacent MAR rift valley shows immature crust (3–3.5 km thick) with low upper mantle velocities.
Kane	160 km	*Detrick and Purdy* [1980]	Well-constrained determination using OBH of 2 to 3-km-thick crust beneath a 50-km length of fracture zone trough. The anomalous fracture zone crust is characterized by low compressional wave velocities and the absence of a normal seismic layer 3.
Kane	160 km	*Cormier et al.* [1984] *Purdy et al.* [1986] *Abrams et al.* [1988]	Extensive OBH explosive and air gun refraction experiment along a 300 km length of the fracture zone. Crust 2–3 km thick common along most of the fracture zone, although thicker crust is present in the eastern transform valley. The crust may be less than 1 km thick at the eastern ridge-transform intersection, although mature crust of approximately normal thickness with ~8 km s^{-1} upper mantle velocities exist further south along the MAR rift valley. The KFZ transverse ridge is underlain by seismically normal crust.
Oceanographer	130 km	*Fox et al.* [1976]	One 35-km unreversed line; porly constrained two-layer slope-intercept solution indicates slightly lower crustal velocities; crustal thickness unknown.
Oceanographer	130 km	*Sinha and Louden* [1983]	Four 100-km-long sonobuoy refraction lines interpreted using two-dimensional ray tracing techniques. Crust 4–5 km thick with anomalously low crustal velocities found in the fracture zone with some crustal thinning extending >30 km from the fracture zone. Crust less than 2.5 km thick found beneath part of the transform valley.
Oceanographer	130 km	*Ambos and Hussong* [1986]	Two long OBS refraction lines across and along the Oceanographer transform. Results similar to those of *Sinha and Louden* [1983], with extremely thin (2–3 km) crust beneath the western transform valley and near-normal crustal thicknesses (4.5 km) elsewhere along the fracture zone. Very thin layer 2 beneath south wall of transform valley.
Tydeman	110 km	*Calvert and Potts* [1985] *Potts et al.* [1986*b*]	Two separate experiments, the first over 75-m.y.-old crust with two-ship multichannel seismic expanding spread profiles, the second over 55-m.y.-old crust using OBS and sonobuoys with explosives shot along and across the fracture zone. Results show slightly thinner than normal crust (5 km), low crustal velocities, and the absence of layer 3.

TABLE 1. (continued)

Fracture Zone	*Offset*	*Reference*	*Summary of Results*
Fracture Zone 1	25 km	*White et al.* [1984]	Reinterpretation of line C of *Sinha and Louden* [1983] using two-dimensional ray tracing. The crust in the fracture zone displays abnormally low seismic velocities which extend for about 20 km on either side of the fracture zone. Crustal thicknesses are about 4.5 km, slightly less than that of the adjacent seafloor.
Blake-Spur	20 km	*Mutter et al.* [1984] *McCarthy et al.* [1988]	Two-ship, wide-aperture multichannel reflection profile reveals a strong, shallow subhorizontal reflector ~1 s below basement that if interpreted as Moho would indicate a crustal thickness of as little as 2–2.4 km. Subsequent reprocessing of this line revealed a deeper possible Moho reflection that would indicate no significant crustal thinning beneath the Blake-Spur fracture zone.
Blake-Spur	20 km	*Minshull et al.* [1991]	Major two-ship multichannel seismic experiment. Interpretation of ESPs across Blake-Spur fracture zone indicates the presence of a 15-km-wide zone of anomalous velocities of 7.2–7.6 km s^{-1} in the lower crust. Moho is associated with the deeper reflector identified by *McCarthy et al.* [1988].
Kurchatov	16 km	*Steinmetz et al.* [1977]	Long-range mantle arrivals from shots above the fracture zone are later by as much as 0.45 s suggesting low crustal velocities or a thicker crust in the fracture zone. Possible error in topographic corrections?
Unnamed fracture zone at 45.5 N, 21 W	10 km	*White and Matthews* [1980]	Delay time experiment using OBH and sonobuoys show 0.2-s decrease in layer 2 delay time over the fracture zone. Unusually high velocities (7.3–7.6 km s^{-1}) found for the lower crust.

Abbreviations are as follows: ESP, expanding spread profile; KFZ, Kane fracture zone; MAR Mid-Atlantic Ridge; OBH, ocean bottom hydrophone; OBS, ocean bottom seismometer.

ocean basins. Crustal velocities can be explained by a wide range of igneous and metamorphic rock types including basalts and gabbros, greenschist facies metamorphic rocks, or partially serpentinized ultramafics, all of which may be present in varying abundances in fracture zones. Thus the Moho within oceanic transforms does not necessarily correspond with the base of the igneous crust defined petrologically, and caution should be exercised in interpreting the seismic results presented below in terms of conventional geological models of oceanic crustal structure.

Nontransform offsets: small ridge axis discontinuities which typically offset spreading centers by a few kilometers or less. These offsets are not associated with through-going strike-slip fault zones and can migrate along the ridge axis. Unlike transform faults, their off-axis traces do not form small circles about the pole of relative motion.

Seismic layer 2: the upper portion of the seismically defined oceanic crust. It is typically associated with a rapid increase in compressional wave velocities from <2.5 km s^{-1} at the seafloor to ~6.6 km s^{-1} at depths of ~2 km (seismic layer 1 refers to sediments overlying volcanic basement). Layer 2 is sometimes subdivided into layers 2A, 2B, and 2C. It is generally interpreted as consisting of extrusive lava flows that grade downward into a sheeted dike complex.

Seismic layer 3: the lower portion of the seismically defined oceanic crust. It is typically about 5 km thick and is associated with low vertical velocity gradients (<1 s^{-1}) and compressional wave velocities of 6.5–7.0 km s^{-1}. Layer 3 is interpreted as plutonic rocks such as gabbro, cumulate gabbro, and cumulate ultramafics.

Serpentinite: a rock consisting almost wholly of serpentine minerals derived from the alteration of previously existing olivine and pyroxene.

Transform fault: one of three fundamental plate boundaries along which lithosphere is neither created or destroyed. The transform faults discussed in this paper are ridge-ridge faults. Although they are often treated as narrow zones of strike-slip motion, transforms at mid-ocean ridges are tectonically complex features characterized morphologically by a band of distinctive ridge and trough topography up to several tens of kilometers wide.

Ultramafics: a general term for basic igneous rocks such as peridotite consisting of olivine with or without other mafic minerals such as amphiboles and pyroxenes.

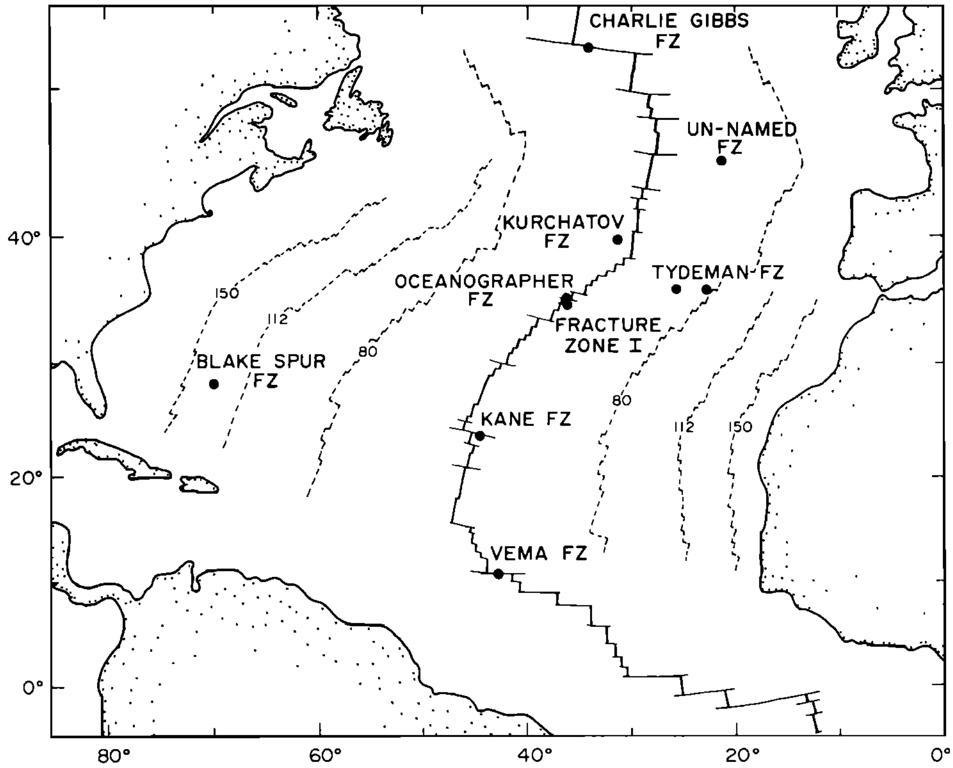

Figure 1. Simplified tectonic map of the North Atlantic showing the location of the fracture zone seismic experiments compiled in Table 1 and discussed in the text.

LARGE-OFFSET FRACTURE ZONES

The largest fracture zones in the North Atlantic offset the Mid-Atlantic Ridge between 100 km and 340 km, corresponding to age offsets of 6 Ma to nearly 30 Ma (Figure 1). Seismic refraction experiments have been carried out at five large North Atlantic fracture zones, the Charlie Gibbs, Oceanographer, Kane, Vema, and Tydeman fracture zones (Table 1). Of the twelve published studies, two used free-floating sonobuoy receivers [*Ludwig and Rabinowitz*, 1980; *Sinha and Louden*, 1983], three used ocean bottom hydrophones (OBH) [*Detrick and Purdy*, 1980; *Cormier et al.*, 1984; *Abrams et al.*, 1988], six used ocean bottom seismometers (OBS) [*Detrick et al.*, 1982; *Ambos and Hussong*, 1986; *Louden et al.*, 1986; *Whitmarsh and Calvert*, 1986; *Potts et al.*, 1986*a*, *b*], and one used a two-ship multichannel expanding spread technique [*Calvert and Potts*, 1985].

These studies have unequivocally established that large Atlantic fracture zones are associated with a crustal structure that differs in fundamental ways from that typically associated with oceanic crust. These differences are readily apparent in Figure 2, which compares the seismograms from a typical refraction profile shot away from the disturbing influence of a fracture zone with a profile shot along a fracture zone trough. The former profile shows the now familiar travel time and amplitude pattern typical of normal oceanic crust, especially the rapid increase in apparent phase velocity out to ranges of 10–15 km, the smaller amplitudes and typical layer 3 phase velocities between 15 and 20 km, and the large-amplitude buildup between 20- and 30-km range corresponding to the crust-mantle triplication. The seismograms on the fracture zone profile are completely different. Arrivals with crustal phase velocities are observed over only a few kilometers' range with no evidence for a typical layer 3 refractor, the crust-mantle triplication is located at 8- to 10-km range, and first arrivals with mantlelike phase velocities are present at ranges of only 10–15 km. These differences reflect crust within oceanic fracture zones that is both unusually thin and associated with an anomalous crustal velocity structure.

Velocity-Depth Functions

A compilation of all available velocity-depth profiles from large North Atlantic fracture zones is shown in Figure 3. In most cases, the crustal structure associated with these fracture zones falls outside the range of velocity structures typically found for oceanic crust in the North Atlantic. However, there are also major differences in both total crustal thickness and crustal velocities within and between fracture zones. The extreme heterogeneity in seismic crustal structure apparent in this compilation is indicative both of the structural and tectonic complexity of major oceanic fracture zones and of the higher degree of uncertainty associated with velocity-depth solutions in the topographically rugged fracture zone setting.

The seismic structures observed in oceanic fracture zones generally fall between two end-members, referred to as "type A" and "type B" crust by *Sinha*

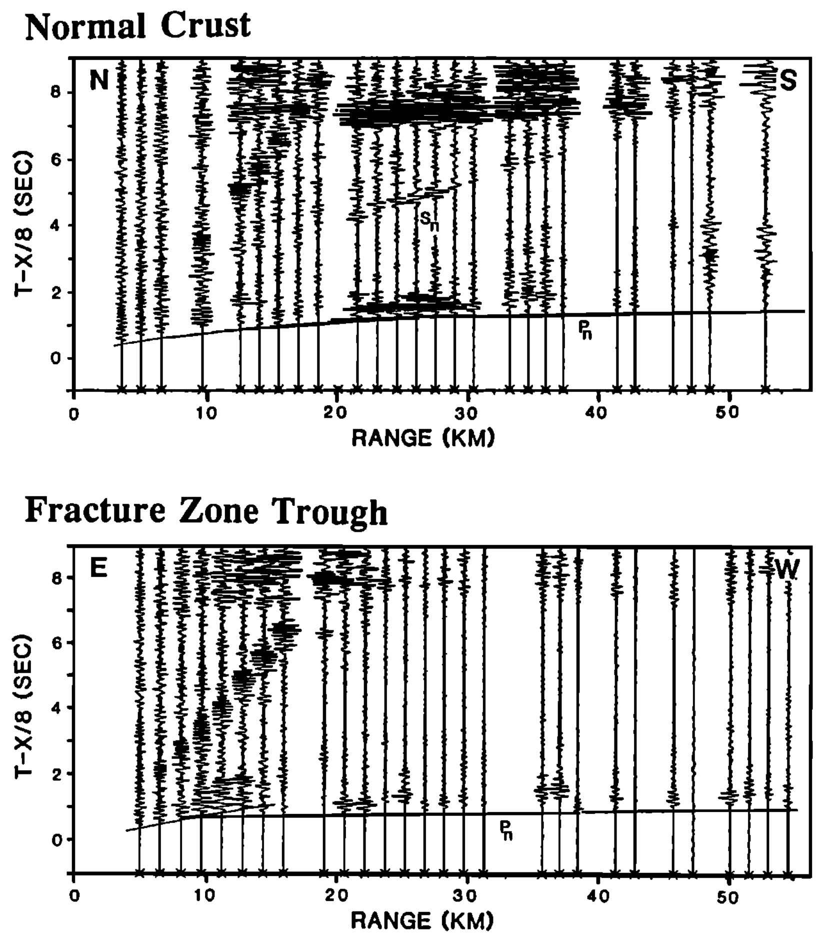

Figure 2. Comparison of topographically corrected seismic record sections representative of "normal" oceanic crust and "fracture zone" crust (modified from *Cormier et al.* [1984]). Note that the apparent phase velocities and the distribution of energy in the seismograms from the fracture zone are completely different from those in seismograms typical of crust formed away from the fracture zone.

and Louden [1983]. Type A crust shows some resemblance to normal oceanic crust, although it is slightly thinner (4–5 km thick) and characterized by lower seismic velocities, especially within the lower part of the crustal section. Type A crust has been found in the Vema transform valley [*Detrick et al.*, 1982; *Potts et al.*, 1986*a*; *Louden et al.*, 1986], along several parts of the Oceanographer fracture zone [*Sinha and Louden*, 1983; *Ambos and Hussong*, 1986], in the eastern portion of the Kane transform valley [*Cormier et al.*, 1984], at the Charlie Gibbs fracture zone [*Whitmarsh and Calvert*, 1986], and in the Tydeman fracture zone [*Calvert and Potts*, 1985; *Potts et al.*, 1986*b*].

The second type of anomalous fracture zone crust (type B) is extremely thin (<3 km thick) and characterized by a crustal velocity structure that bears little resemblance to normal oceanic crust. With seismic velocities typical of the upper mantle present at depths of only 1–2 km below the seafloor, this crust is the thinnest found anywhere on Earth. It is present extensively along the Kane fracture zone trough east of its intersection with the Mid-Atlantic Ridge [*Detrick and Purdy*, 1980; *Cormier et al.*, 1984], in the axis of the Oceanographer transform [*Sinha and Louden*, 1983; *Ambos and Hussong*, 1986], and along the edges of the Vema transform valley [*Potts et al.*, 1986*a*; *Louden et al.*, 1986].

The crust found in oceanic fracture zones is not just a thinner version of normal oceanic crust; its crustal velocity structure is fundamentally different. This is particularly well illustrated by synthetic seismogram modeling of an air gun profile shot along the Kane fracture zone trough (Figure 4). This record section displays large-amplitude, crustal arrivals out to 4- to 5-km range, a sharp decrease in amplitude over the next 1–2 km, then another buildup in amplitude between 7- and 9-km range associated with the crust-mantle triplication and the first appearance of mantle-refracted arrivals. Synthetic seismograms have been computed using the WKBJ method [*Chapman*, 1978] for three simplified, laterally homogeneous crustal models, each having a crustal thickness of 2.6 km. In the first model (Figure 4*c*) the crust is composed of a normal thickness seismic layer 2 with layer 3 completely absent. In the second model (Figure 4*d*) the crust is thinned by largely eliminating seismic layer 2; the crust consists of a few hundred meters of low-velocity material overlying a normal seismic layer 3. The final model (Figure 4*e*) is a typical oceanic velocity structure with the thicknesses of both layers 2 and 3 reduced proportionately so that the total crustal thickness is only 2.6 km.

The two models (Figures 4*d* and 4*e*) with substantial thicknesses of seismic layer 3 (i.e., a layer with veloc-

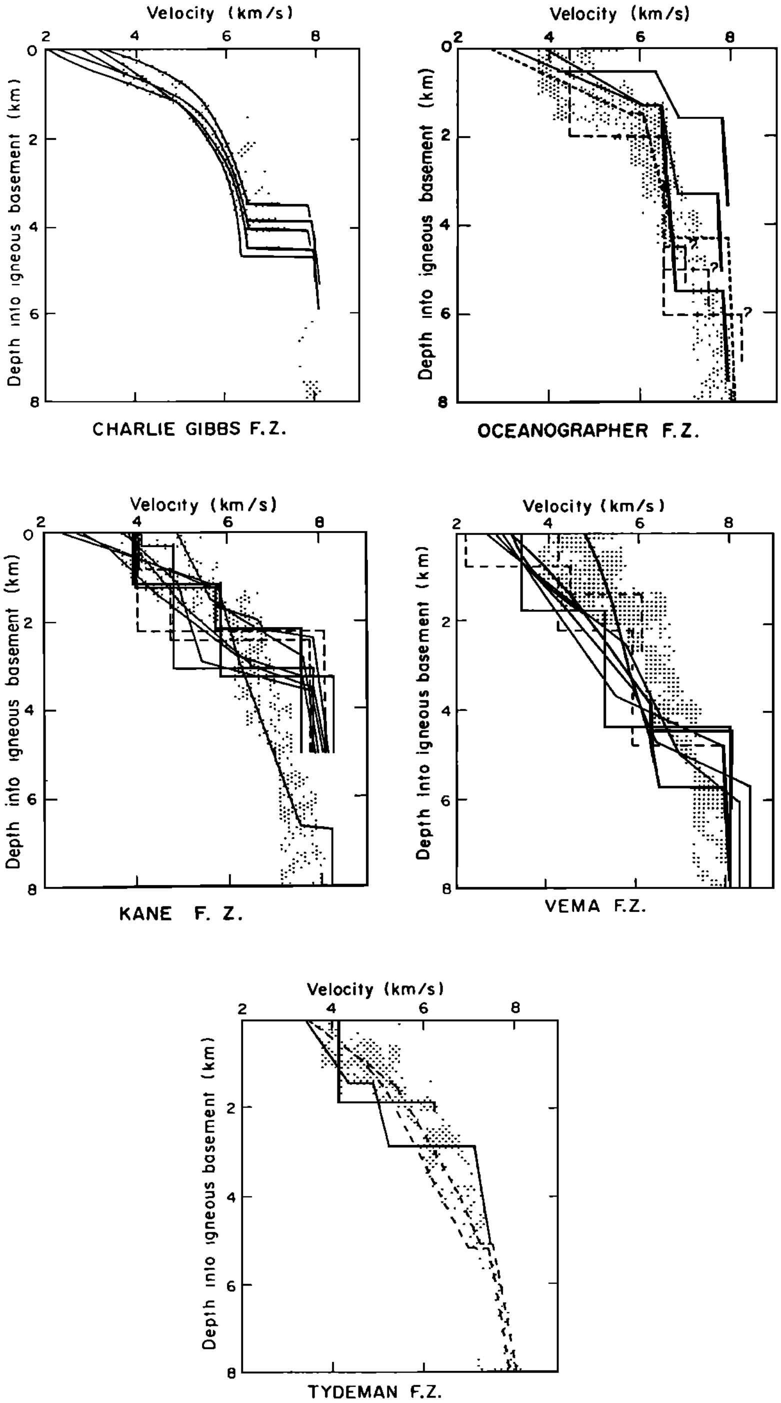

Figure 3. Compilation of seismic velocity functions at five large North Atlantic fracture zones. The shaded area shows the bounds on crustal structures typical of "normal" oceanic crust from *White* [1984]. Sources are as follows: Charlie Gibbs fracture zone from *Whitmarsh and Calvert* [1986]; Oceanographer fracture zone, dashed lines from *Fox et al.* [1976], solid lines from *Sinha and Louden* [1983], and dotted lines from *Ambos and Hussong* [1986]; Kane fracture zone, dashed lines from *Detrick and Purdy* [1980], solid lines from *Cormier et al.* [1984]; Vema fracture zone, dashed lines from *Ludwig and Rabinowitz* [1980], solid lines from *Detrick et al.* [1982], thick solid line from *Potts et al.* [1986*a*]; Tydeman fracture zone from *Potts et al.* [1986*b*] (solid lines, solutions from expanding spread profiles; dashed lines from ocean bottom seismometer lines).

ities of 6.5–6.9 km s^{-1} and low velocity gradients) clearly do not fit the observed data. In both cases the calculated travel times are too early and the crust-mantle triplication is shifted out to shot-receiver ranges well beyond what is actually observed. The best fitting model (Figure 4*b*) for this section, and one that is typical of the anomalous crust found at all major Atlantic fracture zones, consists of very low compressional wave velocities (~2.5 km s^{-1}) in the shallowest crust, relatively high velocity gradients throughout most of the thinned crustal section (1–2 s^{-1}), and the distinct absence of the high crustal velocities and low velocity gradients that characterize seismic layer 3.

Structural Variability Across and Along Major Atlantic Fracture Zones

Recent seismic studies not only have documented the existence of anomalous fracture zone crust, but also have begun to establish the scale of structural variability both along and across a number of large Atlantic fracture zones. These results have provided important constraints on possible mechanisms for creating this anomalous crust and associated features such as the large transverse ridges that border many large Atlantic fracture zones. Figures 5–9 show structural cross sections at five large Atlantic fracture zones plotted at similar scales with identical vertical exag-

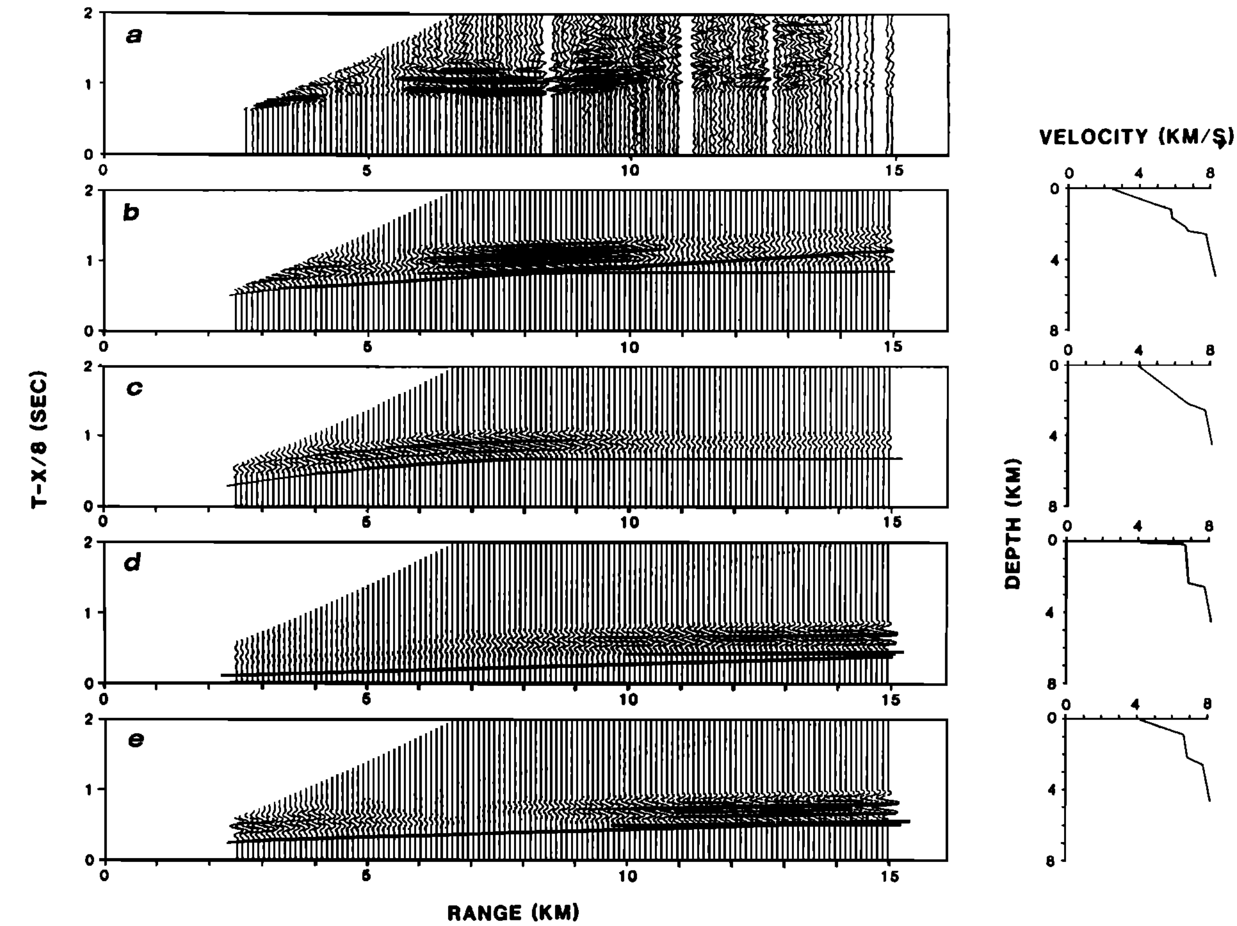

Figure 4. Comparison of (*a*) an ocean bottom hydrophone air gun profile shot along the Kane fracture zone trough with WKBJ synthetic seismograms calculated for the velocity models shown on the right [from *Cormier et al.*, 1984]: (*b*) synthetic seismogram that best approximates the main features of the observed profile, including the large amplitudes out to 4- to 5-km range, the sharp decrease in amplitude over the next 1–2 km and the large increase in amplitude again between 7- and 9-km range; (*c*) synthetic for a thin crust composed of a normal thickness layer 2 with seismic layer 3 completely absent; (*d*) synthetic for a thin crust composed mainly of seismic layer 3; (*e*) synthetic for a typical oceanic crustal section with the thicknesses of both layers 2 and 3 reduced proportionately.

geration to facilitate comparison. Compiled from published results, these sections were derived in nearly all cases by matching travel times using two-dimensional ray tracing techniques [e.g., *Cerveny and Ravindra*, 1971]. In one study [*Potts et al.*, 1986*b*], amplitude modeling was also employed [*Cerveny et al.*, 1977]. Although these solutions are nonunique, in cases where there are a large number of crossing ray paths, the crustal structure can be relatively well constrained with an estimated resolution of a few hundred meters vertically and a few kilometers horizontally.

Charlie Gibbs fracture zone [*Whitmarsh and Calvert*, 1986]. Crustal sections along and across the southern transform fault valley of the Charlie Gibbs fracture zone are shown in Figure 5. The crust beneath the transform valley is only 3–4.5 km thick, significantly less than the crust south of the fracture zone. On profile A this thinning occurs gradually over a distance of about 40 km south of the transform, although the thinnest, most anomalous crust is confined to a relatively narrow zone only about 12 km wide. The crust within the fracture zone is characterized by relatively low seismic velocities, high velocity gradients, and the noticeable absence of seismic layer 3, all typical of the crust found in other large Atlantic fracture zones. The transition between this anomalous fracture zone crust and more normal crustal velocities to the south occurs quite abruptly over a distance of only 1–2 km.

This experiment also provides constraints on the crustal structure of a transverse ridge (Hecate Bank) which forms the northern wall of the southern transform valley. This ridge was formed at a short central spreading segment within the Charlie Gibbs fracture zone and is similar morphologically to other large fracture zone transverse ridges. Moho shallows by about 2 km beneath the ridge, and crustal thicknesses and velocities are similar to those of normal oceanic

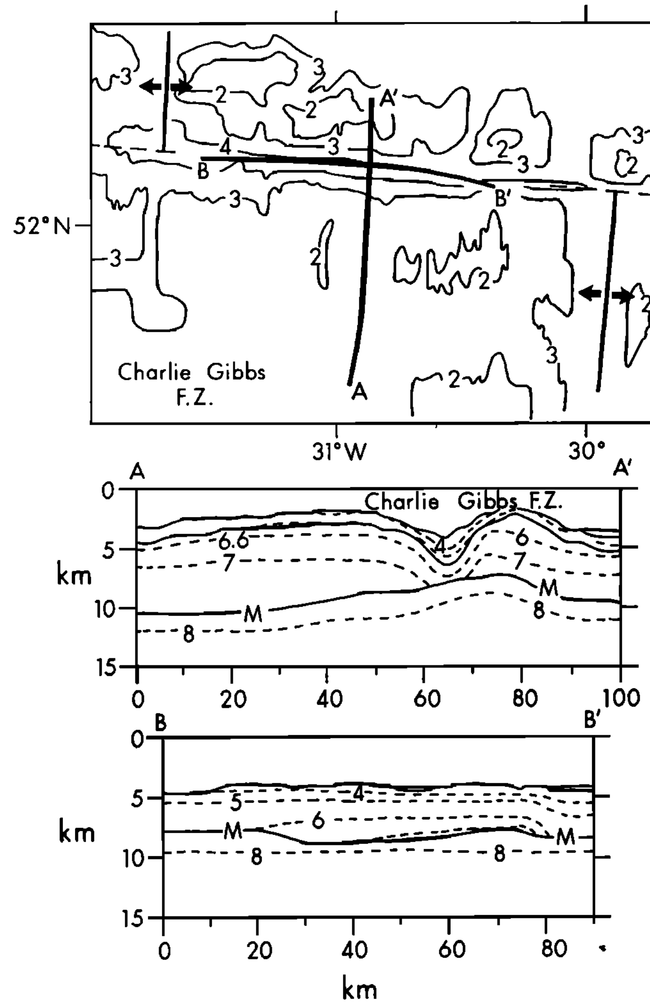

Figure 5. Structural cross sections at a vertical exaggeration of 2.5:1 across and along the Charlie Gibbs fracture zone, from *Whitmarsh and Calvert* [1986]. Profile locations are shown in the accompanying map, also from *Whitmarsh and Calvert* [1986]. Arrows indicate the location of the Mid-Atlantic Ridge rift valley; the dashed line marks the fracture zone trough. Depth contours are in kilometers; seismic velocity contours in profiles are in kilometers per second.

crust, suggesting that this is an uplifted section of essentially normal oceanic crust [*Whitmarsh and Calvert*, 1986].

Oceanographer fracture zone [*Sinha and Louden*, 1983; *Ambos and Hussong*, 1986]. A compilation of published crustal sections along and across the Oceanographer fracture zone is shown in Figure 6. *Sinha and Louden* [1983] have published four refraction lines from this area: profile B across the Oceanographer fracture zone at the western ridge-transform intersection, profiles C and D across the fracture zone about 30 km further west, and profile E within the Oceanographer transform valley. *Ambos and Hussong* [1986] have published two lines, profile A across the Oceanographer transform about midway between the two ridge-transform intersections, and profile F along the eastern transform valley.

All of these profiles reveal some crustal thinning beneath the Oceanographer fracture zone, although extremely thin crust is present only near the western ridge-transform intersection in profiles E and F. The thickest crust (9–10 km) is seen in profile B in the rift mountains south of the Oceanographer transform. In this profile the crust thins gradually over a distance of 25–30 km both to the north toward the Oceanographer fracture zone and to the south toward a small offset fracture zone known as fracture zone I. Some crustal thinning is also observed in profiles C and D beneath the Oceanographer fracture zone west of the ridge-transform intersection, although at its thinnest point the crust is only 1.5–2 km less than the average crustal thickness along the rest of these lines. This thinning appears to extend about 20 km to either side of the fracture zone. Profile A reveals an abrupt 1-km thinning of the lower crust (seismic layer 3) beneath the Oceanographer transform valley and an unusual thinning of layer 2 beneath the south wall of the transform. The latter results in the presence of high-velocity, lower crustal type material at shallow depths beneath the transform valley wall and is consistent with the abundant exposure of gabbroic and ultramafic rocks [*OTTER*, 1985] on these escarpments [*Ambos and Hussong*, 1986].

Profiles E and F indicate considerable structural heterogeneity along and across the Oceanographer transform valley, with total crustal thicknesses ranging from less than 2 km to more than 8 km. The thinnest crust (only 1–2 km thick) is found beneath the western ridge-transform intersection (near the intersection with profile B) and along the southern wall of the transform valley (sampled in profile F). In contrast, crustal thicknesses along most of the eastern half of the transform valley and beneath the northern transform valley wall (sampled by the eastern half of profile E) are >4.5 km to 8 km, comparable to that of most oceanic crust. There is no evidence for significantly thinner crust beneath the eastern ridge-transform intersection (profile F) despite the pronounced crustal thinning observed in profiles B and E near the western ridge-transform intersection.

Kane fracture zone [*Detrick and Purdy*, 1980; *Cormier et al.*, 1984; *Purdy and Detrick*, 1986; *Abrams et al.*, 1988]. Three crustal sections across the Kane fracture zone have been published (Figure 7): profile A along the Mid-Atlantic rift valley south of the eastern ridge-transform intersection [*Purdy and Detrick*, 1986], profile B across the Kane fracture zone and bordering transverse ridge about 90 km east of the ridge-transform intersection (originally published by *Detrick and Purdy* [1980] and reanalyzed by *Abrams et al.* [1988]), and profile C across the fracture zone 170 km east of the Mid-Atlantic Ridge [*Abrams et al.*, 1988]. In addition, structural control exists along a 200-km-long segment of the Kane fracture zone trough extending from the eastern portion of the transform

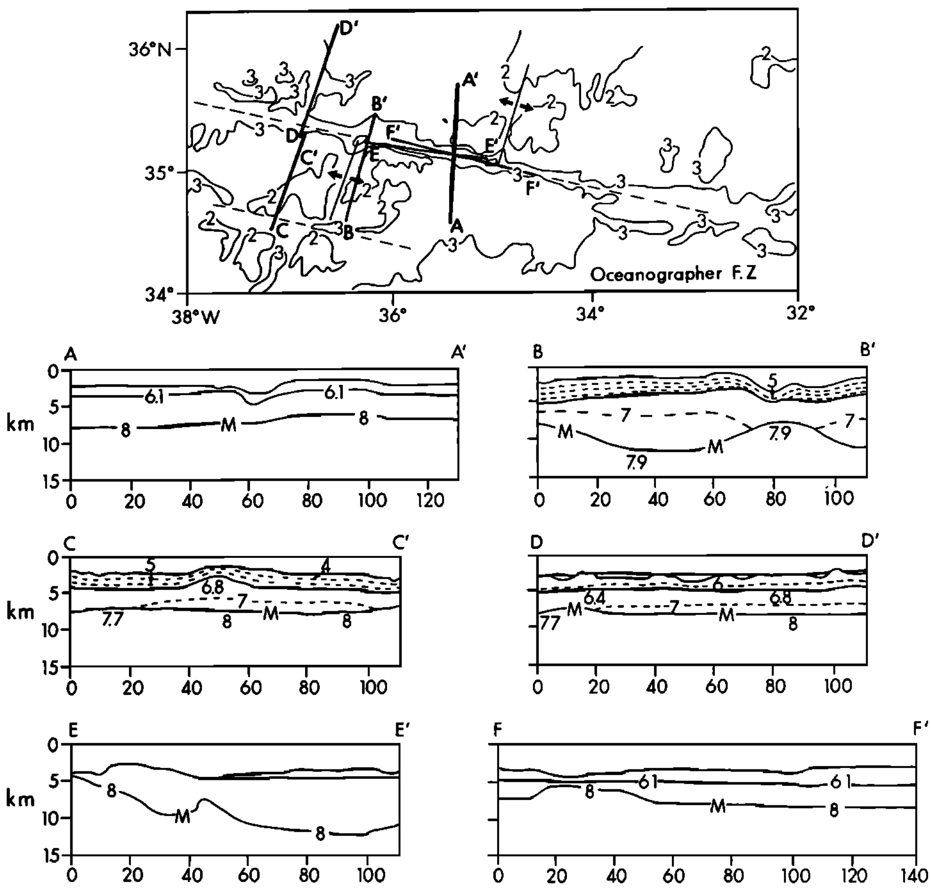

Figure 6. Structural cross sections at a vertical exaggeration of 2.5:1 across and along the Oceanographer fracture zone. Profiles A and F are from *Ambos and Hussong* [1986]; the remaining profiles are from *Sinha and Louden* [1983].

valley, across the ridge-transform intersection out onto crust of about 25 Ma (profile D [from *Cormier et al.*, 1984]).

The most striking feature of the Kane fracture zone is the common existence of extremely thin crust along most of the fracture zone trough. The fracture zone valley east of the ridge-transform intersection is underlain by crust only 1.5–2 km thick for a distance of almost 175 km (profile D). Beneath the nodal deep at the eastern ridge-transform intersection a specially designed delay time experiment indicates that Moho may lie less than 1 km below the seafloor [*Cormier et al.*, 1984], although mature crust of approximately normal thickness with upper mantle velocities of ~8 km s^{-1} is present further south along the Mid-Atlantic Ridge rift valley [*Purdy and Detrick*, 1986]. Thicker crust like that observed within parts of the Vema and Oceanographer fracture zones is present only in the eastern transform valley of the Kane and just east of the eastern ridge-transform intersection.

In most other respects the crustal structure at the Kane fracture zone is similar to that found at other large Atlantic fracture zones. Moho is elevated over a relatively broad zone 15–30 km wide centered on the fracture zone, resulting in a gradual thinning of the crust toward the fracture zone (profiles B and C). The thinnest crust, typically only 2–3 km thick, is confined to a relatively narrow zone 10–15 km wide beneath the fracture zone valley and the south-facing wall of the bordering transverse ridge. The transition to this highly anomalous crust occurs quite abruptly in profile B and more gradually in profile C. In both cases this transition is marked by a significant thinning of the low-gradient 6.1–7.2 km s^{-1} refractor associated with seismic layer 3. Within the fracture zone itself, crustal velocities are unusually low, and crustal velocity gradients are high (1.5–2 s^{-1}) throughout almost the entire crustal section.

The crustal structure of the Kane fracture zone transverse ridge in profiles B and C has been carefully modeled by *Abrams et al.* [1988]. Crustal thicknesses beneath most of the transverse ridge are close to 5 km, only 1–1.5 km less than crust flanking the transform and significantly thicker than the crust found beneath the adjacent fracture zone trough. The velocity structure of the crust comprising the transverse ridge is generally similar to normal oceanic crust in that it appears to have a low-gradient zone in the lower part of the section with layer 3–type velocities and a well-developed crust-mantle boundary. The south wall of the transverse ridge is characterized by relatively high upper crustal velocities (4.5–4.8 km s^{-1}), especially in comparison with the anomalous crust in the fracture zone trough immediately to the south.

Vema fracture zone [*Ludwig and Rabinowitz*, 1980; *Detrick et al.*, 1982; *Potts et al.*, 1986*a*; *Louden et al.*,

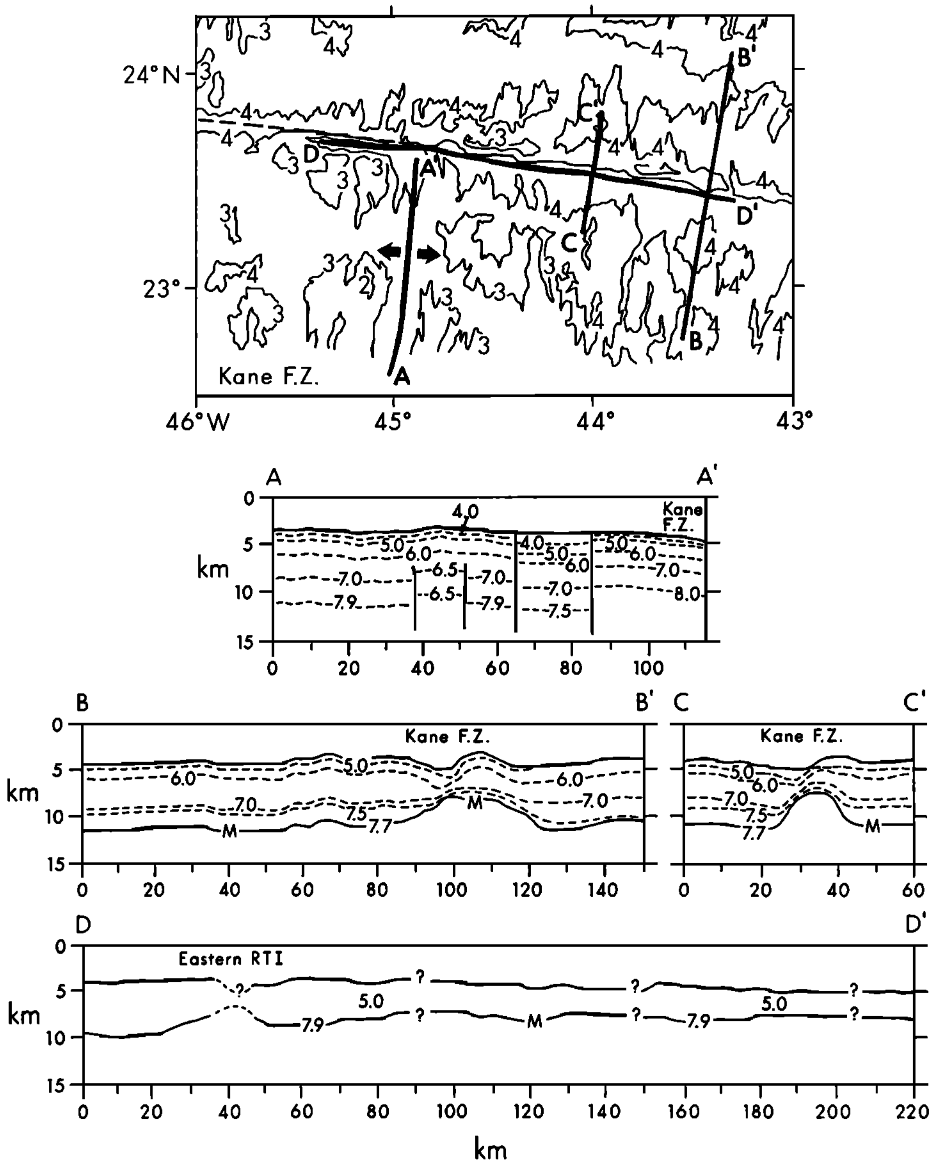

Figure 7. Structural cross sections at a vertical exaggeration of 2.5:1 across and along the Kane fracture zone. Profile A is from *Purdy and Detrick* [1986]; profiles B and C are from *Abrams et al.* [1988], and profile D is from *Cormier et al.* [1984]. Base map is redrawn from *Tucholke and Schouten* [1989]. RTI, ridge-transform intersection.

1986]. The crustal structure beneath the Vema fracture zone and its flanking transverse ridge has been determined from refraction profiles along and across the transform valley (Figure 8). Profile A is from *Louden et al.* [1986]; profiles B and C are from *Potts et al.* [1986*a*]. In addition, sonobuoy and OBS refraction lines have been shot along the transform valley east of the ridge-transform intersection [*Ludwig and Rabinowitz*, 1980; *Detrick et al.*, 1982].

The region of anomalous crust associated with the Vema fracture zone extends about 20 km to either side of the axis of the transform valley, with the crust gradually decreasing in thickness as the fracture zone is approached. Unlike that in the Kane fracture zone, the thinnest crust is found beneath the edges of the Vema transform valley. Under the axis of the transform valley the crust is only slightly thinner than normal, but it is characterized by the highly anomalous velocity structure (low velocities, absence of layer 3) documented at several major Atlantic fracture zones. The median valley north of the Vema transform is underlain by crust 3–3.5 km thick with low upper mantle velocities that increase as the fracture zone is approached [*Louden et al.*, 1986].

The transverse ridge that forms the south wall of the Vema transform on profiles A and B is essentially normal in its seismic velocity structure and overall thickness and is associated with a broad upwarping in the Moho [*Potts et al.*, 1986*a*]. Its structure is thus similar to the transverse ridges associated with both the Kane and Charlie Gibbs fracture zones.

Tydeman fracture zone [*Calvert and Potts,* 1985; *Potts et al.,* 1986*b*]. The only large-offset Atlantic fracture zone that has been studied off-axis, on relatively old crust, is the Tydeman fracture zone located in the central North Atlantic near 36°N. Two separate experiments were carried out: the first over 75-Ma crust with two-ship multichannel seismic expanding spread profiles [*Calvert and Potts*, 1985], and the second over 55-Ma crust using OBS and sonobuoys [*Potts

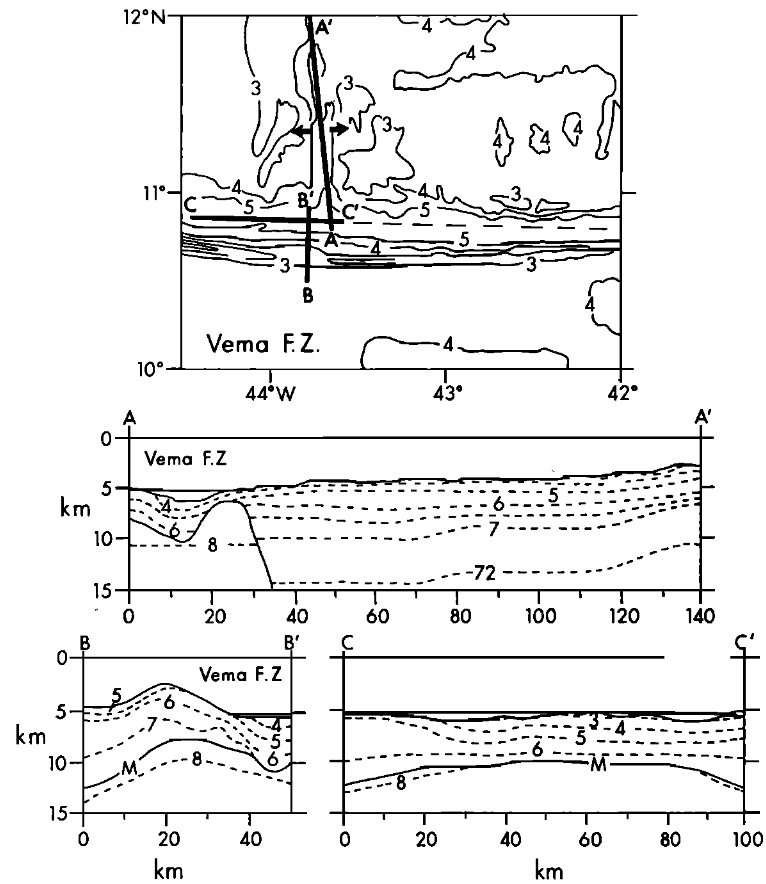

Figure 8. Structural cross sections at a vertical exaggeration of 2.5:1 across and along the Vema fracture zone. Profiles B and C are from *Potts et al.* [1986*a*]; profile A is from *Louden et al.* [1986].

et al., 1986*b*]. Figure 9 shows two crustal sections across the fracture zone constructed from the latter experiment in an area where the age difference across the fracture zone is about 6 Ma.

The crust thins by 1–1.5 km beneath the Tydeman fracture zone but is still close to 5 km thick. Upper crustal velocities within the fracture zone are lower than those in normal crust of comparable age (3.5–3.8 km s^{-1} compared with 4.0–5.8 km s^{-1}), and layer 3 appears to be largely missing. Within the lower crust, material with seismic velocities of 7.2–7.5 km s^{-1} extends over a zone about 15 km wide, comparable to the width of crustal thinning associated with the fracture zone. These unusual velocities (interpreted as "mantle" by *Calvert and Potts* [1985]) have not been reported from the younger portions of most large Atlantic fracture zones. Away from the fracture zone, normal mantle velocities of about 8 km s^{-1} are found.

SMALL-OFFSET FRACTURE ZONES AND NONTRANSFORM DISCONTINUITIES

Although their importance has been recognized only relatively recently with the advent of high-resolution bathymetric mapping of the Mid-Atlantic Ridge [*Macdonald et al.*, 1991; *Sempere et al.*, 1992], small-offset transforms and various nontransform offsets are the most common ridge axis discontinuity along the Mid-Atlantic Ridge. Relatively few seismic constraints exist on this class of ridge axis discontinuity (Table 1). Seismic refraction data are available from *White et al.*'s [1984] reanalysis of *Sinha and Louden*'s [1983] line C1 across fracture zone I, a small-offset (25 km) transform south of the Oceanographer fracture zone, and from *White and Matthew*'s [1980] experiment across an unnamed (approximately 10 km offset) fracture zone in the northeast Atlantic near 45.5°, 21°W. The best studied small-offset Atlantic fracture zone is the Blake Spur fracture zone, a 12-km offset located on 140-Ma seafloor southwest of Bermuda which has been investigated using both multichannel seismic reflection and two-ship expanding spread profiling techniques [*Mutter et al.*, 1984, 1985; *McCarthy et al.*, 1988; *Minshull et al.*, 1991]. However, the strongest evidence for the existence of anomalously thin crust at even small ridge axis discontinuities has come from recent detailed two-dimensional mapping of the gravity field over sections of the Mid-Atlantic Ridge [*Kuo and Forsyth*, 1988; *Lin et al.*, 1990; *Morris and Detrick*, 1990; *Blackman and Forsyth*, 1991].

Seismic Studies

A compilation of published velocity-depth profiles at small Atlantic fracture zones is shown in Figure 10; Figure 11 shows structural cross sections of both frac-

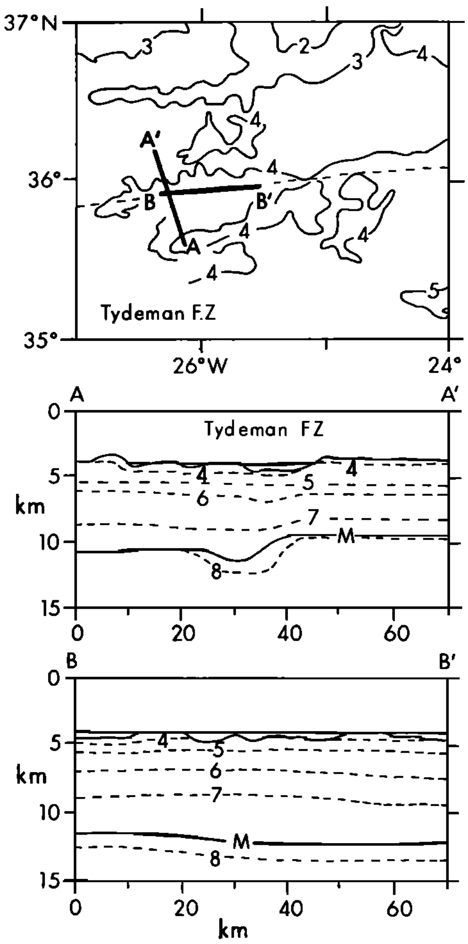

Figure 9. Structural cross sections at a vertical exaggeration of 2.5:1 across and along the Tydeman fracture zone, from *Potts et al.* [1986*b*].

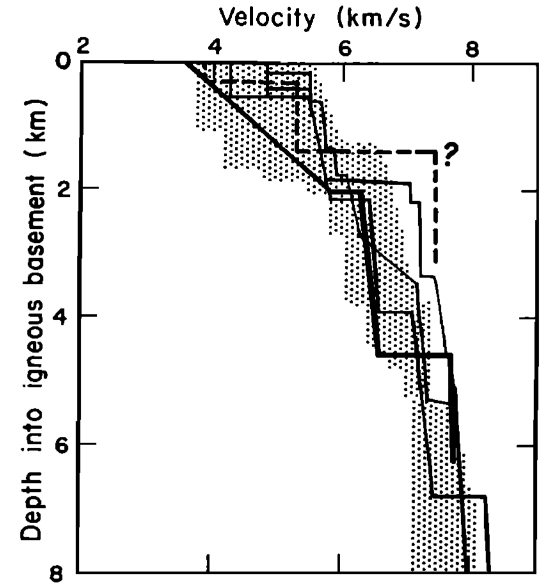

Figure 10. Compilation of seismic velocity functions for small-offset North Atlantic fracture zones. The shaded area is the same as in Figure 3. Dashed line, small-offset, unnamed fracture zone at 45.5°N, 21°W [*White and Matthews*, 1980]; thick solid line, from fracture zone I [*White et al.*, 1984]; thin solid line, Blake Spur fracture zone [*Minshull et al.*, 1991].

ture zone I and the Blake Spur fracture zone. The unnamed fracture zone studied by White and Matthews is characterized by a crustal layer with a velocity of 5.3 km s^{-1} underlain at a depth of only 2.1 km by a 7.3–7.6 km s^{-1} refractor. Originally interpreted as "mantle," this refractor may correspond to material of similar velocity reported from the Tydeman and Blake Spur fracture zones. Crustal thicknesses within fracture zone I are about 4.5 km, somewhat less than the 5- to 6-km thick crust found between fracture zone I and the Oceanographer fracture zone to the north (Figure 11). The crust within fracture zone I, like that beneath many large Atlantic fracture zones, displays abnormally low seismic velocities (6.25–6.55 km s^{-1}). The anomalous crust associated with fracture zone I extends for about 20 km on either side of the fracture zone, similar to the lateral extent of the anomalous crust found at larger offset fracture zones. Finally, although not as well constrained as the crustal velocities, there is also evidence here for somewhat low mantle velocities (7.7–7.8 km s^{-1}) beneath the fracture zone.

The crustal structure of the Blake Spur fracture zone is constrained by both multichannel seismic reflection and refraction data. The multichannel reflection data were collected in two separate studies: the 1981 North Atlantic Transect (NAT) experiment [*NAT Study Group*, 1985] and a more recent two-ship U.S.-British multichannel seismic study [*Minshull et al.*, 1991]. *Mutter et al.* [1985] interpreted the original North Atlantic Transect profile as indicating a gradual thinning of the crust toward the fracture zone over a distance of several tens of kilometers. Immediately beneath the fracture zone, a strong, shallow, subhorizontal reflector is present ~1 s below basement (Figure 12). This event was interpreted by *Mutter et al.* [1984] as Moho, suggesting that the crust is less than half its normal thickness beneath the fracture zone. However, subsequent reprocessing of this profile by *McCarthy et al.* [1988] revealed a deeper reflector that is laterally continuous with the reflection Moho southeast of the fracture zone (Figure 12), leading *McCarthy et al.* [1988] to argue that the crust may actually be thicker than normal beneath the Blake Spur fracture zone.

Figure 11 shows a crustal velocity section across the Blake Spur fracture zone in this same area constructed from the interpretation of seven expanding spread profiles by *Minshull et al.* [1991]. The fracture zone is associated with relatively normal upper crustal velocities underlain by a 15-km-wide zone of anomalous 7.2–7.6 km s^{-1} velocities in the lower crust. Typical upper mantle velocities of >8 km s^{-1} occur at

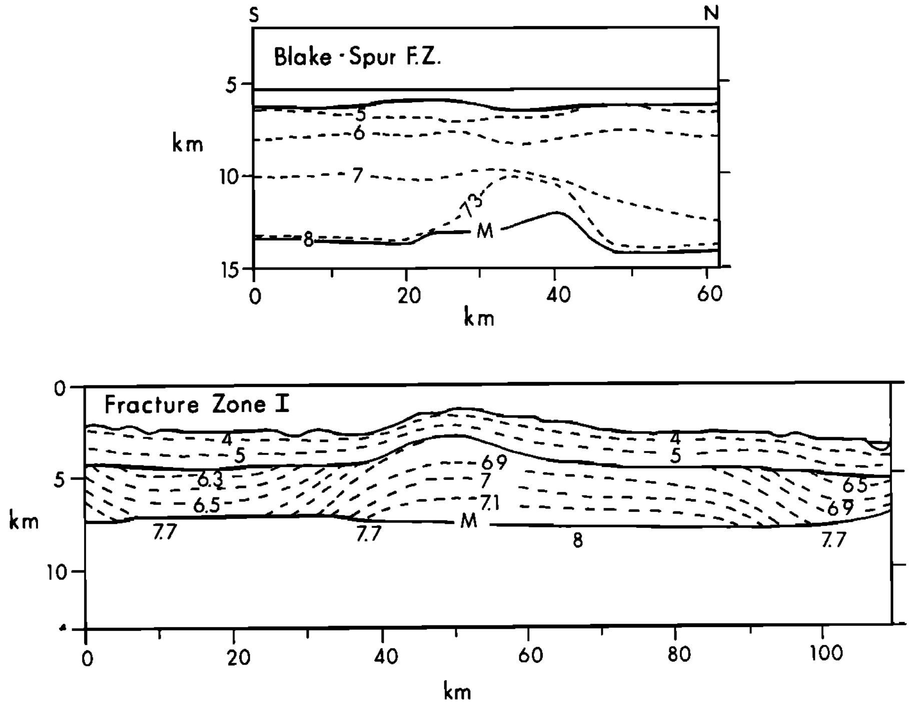

Figure 11. Structural cross sections at a vertical exaggeration of 2.5:1 across fracture zone I south of the Oceanographer fracture zone, from *White et al.* [1984], and the Blake Spur fracture zone, from *Minshull et al.* [1991].

depths of 12–14 km beneath the fracture zone, indicating that the deeper reflector identified by *McCarthy et al.* [1988] is associated with Moho. The shallower reflector of *Mutter et al.* [1984] may coincide with the top of the anomalous 7.2–7.6 km s^{-1} layer. *Minshull et al.* [1991] interpret this structure to have developed from a thin (2–4 km thick) original igneous crust followed by 15–30% serpentinization of a thick prism of upper mantle material.

Gravity Observations

A new perspective on the variation in crustal thickness along the Mid-Atlantic Ridge near small ridge axis discontinuities has come from recent detailed two-dimensional mapping of the gravity field over sections of the Mid-Atlantic Ridge [*Kuo and Forsyth*, 1988; *Lin et al.*, 1990; *Morris and Detrick*, 1990; *Blackman and Forsyth*, 1991]. In these studies, high-resolution multibeam bathymetric maps were used to remove the gravitational attraction of the water-crust and crust-mantle interfaces, assuming a constant thickness crust, in order to detect more subtle subseafloor density and crustal thickness heterogeneity. The resulting anomaly, called the mantle Bouguer anomaly (MBA), is due to the combined effects of variations in crustal thickness and changes in crustal or upper mantle density.

The MBA observed along the Mid-Atlantic Ridge shows a strong dependence on ridge-axis segmentation (Figure 13*a*). The MBA is typically more negative over the central portion of a tectonic segment but becomes

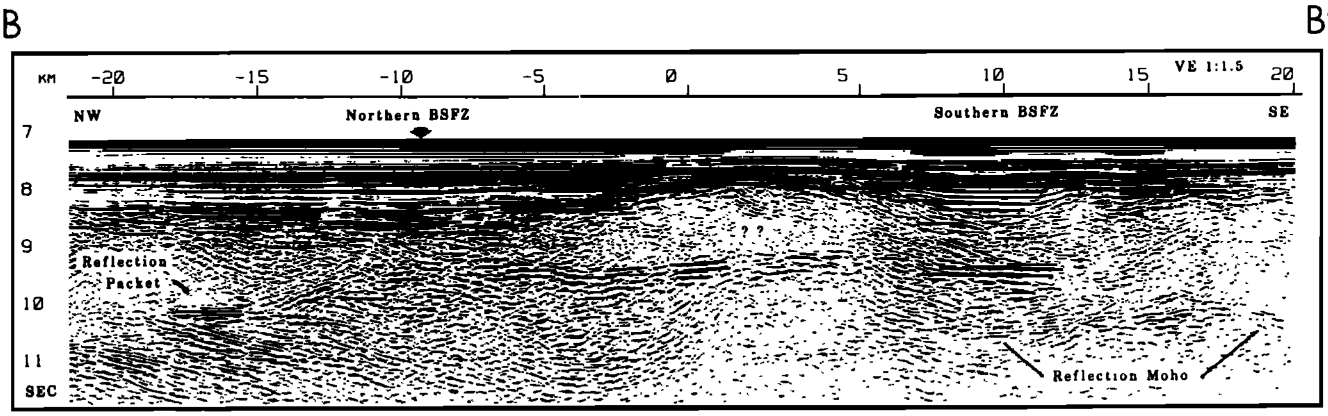

Figure 12. Migrated multichannel reflection profile NAT line 15 across the Blake Spur fracture zone, from *McCarthy et al.* [1988]. The strong, subhorizontal event at about 9.3 s was originally interpreted by *Mutter et al.* [1984] as Moho. However, the weaker event between 10 and 10.5 s correlates better with the position of the Moho shown in Figure 11. The shallower event may correspond with the top of the 7.3 km s^{-1} layer.

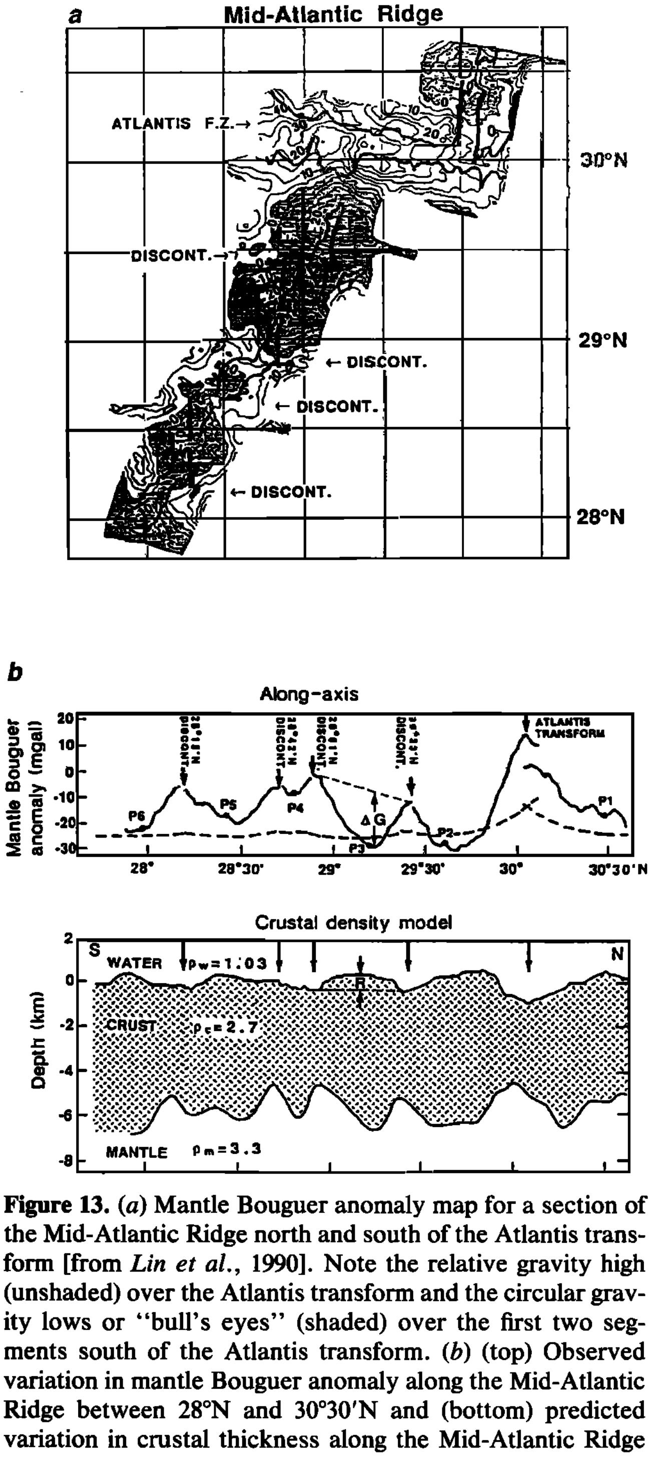

Figure 13. (*a*) Mantle Bouguer anomaly map for a section of the Mid-Atlantic Ridge north and south of the Atlantis transform [from *Lin et al.*, 1990]. Note the relative gravity high (unshaded) over the Atlantis transform and the circular gravity lows or "bull's eyes" (shaded) over the first two segments south of the Atlantis transform. (*b*) (top) Observed variation in mantle Bouguer anomaly along the Mid-Atlantic Ridge between 28°N and 30°30'N and (bottom) predicted variation in crustal thickness along the Mid-Atlantic Ridge required to explain the observed mantle Bouguer anomaly variation [from *Lin et al.*, 1990]. Note that both the Atlantis transform and the small nontransform offsets at 29°23'N, 28°51'N, 28°42'N, and 28°15'N are associated with anomalously thin crust.

more positive toward transform and nontransform segment boundaries. These MBA gravity lows, sometimes referred to as gravity "bull's eyes," are typically centered over tectonically defined spreading center segments and have wavelengths of 20–75 km, amplitudes of up to 50 mGal, and dimensions that are proportional to segment length [*Lin et al.*, 1990].

Although gravity data alone are insufficient to distinguish the relative contributions of crustal thickness variations or changes in crustal or upper mantle densities to these anomalies, there is evidence, based on both theoretical and observational grounds, that these anomalies arise primarily from variations in crustal thickness. *Lin and Phipps Morgan* [1992] have shown, for example, that the combined effects of mantle thermal expansion and melt extraction can explain only about half of the along-axis variation in MBA observed near the Atlantis transform in the central North Atlantic. They predict that these mantle effects will be even less important at smaller nontransform offsets. Refraction work on the southern Mid-Atlantic Ridge between 33°S and 34°S confirm that the amount of the along-axis crustal thickness variation (~3–4 km) is consistent with that inferred from the amplitude of the observed MBA [*Tolstoy et al.*, 1992].

Figure 13*b* shows the magnitude of along-axis crustal thickness variation required to explain the observed MBA along a 250-km-long section of the Mid-Atlantic Ridge south of the Atlantis transform. This section of ridge includes four small nontransform discontinuities as well as the large-offset Atlantis transform. The gravity data are consistent with along-axis variations in crustal thickness of at least 2 km from the middle to the ends of segments. Note that the magnitude of crustal thinning associated with the small nontransform offsets is comparable to that observed near the much larger offset Atlantis transform. Similar observations have been reported by *Kuo and Forsyth* [1988], *Morris and Detrick* [1990], and *Blackman and Forsyth* [1991]. There is also evidence in MBA maps from the Mid-Atlantic Ridge for a significant asymmetry in crustal structure across the rift valley, with the crust created nearest the ridge-transform intersection ("inside-corner") associated with a positive MBA (suggesting thinner crust) and the crust located on the opposite side of the rift valley associated with an MBA low (suggesting thicker crust). Thus the both seismic and gravity data suggest that the crust created within even small Atlantic fractures is anomalously thin and similar in some respects to crust formed within larger Atlantic fracture zones.

SUMMARY OF THE CRUSTAL STRUCTURE OF NORTH ATLANTIC FRACTURE ZONES

On the basis of the results reviewed above, we can summarize what is known about the seismic structure of North Atlantic fracture zones and smaller nontransform offsets as follows:

1. The crust within large Atlantic fracture zones is frequently very thin with upper mantle–type velocities (>7.6 km s^{-1}) found at depths of only 2–3 km below

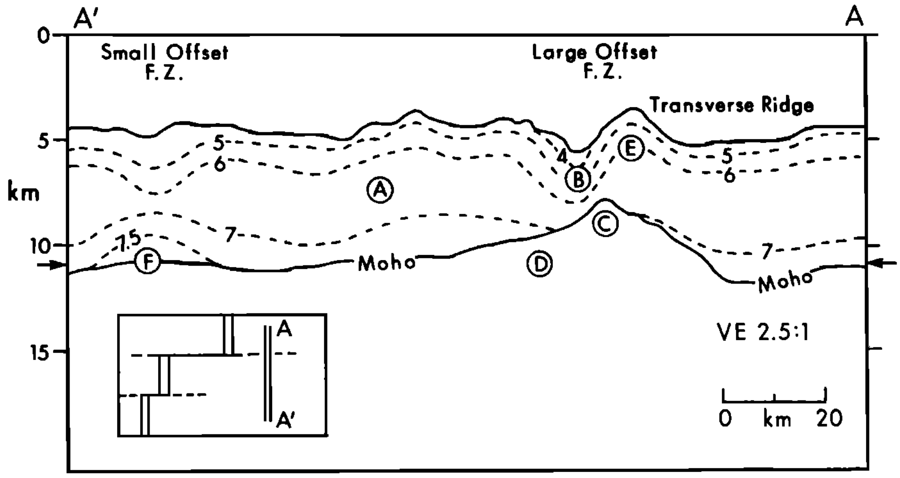

Figure 14. Generalized structural cross-section of a large Atlantic fracture zone based on seismic studies over the past decade. Fracture zone crust is anomalously thin compared to normal oceanic crust (A) and characterized by unusually low compressional wave velocities without a normal seismic layer 3 (B). The thinnest crust is confined to a relatively narrow zone within the transform valley (C), but a more gradual crustal thinning may extend up to 20 km or more from the fracture zone (D). Transverse ridges bordering the fracture zone are associated with relatively normal crustal velocities and thicknesses and a broad upwarping of the Moho that is consistent with their interpretation as uplifted sections of essentially normal oceanic crust (E). Partial serpentinization of the upper mantle occurs under some fracture zones (F). See text for discussion.

the seafloor. However, more normal crustal thicknesses (4.5–5 km) are also present along many parts of these fracture zones.

2. The crust within large Atlantic fracture zones is characterized by unusually low compressional wave velocities and relatively high velocity gradients (1–2 s^{-1}) throughout the entire crustal section. Seismic layer 3 is either thin or absent in the fracture zone, although material with seismic velocities in the range 7.2–7.5 km^{-1} is sometimes present at the base of the crust, especially along the older sections of some fracture zones.

3. Large apparent variations in total crustal thickness occur both along and across major Atlantic fracture zones. Anomalously thin crust has been found in the active transform, along the nontransform fracture zone valley, and beneath ridge-transform intersections. The thinnest crust is often found beneath the deepest portion of the fracture zone valley (e.g., Kane), although in some cases (e.g., Vema) it is located near the edges of the valley. A more gradual crustal thinning may extend up to 20 km or more from the fracture zone. The transition between the very thin crust beneath the fracture zone valley and the more normal crustal thicknesses on either side is typically very abrupt, occurring over a distance of at most a few kilometers.

4. The large, anomalously shallow transverse ridges that frequently border large Atlantic fracture zones are associated with relatively normal crustal velocities and thicknesses and a broad upwarping of the Moho relative to the surrounding crust.

5. Both seismic and gravity data suggest that smaller transform and nontransform offsets on the Mid-Atlantic Ridge are also associated with anomalously thin crust. The lateral extent of the anomalous crust associated with small-offset fracture zones is comparable to that found at major Atlantic transforms.

ORIGIN OF THIN CRUST IN ATLANTIC FRACTURE ZONES

The establishment of the first-order crustal structure of oceanic fracture zones is one of the major accomplishments of marine seismology over the past decade. The similarity in the results from nearly a dozen different seismic experiments at five major Atlantic fracture zones is quite remarkable, especially given the complex topography and tectonics of these major shear zones. Figure 14 shows a hypothetical structural cross section of a large Atlantic fracture zone based on the observations summarized in the previous section. The features shown in this figure (the narrow zone of thin, low-velocity crust beneath the fracture zone valley, the more gradual crustal thinning that extends up to 20 km or more from the fracture zone, and the broad upwarping of the Moho with normal crustal velocities and thicknesses beneath the bordering transverse ridge) characterize the crustal structure of all large Atlantic fracture zones that have been studied in any detail.

Although their seismic structure is now relatively well known, the geological interpretation of the anomalous crust found within Atlantic fracture zones, and how it forms, is still controversial. The low upper crustal velocities in fracture zones have generally been attributed to a thin, highly fractured and altered basaltic and gabbroic section [*Detrick and Purdy*, 1980; *White et al.*, 1984] and, at shallow levels, to the presence of rubble and rock talus from the fracture zone valley walls [*Detrick et al.*, 1982]. The absence of a normal seismic layer 3 with velocities of 6.7 km s^{-1} and low or zero velocity gradients contributes to the anomalously low crustal velocities found in fracture zones. *White et al.* [1984] proposed that these low velocities (5–6 km s^{-1}) could also represent highly serpentinized peridotites, raising the possibility that in some places the mafic crust may be even thinner than the seismically determined crustal thickness. Laboratory data [*Christensen*, 1978] indicate that >60% serpentinization will be required to reduce the compressional wave velocity of peridotite to these values and that these altered rocks will have high (>0.34) Poisson ratios. *Calvert and Potts* [1985] have shown that at least in the case of the Tydeman fracture zone, observed Poisson ratios in fracture zone crust (0.26–0.29) are too low to be explained by serpentinite.

Material with seismic velocities in the range 7.2–7.5 km s^{-1} is sometimes present at the base of the crust, especially along the older sections of some fracture zones (e.g., the Tydeman and Blake Spur fracture zones). These velocities are significantly lower than typical upper mantle velocities (7.7–8.2 km s^{-1}) but are much higher than normal layer 3 velocities. This material could represent an overthickened seismic layer 3B in the fracture zone. Layer 3B has seismic velocities in this range [e.g., *Spudich and Orcutt*, 1980], and layered gabbros and peridotites at the base of the crustal section in the Oman ophiolite are also associated with *P* wave velocities of about 7.4 km s^{-1} and a Poisson ratio of 0.29 [*Christensen and Smewing*, 1981]. Alternatively, this material may represent partially serpentinized peridotites lying below a thin, highly fractured mafic crustal section [*Calvert and Potts*, 1985; *Minshull et al.*, 1991]. A velocity of 7.2–7.6 km s^{-1} at 2 kbar will require 15–30% serpentinization and be associated with a Poisson ratio of 0.30–0.31 [*Christensen*, 1978]. With the difficulty of observing shear waves in fracture zones, definitive in situ Poisson ratio determinations are not available to distinguish between these two different interpretations. However, the occurrence of these anomalous velocities along the older sections of fracture zones [*Calvert and Potts*, 1985; *Minshull et al.*, 1991] and the presence of extremely thin crust without this layer in the younger sections of many fracture zones [e.g. *Cormier et al.*, 1984] favor the interpretation of this material as partially serpentinized upper mantle peridotite which forms as a result of continued off-axis hydrothermal circulation through an originally thin, highly fractured mafic crustal section.

The presence of anomalously thin oceanic crust in Atlantic fracture zones and at smaller nontransform offsets has been attributed both to a reduction in magma supply near these ridge axis discontinuities [*Fox and Gallo*, 1984; *White et al.*, 1984] and to stretching and tectonic thinning of the crust by large-scale detachment faulting [*Mutter and Karson*, 1992]. The relative importance of these two processes in forming anomalous fracture zone crust is still being debated.

Fox and Gallo [1984] suggested that a reduction in magma supply near transforms will be caused by the juxtaposition of an older, colder plate edge across a ridge-transform intersection (the so-called transform fault effect). In support of their model they noted that seafloor depths at ridge-transform intersections and the magnitude of observed along-axis depth variations are related to the length (age offset) across the transform. However, this explanation cannot account for the magnitude of crustal thinning observed seismically at small ridge axis discontinuities, since the transform fault effect should be minimal for the small age offsets associated with these segment boundaries. Numerical modeling of the three-dimensional temperature structure at a ridge-transform intersection for a simple passive, plate-driven flow shows that conductive cooling across the transform will extend only a few kilometers from the fault, even for relatively large age offsets [*Forsyth and Wilson*, 1984; *Phipps Morgan and Forsyth*, 1988]. Thus a simple thermal edge effect model cannot explain the well-documented systematic thinning of the oceanic crust up to several tens of kilometers from many fracture zones.

The existence of a broad zone of reduced magma supply near ridge offsets can be explained by a focusing of mantle upwelling and melt production near the center of ridge segments [*Lin et al.*, 1990]. Focused upwelling near the middle of a segment will result in higher mantle temperatures and upwelling velocities, leading to the accretion of thicker crust near the middle of a segment compared with the ends of segments, where mantle temperatures will, on average, be lower and upwelling will be perturbed by the juxtaposition of an older, colder plate edge [*Phipps Morgan and Forsyth*, 1988; *Sparks and Parmentier*, 1993]. If mantle upwelling is dominated by passive, plate-driven flow, this effect will be most important at large-offset transforms. However, if the flow is driven more by buoyancy forces caused by thermal expansion of the upwelling mantle and compositional density reductions due to the extraction of partial melt, then both laboratory and theoretical studies suggest that focused upwelling may be as important at segments bounded by small offsets as at those bounded by large offsets [e.g., *Whitehead et al.*, 1984; *Lin and Phipps Morgan*, 1992; *Sparks and Parmentier*, 1993]. This could explain the existence of comparable degrees of crustal thinning at both large- and small-offset ridge discontinuities. *Lin and Phipps Morgan* [1992] have suggested that buoyancy-driven flow will be much more important at slow spreading ridges than at fast spreading ridges. If this hypothesis is correct, then thin crust should be found preferentially along ridge offsets at slow spreading ridges like the Mid-Atlantic Ridge.

In the past few years there has been a greater appreciation of the role that tectonic processes may play in creating the crustal heterogeneity observed along slow spreading ridges. *Karson* [1991] and *Mutter and Karson* [1992] have proposed that large-scale detachment faulting along the median valley walls can exhume lower crustal rocks, resulting in large, tectonically induced crustal thickness variations across and along the rift valley near ridge transform intersections. This process may explain the asymmetry in the MBA field observed across the rift valley near some ridge axis discontinuities [*Kuo and Forsyth*, 1988; *Lin et al.*, 1990; *Morris and Detrick*, 1990]. These faults will also provide permeable pathways along which water can flow into the uppermost mantle, leading to serpentinization of the ultramafic section that appears to be an important process in many oceanic transforms [e.g., *Bonatti*, 1976, 1978].

The relative importance of focused upwelling and tectonic processes in creating the anomalous crust observed at oceanic fracture zones in the North Atlantic is still being debated. However, it is clear that these processes are closely interrelated. The temperature structure of the underlying mantle will strongly influence tectonic processes through its effect on the depth to the brittle-ductile boundary and the thickness of the rheologically weak crustal section [*Shaw*, 1992]. Thus if upwelling is focused near the center of segments, the lithosphere near segment midpoints will be hotter and weaker than lithosphere near segment ends, favoring the development of large-scale faults and tectonic stretching near segment offsets. The extremely variable three-dimensional crustal structure documented in gravity and seismic data at North Atlantic fracture zones suggests strongly that faulting and hydrothermal alteration processes are superimposed on a larger-scale reduction in magma supply to form the thin, intensely fractured, and hydrothermally altered basaltic and upper mantle section found at these discontinuities.

SUMMARY AND FUTURE RESEARCH DIRECTIONS

Seismic studies over the past decade have established that large-offset transforms along the slow spreading Mid-Atlantic Ridge exhibit anomalous crustal structures that fall well outside the range of velocity structures typically associated with oceanic crust. The crust within major Atlantic fractures zones is less than 1–2 km thick in places, and the magnitude of the crustal thickness variation from the midpoint to the ends of spreading segments in the Atlantic can be as great as 3–4 km. Although fewer constraints exist on the crustal structure of smaller offset transforms and the numerous small nontransform offsets found on the Mid-Atlantic Ridge, available evidence suggests that these axial discontinuities are also associated with anomalously thin crust. In some places the magnitude of the crustal thinning appears to be comparable to that observed at large-offset transforms. The presence of anomalously thin crust at these ridge axis discontinuities is attributed to a reduced magma supply within a broad region near ridge offsets caused by a focusing of mantle upwelling beneath segment midpoints and tectonic dismemberment of the crust by large-scale detachment faults that preferentially form in the cold, brittle lithosphere near the ends of segments.

Some important questions remain to be answered about the nature and origin of the crust formed in oceanic fracture zones. To date, little information is available on the crustal structure of fast slipping transforms such as those found along the East Pacific Rise. The buoyancy forces that lead to focused mantle upwelling at slow spreading ridges are expected to be less important at fast spreading ridges [*Sparks and Parmentier*, 1993; *Lin and Phipps Morgan*, 1992]. This should result in a more two-dimensional upwelling pattern and a more uniform magma supply along a fast spreading ridge segment. Thus we would expect less variation in crustal thickness along-axis at the East Pacific Rise and a thicker igneous crust proximal to fast slipping transforms. Gravity data from the East Pacific Rise are consistent with this hypothesis [*Lin and Phipps Morgan*, 1992], but seismic data are needed to directly measure the crustal thickness and structure at a fast slipping transform. The importance of small nontransform offsets, at both slow and fast spreading ridges, has only been recently been recognized. Few direct constraints exist on the seismic structure of crust created at these small axial discontinuities, and it is still not known whether or not they display the same range of structures as large-offset fracture zones. Constraining the magnitude of crustal thickness variation at both fast slipping transforms and small nontransform offsets will be important in assessing the role of upper mantle flow and melt migration in creating this first-order heterogeneity in the structure of oceanic crust.

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