THE VEMA FRACTURE ZONE AND THE TECTONICS OF TRANSVERSE SHEAR ZONES IN OCEANIC CRUSTAL PLATES

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Abstract. At 11°N latitude, the Mid-Atlantic ridge is offset 300 km by the Vema fracture zone. Between the ridge offset, the fracture consists of an elongate, parallelogram-shaped trough bordered on the north and south by narrow, high walls. The W–E trending valley floor is segmented by basement ridges and troughs which trend W10°N and are deeply buried by sediment. Uniform high heat flow characterizes the valley area. Seismically inactive valleys south of the Vema fracture, also trending W10°N, are interpreted as relict fracture zones. We explain the high heat flow and the shape of the Vema fracture as the result of secondary sea-floor spreading produced by a reorientation of the direction of sea-floor spreading from W10°N to west-east. This reorientation probably began approximately 10 million years ago. Rapid filling of the fracture valley by turbidites from the Demerara Abyssal plain took place during the last million years.

The large amount of differential uplift in the Vema fracture is not explained by the reorientation model. Since the spreading rate across the valley is small compared to that across the ridge crest, we suggest that it takes place by intrusion of very thin dikes that cool rapidly and hence have high viscosity. Upwelling in the fracture valley will thus result in considerable loss of hydraulic head, according to models by Sleep and Biehler (1970), and recovery of the lost head could produce valley walls higher than the adjacent ridge crest. We further postulate that the spreading takes place along the edges of the fracture zone rather than in the center. This would account for the uniform distribution of heat flow along the fracture valley and for the lack of disturbance of the valley fill. As a consequence, a median ridge should form in the center, where head loss is compensated in the older crust; such a median ridge may be present. The width of the valley should be a function of the angle and time of reorientation, and of the spreading rate; the width so obtained for the Vema fracture is in accordance with the observed width. If this model is correct, the narrowness of the valley walls implies a thin lithosphere of very limited horizontal strength.

1. Introduction

In recent years, rapid growth of geophysical information has resulted in comprehensive concepts regarding the movements of large crustal plates. These concepts are now quite specific (Le Pichon, 1968; Isacks *et al.*, 1968) and may be fruitfully examined in the light of detailed structural analyses of selected critical portions of the ocean floor. Fracture zones, as sites of adjustment of the differential movements of plates, play a major role in these concepts, (e.g., Ball and Harrison, 1970), but details concerning their morphology, geology, and structural history are still sparse. As part of a series of detailed tectonic studies of selected parts of the Mid-Atlantic ridge province, we have examined the intersection of the ridge with the Vema fracture zone at approximately 11° N latitude. In this paper, we present a summary of the available geophysical and

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geological data, a new and detailed bathymetric chart, and models for the structure and genesis of the parts of fracture zones which connect displaced ridge crests. It appears to be too early, in view of the very limited amount of detailed information, to test fully specific hypotheses regarding the dynamics of fracture zones, but the purely geometric concepts first presented by Wilson (1965) are clearly too limited for a full understanding of the tectonics of these features.

The primary data were collected on cruise 1965-1 of R. V. Thomas Washington of the Scripps Institution of Oceanography, and cruises 20 and 31 of R. V. Atlantis II of Woods Hole Oceanographic Institution. Bathymetric data obtained on various cruises of R. V. Vema of Lamont-Doherty Geological Observatory were provided by Bruce C. Heezen. Bathymetric coverage is nearly continuous (Figure 1). Seismic

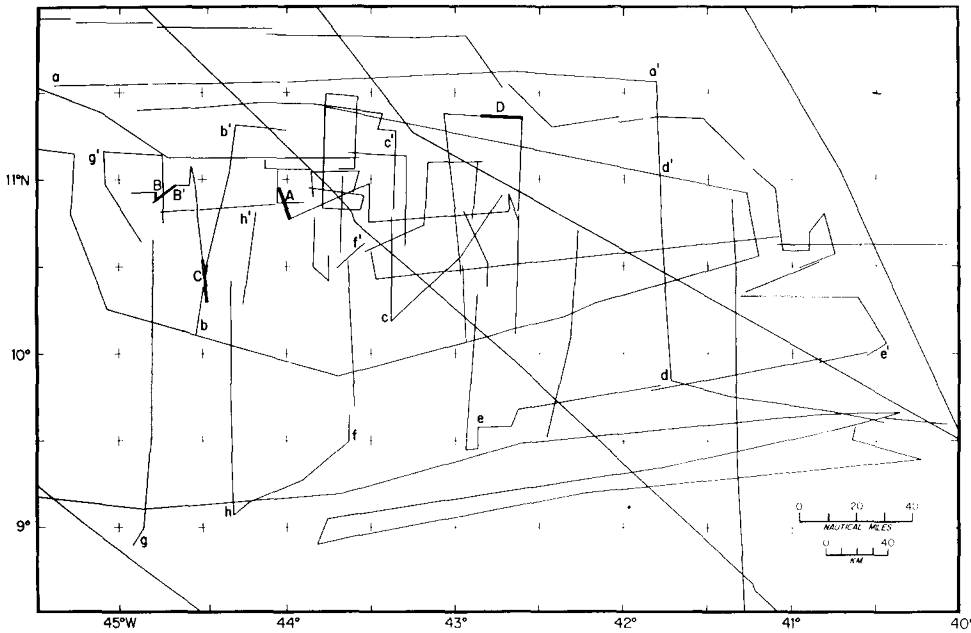

Fig. 1. Location of ship's tracks used in preparation of the bathymetric chart of Figure 2. Profiles of Figure 3 indicated with lower case letters; profiles of Figures 5–6, 8–9 shown as heavy lines.

profiling with a 40 000 J arcer was carried out on part of the Washington cruise, whereas heat-flow data, cores and dredge samples were obtained on the Atlantic cruises. Magnetic data have been collected in the vicinity of the Vema fracture aboard Research Vessels Chain and Atlantis II of the Woods Hole Oceanographic Institution, Washington and Argo of the Scripps Institution of Oceanography, Conrad and Vema of the Lamont-Doherty Geological Observatory and Trident of the University of Rhode Island. Navigation on all cruises was by conventional means, assisted by anchored radar reflector buoys and VLF navigation on the Washington cruise.

2. Relief

The Vema fracture zone, discovered by R. V. Vema in 1956 was first described by Heezen *et al.* (1964b), who suggested that several additional fracture zones might be located somewhat farther south. Van Andel *et al.* (1967) presented further data including seismic reflection profiles that showed the presence of thick sediment fill in the fracture valleys.

Several major topographic features dominate the regional relief (Figure 2): two

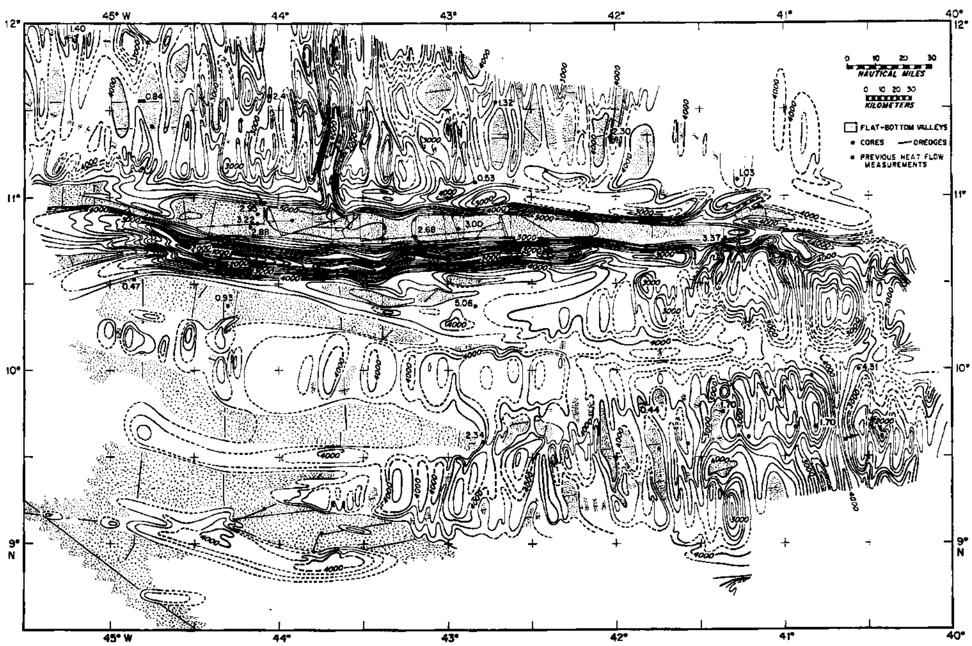

Fig. 2. Bathymetric chart of the Vema fracture region. Contour interval 200 m; based on uniform sound velocity of 1463 m/s (800 fa/s) velocity of sound in sea water. Heat flow values shown in HFU (10^{-6} cal/cm^2/s). Black circles indicate cores and heat flow values obtained during Atlantis II cruises, squares the heat flow values previously available. Stippled pattern represents all larger flat sediment floors.

segments of the Mid-Atlantic ridge, the Vema fracture zone itself, and two, perhaps three, transverse valleys located farther south. The Vema fracture offsets the crest of the ridge over a distance of 320 km. The southern and northern ridge segments possess a well defined median rift (Figure 3) with a maximum depth ranging from 3800–4200 m. The median rifts are bordered by crestal ranges rising 1600–2000 m above the valley floors. The ranges consist of narrow parallel ridges and valleys with an internal relief of 400–800 m and a northerly trend. In the crestal province, the sediment cover is thin or absent. Although it is difficult to determine precisely, the total width of the crestal zone so defined is about 80 km for the northern and perhaps

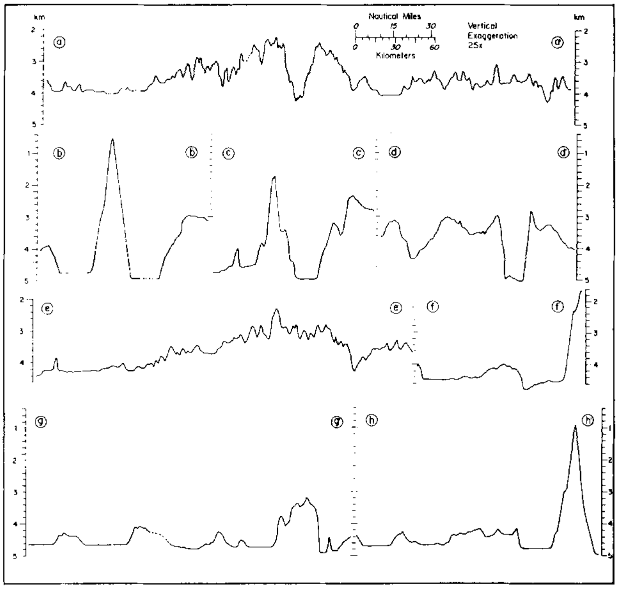

Fig. 3. Topographic profiles. Locations see Figure 1. Vertical exaggeration 25 ×. Profiles a–a′ and e–e′ northern and southern ridge segments; b–b′, c–c′ and d–d′ Vema fracture zone; others from west slope of southern ridge across southern transverse valleys.

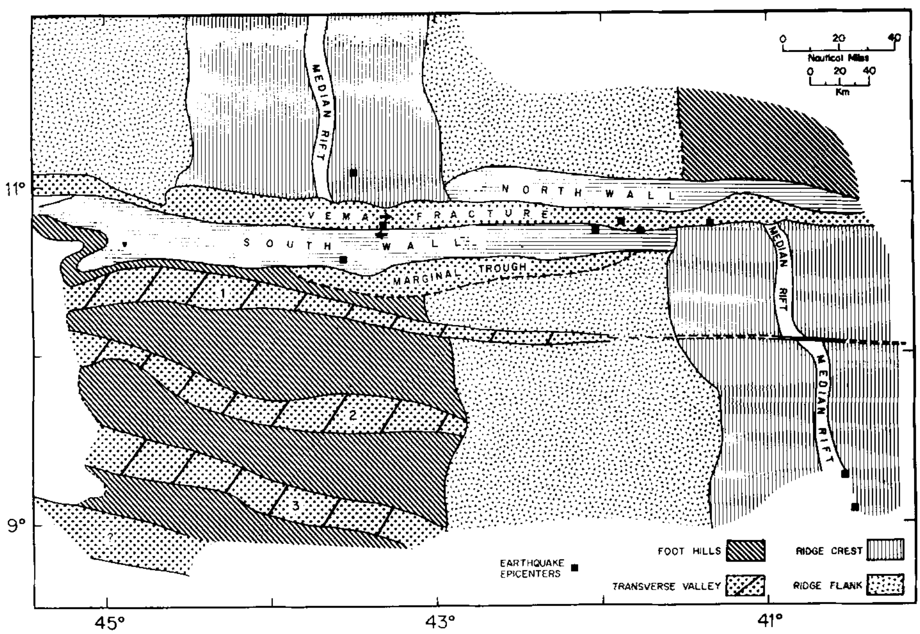

Fig. 4. Physiographic provinces of the Vema fracture region. Black squares are earthquake epicenters (courtesy of Lynn R. Sykes). First motion sense shown with arrows for one epicenter.

100 km for the southern segment (Figure 4); the eastern boundary of the latter has not been determined in our surveys. This is somewhat less than the crestal width farther north in the Atlantic (Heezen *et al.*, 1959, Figure 43; van Andel and Bowin, 1968) and also than that in the South Atlantic (van Andel and Heath, 1970).

The crestal zone is bordered by a flank province with more subdued and longer wave length topography, an internal relief of 200–500 m, and less but still significant N–S linearity. In this zone, a more or less continuous sediment blanket is present and the valleys are nearly always partly filled with sediment. The flank zone extends to approximately 120 km to each side of the ridge axis and is much narrower than the equivalent province farther north or in the South Atlantic. It cannot be easily subdivided into steps as is possible elsewhere in the North Atlantic (Heezen *et al.*, 1959).

Beyond the flank zone occurs a broad foothill province of open roundish hills with an internal relief of 100–200 m and a weak north–south linearity, smoothed and partly leveled by sediments. Further westward, the foothills province grades into the Demerara abyssal plain. The boundaries between the various physiographic provinces (Figure 4) are well-defined by the nature of the relief and the distribution of sediment and are also offset by the Vema fracture. The morphological differences between the zones are significant and many are not produced merely by changes in sediment cover. Simple lateral transfer of structural elements from crest to flank by sea-floor spreading cannot account for many of the differences between provinces, but additional tectonic events are probably required (Schneider and Vogt, 1968; van Andel and Bowin, 1968).

Transverse disturbances interrupt the general north-south trend of the ridge. The most important of these is the Vema fracture, a narrow, west–east trending elongate trough. Between 41°10′ and 44°30′W the floor of this valley is flat and lies at approximately 5000 m. At the western end, near the boundary between ridge crest and flank to the north, the trough abruptly narrows and changes to a more northerly trend (W 15° N). At the eastern end, the flat floor disappears, the valley shoals, and a longitudinal ridge, rather poorly defined by existing data, occupies its center. The eastern end of the valley lies outside the surveyed area.

The walls of the valley are steep with slopes of approximately 15°. A long, narrow ridge with a width of 30 km and a length of 400 km borders the valley on the south side. This ridge crests at 3000 m above the valley floor with portions as shoal as 600 m below sea level. A similar wall of less height (2000 m above the valley floor) lies along the north side of the valley, east of the northern ridge crest. Similar ridges along fracture zone valleys have been found elsewhere in the Atlantic, for example along the Chain and Romanche (Heezen *et al.*, 1964a), Atlantis (Heezen and Tharp, 1965), Oceanographer (Fox *et al.*, 1969a), Gibbs (Fleming *et al.*, 1970), and Ascension fracture zones (van Andel and Von Herzen, unpublished data). The graben-like shape of the fracture zones and the presence of bordering ridges, as well as their transverse asymmetry, appear to characterize many major Atlantic fracture zones. These features distinguish them from those of the Pacific which consist of ridges with different regional sea floor depths on either side, or of complexes of parallel smaller ridges and valleys (Menard, 1964; Menard and Atwater, 1968).

The median rifts of the northern and southern ridge segments open directly onto the fracture floor. The exit of the northern rift segment has been well surveyed; the absence of closure is established without question. Just south of the northern exit, the flat floor of the fracture valley is gently depressed about 150 m below its general level. Otherwise the fracture valley floor itself is generally level, although a few channel-like depressions occur near 42° 30′ and 43° W. West of 44° 40′, the floor shoals slightly to 4950 m. It appears to have continuity with the Demerara abyssal plain, although this continuity is considerably restricted through at least one narrow channel near 45° W.

Other transverse features exist south of the Vema fracture. South of the south wall, an elongate marginal depression occurs which consists of several *en echelon* troughs with flat floors at about 4600 m. A reflection profile across one of these near 43° W shows about 240 m of sediment (based on an assumed velocity in the sediment of 2 km/s). Farther south lie three large, flat-floored transverse valleys. Similar in general dimensions to the Vema fracture, they lack the pronounced bordering walls and trend W 15° N rather than west–east. At least at some locations, they are filled with thick sediments similar to those in the Vema fracture (Figure 8). The northernmost of the three seems to extend into the crestal province of the ridge although the data are somewhat ambiguous; it may simply be aligned with a small dislocation of the crestal zone which has produced a left-lateral offset of approximately 25 km. This part of the transverse disturbance is V-shaped and does not have the box-like cross-section of the Vema fracture and the southern transverse valleys. The middle transverse valley terminates abruptly against the western edge of the crestal ridge province, while the eastern extension of the third lies outside the survey area. The flat floors of all three merge into the abyssal plain to the west at about 4800 m, and rise gently eastward towards the ridge crest.

Earthquake epicenters (made available by Lynn R. Sykes) are located mainly along the south side of the Vema fracture between the two ridge axes (Figure 4). No epicenters occur along the fracture zone beyond these limits. This is commonly the case, although Sykes (1967) and Stover (1968) show that the trace of some fracture zones well beyond the limits of the ridge is marked by sparse epicenters. A first motion solution for one of the epicenters in the Vema fracture is strike-slip; the sense of motion is in agreement with that of a transform fault. All epicenters, with location errors of 10–20 km, are close to the south side of the fracture. A few epicenters occur in and near the northern and southern median rifts.

3. Basement

Reflection profiles (Figures 5, 6) show that the Vema fracture valley contains a thick body of stratified sediment resting on a strong reflector which merges with the exposed slopes of the valley walls. We interpret this reflector as volcanic basement. In the deepest part of the valley, along the southern edge, this reflector is not reached; extrapolation of slopes suggests that the sediment thickness here may be as much as

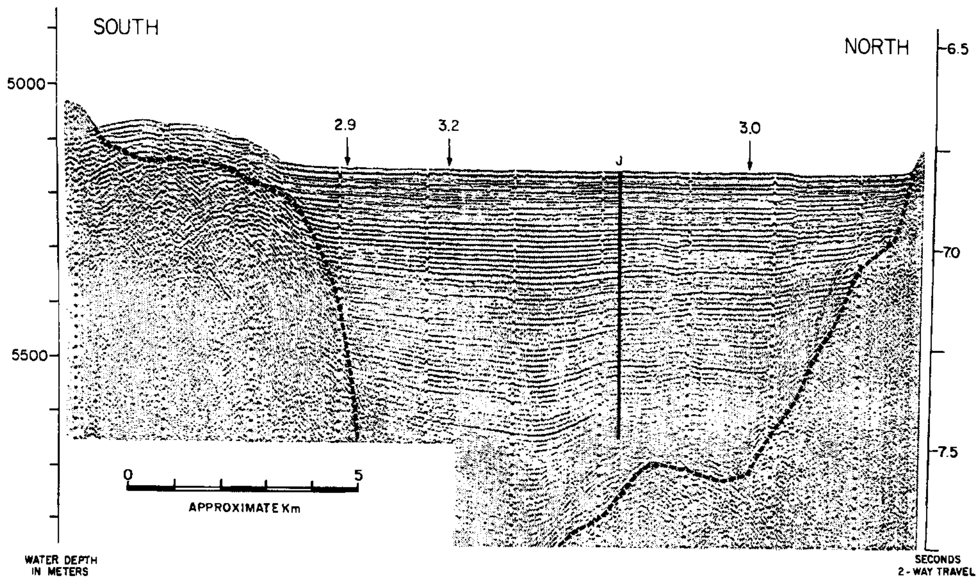

Fig. 5. Seismic reflection profile across center of Vema fracture. Location on Figure 1 (profile A). Heat flow values projected on section. JOIDES drill site (DSDP-26) located near halfway point between 3.0 and 3.2 heat flow values. Vertical exaggeration 40 ×. '*J*' indicates a JOIDES site.

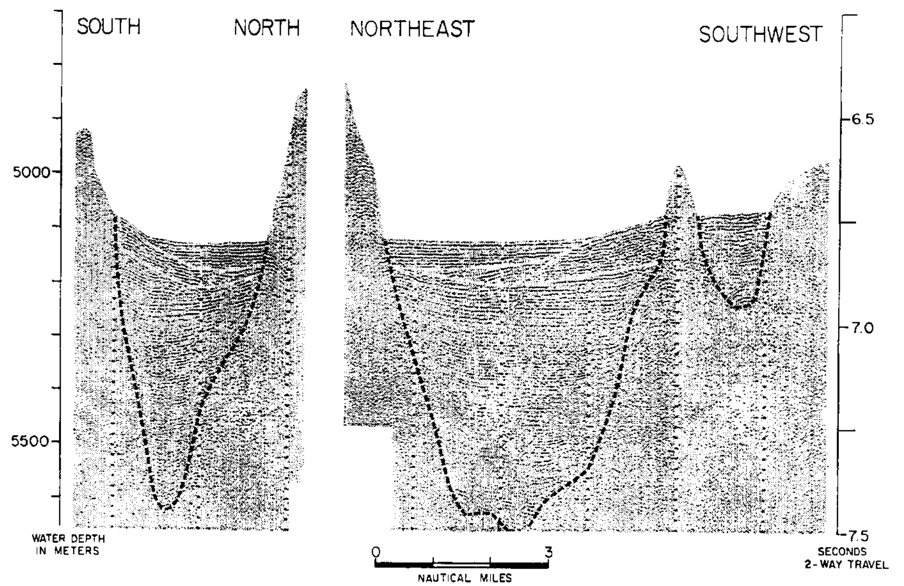

Fig. 6. Seismic reflection profiles across western end of Vema fracture. Locations on Figure 1 (left: profile B, right: profile B′). Vertical exaggeration 30 ×.

1200 m, and that therefore the basement depth is 6200–6300 m below sea level. Underneath the sediment fill the valley floor consists of hills and troughs (Figure 7). A tentative correlation indicates the presence of longitudinal ridges separated by valleys. West of the northern rift segment, these ridges trend approximately W15°N, parallel to the trend of the fracture zone west of the Vema valley proper and to that of the southern transverse valleys. Near 44° 30′ W, the buried ridges emerge and restrict the valley width (Section C, Figure 7). The maximum basement depth of about 5600 m is significantly less than the 6200 m or more observed farther east (Section E). Data of Lamont-Doherty Geological Observatory (courtesy N. Terence Edgar) indicate that basement depth under the Demerara abyssal plain to the west is approximately 5900 m.

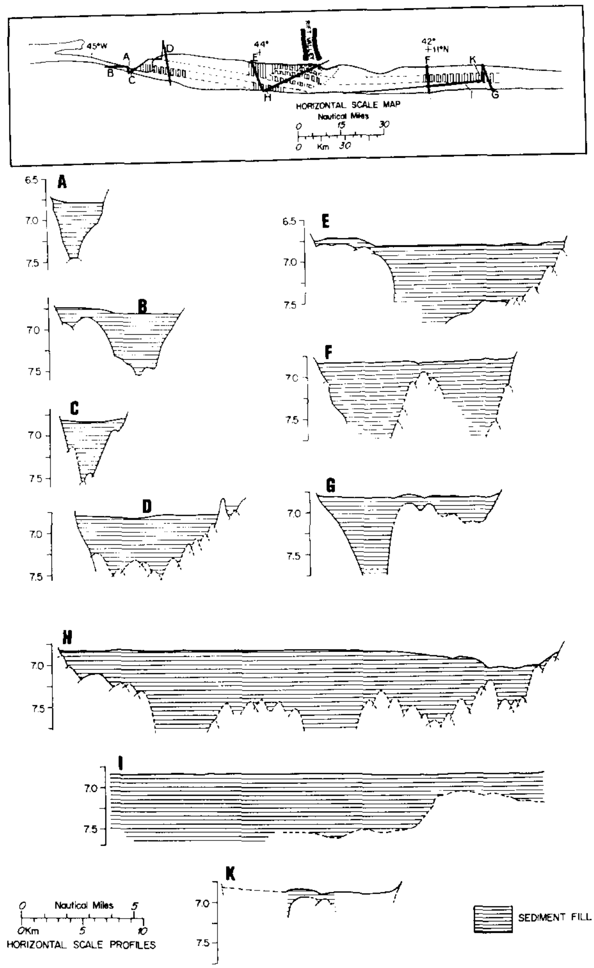

Fig. 7. Basement morphology in the Vema fracture, based on seismic reflection profiles. Map: tracks and inferred buried ridges (vertical hatching); dotted line is 5000 m contour, dashed line 5100 m contour. Sections: basement configuration and sediment fill from reflection profiles; vertical scale in seconds two-way travel time; vertical exaggeration 35 ×.

Thus, the Vema fracture is not only constricted laterally but also has a shallow basement sill at the western end where its trend changes.

A single reflection profile across the northernmost of the southern transverse valleys (Figure 8) shows that it also contains a thick sediment fill, in excess of 900 m, which is indistinguishable from that of the Vema valley itself.

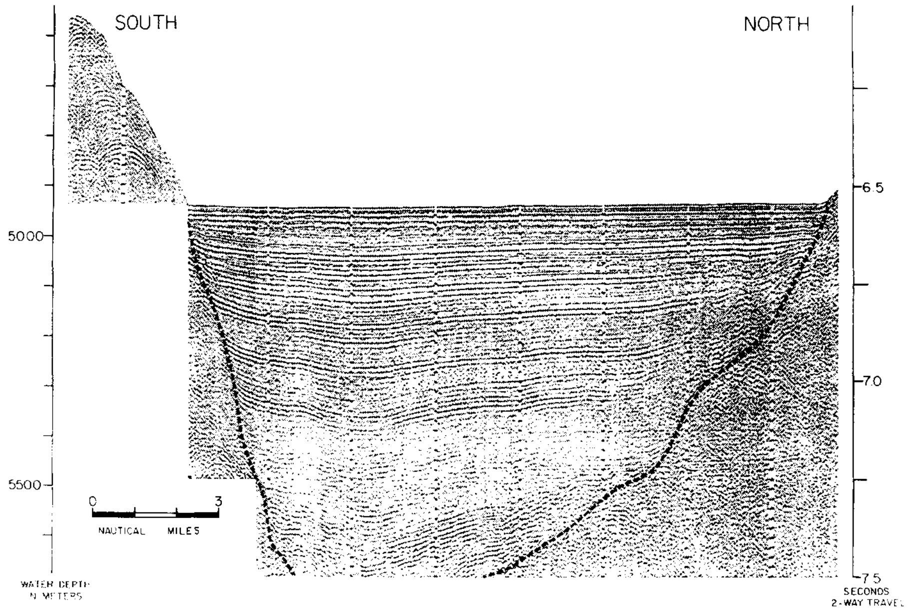

Fig. 8. Seismic reflection profile across first southern transverse valley. Location on Figure 1 (profile C). Vertical exaggeration 40 ×. (From van Andel *et al.*, 1967).

The most impressive topographic features of the region are the walls of the Vema valley, in particular the south wall. This long, narrow, and straight ridge rises 3500 to 5000 m above the depth of basement in the Vema valley and the adjacent Demerara plain. Its steep walls and general shape preclude an origin as a chain of submarine volcanoes, and suggest that it is an uplifted slice of oceanic crust. The same explanation presumably holds for the north wall. Although the crustal structure for this region is not known, data presented by Ewing and Ewing (1959) suggest that an average thickness for the crust of 5 km is a reasonable assumption. Thus, the south wall uplift may represent not only all of layer 2, but also the major part of layer 3.

This hypothesis receives support from the results of rock dredging. Two dredge hauls from the base of the slope, one on the north side of the valley west of the median rift, the other on the south side at 43° 18′ W, yielded serpentinized plagioclase peridotite. Three others from the upper slope of the south wall contained only basalt. According to Melson and Thompson (1971), the peridotites contain high alumina enstatites which originated at high temperature and moderate depth (10 km) and were subsequently emplaced in the lower crust. Uplift of several kilometers then exposed them in the valley walls.

The basalts of the upper part of the wall are similar to those of the adjacent ridge crest segments. Melson and Thompson assume that they are all part of a regional basaltic blanket (layer 2) that is continuously created at ridge crests.

4. Sediment Cover and Recent Tectonic Activity

The sediments of the Vema fracture are evenly and thinly bedded (Figures 5, 6) and identical with those of the first southern transverse valley (Figure 8) and of the eastern Demerara abyssal plain. No major stratigraphic units or key horizons exist. Minor irregularities of bedding suggest sporadic channeling, but in general the process responsible for the deposition of the sediments appears to have operated uniformly over the entire valley floor. Only near the western end of the valley do we find some internal unconformities (Figure 6) with onlapping beds suggestive of erosion, perhaps coupled with vertical movement.

Some broad gentle undulations are probably the result of differential compaction; others may be related to continuing uplift of the longitudinal ridges (Figure 7, profiles F and G). Some sections show dipping contorted bedding near the basement surface that may be the result of penecontemporaneous slumping, for example in the lower center of Figure 5. Broad upwarping over a shallow basement shelf along the south side of the valley may still continue (Figure 5). Thus, there is evidence for moderate vertical movement of the longitudinal ridges and troughs and of the valley walls. On the other hand, there is no indication of horizontal movement or deformation induced by strike-slip motion along the fracture zone. Although the lack of rigidity of a thick body of unconsolidated sediment may perhaps account for the absence of any deformation large enough to be observed by seismic profiling, the evenness of bedding and lack of disturbance are puzzling in such a presumably active tectonic feature. Cores taken in the Vema fracture and in the southern transverse valleys contain an upper layer, approximately 100–150 cm thick, composed of brown calcareous pelagic ooze containing foraminifera and calcareous nannoplankton. This sediment, which is identical with the pelagic blanket covering the uplands of the Mid-Atlantic ridge, probably represents the post-glacial deposition in the area. In the Vema fracture, and in the southern transverse valleys, it overlies a thick section of dark gray to black silty clays and fine sands, composed of continent-derived material with much organic matter. Some beds are composed entirely of plant material (McGeary, 1969). The sands are commonly graded and frequently have small-scale current bedding. The mineral composition and the common presence of reworked shallow water foraminifera identify the Brazilian continental margin and the Amazon River as the principal source (McGeary, 1969). The sediments are identical with those of the Demerara abyssal plain (where the overlying pelagic sequence is lacking). The inference of Heezen *et al.* (1964b) that the fill of the Vema fracture represents a tongue of the abyssal plain and was deposited by turbidity currents thus appears to be substantiated. This, then, also holds true for the southern transverse valleys.

Recently, D. V. Glomar Challenger drilled into the sediments in the Vema fracture valley at 44° W, 10° 54′ N (Site 26, Leg 4), where the total thickness is about 870 m. Cores were taken to a depth of 476 m; the hole bottomed at 600 m in sediment. The deposits consisted uniformly of dark turbidites derived from the Amazon and were of middle to late Pleistocene age (Bader *et al.*, 1970). This gives a sedimentation rate of 120 cm/1000 yr, and an approximate age of a million years for the beginning of turbidite sedimentation. Presumably, the initiation of turbidite deposition in the valley cannot have taken place until the sediment level in the Demerara abyssal plain had reached the top of the 5500 m sill at the western entrance. Consequently the formation of the Vema valley may be significantly older than one million years. The pelagic deposits of the early period have not been reached by the drill, but their nature may be inferred from cores described by Heezen *et al.* (1964b) from the V-shaped portion of the valley beyond the eastern end of the flat floor. Here a totally different lithology is found, consisting of brown pelagic ooze, graded foraminiferal sands and heavy mineral sands of local ultrabasic origin.

Little sediment (probably not more than a few meters) exists on the crest of the northern and southern ridge segments to a distance of approximately 60 km from the ridge axis. Farther down the slope, stratified sediments with level or occasionally with tilted surfaces fill each valley while the slopes are generally bare or covered with thin deposits (Figure 9). Cores consist of brown pelagic ooze interbedded with coarse, sometimes graded foraminiferal sands probably laid down by turbidity currents of local origin (van Andel and Komar, 1969). Normal faults with small vertical throw occur in several valleys. These faults usually reach the surface and are usually of the sense to indicate continuing uplift of the crestal side of the valleys. Thick sediment bodies with tilted stratification occur on some slopes; cores from such deposits are similar to those of the valley floors and show that the deposits were originally deposited horizontally.

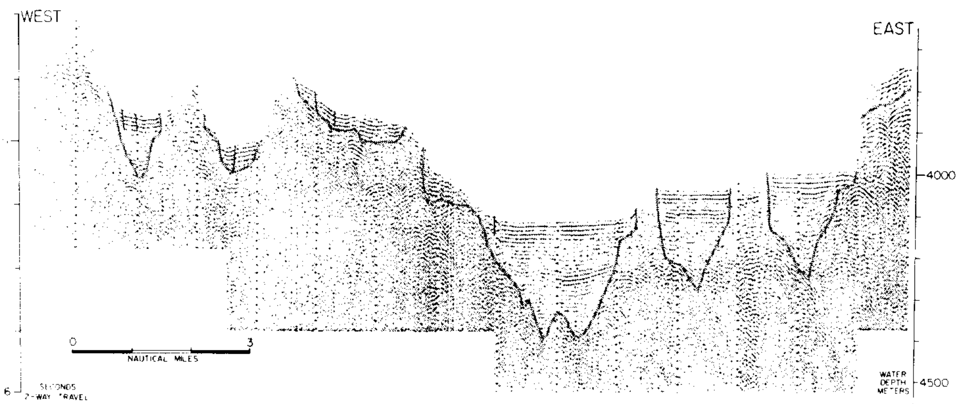

Fig. 9. East–west seismic reflection profile across part of east flank of northern ridge segment. Location on Figure 1 (profile D). Vertical exaggeration 20×. Superimposed dashed boundaries illustrate basement topography.

Significant recent uplift of the ridge crests is also indicated by a marked eastward rise of the flat floors of the southern transverse valleys. Since the deposits in these valleys were laid down by turbidity currents from the Demerara abyssal plain, they must initially have been horizontal or have sloped gently to the east. At present, they rise 500 to 1000 m from the edge of the abyssal plain to the edge of the crestal province of the Mid-Atlantic ridge, implying a recent uplift of the ridge crest of a kilometer or more (van Andel, 1969).

5. Heat Flow

The heat flow measurements in the Vema fracture area have been superimposed on the bathymetric map (Figure 2). The results have been discussed in detail by Von Herzen *et al.* (1970).

As with other sets of measurements close to the axis of mid-ocean ridges, the values in this region show a wide range: 0.44 to 5.06 HFU (10^{-6} cal/cm^2/s), or more than one order of magnitude. However, a plot of all values, except those from the Vema fracture proper, against increased distance from the ridge axis (Figure 10) shows a double-

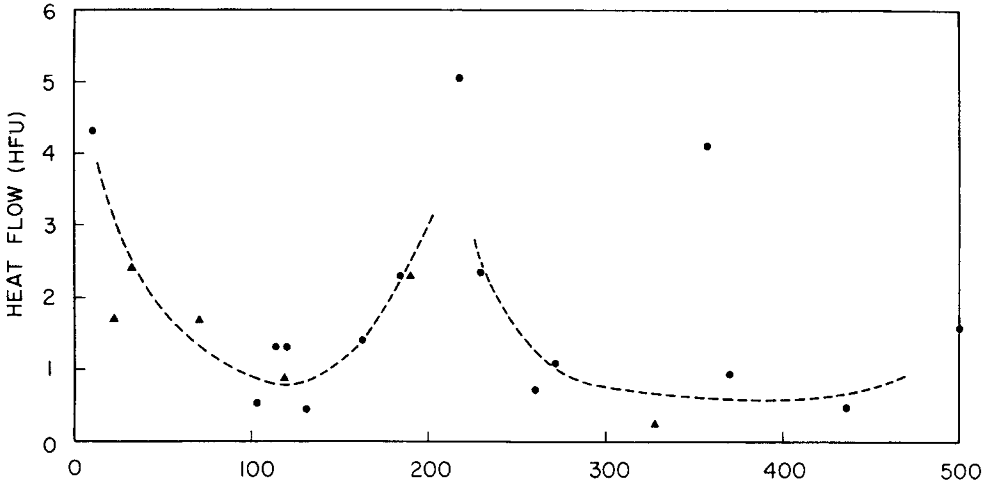

Fig. 10. Heat flow values from the Mid-Atlantic ridge near the Vema fracture zone plotted against distance from the ridge axis. Values for Vema fracture itself are omitted. Triangular symbols are values obtained prior to the study of Von Herzen *et al.*, 1970. (After Von Herzen *et al.*, 1970, Figure 6.)

peaked pattern rather than the characteristic monotonic decrease normally seen (Vacquier and Von Herzen, 1964; Lee and Uyeda, 1965). The secondary peak at about 200 km from the axis is supported by 2.34 and 5.06 HFU south of the Vema fracture and 2.30 HFU north of the fracture (Figure 2). There is no obvious correlation with the topography, and other than the assumption that it may have occurred by random sampling of a complex pattern we have no satisfactory hypothesis for this observation.

The six measurements in the Vema valley have a mean of 3.0 HFU and a standard deviation of only 0.2 HFU. This uniformity is characteristic only for this fracture valley. Combined with the thick fill of the valley, this uniformity in heat flux is analogous to that found in the abyssal plains between the Mid-Atlantic ridge and the continents (Langseth *et al.*, 1966). Numerical studies of Vema fracture models (Von Herzen *et al.*, 1970) show that even a complex structure of the basement in the valley under the thick sediment cover would produce no more than 10% variability across the filled-in portion except near the extreme edges. Thus, the heat flux uniformity is consistent with uniform heat production at depth beneath the fracture zone.

The high rate of sedimentation noted earlier significantly decreases the surface heat flux compared to that at greater depth. The magnitude of equilibrium flux may be of the order of 5–6 HFU, a value comparable to the largest fluxes measured over oceanic ridges. Two fundamentally different causes may exist for such a large flux. One is the oxidation of organic matter which is a significant component of the sedimentary fill; the other is tectonically generated heat, either as a result of friction generated by strike-slip faulting, or in association with a widening of the valley accompanied by the generation of new oceanic crust (van Andel *et al.*, 1969).

In contrast to the high flux in the Vema fracture, the first southern transverse valley is characterized by a lower than normal flux. The sedimentary fill of this valley is similar to the Vema valley, and the mechanism of deposition, deposition rate, and nature of the material are also comparable, but the valley appears to be tectonically quiet. Therefore, the hypothesis of high heat flux associated with active tectonism seems favored by the evidence and, in the absence of evidence for significant shear, the heat flow observations appear to support the rifting hypothesis.

6. Magnetics

Numerous shipboard geomagnetic traverses have been made in the vicinity of the Vema fracture zone. The location of the ship tracks along with their respective anomaly profiles is shown in Figure 11. The total intensity anomalies have been calculated using the International Geomagnetic Reference Field (IAGA, 1969). All data were collected with marine proton precession magnetometers. Inspection of the profiles reveals that the ridge crest north of the Vema fracture zone is clearly marked by a strong positive anomaly greater than 200 γ amplitude. Flank anomalies of generally less than 100 γ amplitude trend parallel to and are symmetrically arranged about the ridge crest. Except for the relatively low anomaly amplitudes, the anomaly pattern here is typical of those observed over other portions of the northern Mid-Atlantic ridge crest (Phillips, 1967 and Phillips *et al.*, 1969). The symmetry and linearity of the pattern is best exhibited by the anomaly features labelled I and II in Figures 11 and 12. The sharp left lateral offset of the anomaly pattern near 12° N latitude suggests that another fracture zone similar to Vema, but with a smaller offset ($\approx$ 100 km) may exist here. South of the Vema fracture, the magnetic coverage is less dense and the available ship tracks are not oriented normal to the presumed north–south trend of the ridge

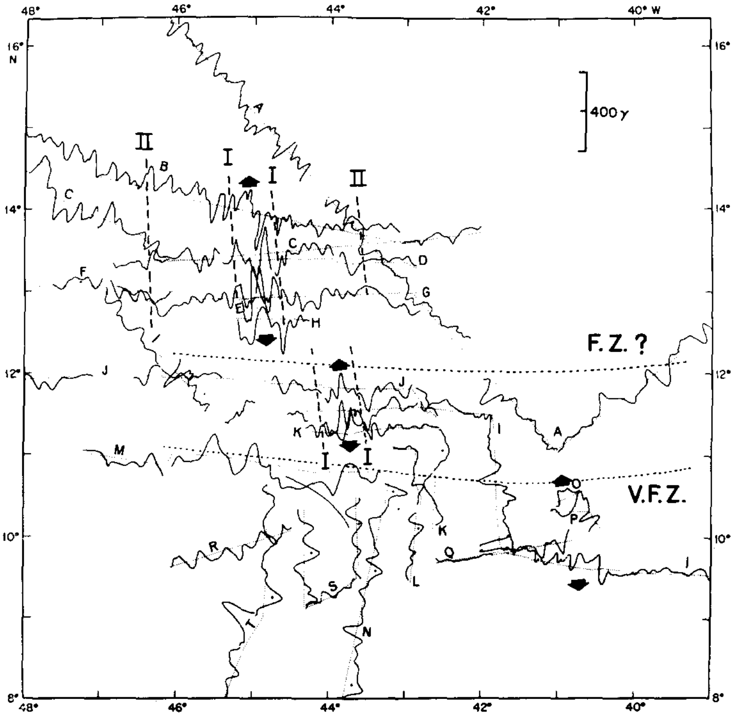

Fig. 11. Magnetic anomaly profiles along ship tracks in vicinity of the Vema fracture zone (VFZ). The bold arrow-tipped lines indicate segments of the Mid-Atlantic ridge crest. The heavy dotted lines represent transverse fracture zones Letters A to T identify the various ship tracks: A and B – Trident; C to H – Chain; I – Argo; J, L, O, P, Q, S and T – Atlantis II; K and R – Thomas Washington; M and N – Conrad.

crest. Thus, it is not possible to recognize the ridge crest from its magnetic signature; let alone to discern any symmetry in the flank anomaly pattern. At distances greater than 100–150 km from the ridge axis both north and south of the Vema fracture, it becomes increasingly difficult to recognize any linear magnetic trends. This might be expected in view of the low amplitude, short wave length of the anomalies and the wide ship track spacing.

The marked linearity and symmetry of the ridge crest magnetic anomalies north of the Vema fracture suggests that the Vine and Matthews (1963) hypothesis for the origin of magnetic anomalies over ocean ridges may be applicable here. Accordingly, various theoretical magnetic anomaly profiles have been computed in an effort to match observed profiles with simulated sea-floor spreading model profiles. Figure 12 shows a comparison of several observed profiles across the ridge crest between 12° and 14° N latitude with a simulated profile generated by a sea-floor spreading rate of 1.2 cm/yr. If this rate is correct, anomalies I and II would correspond to those anomalies of the Heirtzler *et al.* (1968) geomagnetic polarity time scale which are approximately 2.5 and 11 million years old respectively. Although the correlation of the

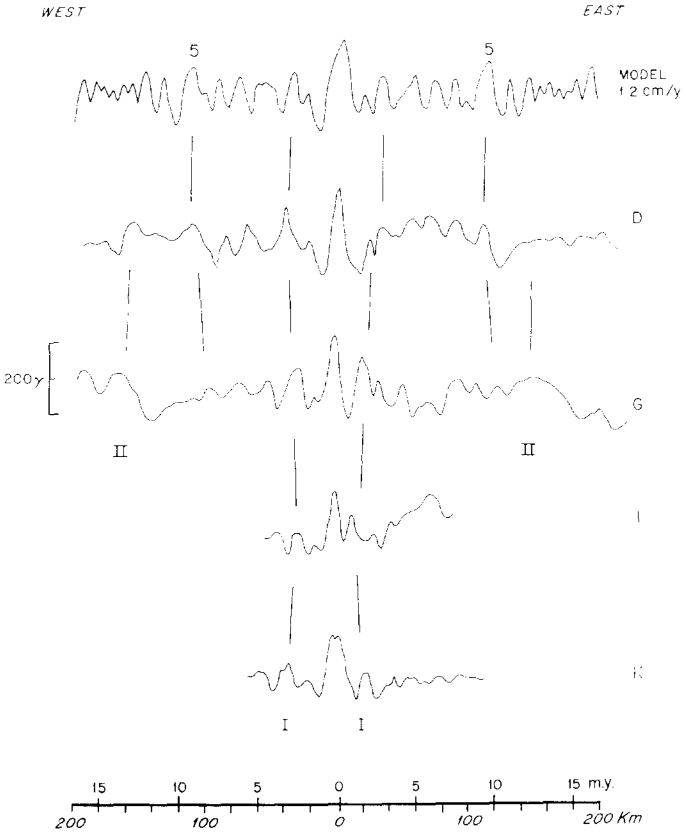

Fig. 12. Symmetrical magnetic profiles. At the top is a simultated profile generated from a sea-floor spreading model for the Mid-Atlantic ridge at 14° N. The upper surface of the magnetic layer is taken at the approximate sea-floor depth of 3.5 km. The layer is 1.7 km thick. The configuration of normally and reversally-magnetized blocks in the layer was determined from the Vine (1968) geomagnetic polarity time scale. The time scale (millions of years ago), shown at the bottom of the figure is related to the distance scale by a 1.2 cm/yr spreading rate. Total intensity of rock magnetization is 0.0035 cgs units, except for central block which is 0.0070. The inclination of magnetization is 26°. The ridge axis strikes 350°. The simulated profile was calculated using a modified version of a computer program described by Talwani and Heirtzler (1964).

anomaly shapes and amplitude are not considered striking, the overall 'fit' is considered good in view of the low anomaly amplitudes inherent with N–S trending ridges in low magnetic latitudes and the generally complex opening history envisioned for the equatorial Atlantic (Funnel and Smith, 1968; Ball *et al.*, 1969; Ball and Harrison, 1970). Geometrical argument requires that the separation of North and South America away from Africa must be accompanied by a gradual increase of the ridge crest length as well as shearing and extensional motion. Such effects might be expected to progressively obliterate the original linearity and symmetry of the magnetic anomaly pattern as the older crust moves away from the ridge crest.

7. Discussion and Conclusions

The existence of fracture zones has been established only recently (Menard and Dietz, 1952). Since they were originally considered to be transcurrent faults, an interpretation supported by then current views on magnetic anomaly patterns, the explanation of their role in ocean floor tectonics encountered severe difficulties. These were resolved by Wilson's (1965) recognition that, as a consequence of sea-floor spreading, they represented a new class of faults with a sense of motion opposite to that of a trans-

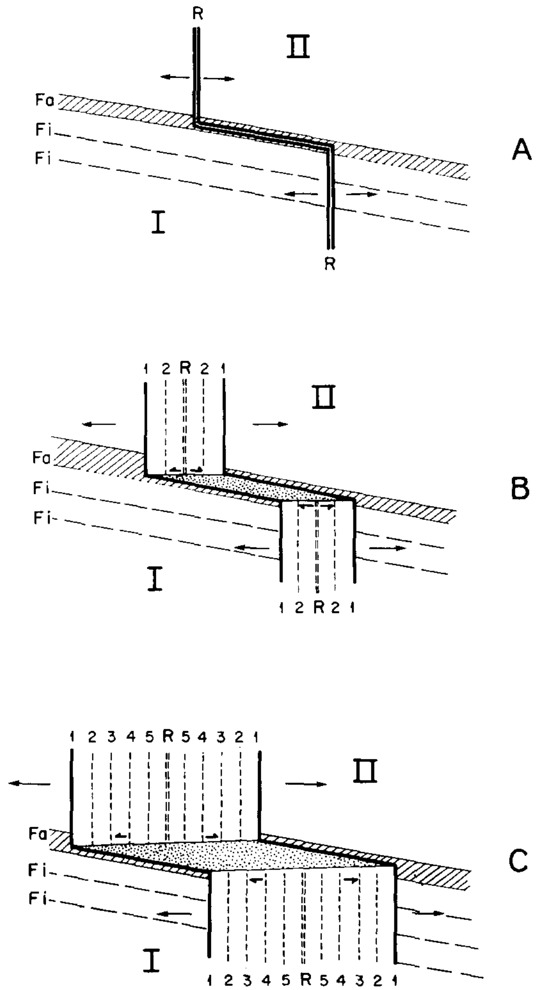

Fig. 13. Model explaining rift nature of Vema fracture. Stage A shows incipient movement by spreading from two ridge crests (R) offset along old fracture zone (F_a) trending at angle to direction of spreading. Other old fracture zones are inactive (F_i). Progressive spreading and formation of new crust (stages B and C) produces an open rift (stippled) along old fracture line. Successive growth stages indicated by numbers. Heavy line marks initial boundaries of blocks I and II. After van Andel *et al.* (1969).

current fault. Since that time, it has been customary to interpret all fracture zones as transform faults.

Several features of the Vema fracture zone suggest that a more complex explanation is required. Because of the high heat flow, the parallelogram shape of the valley, and the parallelism between the southern valleys and the extreme ends of the Vema fracture, we have proposed (van Andel *et al.*, 1969) that the present Vema fracture is the result of a reorientation of plate movements in the Atlantic. This reorientation rendered several fracture zones inactive – the southern transverse valleys – and changed the direction of motion along one of them, now the Vema fracture, from W10°N to W–E. A passive N–S extension of the Vema fracture, with concomitant intrusion of new crustal material, was the consequence of the reorientation (Figure 13). The rate of crustal extension across the Vema fracture is, as a result of the geometric relations between the orientation of the fracture and the direction of spreading, slower than that across the adjacent rift zones. A similar idea was put forth by Menard and Atwater (1968, Figure 4c) to explain magnetic anomaly patterns in the Pacific.

Since the inactive southern transverse valleys terminate against the western edge of the southern crestal province, the age of the crust at this point is also the time of reorientation. Unfortunately, magnetic data for the Vema fracture region are fragmentary and located along poorly oriented tracks. Thus, the age of the crust on the southern ridge segment cannot be determined directly. If we assume, however, that the spreading rate does not vary greatly over a distance of several degrees latitude, data from good traverses between 11° and 14° N can be used to approximate the age of the edge of the crestal province (see Figures 11 and 12). Using an average spreading rate of 1.2 cm/yr the period since anomaly 5, the reorientation must have occurred approximately 10 million years ago. This time has been postulated on other grounds as a time of change in the rate of crustal plate motions in other parts of the Atlantic (Ewing and Ewing, 1967; Schneider and Vogt, 1968; van Andel and Bowin, 1968; Maxwell *et al.*, 1970; van Andel and Heath, 1970). Fox *et al.* (1969b) in charting the trend of a fracture zone near 24° N latitude prefer a change in the poles of rotation of the Atlantic plates at this time. However, Phillips and Luyendijk (1970) report there is little evidence for such a change from a similar study of the Atlantic fracture zone at 30° N latitude. Until good anomaly and fracture zone patterns are available both north and south of the Vema fracture, this problem cannot be solved.

A 10 million year age for the beginning of the Vema fracture valley is much in excess of the estimated 1 million years required to produce the turbidite fill. This discrepancy can be explained if the sill at the western end of the valley served as a barrier for near-bottom transport until the abyssal plain had been filled to that level 1 million years ago. Alternatively, turbidity current transport on the Demerara abyssal plain may not have started until the onset of the Pleistocene. In either case, pelagic sediments up to 10 million years in age should be found above basement rocks in the Vema fracture.

The previous hypothesis for the origin of the Vema fracture (van Andel *et al.*, 1969) did not elaborate on the mechanism of formation of north and south walls and the deep depression between them, and supplied no mechanism for the large amount of differential uplift that they require. Neither does it explain the recent uplift of the northern and southern ridge crests. The latter, however, may not be directly related to the dynamics of the fracture zone. Maxwell *et al.* (1970) and van Andel and Heath (1970) have presented evidence for several phases of ridge uplift elsewhere in the Atlantic, possibly in relation to changes in spreading rate.

High bordering walls and a deep central depression, although unusually well developed in the Vema fracture, are common to all Atlantic fracture zones that have been studied in some detail (Heezen *et al.*, 1964a; Heezen and Tharp, 1965; Fox *et al.*, 1969a; Fleming *et al.*, 1970). Therefore, it seems desirable to develop an explanation of the topography which is generally applicable to the Atlantic fracture zones.

Sleep and Biehler (1970) have proposed a qualitative model to explain some topographic features at the intersections of fracture zones and rifts, based on the viscodynamic behavior of upwelling mantle material in rifts. The viscosity and/or wall friction associated with upwelling material create the median depressions associated with rifts and at their intersections with fracture zones. We will combine this physical model with our model of formation of the Vema fracture (and perhaps others) to explain the general topographic features.

Sleep and Biehler point out that the loss of hydraulic head near the intersection of a ridge crest and a fracture zone should cause a local depression there, and indeed we have evidence for a depression in the Vema valley at the exit of the northern ridge crest into the valley (Figure 2). Directly following the qualitative reasoning of Sleep and Biehler, we may postulate that the subsequent recovery of lost head due to cooling and welding of the new crustal material to the already consolidated plate results in the uplift of this depression along with the surrounding topography. This regional recovery is caused by an increasing thickness and, presumably, increasing strength of the lithosphere as cooling proceeds. Given the condition that the deep fracture zone is welded to the spreading plate, the high walls of the fracture zones are thus accounted for as the product of recovery of hydraulic head loss.

This model encounters several difficulties when applied to the features of the Vema fracture zone. The great height and very limited width of the walls imply only a limited ability of the lithosphere in this area to transmit stresses laterally. This indicates a thin and not very rigid crust, in contrast to the assumption made above of a lithosphere of increased thickness and strength. Secondly, the assumption of lateral cooling, away from the intersections of ridge crests and fracture zone, should produce a systematic variation of heat flow along the fracture valley floor. Instead, we find a remarkable uniformity of heat flux along at least 300 km of the valley floor – an observation which is unique in the distribution of oceanic heat-flow values near ridge crests. Thirdly, the basement topography of at least the western part of the Vema fracture appears to have a marked lineation which is approximately parallel to a line connecting the two rift valleys. Finally, the sediments of the fracture valley are essentially undisturbed, a fact that is difficult to reconcile with spreading centers or the formation of new crust within or attached to the fracture valley.

We propose a modified application of the Sleep and Biehler mechanism, based on a refined version of the model of van Andel *et al.* (1969), which better accounts for these observations. We begin with the re-activation of spreading of a pre-existing ridge-ridge transform fault in a somewhat different direction (Figure 14A). If the rate of spreading from the ridge axes is v, the geometry of the changed direction imposes a rate of spreading perpendicular to the transform fault of $v \sin \theta$, where θ is the difference in

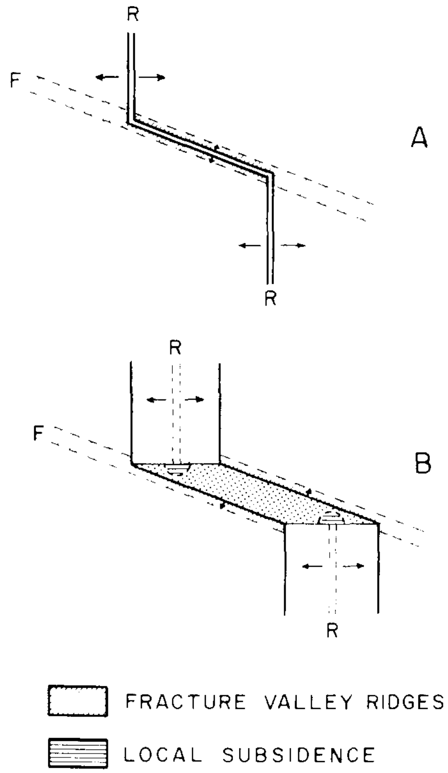

Fig. 14. Diagram of development of fracture zone walls according to model explained in text. (A) Previously existing ridge (*R*) and transform fault (*F*) system at moment of reorientation of spreading direction. (B) Geometry of main topographic features after some time of renewed spreading. Stippled: sediment fill.

angular direction between the old and new spreading directions. As θ is small ($\leqslant 10°$) for the Vema and presumably also for other Atlantic fracture zones, the rate of spreading perpendicular to the transform fault will be much smaller than that normal to the ridge axis. Hence the formation of new crust along the fault will probably occur by intrusion of thinner dikes than along the ridge. Such thin dikes will be associated with higher rates of cooling and higher viscosity. According to the principles of Sleep and Biehler, the greater loss of hydraulic head that accompanies these thinner intrusions will result in the formation of higher ridges along the transverse fault lines than along the ridge crest (and a deeper depression in the fracture zone than in the rift zones). Since we observe that these ridges are narrow, we must also assume that the lithosphere is locally thin.

As the ridge and fracture valley spread (Figure 14B), we assume that active growth of the fracture valley occurs along the margins of the valley at the edges of the valley walls, rather than in the center as is the case for the rift valleys on the ridge crests. This hypothesis satisfies two observations:

(1) It permits the continued existence for the duration of the spreading process of the high and narrow fracture walls as compensation for the hydraulic head loss along their inner margins.

(2) It leaves the sediments in the central part of the fracture valley undisturbed, since adjustments only take place on the valley margins. We note that our model permits the valley walls to extend only to the boundary of the new ridge crest material formed since the onset of re-oriented spreading, as indeed seems to be the case for the Vema fracture (Figure 4). The linear basement ridges, observed in the western part of the fracture valley (Figure 7), may represent periods of intermittent or alternate growth on opposite sides of the valley. Overall symmetrical growth of the whole ridge system, however, implies that, when integrated over a sufficiently long period, the growth of the fracture valley must also be symmetrical. If during part of the spreading period, fracture valley growth occurred only on one side of the valley, this implies that faulting would have to occur principally on the other side in order to accommodate symmetrical ridge growth. The earthquake epicenter distribution of Figure 4 may indicate this. This situation must, from time to time, reverse itself.

There are several consequences of the model which may be tested:

(1) the basal sediments in fracture zones of this type should become generally younger from the center to the edge, opposite to the age sequence of sediments on spreading ridge crests. This can only be tested by deep drilling.

(2) The width of the fracture zone should be directly related to the spreading rate of the ridge, the angle between the old and new spreading directions, and the time elapsed since the rejuvenation of spreading. For the Vema, a spreading rate of 1.2 cm/yr and a change in spreading direction of 10° ten million years ago, give a width of 21 km. This is close to the average width of Vema fracture valley.

(3) As the fracture zone widens, the recovery of lost hydraulic head should produce an uplift in the older central portion resulting in a median ridge. The basement ridge that exists in the portion of the Vema fracture between the north and south rift (Figure 7) may represent this uplift. A prominent central ridge also occurs in the well-surveyed Gibbs fracture zone at 53° N (Fleming *et al.*, 1970), which led these authors to postulate a double fracture zone. In contrast to the conclusion of Fleming *et al.*, we suggest that the Gibbs fracture zone might also be the product of a change in spreading direction, and that the central dividing ridge is produced by adjustment for lost hydraulic head as described in our model. The adjustment by means of uplift would account for the earthquake epicenters located near this median ridge. The outer walls of the Gibbs fracture zone, although not as prominent as those of the Vema

fracture, define a width of 50 to 60 km for the fracture valley system (Fleming *et al.*, 1970, Figure 3), which is more than twice the width of the Vema. The greater width of the Gibbs valley and the prominence of the central ridge as compared to the Vema, would be explained by a higher spreading rate and/or a larger angle of reorientation.

The proposed mechanism should provide a uniformity of heat flow along the fracture valley as we have observed in the Vema. The uniformity *across* the valley is probably the result of conduction from heat sources at depths comparable to or greater than the valley width, combined with the smoothing effect of the thick valley sediments (Von Herzen *et al.*, 1970). Local anomalies associated with vertical transport of magma should exist only at the edges of the fracture zone, but these might be quite narrow and difficult to observe.

The models discussed here provide a possible explanation for the major features of Atlantic fracture zones. Since they have significant consequences for the nature of the oceanic crust (see, for example, the petrologic models proposed by Miyashiro *et al.*, 1970, and Melson and Thompson, 1971), gravity and seismic refraction studies as well as additional detailed structural investigations of other Atlantic fracture zones are needed. Whether the qualitative model can indeed account for the vertical movements of the walls and valley floors can probably be evaluated only when enough field parameters are available to develop them in a quantitative form. The Atlantic fracture zones seem to be systematically different from those in the Pacific, which consist of zones of smaller elongate ridges and troughs. Although absence of reorientation might account for this, it is tempting to relate these differences in morphology to differences in spreading rate between the two ridge systems as well, since the rate of spreading with its effect on cooling rate is an important factor in our models.

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