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FROM SEISMIC REFRACTION STUDIES

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TECHNICAL REPORT

Prepared for the Office of Naval Research
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Geophysical Research, Vol. 85, No. B7, pp. 3759-3777, July 10,
1980".

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Department of Geology and Geophysics
The Crustal Structure of the Kane Fracture Zone From Seismic Refraction Studies

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A detailed seismic refraction experiment was carried out across the Kane Fracture Zone near 24°N, 44°W using explosive and air gun sound sources and eight ocean bottom hydrophone receivers. The shooting lines and receivers formed a ‘T’ configuration across the fracture zone, with two receivers located about 50 km apart in the fracture zone trough and the remaining six receivers positioned 25–30 km apart on either side of the fracture zone. The crustal thicknesses and velocities observed at the receivers located north and south of the Kane Fracture Zone fall within the range of those typically observed for normal oceanic crust. There is no convincing evidence for significantly different crustal thicknesses or upper mantle velocities on either side of the fracture zone despite a 10-m.y. age difference. Anomalously thin crust is present beneath the Kane Fracture Zone trough with total crustal thicknesses of only 2–3 km, about half the thickness of normal oceanic crust. This crust is also characterized seismically by low compressional wave velocities (~4.0 km/s) at shallow depths and the absence of a normal layer 3 refractor. This anomalous crust extends over a width of at least 10 km. Dense, high-velocity mantle type material may also exist at shallow depths beneath the adjacent Kane Fracture Zone ridge. Results from other geological and geophysical studies of fracture zones suggest that this type of crustal structure may be typical of many Atlantic fracture zones. We propose that the anomalously thin crust found within these fracture zones is a primary feature caused by the accretion of a thinner volcanic and plutonic layer within the fracture zone. This anomalous crust, which probably is restricted to a zone no wider than a typical transform fault valley (~10 km) in most cases, is inferred to consist of a few hundred meters of extrusive basalts and dikes overlaying about 2 km of gabbro and metagabbro, possibly interbedded with ultramafics. This anomalously thin crustal section may be extensively fractured and brecciated at shallow levels by faulting in the active transform domain. A relatively narrow zone of thin crust within fracture zones can explain a number of geological and geophysical characteristics of fracture zones including the depth of the transform fault valley and the exposure of deep crustal and upper mantle rocks in the walls of fracture zones.

INTRODUCTION

Marine geophysical surveys over the past decade have shown that transform faults and fracture zones are present along the entire length of the midocean ridge system. In well-surveyed areas, such as the North Atlantic, transform faults are found every 50–100 km (H. Schouten and K. Klitgord, manuscript in preparation, 1980), making them one of the most common tectonic features in the ocean basins. Although most transform faults are often treated as narrow zones of simple strike slip motion, they are actually tectonically complex features with widths of 10–20 km or more and relief of up to several thousands of meters. Morphologically, fracture zones display many similar characteristics. The axis of a ridge/ridge transform is typically defined by a deep, linear trough a few kilometers (~10 km) wide and a few hundred meters deep. This trough remains noticeably deeper than the surrounding sea floor for tens of millions of years and can often be identified in even the oldest portions of the ocean basins. Frequently, bordering this trough are ridges or escarpments rising as much as several kilometers above the transform valley floor. These linear fracture zone ridges parallel the axial trough and extend for distances of several hundred kilometers. Transform faults with small offsets (~20 km) like Fracture Zone A in the French-American Mid-Ocean Undersea Study (Famous) area [Detrick et al., 1973; Choukroune et al., 1978] are typically characterized by relief of about 1 km and widths of less than 20 km, while larger offset fracture zones (~100 km) like the Vema, Romanche, and Oceanographer fracture zones [Heezen et al., 1964; Fox et al., 1969; van Andel et al., 1971] display relief in excess of 3 km and widths of 20 km or more.

Few good geophysical constraints exist on the crustal structure of oceanic fracture zones. Crustal models have been published for the Vema [Robb and Kane, 1975] and Romanche [Cochran, 1973] fracture zones; however, these models were based on gravity data without refraction control. The only published refraction study of a major Atlantic fracture zone is that of Fox et al. [1976] from the Oceanographer Fracture Zone. However, no mantle arrivals were observed in this experiment, and consequently, the total crustal thickness within the fracture zone could not be determined. Better geophysical control on the deeper crustal structure of fracture zones is essential in order to construct more realistic geological models for fracture zones.

Here we describe the results of a seismic refraction experiment carried out in October–November 1977 on a portion of the Kane Fracture Zone near 23°30′N, 44°W (Figure 1). The principal scientific objectives of this experiment were (1) to define the crustal structure of the major morphological features within a large Atlantic fracture zone and (2) to determine whether or not significant age-dependent differences in crustal and upper mantle velocity structure exist on either side of the fracture zone.

KANE FRACTURE ZONE

The Kane Fracture Zone is one of the largest and best mapped fracture zones in the central North Atlantic (Figure...
1. It has been traced from its intersection with the Mid-Atlantic Ridge near 24°N out beyond 80 m.y. B.P. on both flanks of the ridge crest [Rabinowitz and Purdy, 1976; Purdy et al., 1979]. A bathymetry map of the Kane Fracture Zone near its intersection with the Mid-Atlantic Ridge is shown in Figure 1. The sources of the bathymetry data used in compiling this map are R/V Vema, R/V Robert D. Conrad, R/V Knorr, R/V Atlantis II, OSS Discoverer, USNS Kane, D/V Glomar Challenger, HMS Stellius, R/V Kana Keoki, R/V Akademik Kurchatov, R/V Chain, and R/V Washington [van Andel and Bowin, 1968].

The Kane Fracture Zone in this area consists of a deep, narrow trough varying in depth from 4500 m in the active transform section to more than 6000 m on the ridge flanks. The average trend of the transform section of the fracture zone trough is 100°, essentially perpendicular to the southern median valley and in agreement with the present spreading direction predicted by the pole of Minster and Jordan [1978]. East of 45°W the fracture zone is bordered on the north by a ridge which rises about 2 km above the fracture zone trough. The western end of this ridge coincides with the intersection of the southern median valley with the fracture zone trough. The eastern end is located at longitude 42°W. The width of the fracture zone trough and ridge together average about 15–20 km.

Magnetic anomalies identified on either side of the Mid-Atlantic Ridge (MAR) south of the Kane Fracture Zone indicate an average spreading rate during the past 5 m.y. of 14-mm/yr half rate [Purdy et al., 1978]. However, there has been a consistent 7:10 east-west asymmetry in spreading which began 35 m.y. B.P. [Schouten et al., 1979]. This asymmetric spreading and a westward jump in the MAR north of the Kane Fracture Zone about 11 m.y. B.P. have gradually increased the offset of the Kane Fracture Zone from 100 km 20 m.y. ago to its present 160-km left lateral offset.

THE EXPERIMENT

The seismic refraction experiment reported here was located on an inactive portion of the Kane Fracture Zone approximately 60 km east of its intersection with the southern
rift valley (Figure 1). The receivers used were ocean bottom hydrophones (OBH's) developed at the Woods Hole Oceanographic Institution [Koelsch and Purdy, 1979]. Eight OBH's were deployed in a T configuration across the fracture zone (Figures 1 and 2). Three instruments (OBH 1, 2, and 3) were located south of the Kane Fracture Zone on sea floor about 7 m.y. old. Three additional instruments (OBH 6, 7, and 8) were deployed north of the fracture zone on 17 m.y.-old crust. The remaining two receivers (OBH 4 and 5) were placed about 50 km apart within the fracture zone trough.

This experimental design was chosen to provide good seismic constraints on the crustal structure on either side of the Kane Fracture Zone as well as beneath the fracture Zone suggesting the presence of a small right lateral offset at about 24°40'N. The WNW trending trough north of the Kane Fracture Zone between 43°W and 45°W is probably the eastern extension of this fracture zone (Figure 1). Another small offset fracture zone may also exist south of the Kane Fracture Zone near 22°30'N. Since we wanted to ensure that the Kane Fracture Zone was the only major structural discontinuity within the bounds of the experiment, we were careful not to cross these smaller fracture zones during the course of the experiment.

Shots were fired along the lines joining OBH 1 and 8 and OBH 4 and 5 using both explosive charges and a 300- or 1000-in. 3 air gun. The explosive sound source was TNT with charge sizes of 14.5 and 116 kg (32 and 256 lb) and a shot spacing of about 1.7 km and 4.8 km, respectively. A few 232-kg (512 lb) charges were fired at the ends of each shooting line. The air gun data were collected primarily to improve resolution of the shallow crustal structure at each instrument. The shot spacing for the 300- and 1000-in. 3 air guns was 30 m and 100 m, respectively.

**DATA PROCESSING**

All eight receivers recorded excellent quality data (Figures 3, 8, and 10). Charge sizes of 14.5 kg produced clear impulsive first arrivals out to 50 km, and the larger 116- and 232-kg shots were recorded out to ranges of 150 km. The quality of the 1000-in. 3 air gun data was also excellent; however, the amplitude of the 300-in. 3 air gun signal was inadequate, and these data were generally uninterpretable. The 1000-in. 3 air gun data were only obtained for the three receivers north of the fracture zone.

Receiver locations were determined from a combination of satellite navigation and minimum ship-receiver distances estimated by ranging on the acoustic transponders at each OBH. Errors in receiver location are estimated to be less than 0.25 km. Shot times were recorded on the R/V Atlantis II using a towed hydrophone streamer. Shot depths were estimated from the observed bubble pulse period. These shot depths were used to correct the shot instants for shot-to-ship travel time (shot-instant correction) assuming a near-surface water velocity of 1537.9 m/s. Shot-receiver ranges were determined from direct water wave travel times using an assumed average water velocity of 1513.7 m/s. Both this velocity and the near-surface water velocity were determined from a nearby CTD cast. Record sections were constructed for each receiver and used to pick refracted wave arrival times. The high signal/noise ratio and clear, impulsive nature of most arrivals generally permitted these travel times to be picked to within 0.025 s.

Although the topography along the shooting line was relatively subdued (except where crossing the fracture zone, Figure 2), topographic corrections to the observed travel times were still necessary. The two standard methods for computing topographic corrections were applied, and the results compared. In the first, a horizontal reference level was chosen at the mean depth of the shooting line, and all travel times were corrected for variations of the sea floor above and below this datum [Whitmarsh, 1975]. The implied assumption in this correction is that all reflectors beneath the sea floor are horizontal. The magnitude of this correction is small, and thus errors in the chosen value of the ray parameter result in small errors in the value of the correction. However, it is very dependent on the assumed value of the velocity of the shallowest layer, which is generally poorly known. In the second type of correction the delay time corresponding to the path between the shot and the sea floor at the ray entry point is removed from the total travel time. This correction effectively places all shots and receivers on the sea floor, and the implicit assumption here is that all reflectors parallel the sea floor. This correction is independent of the velocity of the shallowest layer, but its magnitude is much larger, and thus errors in the chosen value of the ray parameter result in larger errors in the correction than in the first method.

Neither technique produced completely smooth travel time–distance relationships even for data outside the fracture zone province. This undoubtedly reflects the inaccuracies in computing the topographic corrections as well as lateral heterogeneity in crustal structure. The major source of computational error in both techniques is in the estimation of the depth of the ray entry point. This depends upon an accurate knowledge of both the sea floor topography and the ray parameter p. Best estimates of p were made by picking first arrivals on the uncorrected record sections; depths at the ray entry points corresponding to these values of p were then picked from the original 3.5-kHz echo-sounding records. At some locations, side echoes and hyperbolic echoes on the 3.5-kHz records introduced uncertainties in water depths of as much as 100 m (the corresponding uncertainty in the topographic correction is ±0.06 s). Along the section of the shooting line which crosses the fracture zone ridge, these uncertainties reach as much as 200–400 m (0.12–0.25 s). However, for the majority of our data we are confident of depths to 20 m (±0.01 s).

All data presented in this paper have been corrected using the second technique, i.e., the water delay correction. We chose this not only because we had more confidence in our estimates of p than in our guesses at the shallow basement velocities but also because it produced more consistent results and facilitated comparisons of crustal thicknesses at different re-
Fig. 3. Topographically corrected record sections for OBH 1, 2, and 3 for shots south of the Kane Fracture Zone. All traces are unfiltered. Amplitudes have been corrected by normalizing for shot size and applying a range-dependent amplification. This correction is $(R/R_0)(W_0/W)^{0.58}$, where $R_0 = 10$ km, $W_0 = 14.5$ kg (32 lb), and $W$ is the charge weight of a shot a distance $R$ from the receiver. Solid lines are least squares fits to the first refracted arrivals. Dashed lines are the expected arrival times for the later phases calculated from the velocity model determined by the first arrivals. Phases are labeled following the convention illustrated in Figure 4.
receivers. Topographic corrections are the biggest single source of error in the data set.

**Crustal Structure South of Kane Fracture Zone**

Record Sections

Record sections corrected for topography are shown in Figure 3 for OBH 1, 2, and 3 for shots south of the Kane Fracture Zone. The nomenclature used in identifying the various phases follows that of Spudich [1979] and is illustrated in Figure 4. The solid lines superimposed on the record sections were calculated from a least squares fit to the first arrival travel time picks. The dashed lines are the expected arrival times for later phases calculated from the velocity model determined by the first arrivals.

These record sections show several important similarities suggesting that the crustal structure south of the Kane Fracture Zone is laterally homogeneous on a scale of tens of kilometers. Eight different refracted and reflected phases can be identified from travel time analysis and synthetic seismogram modeling: P_n, P_o, P_m,P, P_mPP_m,P, S_m,S, and P_oWW. The crustal phases display clear, impulsive first arrivals out to ranges of 50 km. Large increases in first-arrival energy occur between 25- and 35-km range with a rapid decrease in energy beyond 35 km. Beyond 35-km range, large-amplitude mantle reflections (P_m,P) are not apparent on any of the record sections. However, on OBH 1 and 3 a small-amplitude phase arriving about 2.0 s after P_m is interpreted as an intracrustal multiple (P_mPP_m,P) reflected from the Moho. Despite the absence of significant sediment cover in this area, prominent S wave arrivals are apparent on all three record sections. Most of this energy is associated with a shear wave mantle reflection (S_m,S) with P to S conversion at the water/basement interface. Refracted upper mantle shear waves (S_o) are not observed, and S_m is generally small in amplitude and difficult to identify owing to interference from earlier arrivals.

Travel Time Interpretations

The observed travel times at each receiver have been interpreted using both conventional slope-intercept methods [Ewing, 1963] and travel time inversion techniques [Bessonova et al., 1974; Kennett and Orcutt, 1976]. A summary of the least squares derived apparent velocities and intercepts is given in Table 1. Also included in Table 1 are the layer thicknesses determined from a standard unreversed structural solution at each receiver. The corresponding velocity/depth functions are plotted in Figure 5. It was necessary at all three receivers to assume the presence of an unobserved shallow layer to make the refraction solutions consistent with the known water depth. Except at OBH 2 South the thickness of this assumed layer is always less than 1 km (Table 1), and varying the velocity of this layer ±0.5 km/s changes the total crustal thickness by less than 100 m.

The apparent velocities observed at OBH 1, 75 km south of the Kane Fracture Zone (5.17, 6.75, and 7.73 km/s), can be associated with layers 2B and 3A and the Moho, respectively. The mantle intercept time (1.09 s) and the total crustal thickness (5.25 km) at this receiver are typical of oceanic crust of this age [Orcutt et al., 1976; Talwani et al., 1971; Keen and Tramontini, 1970]. At OBH 2 a thinner layer 2 and thicker layer 3 are present. Mantle arrivals at this receiver were observed only about 10-km range, and consequently, the apparent velocities and intercepts are not well constrained. In particular, it is not clear whether the 7.2-km/s phase observed south of this receiver is a refracted head wave from the lower crust (layer 3B) or upper mantle. An unreversed structural solution for shots north of OBH 2 indicates a total crustal thickness of about 4.4 km. At OBH 3 the observed apparent velocities (4.97, 6.33, and 7.55 km/s) have been identified with layers 2B and 3A and the Moho, respectively. The mantle intercept time at this receiver (0.90 ± 0.03 s) is well constrained and about 0.19 s less than that observed at OBH 1. This is reflected in an apparent decrease in total crustal thickness from 5.2 km at OBH 1 to 3.9 km at OBH 3, the receiver closest to the fracture zone.

Also shown in Figure 5 are the bounds on all possible velocity-depth functions which fit the observed travel times calculated using the τ-p travel time inversion technique [Bessonova et al., 1974; Kennett and Orcutt, 1976]. This method is useful in that it determines a velocity-depth function which varies continuously with depth and also provides quantitative estimates of the resolution of the observed travel times. In applying this technique we have followed the method outlined by Kennett and Orcutt [1976]. Plots of reduced travel time (T-px) versus distance were calculated for different values of ray parameter p by fitting a weighted cubic spline curve to the observed travel times. A τ(p) function was constructed from the maxima of these curves. Errors in τ were estimated by taking the mean error in travel time observations at ranges corresponding to these maxima. Estimates of arrival time picking errors range from ±0.025 to 0.075 s, and errors in topographic corrections range from ±0.01 to ±0.06 s. The resulting extremal
TABLE 1. Apparent Velocities, Intercepts, and Layer Thicknesses
Observed at OBH 1, 2, 3, 6, 7, 8

<table>
<thead>
<tr>
<th>Layer</th>
<th>Velocity, km/s</th>
<th>Intercept, s</th>
<th>Thickness, km</th>
</tr>
</thead>
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<tr>
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<tr>
<td>P2</td>
<td>4.50†</td>
<td>0.41†</td>
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<tr>
<td>P3</td>
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<td></td>
<td>7.73 ± 0.03</td>
<td>1.09 ± 0.02</td>
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<tr>
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<td>1.48†</td>
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<tr>
<td>Pn</td>
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<td>Pn</td>
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<td>7.80 ± 0.05</td>
<td>1.04 ± 0.03</td>
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</table>

* Layer thicknesses determined from an unreversed structural solution at each receiver.
† Assumed.

bounds on the velocity/depth functions at each receiver are shown in Figure 5.

These bounds are similar to those previously derived for marine refraction profiles over normal oceanic crust [Kennett and Orcutt, 1976; Orcutt et al., 1976]. Gradients in the upper 2 km are steep (~1.0 s⁻¹), decreasing to much smaller values (0.1–0.2 s⁻¹) at greater depths. This sharp change in gradient occurs at a velocity of about 6.0 km/s and corresponds to the transition between seismic layers 2 and 3.

Even with the close shot spacing of the explosive data (<2 km) and the excellent data quality, travel time data alone do not provide particularly good resolution of the velocity structure at these receivers. Additional constraints on the crustal structure south of the Kane Fracture Zone were determined by applying delay time function and amplitude modeling techniques.

Delay Time Solution

One hundred and twenty three mantle arrivals were observed at OBH 1, 2, 3, and 4 from shots south of the Kane Fracture Zone. In order to interpret this large data set we have used the delay time function technique [Morris et al., 1969; Rafti et al., 1969]. This method has two major advantages over conventional reversed solutions. First, the layer boundaries are not required to be planar surfaces but are allowed to undergo gentle undulations as described by low-order polynomial or Fourier functions of position. Second, the solution is not restricted to receiver pairs, travel times from any number of receivers can be combined to produce a single solution.

The general analytical procedure used was that described by Morris [1972]. We assumed lateral and vertical velocity homogeneity and chose to represent the delay time surfaces as a linear combination of polynomial and double Fourier series. Travel times of mantle arrivals from all four receivers were combined into a single data set, and delay time surfaces of progressively higher order were calculated. The difference $R$, between observed travel times and solution-predicted travel times was used to calculate the standard error about the regression, $\sigma$, from

$$\sigma = \left( \frac{\sum R_i^2}{N - M} \right)^{1/2}$$

where $N$ is the number of observed travel times and $M$ is the total number of coefficients in the least squares solution [Morris, 1972]. The quantity $\sigma$ indicates the overall quality of the fit and was used to judge whether higher-order polynomials significantly improved the solution.

The calculated mantle delay time surfaces and residuals are shown in Figure 6. An increase in the number of terms in the function representing the delay time surface allows the surface to vary more rapidly and thus corresponds to more complicated structures. The higher-order solutions shown in Figure 6 are very stable, and the standard errors are reduced from 0.051 s to 0.034 s for the $M = 1$ solution. The best fitting mantle delay time surface ($M = 11$) slopes gradually upward from OBH 1 to OBH 3, the mantle delay times decreasing from about 0.56 s at OBH 1 to 0.50 s at OBH 3. This reflects the decrease in total crustal thickness toward the fracture zone which was suggested by the unreversed slope-intercept interpretations at each receiver. For an average crustal velocity of 6.0 km/s this decrease in mantle delay time corresponds to a

![Fig. 5. Velocity/depth functions determined for OBH 1, 2, and 3 south of the Kane Fracture Zone. Plotted on the left are the velocities and layer thicknesses calculated from a conventional, unreversed slope-intercept solution at each receiver. On the right are the extremal bounds on all possible velocity/depth functions which fit the observed travel times at each receiver.](image)
Fig. 6. Mantle delay time surfaces and residuals calculated for 123 mantle arrivals observed at OBS 1, 2, 3, and 4 south of the Kane Fracture Zone. Receiver locations are indicated by triangles, and shot locations by circles. Delay time surfaces are represented by a linear combination of polynomial and double Fourier series, where $M$ is the number of coefficients in the least squares solution (for example, $M = 5$ corresponds to a first-order polynomial and two Fourier terms). The calculated mantle velocity $V$ and root mean square difference $RMSQ$ between observed travel times and solution-predicted travel times are shown for each surface. The fit does not improve beyond the $M = 11$ solution.
change in total crustal thickness of about 0.5 km. The solution velocity (7.70 km/s) is not unusual for very young crust near spreading centers [Talwani et al., 1971]. Mantle velocities of less than 8.0 km/s have been reported from previous refraction work at International Program of Ocean Drilling (IPOD) site 396 [Husson et al., 1978] on slightly older crust just south of the Kane Fracture Zone (see Figure 1).

Amplitude Modeling

The bounds calculated on possible velocity/depth functions by inverting the observed travel times at OBH 1, 2, and 3 proved disappointingly large (Figure 5). A number of investigators have shown that greater constraints can be placed on the actual crustal structure by fitting both the observed travel times and amplitudes using synthetic seismogram techniques [Ocuiti et al., 1976; Kennet, 1977; Spudich, 1979; Lewis and Sydnor, 1979]. We have used this method to examine the nature of the crust/mantle boundary south of the Kane Fracture Zone.

The synthetic seismograms were calculated using the reflectivity method [Fuchs and Muller, 1971], modified for the marine case and permitting different structures immediately beneath the source and receiver [Kennett, 1975]. We have attempted to model only the OBH 1 record section; however, since similar amplitude relationships were observed at OBH 2 and 3, our results should be applicable to those receivers as well. We have concentrated on explaining the gross amplitude characteristics displayed by the seismograms. In particular, we were interested in understanding the origin of the large-amplitude increase observed near 30 km. We did not attempt to model amplitudes at ranges of less than 10 km.

The layered model shown in Figure 5 was used as the start-
ing point for our analysis, and various classes of models were
tried using a simple ray-tracing algorithm. Synthetic seismo-
grams were then calculated for those models which seemed
likely to reproduce the observed amplitudes. We found this to
be a very useful approach which minimized the number of
synthetic seismograms that needed to be calculated. Gener-
ally, we found, as did Malecek and Cloves [1978], that careful
positioning of the distance range over which triplications in
the travel time-distance curve occur is the most effective
method of controlling relative amplitude changes.

Our preferred model for OBH 1 and the calculated syn-
thetic seismograms are shown in Figure 7. A crust/mantle
transition 1.0–1.5 km thick is suggested, although fine struc-
ture within the transition zone cannot be resolved. A sharper
Moho velocity discontinuity can be discounted because of the
small amplitude of the mantle reflection (PnP), and a signifi-
cantly wider Moho transition would result in much smaller
amplitudes in the 20- to 35-km range than are actually ob-
served. The increase in velocity in the lower part of layer 3 is
not well constrained but is needed to maintain amplitudes be-
 tween 15 and 20 km. The preferred model also accounts for
most of the major second arrival energy apparent on OBH 1 —
the large-amplitude S wave arrivals between 15 and 35 km
and the intercrustal multiple (PnPnPP) apparent beyond 50
km. The S waves were not modeled separately but were calcu-
lated using the best fitting P wave model and a Poisson’s ratio
of 0.25. Most of the larger-amplitude S wave energy appears
to be associated with a shear wave mantle reflection (SvS).

Seismograms with amplitude characteristics similar to those
observed on OBH 1, 2, and 3 have been reported elsewhere
[Malecek and Cloves, 1978; Spudich, 1979; Lewis and Snyd-
smann, 1979] and also interpreted in terms of a crust/mantle
transition width of at least 0.8 km. The location of the crust/
mantle triplication and its associated amplitude increase are
naturally dependent on the total crustal thickness at the site.
The thinner the crust, the closer to the receiver this amplitude
increase will be observed. South of the Kane Fracture Zone,
large amplitudes are centered at about 30 km for OBH 1 and
about 27 km at OBH 3. This is consistent with the slight de-
crease in total crustal thickness between these two receivers
inferred from our travel time analysis.

Summary

The crustal structure of the 7-m.y.-old sea floor south of the
Kane Fracture Zone is well constrained by both travel time
analysis and amplitude modeling. Total crustal thicknesses
decrease from 5.3 km at OBH 1 to 3.9 km at OBH 3, the re-
civer closest to the fracture zone. Crustal velocities are within
the range of those typically observed for normal oceanic crust.
A Moho transition 1–1.5 km thick is required to explain the
distinctive amplitude characteristics displayed by the seismo-
grams at all three receivers. The similarity of the seismograms
observed at these receivers suggests that the deeper crustal
structure south of the Kane Fracture Zone is relatively ho-

mogeneous over a distance of almost 60 km.

Crustal Structure North of Kane Fracture Zone

Record Sections

Record sections for OBH 6, 7, and 8 north of the Kane
Fracture Zone are shown in Figure 8. Combined record sec-
tions with seismograms from explosive and 1000-in. 3 air gun
shots are shown in Figure 8a. The 1000-in. 3 data are also pre-

sented in Figure 8b at an expanded scale. In contrast to the
generally homogeneous structure south of the fracture zone,
distinct offsets exist in the observed travel time–distance
curves north of the fracture zone, indicating a more hetero-
geneous crustal structure. These discontinuities occur about
8.0 km north of OBH 6 and about 10 km south of OBH 7.
Both appear to be related to the presence of a north-south
trending trough a few kilometers in length between OBH 6
and 7. This bathymetric low is apparent in Figure 1 as a
northward bend in the 4000-m contour between OBH 6 and 7.
Travel times from shots over this trough are consistently early
by as much as 0.2 s. Neither water delay nor flat datum topo-
graphic corrections could satisfactorily explain the earliness of
these arrivals.

The large increases in first-arrival energy observed at the
receivers south of the Kane Fracture Zone between 25 and 35
km range do not exist on these record sections. The 1000-in. 3
air gun data (Figure 8b) display good phase coherence over
distances of a few kilometers (for example S0 on OBH 7 south
between 7.5 and 9.5 km and OBH 6 between 8.0 and 9.5 km),
but these data are often separated by intervals over which the
arrivals are weak and incoherent (for example, OBH 7 North
between 8.0 and 10.0 km). In some cases these differences can
be related to the nature of the sea floor at the ray entry point.
If the bottom is flat or gently sloping over a distance of a few
kilometers, the arrivals are coherent; if the bottom relief is
rugged, the arrivals are incoherent. In other cases the origin of
the changes in amplitude displayed by the seismograms in
Figure 8 is far from clear. S wave arrivals exist on seismo-
grams at all three receivers; however, these arrivals are not as
well developed as the S waves observed south of the Kane
Fracture Zone, and like the compressional wave arrivals they
display no consistent range-dependent amplitude character-

Travel Time Interpretation

A summary of the least-squares derived apparent velocities
and intercepts for these receivers is given in Table 1. Also in-
cluded in Table 1 are the layer thicknesses determined from a
standard unreversed structural solution at each receiver. The
corresponding velocity/depth functions are plotted in Figure
9. Also shown are the extremal bounds on the observed veloc-
ity/depth function at OBH 8 calculated using the method out-
lined earlier. Bounds were not calculated for OBH 6 and 7 be-
cause of the obvious lateral heterogeneities in crustal structure
near these receivers.

OBH 6, 7, and 8 had to be deployed closer together than
originally intended in order to avoid crossing a minor fracture
zone located about 70 km north of the Kane Fracture Zone
(see Figure 1). Consequently, our control on the total crustal
thickness and true mantle velocity north of the fracture zone
is much poorer than that south of the fracture zone. At OBH
8, 67 km north of the Kane Fracture Zone, the observed ap-
parent velocities (5.35, 6.75, and 7.80 km/s) are typical of nor-
amal oceanic crust. The mantle intercept time (1.04 s) is similar
to that observed at OBH 1, suggesting approximately the same
total crustal thickness (5–6 km). However, no mantle arrivals
were observed at OBH 7 from shots north of the fracture zone,
and at OBH 6, estimates of the total crustal thickness are com-
plicated by a discontinuity in the travel time–distance curve at
8.0-km range. If a shallow crustal structure similar to that ob-
served at OBH 8 is assumed, the total crustal thickness is 4.5
km, similar to what was observed at OBH 2 and 3 south of the
fracture zone. There is some evidence for crustal thickening of
Fig. 8. Topographically corrected record sections for OBH 6, 7, and 8 for shots north of the Kane Fracture Zone. (a) Combined record section showing seismograms from both explosive and 1000-in.³ air gun shots. Only every fourth air gun shot is plotted. (b) Record sections of 1000-in.³ air gun data plotted at an expanded scale. Otherwise, the same as for Figure 3.
1.0–1.5 km between OBH 6 and 8, which can explain the low apparent mantle velocity (7.56 km/s) observed at OBH 6.

Because so few mantle arrivals were observed from shots north of the Kane Fracture Zone, a delay time solution was not attempted. Synthetic seismogram modeling was also not done because of the inconsistent amplitude relationships exhibited by the record sections at these receivers. However, additional constraints on the crustal structure north of the Kane Fracture Zone can be derived from the intercepts of mantle arrivals observed at OBH 6, 7, and 8 from shots south of the fracture zone and the apparent mantle velocities observed at OBH 1, 2, 3, and 4 from shots north of the fracture zone. Table 2 summarizes these results.

The mantle intercept times observed at OBH 6, 7, and 8 for shots south of the fracture zone are the same to within the limits of their standard errors. The mean intercept time (1.03 s) is approximately twice the average mantle delay time calculated south of the fracture zone (Figure 6). This is consistent with the existence of similar crustal thicknesses north and south of the Kane Fracture Zone. The apparent mantle velocities observed at OBH 1, 2, 3, and 4 from shots north of the fracture zone are also essentially the same. The mean velocity (7.62 km/s) differs by only 0.1 km/s from that determined south of the fracture zone. These results and the low apparent mantle velocity observed at OBH 8 both indicate that mantle velocities north of the Kane Fracture Zone are not significantly higher than those observed south of the fracture zone.

Summary

While the crustal structure north of the Kane Fracture Zone area is not as well constrained as that south of the fracture zone, there is no convincing evidence for significantly different crustal thicknesses or upper mantle velocities. This is somewhat surprising in that the sea floor north of the fracture zone is 10 m.y. older than that south of the fracture zone. The older crust north of the fracture zone, in fact, has some seismic characteristics that are sometimes associated with younger crust including less well defined velocity interfaces, pronounced structural heterogeneity, and the absence of a well-developed crust/mantle triplication. At OBH 8, north of the fracture zone, the total crustal thickness is 5.3 km, nearly identical to that observed at OBH 1 75 km south of the fracture zone. The crust appears to thin slightly toward the fracture zone on both sides; however, within 15 km of the fracture zone trough, total crustal thicknesses are probably not more than 1.0 km less than that observed at the most distant receivers. The crustal velocities and thicknesses on both sides of the Kane Fracture Zone appear to fall within the range of those typically observed for normal oceanic crust.

CRUSTAL STRUCTURE WITHIN THE KANE FRACUTRE ZONE

Interpretation of a Reversed Line in the Fracture Zone Trough

A reversed refraction line about 50 km in length was shot along the Kane Fracture Zone trough between OBH 4 and 5 (Figure 1) in order to determine if the crustal structure here differs in any significant way from the crust on either side of the fracture zone. Topographically corrected record sections for this line are shown in Figure 10.

The seismograms recorded at OBH 4 and 5 are quite differ-
Fig. 10. Topographically corrected record sections for OBH 4 and 5 for reversed line shot along Kane Fracture Zone trough. Otherwise, the same as for Figure 3.
Fig. 11. Travel time–distance graphs for OBH 4 and 5 with two alternative least squares solutions which fit the observed travel times. Light dotted lines are expected travel times for Raitt’s [1963] standard mean oceanic crust model.

tent: at OBH 5 the compressional wave arrivals are characterized by long reverberative wave trains, while at OBH 4 the arrivals are more impulsive. These are not instrumental differences: their origin is not known. No significant S wave energy is apparent on either receiver. Corrected travel time–distance graphs are shown in Figure 11 for these receivers with two alternative least squares solutions which fit the observed travel times. For comparison, travel time–distance curves for Raitt’s [1963] standard mean oceanic crustal model are also shown. The apparent velocities and intercepts for our preferred solution are listed in Table 3. The layered velocity models derived from these travel times are shown in Figure 12 together with the extremal bounds on all possible velocity functions that will fit the observed travel times. Owing to the uncertainty in determining ray entry points, the errors involved in making topographic corrections are significantly larger within the fracture zone than over normal oceanic crust, and this is reflected in much wider bounds at OBH 4 and 5.

The crustal structure within the Kane Fracture Zone is clearly anomalous, with a total crustal thickness of as little as 2–3 km, about half the thickness of normal oceanic crust. Velocities of 8 km/s may occur at depths as shallow as 2 km beneath the floor of the fracture zone. At 40-km range, arrivals within the fracture zone are early by more than 0.2 s in comparison to what would be expected if normal crustal thicknesses existed beneath the Kane Fracture Zone trough (Figure 11). Aside from its thickness, two other characteristics of the Kane Fracture Zone crustal structure distinguish it from normal oceanic crust. First, a low-velocity (~4 km/s) refractor occurs as a first arrival out to ranges of almost 10 km. This low velocity may be due to fracturing and brecciation of the crust by faulting within the active transform domain. Fault gouge with low compressional wave velocities has been reported to extend to depths of several kilometers along the San Andreas fault zone in California [Wang et al., 1978]. The extensive fracturing and brecciation of the shallow crust may also explain the lack of significant S wave energy on these record sections. R. S. White and R. Stephen (manuscript in preparation, 1980) have shown that at phase velocities above the basement P wave velocity the efficiency of shear wave conversion is significantly reduced. Extensive fracturing of the upper 50–100 m by faulting in the active transform section will lower the basement P wave velocities enough that converted shear waves from the lower crust and upper mantle will not be observed. Extensive fracturing of the shallow crust will also significantly increase the attenuation of shear wave energy.

Another anomalous characteristic of the Kane Fracture Zone crust is the absence of a refractor with a velocity typical of seismic layer 3. The crust within the fracture zone can be modeled with a single uniform gradient of ~1.3 s⁻¹ changing at depths of 2–4 km below the sea floor into a lower gradient typical of the upper mantle. This type of crustal structure, with no major velocity discontinuity at depth, is consistent with the lack of significant reflected wave energy (P or S) on these seismograms and the very gradual decrease in the amplitude of Pp arrivals with increasing range.

Reflected head waves from oceanic layer 3 are the most commonly observed and most consistent of all crustal refractions [Lewis, 1978]. The absence of this refractor within the Kane Fracture Zone is the major characteristic which distinguishes this crust from normal oceanic crust. Layer 3 is generally thought to be composed of an assemblage of 3–5 km of gabbro, metagabbro and, at deeper levels, cumulate gabbro with interbedded ultramafic units [Miyashiro et al., 1969; Melson and Thompson, 1971]. Since gabbro is the most common rock type recovered from the walls of the Kane Fracture Zone [Thompson, 1973; W. Bryan and H. Dick, personal communication, 1979], it is unlikely that layer 3 is completely missing beneath the fracture zone trough. The anomalous crust within the fracture zone is, however, probably composed of an unusually thin gabbroic layer (2 km thick) overlain by a few hundred meters of extrusive basalts and dikes. At shallow levels the velocity of the gabbroic layer may be lowered by fracturing and brecciation within the active transform domain or the emplacement of low-velocity serpentinite.

**Interpretation of Off-Line Shots**

Additional constraints on the lateral extent of the anomalous crust observed beneath the Kane Fracture Zone trough can be determined from examining the travel times of manile arrivals recorded at OBH 5 from shots along the shooting line.
perpendicular to the fracture zone (Figure 1). These travel times can be written as the sum of a shot and receiver delay:

\[ T = \Delta / v_s + \tau_s + \tau_R \]  

(2)

with

\[ \tau_s = \sum_{i=1}^{n} Z_i (1 - V_i^2 / V_s^2)^{1/2} / V_i \]

and a similar expression for \( \tau_R \). Here \( \Delta \) is the shot/receiver range, \( V_s \) is the mantle velocity, and \( \tau_s \) and \( \tau_R \) are the mantle delay times at the shot and receiver, respectively. \( Z_i \) and \( V_i \) are the thickness and velocity, respectively, of the \( i \)-th crustal layer beneath the shot (or receiver).

We have used a constant mantle velocity of 7.7 km/s (determined from our previous analysis) and subtracted the first term on the right side of (2) from the observed travel times. What remains is the total mantle delay time—the sum of the mantle delay times at the shot and receiver. Since all of the shots are recorded at the same receiver, OBI 5, we can in effect examine the relative change in shot delay along the main shooting line perpendicular to the fracture zone. With our shot spacing of less than 2 km this method provides a way of determining changes in crustal structure near the fracture zone over distances of only a few kilometers.

The calculated mantle delay times (\( \tau_s + \tau_R \)) are plotted against distance from the fracture zone in Figure 13. With the exception of a few shots around OBH 2, mantle delay times south of the fracture zone vary smoothly, decreasing from 1.2 s near OBH 1 to about 1.0 s at OBH 3. This small decrease in mantle delay time toward the fracture zone is consistent with the 0.5- to 1.0-km decrease in total crustal thickness between OBH 1 and 3 estimated from our previous analysis. As the fracture zone trough is approached, the mantle delay times decrease dramatically over a distance of only 5 km. Within the fracture zone trough, total mantle delay times are only 0.5 s, about half their value south of the fracture zone, reflecting the much thinner crust within the fracture zone. North of the Kane Fracture Zone, mantle delay times remain nearly 0.4 s less than the delay times south of the fracture zone. This surprising delay time offset across the fracture zone is also apparent in the different mantle intercept times observed at OBH 4 for shots north and south of the fracture zone (Table 2).

The total crustal thickness north of the Kane Fracture Zone does not appear to be significantly different from that observed south of the fracture zone (Figure 9 and discussion in previous section). Consequently, the shot delays (\( \tau_s \)) should be approximately the same for all shots except those over the fracture zone trough. The fact that they are not is surprising. From (2) it is obvious that the smaller mantle delay times north of the fracture zone must be due to either higher mantle velocities or the fracture zone (\( V_s \sim 8.0 \) km/s) or a smaller value for the receiver delay at OBH 5.

A reversed determination of the true mantle velocity north of the Kane Fracture Zone could not be made, but existing data do not support the existence of higher mantle velocities north of the fracture zone. Low apparent mantle velocities (7.55–7.67 km/s) were observed at OBH 1, 2, 3, and 4 from shots north of the fracture zone, and only a slightly higher velocity (7.80 km/s) was observed at OBH 8 from these same shots. This appears to preclude significantly higher mantle velocities north of the Kane Fracture Zone. Anisotropy is another possible source of error, since the shot-receiver azimuth of the off-line shots varies from N-S to almost E-W. Anisotropy in upper mantle velocities of \( \sim 0.3 \) km/s has been well documented in the Pacific, where the maximum velocity generally parallels the spreading direction [Ratti et al., 1969; Morris et al., 1969]. Anisotropy in this sense cannot, however, explain the delay time offset observed across the Kane Fracture Zone, since it would result in a delay time pattern that would be symmetric across the fracture zone.

A basic assumption in this analysis is that the receiver delay at OBH 5 is not azimuthally dependent. This assumption is generally valid but will not necessarily be true if large lateral changes in crustal structure exist near the receiver. In this portion of the Kane Fracture Zone there appears to be a very abrupt transition from the normal crustal thicknesses south of the fracture zone to the anomalously thin crust beneath the fracture zone trough. If dense, high-velocity mantle type material exists at shallow depths beneath the Kane Fracture Zone ridge, then the receiver delays at OBH 4 and 5 could be significantly less for shots north of the fracture zone than for shots south of the fracture zone. Mantle arrivals from shots north of the fracture zone will be early in relation to mantle arrivals from shots south of the fracture zone because they have passed through the anomalously thin crust within the fracture zone (Figure 14). This can explain the delay time offset across the fracture zone observed at OBH 5 (Figure 13) and the much smaller mantle intercept time observed at OBH 4 from shots north of the fracture zone compared to south of the fracture zone (Table 2). While these results suggest that the anomalously thin crust found beneath the Kane Fracture Zone trough extends northward beneath at least part of the
fracture zone ridge, few other seismic constraints can be placed on the crustal structure of this feature.

**Summary**

Anomalously thin crust is present along at least a 50-km-long segment of the Kane Fracture Zone near 44°W. Total crustal thicknesses beneath the fracture zone trough may be as little as 2-3 km, about half the thickness of normal oceanic crust (Figure 14). An analysis of mantle delay times across the fracture zone suggests that the transition from normal crustal thicknesses south of the fracture zone to the anomalously thin crust beneath the fracture zone trough occurs over a horizontal distance of only about 5 km. This anomalous fracture zone crust extends over a width of at least 10 km, and dense, high-velocity mantle type material may also be present at shallow depths beneath part of the adjacent Kane Fracture Zone ridge.

**Discussion**

Several other geophysical studies suggest that the crustal structure within fracture zones may be anomalous. On the basis of an analysis of gravity and magnetic data without refraction control, Cochran [1973] and Robb and Kane [1975] proposed that dense, ultrabasic rocks reside at shallow depths along both the Romanche and Vema fracture zones. Rowlett et al. [1979] found a decrease in travel times on a refraction line which extended across Fracture Zone A in the Famous area. They interpreted this as indicating a zone of higher than normal velocity material beneath the fracture zone. White and Matthews [1980] report a decrease in layer 3 delay times over a 7-km-wide zone associated with a small offset fracture zone near King’s Trough in the northeastern Atlantic. They suggest that the crustal structure immediately beneath the fracture zone is anomalous, with higher velocities than normal oceanic crust at comparable depths. White and Matthews also argue...
that a narrow zone of anomalous crust associated with Fracture Zone B in the Famous area can explain why refracted waves which have crossed this fracture zone are early by about 0.15 s on Whittmarsh’s [1975] line 4a shown by Fowler [1976].

These results indicate that the anomalous crustal structure found within the Kane Fracture Zone may be typical of parts of several other North Atlantic transform faults and fracture zones. We propose that the anomalously thin crust found within these fracture zones is a primary feature caused by an alteration in the normal processes of crustal genesis at a ridge/transform intersection. With a large offset transform like the Kane Fracture Zone the ridge magma system must be effectively terminated at the fracture zone. Near the end of the magma chamber the amount of basaltic liquid derived from partial melting is likely to be decreased, resulting in volcanism that is either less frequent or less voluminous within the fracture zone and the formation of a thinner crustal layer. The abundance of gabbros exposed in the walls of the Kane Fracture Zone [Thompson, 1973; W. Bryan and H. Dick, personal communication, 1979] and our seismic refraction results suggest that the crust which forms is composed of only a thin carapace of extrusive basalts and dikes a few hundred meters thick underlain by about 2 km of gabbro and metagabbro. The width of this anomalous crust may be quite narrow, possibly restricted to a zone no wider than a typical transform fault valley (~10 km) in most cases.

White and Matthews [1980] have suggested a similar mechanism to explain the anomalous crustal structure which they found associated with a very small offset fracture zone in the northeastern Atlantic. Their results suggest that the magma systems on either side of even very small offset transforms remain distinct and the accretion process within a narrow zone associated with the fracture zone is altered. Thus the juxtaposition of a much older plate edge at the ridge/transform intersection of a large offset fracture zone may not be required; the formation of anomalous crust within fracture zones may be primarily due to the termination of shallow level magma systems at a ridge/transform intersection, regardless of the age difference across the fracture zone.

The concept of thinner crust within fracture zones is appealing in that it is able to explain a number of different geological and geophysical observations regarding fracture zones [Fox, 1978]. Thinner crust can explain the observed gravity anomalies across the Romanche [Cochran, 1973] and Vema fracture zones [Robb and Kane, 1975]. It also can reconcile the recovery of deep crustal rocks from fracture zone walls which are composed of relatively small throw faults. Multibeam echo-sounding surveys of several major North Atlantic fracture zones (H. Needham et al., manuscript in preparation, 1980) and deep-tow and submersible observations of Fracture Zone A in the Famous area [Detrick et al., 1973; Choukroune et al., 1978] have clearly demonstrated that fracture zone walls are not composed of a single large fault but are the product of a large number of small throw faults (~500 m) linked together by terraces and talus ramps. Francheteau et al. [1976] have shown that gabbros and other deep crustal rocks are often recovered from high on these escarpments, which is difficult to explain if oceanic crust of normal thickness (~5–6 km) is exposed in the fracture zone walls. If crustal thicknesses within fracture zones are only 2–3 km, these deep crustal rocks are much more likely to be exposed. Submersible observations within the Cayman Trough, in close proximity to a large-offset transform, confirm that an attenuated volcanic and plutonic section is present [Caytough, 1979].

Thin crust within fracture zones may also provide an explanation for why transform fault valleys are several hundred meters to a kilometer deeper than the surrounding sea floor. Near the ridge/transform intersection of a large offset fracture zone the depth of the transform valley may be partly controlled by dynamic processes [Sleep and Biehler, 1970]; however, the bathymetric distinction between the fracture zone trough and the adjacent sea floor persists even in sea floor many tens of millions of years old, where these processes are unlikely to be important. Using a simple Airy isostasy model and densities for crust and mantle rocks of 2.80 and 3.39 g/cm³, respectively, we would expect fracture zone crust 2.5–3.0 km thick to be about 700–800 m deeper than normal oceanic crust 5–6 km thick. Thus thinner crust within fracture zones provides a simple mechanism for creating and maintaining a transform fault valley. Extension across the transform fault [van Andel et al., 1969] is not required to produce this morphology.

The crustal structure of the Kane Fracture Zone between 40°W and 45°W is complicated by the presence of an elongate ridge which borders the fracture zone on the north. While we have few seismic constraints on the crustal structure of this ridge, our results do suggest that dense, high-velocity mantle type material may exist at shallow depths beneath this feature (Figure 14). Similar ridges border parts of the Vema and Romanche fracture zones in the equatorial Atlantic [van Andel et al., 1971; Bonatti et al., 1971]. Bonatti and Honnorez [1976] have proposed that the transverse ridges bordering these fracture zones are essentially uplifted blocks of crustal and upper mantle material. They speculate that the circulation of seawater deep within fracture zones serpentinitizes upper mantle peridotite, lowering its density in relation to crustal gabbro and metagabbro. A gravity-induced ultramafic diapir forms which intrudes the overlying crust uplifting these fracture zone ridges. In the Romanche Fracture Zone near 18°W, Bonatti and Honnorez [1976] claim that over 4 km of serpentinitized peridotite are exposed in the southern wall of the fracture zone.

The diapiric intrusion of partially hydrated upper mantle peridotite along fracture zones has frequently been suggested primarily on petrologic grounds [Miyashiro et al., 1969; Bonatti, 1971; Bonatti and Honnorez, 1971; Melson and Thompson, 1971]. The presence of anomalously thin crust within fracture zones would make it easier for water to penetrate into the upper mantle and form these diapirs. Much of the serpentinite recovered from fracture zones may have been forced upward as narrow slivers or screens along preexisting fault zones [Francheteau et al., 1976]; however, large ultramafic intrusions along fracture zones could also form, and the ridges bordering many large Atlantic fracture zones may be the surface expression of these intrusions. The origin of the Kane Fracture Zone ridge is not known, but it could be the surface expression of a large ultramafic intrusion.

If the anomalous crustal structure observed within the Kane Fracture Zone near 44°W is representative of the crustal structure along many Atlantic transform faults and fracture zones, then our view of the gross structure of the ocean basins must change. These narrow zones of anomalously thin crust associated with fracture zones would be a major feature of the seismic structure of the ocean basins. Ribbons of relatively homogeneous, so-called normal crust 50–100 km wide will ex-
tend away from spreading centers, each ribbon formed at a discrete ridge crest segment and each separated by the thin crust underlying the fracture zone troughs. The anomalous crust associated with fracture zones will generally be confined to a zone no wider than a typical transform fault valley (~10 km). If the transform fault valley is in isostatic equilibrium, the mass excess associated with the elevated Moho will be compensated by the mass deficiency associated with the greater water depths in the fracture zone trough, and as a result there may be no significant gravity anomaly over the fracture zone. Only where large ultramafic intrusions have occurred along fracture zones will large gravity anomalies exist. As a result, detailed seismic refraction studies of fracture zones like that reported here are probably the best method of determining the presence of this anomalous crust.

CONCLUSIONS

On the basis of this study we draw the following conclusions:

1. The crustal thicknesses and velocities observed on either side of the Kane Fracture Zone near 44°W fall within the range of those typically observed for normal oceanic crust. Despite a 10-m.y. age difference across the fracture zone, there is no convincing evidence for significantly different crustal thicknesses or upper mantle velocities on either side of the fracture zone. The crust does appear to thin slightly toward the fracture zone on both sides; however, within 15 km of the fracture zone trough, total crustal thicknesses are probably not more than 1 km less than that observed at the most distant receivers.

2. Anomalously thin crust is present beneath a 50-km-long segment of the Kane Fracture Zone trough near 44°W. Total crustal thicknesses are only about 2-3 km, about half the thickness of normal oceanic crust. The crust is also characterized seismically by low compressional wave velocities (~4.0 km/s) at shallow depths and the absence of a normal layer 3 refractor. The transition from normal crustal thicknesses south of the Kane Fracture Zone to the thin crust beneath the fracture zone trough is remarkably abrupt, probably occurring over a horizontal distance of less than 5 km.

3. From these results and the abundant exposure of gabбро and other deep crustal rocks in the walls of the Kane Fracture Zone we infer that the crust within the fracture zone is composed of a thin carapace of extrusive basalts and dikes a few hundred meters thick overlaying about 2 km of gabбро and metagаббро, possibly with interbedded ultramafics. This anomalously thin crustal section may be extensively fractured and brecciated at shallow levels by faulting in the active transform section of the fracture zone.

4. Results from other geological and geophysical studies of fracture zones suggest that this type of crustal structure may be typical of parts of many Atlantic fracture zones. We propose that the anomalously thin crust found within these fracture zones is a primary feature caused by the accretion of a thinner volcanic and plutonic layer at the ridge/transform intersection. A relatively narrow zone (~10 km wide) of thin crust within fracture zones can explain a number of geological and geophysical characteristics of fracture zones including the depth of the transform fault valley and the exposure of deep crustal and upper mantle rocks in the walls of fracture zones.

5. Few seismic constraints exist on the crustal structure of the ridge which forms the northern wall of the Kane Fracture Zone between 42° and 45°W; however, an analysis of mantle delay times across the fracture zone suggests that dense, high-velocity mantle type material may exist at shallow depths beneath this feature.

Acknowledgments

This research was sponsored by the Office of Naval Research under contract N00014-74-0262 NR083-004 to the Woods Hole Oceanographic Institution. We thank the officers, crew, and scientific complement of the R/V Atlantis II during cruise 96, legs 2 and 3, whose hard work and cooperation made this experiment possible. In particular, we thank E. T. Bunce, Chief Scientist during leg 2, who carried out important preliminary survey work. The flawless performance of the ocean bottom hydrophones and the excellent quality of the data they recorded are due to the technical skill and dedication of D. Koelsch and C. Grant, who built and maintained these instruments. C. Dean, G. Power, and G. Glass all helped to develop a working data reduction system for the refraction data, and L. Gove has been instrumental in streamlining this system and adapting it for use on the VAX 11/780. C. Dean, W. Lau, and A. Sundberg assisted in various aspects of data reduction and interpretation. H. Hayes, M. Coffin, C. Peters, and P. D. Rabinowitz assisted in compiling and contouring the bathymetry shown in Figure 1. We thank R. Stephen for modifying and improving the reflectivity synthetic seismogram method to use in this work and for his invaluable guidance in its application. R. S. White, R. Stephen, H. Schouten, P. J. Fox, W. Bryan, and R. Parker offered suggestions at various stages which improved this work. Woods Hole Oceanographic Institution contribution 4426.

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Hussong, D. M., P. B. Fryer, J. D. Tuthill, and L. K. Wipperman, The


(Received September 4, 1979; accepted March 28, 1980.)
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