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THE TECTONIC FABRIC OF THE EQUATORIAL ATLANTIC AND ADJOINING CONTINENTAL MARGINS: GULF OF GUINEA TO NORTHEASTERN BRAZIL

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TO MY DEAR PROFESSOR OTHON HENRY LEONARDOS FOR HIS TEACHINGS, ENTHUSIASM AND KINDNESS THROUGHOUT THE YEARS THAT I BECAME A GEOLOGIST

TO MY PARENTS

TO MY WIFE

ABSTRACT

THE TECTONIC FABRIC OF THE EQUATORIAL ATLANTIC AND ADJOINING CONTINENTAL MARGINS: GULF OF GUINEA TO NORTHEASTERN BRAZIL

MARCUS AGUIAR GORINI

Fracture zones are very prominent and generally linear basement features which bound segments of oceanic crust and offset the mid-oceanic ridge. Major fracture zones offset the Mid-Atlantic Ridge in the equatorial region by distances ranging from 300 to 900 km. St. Paul's, Romanche, Chain, Jean Charcot - Fernando de Noronha and Ascension fracture zones were mapped from the African to the Brazilian coast across the equatorial Atlantic. Each of these fracture zones display a ridge and trough morphology that can be traced laterally (when available data permitted) from the offset region of the Mid-Atlantic Ridge to the continental margins on either side of the Atlantic. Mapping of the basement features (ridges and troughs) in the equatorial Atlantic reveals that a marked east-west basement fabric exists for the entire ocean floor and that the Mid-Atlantic Ridge axis is asymmetrically located toward the west. The fracture zones in the equatorial Atlantic vary considerably in width ($\geq$ 50 km), complexity, trend and morphology along their strikes. They extend nearly continuously from the Brazilian shield to the West African shield and, divide the ocean floor into segments bounded by linear ridges and intervening troughs. In the continental shelves, horst and graben structures occur laterally along the continuation of fracture-zone trends and half-graben basins occur in the continental shelves in the crustal segments between these fracture-zone trends. The fracture zone trends were established at the on-

set of rifting; these trends did not necessarily originate along old weakness zones in the Precambrian shields and platforms.

Marginal fracture ridges which occur in the continental margin of the equatorial Atlantic are very prominent physiographic features which are lateral continuations of the transverse ridges of fracture zones in mid-ocean. The very high relief and the youthfulness of volcanism in some of the marginal ridges suggest that tectonic adjustments have been taking place along fracture zones at distances far from the offset region of the Mid-Atlantic Ridge axis.

Other important structural features which are intimately related ro rifting in the equatorial Atlantic and which are examined in detail in the present study are the Marajō system of grabens, the Benue Trough and the Cameroon structural trend. The Marajō system of grabens (in the Amazon area) of Brazil originated from the southward propagation of the rifting direction that was probably associated with the opening of the North Atlantic Ocean. The Benue Trough in Africa is flanked by fracture-zone directions that imprinted a horst-and-graben setting in the basin. Folding in the Benue Trough is probably caused by vertical tectonism associated with the uplift and differential tilting of buried basements horsts. This vertical tectonism may have been caused by adjustments to changes in transform motion between the African and South American plates. The Cameroon Trend is a horst-like feature whose origin may be associated with the break-up of the continents and is probably linked with the reactivation of a Precambrian lineament that is represented in both sides of the Atlantic by the Pernambuco lineament in Brazil and by the Ngaoundērē fault zone in West Afri ca. The tectonic reactivation of this lineament caused the occurrence of magmatism and the development of horst-like features that acted as a physiographic barrier to salt deposition in the South Atlantic during the Aptian. To the north of the lineament no salts were deposited whereas a large salt basin existed to the south. The Cameroon Volcanic Line has been reactivated since the late Mesozoic by complex magmatism (granites, basalts and alkaline rocks) and block faulting. On the Brazilian side, the Pernambuco lineament was also reactivated in the late Mesozoic. The Cameroon Trend is apparently a landward continuation of the Ascension Fracture Zone.

The remarkable geological fit of Africa and Brazil in the equatorial Atlantic suggests that the rifting and subsequent drifting of the two continents did not involve appreciable crustal distortion of the two continents.

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INTRODUCTION

The oceanic fracture zones constitute an important feature of the plate tectonic hypotesis because they are believed to be regions where slip motions between two plates are accomodated. Ever since Wilson (1965) introduced the term "transform fault", one basis for defining plate motions has been the assumption that the strike of a fracture zone lies along a small circle about a "pole of rotation" (Le Pichon, 1968). Consequently, systematic changes in the strike of a fracture zone have been taken to indicate changes in relative motion of the adjacent plates (Pitman and Talwani, 1972, Le Pichon and Hayes,1971) . However, the definition and the mapping of fracture zones in the ocean floor through topographic, seismic, and magnetic data has not been carried out through application of uniform criteria. Therefore, fracture zones have been alternately defined and/or mapped as linear scarpments or linear troughs and ridges depending upon which feature was most easy to follow. The need for a better understanding of the structures which constitute a fracture zone is apparent from the fact that the majority of published information concerns only a small portion of the entire length of a given fracture zone such as the offset regions within the mid-ocean ridge crest.

The equatorial-Atlantic floor displays several fracture zones which cause large offsets of the ridge axis. Extensive seismic profiling has been carried out in this area since 1970 by various oceanographic laboratories. This data has permitted a study of the features associated with fracture zones and of the lateral variations within a fracture zone. In the present study the equatorial Atlantic fracture zones are shown to consist of unique topographic and basement features. These fracture zones can be traced away from the ridge crest into the continental margins of Africa and Brazil. The relationship of these fracture zones to the structure and geologic history of the marginal basins in the continental areas of Africa and Brazil shows that fracture zones have played an important role in the early rifting history of these continents.

The equatorial Atlantic ocean floor has been an area of oceanographic surveys since the middle of 19th Century. During the cruise in 1832 of H.M.S.Beagle, Darwin landed on Saint Paul's Rocks where he collected rock samples. The H.M.S. Challenger (1872-1873) also visited the islets and her scientists collected several rock samples that were studied in detail.

The scientific interest in the equatorial Atlantic floor started in 1883 when the French research vessel Romanche discovered an anomalous depth of 7370 m at 00º11'S and 18º15'W (Heezen et al., 1964). subsequently confirmed by other oceanographic cruises, this region of very great depth in the Atlantic Ocean came to be known as the Romanche Trench.

The Romanche Trench aroused considerable scientific interest because oceanographic studies of the deep circulation predicted that a sill depth of more than 4000 m should exist in its vicinity which would enable the Antarctic Bottom Water to flow from the western to the eastern Atlantic basins. The R/V Chain, R/V Vema and R/V Meteor conducted surveys in the region of the Romanche Trench that culminated in understanding of the topography, hydrography and sedimentation close to that topographic feature (Heezen et al., 1964; Metcalf et al., 1964;Tomczak and Annutsch, 1970). A maximum depth of the Romanche Trench of 7856 m was established by Vema cruise 12; this location was named Vema Deep (Heezen et al., 1964), and represents the deepest portion of the Atlantic Ocean except for the floor of Puerto Rico Trench.

Heezen et al. (1964) presented a detailed topographic map of the equatorial Atlantic between 10ºW and 20ºW; 3ºN and 3ºS. This work showed an orthogonal relationship between the mid-oceanic ridge axis and the fracture zones and defined the Romanche and Chain fracture zones. Subsequently, Heezen and Tharp (1965) named Saint Paul's Fracture Zone, after the St. Paul's Rocks, to the north of Romanche Fracture Zone.

Tomczak and Annutsch (1970) on the basis of an extensive bathymetric and hydrographic survey by R.V.Meteor in 1965, presented a detailed topographic map of the Romanche Fracture Zone from approximately 17º45'W to 24º15'W and from 1ºN to 3ºS. The area to the south of Romanche Fracture Zone, including the westward continuation of Chain Fracture Zone, was also shown.

These hydrographic and bathymetric surveys constituted the basic information for attempts at combining the studies of the adjacent continental margins with structures in the mid-ocean. The existence of very prominent fracture zones in the mid-oceanic region and prominent ridges in the continental margins (Hayes and Ewing, 1970; Burke, 1969; Fail and others, 1970) led to several important papers that dealt with the relationship of mid-oceanic and continental margin structures. The westward continuation of Romanche Fracture Zone to 33ºW was inferred by Hayes and Ewing (1970) (Fig. 1). Le Pichon and Hayes (1971), and Francheteau and Le Pichon (1972) predicted that Romanche Fracture Zone extended into one of the segments of the North Brazilian Ridge and also into a marginal basin in the Brazilian continental shelf.Cochran (1973) also considered the possibility of Romanche Fracture Zone extending into one of the segments of the North Brazilian Ridge,although he pointed out that no sufficient data was available at that time. Bryan et al. (1973) considered the equatorial fracture zones to be wides zones of fracturing and traced the Romanche Fracture Zone into the North Brazilian Ridge. Gorini and Bryan (1974) also showed that Romanche Fracture Zone could be continuously traced into the southern segment of the North Brazilian Ridge. Gorini et al. (1974) also indicated that Fernando de Noronha Ridge (including Fernando de Noronha Islands and Atol das Rocas) actually lies along the continuation of a fracture zone (Fernando de Noronha Fracture Zone) to the south of the Romanche Fracture Zone.

The eastward continuation of the equatorial Atlantic fracture zones was tentatively demonstrated by several workers in the African side. Burke (1969)and Burke et al. (1971) related seismic activity in the Accra region of Ghana as possibly caused by the association of a prominent fault in the continent with an oceanic fracture zone that he inferred to be the Chain Fracture Zone.He also presented a map showing the extensions of what he considered to be the Romanche,Chain and the 8ºS fracture zones.Using detailed bathymetric,magnetic and seismic reflection data Fail et al.(1970),tentatively traced the Romanche Fracture Zone into the slope off the Ivory Coast-Ghana Ridge and speculated about the possibility that the trend continued into the Benue Trough. Chain Fracture Zone was followed as far east as 5ºW . Arens et al.(1971) associated the trend of the continental slope off the western Ivory Coast and Liberia,with the Saint Paul's Fracture Zone.

Delteil et al.(1974) using more detailed surveys established that the continental margin of the Gulf of Guinea is divided into structural blocks bounded by the equatorial Atlantic fracture zones. They also recognized the tectonic importance of the Romanche Fracture Zone in the early geological history of the Ghanaian continental shelf and traced farther eastward, up to 2ºE, the extension of the Chain Fracture Zone. Delteil et al. (1974) also defined a new lineament under the continental rise as the Jean Charcot Ridge which they were able to trace to approximately 4.5ºE. Mascle (1975) later extended the Chain and Jean Charcot lineaments into the Benue Trough.

Behrendt et al. (1974) studied the Liberian continental margin and defined three ridges in the margin which they thought may possibly connect with the eastward extension of the Saint Paul's Fracture Zone.

Emery et al.(1975) presented an overall compilation about the continental margin off Western Africa and showed maps displaying the extention of equatorial Atlantic fracture zones into the African continental margin and continent.

Sykes (1967) was the first to demonstrate the nature of the relative motion of Romanche Fracture Zone by using earthquake mechanisms. Bonatti et al. (1970), Bonatti (1971), Bonatti and Honnorez(1971), Bonatti (1973), and Bonatti and Honnorez (1976)made extensive petrological studies of dredge samples from the Romanche and Saint Paul's frac ture zones and Melson et al. (1972) presented detailed petrologic work on the Saint Paul's Rocks.

Whereas the previous studies have either considered only the mid-ocean region, or one of the two adjacent margins, this study considers the entire extent of the equatorial Atlantic ocean floor. Moreover, previous workers have had difficulty in tracing the fracture zones laterally because they considered only one ridge, or one trough or one scarp as representing the fracture zone. However, in this study each fracture zone is shown to be a wide zone (up to approximately 100 km) of fracturing, and the entire zone is traced laterally by seismic-profile crossings. This lateral correlation has permitted the mapping of equatorial fracture zones and the areas in between across the entire equatorial Atlantic from 3ºN to 4ºS latitudes. The mapping of the tectonic fabric of the equatorial-Atlantic ocean floor forms the main basis of this study.

The equatorial Atlantic fracture zones divide the ocean floor into individual crustal segments. Each segment is bounded to the north and south by a fracture zone and extends continuously from Africa to Brazil. These segments vary appreciably in widths according to the variations in the strike of the bounding fractures zones. The fracture zones themselves vary appreciably in width along their strikes. In some cases, the width of the fracture zone is comparable to the widths of the crustal segments it bounds.

The "zone of fracturing" as defined here consists of laterally continuous ridges and troughs. The ridges are pronounced physiographic features in the mid-ocean and have been termed as transverse ridges by

Bonatti (1973). The transverse ridges can be traced into the continental margins through linear alignments of seamounts and buried basement highs which include some oceanic islands.

The present work consists of two parts. The first part deals with the tectonic fabric of the equatorial-Atlantic ocean floor. The physiography, structure, and petrology of the equatorial fracture zones and the significance of earthquake epicenter distribution are described in detail. On the basis of these details a discussion on the tectonics of the fracture zones is presented. Part II concentrates on the geology and tectonic structure of the continental margins and of the bordering continental areas of the Gulf of Guinea and Northern Brazil. The structures of the continental margins on both sides of the equatorial Atlantic are compared with the mid-oceanic structures. The geological evolution of the Mesozoic marginal basins and the structural trends in Africa and Brazil area compared and contrasted. The structures in the continental margins of Brazil and Africa permit a geological and tectonic fit of Brazil and Africa which is more realistic than the bathymetric fits presented by earlier workers.

PART I - THE EQUATORIAL ATLANTIC OCEAN FLOOR

CRITERIA USED IN TRACING FRACTURE ZONES

Oceanic fracture zones are zones of very prominent and generally linear basement features which bound segments of oceanic crust and offset the mid-ocean ridge. A fracture zone in mid-ocean is revealed in a seismic profile by a series of prominent peaks and intervening troughs together with a difference in average basement level across the zone of peaks and troughs.

The basement level is a function of distance from the axis of the mid-ocean ridge. Thus, an offset in the ridge axis results in a difference in basement level across the fracture zone which is a function of the magnitude of the offset and the position of the profile relative to the offset.

Typical cases are illustrated schematically in Fig. 2.In the special case of a profile in the center of an offset (Fig.2) the basement-level difference vanishes. Similarly, at distances large compared to the width of the offset the difference is small. More generally (Fig. 2a and d), the basement-level change is pronounced enough to be an important diagnostic tool. This simple picture is often complicated near the continental margins where differing thickness of sediments across a fracture zone may lead to differential subsidence.

For the most part the fracture zones were traced from profile to profile by first locating the zone of basement-level change associated with a fracture zone and then mapping the topographic features occurring along the trend of change.

ROMANCHE FRACTURE ZONE

Romanche Fracture Zone Between the Offset Portions of the Mid-Atlantic Ridge Axis.

Seismic profile crossings between the axial segments of the Mid-Atlantic Ridge between 2ºN and 0º latitude reveal a wide zone (about 100 km) where the topography is characterized by prominent peaks and troughs (Figs. 3 and 4, profiles 39-45). This zone of alternating peaks and depressions generally separates regions of smoother topography which lie at different depths on either side (Fig. 4, profiles 39, 40 and 43). The wide zone of highly contrasting relief is thought to constitute the Romanche Fracture Zone proper. The lateral correlation of the prominent topographic features seen on seismic profiles indicates a complex of linear ridges and troughs that depicts the lateral extension of Romanche Fracture Zone (Fig. 5). These linear ridges and troughs constitute the transverse ridges and valleys of the Romanche Fracture Zone (Bonatti, 1973).

In the eastern part of the offset region, profiles 39-41 (Figure 4) show the fracture zone symmetrically divided by one prominent transverse valley. On the other hand, in the western part of the offset region (Figure 4, profile 43) two prominent transverse valleys are displayed. The southern trough of profile 43 extends at least as far east as profile 42 (Figure 3) where it is markedly shallower than the transverse valley to the north.

When earthquake epicenters are plotted on the tectonic map of Figure 5 the majority of epicenters lie in the region of the very continuous deeper valley. Sykes (1967) determined fault-plane solutions for an earthquake which plots in this transverse valley of the Romanche Fracture Zone.

The earthquake epicenters do not fall in the other transverse valley situated to the south. A gap in the earthquake epicenters occurs in the western side of the offset region (Fig.5) where the main transverse valley laterally ends against a transverse ridge. The epicenters only appear again close to the western axial segment of the ridge.

The topographic map in Figure 6 shows the morphological complexity of the Romanche Fracture Zone. The main (northern) transverse valley exceeds ddpths of 6500m in two separate regions: in one region a maximum depth of 7856m has been recorded (Vema Deep) (Heezen et al. 1964). The southern transverse valley is approximately 5000m deep throughout and loses its marked topographic expression near 19ºW longitude. The transverse ridges are also very prominent features in relation to the ocean floor on either side of the fracture zone. Their depths are shallower than 2000m and they are clearly delineated by the 3000m isobath (Fig.6).

The axial segments of the Mid-Atlantic Ridge are displaced by approximately 940 km and are intercepted by troughs associated with the fracture zone. It is very significant that with the exception of the region close to the rift-valley, the topographic contours tend to increasingly align in an east-west direction away from the ridge axis (Figs. 5 and 6). The eastern axial segment abuts against the main transverse valley (Figs. 5,6) and in that region the trough shoals considerably. Across from the intersection with the ridge axis, the transver se ridge situated to the north of the transverse valley shoals to a depth of 1000m.

Romanche Fracture Zone Outside of the Offset Portions of the Mid-Atlantic Ridge Axis.

Profile 39 (Figures 3,4) is located within the offset region of the ridge and displays the features of the Romanche Fracture Zone. This profile has been used to trace the Romanche Fracture Zone farther eastward. All the features seen on profile 39 can be compared with profile 37 (Figs. 3 and 4) situated outside the region between the axial segments. However, in profile 37 the depth of the trough region of Romanche Fracture Zone is considerably less than in profile 39. Furthermore, in profile 37 the region separating the Romanche Fracture Zone from the St. Paul's Fracture Zone is considerably narrower than the corresponding region in profile 39. Eastward from profile 37, the Romanche Fracture Zone can be traced by the lateral correlation of geomorphological features such as ridges and troughs that occupy the same relative position with respect to a central trough with flat-lying sediments that marks the abrupt difference in depth of the oceanic basement across the fracture zone (Figs. 3 and 4, profiles 36-27, 25, 24, and 22-17). The abrupt difference in depth of the oceanic basement is followed on profiles 21 through 17, and in these profiles the basement level difference has probably been accentuated by the presence of thick stratified sediments to the north of the transverse ridge which marks the southern boundary of the fracture zone. Profiles 17-14 (Figure 4) show this transverse ridge to be buried by continental-rise sediments. This ridge separates the continental rise of the Ivory Coast and Ghana from the Guinea abyssal plain to the south and lies along the southwestern continuation of the prominent Ivory Coast-Ghana Ridge seen in profiles 13-11, which show the ridge abruptly separating the continental rise from the Guinea abyssal plain. Because of the striking lateral correlation seen on profiles 21 through 14, the southern transverse ridge of the Romanche Fracture Zone is thought to continue into the Ivory Coast-Ghana Ridge (Fig. 7).

The trough partially filled with flat-lying sediments to the north of the prominent southern transverse ridge of Romanche Fracture Zone (fig. 4, profiles 32-30 and 27) may occupy the same relative position as the main transverse valley of the Romanche Fracture Zone within the offset region because it is the deepest trough and has a more or less symmetrical position in relation to the abrupt difference in the levels of the oceanic basement on either side of the trough (Fig.7). There is no clear evidence on the seismic profiles of the presence of a transverse ridge to the north of the above-mentioned trough (with the exception of profile 32) although profiles 31,30, and 27 (Figure 4)show a broad basement high adjoining the trough. Profiles 25,24 and 22-19 also suggest a basement high to the north of the topographically prominent ridge. Consequently, the counterpart of the northern transverse ridge inside the offset region can be tentatively followed with the present data as far east as profile 19 (Figs. 3 and 4).

The bathymetry (Fig. 8) further substantiates the correlations suggested earlier. The main transverse valley of the fracture zone immediately outside of the offset region is interrupted by several sill regions but maintains its continuity towards the east. In the vicinity of

15ºW, the main transverse valley lines up with a prominent trough that is enveloped by the 5000m contour (Figure 8).

The northern transverse ridge extends as a topographically continuous ridge as far east as 13ºW longitude, as seen by the 4500m isobath (Figure 8). After a topographic gap, a broad high area encircled by the 4500m contour is seen immediately to the north of the trough encircled by the 5000m isobath. To the east of 9.5ºW longitude, the continental-rise morphology is dominant and the northern transverse ridge is no longer seen in the topography.

The southern transverse ridge of the fracture zone is not as clearly seen in the topography immediately to the east of the offset region. Several apparently isolated topographic highs border the southern edge of the main transverse valley as far east as 13ºW longitude (Fig. 8). Eastward from 13ºW the ridge is continuous and is delineated by the 4500m isobath; in places it reaches depths as shallow as 3000m (Fig. 8). The ridge as a topographic feature extends to 7ºW longitude.

For the westward continuation of the Romanche Fracture Zone I have used the features seen on seismic profiles (Figs. 3 and 4)located between the offset ridge axis.

Features seen on profiles 42 and 43 can be easily followed westward through profiles 44-48 and 51-55. The prominent ridge located immediately to the north of the main transverse valley of Romanche Fracture Zone can be traced in seismic profiles as far west as 33º where a gap is present in the ridge topography (profiles 57-59). Profile 59 (Figs. 3 and 4) located across this topographic gap shows a basement high covered by a thin cover of continental-rise sediments. This shallow oceanic basement marks an abrupt change in the level of the oceanic basement and is in the same relative position as the northern transverse ridge of the fracture zone. The termination of acoustic reflectors against this basement high (Profile 59) demonstrates that it is part of a continuous ridge that acted as a barrier for the continental-margin sediments. The fact that this shallow oceanic basement lies along the westward continuation of the northern transverse ridge of the Romanche Fracture Zone, and is located along the eastward prolongation of the North Brazilian Ridge which also acted as a sedimentary barrier (Hayes and Ewing, 1970) strongly suggests that the North Brazilian Ridge is linearly continuous with the northern transverse ridge of the Romanche Fracture Zone (Fig. 4, profiles 58-63; Fig. 7). Profiles 60-63 (Fig. 4) show conspicuous ponding of continental-rise sediments against the North Brazilian Ridge. Profiles 60-65 depict the North Brazilian Ridge as a double-peaked feature, also noticed by Hayes and Ewing (1970).

These basement features of the North Brazilian Ridge are very similar to features of the northern transverse ridge of the Romanche Fracture Zone within the region between the ridge axes (compare profile 60 with 41-43, Figure 4). In the topographic map of Figure 8, the northern transverse ridge of Romanche Fracture Zone in the western equatorial Atlantic is depicted by the 4000m contour as far west as 32°W.

In general, the main transverse valley of the fracture zone is followed westward as the region where the level of the oceanic base ment changes appreciably. A trough to the south of the prominent northern transverse ridge is observed in profiles 45-48 and 51-55 (Fig. 4). In the continental rise, the available seismic profiles did not have enough penetration to actually show the trough. Nevertheless,a sediment isopach map of the area (Fig. 9) suggests the presence of a linear area of accumulation as shown by the 2.5,2.0, and 1.5 sec contours that is bounded to the south by a basement high. This area is parallel to and is located to the south of the northern transverse ridge of Romanche Fracture Zone and its counterpart to the west, the North Brazilian Ridge. Consequently, the counterpart of the main transverse valley of the fracture zone in the mid-ocean is suggested as deeply buried under continental-rise sediments in the Brazilian margin (Fig.7).

Prominent transverse ridges on profile 43 are tentatively traced westward as subdued topographic features and buried basement highs (Fig. 4, profiles 45-48, 51-55). The prominent southern trough of the fracture zone seen on profile 43 is also traced westward to profile 55. It is present in Figure 9 where a trough region is bounded to the north by a basement high enveloped by the 2.0 sec contour.

In general, in the manner that the northern transverse ridge of the fracture zone is subdued in the eastern equatorial Atlantic,the southern transverse ridge is subdued topographically in the western equatorial Atlantic. As the mid-ocean ridge is offset left-laterally, the subdued transverse ridges lie on that side of the fracture zone which allows for the greater distance away from the ridge axis.

Petrology

Extensive dredge samples have been obtained from the Romanche Fracture Zone by E. Bonatti and co-workers (Bonatti et al. 1970;Bonatti and Honnorez, 1971; Bonatti et al., 1971; Bonatti, 1971 and Bonatti , 1973).Recently, Bonatti and Honnorez (1976) considered all the dredge results for the Romanche as well as the Vema Fracture Zone.Their data show a percentage distribution of rock types for all the dredge sites in the Romanche Fracture Zone (Fig. 10). A majority of the samples are from the region within the offset ridge axis of the Mid-Atlantic Ridge and dredge sites were mostly concentrated in survey areas where north-south traverses recovered rocks at relatively close locations.

The petrologic types recovered in the Romanche Fracture Zone are extremely varied and consist of serpentinized peridotites, serpentinites, metaserpentinites, basalts, metabasalts, gabbros,metagabbros, mylonites, breccias and sedimentary rocks (Bonatti and Honnorez,1976). Because of the importance of the petrology on the tectonic and geological aspects of a fracture zone, a compilation on the dredge results in the Romanche Fracture Zone will follow. This compilation is essentially based on data presented by Bonatti and Honnorez (1976) and on earlier published information by Bonatti and his co-workers. The dredge locations are shown in Fig. 10, Figure 11 tabulates the frequency of occurrence of various types of rocks in the two walls of the Romanche Fracture Zone.

Serpentinized Peridotites, Serpentinites and Metaserpentinites - Serpentinized peridotites, serpentinites and metaserpentinites were one of the most abundant rock-type recovered from dredges in the area (Fig. 10). Dredges in which these ultrabasic rocks were abundant were raised from depth ranges of 5900-1300m. Serpentinized peridotites, serpentinites and metaserpentinites as secondary constituents of dredge hauls were collected in depths as deep as 7300m and were present in the majority of the recovered dredges. The occurrence of serpentinized peridotites and serpentinites very close to each other and in the same dredge haul suggests a varied degree of serpentinization. In general, serpentinized peridotites, serpentinites and metaserpentinites suggest cataclasis.

Peridotites are serpentinized to varying degrees and consist of: lherzolite, wehrlite, plagioclase-harzburgite and plagioclase-peridotites (Bonatti and Honnorez, 1976). Serpentinized dunites were collected in just onde dredge. Serpentinites were also recovered as breccias, as sedimentary serpentinites and even as serpentinitic conglomerates.

Serpentinites were recovered from all depths in the Romanche Fracture Zone. This fact together with the abrupt lateral variation of petrologic types points out the intrusive character of the serpentinitic rocks (Melson et al., 1972; Bonatti et al., 1974; Bonatti and Honnorez, 1976). As pointed out by Bonatti and Honnorez (1976), a succession of dredges taken from the southern transverse ridge adjacent to the Vema Deep demonstrated a remarkable asymmetry in petrographic constitution (Fig. 11a, location on Fig. 10). The southern transverse ridge was almost entirely constituted by serpentinites whereas the northern transverse ridge had serpentinites as secondary constituents.

Basalts, Metabasalts, Diabases and Metadiabases - Basalt was the second most abundant petrological type recovered from the Romanche Fracture Zone (Figs. 10 and 11). It was generally recovered as the main constituent of the dredges rocks in the 5300-3500m depth range.Basalts were generally a minor constituent in depths greater than 5300m and presented varied structures and textures. They constituted pillow basalts, vesicular basalts and some displayed variolitic texture (Bonatti and Honnorez, 1976). Olivine-basalts were also recovered and some showed several degrees of weathering. Basalts did not show evidences of mylonitization although basaltic breccias of uncertain origin were present.

Metabasalts have similar depth range of occurrence as the basalts although at least in one dredge, dominant metabasalts have been recovered between 5800m and 5000m. The shallowest recovery of predominantly metabasalts ranged in depth from 2500m to 3300m. Metabasalts are generally metamorphosed in the greenschist-facies and often present the same textures as the fresh basalts (Bonatti and Honnorez, 1976).

Diabases were collected in some dredges and appear as dominant constituent together with basalts and hyaloclastites in one dredge(5300-5100m). It is generally not as abundant as basalts and metabasalts. The shallowest depth of recovery of diabases was 3900m and the deepest was 6000m. Olivine-diabases were recognized in one dredge. Metadiabases generally showed a greenschist facies of metamorphism and few of them showed evidences of cataclasis.

Of the total of 42 dredges only 3 recovered alkaline basalts indicating that they constituted only a minor portion of the bulk of the recovered rocks. Dredged alkaline basalts were in two dredges containing mainly metaserpentinites and serpentinites (5200-4300m). The third occurrence of an alkali basalt was present in a dredge that dominantly recovered fresh basalts with no serpentinized peridotites, serpentinites or metaserpentinites (4800-3500m). The metabasalts and metadiabases dredged in the Romanche Fracture Zone do not differ from the greenschist-facies metabasalts and metadiabases recovered from the axial region of the Mid-Atlantic Ridge (Miyashiro and others, 1971).

Gabbros and Metagabbros - Gabbros and metagabbros appear only in four dredges as dominant constituents. They predominate over other rock types (more than 80% of the total weight) only in two dredges. The depth ranges of these two dredges were 5700-3900m and 7300-5900m. Predominantly gabbros and metagabbros were recovered in depths greater than 4000m although, as a minor constituent, they have been recovered from depths as shallow as 3200m.

Gabbros, leuconorites, olivine-gabbronorite, troctolites, ferrogabbros and microgabbros and other gabbroids were numerous in the Romanche Fracture Zone dredges (Bonatti and Honnorez, 1965). Evidence of metasomatic processes were given by the occurrence in several dredges of rodingites, partly rodingitized metagabbroids and prehnitized metagabbros. Brecciation and cataclasis are very frequent in the specimens of gabbroic rock.

The gabbroic rocks are commonly associated with serpentinized peridotites, serpentinites and metaserpentinites. Their occurrence within the offset region of the Mid-Atlantic Ridge in the deepest regions of the fracture zone suggests that they are generally confined to the base of the transverse ridges in association with intrusive bodies of serpentinized ultrabasics.

Alkali-gabbros (nepheline-gabbro) are present in very minor quantities in the Romanche Fracture Zone (Honnorez and Bonatti, 1970). The two dredges that collected alkali-gabbros predominantly consisted of basalt and diabase in one and of metagabbros in the other and both had serpentinized peridotites and serpentinites as minor constituents.

Sedimentary Rocks - Sedimentary rocks were recovered in varying proportions along the strike of Romanche Fracture Zone in the offset region of the Mid-Atlantic Ridge. In general, sedimentary rocks are minor constituents of the dredges. Breccias of uncertain origin that are generally classified as basaltic, serpentinitic, carbonate-altered basaltic, quartz-limestone breccias and metabasaltic breccia with ooze sediments were the most frequent sedimentary rocks recovered. Serpentinized peridotite-carbonate breccias that were recovered from the northern slope of the southern transverse ridge of Romanche Fracture Zone were discussed by Bonatti et al., (1974). The rocks discussed by these authors were collected in the 5700-5340m depth range in the lower portion of a continuous slope to 7000m (Fig.12a, dredge 13). Dredge ≠ 13 consisted entirely of peridotites, serpentinized peridotites and sedimentary serpentinites. Bonatti et al., (1974) gave three possible explanations for the origin of the ultramafic-carbonate breccias: (a) weathering plus subsequent carbonatic cementation; (b) talus breccia plus subsequent carbonatic cementation; and (c) tectonic origin with subsequent carbonatic cementation, which the authors favored.

The occurrence of graywackes in five dredges from the Romanche

Fracture Zone is quite remarkable. Some of the graywackes present distinct layering and a few of them apparently show evidence of fold structures (E. Bonatti, 1976, personal communication). The graywackes generally contain abundant quartz. Four of the dredges in which graywackes have been reported (Bonatti and Honnorez, 1976) range in depths from 6980 to 3500m(Figs. 12a and 12b). The figures show that, with the exception of dredge number 51 (Fig. 12b) all were dredged from steep slopes of the southern and the northern transverse ridges of the Romanche Fracture Zone (see Fig. 10 for location). The fifth dredge was about 110 km to the west of the others and was located in the south-facing slope of the northern transverse ridge of Romanche Fracture Zone.

The abundant quartz and, in some specimens, common biotite (E. Bonatti, 1976, personal communication) suggest terrigenous sialic source for these minerals which seems to be not compatible with the geologic setting of these graywackes. Surrounded by the high topography of the Romanche Fracture Zone and far away from the western and easternmost extensions of the abyssal plains of the eastern and western equatorial Atlantic basins, and being barred from the Sierra Leone Basin to the north by east-west ridges associated with the Saint Paul's Fracture Zone, it is very difficult to associate the topographic setting and the primary sedimentation of these graywackes with the present-day morphology of the ocean floor and recent terrigenous sedimentation in the equatorial Atlantic (Fig. 8).

The primary sedimentary character of the graywackes, the small percentage of the total of the dredged rocks that they form and the wide range of depths from which the graywackes were collected suggest that tectonism post-dating the sedimentation was the main factor in controlling the vertical distribution of the graywackes. The present-day geomorphology of the equatorial Atlantic shows that the abyssal plains on the Brazil and Gulf of Guinea basins extend considerably eastward into the deepest areas which generally coincide with the troughs of the quatorial Atlantic fracture zones. Because of their terrigenous sediments were probably carried down slope by turbidity currents,these tongues of abyssal plain contrast rather remarkably with the hemipelagic character of the sediments lying outside the trough areas.

During its early history, the equatorial Atlantic Ocean was narrower and consequently terrigenous sources were closer to the fracture zones. The sediments that form the dredged graywackes were probably incorporated in turbidity currents that followed the deepest paths available frm the continental margins to the abyssal regions. These deepest paths most probably coincided with the floors of troughs associated with fracture zones. These sediments were probably "reworked"by local turbidity currents that incorporated local sediments and gave the observed complexity to their original mineralogy (Fig. 13). These local turbidity currents are probably very frequent in mid-oceanic regions associated with fracture zones as seen by turbidites recovered from the Vema Deep area (Heezen et al., 1964) and by troughs covered by flat-lying sediments.

If the model of the sedimentation for the formation of graywackes is correct, inversion of relief of features of the Romanche Fracture Zone has occured. Consequently, vertical movements are indeed ve ry important in fracture zones.

Sedimentary serpentinites from Romanche Fracture Zone were reported by Bonatti et al. (1973). They were present in three dredges (≠ 13, 33, and 53, Fig. 12). Dredge 33 has a depth range of 5200m to 4800m. These three dredges show generally ultrabasics as the predominant lithologic type. The fact that these sedimentary serpentinites present layering and graded bedding in some samples suggest that they were deposited by gravity sliding, slumping and turbidity currents, as pointed out by Bonatti et al. (1973). These sedimentary serpentinites are very similar in structure and mineralogy to sandy layers studied in cores that were recovered in the close proximity of the Vema Deep (> 7000m) by Heezen et al. (1964). These sandy layers were generally graded and contained mineral fragments of ultrabasic affinities. The ultrabasic affinities of the sedimentary serpentinites indicate very local source areas and a confined lateral extent.

The fact that sedimentary serpentinites have been recovered from very steep slopes of transverse ridges and in several depth ranges suggest that they probably are not "in situ". Thus, a complex history of erosion (tectonic?), deposition, consolidation and tectonic uplift was probably involved in the formation and "emplacement" of sedimentary serpentinites.

Oolitic limestones have been dredged in the northern transverse ridge of the Romanche Fracture Zone in depth range of 900-1134m (Bonatti et al.,in press; dredge ≠ 8, Fig. 9). Corals and microfossils present in the sample were sufficient to give a good estimate of the possible range in age of the sample (Pliocene). Dissolution and recrys tallization had taken place in some of the samples, and geochemical data support that at least some portions of the ridge had been above sea level in recent past. This assertion is corroborated by very shallow-water corals present in the samples (Bonatti et al., in press). The rate of subsidence calculated from the minimum depth of the dredge (0.2mm/year) and the paleontological data is higher than the one predicted for the subsidence of the oceanic crust, according to a gradative heat loss and consequent increase in density with distance away from the ridge axis (Sclater et al., 1971 ; Bonatti et al., in press). These results point to the fact that there is a decoupling between fracture-zone tectonics and mid-oceanic ridge processes and that vertical tectonism in fracture zones is very important.

SAINT PAUL'S FRACTURE ZONE

Saint Paul's Fracture Zone Between the Offset Portions of the Mid-Atlantic Ridge Axis

The axis of the Mid-Atlantic Ridge between 1º N and 0º latitudes and 25ºW and 30.5ºW longitudes is offset by about 630 km along the St. Paul's Fracture Zone, a zone of complex and highly variable morphology of the ocean floor. This fracture zone contains Saint Paul's Rocks and generally marks a change in the oceanic basement level across the fracture zone. This oceanic basement on each side of the fracture zone generally shows distinct differences in sediment thickness and thus implies different crustal ages (Figs. 3 and 4, profiles 46-48, 51 and 53).

Seismic profiles 46-48 and 51 (Figure 4) reveal a prominent ridge that corresponds to the northern limit of the fracture zone. This northern limit separates the pelagic-sediment covered basement to the north from the zone of complex morphology to the south. Saint Paul's Rocks are located on this ridge. The fracture zone proper is constituted by alternating ridges and troughs that are sometimes laterally traced into troughs and ridges respectively (Fig. 14). The troughs have variable thicknesses of sediment filling (some have no sediments)and their floors are generally flat. The transverse ridges generally contain less sedimentary cover than the associated oceanic basement either to the north or to the south. The boundary of the fracture zone is also generally considered to be the ridge that separates the pelagic-sediment covered basement to the south from the zone of complex topographic features (Figs. 3 and 4, profiles 47 and 53). This boundary is not clear in all the profiles due to the highly fractured state of the oceanic crust in some areas to the south of the fracture zone as seen by alternating ridges and troughs (Fig. 4, profiles 48, 51 and 55, and Fig. 14). Unlike the Romanche Fracture Zone, the Saint Paul's Fracture Zone within the offset region of the ridge axis does not have a main trough and its transverse ridges are not as high as those of the Romanche Fracture Zone; an exception is the pedestal ridge of the St. Paul's Rocks and a seamount at 25ºW longitude (Figs. 15 and 4, profiles 46 and 51). Generally, the 3000 and 3500m bathymetric contours show the transverse ridges and the 4000m contour, the transverse valleys (Fig. 15). The northernmost transverse ridge of Saint Paul's Fracture Zone forms a prominent topographic feature along the entire extent of the offset region of the ridge axis, where as the other topographic features of the fracture zone vary considerably. Profile 48 (Fig. 4) at approximately equal distances from both axial segments of the Mid-Atlantic Ridge and profile 47 that is close to the eastern axial segment reveal prominent ridges and troughs. In contrast, profile 51 shows that, with the exception of the ridge associated with St. Paul's Rocks, the ridges are subdued in relief and do not contrast very much in topography with the features associated with the oceanic basement to the south. In general, the transverse ridges display a step-like configuration towards the shallower oceanic basement to the north (Fig. 4, profiles 51, 53, 54 and 55).

Saint Paul's Fracture Zone has a wide scatter of earthquake epicenters (Fig. 14). These are widespread in the area of the fracture zone and are distributed along the offset region of the ridge. In contrast with Romanche Fracture Zone, the majority of the epicenters are not necessarily located along transverse valleys (Fig. 14). Only in the vicinity of 28ºW (Fig.14) do the epicenters occur along a transverse valley. The presence of one epicenter immediately to the north of the prominent seamount depicted in profiles 46 (Fig. 4) is noteworthy. The ridge associated with Saint Paul's Rocks contains 5 epicenters that plot in an east-west direction (Fig. 14).

Saint Paul's Fracture Zone Outside of the Offset Portions of the Mid-Atlantic Ridge Axis

The topographic features that characterize Saint Paul's Fracture Zone outside the offset region and towards the eastern equatorial Atlantic consist of alternating ridges and troughs. The troughs are generally partially filled with flat-lying transparent sediments (Figs. 3 and 4, profiles 45, 43, 44 and 42). The contrast in depths of the oceanic basement on both sides of the fracture zone is apparent on profiles 45, 43 and 44, where the southern side is shallower than the northern side. Profiles 40 and 39 (Figs. 3 and 4) suggest that the actual zone of fracturing is very wide. The latter profiles show a prominent trough almost completely filled by sediments and with distinct reflectors to the north of the features that have characterized the fracture zone to the west of 21ºW longitude. This trough with flat-lying sediments represents a transition from the fracture zone to a pelagic-sediment-covered oceanic basement to the north.

The very wide zone of fracturing of profiles 40 and 39 is narrowed considerably to the east as seen in profiles 38 and 37 (Figs. 3 and 4). In that region between 18º and 16ºW, the crustal segment between the Romanche and Saint Paul's fracture zones reaches its narrowest width in the entire equatorial Atlantic (Fig. 7).

In contrast with profile 38 (Fig. 4), two prominent troughs separated by an intervening ridge are visible in profiles 37 and 36. Profile 35 shows a wide zone of topographic features that has its northern boundary in a ridge that separates the heavily sedimented area of the Sierra Leone abyssal plain from the fracture zone proper (Grand Cess Ridge: Behrendt et al., 1974). The fracture zone progressively widens eastward and can be traced almost feature by feature in the seismic profiles (Figs. 3 and 4, profiles 35-20).

The Grand Cess Ridge, which corresponds to the northernmost ridge in the latter seismic profiles, is continuous into the upper continental rise of Liberia. An intervening ridge separating two trough regions is also seen in profiles 34, 33 and 31-27, in the same way that a southern ridge is also followed into the Cape Palmas region. Eastward from profile 27 (10ºW longitude) the Grand Cess Ridge has a steep southern slope that extends towards the continental shelf of southern Liberia. The intervening ridge, named the Cape Palmas Ridge by Behrendt et al., (1974) in the continental rise west of Cape Palmas, is followed as a buried feature into the continental slope to the south of Cape Palmas (Fig. 4, profiles 26-21). Eastward from profile 23 (Figs. 3 and 4), the Cape Palmas Ridge is in line with the continental slope of western Ivory Coast (Fig. 4, profiles 23-20, Fig.8).

The southern transverse ridge of Saint Paul's Fracture Zone is present as a topographic feature as far east as 10ºW and continues eastward under the continental rise to the west of Cape Palmas as a buried ridge that was named St. Paul's Ridge by Behrendt et al., (1974). This ridge extends eastward as far as 7ºW (Behrendt et al., 1974).

The topographic features that form the Saint Paul's Fracture Zone outside the offset region of the Mid-Atlantic Ridge axis and towards the western equatorial Atlantic also consist of alternating ridges and troughs that are generally between the two distinct topogra phic levels of the oceanic basement (Fig. 4, profiles 53-55 and 57).The varying topographic levels of adjacent troughs with flat-lying sediments within the fracture zone suggest that the troughs are bounded by continuous ridges (Fig. 7). These transverse ridges have variable depths and morphologies.

Westward from profile 57 (Fig. 3), the available seismic sec tions do not show in detail the oceanic basement under abyssal plain sediments. Nevertheless, the northern boundary of Saint Paul's Fractu-re Zone is visible in all profiles that cross the zone of fracturing because it is a prominent feature that marks the sharp boundary between two distinct levels of the oceanic basement (Figs. 3 and 4, profiles 58, 63, 66 and 67). The numerous transverse valleys of the Saint Paul's Fracture Zone within the offset region of the Mid-Atlantic Ridge axis seem to be represented in the western equatorial Atlantic by only two major trough areas. These troughs generally contain acoustic stratigra phies that differ from each other (profile 63) and are separated by an intervening ridge (Fig. 4; profiles 57, 63, 66 and 67). This interven-ing ridge occasionally form seamounts (profiles 57 and 67) but is gene rally subdued in relief or bulied by sediments (profiles 63 and 66).

The southern boundary of the fracture zone corresponds to a ridge that separates a deeper oceanic basement to the south from the zone of anomalous basement features that characterize the Saint Paul's Fracture Zone to the north (Figs. 3 and 4, profile 63). This ridge is not obvious in the poor seismic sections between 34º and 37º longitude but can be seen in profiles 63, 64 and 67 (Fig. 4). In profile 67 and in all seismic sections westward, the southern transverse ridge of the Saint Paul's Fracture zone is in line with the prominent northern east-west segment of the North Brazilian Ridge (Profiles 68-75). The buried northern transverse ridge of Saint Paul's Fracture Zone seen on profiles 68 and 71 is correlated to a prominent seamount on profile 70. This latter profile also shows distinct acoustical stratigraphies to the north and to the south of the seamount, which suggests that sediment accumulation in the two regions took place independent of each other. This latter conclusion implies that the northernmost seamount of profile 70 belongs to a continous ridge. This ridge is probably continuous under continental rise and abyssal plain sediments into the buried ridges shown in profile 76; and are laterally correlated into the seamount observed on profile 77 (Fig. 4). This latter seamount corresponds to the westernmost extension of the North Brazilian Ridge (Hayes and Ewing 1970).

Profiles 57, 63, 67, 68, 70 and 71 suggest that Saint Paul's Fracture Zone displays a continuous intervening ridge that may be traced as far west as 41ºW longitude. This ridge is a subdued topographic feature in profile 68, a buried basement high separating trough regions in profile 71 and a prominent seamount in profile 70. To the west of profile 70, almost all of the seismic sections show no basement features. However, profile 76 (Fig. 3 and 4) shows a prominent peak that separates areas of distinct acoustic stratigraphies. This peak is situated to the north of the northern east-west segment of the North Brazilian Ridge and may represent the westernmost recognizable extent of the intervening ridge of the fracture zone.

To the west of 46ºW longitude lies the wide continental shelf and slope associated with the region of very high sedimentary rates of the Amazon Cone (Damuth and Kumar, 1975).

Topographically, the Saint Paul's Fracture Zone in the western equatorial Atlantic separates the flanks of the Mid-Atlantic Ridge that are enveloped by the 4250m bathymetric curve from a deeper region to the south between 30º and 38ºW longitude (Fig. 8). This deeper region contains an abyssal plain (Ceará) that is interrupted by east-west ridges. Tongue-like extensions of the Ceará abyssal plain, which are enveloped by the 4400m isobath, extend into the flanks of the Mid-Atlantic Ridge (Fig. 8). These tongue-like extensions of abyssal-plain topography coincide with the transverse valleys of the fracture zone.

To the west of about 38ºW, the several aligned seamounts of the northern east-west segment of the North Brazilian Ridge are the topographic expression of the southern transverse ridge of the fracture zone (Fig. 8). Isolated seamounts encircled by the 4000m and 3000m bathymetric curves aligned in an approximate east-west direction are the only evidences of the northern transverse ridge of Saint Paul's Fracture Zone in the smooth topography of the continental rise and abyssal plain (Fig. 8).

Although in the Brazilian side there was not enough data to continuously trace all the features associated with Saint Paul's Fracture Zone, to the west of 41ºW longitude, I think it is reasonable to suppose that the northern and the intervening transverse ridges are continuous features because of the similarities with the African side where three prominent ridges are correlated continuously from 15ºW eastward, as discussed earlier. The most prominent transverse ridge in the African side is the northern one (Grand Cess Ridge) whereas the most prominent ridge in the Brazilian side is the southern transverse ridge of the fracture zone (southern east-west segment of the North Brazilian Ridge).

In the tectonic map of Figure 7, Saint Paul's Fracture Zone has a wide zone of generally east-west aligned ridges and troughs and is traced across the equatorial Atlantic. In both sides of the ocean, the fracture zone is widest close to the continental margins of Africa and Brazil (140-170 km). The strike of the zone of fracturing is essentially east-west with a pronounced change to an east-northeast trend at about 16ºW longitude towards the African margin. The tectonic map also shows the lateral variation of transverse ridges into troughs and vice-versa, as in the case of the ridge that contains the St. Paul's Rocks.

Petrology

Very few dredges have been recovered from the Saint Paul's Fracture Zone (Fig. 10). Three dredges have been taken from the slopes of the Saint Paul's Rocks pedestal. Two of them, reported by Bonatti et al., (1970 and 1971) at depth ranges of 2800-3000m (dredge nº. 4, Fig. 10) and 900-1100m (dredge nº 3, Fig. 10) recovered serpentinized lherzolites that displayed varying degrees of mylonitization. The third dredge (dredge nº 5, Fig. 10) was taken in the 2950 - 1975m depth range and was part of a series of dredges carried out by the R/V Atlantis II in the vicinity of Saint Paul's Rocks. This dredge was reported by Melson et al., (1967) and recovered a vesicular alkali-basalt that had nepheline in the norm. The alkali-basalt contains abundant small olivine nodules and partly "digested" mylonitized spinel peridotite inclusions (Melson et al., 1967). According to the latter authors, the alkali-basalt represents a volcanic flow that was erupted directly on a floor of spinel peridotite mylonites.

Another five dredges were recovered from the transverse ridges of the Saint Paul's Fracture Zone. Four of them were taken by Neil Opdyke on the R/V Vema (Fig. 10, dredges 6, 7, 8 and 9). Three of these latter dredges were recovered in depth-ranges of 3700-2600m (dredge nº 6), 5140m to shallower depths (dredge nº 8) and 4090m to shallower depths (dredge nº 9), and they all recovered ultrabasic rocks. The ultrabasic rocks in one of the latter dredges (dredge nº 8) had slickensides. The fourth dredge taken by the R/V Vema (dredge nº 7) recovered only two small pieces of fresh basalt in a depth range of 4250-3580m. The fifth dredge (Fig.10, dredge nº 1) was reported by Bonatti and Honnorez (1976) to consist of metadiabases and metabasalts (88% of the total weight), serpentinized peridotites with evidences of cataclasis and sedimentary serpentinites and carbonates (4000-3700m depth range). Dredges 1 and 6 were taken very close to each other and in the same transverse ridge and show ultrabasic rocks at shallower depths than metadiabases.

A great deal of information on petrology of a transverse ridge of a fracture zone comes from the study of the Saint Paul's Rocks.

Saint Paul's Rocks - Also named as Saint Peter and Saint Paul's Rocks or "Rochedos de São Pedro e São Paulo", these tiny islets are isolated in the equatorial Atlantic and reach only 23m above sea level (Fig. 16). The islets constitute the highest peaks of an ENE-WSW elongated submarine mountain that rises 3500m above the ocean floor (Fig. 15).

Saint Paul's Rocks were first visited by the H.M.S.Beagle in 1832 and Charles Darwin was the first to recognize the non-volcanic character of the rocks. The islets were further visited by the H.M.S. Challenger, Meteor, Quest, Owen, Vema, Chain and other oceanographic vessels. Washington (1930) pointed out that Saint Paul's Rocks were mainly constituted by peridotite mylonites and Guimarães (1932) described a nepheline-basalt and tuffs recovered from one of the islets. The rocks are mostly peridotites that present a cataclastic foliation (Tilley, 1947; Washington, 1930).

Melson et al., (1972) made the most comprehensive petrological study to date of the "Rochedos de São Pedro e São Paulo" and classified the rocks of the islets as peridotite-mylonites, brown-hornblende mylonites and clinopyroxene-plagioclase mylonites. The peridotite-mylonites are the most common and present the following average normative mineralogy: 64% olivine, 22% enstatite, 2% diopside, and 8% plagioclase. Mineralogic evidence suggests that mylonitization had taken place during the emplacement of the intrusion of the peridotitic body in temperatures above 500ºC, in the majority of the cases (Wiseman, 1965; Melson et al., 1972). The emplacement of the peridotitic intrusion has taken place in solid state and in a way as fast as to allow the preservation of high P and T mineral assemblages (Melson et al., 1972). The crystallization of primary and secondary minerals of the peridotitic intrusion occurred at different depths (Melson et al., 1972).

The peridotites of the Saint Paul's Rocks studied by Wiseman (1965) and Melson et al., (1972) although cataclastic, do not show evidences of serpentinization. The lack of serpentinization suggested that the cataclasis took place in an anhydrous state. However, dredges 3 and 4 (Fig. 10), as mentioned earlier, recovered mylonitized serpentinized peridotites in the slopes of the pedestal of Saint Paul's Rocks. Consequently, differential serpentinization along the ultrabasic intrusion is implied.

The dredges from the submarine ridge that contains Saint Paul's Rocks showed that indeed the ridge is mainly composed of spinel-peridotite-mylonites and alkaline-ultrabasic-brown-hornblende mylonites, the two major rock types of the islets (Melson et al., 1967). The recovery of an alkali-olivine-basalt (Fig. 10 dredge 5) as mentioned earlier, from the pedestal of the islets (Melson et al., 1967) is also compatible with the nepheline-basalt and volcanic tuffs collected in one of the islets by the Brazilian Navy ship "Belmonte" in 1930 and described by Guimarães (1932). Although a volcanic rock has never been reported from Saint Paul's Rocks, excluding the Guimarães description , still to date not all the islets have been visited or sampled.

The foliation of the ultrabasic mylonites of Saint Paul's Rocks is quite pervasive (Fig. 17). Measurements of this foliation presented by Melson et al., (1972) shows an essentially north-south strike with dips toward the east of 60º and 70º (Fig. 15). The north-south strike of the foliation is surprising because it is expected that the foliation would follow the pervasive east-west topographic trends of the ridges and troughs of the Saint Paul's Fracture Zone. However , because Saint Paul's Rocks and its pedestal ridge mainly corresponds to a solid state ultrabasic intrusion it is possible that a great deal of the mylonitization was caused by its diapiric emplacement that may have involved rotations of the original structures of the serpentinitic-peridotitic body.

The pedestal ridge of Saint Paul's Rocks is in the strike of two troughs (Fig. 14). The trough to the east of the ridge is bare of sediments and thus it may correspond to a very recent feature (Fig. 4, profiles 48 and 51). The youthfulness of the pedestal ridge as a tectonic entity is exemplified by several earthquake epicenters that plot on it and by an earthquake that was felt "in situ" by the crew of the Brazilian Navy ship "Belmonte", when they were on the location to install the automatic lighthouse that now stands there in ruins (Soares, 1968).

In contrast with the youthfulness of the ridge of Saint Paul's Rocks considered here as a tectonic entity, a radiometric age of 4.5 b.y. for the rocks in the islets was reported in a Carnegie Institution Report (Wright, 1965). Rb/Sr ratios also support an old age for the peridotites (Hart, 1964). Melson et al. (1972) determined K-Ar radiometric ages on a hornblende concentrate from a brown-hornblende mylonite. The results pointed out an 835 m.y. age for the rocks that may be real, because the Ar is present in much larger quantities in the hornblende than in the whole rock. In conclusion, Saint Paul's Rocks may represent rocks from a very old and deep-seated part of the upper mantle that were brought up to the surface by a very recent tectonism.

CHAIN FRACTURE ZONE

The work of Heezen et al. (1964) defined a fracture zone to the south of Romanche Fracture Zone named after the R/V Chain in the region between 10º and 20ºW in the equatorial Atlantic.

Chain Fracture Zone Between the Offset Portions of the Mid-Atlantic Ridge Axis

Chain Fracture Zone displaces left-laterally the axial segments of the Mid-Atlantic Ridge by about 300 km. Only two seismic profiles in the offset region of the Mid-Atlantic Ridge axis were available for this work but bathmetric data further delineated the trend of the fracture zone.

Profiles 36 and 37 (Figs. 3 and 4) show a zone of anomalous topography with an intervening trough deeper than 5000m (7 sec two-way travel time). This zone separates two distinct levels of the oceanic basement and also contains northern and southern transverse ridges that are not appreciably higher than the average topographic level of the adjacent oceanic basement. Profile 37 (Fig.4) shows that the fracture zone separates segments of oceanic basement with distinctly different thicknesses of sedimentary cover and that the transverse valley has no appreciable sediments.

The topography of the offset region shows the prominent transverse valley enveloped by the 4500m and 5000m bathymetric contours (Fig. 18). Similar to the main trough of the Romanche Fracture Zone, the trough of the Chain Fracture Zone has a sill as shallow as 4000m, close to the region of its southern intersection with the Mid-Atlantic Ridge axis (Fig. 18). The northern transverse ridge is only broadly delineated by the 3500m contour and, very close to the Ridge axis to the north, by the 3000m isobath. The southern transverse ridge close to the Ridge axis to the south is delineated by the 3000m and 3500m bathymetric contours (Fig. 18).

The tectonic map of Figure 19 shows that most of the earthquake epicenters are located in or very close to the transverse valley of the fracture zone. The epicenters are either restricted to the region between the two axial segments or are located along the rift-valley of the Mid-Atlantic Ridge. However, two of the epicenters are located in the crustal segment between Romanche and Chain fracture zones far away from the Ridge axis. Close to the axial segment to the north of the fracture zone, several epicenters are located close to each other and their distribution encompasses the transverse valley, the northern transverse ridge and the axial region of the Mid-Atlantic Ridge (Fig.19).

The strike of Chain Fracture Zone is east northeast-west southwest and is essentially parallel to the strike of the main trough region of Romanche Fracture Zone, to the north.

Saint Paul's Fracture Zone Outside of the Offset Portions of the Mid-Atlantic Ridge Axis

Seismic profiles are not available between 13ºW and 9ºW but the continuity of Chain Fracture Zone is evidenced by bathymetric data (Fig. 8). Profile 21 (Fig. 4) shows a trough between two prominent peaks that can be correlated laterally to profiles 20 and 17. The two latter profiles show that the Chain Fracture Zone marks distinct basement-level changes and sedimentary covers which characterize different crustal segments. Profiles 20 and 17 (Fig. 4)also shows the fracture zone with a distinct southern transverse ridge and either a subdued (Profile 20) or a buried northern transverse ridge (Profile 17). Because of the increasing proximity of the continental margins, discriminating the basement features associated with the fracture zone becomes difficult. However, the fracture zone is inferred by the presence of basement features that indicate a shallower basement to the south (Profiles 14-12, and 10); or by differences in acoustic stratigraphies across the inferred zone of fracturing (Fig. 4, profile 15). Profiles 9 and 6 off Niger Delta show different basement levels that confined the lowermost reflector of each profile to the crustal segment to the north of a step in the basement. Consequently, Chain Fracture Zone was traced as far east as profile 6, at about 2ºE longitude (Fig.7). The trace of the fracture zone from the mid-ocean to the Gulf of Guinea is in agreement with the traces of Chain Fracture Zone in the eastern equatorial Atlantic presented by Delteil et al. (1974) and by Emery et al.(1975). Eastward from 2ºE the inference of the fracture zone is only conjectural. Along the strike of the hypothetical prolongation of the fracture zone, a linear high area (Okitipupa ridge) separates two sedimentary basins in the continent.

Chain Fracture Zone can be traced westward from the offset region of the Mid-Atlantic Ridge by seismic profiles that are considerably separated from each other. The lateral correlation among these profiles is supported by bathymetric data published by Tomczak and Annutsch (1970) between 20ºW and 24.5ºW longitude that complemented the work of Heezen et al. (1964) between 10º and 20ºW.

Profile 39 (Figs. 3 and 4) depicts a deep trough that separates two basement blocks to the south of Romanche Fracture Zone. Profile 39 also shows an abrupt difference in oceanic basement level to the south of the Chain Fracture Zone proper. This feature apparently is very local and if present in profile 36 to the east is rather subdued; it may correspond to the small offset of the Mid-Atlantic Ridge axial segment to the south of Chain Fracture Zone, as suggested by earthquake epicenters. Based on this offset, Emery et al. (1975) suggested a new fracture zone (the Benue Fracture Zone) that would extend far into the Gulf of Guinea.

Profile 43 shows the Chain Fracture Zone as prominent peaks and troughs separating two different levels of the oceanic basement that are very similar to the features seen on profiles 49 and 50. In these two latter profiles a prominent trough is partially filled with flat-lying sediments and has an acoustic stratigraphy of its own. Westward from profile 50, Chain Fracture Zone is buried by abyssal sediments but its basement features including the site of its intervening trough can be followed in profiles 52, 54 and 55 (Figs. 3 and 4). Westward from profile 55 Chain Fracture Zone is not directly recognized with the available seismic profiles under the thick continental margin sediments although profile 56 may indicate its presence as far west as 32°W.

Topographically, the portion of Chain Fracture Zone outside the offset region in the western equatorial Atlantic is mainly followed by deep elongated regions (5000m to 5500m) which correspond to its transverse valley (Fig. 8). The transverse valley of Chain Fracture Zo ne is flanked by transverse ridges (enveloped by the 4000m and 4500m isobaths), that do not have too much relief contrast in relation to the associated crustal segments (Fig. 8). The topographic expression of Chain Fracture Zone ceases at about 28ºW longitude.

The tectonic map of the equatorial Atlantic (Fig. 7) shows that between 18ºW and 0º longitude, Chain Fracture Zone has an east-northeast-west-south-west strike. To the east of 0º longitude, Chain Fracture Zone is generally northeasterly oriented. To the west of 18ºW longitude, Chain Fracture Zone has an east-west trend.

Chain Fracture Zone changes its relative distance from the Romanche Fracture Zone to the north along its strike (Fig. 7). In the offset region it is 220 km away from Romanche Fracture Zone and at about 30ºW longitude is only 136 km away (Fig. 7).

FERNANDO DE NORONHA - JEAN CHARCOT FRACTURE ZONE

The discovery by French workers of an important lineament to the south of Chain Fracture Zone which consisted of a buried ridge in the continental rise of the Gulf of Guinea led to the speculation that this lineament represented a fracture zone in mid-ocean (Delteil and others,1974). The lineament, the Jean Charcot Ridge,is not only a prominent ridge as seen in profile 6, but also marks a pronounced basement level change. This implies that the ridge indeed corresponds to a fracture zone trend (Figs. 3 and 4, profiles 9, 11-15, and 20). Seismic profiles are not available between 5ºW and 12ºW longitude but profiles 36 and 37 (Figs. 3 and 4) suggest that a prominent fracture zone, to the south of Chain Fracture Zone, may well correspond to the Jean Charcot lineament in mid-ocean. This last assertion is in agreement with the so-called Charcot Fracture Zone shown in Emery et al. (1975).

Fernando de Noronha seamount chain was visualized as a continuous ridge when Lamont-Doherty Geological Observatory made geophysical surveys in the continental margin of northern Brazil. The lineament of Fernando de Noronha together with the North Brazilian Ridge confined the sedimentation of a part of the northeastern Brazilian margin (Gorini and Bryan, 1974). Further studies demonstrated that the guyots, volcanic islands, seamounts and basement highs of Fernando de Noronha Ridge were continuous with a fracture zone to the east of 31ºW, hence named as Fernando de Noronha Fracture Zone (Figs. 3 and 4, profiles 49, 50, 52-54, 56-58, 60 and 62), (Gorini et al., 1974). The fracture zone was mapped as far east as 25ºW and it could not be followed farther eastward because of a lack of closely spaced crossings. Profile 43 shows a topographically complex zone between two markedly distinct basement levels, thus suggesting a fracture zone. This fracture zone is correlated laterally to the probable fracture zone depicted in profile 39 and may also correspond to the fracture zone seen in profiles 36 and 37. Consequently, Fernando de Noronha Fracture Zone is believed to be in the same lineament on the Brazilian side as the Jean Charcot Fracture Zone on the African side both have the same relative positions with respect to Chain Fracture Zone on either side of the Atlantic (Fig. 7). However, as it will be discussed later, in the early rifting history of Brazil and Africa, the two lineaments had distinct geographic locations.

ASCENSION FRACTURE ZONE

The offset region of the Ascension Fracture Zone, named after the Ascension Island, has been described in detail by van Andel et al. (1973). Contrary to other fracture zones studied in this work, Ascension Fracture Zone offsets right-laterally the axes of the Mid-Atlantic Ridge. The total offset is about 230 km and the region between the axis of the Mid-Atlantic Ridge is situated between 11°25'W and 13°25'W longitudes, at about 7°S latitude.

Seismic profiles analyzed by van Andel et al. (1973) show the fracture zone to be formed by broad ridges and troughs (Fig. 20). The broad ridges generally differ from the oceanic basement situated either to the south or to the north by an absent or thin sedimentary cover whereas the troughs contrast with the adjacent regions because they generally contain flat-lying and thicker sediments (Fig. 20, profiles 1-7). Profile 7 (Fig. 20) shows the fracture zone constituted by a trough bounded to the south by a topographically prominent crustal block and to the north by an oceanic basement that does not have high marginal topographic features. The crustal block to the south of the prominent trough (marked with an arrow in profile 7) is separated from the oceanic basement to the south by another trough. The features of the Ascension Fracture Zone displayed in this latter profile can be correlated laterally with profiles 6 through 1 (Fig. 20).

The troughs generally are characterized by magnetic lows and the ridges by magnetic highs. The mapping of these relatively low and high magnetic values by van Andel et al. (1973) also shows the lateral continuity of the prominent trough of the fracture zone seen in the profile 7 and of the other features that were laterally correlated in the seismic sections (Fig. 20).

Topographically, the linearly continuous trough of the Ascension Fracture Zone is enveloped by the 3500m bathymetric curve and exhibits depths as great as 4500m in the region between the axial segments of the Mid-Atlantic Ridge (Fig. 21). To the west, sill depths shallower than 4000m separate the region of the trough that is deeper than 4500m from depths greater than 4000m (Fig.21). The transverse ridge to the south of the main transverse valley of the fracture zone sometimes have high areas encircled by the 2500m, 3000m and 3500m bathymetric contours (Fig. 21).

Earthquake epicenters are confined to the axial regions and to the region between the two axial segments of the Mid-Atlantic Ridge. In the offset region, the epicenters show a preferred distribution along the trend of the trough that separates the southern transverse ridge seen on profile 7 (Fig. 20) from the oceanic basement to the south, as pointed out by van Andel et al. (1973) (Fig. 30). Other epicenters are located along the main trough of the fracture zone but are also scattered within the fracture zone proper (Fig. 22).

Ascension Island is situated 50 km south of the main trough of the fracture zone and 100 km from the median valley of the Mid-Atlantic Ridge(Fig.22)The island is essentially volcanic with diversified petrographic types that include basalts, trachydolerites, andesites (?), trachyandesites, trachytes and quartztrachytes (rhyolites) (Mitchell-Thomẽs,1970). Xenoliths of plutonic rocks that include dunites, peridotites, gabbros, syenites, diorites and alkali-granites are also present in the island (Baker, 1973). The presence of granites as xenoliths has been interpreted as evidence for the nature of the possible basement rock of the island by Daly (1925 in Mitchell-Thomés 1970). However, Tilley (1950, in Baker, 1973) and Roedder and Coombs (1967 , also in Baker, 1973) explained the granitic blocks as subvolcanic equivalents of alkalic rhyolites of the island.

Ascension Island apparently is an isolated seamount with respect to the median valley of the Mid-Atlantic Ridge and with respect to the known features of the Ascension Fracture Zone. However, its close proximity with the wide zone of fracturing and the alkaline nature of the rocks of the island may be indicative of a crustal relationship between the origin of the island and the Ascension Fracture Zone.

No actual mapping of the Ascension Fracture Zone was done that would extend the fracture zone into the African and Brazilian sides of the Atlantic. Emery et al. (1975) considered that the Ascension Fracture Zone extended as far east as the west of the Annobon Island, in the Gulf of Guinea. A few seismic profiles (Fig. 4, profiles 36, 15, 20, 12, 14) indicate a prominent fracture zone to the south of the Jean Charcot Fracture Zone that may extend into the vicinity of and to the west of Fernando Poo-Annobon Ridge. This ridge probably corresponds to the trace of the Ascension Fracture Zone in the continental margin of the Gulf of Guinea. Towards the Brazilian side there is virtually no published information on the probable continuation of the Ascension Fracture, and only a few and widely spaced seismic profiles show that most probably the fracture zone is a prominent feature that extends into the Brazilian margin in the latitude of the city of Recife (Figs. 3 and 4, profiles 39, 43 and 55).

Comments on the Topographic, Tectonic and Petrological Character of the Equatorial Atlantic Fracture Zones

The main trough of Romanche Fracture Zone in the offset region of the Mid-Atlantic Ridge axis is V-shaped and is interrupted by peaks or sill depths in only a few places (Fig.6). If the transform motion is taking place underneath the main trough of Romanche Fracture Zone, or in close proximity to it, it is reasonable to suppose that the trough and the adjacent transverse ridges are the surface expression of the tectonism associated with this relative movement.

Horizontal motions in fracture zones in mid-ocean are associated with the differential relative motion of the crustal segments bounded by a fracture zone. A well-known characteristic of strike-slip faults is to display, along their strike, features that are compressional in nature, and features that characteristically present normal fault pattern, with the development of horst and grabens (Allen, 1965). Fracture zones have very linear topographic and basement features across the entire equatorial Atlantic. Within the offset portions of the Mid-Atlantic Ridge axis, transverse ridges border transverse valleys for hundreds of kilometers. This topographic character can be compared to a horst and graben tectonic setting. Such long and linear tectonic features are difficult to explain purely on the basis of tectonism associated with the strike-slip (transform motion) character of the fault.

Dredge results suggest that diapirism of low-density ultramafic rocks is also a very important process within fracture zones (Bonatti, 1973; Bonatti and Honnorez, 1976; Melson and Thompson, 1972). In Romanche Fracture Zone, as discussed previously, the serpentinized peridotites and serpentinites and their metamorphic equivalents make up a large portion of the rocks recovered in the majority of dredge hauls. These types of rocks have been dredged from all depths on walls of transverse ridges. The intrusive character of these rocks is inferred from the fact that they were originally constituted by high P and T minerals and occur in association with fresh basalts, basaltic volcanic glass and other rocks. The widespread occurrence of these rocks in a horizontal and vertical sense, points out that they penetrated everywhere in the zone of fracturing. The mylonitic character of these ultramafics is quite pervasive in the great majority of the dredged samples. The Saint Paul's Rocks evidence, showing that the mylonitic foliation does not strike in accordance with the direction of the expected relative motion of the crustal segments bounded by Saint Paul's Fracture Zóne, can be explained by a diapiric emplacement.

A diapiric body occupies a volume that is made at expenses of the space occupied by the other rock types that it intrudes and displaces; a space that was created by compressive stresses, and/or a space that was formed by extensional forces. The diapiric body has lower density than the surrounding rocks and therefore tends to flow upward into physically favorable places. Sites of concentrated intrusions of relatively low-density diapiric masses are favored by tectonic conditions such as lateral compression or tension. Vertical migration of a diapiric body is accomplished by brecciation, mylonitization, faulting, folding and upwarping. The prolonged action of a diapiric mass can tectonically detach a section of intruded rocks especially if lateral weakness zones are already present in the vicinity. The vertical displacement of the individualized section will be directly related to the intensity of the diapiric flow, to the tectonism that is favoring diapirism, or to a combination of the two processes. Also, perforation with consequent brecciation and domal uplift may lead to gravitational gliding of the uplifted rocks, tending to cause the diapiric mass to outcrop. In conclusion, it is quite possible to have outcrops of relatively low-density ultramafic diapiric masses in the ocean by the disruptive action of the intrusion ; by a consequent lateral extension of the host rocks due to the increasing volume of the intrusive and consequent filling of fractures; by the lateral gliding of overlying rocks; and/or by the subsequent erosion of the cover rocks of the intrusive mass, that may even be accomplished by marine erosion at the sea level.

A model for the tectonics and the origin of the morphology of fracture zones that explains the seemingly abundant diapirism of ultramafics of relatively low density, and the existence of such linear features as the transverse ridges and valleys, is difficult to make based solely on the overall available geological information. A major difficulty for example is to reconcile the fact that abundant serpentinized peridotites and peridotites have penetrated the rocks of the transverse ridges and apparently were not able to modify the V-shaped topography of the prominent transverse valley of the Romanche Fracture Zone. This last observation is hard to explain if we assume that a deep fault zone underneath the trough region of Romanche Fracture Zone is the probable major conduit of diapiric masses (Fig. 23). Instead of being abundant in the region of the transverse ridges, why did the diapiric intrusions not invert the V-shaped profile of the trough region, the least resistant surface that would correspond to the ocean floor in the vicinity of the bottom of the trough? In some cases though, diapirism of serpentinized peridotites and peridotites have penetrated in trough regions. One such example is the Saint Paul's Rocks in the Saint Paul's Fracture zone where the ridge pedestal of the islets lies longitudinally between two trough regions and earthquake epicenters occur along this ridge (Fig. 14); the main trough of Romanche Fracture Zone has some topographic peaks between the two deepest areas (Fig. 6); and E. Bonatti (personal comm., 1976) dredged serpentinized peridotites and serpentinites in one peak in the trough region. These examples suggest that diapiric masses lie in the vicinity of the trough regions and are at very shallow crustal depths.

Because transverse ridges are very linear features and abundantly intruded by serpentinitic rocks, it is probable that diapirism of relatively low-density ultramafics is an important tectonic factor for the origin and maintenance of the high topography of the transverse ridges (Figs. 23 and 24). The reason why diapirism apparently did not invert the V-shaped relief of the trough along the seismically active trough region of the Romanche Fracture Zone, is not understood on the basis of available data.

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The petrological complexities associated with diapiric intrusion into transverse ridges can be visualized in a schematic block diagram of a transverse ridge of Romanche Fracture Zone (Fig.25).

The complex topographic, tectonic and petrologic nature of fracture zones is further complicated when the direction of the main transform motion changes. The uppermost surface of the deep fault that has accomodated the transform motion associated with the old trend (site of the main trough of the fracture zone) can be easy conduit for diapiric masses and/or eventual magmatism that may intrude the fault surface (Fig. 26). After encountering the "free surface" of the ocean floor the intrusive diapir can rise several thousand meters above the sea floor to near or above the sea surface (e.g. the southern transverse ridge of Vema Fracture Zone (?)). The sediments that once were in the deepest parts of the trough may be transported upward by the diapiric intrusion (quartziferous limestones dredged on top of the southern transverse ridge of the Vema Fracture Zone and described by Honnorez et al., 1975) and may be involved in complex deformation due to the lateral compression caused by the vertical motion of the diapir (deformed graywackes of Romanche Fracture Zone dredged by Bonatti and his co-workers). The former transverse ridges would probably also be intensely modified topographically, especially if the angle of change in the transform direction is great enough to affect a large area of the ocean floor (such as in the case of Romanche Fracture Zone). The former sites of transverse ridges could become sites of troughs of the fracture zones associated with the new trends of transform motion and vice-versa (Fig. 26). Added petrological complexity will also be a probable consequence.

Serpentinitic intrusions are widespread in the equatorial Atlantic fracture zones as shown earlier in this work. The petrological and geochemical nature of these serpentinites suggests that these serpentinites were probably derived from an ultrabasic source very different from the one that had given rise to the tholeiites, characteristic of the spreading axes. The fact that the ridge axial segments end abruptly

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in fracture zones that offset these segments is further evidence that the level of the anomalous mantle (velocity-wise) associated with the ridge axis and the level of origin of diapiric ultramafic masses either differ considerably, or that the physical circumstances of a sheared zone is very different from a tensional zone (ridge axis) as far as magma generation is concerned.

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The entire zone of fracturing associated with the transform motion can then be understood as enveloping a longitudinal narrow zone in which the main transform motion has been taking place, and a northern and southern elongated areas in which vertical forces are accommodated as a result of diapirism of relatively low-density ultramafics, and as a result of dip-slip components of the main horizontal motion.

Outside of the offset region of the ridge axis the topographic contrast between the two transverse ridge blocks tends to increase sharply, immediately away from the offset region, indicating that the transverse ridges are coupled at least partially with the subsident oceanic crust away from the offset region.

Нужно проверить, насколько закономерно нарастание контраста глубин по обеим сторонам разломов с удалением от «офсета»

TECTONIC CHARACTER OF FRACTURE ZONES OUTSIDE THE OFFSET REGION OF THE MID-ATLANTIC RIDGE AS EXEMPLIFIED BY ROMANCHE FRACTURE ZONE

There are no earthquake epicenters located in fracture zones far away from the offset region of the Mid-Atlantic Ridge. The general tendency outside of the offset region is that shallower crustal segments contain shallower transverse ridges and deeper crustal segments contain deeper transverse ridges.

Because at any specific north-south traverse the crustal segments are generally at different distances from their correspondent spreading axes, the rate of subsidence is different for the two adjacent crustal segments. This difference in rates of subsidence is greatest close to one of the ridge-axis segments and gradually decreases laterally away from the ridge axis so that at great distances from the ridge axis subsidence rates are approximately equal (Sclater et al., 1971). Consequently, the topographic features associated with a fracture zone away from the offset region are submitted to differential vertical motions that tend to die away with distance from the mid-ocean region. The deep fault-zone that was the main site of transform motion in the offset region will be the most probable place in which the differential vertical motion will take place. The maintenance of the very deep fault-zone by differential subsidence of the adjacent crustal segments has very important geological consequences. The deep fault-zone can thus become possible natural conduits for vertical migration of magmas; the most probable surface to accomodate further differential subsidence between the two adjacent crustal segments; and the most probable surface to yield to additional tectonic forces.

By analyzing the depths of summits of transverse ridges along the strike of Romanche Fracture Zone, the model for fracture zones as put forward earlier is clearly oversimplified. One of the transverse ridges of Romanche Fracture Zone in the continental margins on both sides of the Atlantic, where its depths should have been the greatest, form very prominent marginal ridges.

The arguments discussed previously can be combined to present an attempt to understand the geological role of a fracture zone outside the offset region of the Mid-Atlantic Ridge. The main trough of the fracture zone constitutes the deepest area of the adjacent crustal segment, and consequently is generally the first to receive sediment influx from local sources and from far-away terrigenous sources. Because one of the crustal segments is originally deeper than the other in any traverse close to one segment of the ridge axis, the deeper crustal segment receives terrigenous sedimentation whereas the adjacent crustal segment does not receive sediments. Subsidence due to sediment loading tends to increase the difference in level between the crustal segments bounded by a fracture zone. The differential subsidence causes further vertical ajustments that probably take place in the deep fault that is the zone of contact between the adjacent crustal segments (Fig.27). Consequently, the deep fault zone can be constantly active and, sometimes, pockets of magma situated underneath the lithosphere can have a vertical conduit to the surface, or magma may be generated by further tectonism within the deep fault zone. Although the surface expression of magmatism may not necessarily be located in the main trough, it will certainly occur within the limits of the fracture zone. Magmatism of such origin may have formed the Fernando de Noronha Islands, and the islands of the Gulf of Guinea.

Evidence for possible horizontal motions in fracture zones outside of the offset region of the Mid-Atlantic Ridge is only indirect and come from theoretical reasoning. If the divergent motion that occurs at the ridge axis has a constant rate throughout geological time and is equal in magnitude for both crustal segments that are separated by a fracture zone, no relative motion is expected in the region outside the offset zone (Fig. 28a). If the magnitude of the divergent motion (constant rate) in both adjacent crustal segments is different, a differential relative motion outside of the offset region is expected (Fig.28b). Note that the associated transcurrent motion is in the opposite sense of the transform motion in one side but is in the same sense of the transform motion on the other side (Fig.28b).

Spreading rates change considerably with geological time (Pitman and Talwani 1972; Hayes and Pitman, 1973). Consequently, the divergent motion is generally not at constant rate for long periods of time. These variable spreading rates and the way that the crustal segments laterally absorb these changes are of paramount importance in deducing the state of stress that the correspondent crustal segment is subjected to.

Comments About the Origin of Marginal Ridges

The marginal ridges of the continental margins of the equatorial Atlantic are characterized by tall seamounts which rise to near sea level and in some instances outcrop as oceanic islands. The seamounts are generally shallower than 3000m and some are flat-topped indicating truncation by wave erosion. Most of the seamounts have their pedestals on elongated basement ridges that are oriented along the strikes of the marginal ridges. The oceanic islands that are in the lineament of the marginal ridges are composed largely of alkaline rocks and commonly their rocks contain peridotite nodules that were brought up from large depths (> 60 km). These islands are generally of Recent age.

Because of the high topography of the marginal ridges and their association with volcanic islands of Neogene or younger age, it is feasible that marginal ridges could be more easily explained by being formed only by Recent volcanism associated with intraplate tectonism. However, this does not seem to be the case because sedimentary barriers which individualized continental margin sectors have been present in the continental margins since the beginning of their sedimentary history as suggested by Fig. 29. The Neogene alkaline volcanism of islands associated with the marginal ridges is consequently better explained by magmatic and/or tectonic reactivation of the lineament associated with the marginal ridges. Because the marginal ridges are traced into fracture zones, we must assume that tectonic and/or magmatic reactivation of portions of fracture zones far from the offset region of the ridge axis gives rise to the oceanic islands and, at least many of the tallest seamounts of the marginal ridges.

Tectonic and/or magmatic reactivation of fracture zones in the continental margins may result from the confinement of a sedimentary prism to a crustal segment bounded by fracture zones (a continental margin sector). The sediment load, as a consequence of the confined sedimentation, may cause the subsidence of the oceanic crust and thus accentuate the difference in levels of adjacent segments of oceanic crust.

We can therefore speculate about the origin of the foundations of the marginal ridges essentially based on the model proposed in this work for the geological processes associated with fracture zones. It is possible that because of the differential subsidence, diapirism of relatively low-density ultramafics was triggered by the more pronounced subsidence of one of the adjacent crustal segments. The diapirism was preferentially concentrated on the more passive crustal segment thus uplifting the blocks of the former transverse ridges. Because the diapirism involves displacement of mass in the upper mantle it is also feasible that alkaline magma may have originated as a consequence of thermodynamic disequilibrium. This magma eventually reached the ocean floor surface and built up volcanic seamounts and islands, probably in the same way that alkaline magmas may have formed in fracture zones within the offset region of the mid-oceanic ridge. This model can be easily tested by intensive dredgings in the marginal ridges of the equatorial Atlantic. A relatively low-density ultramafic fondation of marginal ridges is in accordance with the suggestion of Bonatti and Honnorez (1976) and with gravity data (Cochran, 1973). The serpentinitic and/or serpentinized peridotites and/or peridotites base for the marginal ridges proposed in the aforementioned model contrasts markedly with Le Pichon and Hayes'(1971) model which implies a petrology for the marginal ridges strongly influenced by derivatives of continental crust. The other fundamental contrast between the proposed model presented here and Le Pichon and Hayes'(1971) is that the marginal ridges are prominent topographic features at present because they have resulted from tectonic reactivation of fracture zones and not necessarily because they were formed as prominent features at the initial separation of Africa from Brazil. The verification of the onset of the tectonic and/or magmatic reactivation of the fracture zones in a manner that would lead to the development of the prominent marginal ridges observed today is not possible using the available seismic profiles because of insufficient resolution.

Consequences of Changes in Transform Fault Directions

The tectonic map of the equatorial Atlantic shows that the Romanche Fracture Zone has a pronounced change in strike in the offset region of the Mid-Atlantic Ridge (Fig. 7). This change in strike of the Romanche Fracture Zone is supported by: (a) the morphology of the zone of fracturing that has a prominent but probably fossil trough; (b) the earthquake epicenters that are located only in the northern trough; and (c) the asymmetry in strike of the western and eastern portions of the fracture zone in the offset region of the Mid-Atlantic Ridge. Because the trend of the portion of the Saint Paul's Fracture Zone to the north of the area of the offset region of Romanche Fracture Zone does not change, the crustal segment between the two fracture zones is remarkably narrowed and the two fracture zones almost join each other (Fig. 7). Thus, it is evident that a great number of the features associated with Romanche Fracture Zone was formed at the expense of the crustal segment between the Romanche and Saint Paul's Fracture Zones.

The narrowing of the crustal segment can be explained either by crustal consumption or possibly by the fact that the southern adjacent crustal segment was formed partly at expenses of the crustal segment to the north. Crustal consumption would imply that part of the crustal segment between the Romanche and Saint Paul's fracture zones was probably subducted following compressive stresses associated with the change in transform direction of the Romanche Fracture Zone. Crustal consumption would probably cause extensive deformation in the sedimentary cover of the oceanic basement in the immediate vicinity. However, to the south of Saint Paul's Fracture Zone, the observation of seismic profiles does not lead to tectonic implications of deformational structures due to the general lack of unconsolidated sediments with the exception of profile 33 (Fig. 4) that shows dips at angles not seen in all other seismic profiles. However, seismic profiles reveal that the sedimentary cover does not change its thickness regularly with distance from the associated Mid-Atlantic Ridge axis. There are places that thicker sediments lie to the west of thinner sedimentary cover. This implies tectonism for the removal of part or all of the sedimentary cover in some places. The removal of the sediments is further substantiated by the common presence of flat-lying troughs probably sedimented by turbidites.

It is also possible to explain the narrowing of the crustal segment between Romanche and Saint Paul's fracture zones by assuming that the adjacent crustal segment between Romanche and Chain Fracture Zones is partly formed at the expense of the crustal segment to the north. The segment between Romanche and Chain fracture zones is widest at its present position between the offset region of the Mid-Atlantic Ridge (Fig. 7). The bathymetric map of the equatorial Atlantic shows that the Mid-Atlantic Ridge axis of the crustal segment between Romanche and Chain fracture zones is the longest and that abutts against the main trough region of Romanche Fracture Zone most probably causing its shoaling (Fig.8). The orthogonality of the trends of the features associated with the Romanche Fracture Zone and the Ridge axis to the south substantiates that most probably the Mid-Atlantic Ridge axial region used up some area of the adjacent crustal segment to the north. Because the Romanche Fracture Zone changes its strike to a more east-northeast direction within the offset region it is also reasonable to conclude that at least part of the topographic features of the fracture zone were created at the expense of part of the surface area of the crustal segment between Saint Paul's and Romanche fracture zones (Fig. 7). Consequently, the wide crustal segment between Chain and Romanche fracture zones can also be explained as originating from northward extension of the sea-floor spreading axis, into the already formed crustal segment to the north, following a change in the transform fault direction. The two fracture zones (Romanche and Chain) then bounded a wider crustal segment that would be constituted by partly inherited, partly tectonized and newly developed structures in the oceanic crust.

Which of the two hypotheses to explain the narrowing of the crustal segment (crustal consumption or tectonic reactivation) is correct, is difficult to assess with the available data but I think that the simplest explanation is that the narrowing is apparent and is caused by an anti-clockwise migration of the main trough of Romanche Fracture Zone with an associated northward migration of the spreading axis to the south, in the manner depicted by Figure 26.

Tectonism and tectonic reactivation of a portion of a crustal segment can be a consequence of changes in transform fault directions that most probably are caused by accomodations to a new direction of relative motion between the associated plates. The region of the ocean floor that will be affected by tectonic reactivation will be dependent on the proximity of the offset region and to the amplitude of change of the transform fault direction. The immediate implication of the tectonic reactivation of crustal portions in the ocean floor is the possibility that this may have happened in the continental margins during earlier stages of development of the Atlantic ocean floor (when the continents were still close together). At those times, complex structures such as folds, normal faulting, reverse faulting (relatively low-density ultrabasic diapirism plus magmatism), and may be low-grade metamorphism of the sedimentary cover may have taken place following adjustments to new transform fault directions.

Asymmetry in the Equatorial Atlantic

Le Pichon and Hayes (1971) reconstructed the flow lines that theoretically represented the total relative motion between South America and Africa up to their present positions from Bullard et al's fit (1965) that is given by a 57.0º angular rotation of Africa with respect to South America of points of known counterparts on both continents around a pole of rotation (44ºN, 30.6ºW). Le Pichon and Hayes (1971) considered the available information on the structures of the mid-oceanic region and of the continental margins of the equatorial Atlantic and originated essentially two poles of rotation from which the flow lines of Africa and South America could be calculated. The earliest pole of rotation identified by Le Pichon and Hayes was determined essentially from the trends of the prominent marginal ridges that are roughly symmetrical on both sides of the equatorial Atlantic. The symmetry is apparent because of these prominent ridges on the continental margins of both continents that pass oceanward through a buried and topographically subdued region. From other topographic trends in the continental margins and in the mid-oceanic region of the equatorial Atlantic,as defined by the earlier works of Heezen et al. (1964), Fail and others , (1970), Arens and others (1971) and Hayes and Ewing (1970), Le Pichon and Hayes (1971) calculated the location of the later pole of rotation and thus the flow lines of the two continents for the time after the marginal ridges were already formed to the present (Fig. 30).

Intrinsic to the method used by Le Pichon and Hayes (1971)was the assumption of a net symmetric sea-floor spreading situation for the equatorial Atlantic because the angular rotations, starting from points of known counterparts in the Bullard et al's fit (1965)of South America and Africa, were always the same for both continents. Although it is a simplified representation of the relative motion between Africa and South America, the total angular rotations of both continents with respect to the early pole and to the later average pole of rotation of Le Pichon and Hayes (1971) provide us with a way to test the symmetry of the axial segments of the Mid-Atlantic Ridge. Instead of reconstructing the flow lines of Africa and Brazil from points of known counterparts in the Bullard et al's fit, points in the axial regions of the Mid-Atlantic Ridge were chosen and were rotated with similar angles using the average and early poles of rotation of Le Pichon and

Hayes (1971). These rotations originated flow lines that should have matched the flow lines originated by the method of Le Pichon and Hayes (1971) if a net symmetrical spreading is assumed.This evidently was not the case (Fig. 30).The remarkable asymmetry of the position of the axial segments of the Mid-Atlantic Ridge, suspected on the firts obser vation of the tectonic map of the equatorial Atlantic ocean floor (Fig. 8), is now confirmed.The theoretical trace of the Romanche Fracture Zo ne delineated from a point on the ridge axis rotated towards the conti nents, shows a rotational asymmetry of 5º (≥ 520 km) to the west in re lation to the average pole of rotation of Le Pichon and Hayes (1971) (Fig. 30). Besides the visible asymmetry, the theoretical flow line reconstructed from a point of intersection of the ridge axis with Romanche Fracture Zone differs more markedly from the actual trace of the fracture zone (Fig. 30).

The asymmetry of the Mid-Atlantic Ridge axial segment to the north of Saint Paul's Fracture Zone is also large (≥ 435 km) and the theoretical flow line seen on Figure 30 is for the most part in disagreement with the actual trend of the Saint Paul's Fracture Zone.

Together with the asymmetry of the position of the Mid-Atlantic Ridge axis and may be related to its origin is the fact that the present-day offset of the axial segments of the Mid-Atlantic Ridge associated with Romanche Fracture Zone (940 km) is 2.5 times greater than the supposedly original offset (370 km) when Africa and Brazil are considered together. The offset associated with Saint Paul's Fracture Zone (630 km) is about 2 times greater than the original offset (330 km).

The marked asymmetry of the Mid-Atlantic Ridge axis as evidenced by the flow lines of Saint Paul's and Romanche fracture zones implies that the asymmetry of the ridge axis can be explained either by a westward jump or jumps of the axial segments, or by asymmetric sea-floor spreading with faster rates in the African plate. The other implication is that Le Pichon and Hayes'(1971) reconstruction of flow lines in the equatorial Atlantic can be misleading for age-interpretations of crustal portions.

Around 25ºW longitude, and to the north of 4ºN latitude there is an elevated basement area that Emery and others (1975) considered to be an integral part of the Sierra Leone Rise (Fig. 30). This area of high oceanic basement is the only evidence that suggests a possible former spreading center that would have jumped to the present-day axial region. However,because this region seems to be the counterpart of the Ceará Rise in the Brazilian margin and they both occupy a symmetrical position in relation to Africa and Brazil they may have originated by abnormally high volcanism (Kumar and Embley, 1977). Because fracture zones are so continuous across the equatorial Atlantic, if ridge jumps actually have occurred the original fracture zones controlled the lenght of the spreading axes and these changed their positions within the boundaries (fracture zones) of the respective crustal segments.

If asymmetric spreading caused the asymmetry, it is implied that faster spreading rates had been associated with the African plate throughout the evolution of the equatorial Atlantic. Because no magnetic anomalies are mapped consistently from the Mid-Atlantic Ridge axis in the equatorial Atlantic, the hypothesis of an asymmetric sea-floor spreading remains untested. Although asymmetric spreading has been reported in the mid-oceanic ridge system south of Australia by Weissel and Hayes (1971), and Hayes (1976), magnetic anomaly maps of the North Atlantic (Pitman and Talwani,1972) and the South Atlantic (Larson and Ladd, 1973; Ladd, 1974 and Pitman et al., 1974) establish a reasonably symmetric pattern of the magnetic anomalies on both sides of the Mid-Atlantic Ridge axis. Areas that have been studied in more detail and that were explained as caused by processes of asymmetric apreading are the Reykjanes Ridge (Talwani et al., 1971) and the Jan Mayen Ridge (Vogt et al., 1972). Other evidence that suggests the possibility that asymmetric spreading has occurred in the equatorial Atlantic is the fact that there is an asymmetry in lengths of the marginal ridges on both sides of the Atlantic (Le Pichon and Hayes, 1971). However, this asymmetry may be only apparent because Le Pichon and Hayes considered the end of the marginal ridges where they become buried by sediments.Because they are continuous features, different sedimentary processes on both continental margins may have caused the asymmetrical burial of the marginal ridges.

PART II - THE CONTINENTAL MARGINS OF GULF OF GUINEA (AFRICA) AND OF NORTHERN BRAZIL

INTRODUCTION

The continental margins of northeastern Brazil and the Gulf of Guinea area of Africa are divided in several sectors by prominent marginal ridges. These ridges either outcrop as volcanic islands and linear chains of seamounts or constitute buried basement highs. The sectors are the following: the Liberia sector in Africa corresponding to the Amapá sector on the Brazilian side; the African Ivory-Coast sector which is the counterpart of the Pará-Maranhão sector in Brazil; the Ghana-Togo-Dahomey-Nigeria sector which is the counterpart of the Ceará sector in Brazil; and the Fernando Poo-Cameroon sector which corresponds to the Rio Grande do Norte-Pernambuco sector in Brazil (Fig. 3). These sectors contain sedimentary basins that evolved independently from each other. Nevertheless, they have a similar tectonic framework and a comparable sedimentary history (Gorini and Bryan, 1974). Each continental margin sector will be discussed in detail with respect to its physiography and tectonic framework and to the geology of the bordering cratonic areas. The objective of this detailed discussion is to present comprehensive information on the geology of the African and Brazilian margins in order to provide a background for comparing and contrasting the two margins. Such a background is necessary in order to understand the influence of early-rifting structures on the origin and evolution of the tectonic fabric of the equatorial Atlantic floor.

CONTINENTAL MARGIN OF GULF OF GUINEA

Liberia Continental Margin Sector

Physiography

The continental margin off Liberia has a smooth continental rise that grades westward into the Sierra Leone abyssal plain. The continental rise towards the south is wider than in the northern part of the sector and it has an area of low-gradient named the Liberia Plateau (Behrendt et al., 1974). The continental rise in the vicinity of the Liberia Plateau gives way to a pronounced slope to the south (Fig.31). This slope has an ENE-WSW strike and is parallel to a series of buried basement highs and/or seamounts that constitute the prominent Grand Cess Ridge (Fig. 31; Behrendt et al., 1974). Sediment ponding behind the Grand Cess Ridge is commom and the evolution of the continental margin in the southern portion of the Liberia sector has been controlled by the damming of sediment behind the ridge (Schlee et al., 1974; Fig. 4, Profiles 23-35). Consequently, the Grand Cess Ridge is considered to be the southern boundary of the Liberia sector. The northern part of the sector will not be considered in this work.

The Grand Cess Ridge is traced westward into the northern transverse ridge to the Saint Paul's Fracture Zone as discussed before. The Grand Cess Ridge continues as a physiographic boundary as far west as 19ºW, separating the lower continental rise off Liberia and the Sierra Leone abyssal plain to the north from the flanks of the Mid-Atlantic Ridge to the south (Fig. 8).

General Geology and Tectonic Framework

Shield rocks predominate throughout Liberia and their radiometric ages fall in three radiometric provinces (Hurley et al., 1971 ; and Hurley and Rand, 1973). The Liberian-age rocks (ca. 2700 m.y.) are present in northwestern Liberia, the Eburnean-age rocks (ca.2000 m.y.) in southeast Liberia and the Pan-African age province (ca. 550 m.y.)is restricted to a narrow outcrop belt in the western littoral zone of Liberia (Fig. 32). A shear zone dipping 50 to 70°SW recognized by Thorman (1972) and a steep gravity gradient recognized by Behrendt and Wotorson (1970) mark the boundary between the Liberian and the Pan-African age provinces. The Pan-African-age rocks mainly belong to a mafic granulite facies whereas the Liberian and Eburnean-age rocks are mainly amphibolite-grade granitic gneiss and metasediments (Behrendt et al., 1974). Sedimentary rocks of Paleozoic and Mesozoic age are present close to the littoral zone in sedimentary embayments. Tholeiitic diabase dikes are common in northwestern Liberia (Dalrymple et al., 1975) (Fig. 32). These dikes have a northwest-southeast trend that parallels the coastline. K/Ar radiometric ages show variable ages that range from 186 to 1,213 m.y. for dikes intruding Precambrian rocks . Diabase dikes intruding Paleozoic sedimentary rocks near the coastline all fall within the range of 173 to 192 m.y. (Dalrymple et al., 1975). Despite the apparent age difference between the dikes that intrude the Precambrian rocks and those that cut the sedimentary Paleozoic rocks , on the basis of paleomagnetic directions Dalrymple et al. (1975) suggests that both set of dykes may have been contemporaneous. The observed age differences are thought to be caused by an excess of ^{40}Ar in dikes that cut Precambrian rocks.

The tholeiitic diabase dikes probably constituted the feeders for amygdaloidal basalt flows that unconformably overlie a Paleozoic(?) sandstone (Schlee et al., 1974). Aptian-Cenomanian sedimentary rocks present in the sedimentary embayments contain, among others,pebbles of diabase suggesting that their deposition post-dated the diabase intrusioning (Schlee et al., 1974). As revealed by drilling, the relationship of the intrusives with the sedimentary succession is very similar in both the offshore and onshore sedimentary basins. Diabase dikes and sills intrude Paleozoic rocks that are unconformably overlain by basalt flows. The extrusive and intrusive rocks are then overlain by sedimentary rocks of Cretaceous age (Schlee et al., 1974).

The continental shelf off Liberia contains three main sedimentary embayments which are elongated in a direction subparallel to the coastline. They contain at least 2000m of sediments that thicken seaward to at least 3000m near the continental shelf edge (Behrendt et al., 1974; Fig. 33). Shallow basement occurs elsewhere in the continental shelf. Magnetic lineaments in areas of shallow magnetic basement in the continental shelf were related to doleritic dikes that are essentially parallel to the dikes that cut the Paleozoic Precambrian rocks (Fig. 33). The free-air gravity anomaly and isopach maps suggest main centers of deposition to lie in the upper continental-rise area (Behrendt et al., 1974). The sediments have been inferred to be thickest under the continental slope. Farther seaward from the continental slope, the sediments seem to thin. Thus, the depocenter of these basins in the continental slope-upper continental rise is subparallel to the shelf edge, to the elongation of shelf basins, to the coastline and to the doleritic dikes of Jurassic age (Behrendt et al., 1974; Schlee et al., 1974). In contrast, immediately to the north of the Grand Cess Ridge, sediments have infilled elongated depressions which are subparallel to the ridge and consequently these troughs are transverse to the direction of the edge of the continental shelf (Fig. 34).

Discussion

The tectonic framework of the part of the Liberia margin considered in this work seems to be influenced essentially by two main structural directions. One direction is northwest-southeast and coincides approximately with the direction of the continental shelf-edge, with the coastline and with the elongation of the depocenters of the continental margin. This direction is also coincident with the strike of the doleritic dikes of Liberia and with the direction of the shear zone that corresponds to the radiometric-age boundary between Pan-African and Liberian rocks. However, this northwest-southeast direction of apparent continental breakup for Liberia contrasts markedly with the northeast-southwesterly oriented main structural trends of the shield (Fig. 32). Thus, the northwest-southeast trend of the Mesozoic or younger structures most probably has developed following main structural trends of Pan-African age that were reactivated in the Jurassic. In this northeast-southwest structural trend, faulted blocks and faults formed localized centers of deposition. The sediments in basinal areas of the continental shelf prograded seaward apparently without major obstacles and built up a continental margin sedimentary prism.

The second structural trend is associated with the trend of Grand Cess Ridge. This ridge was present as a continuous barrier for sediments since the beginning of the continental margin deposition because it separates distinct continental-margin sedimentary provinces. Linear depressions subparallel to the Grand Cess Ridge and between two linear highs are evidences for confined preferential sedimentation in the margin along an east-northeast west-southwest direction(Fig.34). The continental shelf close to the projected landward extension of the Grand Cess Ridge is almost bare of sediments and a possible fault is inferred in the continental shelf and in the littoral zone of Liberia, as its probable landward expression (Behrendt et al., 1974 ; Schlee et al., 1974). Consequently, a structural high in the continental shelf appears to correspond to the continuation of the Grand Cess Ridge.

The 173-192 m.y. old diabase intrusions and possibly contemporaneous basalt flows (Hettangian to Early Bajocian) although present onshore and offshore, were most probably post-dated by younger volcanics of early Cretaceous age evidenced by ages of volcanic rocks in one offshore drill hole (Schlee et al., 1974). In the offshore area, Cretaceous sedimentary rocks generally overlie basalts and flows as mentioned before. These sedimentary rocks were dated on the basis of spores, pollen, gastropods and ostracods as Albian or older (Schlee et al., 1974). The time gap (50 m.y.) between the Jurassic magmatism and the early-Cretaceous volcanism and sedimentary deposition may delineate a time of development of the tectonic framework of the basinal areas close to the coastline in which no appreciable sediment accumulation has been preserved.

Thus, the Liberia continental margin sector corresponds to a sedimentary basin which had its tectonic framework established since the early-rifting of North America and Africa. This tectonic framework is mainly represented by basinal areas that extend in a northwest-southeast direction that are bordered by the rim of the Western African craton in Liberia and by an east-northeast-west-southwest ridge (the Grand Cess Ridge), that is traced laterally into a tectonic high in the continental shelf. Sedimentation off western Liberia was apparently open westward and built up a continental rise that grades laterally into an abyssal plain. This open sedimentation and the seaward thickening of the basinal areas of the continental shelf suggest that these basins had no major barriers or dams and thus fitting the pattern of half-graben or open-type of basins.

In contrast, sediments were ponded to the north side of the Grand Cess Ridge and prograded seaward along the strike of the ridge . This longitudinal progradation caused the continental rise to become wider in the southeastern part of the section. Elongated troughs parallel to the Grand Cess Ridge accumulated more sediments than the adjacent regions. Because the Grand Cess Ridge is traced laterally into the Saint Paul's Fracture Zone, it is apparent that fracture zone tectonics played an important role in developing the tectonic structures present in southeastern Liberia.

Ivory Coast Continental Margin Sector

Physiography

The Ivory Coast continental margin sector is situated offshore from southeast Liberia, the Ivory Coast and west-southwest Ghana. The

Ivory-Coast sector is constituted by a continental margin that prograd ed south, southwest and westward. The morphology of the sector consist ing of continental margin and abyssal provinces is bounded by the Grand Cess Ridge to the north, the Ivory-Coast-Ghana Ridge and the southern transverse ridge of Romanche Fracture Zone to the south, and the flanks of the Mid-Atlantic Ridge to the west (Fig. 31).

The Ivory-Coast sector contains two distinct morphological re gions that merge together. One region is relatively narrow and is lo cated west of Cape Palmas;the other region corresponds to the remainder of the continental margin that is off the Ivory Coast and southwest Ghana. The area situated to the west of Cape Palmas displays iso baths that are subparallel to the narrow continental shelf south of the Grand Cess Ridge, and that curve around the shelf edge off Cape Palmas (Fig. 31). The continental slope of this region is topographically marked by several irregularities and has at least two seamounts at its southeastern end (Figs. 3 and 4,profiles 20-22). This area ex tends westward into a confined and narrow abyssal plain. The abyssal plain coincides with the northern trough of the Saint Paul's Fracture Zone which is confined between the Grand Cess Ridge and the transverse ridge immediately to the south (Fig. 4, profile 27). Westward from approximately 10.5ºW longitude, the abyssal plain probably does not con tinue into the generally flat topography of the trough of the Saint Paul's Fracture Zone because the latter is slightly shallower in the west and both have distinct acoustic stratigraphies (Fig. 4, profiles 27-31, 33 and 34). The southern portion of the continental rise westward from Cape Palmas is increasingly more geomorphologically related to the rest of the Ivory Coast continental margin sector toward the east (Fig. 31).

The continental margin off the Ivory Coast and Ghana has a shelf-edge subparallel to the coastline except off Cape Three Points. The continental shelf is widest southeast of Cape Three Points and gives way to a steep continental slope that descends directly to the Guinea abyssal plain (Fig. 31). No continental rise is developed in this region. The continental shelf is generally smooth and featureless (Martin, 1971) and the shelf break occurs between 100 and 130m. "Le Trou Sans Fond", a submarine canyon, incises the continental shelf and is located at the inflexion point of the change in the direction of the coastline of the Ivory Coast and Ghana (Martin, 1971).

The continental slope and the upper continental rise of the Ivory Coast sector widens in the area immediately to the north of the Ivory Coast - Ghana Ridge to form a low-gradient area named the Ivory Coast marginal plateau (Delteil et al., 1974). The plateau lies between 1500 and 2000m and is the result of ponding of sedimentation by the linearly continuous Ivory Coast - Ghana Ridge to the south (Fig.31). This preferential and semi-confined sedimentation and the bulging configuration of the isobaths immediately to the north of the ridge, suggest a west-southwest direction of sedimentary progradation. This contrast with a main north-south direction of progradation for the major part of the Ivory-Coast sector. The continental rise morphology extends into the Guinea abyssal plain through a topographic gap between the Ivory Coast-Ghana Ridge and the southern transverse ridge of Romanche Fracture Zone (Figs. 3 and 4, profiles 14-17). Consequently, the lower continental rise in the area within the topographic gap of the Roman-che Fracture Zone is shared by the Ivory Coast sector and by the Ghana-Togo-Dahomey-Nigeria continental margin sector to the south.

The Guinea abyssal plain is bounded to the north by the lower continental rise in the region of the topographic gap of the Romanche Fracture Zone and by the southern transverse ridge of Romanche Fractu-re Zone at 6.5ºW (Fig. 31). A small elongated abyssal plain that coincides with the deepest region of the Ivory-Coast sector has developed immediately to the north of the southern transverse ridge of the Roman-che Fracture Zone in the main trough of this fracture zone (Fig.31).

General Geology and Tectonic Framework

Shield rocks which show northeast-southwest foliations form the bulk of the rocks that border the Ivory Coast continental-margin sector. Most of these rocks belong to the Birrimian shield that has been dated as 1600-2050 m.y. (Grant, 1969) and that constitutes the Eburnean radiometric province (ca. 2000 m.y.) (Fig. 35; Hurley et al., 1971; Hurley and Rand, 1973). These Eburnean-age rocks are composed of metamorphic schists that are intruded by granites and granodiorites which are aligned in a northeast-southwest direction along the main foliation trend of the shield. These schists contrast with the gneissic rocks of Liberia to the west and with the crystalline massif of southern Ghana (Spengler and Delteil, 1966). An important fault in the vicinity of Accra, with a north-northeast - south-southwest strike, coin-cides with the radiometric-age boundary between Eburnean and Pan-Afri-can age rocks (Fig. 35; Grant, 1969; Hurley, 1972; Hurley and Rand, 1973). The Pan-African age rocks, to the east of Accra fault, constitute the Dahomeyan and Nigerian basement complex. The north-northeast -south southwest fault zone is complex and includes some areas of thrust faulting. The zone of thrust faulting is located between metasedimentary groups to the east and sedimentary rocks to the west. The sedimentary rocks are of late Precambrian to early Paleozoic age and belong to the so-called Voltaian Group. The metasedimentary rocks constitute the Buem and Togo formations. The Voltaian group and the Buem and Togo formations were interpreted by Grant (1969) as belonging to a single sedimentary group that to the east of the fault zone became deformed and metamorphosed because of tectonic events associated with the Pan-African thermotectonic event (Buem and Togo formations). To the west, the sediments are lying on a stable platform (Voltaian Group) (Fig. 35). Burke (1969) reports recent earthquake activity in the vicinity of Accra which may be connected with recent tectonic movements along blocks bounded by this fault zone. Isostatic gravity anomalies associated with the Accra fault zone show a steep gradient that coincides with the zone of transition from Eburnean to Pan-African age rocks.A crustal model to fit the gravity data implied two different crustal thicknesses across the Accra fault zone (Crenn, 1957; Grant, 1973). Consequently, geological, geophysical and seismological data point out that the Accra fault zone is a deep structure in the Western African shield and, as such, it has been a weakness zone since the Pan-African thermotectonic event.

The oldest sedimentary rocks (lower Paleozoic) that outcrop in the Ivory-Coast sector close to the coastline are those in the littoral zone of Ghana at Takoradi, Sekondi, Elmina and Accra (Machens , 1973).

Mesozoic and Tertiary rocks of the Ivory-Coast sector outcrop mainly in the Ivory Coast and in southwestern Ghana and are related to the Ivory Coast Basin (Spengler and Delteil, 1966). Other areas of exposed Mesozoic and Tertiary rocks are in the littoral zone of southern Ghana and to the east of the Volta - delta region (Machens, 1973). The area to the west of the Volta delta corresponds to the westward continuation of the Togo-Dahomey Basin that will be discussed later. Other sedimentary basins in the Ivory-Coast sector include the offshore-Ghana Basin (Delteil et al., 1974) and a confined basinal area immediately to the southeast of Cape Palmas. This basin in the Liberian continental shelf is probably the continuation of the trough seen in profile 20 (Fig. 4). This basin probably is very restricted because most of the continental shelf of the western Ivory Coast is relatively bare of sediments (Lucien Montadert, pers. comm., 1975).

Ivory Coast Basin - The Ivory-Coast Basin is a marginal basin along a narrow zone of the littoral of the Ivory Coast and southwestern Ghana. The portion of the basin in the coastal region is separated from a shallow platform by a normal fault (Ivory Coast fault),(Spengler and Delteil, 1966). To the north of the fault the shield is thinly covered by Mesozoic and Tertiary sediments whereas, immediately to the south, sediments as thick as 5000m have been reported (Spengler and Delteil, 1966). This marginal fault extends along the coastal plain of the Ivory Coast and is generally oriented in an east-west direction. Close to the city of Abidjan, there is a slight change in the strike of the fault. Subsidiary faults parallel to the Ivory Coast marginal fault were suggested by Spengler and Delteil (1966) (Figs. 36 and 37).

The basin contains faulted blocks that extend in a northeast - southwest direction parallel to the main structural direction of the rocks of the Birrimian shield (Spengler and Delteil, 1966). However, these structures are secondary in relation to the dominant east-west trend of the marginal fault and only reflect local reactivation of old structural trends of the shield.

The oldest sediments present in the Ivory Coast Basin are detrital sedimentary rocks that directly overlie Precambrian rocks. These sedimentary rocks are of variable thicknesses (472-2000m) along the basin and consist of sandstones, conglomerates and variegated shales with some intercalations of black shales (Spengler and Delteil, 1966). This basal series ranges in age from late Jurassic-early Cretaceous to Aptian-Albian. The first marine transgression is dated as Aptian-Albian in age. A period of marine regression is believed to have occured at the end of early Cretaceous which was followed by a period of marine transgression at the beginning of the late Cretaceous. The lower Tertiary sedimentation was not extensive in the Ivory Coast Basin at least in its onshore portion, and Oligocene rocks have not been reported yet (Fig. 37; Spengler and Delteil, 1966). Miocene marine sediments are restricted to the vicinity of Abidjan. According to drilling and seismic profile data, Miocene sediments seem to lie discordantly over the Cenomanian in the continental shelf. Mesozoic magmatism has not been described in the Ivory Coast Basin (Spengler and Delteil, 1966) but its "absence" may reflect the lack of ubiquitous drilling data in the offshore region of the basin.

Drilling data and some seismic profiles of deep penetration enabled Spengler and Delteil (1966) to conclude that the Ivory Coast Basin is an open type of basin. Indeed, most of the sediments that built up the sedimentary prism of the Ivory Coast continental margin sector bypassed the Ivory Coast Basin and its offshore extension apparently without being barred by offshore tectonic highs. Consequently , the continental margin sediments constitute the seaward progradation of the Ivory Coast Basin.

The westward and southeastward continuation of the Ivory Coast Basin in the continental shelf is not known. The Ivory Coast fault ends in the littoral zone east of Fresco and west of Axim (Fig.36). In the western side of the Ivory Coast, the continental shelf break and the relatively steep continental slope are apparently in line with a ridge (St. Paul's Ridge of Behrendet et al., 1974) that is traced laterally into one of the transverse ridges of Saint Paul's Fracture Zone (Fig. 7). This lineament may have controlled the linear character of the shelf break in this region. There is no specific data which shows that this lineament extends into the western extremity of Ivory Coast fault, although the similarities in strike of the two features are suggestive of mutual dependence. In contrast, the eastern extremity of the Ivory Coast Fault has an almost orthogonal relationship with the main trend of the flexure zone that corresponds to the northern boundary of the offshore-Ghana Basin, and with the strike of the Ivory Coast-Ghana Ridge.

<u>Offshore-Ghana Basin</u> - The offshore-Ghana Basin is a confined basin between the littoral and a shelf-edge high in the continental shelf off southern Ghana. A deep marginal fault separates the littoral

region from the basinal area. The oldest sediments are early Paleozoic in age (Cudjoe et al., 1973, in Delteil et al., 1974; Kevin Burke,pers. comm., 1975). These sediments are overlain by a Mesozoic to Tertiary sequence. The offshore-Ghana Basin has been interpreted as a horst and graben structure generally caused by NE-SW or NW-SE trending faults (Cudjoe et al., 1973, in Delteil et al., 1974).

The existence of a shelf-edge high in the southern Ghanaian continental shelf is inferred by several lines of evidence. The continental slope off southern Ghana is the steepest slope in the Gulf of Guinea and links the continental shelf with the lowermost part of a poorly developed continental rise to the south, suggesting that no appreciable sediment progradation has taken place from the continental shelf southward (Figs. 31 and 4, profiles 8-10). This steep slope is continuous into the Ivory Coast-Ghana Ridge which, as far as seismic profiles and the submarine morphology have shown, is a linear barrier for sediments (Delteil et al., 1974; Fig. 4, profiles 11-14). Core samples from the Ivory Coast-Ghana Ridge close to the continental shelf recovered coarse detritic marine sediments dating from the middle to late Albian (Fig. 31; Delteil et al., 1974). The coarseness of these sediments contrasts with the argillaceous sediments of Albian age present to the north of the ridge and in the continental shelf off the Ivory Coast (Delteil et al., 1974). These core samples were recovered below 7m of Recent sediments suggesting that the Ivory Coast Ridge is a structural high. Being a barrier for the continental-margin sediments and also containing the coarse Albian sediments, the Ivory Coast-Ghana Ridge has been interpreted as a high zone since the middle to late-Albian time (Delteil et al., 1974).

Indirect evidence suggesting that a high area existed close to the edge of the Ghanaian continental shelf since the beginning of the evolution of the offshore-Ghana basin is that sedimentary-facies changes in the Mesozoic sequences of the basin were generally from the east towards the west (Kazumi Miura, pers. comm., 1974). This is better explained by an offshore barrier that remained high throughout the evolution of the offshore-Ghana Basin and that prevented appreciable sediment progradation from the north to the south and instead prograded along the east-west axis of the basin.

In consequence of this east-west sediment progradation, conspicuous sediment ponding along the northern side of the Ivory Coast-Ghana Ridge and the accumulation area of the Ivory Coast Plateau were created.

Dredge results (Delteil et al., 1974) at the eastern end of the Ivory Coast-Ghana Ridge in the continental slope south of the Cape Three Points (Fig. 31) showed schists and micaschists, and coarse-grained, feldspathic, micaceous sandstones and micaceous, ferruginous sandstones that were comparable to Devonian sandstones in Takoradi. These results indicate that the Ivory Coast-Ghana Ridge in the dredged region is formed by continental rocks (Delteil et al., 1974).

A basinal area is present immediately to the north of the Ivory Coast-Ghana Ridge in the continental rise (Delteil et al.,1974). This basin is elongated and parallel to the ridge (Figs. 7 and 4, profiles 11 and 12). It probably represents the westward continuation of the offshore-Ghana Basin. Because the tectonic high of the Ghanaian shelf is continuous with the Ivory Coast-Ghana Ridge and because the ridge is most probably traced laterally into the southern transverse ridge of Romanche Fracture Zone, the offshore-Ghana Basin and its prolongation in the continental rise occupies the same relative position as the main Romanche Fracture Zone trough in relation to the transverse ridge (Fig.7).The marginal flexure zone in the littoral zone of southern Ghana which corresponds to the northern boundary of the offshore-Ghana Basin was probably a linear high limited by normal faults. The Paleozoic sedimentary cover of the crystalline basement was entirely removed by erosion with the exception of regions close to the littoral, adjacent to the flexure zone. This linear high has the same relative position of the northern transverse ridge of Romanche Fracture Zone in relation to the offshore-Ghana Basin and to the main trough of the fracture zone.

Discussion

The Ivory Coast continental margin sector is tectonically bounded by three important structural lineaments. Two of them are sub-parallel to each other and correspond to the northern and southern boundaries of the sector. The northern boundary of the sector is associated with the Grand Cess Ridge and the southern limit is marked by the Ivory Coast-Ghana Ridge. The third structural boundary coincides with the Ivory Coast fault.

Because a series of ridges and troughs associated with the St. Paul's Fracture Zone acted as barriers and conduits,respectively , for terrigenous sediments, the continental margin westward from Cape Palmas was semi-isolated from the rest of the continental margin sector up to the time when sill depths were reached by sedimentation.

The Liberia and the Ivory Coast continental margin sectors have many similarities in tectonic settings and in geomorphology. The Liberia and the Ivory Coast marginal plateaus for instance, have identical areas of occurence to the north of the prominent Grand Cess and Ivory Coast marginal ridges respectively, as pointed out by Delteil et al., (1974). Graben basins are associated with fracture-zone directions whereas the half-graben basin is located between fracture zones.

Basic volcanic rocks dated at 160 to 165 m.y. have been reported by Cudjoe et al. 1973 (in Delteil et al., 1974) on the continental shelf of Ghana. This volcanism contrasts in age with the first Aptian-Albian marine transgression in the Ivory Coast Basin and probably in the offshore-Ghana Basin, as suggested by core samples taken from the Ivory Coast-Ghana Ridge. These volcanic rocks probably were the result of the first tectonic movements that imprinted the horst and graben framework of the offshore-Ghana Basin during the time when Africa and Brazil were together.

The northeast-southwest structural trend of the shield rocks of the Ivory Coast did not influence the localization of the main structural Mesozoic directions of the Ivory Coast continental margin sector. In general, the Mesozoic structures cut the old weakness zones of the shield at sharp angles. The northeast-southwest trend of the shield influenced only locally the basement configuration (e.g. Ivory Coast Basin and offshore-Ghana Basin). The North-northeast-south-southwest Accra fault zone that is probably a suture zone between two cratons is cut at almost right angles by the lineament represented by the offshore-Ghana Basin.

In the lower continental rise of the Ivory Coast continental margin, an abrupt transition from a thickly covered acoustic basement to a considerably shallower basement was mapped as the so-called continental rise fault (Arens et al., 1970). These authors tentatively interpreted the continental rise fault as the possible boundary between oceanic and continental crust. Le Pichon and Hayes (1971), in their reconstruction of the equatorial Atlantic for 80 m.y.b.p., showed that the continental rise fault coincided with the northwest-southeast extension of the North Brazilian Ridge (Hayes and Ewing, 1970). An interpretation for the origin of the continental rise fault of Arens et al. (1970) and of the northwest-southeast segment of the North Brazilian Ridge will be given later.

As discussed earlier, the southern transverse ridge of Romanche Fracture Zone was tentatively traced into the Ivory Coast-Ghana Ridge. This ridge represents a tectonic high in the continental rise and tentatively followed, as a continuous feature, the tectonic high of the continental shelf edge of Ghana. By examining the morphology of Romanche Fracture Zone in mid-ocean and the horst and graben character of the offshore-Ghana Basin it is very tempting to conclude that the marginal basin of the graben type was a consequence of fracture zone tectonics. Because the offshore-Ghana Basin is entirely confined in continental crust, the latter conclusion implies that fracture zone tectonics do not discriminate continental from oceanic crust.

Ghana-Togo-Dahomey-Nigeria and Fernando Poo-Cameroon Continental Margin Sectors

Physiography

The continental margin off Ghana east of Volta delta, off Togo, Dahomey and southwestern Nigeria has a smooth continental rise that grades laterally into the Guinea abyssal plain. The Guinea abyssal plain extends westward into the flanks of the Mid-Atlantic Ridge and farther westward as tongue-like extensions of abyssal-plain regions which coincide with generally east-west aligned troughs. The northern boundary of the Ghana-Togo-Dahomey-Nigeria (GTDN) sector is placed in the continental slope off Ghana, westward from Accra, and in the Ivory Coast-Ghana Ridge (Fig. 31). West of Ivory Coast-Ghana Ridge, the northern boundary of the GTDN sector is not obvious because, as it was shown earlier, sill depths have been crossed along the connection of the Romanche Fracture Zone with the Ivory Coast-Ghana Ridge. Westward from the buried extension of Romanche Fracture Zone, the southern transverse ridge of the same fracture zone continues to be a real boundary between the Ivory Coast and the adjacent crustal segment to the south.

The GTDN sector is individualized from the Fernando Poo-Cameroon continental margin sector to the south by a buried ridge, the Jean Charcot Ridge (Fig. 31). From north to south isobaths representing depths greater than 1000m do not parallel the bulge of the Niger delta in the way the isobaths shallower than 1000m do. This is demonstrated off Niger Delta by a pronounced bulge in the isobaths between 1500m and 3500m (Fig. 31). Although it may be interpreted as due to irregularities in the topography caused by intensive diapirism of salt (Mascle et al.,

1973) or mud (Delteil et al., 1974, 1975) the bend of the isobaths is probably very significant because it is coincident with the location of the Jean Charcot Ridge. The Fernando Poo-Cameroon continental margin sector lies between the Jean Charcot and the Fernando Poo-Annobon Ridge (Fig. 31).

The continental shelf off southeastern Ghana bulges around the Volta River delta but is essentially parallel to the littoral zone throughout Togo, Dahomey and Nigeria (Fig. 31). It is generally wider off the Niger Delta and it is indented by at least four prominent submarine canyons in the GTDN and in the Fernando Poo-Cameroon continental margin sectors. The Avon Canyon is located where the east-west oriented shelf-edge changes its strike to a northwest-southeast direction in the region to the east of Lagos (Fig. 31). The Mahin canyon indents the continental shelf very close to the Avon canyon and the two may merge on the continental rise (Brower, 1973). The Kwa Ibo and the Calabar canyons (Houbolt, 1973; Brower, 1973 and Allen, 1964) are present to the west of Fernando Poo Island and extend into the continental rise following the trend of the Fernando Poo-Annobon Ridge (Fig. 31). These two canyons seem to end at about 2800m depth (Houbolt, 1973). Seaward from the terminus of these two canyons, a channel (Principe Channel)is continuous westward to 1ºE of longitude and to a depth of approximately 4600m (Houbolt, 1973) (Fig. 31). The location of the Kwa Ibo and the Calabar canyons is coincident with the region of the Cameroon and Nigeria where the littoral and the shelf edge change their strikes appreciably (Fig. 31).

The topographic gradient of the continental rise of the GTDN sector becomes essentially east-northeast-west-southwest in depths greater than 4000m and grades smoothly into the Guinea abyssal plain. The strike of the topographic gradient of the continental rise of the Fernando-Poo-Cameroon continental margin sector is northeast-southwest in the upper rise and becomes more east-northeast-west-southwest in the lower continental rise.

The Fernando Poo-Annobon Ridge consists of various islands in the Gulf of Guinea and intervening seamounts and basement highs (Figs. 3 and 4, profiles 1-4). It has been an effective barrier in preventing the coalescence of the Fernando Poo-Cameroon sector with the continental margin off Gabon, Equatorial Guinea and Cameroon. Sill depths have been crossed in the upper continental rise, but southward, the ridge has been more effective as a sediment barrier.

General Geology and Tectonic Framework

The Dahomeyan and Nigerian basement complex lies to the east of the Accra fault in Ghana. This basement complex, although in some places gives radiometric ages similar to the Birrimian shield (1600-2050 m.y.) situated to the west of the Accra fault, generally yields radiometric ages in the range of 500-700 m.y. (Fig. 38; Grant, 1973). Close to the Accra fault or thrust zone, the Dahomeyan shield is composed of granulite-facies rocks, eclogites (Ghana), pyroxene-bearing basic gneisses (Togo and Dahomey), and occasionally ultrabasic rocks (Kabrē massif in Togo) (Grant, 1973). Farther east, the Dahomeyan and Nigerian basement complex is mostly constituted by amphibolite-facies gneisses such as biotite-hornblende gneisses, amphibolites and meta-basic rocks, migmatites, granites, gneissic potassium-rich syenites and metasedimentary rocks (Grant, 1973).

The foliation of high-grade metamorphic rocks is essentially north-northeast-south-southwest and northeast-southwest and metasedimentary belts generally follow the trends of that foliation. These rocks yielded radiometric ages correlatable with the time span of the Pan-African thermotectonic event of Kennedy (1965, see also Grant, 1973). In some areas, there are rocks that show evident poliphase deformational history that contrast with the structural history of rocks elsewhere. Intruded in these poliphasic deformed rocks there are granitic orthogneisses that had given radiometric ages older than 1650 ± 220 m.y. (Jacobson et al., 1963). Consequently, the Nigerian-Dahomeyan shield is not constituted by rocks of Pan-African age only. In some areas old basement rocks were clearly tectonically reactivated during the Pan-African thermotectonic event (McCurry, 1971; Grant, 1973). In others, shield rocks had their radiometric clocks reset by a strong thermal event (500-700 m.y.) and in still other regions, geosynclinal belts were deformed and incorporated in the cratonic area by events of Pan-African age (McCurry, 1971; Grant, 1973).

The granites that intruded the Dahomeyan-Nigerian basement complex are essentially the so-called Older and Younger granites (Jacobson et al., 1963; Black and Girod, 1970). The Older granites have been emplaced in the basement rocks during the Pan-African thermotectonic event. The emplacement of the Older Granites was followed by widespread migmatization and granitization. All the known occurrences of Older granites point to a synkinematic emplacement (Jacobson and others, 1963). They consist of granites, granodiorites and quartzdiorites that reach batholitic proportions in some areas.

The Younger granites correspond to non-orogenetic intrusions (Black and Girod, 1970). They have been interpreted as magmatic intrusions and the majority of the occurrences contain volcanic and/or shallow crustal level acidic members together with plutonic representatives. The Younger granites correspond to the late Paleozoic-Mesozoic granites of Niger and Nigeria and to the Tertiary granites of the Cameroon (Black and Girod, 1970, Fig. 39).

The Younger granites span in age from the late Paleozoic (Aïr region) to Tertiary (Cameroon). In Niger and Nigeria they correspond to more than 60 massifs roughly lying in a north-south strip (Black and Girod, 1970; Figure 39). Radiometric ages of Younger granites in Nigeria have fallen in the mid-Jurassic (160 m.y.) (Jacobson and others, 1963). The granitic complexes in Aïr were dated by K/Ar method as 295 m.y. (Black and Girod, 1970). In the Cameroon and Tchad a group of more than twenty small plutons aligned in a northeast-southwest direction are inferred to be Tertiary from datings in two of these massifs (Lasserre, 1966; Fig. 40). In the Hoggar area ring complexes probably of Cretaceous age have been described (Black and Girod, 1970). In the Tchad, the Tibesti massif has been the center of intense volcanism from the early Tertiary to the Quaternary which includes basaltic and rhyolitic magmas with widespread ignimbrites (Vincent, 1970).

The Mesozoic (mid-Jurassic)granites of Nigeria form ring complexes in which ring dikes, cone sheets, and volcanic cauldron subsidence are common (Black and Girod, 1970; Jacobson and others, 1963;

Jacobson and others, 1958; Lasserre, 1966). These post-orogenetic granitic centers generally contain plutonic, hypabissal and volcanic acidic roeks together with minor gabbro, diorite,andesite, diabase and basalt. Rhyolites and ignimbrites generally were extruded in vast quantities before plutonic equivalents were emplaced by cauldron subsidence (Jacobson et al., 1958; Black and Girod, 1970). The rhyolites and the welded tuffs are now only partially preserved in areas that subsided along ring-faults (Black and Girod, 1970).

The sites of intrusion of the Younger granites were coincident with strong uplift as evidenced by acidic volcanics lying immediately above the basement. These sites, even today, represent plateau areas (i.e. Jos Plateau) (Black and Girod, 1970). Although the majority of these post-orogenetic granitic intrusions are roughly contemporaneous, when each area of occurrence is considered as a unit , each center of magmatism apparently corresponded to a unique magmatic chamber during its emplacement (Black and Girod, 1970). This conclusion arose because, apparently, igneous activity in each volcanic center ended before the initiation of a new magmatic center and because the basement separating individual massifs was rarely intruded.

Petrographically, the Younger granites are constituted by rhyolites,quartz-syenites, peralkaline granites, hastingsite-granites and biotite-granites that make 95% of all petrological types. The remaining 5% are represented by anorthosites and olivine-gabbros (Black and Girod, 1970).

Partial fusion of crustal and sub-crustal rocks was probably responsible for acidic and primary intermediate magma generation that gave rise to the Younger granites. Black and Girod (1970) argued that besides the cauldron subsidence evidence, the existence at depth of immense basaltic magma reservoirs as the source for granitic differentiates seems unlikely in view of the rarity of associated basaltic lavas in a terrain sliced by pre-Ordovician wrench faults which provided readily-opened conduits for the rise of basalts, as shown by the Tertiary-Recent volcanism. Characteristic of the Younger granites was a widespread mineralization of tin, tantalum and niobium. Black and Girod (1970) also pointed out the coincidence of the presence of cassiterite both in the Younger granites and in pegmatites of the Nigerian basement complex, and its absence in similar igneous provinces in several parts of the world. Tugarinov (1968) found isotopic lead ratios leading to a Katangan-Damaran (Pan-African) age in feldspar from Younger granites that further suggests the possibility of partial fusion of crustal and sub-crustal rocks, for the origin of the Younger granites.

The period of acidic magmatism in mid-Jurassic (160 m.y.) in Nigeria and Niger was roughly contemporaneous with diabase intrusions in Liberia and probably with basaltic magmatism in the offshore-Ghana Basin at 160 m.y.

Tertiary granites in ring-dyke complexes and boss-like outcrops are aligned in a northeast-southwest direction in the Cameroon and Tchad. This strike follows the foliation direction of the shield and the marked northeast-southwest direction of Annobon-Fernando Poo and Mount Cameroon lineament with which the Tertiary granites merge. Radiometric ages of 54 $\pm$ 22 m.y., 33 $\pm$ 5 m.y. and 38 $\pm$ 4 m.y. have been found for the granitic massif of Poly. The Mayo Darlē massif has yielded ages of 56 $\pm$ 8 m.y. and 65 $\pm$ 12 m.y. The Golda Zuelva granite complex was found to be 38 $\pm$ 5 m.y. (Lasserre, 1966; Fig. 40). These Tertiary granites correspond to the youngest granitic intrusions in Western Equatorial Africa.

Other magmatic intrusions and extrusions from Mesozoic through the Recent are widespread within the Nigerian basement complex. Pre-Albian volcanics have been confirmed in the Abakaliki region of the Benue Trough (Uzuakpunwa, 1974). These volcanics consist of pyroclastics (agglomerates and tuffs). The agglomerates contain basalt clasts and other rocks in a finegrained andesitic matrix. The tuffs are generally lithic in character and contain angular rock-fragments that include basalt and earlier pyroclastics. These pyroclastics are overlain by shales of Albian (middle Albian?) age (Uzuakpunwa, 1974). Contrary to the interpretation of Okezie (1965, in Uzuakpunwa, 1974) who believed that these volcanics were products of late Cretaceous volcanism, the pre-Albian age is also evidence against the suggestion of andesitic magmatism in the Santonian time in the Abakaliki region of the Benue Trough, put forward by Burke et al. (1970) using Okezie's interpretation.

In the Benue Trough, thin occurrences of tuffs and possibly lavas were found interbedded with sediments older than Albian (?) - Cenomanian (Carter et al., 1963). Lead-zinc mineralization, probably associated with a large number of intermediate to basic intrusions, is dominant in the Albian shales and fewer occurrences have been found in the Turonian strata. No mineralization has been found in post-Turo nian sediments (Nwachukwu, 1972). The erosion surface which characterizes the Abakaliki folded belt where it becomes buried by younger sediments and the Calabar flank was locally overlain with efusives(tuffs and lava flows) as shown by drilling wells (Fig. 38; Murat, 1970).One of these wells penetrated an effusive sequence more than 1300m thick in the region of the Calabar flank. Very rare basic intrusions of late Cretaceous-Eocene age can also be seen cutting through upper Senonian sediments on the western limb of the Abakaliki uplift (Murat, 1970).

Jos Plateau, Biu, Southern Cameroon and Ngaoundéré are regions of Tertiary to Recent volcanism of alkaline character. Southern Aïr, Hoggar and Tibesti were also affected by the alkaline magmatic activity (Fig. 41). The alkaline magmatism resulted in widespread volcanism. Generally, the extrusions were contemporaneous with important epeirogenic movements that uplifted the sites of magmatic concentration (Jos Plateau-1000m of uplift; Aïr-500m of uplift) (Black and Girod, 1970). According to Black and Girod (1970),the volcanoes of the Jos Plateau occur along the more recent north-northeasterly faults. The same authors pointed out that, both in Niger and Nigeria,volcanic craters of Tertiary to Recent magmatism are located on Younger granite ring-faults and on the intersection of these faults with pre-Ordovician transcurrent faults of the basement.

The tectonic setting and the magmatic character of the rocks extruded in the Tertiary-Recent times is very similar. The alkali-olivine-basalt-trachyte association is common to these volcanic rocks. They are widespread in area of the Western African craton (Black and

Girod, 1970; Fig. 41). The rocks are characterized by the predominance of basic types (basanites, basanitoids, ankaratrites); by the scarcity of intermediate types (hawaites, mugearites); and by the peralkaline nature of some of the most differentiated members (peralkaline phonolites, comendites, pantellerites) (Black and Girod, 1970).

Mount Cameroon is an active volcano (De Swardt,1954;Hedberg, 1968) and includes lavas which are particularly deficient in silica and sometimes potassic (nephelinites, leucitites and hauynophyres) (Black and Girod, 1970). To the northeast of Mount Cameroon, the Adamoua plateau region (Sarcia and Sarcia, 1952) is partially covered by basic volcanics of Tertiary to Recent age. These volcanics follow a general northeast-southwest trend, as far north as Mount Kambo. From Mt. Kambo northward, the Tertiary to Recent volcanism spreads in area and forms several centers in the Cameroon (the Ngaoundéré being the largest one, Fig. 38). Andesites, as a minor constituent of these volcanics in the Cameroons, have been often referred in the literature (Sarcia and Sarcia, 1952; Furon, 1960; Nwachukwu, 1974;Murat, 1970).

The Tertiary to Recent alkaline volcanism is concentrated generally in areas that were previously intruded by the Younger granites (Mid-Jurassic and Tertiary). This alkaline volcanism is not an event restricted to the cratonic areas in equatorial Africa, it is also present in the islands of the Gulf of Guinea that make up the oceanic continuation of the Cameroon Trend (Furon, 1960).

The sedimentary basins of Mesozoic age that are present within the craton are either roughly parallel or roughly transverse to the foliation trends of the shield. These sedimentary basins are the Dahomey, Benue-Niger Delta and the Douala (Fig.38).

Dahomey Basin or Togo-Dahomey Basin - The Dahomey Basin is a sedimentary embayment located in Togo, Dahomey (now Benin) and westernmost Nigeria and may extend westward into the offshore-Ghana Basin. Southeastward the basin is separated from the Niger Delta Basin by the Okitipupa ridge which corresponds to a high area in the westward continuation of the Benin Flank (Burke et al., 1970; Murat, 1970; Fig. 38).

The tectonic framework of the Dahomey Basin is poorly known. The basin shoals considerably toward the cratonic area and Furon(1960) reported a very fast transition from the thin cover of the shield (300m) to sediments as thick as 1800m at Cotonou. Sediments, as thick as 2400m (Porto Novo), are present in some parts of the littoral (ASGA-UNESCO, 1968) and probably they attain considerable thicknesses in the offshore portion of the basin. Two major faults of northeast-southwest and north-northeast-south-southwest directions have been mapped in the basin (ASGA-UNESCO, 1968; Murat, 1970; Fig.38). These faults are subparallel to the foliation of the basement complex.

The shape of the Togo-Dahomey Basin may reflect its fundamental structural orientation. The western side of the basin has essentially an east-northeast-west-southwest direction whereas the eastern side of the basin has a northwest-southeast direction. The first direction, as mentioned before, is in line with the northern boundary of the offshore-Ghana Basin and may represent its continuation in Togo and Dahomey,as suggested by structures mapped by Delteil et al., (1975). Another possibility is that the flexure zone of the littoral of Ghana may be linked with normal faults in the area close to the littoral in the region of Cotonou and Porto Novo, as suggested by the pronounced thickening of the sediments in those two areas (Fig. 38). It is not known if the shelf-edge high of the offshore-Ghana Basin is continuous into the Togo-Dahomey Basin. The northwest-southeast direction in the eastern side of the Dahomey Basin is roughly transverse to the trend of the Okitipupa ridge-Benin Flank of the Benue Trough-Niger Delta basins and roughly coincides with the direction of the Calabar Flank to the south (Fig. 38).

The Dahomey Basin appears to have a structure very similar to that of the Ivory Coast Basin. To the north, it may be bounded or is in continuation with Romanche Fracture Zone structures and, to the south, it is bounded or interrupted by a ridge that is in the same direction as ridges or highs that are associated with fracture zones in the Gulf of Guinea (Fig. 7). A major fault, similar to the Ivory Coast, may be present along the continental shelf but very close to the littoral (Fig. 42).

Campanian-Maestrichtian marine sediments are the oldest sediments reported in the Togo-Dahomey Basin on land, and these overlie continental deposits (Furon, 1960; Reyment, 1966; Machens, 1973). As far as the literature information is concerned, the offshore stratigraphy of the basin is virtually unknown. However, the presence of older sediments in the offshore region is very probable, because there is no reason to suppose that the origin of the Dahomey Basin is not contemporaneous with other marginal basins of the Gulf of Guinea.

Benue Trough - Niger Delta Basin - The Benue Trough is situated in Nigeria along the courses of the Benue and Niger rivers. It is an elongated sedimentary basin containing sediments as old as Albian.

The Benue Trough encroaches into the African shield in a northeast direction. It is a very linear feature 130-150 km wide and either terminates in the north of the Biu Mountains or might extend into the Tchad Basin (Fig. 38; Furon, 1960). Southestward, the trough merges into the Anambra and Niger Delta basins that by themselves merge into the continental margin of Nigeria, Dahomey, Togo and Ghana (Fig. 38).

The Benue Trough is bordered by rocks of the Nigerian basement complex. The flanks of the basin correspond to an abrupt transition from the exposed shield to the very deeply covered basement of the trough (more than 6000m of sediments). The northern border of the Benue Trough is oriented in a northeast-southwest direction and is called Benin flank (Fig.38). The southern border of the Benue is also represented by an abrupt lithologic transition from the shield rocks to the sediments of the trough (Fig.38). To the southwest, this flank is in line with the complex region of Abakaliki.

Several sedimentary embayments occur along the Benue Trough that are transverse to its main axis: (a) The Middle-Niger embayment with sediments dating from the Late Cretaceous (Adeleye and Dessauvagie, 1971); (b) the Calabar flank; (c) the Mamfe embayment; and (d) the upper Benue embayment (Fig. 38). The Mid-Niger embayment contains the middle course of the Niger river and probably is in line with the northern boundary of the West-African craton where the Gao Rift is located (Furon,1960). This embayment is abruptly interrupted to the south by the thicker successions in the Benue Trough. The Mamfe embayment, to the east of the Abakaliki folded belt, contains volcanics aligned in a northwest-southeast direction and it has approximately the same trend as the Mid-Niger embayment (ASGA-UNESCO, 1968). This embayment is separated from the Calabar flank to the south by the Oban hills that consist of outcropping shield rocks. The Calabar flank constitutes a region where the transition from the shield to the deeply covered basement is abrupt and certainly represents an area of intense normal faulting with down-to-the-basin blocks that are flanked by the Oban massif horst (Fig. 38).

The Benue Trough contains sediments at least as old as early Cretaceous, Albian-age sediments overlie continental sediments and volcanic rocks (Reyment, 1966; Uzuakpunwa, 1974). Basic volcanics are present in the Benue Trough as demonstrated by ancient centers of volcanism (ASGA-UNESCO, 1968) and intercalations of volcanic rocks in the sedimentary succession (Murat, 1970; Uzuakpunwa, 1974).

The first marine transgression took place in the mid-Albian (Reyment, 1966; Murat, 1970). The depocenter of the basin from the early Cretaceous to the Santonian-Campanian time was close to the southern flank of the trough, near the present position of the Abakaliki folded belt (Murat, 1970; fig. 43a). To the north, towards the Benin flank, a shallow platform, the Anambra platform, remained stable up to the same time. In the Santonian-Campanian time, tectonism caused folding in the sediments of the trough and transformed the trough into a positive area that underwent considerable erosion thereafter (Reyment, 1966; Murat, 1970; Burke et al., 1970). In consequence, the de pocenter of the basin changed to the former site of the Anambra platform that has since accumulated a great thickness of sediments (Anambra Basin) (Stoneley, 1966; Murat, 1970; Fig. 43b). At the end of the Eocene, a thick deltaic complex was deposited in the Anambra Basin, and this preceded the development of the Niger Delta Basin following an Oligocene period of subsidence in the Anambra Basin (Murat,1970).

The Santonian folding episode is presently represented by fold axes running parallel to the Benue Trough (Fig. 38). Dips are generally 5º to 25º and rarely exceed 30º (Burke et al., 1970). The folds are in some places continuous for up to 60 km (Burke et al,1970). The folds are asymmetric, and minor reverse faults were detected in a few outcrops (Burke et al., 1970). In the northern extremity of the Benue Trough, close to Biu Mountains, the fold axes appear to be parallel or in continuation with faulted blocks of shield rocks from which the sedimentary cover has been eroded away (Fig. 44). These blocks are horsts and their elongation and bordering faults are essentially parallel to the fold axes of the overlying sedimentary rocks. The Upper Benue embayment has sediments as old as Albian (Grant,1971) and an axis approximately transverse to the Benue Trough. The folding axes of the sedimentary rocks are essentially parallel to the flanks of the embayment and consequently exist at a very high angle to the main direction of the fold axes in the Benue Trough (Figs.38 and 44).

Although the folding episode in the Benue Trough has been generally referred to in the literature as a single folding event (Burke et al., 1970; Murat, 1970; Reyment, 1966; Adeleye, 1975), Nwachukwu (1972) presented evidence to suggest that separate

Cenomanian and Santonian deformations had actually taken place.

Intermediate and basic intrusions in the Benue Trough dating from Albian (or older) to Turonian have been reported in the literature (Cratchley and Jones, 1965; Reyment, 1966; Burke et al., 1970; Murat, 1970; Grant, 1971; Nwachukwu, 1972; Uzuakpunwa, 1974; Adeleye, 1975). Andesitic magmatism contemporaneous with the folding episode of the Benue Trough has been suggested by Burke for the Abakaliki region on the basis of the presence of andesites reported by Okezie (1943, in Burke et al., 1970). However, Uzuakpunwa (1974) concluded that the Abakaliki pyroclastics of an andesitic matrix were most probably related to volcanism associated with the early rifting of South America and Africa, prior to the sedimentation of Albian shales as discussed earlier.

The Niger Delta Basin is considered to occupy the area from the Benin flank to the Cameroon. An important linear high in the southwestward continuation of the Abakaliki anticlinorium is inferred to exist in the Niger Delta Basin from a detailed isopach map of the Benin formation of Miocene age (J.C. Mueller from CONOCO, personal communication). This map shows a thinner Benin formation along a stretch that extends from the Abakaliki folded belt to the littoral zone in the area of a gravity high mapped by Hospers (1965)(Fig.38). This linear high may have played a fundamental geological role in dividing the Niger Delta Basin during the pre-Benin formation time. This linear high is in line with the Jean Charcot Ridge in the continental margin that most probably separated, from the beginning of their sedimentation, two separate continental-margin sectors as discussed earlier.

The Niger delta has been prograding the littoral zone since the Early Tertiary (Murat, 1970; Hospers, 1965; Fig. 45). Sediments bypassing the subaerial part of the delta were responsible for building up the huge sedimentary wedge of the continental margin and for constructing the vast Guinea abyssal plain. The late Quaternary history of the Niger Delta has been studied in detail by Allen (1965) and its influence in the continental shelf and slope has been shown by Allen (1964).

Discussion - Because Benue Trough is located close to where the predominantly north-south trending continental margins of the South Atlantic curve sharply to an east-west trend, the trough has attracted considerable interest because its origin was thought to be fundamental to the understanding of the opening of the Atlantic Ocean.

Cratchley and Jones (1965) interpreted the Benue Trough as a rift-valley. Wright (1968) thought of the Benue Trough as a tensional graben system which originated because of the partial relief of distorting stresses that accumulated as the southern portions of Africa and South America were splitting apart. The Benue folded system was believed to have originated when the continents (Africa and South America) finally separated, and the southern half of Africa swung back therefore compressing the sediments deposited in the Benue Trough.

Burke et al. (1970) interpreted the Benue Trough as the result of tensional stresses which originated from misfit motion within the African plate. This misfit motion was caused essentially by different spreading rates of the proto-North and South Atlantic ocean floors (Le Pichon, 1968). Because of the misfit motion within the African plate, an RRR triple junction (McKenzie and Morgan, 1969) developed in the Gulf of Guinea area. One of the ridges corresponded to the site of the Benue Trough that was subjected to sea floor spreading processes. The geological setting was probably similar to the modern Red Sea - Gulf of Aden region. Oceanic crust may have formed the floor of the trough during earlier stages of trough formation. A great deal of sediments were deposited in this confined depression and prograded longitudinally southwestward along the depocenter of the basin. When the two continents finally separated, the West African shield partly swung back and probably caused an episode of consumption of the oceanic crust developed during the spreading phase of the trough. This episode was accompanied by andesitic magmatism and would also have been responsible for the assumed Santonian phase of folding. Similar to Burke et al. (1970), Grant (1971) also interpreted the Benue Trough as having originated from the Cretaceous separation of Africa and Brazil. However, Grant (1971) considered the Benue Trough as the site of an RRF triple junction (McKenzie and Morgan, 1969). Being an unstable triple junction when active, the RRF type could have explained the ephemeral nature of the spreading movements under the Benue Trough and it could have originated strains in the African plate that would be partly accomodated by elastic distortion. After the spreading had ceased under the Benue Trough, the partial recovery of the elastic distortion would have caused the Santonian and younger folding movements in the Benue Trough.

These hypothesis on the origin of the Benue Trough are based on several inconclusive arguments. The fact that the andesitic magmatism in the Abakaliki region (Burke et al., 1970) actually predates the folding episode (Uzuakpunwa, 1974) as mentioned earlier, makes

Burke et al.'s (1970) arguments of andesitic magmatism associated with an episode of crustal consumption in the Benue Trough very difficult to accept. Although hypothetical Andean-type of andesitic magmatism may explain andesitic rocks in the Benue Trough and in the Cameroon, it does not explain the association of andesites with Tertiary volcanics in the Cameroon (Lasserre, 1968; Nwachukwu, 1972; Sarcia and Sarcia, 1952), in the islands of the Gulf of Guinea (Hedberg, 1968; Mitchell-Thomes, 1970; Machens, 1973) and in the northeastern Brazil (Cobra, 1967).

Magnetic anomalies of opposite polarities running parallel to the Trough in the region of the Lower Benue depression were reported by Burke et al. (1971). These anomalies were interpreted by the same authors as probable representatives of sea-floor spreading anomalies but this interpretation is not unique.

The episode of folding interpreted as a single eventy by Wright (1968), Burke et al. (1970) and by Grant (1971) as Santonian-Maestrichtian may have constituted in fact two separate folding events of post-Albian-pre-Turonian age and of post-Turonian-pre-Maestrichtian age (Nwachukwu, 1972), as mentioned earlier. These two episodes of folding are evidence against the models of Burke et al. (1970) and Wright (1968) of the Benue Trough, because these models imply that the folding event must have occurred at the time of the final separation of Africa and Brazil and, most probably as a single event.

High Bouguer gravity-anomaly values flanked on either side by elongated negative anomalies in the Benue Trough as reported by Cratchley and Jones (1965) in the central part of the Benue Trough, are not diagnostic of the nature of the basement floor of the Benue Trough.

They were interpreted by these authors as due to the combined effects of crustal thinning,basic to intermediate igneous rocks, and possible shallow crystalline basement. Burke et al. (1970) pointed out the similarity of the gravity profile of the Benue Trough with the gravity profile of the Red Sea.

The Benue Trough is more or less parallel to the east-west continental margin sectors of the Gulf of Guinea and of the northern Brazil, and it is more or less transverse to the prevailing north-south Mesozoic structures in the South Atlantic.

Douala Basin - Douala Basin corresponds to the sedimentary embayment that is limited to the north by the Cameroon Trend and that extends as far south as 3ºN, close to the city of Kribi (Fig. 38).The basin although believed to be continuous into the Calabar flank basinal area by several earlier workers (Reyment, 1966; Reyre, 1966 and Hedberg, 1968), is here thought to be tectonically bounded by the Fernando Poo-Cameroon lineament. Douala Basin continues offshore in the continental shelf where it may contain sedimentary sequences greater than 7000m (Hedberg,1968).

The basal sediments of the Douala Basin are massive, cross-bedded sandstones and pebble conglomerates which rest unconformably on the crystalline basement and constitute the Mundeck formation (Hedberg, 1968). These sediments are continental in origin and are non-fossiliferous. Lagoonal, littoral and neritic sediments that range in age from Cenomanian (possibly Albian) to Maestrichtian (Hedberg, 1968) or from Turonian to Maestrichtian (Reyre,1966) conformably overlie the Mundeck formation. Deposition of the Mundeck formation was probably during the pre-Aptian time (Reyre, 1966). This is supported by Reyment (1956) who demonstrated that the Mundeck equivalent in the Calabar area of Nigeria is conformably overlain by marine shales of Albian age. Other evidence that indirectly supports the pre-Albian age for the basal sediments of the Douala Basin is based on the fact that recent drilling from the basin has found evaporites in the offshore region (Grunau et al., 1975). These evaporites if correlatable with the evaporites of the marginal basins in the South Atlantic (Asmus and Ponte, 1973; Leyden et al., 1976) are of Aptian age and the salt basins around Africa should include the Douala Basin as well. Because these evaporites of the South Atlantic overlie continental deposits similar to the Mundeck formation, it is very probable that, at least in the offshore portion of the basin, the basal sediments are Aptian or pre-Aptian in age as suggested by Reyre (1966) for the onshore portion of the Douala Basin. Consequently, in the offshore portion of the Douala Basin, marine sediments of Albian age probably exist.

Several wells have been drilled in the onshore portion of the Douala Basin and they all showed a quick thickening of the sedimentary succession between the boundary with the shield and the littoral zone (Hedberg, 1968; ASGA-UNESCO, 1968). A deep well situated in Lagbaba (Hedberg, 1968) bottomed in Turonian sediments at a depth of 172m (Fig. 46). In the littoral zone, at Suallaba, a 2,697m well reached Maestrichtian sediments (Hedberg, 1968; Fig. 46). Belmonte (1965, in Hedberg,1968) estimated that the Douala Basin attains a depth of 7000 to 8000m between the Sanaga and Wouri Rivers. The sediment isopachs of the basin trend parallel to the boundary with the shield (ASGA-UNESCO,1968). This boundary and the isopachs bend from north-south to northwest-southeast and this bend is also visible in the continental shelf-edge (Fig.38). The thickening of the basin towards the littoral zone and the continental margin suggests that no major barriers for the sediment progradation were present in the offshore portion of the Douala Basin.

Hedberg (1968) pointed out that wells drilled in the offshore portion of the Calabar flank basinal area did not show any evidence of thinning towards the Fernando Poo Island (one well was drilled only 12 km from the island). All the wells showed a seaward thickening of sedimentary formations. Hedberg also pointed out that Bouguer gravity anomalies as low as +41 mgals occur in the northern part of Fernando Poo Island. These values contrast with positive anomalies as high as +140 mgals in the southern portion of Fernando Poo and as high as 100 mgals for the Mount Cameroon (Fig. 46). Based on the well information and gravity data, Hedberg (1968) also suggested that the Douala Basin was continuous with the Calabar Basin. However, the Douala Basin in the offshore region contains predominantly shales (S. Whitten from Conoco, pers. comm., 1975) which contrasts markedly with the sedimentary sequence in southeastern Nigeria and off Calabar flank.This latter sequence is mostly constituted by sands that are rich in oil (S. Whitten from CONOCO, pers. comm., 1975). A seismic section along the continental shelf from the southeastern portion of the Niger delta towards the Douala Basin shows distinct acoustic stratigraphies that are bounded by shallow basement area that corresponds to the continuation of the Mount Cameroon-Fernando Poo-Annobon trend. According to

S. Whitten of CONOCO (pers.comm., 1975), the shallow basement area demonstrates that Fernando Poo-Mount Cameroon trend has formed an effective barrier since Cretaceous time and which separates the structural and stratigraphic development of Douala from the rich-in-oil Niger Delta Basin to the north. The present author fully agrees with this interpretation and suggests that the Cameroon Trend was probably a geologic structure that was formed during the early rifting history of Africa and South America and was subsequently tectonically reactivated several times as suggested by the associated magmatism and faulting in the Douala Basin (Miocene; Hedberg, 1968).

The strong gradient of isopachs in the east-west direction in the eastern part of the Douala Basin (ASGA-UNESCO, 1968) supports the existence of a probable north-south striking system of normal faults that down-faulted basement blocks towards the west. Another important trend of the Douala Basin that is shown by the shelf break is the NW-SE structural direction that is essentially transverse to the Cameroon Trend, is in the same direction as the Calabar flank and is probably represented by normal faults. This direction is also represented in the Mamfe embayment and in subsidiary faults of the Cameroon Trend. Consequently, the author considers that the Douala Basin is tectonically formed by a system of north-south, northwest - southeast and northeast - southwest faults (Fig. 42). In short, the Douala Basin appears to be a half-graben basin (open type) controlled by the Cameroon Trend and faults of NW-SE and N-S directions. The Cameroon Trend represented by the Fernando Poo-Annobon Ridge continued to be an effective barrier in the continental rise for sediments coming from the Douala Basin area.

Cameroon Trend - The Fernando Poo-Annobon Ridge is constituted by the oceanic islands of Fernando Poo, Príncipe, São Thomé and Annobon, seamounts and buried basement highs. The ridge has a continuous northeast-southwest strike and islands like São Thomé and Fernando Poo have their long axes oriented in that direction (Figs.7 and 31). The ridge extends into the contiment as a line of volcanoes and volcanic rocks in the Cameroon region (Fig. 38). The lineament corresponds to the so-called Cameroon Volcanic Line (Furon, 1960) or Cameroon Trend (Hedberg,1968).

These islands are volcanic in nature and of very recent age. Fernando Poo Islands is as old as 1.1 m.y. (Hedberg, 1968) and its volcanic rocks are essentially basaltic in composition (Hedberg,1968; Mitchell-Thomes, 1970; Baker, 1973). Princípe Island is constituted by volcanic rocks of an alkalic-calcic and an alkaline series (Mitchell-Thomes, 1970) that are probably of a pre-Miocene age (Hedberg , 1968). Oil seepages from fractures in the volcanics have been reported in two localities in Princípe Island (Hedberg, 1968; Mitchell-Thomes, 1970; Baker,1973). São Thomé Island is mostly constituted by olivine basalts, phonolites and subordinate trachytes and andesites (Mitchell Thomes, 1970; Cotello Neiva, 1956 in Hedberg, 1968). Radiometric datings in São Thomé Island gave an age range of 0.1 to 3.0 m.y. (Hedberg, 1968) and an albitized trachyte was dated at 15.7 $\pm$ 0.8 m.y. (Grunau et al., 1975). Hedberg (1968) has mapped an occurrence of quartziferous sandstones and shales (1 km^2) that was quite surprising because of the oceanic setting of São Thomé (Fig. 31). These sediments correspond to two sedimentary units. The lower unit with rare but not diagnostic fossils, and the upper unit with radiolarians and foraminifera. Grunau et al. (1975) considered the lower unit as possibly of Cretaceous age, and the upper unit as post-Paleocene age. Hedberg (1968) has concluded from primary structures and textures that the lower unit was of shallow-water origin, and that its heavy-mineral assemblage and primary structures were very similar to the ones of a Cretaceous basal sandstone of Gabon and Cameroon. J.Houbolt (pers. comm., 1975) pointed out the similatities in sedimentary primary structures of the sandstones and shales of São Thomē Island with sands and muds recovered in the floor of and in the close vicinity of the Princīpe Channel and other canyons in the Gulf of Guinea (Houbolt, 1973; Fig. 31). Consequently, on the basis that indeed the sedimentary formation of São Thomē Island represents continental rise or even abyssal sediments, the island seems not to be a volcanic pile only but rather a tectonic block of continental rise that was uplifted after it was partially covered by terrigenous sediments. Widespread magmatism along fractures eventually covered this block and thus volcanics constitute the bulk of the surface of the island.

Annobon Island is entirely constituted by basaltic rocks that are characterized by the abundance of olivine phenocrysts and scarcity of plagioclases. These rocks are probably Miocene or younger in age (Hedberg, 1968).

Basement highs between the islands are generally detected on seismic profiles in regions of the continental rise where the basement is covered by sediments (Fig. 4, profiles 1-4). Several seamounts are also present between islands. Southwestward from Annobon Island as far as 3ºS there are seamounts that are clearly in the continuation of the Fernando Poo-Annobon Ridge (Fig. 31). Close to 4ºS a prominent seamount is the only topographic evidence of the possible continuation of the ridge to the south. Southwestward from 4ºS data are sparse and the high relief of the Guinea Ridge prevents a clear understanding of the relationship of the Fernando Poo-Annobon ridge to the surrounding topography.

However, seismic profiles demonstrate the presence of a prominent fracture zone in the vicinity of the topographic "end" of the Fernando Poo-Annobon Ridge (Fig. 4, profiles 1, 14, 15 and 20). This fracture zone is thought to be the Ascension Fracture Zone (Emery et al., 1975). Because the most important fracture zones in the equatorial Atlantic continue as prominent marginal ridges on both sides of the ocean, and because of the close proximity of the Ascension Fracture Zone to the southernmost extension of the Fernando Poo-Annobon Ridge, the ridge is believed to lie along the continuation of the Ascension Fracture Zone in the continental rise of Gulf of Guinea.

The Fernando Poo-Annobon Ridge is linked to the continental Cameroon Trend by means of an intervening high in the continental shelf between Fernando Poo Island and Mount Cameroon, as discussed earlier (Fig. 38). This high was created during the early rifting history of Africa and Brazil and probably acted as a geomorphological barrier to prevent the deposition of Aptian salt to the north of Douala Basin. Although salt diapirs have been reported in the continental margin off Niger Delta by Mascle et al. (1973) they actually correspond to diapirs originated from layers younger than Aptian and probably are mud diapirs (Delteil et al., 1975).

In the continent, a line of volcanoes and volcanic rocks in the Cameroon extends deeply inland, northeastward from Mount Cameroon (Fig. 38). These volcanics are alkaline in nature and, as far north as Mt. Kambo, they are continuous and essentially located in a northeast-southwest direction, the same trend that characterizes the Annobon-Fernando Poo-Mount Cameroon lineament. Northeastward from Mt. Kambo, these basic volcanics are widespread in area in the Adamoua Plateau, and are present as far east as the region around Ngaoundéré (Sarcia and Sarcia, 1972; Fig. 38). Nevertheless, the northeast-southwest trend of the Cameroon Volcanic Line to the north of Mt. Kambo is in line with separated occurrences of basic volcanics that extend as far north as Mandaba Mountains and also in line with remarkably linear occurrences of Tertiary granites dated by Lasserre (1966) (Fig. 40). Consequently, the Cameroon Trend is extended as far north as 11°N in the region of Mandaba Mountains (Fig. 40). How far northeastward the Cameroon Trend is supposed to extend from Mandaba Mountains is speculative. More geophysical studies are necessary in the Tchad Basin to define major basement structures.The inference that the Cameroon Trend extends into the Tibesti Mountains (Furon, 1960)is based on the facts that the Tibesti Mountains are in the northeastward continuation of the strike of the Trend and the magmatism of the Cameroon and Tibesti is similar. Although the magmatic similarity is remarkable between the two regions, this fact by itself does not explain the seemingly similar magmatism in the region of the Jos Plateau and Aïr that evidently does not have linear connections with the Cameroon Trend (Fig. 41).

Southwestward, the Cameroon Trend through the islands of the

Gulf of Guinea was inferred to extend into the island of Saint-Helena in the South Atlantic by Furon (1960). The present author believes that the coincidence of Saint-Helena Island in lying in the direct prolongation of the strike of the Trend is purely coincidental.

Furon (1960) following Gèze (1943) interpreted the Cameroon Trend as a rift-valley in which its axial portion constitutes a tectonic line. This tectonic line is formed by several parallel fractures that constituted conduits for magma outpourings that formed the volcanic massifs of Cameroon, Manengouba and Bambouto (Fig. 47). Machens (1973) interpreted the Cameroon Volcanic Line as a deep, median fracture line in a particularly strong warped block and tended to favor the origin of the graben-like depression as due to the sinking of a central strip as a result of the emptying of the magma chamber. Reyre (1966) showed the Cameroon and Adamaoua uplift as a region of high altitudes in the Cameroon (higher than 1500m) that was bounded to the south by a continuous fault or flexure zone. This latter author pointed out that the weak seismicity of the volcanic lineament of Cameroon rules out the possibility that the lineament represents a strike-slip fault. Reyre (1966) also pointed out the very deep origin of the Cameroon Trend due to the fact that the lineament is continuous from the continent to the ocean and consequently is independent from the Mohorovicic discontinuity. Hedberg (1968) concluded that there is considerable evidence of uplift along the length of the Trend. Magmatism along the Cameroon Trend has been occurring since at least the Cretaceous and demonstrated by basal basaltic tuffs that contain plant remains believed to be representatives of the Upper Cretaceous (Senonian) (Guillemain in Furon, 1960; Machens, 1973).

Mount Cameroon is an active volcano and constitutes a typical strato-volcano that is at least 4000m high (Furon, 1960). The mountain is essentially aligned in a northeast-southwest direction and contains several craters also aligned in that direction (Hedberg, 1968). Its last eruption was in 1954 as reported by De Swardt (1954; Fig.48).Other eruptions occurred in 1909 and 1922.

Three phases of volcanism in the Cameroon have taken place : a) the first phase was characterized by basaltic and andesitic flows (late Cretaceous-early Tertiary); b) the second was mostly constituted by phonolites and trachytes (Miocene); c) the third phase of Plio-Pleistocene age renewed the basaltic volcanicity (Gèze, 1943). Not included in these phases are the Tertiary granites reported by Lasserre (1966). These granites range in age from 54 ± 22 m.y. to 33 ± 5 m.y. and are intruded in the shield as ring-like structures for their outcrop area. Basic volcanism in Mount Koupé as old as 35 m.y.,as reported by Tchoua (1970), suggests the contemporaneity of basic and acidic volcanism in the Cameroon Trend (Fig. 38).

The Adamoua Plateau (Sarcia and Sarcia, 1952) is a vast shield region with widespread volcanic, hypabissal and plùtonic rocks of Mesozoic and younger age. A very marked lineament of Precambrian age,the Ngaoundéré fault zone, corresponds to a complex zone of fracturing that has been reactivated in the Mesozoic and Tertiary as seen by the Sud-Adamoua Graben (Le Marechal and Vincent, 1970) and by volcanic rocks (Fig. 38). The Ngaoundéré fault zone has an east-northeast-west-southwest direction with several branches of mylonitization zones that merge west-southwestward into a single zone of mylonitization (ASGA-UNESCO, 1968). This fault zone merges into the Cameroon Trend volcanics (Fig. 38). The Ngaoundéré fault zone is probably buried by the volcanics of the Cameroon Trend toward the southwest and may extend into the littoral region of the Gulf of Guinea. Thus, the Cameroon Trend is believed to represent a structure originated during the early rifting history of Africa and Brazil in consequence of the reactivation of an important Precambrian weakness zone, the Ngaoundéré fault zone. The rectivation of this structure developed an important fracture zone in mid-ocean (Ascension Fracture Zone). The periodicity and the recent magmatism of the Cameroon Trend suggest that the Trend is still an active tectonic line that disregards the ocean-continent boundary. The contemporaneity of the magmatism along the lineament rules out a hot spot origin. Most probably the Cameroon Trend in the ocean and in the continent corresponds to a complex of horst and graben features with associated fractures that constituted easy magma conduits for a variety of magma types (basaltic, alkaline, calc-alkaline and granitic).

Sediment Dispersal in the Gulf of Guinea

The tectonic framework of the bordering continent, the original offsets of rifting and subsequent spreading, and the fracture zones controlled the way the terrigenous sediments built the continental margin of the Gulf of Guinea. The bulk of terrigenous sediments prograded seaward from the marginal basins of half-graben type. Terrigenous sediments coming from continental areas associated with marginal basins of graben type prograded longitudinally along the axis of the basins and were barred to the south by prominent marginal ridges that lie along the trends of the fracture zones. Because of these ridges, the sediments ponded behind them and in some cases marginal plateaus such as the Liberia and the Ivory Coast plateaus have developed in this manner. The sediments confined between two marginal ridges prograde farther seaward toward the flanks of the Mid-Atlantic Ridge than in areas off the half-graben basins. Prominent troughs associated with fracture zones were filled with terrigenous sediments coming from the continental rise and constituted tongue-like abyssal plains (Fig.49).

The several levels of the crustal segments bounded by Romanche and Chain fracture zones (deeper), by Chain and Jean Charcot fracture zones (shallower) conditioned the evolution of the sedimentary progradation off the Niger Delta. Because the segment bounded by Romanche and Chain fracture zones was deeper, it was the first to receive sediments from the Anambra Basin area and this can be shown by the deepest reflectors in the area that are bounded to the south by the Chain Fracture Zone (Fig. 4; profiles 6, 8 and 9). Only after sill depths had been crossed, the terrigenous sediments prograded over the crustal segment between Chain and Jean Charcot fracture zones (Fig.4; profiles 6, 8-10, 12 and 13). This time may coincide with the development of the wide Guinea abyssal plain that continuously prograded towards the flanks of the Mid-Atlantic Ridge, with the progradation of the continental rise. The Jean Charcot Fracture Zone, that barred to the south the sedimentation in the onshore-Anambra Basin,channelled terrigenous sediments to the continental margin to the north of it. This probably happened up to the Miocene, when the high area to the south of the Anambra Basin (Jean Charcot high) was overcome by the sedimentation. Consequently, the continental margin barrier of the Jean Charcot Ridge was also probably overcome by sediments coming from the modern Niger delta. These sediments were barred to the south by the Fernando Poo-Annobon Ridge but still went towards the Guinea abyssal plain flanking irregularly the high grounds of the Guinea Ridge (Fig. 49).

CONTINENTAL MARGIN OF NORTHERN BRAZIL

In this section, the continental margin sectors of the Brazilian side at the equatorial Atlantic are discussed in a detailed manner similar to the discussion of the African side. At the end of the descriptive part of each sector a comparison of the geological and tectonic features of the correspondent parts of Brazil and Africa is made. The shield geology of northern Brazil is summarized in detail in the appendix.

Amapá Continental Margin Sector

Physiography

The continental margin to the north of the northern east-west segment of the North Brazilian Ridge presents a morphology that contrasts with the continental margin to the south. After bulging around the Amazon cone (Heezen and Tharp, 1961; Damuth, 1973; Damuth and Kumar, 1975), the bathymetric contours of depths greater than 3750m, shift appreciably to the east, around the northern east-west segment of the North Brazilian Ridge (Fig. 50). This shift in bathymetric curves demonstrates the importance of the North Brazilian Ridge as a geomorphologic boundary for the continental margin off Amapá, Pará and Maranhão states of Brazil. Consequently, that segment of the North Brazilian Ridge is considered to be the southern limit of the Amapá continental margin sector.

The widest part of the continental shelf in northern Brazil lies off the Amazon River, in the Amapá continental margin sector(Fig. 51, Zembruscki et al., 1971). The Amazon canyon, which at present is an inactive feature on the outer continental shelf (Zembruscki et al., 1971; Gorini et al., 1971), was the main feeder canyon of terrigenous sediments that formed the Amazon submarine cone (Fig. 51, Damuth and Kumar,1975). Towards the east, the Amazon cone grades laterally into the Ceará abyssal plain. The Ceará abyssal plain extends eastward to the flanks of the Mid-Atlantic Ridge, is limited to the north by the morphology of the Ceará Rise (Damuth, 1973; Embley and Hayes, 1972), and to the south by the North Brazilian Ridge (Fig. 51).

Seamounts interpreted by Hayes and Ewing (1970) as the westernmost extension of the North Brazilian Ridge are present in the continental slope and upper rise off Belém (Fig. 50). These seamounts are located to the south of the sedimentary bulge of the Amazon cone.

The northern east-west segment of the North Brazilian Ridge was tentatively traced into the southern transverse ridge of the Saint Paul's Fracture Zone, as discussed earlier. The Amapá continental margin sector corresponds to the Liberia sector in Africa (Fig. 3). Both are bounded to the south by prominent marginal ridges that correspond to transverse ridges of Saint Paul's Fracture Zone (Fig. 7).

General Geology and Tectonic Framework

The geologic structures present in the Amazon continental shelf and in the region of the bordering continent will be discussed in detail.

Marajó Basin - The Marajó Basin is separated from the Amazon Basin by the Gurupá horst and from the Parnaíba Basin by the Tocantins arch (Fig. 52; Aguiar et al., 1969; Rezende and Ferradaes, 1971).

The Marajó Basin is essentially a complex system of grabens that extend into the continental shelf off Amazon River mouth (Fig. 53). The graben system starts deep in the Parnaiba Basin, follows a northwest direction up to the Marajó Island (Grajaú, Cametá and Limoeiro grabens) and from there, it continues in a northeast trend up to the inner Amazon continental shelf (Mexiana graben). At this point the graben system bifurcates into a north-northeast branch (west-Mexiana graben) and into a northeast branch (east-Mexiana graben)(Fig.53) (Rezende and Ferradaes,1971).

The Marajó Basin in the continental shelf is bounded to the east by the Pará-Maranhão platform and to the west by the Amapá platform (Fig. 53). The Ferrer-Urbano Santos Arch separates the Marajó graben system from the São Luiz-Bragança-Vizeu basins (Rezende and Pamplona, 1970). The Marajó Basin is essentially Mesozoic in age and the basal sediments of Upper Jurassic (?)-Cretaceous age rest either on Paleozoic erosional remnants or on shield rocks. Basaltic magmatism is represented by dykes intruded in Paleozoic sediments or in the Precambian basement (Rezende and Ferradaes, 1971). This magmatic event was attributed by Rezende and Ferradaes (1971) to be early to late Jurassic in age. The most important sedimentary accumulations are in the Limoeiro and Mexiana grabens and these may be as thick as 6 km. Farther seaward, in the continental shelf, the sediments are generally thicker and may reach more than 10 km according to estimates based on aeromagnetic data in the area of the so called Amazon low (Fig. 53; Rezende and Ferradaes, 1971; Ponte and Asmus, 1976).

The Cretaceous sedimentation in the graben system was domi nantly continental in the Grajaú, Cametá and Limoeiro grabens; transitional from continental to marine in the Mexiana graben; and,entirely marine in the west-Mexiana and east-Mexiana grabens (Fig. 53). The Mesozoic sedimentation in the Marajó Basin started in the pre-Albian and possibly in the Jurassic (Aguiar et al., 1969). Albian sediments were found only in two wells in the Mexiana graben whereas upper Cretaceous sedimentation (Maestrichtian to Campanian) has been confirmed by paleontological determinations in most of the area of the basin (Aguiar et al., 1969).

Amapá and Pará-Maranhão platforms - The Amapá and Pará-Maranhão platforms are distinguished from the basinal areas by the thickness of their sedimentary cover. As defined by Rezende and Ferradaes (1971) , the two platforms constitute areas of not more than 2000m of sediments that include several small centers of accumulation of sediments of Cretaceous or Jurassic-Triassic (?) age (Fig. 53). Because the Tertiary sedimentary cover (1000-1500m of thickness) is not tectonically influenced by the older centers of deposition, the platform areas are thought to have been very stable (Rezende and Ferradaes , 1971). The platform areas are separated from the graben basinal areas by marginal normal faults with substantial throws (> 2000m) (Figs. 52 and 53). The fault zone is generally marked by down-to-the-basin faulted blacks (Ponte and Asmus, 1976).

Gurupá Arch - The Gurupá arch is a horst bounded by high-angle normal faults on the west and by a step-like fault system on the east (Fig. 54, section C-D; Bigarella, 1973). The Gurupá arch has been a high area since the late Carboniferous. Erosion in the entire Marajó area, as a result of a regional uplift, occurred up to the early Cretaceous time (Aguiar et al., 1969; Bigarella, 1973). Prior to the regional uplift of the Marajó area, the Parnaiba and Amazon Basins were interconnected. With the faulting in the Jurassic (?)-early Cretaceous that resulted in the development of the Marajó graben system, the Gurupá arch was individualized as a horst (Aguiar et al., 1969; Rezende and Ferradaes, 1971). To the north of the Gurupá horst, the Limoeiro graben has a gravity signature that corresponds to a high Bouguer anomaly value flanked by two negative values. The high Bouguer gravity values coincide with the axis of the highest sedimentary accumulation in the graben (Aguiar et al., 1969; Rezende and Ferradaes, 1971 and Ponte and Asmus, 1976). The inversion of the gravimetric signature in a center of accumulation led Aguiar et al., (1969) and Rezende and Ferradaes (1971) to imply the presence of more dense basic rocks that were intruded in the basement and even in the basal portions of the basin, in the early Cretaceous time (Fig. 54, Section E-F).

Amazon sedimentary low and the Amazon Cone - The thickest accumulation of sediments in the Amazon continental shelf corresponds to the so-called Amazon low (Figs. 53 and 54,sections A-B, G-H and I-J). Sediment thicknesses between 12 and 15 km have been estimated in the low by aeromagnetic and gravity data (Rezende and Ferradaes, 1971; Rezende, 1971) although Houtz et al., (in press) indicated no more than 6-8 km. Most of these sediments were deposited from the Paleocene-Eocene up to the Recent. In the southern border of the Amazon low,structures such as growth faults and roll-over anticlines have been suggested on the basis of seismic profiles (Rezende and Ferradaes, 1971). In the northern flank of the Amazon sedimentary low, the existence of structural highs was inferred from seismic data, and Rezende and Ferradaes (1971) tentatively correlated them with the westward continuation of the North Brazilian Ridge in the continental shelf. The axis of the west-Mexiana graben divides the Amazon low symmetrically. The shape of the sedimentary low area, as shown by Rezende and Ferradaes (1971), Rezende (1971), and Ponte and Asmus, (1976) denotes a parallelism of the northern part of the low with the marginal fault of the Amapá platform whereas the southern part of the Amazon sedimentary low has essentially an east-west orientation (Fig. 53). Seaward from the Amazon low the thick Amazon cone probably contains up to 13 km of sediments in depths shallower than 3900m (Fig. 54, section A-B; Edgar and Ewing,1968; Damuth and Kumar,1975). Gravity data from Cochran (1973) also indicated a considerable thickness of 11 to 12 km for the upper Amazon cone. The minimum thickness of sediments on the upper cone is 9 to 11 km (Damuth and Kumar,1975).

The axis of the west-Mexiana graben also coincides with the location of the prominent Amazon canyon in the continental shelf. The canyon occupies a central position in relation to the Amazon cone (Fig. 51). Several buried paleo-canyons probably the predecessors of the Amazon canyon, also lie in the axis of the west-Mexiana graben in the continental shelf (Rezende,1971; Rezende and Ferradaes, 1971).These paleo-canyons were probably the main feeders of several prograding wedges of sediments that ultimately constituted the present-day Amazon cone (Rezende and Ferradaes, 1971). Consequently, the Amazon cone deposition was most probably tectonically controlled by the western Mexiana graben, as proposed by Rezende and Ferradaes (1971).

The existence of possible sedimentary barriers in the upper Amazon cone was suggested by seismic surveys carried out by Petrobras in the area. Complex structures associated with intensive deformation of sediments of Oligocene age and local uplift of Miocene sediments were reported by Rezende and Ferradaes (1971) (Fig.54, sections A-B and C-D). These same authors implied that these complex structures were the result of tectonism in high areas that were associated with the possible westward prolongation of the North Brazilian Ridge as also suggested by Kumar et al., (1974). However, because of the complexity of the structures present in the Amazon cone area it is very difficult to discriminate the structures originated from tectonism of the basement from structures associated with compaction and progradation of a sequence deposited in response to a large sedimentary accumulation.

Discussion

In the Mesozoic, probably in the Jurassic-early Cretaceous time, the present-day area of the Amazon River mouth underwent a series of tectonic events that were effective up to the Tertiary (Rezende and Ferradaes, 1971). These tectonic events consisted essentially of the development of the Marajó graben system. This graben system consisted of several aligned grabens that received a considerable amount of terrigenous sediments from the Parnaíba Basin or from the Guaporé Platform to the south (Fig. 52; Rezende and Ferradaes, 1971). Only after the Oligocene did the Amazon River contribute sediments to the graben system (Rezende and Ferradaes, 1971). The Marajó Basin was developed in a tensional tectonic setting, in a region that was submitted to regional uplift after the Upper Carboniferous (Bigarella, 1973). As a result of erosion, few remnant Paleozoic sedimentary layers are preserved underneath the Cretaceous rocks. These Paleozoic rock remnants prove that the Amazon Basin extended towards the east and was probably a continuous basin prior to the development of the Gurupá horst (Rezende and Ferradaes, 1971).

The northwestern continuation of the west-Mexiana graben goes into one of the arms of a deep fault that characterize the western boundary of the Amazon sedimentary low of Rezende and Ferradaes (1971), Rezende (1971) and Ponte and Asmus (1976). When the two continents are considered together (Fig. 55) this arm of the fault is essentially parallel to the trend of the littoral of Liberia. This trend, as discussed before, represents the direction of rifting in Liberia (Fig. 55). Consequently, the Marajó system of grabens may represent structures originated by the propagation of the rifting direction of Liberia towards the continental area of Amapá and Pará, in Brazil. The deflection of the axis of the Mexiana graben towards the Limoeiro graben is probably associated with deep structures of the Gurupá Arch and with reactivated deep structures in the Parnaíba Basin.

The general configuration of the Amazon sedimentary low coincides with the curvature of the continental slope around the Grand Cess Ridge and with the rifting direction of Liberia (Fig.55). The east-west strike of the northern transverse ridge of the Saint Paul's Fracture Zone (westernmost seamounts of the North Brazilian Ridge), if followed westward, coincides with the region immediately to the south of the Amazon sedimentary low in the continental shelf.

When the two continents are considered together, this region coincides with the area where the Grand Cess Ridge (northernmost transverse ridge of Saint Paul's Fracture Zone in the African side) overlaps (Figs. 53 and 55). Consequently, the borders of the Amazon sedimentary low may well represent the exact point where that part of Africa and Brazil split apart from each other. This is in agreement with the hypothetical sialic crust limit of Rezende (1971) that was placed at the borders of the Amazon sedimentary low. The geological fit of the two continents (Fig. 55) also shows the correspondence of the radiometric boundaries of Pan-African age in Liberia and the one that borders Pan-African age rocks to the south of the São Luis cratonic area, in Brazil (Fig. 52).

The tectonic direction of the east-Mexiana graben has no counterpart in the African side but it is significant that the graben follows the curvilinear trend of the marginal fault of the Pará-Maranhão platform. When the two continents are considered together, it is noticed that the marginal fault of the Pará-Maranhão platform, when followed eastward, becomes parallel to the marginal fault of the Ivory Coast Basin (Fig. 55). The prolongation of the northern east-west segment of the North Brazilian Ridge (one of the transverse ridges of Saint Paul's Fracture Zone) coincides with the western extremity of the marginal fault of Pará-Maranhão platform (Fig. 53). Similarly, in the African side, the Saint Paul Ridge (one of the transverse ridges of Saint Paul's Fracture Zone) is in strike with the shelf-edge off Cape Palmas and western Ivory Coast and, consequently, is parallel to the strike of the western end of the Ivory Coast marginal fault, as discussed in Part I.

The rifting direction of Liberia was probably originated at the time of the rifting in North America, Northwestern Africa and Western Europe and consequently predates the rifting in the South Atlantic. Because the Marajó graben system is believed to have originated by the propagation of the Liberian direction of rifting, it is possible that this basinal system may have originated earlier (Jurassic or even Triassic ?) than the basins in the South Atlantic and in the equatorial Atlantic (late Jurassic (?) - arly Cretaceous).

Because fracture zones in mid-ocean are traced as highs in the continental shelves in the equatorial Atlantic, a similar situation may exist in the Amazon continental shelf that would be comparable to the continental margin of the Gulf of Guinea as discussed earlier in this work. It is expected that the northernmost transverse ridge of the Saint Paul's Fracture Zone corresponds to a deeply buried high that bounded to the south the approximately east-west arm of the fault that delineates the Amazon sedimentary low (Fig. 56). The southernmost transverse ridge of the Saint Paul's Fracture Zone (North Brazilian Ridge) is in strike with the western portion of the marginal fault of the Pará-Maranhão platform. The trace of this fault apparently continues eastward in the continental shelf as seaward-dipping faults with very similar configuration to the marginal fault of the Ivory Coast. The geological role of the intermediate transverse ridge of the Saint Paul's Fracture Zone, that on the African side is represented by a continuous ridge up to the continental shelf (Cape Palmas Ridge) and probably mingles with the southern transverse ridge of the same fracture zone, is not certain on the Amazon continental shelf.

The Amazon cone had its sedimentary evolution controlled by the west-Mexiana graben as discussed earlier. The probable high associated with the northern transverse ridge of Saint Paul's Fracture Zone may have acted as a complete or partial sedimentary barrier that in the past, prevented the seaward continental margin progradation (Fig. 56; Kumar et al., 1976). Although probably effective during the early geological development of the margin, the barrier was probably overwhelmed by a large amount of terrigenous sediments associated with the building up of the Amazon Cone, in the late Tertiary (Fig.56) (Damuth and Kumar,1975).

The axis of the Amazon Cone is approximately symmetrical to the deeply buried boundaries of the Amazon sedimentary low. Although controlled by the west-Mexiana graben, the continuous sedimentary progradation of the cone took place approximately at the intersection of the north-south and east-west arms of the above mentioned deeply buried boundaries, as seen by the location of the Amazon canyon and several buried paleocanyons.

PARÁ-MARANHÃO CONTINENTAL MARGIN SECTOR

Physiography

The Pará-Maranhão sector is bounded to the north by the Saint Paul's Fracture Zone which, in turn, connects with the northern most east-west segment of the North Brazilian Ridge. To the south, this sector is bounded by Romanche Fracture Zone which, in turn, connects with the southern east-west segment of the North Brazilian Ridge (Fig. 50). Near the margin, the Pará-Maranhão continental margin sector is limited to the south by the continental slope off eastern Maranhão and Piauí states and by the steep slope of the Ceará Terrace.

The continental shelf of the Pará-Maranhão continental margin sector widens considerably towards the west. The shelf-edge off northwest Ceará, Piauí and eastern Maranhão trends east-west and curves appreciably to northwest off the city of São Luís (Fig. 50). The continental slope is steepest off northwestern Ceará and Piauí states where it is east-west oriented and aligns with the slope of the Ceará Terrace which lies further to the east. The bathymetric curves in the continental rise of the Pará-Maranhão sector show a general southwest northeast gradient.

General Geology and Tectonic Framework

The offshore-Barreirinhas-Piauí Basin constitutes a very linear basinal system in the continental shelf of northwest Ceará, Piauí and eastern Maranhão (Fig. 57). The offshore-Barreirinhas Basin is the seaward continuation of the Barreirinhas and São Luis basins in the littoral and in the continent. São Luís basin is separated from the shallower Bragança-Vizeu Basin to the west by an intervening cratonic area (Fig. 57). Limiting to the south the Bragança-Vizeu, São Luís, Barreirinhas and Piauí basins and separating them from the Paleozoic Parnaíba Basin, the Ferrer-Urbano Santos Arch is one of the main tectonic features of the region (Fig.57; Mesner and Wooldridge , 1964; Rezende and Pamplona, 1970).

The Pará-Maranhão platform in the continental shelf of Pará and Maranhão constitutes the western tectonic boundary of this continental-margin sector. The continental shelf edge bordering the Pará-Maranhão platform parallels a marginal fault that separates the platform area from a deep accumulation region (more than 4km of sediments) in the continental shelf that grades laterally into the sedimentary prism of the continental rise of the Pará-Maranhão continental margin sector (Fig. 52). I have named the sedimentary basin that encompasses the continental rise and part of the continental shelf adjacent to the marginal fault of the Pará-Maranhão platform, the Pará-Maranhão Basin.

Barreirinhas Basin - The Barreirinhas Basin is located in the littoral zone of Maranhão State and is separated from the São Luís Basin to the west by the Rosário horst (Fig. 57; Mesner and Wooldridge, 1964 ; Pamplona, 1969; Rezende and Pamplona, 1970; Miura and Barboza,1972) . The basin extends offshore where it is in line with the Piauí Basin, off Piauí state, to the east (Fig. 57; Miura and Barboza, 1972; Asmus and Ponte, 1973; Ponte and Asmus, 1976).

Onshore Barreirinhas Basin is a fault-bounded basin in which several parallel and subparallel normal faults mark the transition from the Ferrer-Urbano Santos Arch to the deep basinal area in the north (Fig. 58). In the deep basinal area more than 6000m of sediments accumulated since the early Cretaceous. In the tectonic transition from the Ferrer-Urbano Santos Arch to the depocenter of the basin,platform areas, bounded by normal faults, individualized local sedimentary facies as shown by drilling in the onshore Barreirinhas Basin (Fig. 58; Pamplona, 1969; Carozzi et al., 1973). The depocenter of the Barreirinhas Basin in the offshore region is essentially east-west and is in line with Piauī Basin to the east (Fig. 57). The deep basinal area with sediments thicker than 6 km is separated from the continental rise sedimentary prism by an east-west oriented high area known as the Atlântico High (fig. 57; Miura and Barbosa, 1972).

The basal sediments of Barreirinhas Basin correspond to the Paleozoic sequence of Parnaiba Basin (Fig. 59; Bigarella, 1973; Rezende and Pamplona, 1970). The Mesozoic sedimentation started in the early Albian and lies either on Paleozoic sediments or directly over shield rocks. The Cretaceous sedimentation corresponds to two main sedimentary groups (Pamplona, 1969). The oldest is called Canarias Group and underlies the younger Caju Group. The Canarias Group is essentially constituted by terrigenous sediments largely deposited in a deltaic environment and range in age from lower Albian to upper Albian - Cenomanian based on pollen stratigraphy (Pamplona, 1969).

The basal sediments of the Canarias Group consist of a polymictic conglomerate interbedded with sandstones and shales. The conglomerate very often contains pebbles of diabase, basalt and shield rocks. Pollen stratigraphy of the Canarias Group demonstrated that sedimentary facies are time transgressive and represent a prograding sedimentary environment (Rezende and Pamplona, 1970),

The source area for the sediments of the basal Canarias Group was essentially from the west (Pará-Maranhão platform), probably coming from the area of São Luís Basin (Rezende and Pamplona, 1970). The Ferrer-Urbano Santos Arch was an effective source area for terrigenous sediments only at the time of the basal sedimentation of the Canarias Group. This is demonstrated by local fan-like deposits which laterally grade into and are subsequently buried by deltaic sediments along the extent of the Arch in the basin (Rezende and Pamplona, 1970; Pamplona, 1969). The sedimentary facies changes of the Canarias Group were, from the west to east, coarse to fine, and shallower to deeper environments (Rezende and Pamplona, 1970; Pamplona, 1969; Miura and Barboza,1972).

The end of the Canarias Group deposition is marked by shales with increasing percentage of $CaCO_3$ that grade upward into the predominantly calcareous Caju Group. The Caju Group consists of three formations that are transitional or disconformable in relation to each other.These three formations are constituted by marls, calcilutites , calcarenites,shales and minor quartz-sandstones.On the basis of planktonic and benthonic marine fauna, their ages range from late Albian to Santonian (Pamplona, 1969).

A Campanian-Oligocene sequence of sediments lies unconformably on the Caju Group in the continental shelf (Miura and Barboza,1972). This sequence consists of shales with minor quartziferous sandstone intercalations and is disconformably overlain by the so-called Miocene

Pirabas-Tibau sequence (Miura and Barboza, 1972). In the onshore portion of the Barreirinhas Basin, the Caju Group is directly overlain by the Pirabas-Tibau sequence. The Miocene Pirabas - Tibau sequence is subsequently overlain by the Barreiras Group of Miocene or younger age, at least in the area close to the littoral.

Piauí Basin - The basin is located off the Parnaíba platform that is a local name for the eastward extent of the Ferrer-Urbano Santos Arch (Fig. 57). Its basal sequence corresponds to very coarse terrigenous sediments that become finer northward (Miura and Barboza,1972). This indicates that these sediments probably correspond to the time of the basal sedimentation of the Canarias Group in the Barreirinhas Basin, the Ferrer-Urbano Santos Arch was also a local source of terrigenous sediments for the Piauí Basin as well.

Piauí Basin has the greatest sedimentary accumulation of the Cretaceous east-west basinal system of the Pará-Maranhão continental margin sector and is linked with the Ceará Basin to the east by an intervening graben that is bounded to the north by the easternmost extension of the Atlântico High (Fig. 57).

São Luís Basin - The São Luís Basin is structurally linked with the Barreirinhas Basin but an intervening horst structure prevented open communication between the two basins during the deposition of the Caju Group (carbonates) (Fig. 57; Rezende and Pamplona, 1970; Asmus and Ponte, 1973).

São Luís Basin is mostly filled with continental sediments of the Itapecurú formation (Rezende and Pamplona, 1970). The total thickness of sediments in the basin is 4500m (Mesner and Wooldridge ,

1964) and the Itapecuru formation is up to 2000m thick (Rezende and Pamplona, 1970). In most of the area, the Itapecurū formation rests on basement rocks of the São Luīs cratonic area. In other areas, Cretaceous rocks rest on Paleozoic sediments (Mesner and Wooldridge,1964). Rezende and Pamplona (1970) also show the Itapecuru formation resting on a gypsiferous formation (Codō formation of an Albian-Cenomanian Age) which is possibly contemporaneous to the upper part of the Canarias deposition. Because of that stratigraphic relationship, the Itapecurū formation was interpreted to be younger than the Canarias Group and contemporaneous to the Caju Group Deposition (Upper Albian to Santonian) of Barreirinhas Basin. Consequently, the area of the São Luīs Basin was interpreted not as a basin but as a high region that was the probable source for terrigenous sediments during the time of deposition of the Canarias Group (lower Albian-upper Albian) (Rezende and Pamplona, 1970). However, if Itapecurū formation is time transgressive, it could have been deposited since the early Albian or older time and only had covered the Codō formation at the end of the Canarias Group sedimentation. Therefore, São Luīs Basin would have been formed contemporaneously with the Barreirinhas-Piauī Basin.

More recently, Ponte and Asmus (1976) confirmed that São Luīs Basin does contain Albian and possibly even older sediments (Neocomian (?)). The Itapecurū formation was apparently contemporaneous with the Canarias and the Caju Group deposition in the Barreirinhas Basin (Ponte and Asmus, 1976; Rezende and Pamplona, 1970). The basal sedimentation of the Itapecurū formation may have prograded from the São Luīs Basin to the Barreirinhas Basin and may have contributed extensively to the deltaic complex of the Canarias Group. Consequently, the Rosário horst may represent a later feature.

The establishment of an Albian or older age for the origin of the São Luís Basin emphasizes the contemporaneity in origin of the Piauí-Barreirinhas and São Luís basinal system and consequently of the Ferrer-Urbano Santos Arch. The relative position of the arch and the basinal system probably did not change appreciably through the evolution of the basins.

São Luís Basin is bounded to the north by a flexure zone characterized by fault-bounded blocks (Fig. 60, sections I and II). This flexure zone is the transition from the basinal area to the Pará-Maranhão platform. Because it is also bounded to the south by a horst feature (Ferrer-Urbano Santos Arch), the São Luís Basin is a graben (Mesner and Wooldridge, 1964; Rezende and Pamplona, 1970; Ponte and Asmus, 1976).

The Bragança-Vizeu Basin with similar Itapecuru sediments of continental origin is probably contemporaneous with Piauí-Barreirinhas São Luís system of marginal basins (Fig. 57; Rezende and Pamplona, 1970). The total sediment thickness of the Bragança-Vizeu is considerably less than that of the São Luís Basin and probably does not reach more than 2000m of sediments. The Bragança-Vizeu Basin was probably linked to São Luís Basin during the deposition of the Itapecuru formation (Rezende and Pamplona, 1970).

After the end of the Canarias Group deposition, which was probably associated with the uplift of the intervening horst between São Luís and Barreirinhas basins and with the subdued topography of the Ferrer-Urbano Santos Arch, the Itapecurú sedimentation spread away from the São Luís Basin, spilling over the Ferrer-Urbano Santos Arch to the south, and contributed sediments to the Parnaíba Basin. This marks the time when the predominantly carbonatic deposition (Caju Group) started in the Barreirinhas Basin.

Pará-Maranhão Basin - The Pará-Maranhão Basin is flanked to the west by the marginal fault of the Pará-Maranhão platform in the continental shelf. In the continental rise the basin is bounded to the north and to the east by the northern east-west and the northwest-southwest segments of the North Brazilian Ridge, respectively. East of the marginal fault in the continental shelf the sediments thicken rapidly from 2000m to more than 4000m (Ponte and Asmus, 1976). The marginal fault of the Pará-Maranhão platform starts in the south, in the vicinity of the so-called "sinclinal maior" of Miura and Barboza (1972), where the fault has an approximate orthogonal relationship with the prevailing east-west structures of the Piauí, Barreirinhas and São Luís basins (Fig. 57). The north-northwest initial direction of the marginal fault changes to a more northwesterly oriented direction and to an east-west trend in the continental shelf of Pará state. The northernmost part of the marginal fault coincides with the area in the upper continental rise where the westernmost extension of the northern east-west segment of the North Brazilian Ridge (one of the transverse ridges of Saint Paul's Fracture Zone) is present (Fig. 7).

No specific studies about the continental shelf off the western margin of Pará-Maranhão basin are currently available in the published literature. However, this part of the continental shelf apparently had no major barriers to sediment progradation, because the sediments thicken seaward and a large sedimentary prism was built in the Pará-Maranhão continental margin. The acoustic stratigraphy of the lower continental rise generally shows continuity towards the shallower portions and has a distinct character in relation to the areas outside of the North Brazilian Ridge indicating that the Pará-Maranhão Basin may have evolved independently and was semi-isolated (Figs. 3 and 4, profiles 69, 71-73, 75, 78 and 79).

The Pará-Maranhão Basin was formed during the early rifting of Africa and Brazil. The basal sedimentation of the basin is probably similar to the Canarias Group deposition of the Barreirinhas Basin, and was probably derived mainly from the Pará-Maranhão platform. The basin may correspond to an open-type of basin (or half-graben type).

Ferrer-Urbano Santos Arch - The Ferrer-Urbano Santos Arch, as mentioned before, is a structure that separates the Paleozoic Parnaíba Basin from the Bragança-Vizeu, São Luís, Barreirinhas and Piauí basins (Fig. 57, Rezende and Pamplona, 1970; de Oliveira and de Castro, 1971).

Because the basal sediments of Barreirinhas Basin are Paleozoic and because there are no Paleozoic sediments covering the region of the arch itself (once an integral part of the Parnaíba Basin) a very linear tectonism is suggested for the seemingly contemporaneous evolution of the high (erosion) and of the basinal area (subsidence) (Fig. 59). The strong denudation of the arch area is also suggested by the coarse fan-like deposits that interrupted the deltaic sediments of the Canarias-Group deposition in the basal portion, and by the present-day outcrop of shield rocks between the once continuous basal portion of Barreirinhas Basin and the Parnaíba Basin (Fig. 57). The association of coarse pebbles of diabase and basalt with these sediments suggests the possibility that the area of Ferrer-Urbano Santos Arch was affected by volcanism that was probably contemporaneous with the volcanics and basic intrusions of the Parnaiba Basin.

Cretaceous sediments onlap Paleozoic, Triassic and Jurassic sediments towards the arch, in the area to the south of the Ferrer Urbano Santos Arch (Mesner and Wooldridge, 1964). At the time the basal portion of the Canarias Group of the Barreirinhas Basin was deposited, the paleotopography of the arch itself was probably irregular with spots of high and low areas in relation to the contemporaneous aggradation level. The low areas enabled the sea to invade the Parnaîba Basin at that time through narrow regions to deposit sediments in a semi-restricted environment (gypsiferous layers of Codō formation; Rezende and Pamplona, 1970). The highest isopach-gradients and thicker areas of the Codō formation are immediately to the south of the high and indicate anomalous subsidence of the southern border of the Ferrer-Urbano Santos Arch (Rezende and Pamplona, 1970).

After the Canarias Group sedimentation in the Barreirinhas Basin had ceased, the Ferrer-Urbano Santos Arch remained a relatively positive area in relation to the regional sea level and to sediment accumulation, but apparently it did not constitute, even locally, a source area of terrigenous sediments during the deposition of the calcareous Caju Group.

The Ferrer-Urbano Santos Arch to the south of the Bragança-Vizeu and São Luîs basins coincides with the radiometric boundary of the so-called São Luîs cratonic area of Almeida et al. (1973) . This cratonic area has yielded radiometric ages between 1800 and 2200 m.y. that contrasts with the shield rocks to the south that mainly constitute the basement of Parnaiba Basin, and that were dated in the 450-700 m.y. age range (Fig. 57; Almeida et al., 1973; Hurley and Rand, 1973). The radiometric boundary also coincides with a prominent gravimetric high that delineates the Ferrer-Urbano Santos Arch and that markedly contrasts with the distinct negative signatures of the Bragança-Vizeu and São Luís basins in a Bouguer gravity anomaly map of the area (Fig. 61; de Oliveira e de Castro 1971).

The radiometric boundary of the São Luís cratonic boundary may be also characterized by mylonitization zones (R. Sadowski, pers. communication, 1975). Therefore, the Ferrer-Urbano Santos Arch in the region here considered, and the Bragança-Vizeu and São Luís basins were developed by Mesozoic tectonism in an old Precambrian weakness zone.

To the east of the São Luís Basin, the proposed radiometric boundary of the São Luís cratonic area (Almeida et al., 1973) cuts the trend of the onshore-Barreirinhas Basin almost at right angles (Fig. 57). In spite of that, the Ferrer-Urbano Santos Arch continues to the east abandoning the probable old cratonic boundary and bounding to the south the Barreirinhas and Piauí basins. The trend of this portion of the arch cuts at high angles the foliation of the shield rocks (Fig. 57). In summary, the Ferrer-Urbano Santos Arch is an early Cretaceous or late Jurassic structure that was formed in response to the rifting and drifting of South America with respect to Africa. The structure is partly constituted by a newly developed structural trend (eastern part) and partly by a rejuvenated old weakness zone (western part).

Atlântico High - Very complex structures in the offshore-Barreirinhas Piauí Basin were interpreted from multi-channel seismic work done by Petrobrás to be reverse faults and folds close to the continental-sheld edge. The reverse faults dip seaward and their strikes together with those of the fold axes are essentially east-west (Figs. 57 and 60; Section 4; Miura and Barboza, 1972).

Drilling data from the area of folding and reverse faulting confirmed the complexity of the structures and showed that some of the sedimentary groups of Barreirinhas basin were uplifted along an east-west trend, thus characterizing a high area (Fig. 60, Section 4). The youngest sediments that apparently were involved in the folding were dated as Santonian (Miura and Barboza, 1972). Campanian sediments unconformably overlie the Santonian sediments and apparently were not involved in the folding. Miura and Barboza (1972) considered the structures of the Atlântico High as being formed probably during the Coniacian-Santonian time.

The structures diaplayed in the Atlântico High are unique essentially because they parallel the depocenter of the Barreirinhas and Piauí basins and the normal faults of the southern border of the basins.

Igneous rocks, probably basic volcanics, were drilled in the area of the eastern extremity of the Atlântico High in the continental shelf and they are the probable cause of a magnetic high on the continental shelf of this area (Fainstein et al., 1975). This also indicates that some magmatic activity took place in the Atlântico High.

The geological role of the Atlântico High during the early sedimentation history of the Barreirinhas-Piauí is not clear. Apparently, the Atlântico High was never a source area for terrigenous sediments of the basal Canarias Group of the basin (Miura and Barboza, 1972). Prodeltaic sediments of that group were involved in the folding and they do not suggest evidences of high areas in the vicinity, at their time of deposition, according to drilling results (Miura and Barboza, 1972). This suggests the possibility that the Atlântico High is a late Cretaceous structure. However, a west-east direction of sedimentary facies change predominated during the Canarias Group deposition and studies of the carbonatic sequence of the younger Caju Group of Barreirinhas Basin also showed a deeper-water facies generally in the eastern side of the basin, in contrast with a shallower-water facies to the west (Miura and Barboza, 1972; Carozzi et al., 1973). The fact that the continental slope off the Atlântico High is the steepest and the deepest in the Pará-Maranhão continental margin sector and the indication that the continental rise has a southwest-northeast gradient rather than a south-north gradient, supports the conclusion that no appreciable south-north sedimentary progradation has taken place from the continental shelf region where the Atlântico High is present (Figs. 50 and 4, profile 72). Thus it is also possible that the Atlântico High had been a high area since the deposition in the Barreirinhas-Piauí Basin and, as a barrier for the south-north continental margin progradation, it would have caused a longitudinal sedimentary facies change from the main source area to the west, toward the east. Nevertheless, during the Campanian-Oligocene time and also during the Miocene there are clear indications of northward progradation of terrigenous sediments across the Atlântico High (Miura and Barboza, 1972).

The continental slope that parallels the Atlântico High is in line with the slope of the Ceará Terrace to the east (Fig. 50).The Ceará Terrace contains sedimentary peaks that display complex structures such as folds and faults (Fig. 4; profiles 65 and 66;Gorini and Bryan, 1974). These complex structures, are most probably aligned in an east-west direction, along the strike of the Atlântico High in the continental rise.

The westward prolongation of the Atlântico High is interrupted in the continental shelf by the so-called "sinclinal maior", an area of thick deposition (Fig. 57; Miura and Barboza, 1972). The "sinclinal maior" coincides with the area where the predominant east-west structures of Piauí-Barreirinhas and São Luís basins meet with the predominant northwest-southeast structures of the marginal fault of the Pará-Maranhão platform (Fig. 57). In the "sinclinal maior" there are growth faults and anticlines of compensation (Miura and Barboza , 1972). The counterpart of the Atlântico High in the São Luís cratonic area may be represented by the flexure zone associated with the northern border of the São Luís and Bragança-Vizeu basins.

<u>Northwest-Southeast segment of the North Brazilian Ridge</u> - This feature will be considered here because it is an integral part of the physiography and of the tectonic framework of the Pará-Maranhão continental margin sector. This segment of the North Brazilian Ridge (Hayes and Ewing, 1970) corresponds to a complex of seamounts and

basement highs that structurally connect the two east-west segments of the North Brazilian Ridge (Figs. 7 and 50). The peaks of some of the seamounts are as shallow as 900m. This segment abruptly separates the continental rise of Pará-Maranhão to the west from the Ceará abyssal plain to the east. Seismic profiles across this segment of the North Brazilian Ridge show distinct acoustical stratigraphies and a marked ponding of sediments behind the ridge segment (Fig. 4,profiles 78 and 79). Sediment isopachs (Kumar et al., in preparation)show that the sediments to the west of the norteast-southwest segment are 2 km thicker than the sediments lying to the east of the same segment.

The seamount complex that makes up the bulk of this segment of the North Brazilian Ridge is covered by sediments that make the topmost part of the ridge approximately flat (Fig. 50, Bader et al., 1970; Hayes and Ewing, 1970). Evidences that the seamount complex was close to the sea level some time during the Miocene were given by two DSDP Leg IV holes that were drilled in 2000m of water. These holes penetrated 59.4m and 58.5m into the sedimentary cover of the seamount complex. Fragments of shallow-water algal limestone were recovered underneath late-middle Miocene-late Pleistocene oozes (foraminifera and nannoplankton)(Bader et al., 1970). The algal limestone is similar to limestones associated with reef growth in the fringes of some islands in the Atlantic (Fernando de Noronha, Atol das Rocas,and Trindade) and similar to algal limestones that floor the outer continental shelf of the eastern and northern Brazil and several flat-topped seamounts of Trindade Seamount Chain in the eastern Brazilian continental margin (Mabesoone and Coutinho, 1970; Zembruscki et al,1971;

Gorini, 1969; Kempf et al., 1968).

These algal limestones overlain by carbonatic oozes of late-middle Miocene to late Pleistocene age were recovered at a depth of about 2000m. Thus, subsidence rates ranging from $8.5cm/10^{-3}$ yrs. to $9.1cm/10^{-3}$ yrs. were calculated considering the initial depths of 140m and 0m, respectively and assuming that the subsidence has been taking place since the beginning of the Miocene. Even higher subsidence rates are possible if it is assumed that the oldest sediments overlying the algal limestones are middle Miocene in age ($14.7cm/10^{-3}$yrs or $13.8cm/10^{-3}$ yrs if the 140m depth for algae growth is considered).

Discussion

The geological fit of Africa and Brazil - When the South American-African continents are fitted together, the Pará-Maranhão Basin faces the Ivory Coast Basin (Fig. 62). It is remarkable that both basins are bounded by structures in the continents believed to be associated with the Saint Paul's Fracture Zone to the north and with the Romanche Fracture Zone to the south. The Barreirinhas-Piauí Basin faces the offshore-Ghana Basin and their offshore highs come together in the fit of the two continents (Fig. 62).

Several remarkable geological coincidences exist between the offshore-Ghana and Barreirinhas basins:

i) The only occurrence of Paleozoic sediments (Devonian) in Western Equatorial Africa is preserved in the littoral zone of Ghana (Takoradi, Sekondi, Accra and Elmina) in the northern border of the offshore-Ghana Basin (flexure zone of the littoral zone of Ghana)(Fig.

62). Sediments of Paleozoic age constitute the basal portion of that basin that is overlain by Cretaceous sediments (Delteil et al.,1974) and the Paleozoic sedimentary succession of Parnaiba Basin (Almeida and Black, 1968) constitute the basal sediments of Barreirinhas Basin. Such coincidence implies that the Paleozoic Parnaiba Basin extended into the African side probably in a narrow strip of Ghana, as Ponte et al. (1971) proposed. The fact that no Paleozoic sediments are found in the interior of Ghana and Ivory Coast indicates accentuated erosion in the shield region probably associated with tectonic movements related to the early rifting of Africa and Brazil.

ii) The offshore-Ghana Basin is a graben structurally bounded by an offshore high and by a flexure zone that more or less coincides with the littoral of Ghana, Togo and Dahomey (?) (Fig. 62).

The Barreirinhas-Piauī Basin is also a graben bounded by an offshore high (Atlântico High) and by a flexure zone in the littoral zone that corresponds to a very linear high (Ferrer-Urbano Santos Arch) (Fig. 62).

iii) The offshore-Ghana Basin had its general pattern of sedimentary facies change from the east towards the west (Kazumi Miura , personnal communication, 1974) and no appreciable progradation of sediments has occurred in a sense transverse to the basin (one exception is the Volta delta region).

The Barreirinhas-Piauī Basin had its general pattern of sedimentary facies changes from the west towards the east (Pamplona,1969; Rezende and Pamplona, 1970; Miura and Barboza, 1972) and no appreciable sedimentary progradation from the south to the north occurred in the basin, at least up to the Santonian-Campanian time.

iv) The offshore-Ghana Basin is open towards the west and most probably sediments were transported towards the west along the axis of the basin. This westward sediment progradation built up the accumulation area (Ivory Coast Plateau) northward from the Ivory Coast-Ghana Ridge.

The Barreirinhas-Piauí Basin is open towards the east, and most probably sediments migrating along the depocenter of the basin from the west towards the east were responsible for the development of a small marginal plateau, the Ceará Terrace. The Ceará Terrace is probably limited to the north by a high associated with the North Brazilian Ridge and with the eastern extension of the Atlântico High.

v) The continental shelf high of the area of the offshore-Ghana Basin lies along the strike of the Ivory Coast-Ghana Ridge that is laterally followed into the southern transverse ridge of the Romanche Fracture Zone. The continental shelf high of the area of the Barreirinhas Basin is on the strike of the North Brazilian Ridge which is traced laterally into the northern transverse ridge of the Romanche Fracture Zone (Fig. 7).

vi) The Accra Fault that marks an important radiometric, structural and metamorphic boundary in the African side (Hurley and Rand, 1973; Grant, 1973) is abruptly cut by the prevailing east-west structures of the offshore-Ghana Basin in the Ghanaian continental shelf (Fig. 62). The probable geological representation of the fault in the continental shelf may correspond to basement highs of secondary tectonic role in the basin (Burke, 1969; Delteil et al., 1974).

The radiometric boundary of the São Luís cratonic area, in the Brazilian side, is also cut abruptly by the basinal east-west axis of Barreirinhas and Piauí basins. Nevertheless, to the south of the São Luís and Bragança-Vizeu basins the radiometric boundary, as a fault zone, was reactivated and located on the western extension of the Ferrer-Urbano Santos Arch. The radiometric boundary may also correspond to an important structural and metamorphic boundary. Whether this boundary continues into the continental shelf off onshore-Barreirinhas Basin is not known. However, the continuation of the Accra fault and the boundary of the São Luís cratonic area, when considered together lie in close proximity to each other (Fig. 62).

The Ivory Coast Basin is bordered by the marginal fault of the Ivory Coast. The Ivory Coast Basin merges to the south with the offshore-Ghana Basin, after the Ivory Coast marginal fault has changed its strike appreciably to a point where an almost orthogonal relationship with the main structures of the offshore-Ghana Basin exists (Fig. 62). The sedimentary progradation that built up the Ivory Coast continental margin essentially came from the basinal area of the Ivory Coast Basin. The Pará-Maranhão Basin is bounded westward by the marginal fault of the Pará-Maranhão platform (Fig. 62). The Pará-Maranhão Basin in the continental shelf merges with the Barreirinhas Basin to the south through an area of thick deposition ("sinclinal maior") after the marginal fault of Pará-Maranhão platform had changed its strike appreciably to bear an almost orthogonal relationship with the main structures of the Barreirinhas-Piauí Basin (Fig. 62). The sedimentary progradation that construted the Pará-Maranhão continental margin sector essentially came from the Pará-Maranhão Basin in the continental shelf.

Comments on the origin of marginal basins - From the beginning of the tectonism that preceded the actual separation of Brazil from Africa, the region of Barreirinhas-Piauī system of basins was linearly uplifted and eroded in one part (Ferrer-Urbano Santos Arch) and linearly subsided in the other (Piauī and Barreirinhas basins) with the preservation of their Paleozoic cover. Consequently, during the early rifting time between Africa and Brazil, the site of the Barreirinhas-Piauī Basin was probably not submitted to a domal phase of uplift in the way that it is generally attributed to the Red Sea rift (Falvey, 1974; Ross and Schlee, 1973).

As mentioned earlier, the tectonic setting of Barreirinhas-Piauī Basin is the mirror image of the tectonic setting of the offshore-Ghana Basin, in the African side (Fig. 62). Because the two basins are located in the trend of Romanche Fracture Zone and their tectonic features are correspondent to the topographic features (that are indeed tectonic entities) of the fracture zone in mid-ocean (Fig. 63), the area of the two basins probably corresponded to the site of the Romanche Fracture Zone in continental crust when Africa and Brazil were still together. Consequently, the area of the two basins probably represented the site of the main transform motion during the initial stages of seafloor spreading.

The horst and graben tectonic setting of the offshore-Ghana and Barreirinhas-Piauī basins is similar to the tectonic setting of fracture zones within the offset region of the Mid-Atlantic Ridge axis. Because the offshore-Ghana and the Atlântico highs are close to the continental-shelf edge, and abruptly separate distinct physiographic provinces on the two sides of the equatorial Atlantic, and because of their juxtaposition in the fit of the two continents, they probably represent the crustal boundary between the oceanic and continental crusts (Fig. 62). Consequently, the site of the main transform motion between the two continents may have been located between the two offshore high areas (Fig. 64).

Because the site of the main transform motion of the Romanche Fracture Zone within the offset region of the axial region of the Mid Atlantic Ridge is represented by a deep trough, we can assume that either the site of the main transform motion of continent against continent was also represented by a deep trough, and, in this case, a common basinal area for the Barreirinhas-Piauī and offshore-Ghana Basin was established; or that the two high areas were always tectonic and sedimentary boundaries, in contrast with the mid-oceanic evidence, that individualized from the beginning two independent basinal areas (Fig. 64, 1st and 2nd models).

The first model implies that the Ghana shelf-edge high and the Atlântico high are later tectonic features and probably originated after the transform motion continent against continent had ceased. The origin of these features might have been a consequence of the rebound of the continental crust after the cessation of tectonic constraints mainly associated with the transform motion. This may have been concomitant to low-density ultramafic diapirism along the fault zone where the main transform motion took place. This model is compatible with the indirect geological evidence of the sedimentation of the basal Canarias Group in the Barreirinhas Basin that precluded an offshore high during the Aptian-Albian through Cenomanian time.

The second model is compatible with the geologic evidence of the offshore-Ghana Basin and implies a high area separating this basin from the Barreirinhas-Piauī Basin since the beginning of the rift between Africa and Brazil. This intervening high area was probably never a source area for the two adjacent basins and acted as a main barrier for sediment progradation in a direction transverse to the basinal axes of the two adjoining basins.

The folds and reverse faults of the Atlântico High in the Piauī-Barreirinhas Basin can be explained on either model of Figure 64 by the tectonic uplift of a sediment-covered basement horst. The evidence that the basement high of the continental-shelf edge is constituted by shield rocks is given by dredge results in the continental slope off Cape Three Points in Ghana (Delteil et al., 1974). Another attempt to explain the structures of the Atlântico High involved folding associated with the shear motion of the continents (Miura and Barboza, 1972).

Because the Pará-Maranhão Basin is also the mirror image of the Ivory Coast Basin and both are half-graben basins, it is probable that they might have formed one continuous basin during the early rifting history of Africa and Brazil. Because basal sediments of the Ivory Coast Basin consist of coarse clastics that are widespread in the basinal area, it is inferred that basal clastics are also of widespread occurrence in the Pará-Maranhão Basin. This would indicate that the bordering cratonic areas of the two basins were very high at that time. With the consequent drifting of Africa from Brazil and the formation of oceanic crust in the middle portion of the basin, the original graben finally split into two half-graben basins that now constitute the Pará-Maranhão and the Ivory Coast basins (Fig. 65).

Fracture zone and rifting directions were equally fundamental in establishing marginal basins in the Pará-Maranhão and Ivory Coast continental margin sectors. These directions were imposed by a stress system that fundamentally provoked rifting and fracture zone directions. Old weakness zones with directions that were not favorable for reactivation by the dominating stress system were not reactivated, and consequently, new structures were formed at high angles with these old lineaments.

<u>NW-SE segment of the North Brazilian Ridge and the continental rise fault</u> - Le Pichon and Hayes (1971) demonstrated, in their reconstruction of the equatorial Atlantic for 80 m.y.B.P., that the site of the continental rise fault of Arens et al. (1970) of the Ivory Coast continental margin approximately coincided with the northwest-southeast segment of the North Brazilian Ridge. Although the continental rise fault does not have a topographic expression as its possible counterpart in the Brazilian margin, both represent a hinge zone between re gions of contrasting thicknesses of sedimentary covers. They both approximately parallel the continental shelves of the Ivory Coast and of the Pará-Maranhão continental margin sectors. They do not represent a continental-oceanic boundary because Houtz et al. (in press) have found oceanic crustal velocities on the landward side of the North Brazilian Ridge. The topographic expression of the northwest-southeast

segment of the North Brazilian Ridge may be very recent in age.

Although a basement-depth contrast might have existed from the onset of the structure, a contrast in depth of basement might have been created in a tectonized zone of the oceanic crust, that acted as a hinge zone between a heavily sedimented area to the west and a "normal" oceanic area to the east. The differential subsidence between areas of oceanic crust on both sides of these structures may have triggered magmatism and/or diapirism of low-density ultramafics that may have formed the prominent seamount of the northwest-southeast segment of the North Brazilian Ridge. The reason why a seamount complex did not develop in the African side may reflect asymmetry of the geometry of the tectonized part of the oceanic crust on both sides of the Atlantic.

A geographical coincidence exists in the location of the northwest-southeast segment of the North Brazilian Ridge in the Brazilian margin, and the continental rise fault in the African margin.The geographic and tectonic setting of both features in the African and Brazilian sides is remarkably similar to the Mid-Atlantic Ridge axial segment with respect to the bounding Romanche and Saint Paul's fracture zones (Fig. 7).

In both sides of the Atlantic, the Romanche Fracture Zone which is represented by the Ivory Coast-Ghana Ridge and the North Brazilian Rdige shows changes in strike as pointed out by Le Pichon and Hayes (1971) (Fig. 7). These different strikes led these authors to consider two poles of rotation to account for the geometry of the opening of the South Atlantic. The early pole of rotation of Le

Pichon and Hayes (1971) is valid as far seaward as the topographic extremities of these marginal ridges where they become buried basement features. The tectonic map of Figure 7 indeed shows that there is a change in strike of the features associated with Romanche Fracture Zone in the junction with the marginal ridges.

If the tectonic map of the equatorial Atlantic is considered with the traces of the marginal ridges only, and the northwest-southeast segment of the North Brazilian Ridge is made coincident with the trace of the continental-rise fault of the Ivory Coast, the reconstruction of Le Pichon and Hayes (1971) of the equatorial Atlantic at 80 m.y. is the result (Fig. 63). In this reconstruction, the northwest-southeast segment of the North Brazilian Ridge and the continental-rise fault of Ivory Coast have a symmetrical position in relation to the adjacent continental shelves and are approximately parallel to the marginal faults of Ivory Coast and Pará-Maranhão basins that most probably reflect the original direction of rifting between Africa and Brazil. These facts suggest that the common direction of the continental-rise fault and the northwest-southeast segment of the North Brazilian Ridge corresponded to an early sea-floor spreading direction.

The reason they constitute hinge zones and as such delineate the boundary of two different basement levels may lie on tectonic events that may have taken place in response to changes in transform fault directions. These changes are shown by changes in strike of the North Brazilian and the Ivory Coast ridges in relation to the strike of the Romanche Fracture Zone (Fig. 7). Comparable and possibly contemporaneous changes in strikes are shown by the Fernando de Noronha and the Jean Charcot ridges (Fig. 7). This change seems to be contemporaneous in the equatorial Atlantic as shown by Le Pichon and Hayes (1971). Such a change in transform fault direction may cause tectonic reactivation of fracture zones within and outside the offset regions of the Mid-Atlantic Ridge. As discussed in Part I of this work,it is possible that with the tendency of the spreading axis to become orthogonal to the transform fault direction, a tectonized portion of the oceanic crust will be developed in close vicinity to the spreading axis which reflected the direction transverse to the former transform direction. This tectonized region constitutes a weakness zone that tends to yield to subsequent differential stresses. Sediment load may have played an important role in causing differential subsidence of the oceanic crust. The weakness zone tends to accomodate the differential stresses on the two sides of the zone by yielding to greater sediment load on one side. The resulting differential subsidence may interact deeply with the upper mantle and thus may trigger magmatism and/or low-density ultramafic diapirism giving rise to topographic or basement features.

Cearã Continental Margin Sector

Physiography

The Cearã continental margin sector off Cearã and Rio Grande do Norte states is bounded by the southern east-west segment of the North Brazilian Ridge and its continuation into the northern transverse ridge of the Romanche Fracture Zone, and by the Fernando de Noronha Ridge to the south (Fig. 50; Gorini and Bryan, 1974). Terrigenous sediments are ponded behind these two marginal ridges and consequently the continental rise is considerably shallower than the adjacent abyssal region to the north of the North Brazilian Ridge or to the south of the Fernando de Noronha Ridge (Figs. 3 and 4; profiles 56,58, 59, 60, 62 and 63; Hayes and Ewing, 1970; Gorini and Bryan, 1974).

The topographic gradient of the continental rise is west-southwest to east-northeast (Fig. 50). Major channels are present in the middle of the continental rise and these constituted the main pathways for terrigenous sediments to bypass the continental rise and to reach the Cearã abyssal plain to the north and the abyssal regions to the east (Damuth, 1973). One of the prominent channels is the Equatorial Mid-Ocean Canyon that now is a relict feature (Fig. 66; Damuth and Gorini, 1976).

The continental shelf of the Cearã sector is very uniform in width and its strike is northwest-southeast. The continental slope is interrupted off the city of Fortaleza by the so-called Cearã Plateau and by the Cearã Terrace, to the north (Fig. 50). The Cearã Plateau (Ealey, 1969) is actually a flat-topped seamount very close to the continental shelf-edge off Fortaleza (Fig. 67, profiles 3 and 4) and lies to the west of another seamount that is almost entirely buried by continental rise sediments (Fig. 50).

The line of seamounts and basement highs off northeast Brazil that includes the Fernando de Noronha Islands and the Atol das Rocas Islet form the Fernando de Noronha Ridge (Gorini and Bryan, 1974). The peaks of some of these seamounts are very shallow (300m deep) flat-topped and elongated along the east-west strike of the ridge (Figs. 50 and 67; profiles 5-8). The Fernando de Noronha Islands are the peak of a roughly circular seamount as revealed by the 4000m bathymetric curve. In depths shallower than 4000m, the seamount has a marked east-west orientation (Fig. 50). The islands are volcanic and are formed by alkaline rocks (Almeida, 1958). Radiometric datings yielded 1-12 m.y.for rocks of the islands (Cordani, 1970).

The continuity of the Fernando de Noronha Ridge is demonstrated by the ponding of sediments against its northern side and by the tectonic control that seems to be imposed on the Equatorial Mid-Ocean Canyon (Fig. 66; Gorini and Bryan, 1974; Damuth and Gorini, 1976).

Although Fernando de Noronha Ridge terminates as a topographic feature at about 31ºW, it apparently extends eastward in subsurface as a lineament that delineates two distinct basement levels and that may bound two different crustal segments, as discussed in Part I of this work (Fig. 66). Thus, similar to the east-west segments of the North Brazilian Ridge, Fernando de Noronha Ridge is traced into the Fernando de Noronha Fracture Zone (Gorini et al., 1974). The trough associated with the Fernando de Noronha Fracture Zone,to the east of 30ºW, was partially filled by sediments coming from the Equatorial Mid-Ocean Canyon (Fig. 66). The partial filling of this trough extended a low gradient area farther eastward, into the flanks of the Mid-Atlantic Ridge.

General Geology and Tectonic Framework

In most of the shield bordering the Ceará continental-margin sector a strong northeast-southwest foliation of the metamorphic rocks is present (Fig. 52, DNPM, 1971). This predominant trend of the shield rocks contrasts with the essentially east-west border of the continental shelf of the northern part of Rio Grande do Norte State, with the east-west trend of the Fernando de Noronha Ridge and with the northwest-southeast strike of the shelf-edge off Cearã State.

Several important structural lineaments are present in the shield (Kegel et al., 1958; Kegel, 1965), and some of them like the Sobral-Pedro II and the Jaguaribe lineaments were reactivated in the Mesozoic (Fig. 57).

The pericratonic basin of Potiguar or Rio Grande do Norte was originated in the Mesozoic. Erosional remnants of very extensive Cretaceous basins constitute isolated sedimentary patches throughout the northeastern Brazilian shield (Ghignone, 1971).

The littoral region of Cearã and Rio Grande do Norte is extensively covered by Tertiary and Quaternary sediments laid down on regions of relative tectonic stability that constitute the Fortaleza and Natal platforms. These platform areas are separated from the Cearã and offshore-Potiguar basins by normal faults (Fig. 57).

Cearã Basin - Cearã Basin is an entirely offshore basin situated in the continental shelf, to the north of the city of Fortaleza (Fig. 57). The basin is bordered landward by the Fortaleza platform which is the eastward continuation of the Parnaība platform. The marginal fault of the Cearã Basin approximately parallels the shelf-edge off Cearã State and merges with the marginal fault of the Parnaība platform to the west (Fig. 57).

The northern end of the Cearã Basin corresponds to the easternmost extension of the Atlântico High and its continuation into the slope of the Cearã Terrace (Fig. 57). The Cearã Basin is structurally connected with the Piauí Basin (Miura and Barboza, 1972). Off the city of Fortaleza, Ceará Basin is interrupted and separated from the offshore-Potiguar Basin to the south, by the Fortaleza High (Miura and Barboza, 1972). To the east, Ceará Basin merges (Fig. 57) into the Fernando de Noronha Basin.

Very little has been published about the sedimentary succession of the Ceará Basin. Miura and Barboza (1972) pointed out that the Ceará Basin has at least 2000m of terrigenous sediments in its basal sequence and more than 3000m of total sediment thickness (Ponte and Asmus, 1976). In general, the basin shallows toward the southeast, where the Fortaleza High is present.

Complex structures have been recognized in the offshore continuation of the Sobral-Pedro II lineament of the shield (Fig. 57). Ojeda y Ojeda in Ponte and Asmus (1976) mapped folds and reverse faults in the continuation of the Sobral lineament into the continental shelf. These folds form an anticlinal structure that has a northeast-southwest axis (Ponte and Asmus, 1976). This strike of the fold axis corresponds to the trend of the Sobral-Pedro II lineament in the shield region. According to Ponte and Asmus (1976) the offshore continuation of Sobral-Pedro II lineament is the boundary of two contrasting tectonic settings in the Ceará Basin. To the west of the lineament, the basement of the basin is deep and dips landward; to the east of the same lineament, the basement is shallow and dips oceanward.

Other folded structures are present in the northern part of the Ceará Basin as shown by Miura and Barboza (1972) and Ponte and Asmus (1976) (Fig. 60, section 5). These folded structures are parallel to the east-west fold structures of the Atlântico High in the Piauí-Barreirinhas Basin.

The east-west folded structures of the Piauí-Barreirinhas and the Ceará basins which include reverse faults are similar to the structures associated with the offshore continuation of the Sobral-Pedro II lineament, although the latter has a northeast-southwest trend. The age of the folding episode that developed the complex structures of the Atlântico High in the Piauí-Barreirinhas Basin was tentatively placed as Coniacian-Santonian by Miura and Barbozá (1972), whereas the folds and reverse faults associated with the offshore continuation of the Sobral lineament were interpreted by Ojeda y Ojeda (in Ponte and Asmus, 1976), on the basis of structural and stratigraphic relationship, as early Cretaceous to Albian and, locally, possibly as young as Turonian.

The Ceará Basin is open eastward into the Fernando de Noronha Basin (Figs. 50 and 57). Terrigenous sediments that bypassed Ceará Basin also caused sedimentation in the area of the Ceará Terrace and in the region between the North Brazilian Ridge and the Ceará Plateau in the upper continental rise (Fig. 50).

Eastward from the Fortaleza High, the continental shelf contains the seaward extension of the Potiguar Basin.

Potiguar or Rio Grande do Norte Basin - The Potiguar or Rio Grande do Norte basin is a pericratonic basin that has evolved since the early Cretaceous in the Rio Grande do Norte and Ceará States (Kegel, 1957 ; Sampaio and Schaller, 1968).

In the continent, the Potiguar Basin is structurally divided in two distinct regions. The western part of the basin corresponds to the area of greatest sedimentary accumulation(more than 2500m of sediments)whereas the eastern part of the basin corresponds to an area of lesser than 500m of sedimentary accumulation(Natal platform) (Fig.57 ; Miura and Barboza, 1972; Asmus and Ponte, 1973).

The area of greatest sedimentary accumulation of Potiguar Basin is confined tectonically between two important marginal faults.The western fault is the northward continuation of the Jaguaribe lineament (Fig.57;Kegel, 1965; Ponte and Asmus,1976).The Jaguaribe lineament in the basinal area is probably represented by normal faults that separate a shallow basement region to the west,from the basinal area to the east(Fig.57; Miura and Barboza,1972).This marginal fault is essentially parallel to the foliation of the shield.The eastern boundary of the depocenter of the Rio Grande do Norte Basin is associated with a marginal fault that separates the Natal platform to the east,from the thicker sedimented area,to the west.This fault [Macau Fault of Miura and Barboza (1972)] is also approximately parallel to the dominant shield trend (Fig.57).

The Potiguar Basin ends to the south by the so-called Apodi scarp of more than 500m of relative altitude that fringes most of the southern boundary of the basin(Kegel,1957).East-west oriented dolerite dykes that are remarkably linear in outcrop are immediately to the south of the Apodi scarp(Fig.68;Rolff,1965; Ferreira and Albuquerque , 1969).The lineament of dykes was named Cabugi after the Cabugi peak in the vicinity of the city of Lages(Soares,1968 in Ferreira and Albuquerque,1969).These dykes are concentrated in a zone 10-15 km wide and are continuous for at least 100 km (Fig.68).They dip at very high angles and are generally 40-60m wide but locally may attain 400m in width. The dolerite dykes cut the marked northeast-southwest structural trend of the shield at high angles without apparent horizontal displacement of the structures (Fig.68;Rolff, 1965).Chilled margins and a progressive increase in crystal size towards the middle of the dykes are common features.The dikes are 125-130 m.y. old and are either constituted by olivine-bearing or olivine-free diabases (Cordani, 1970; Sial, 1975).

According to Rolff (1965),the Cabugi dikes may extend from the vicinity of the Jaguaribe lineament in the west,to the littoral region situated to the north of the city of Natal,(a distance of more than 200 km; Fig.57). Because the dikes represent a preferred direction of tension and because they mark the approximate outcrop boundary of the sedimentary rocks of the Potiguar Basin to the north, the Cabugi lineament corresponds most probably to the southern tectonic boundary of Potiguar Basin and consequently to a flexure zone of the shield.

The Cabugi lineament was reactivated in the Miocene with the intrusion of several plugs and/or the development of volcanic centers (Fig.68).These intrusions correspond to a volcanic suite composed of ankaratrites, basanites, and olivine-basalts with basanitic or tholeiitic affinities (Sial,1974) which were dated as 20 m.y.(?) old (Cordani,1970). The occurrence of xenoliths of peridotite nodules is very common in these volcanic centers (Rolff,1965; Leonardos,Jr.and de Araũjo,1970).These peridotite nodules are of sheared spinel-lherzolites and harzburgites (Sial,1974; Leonardos,Jr.and de Araũjo,1970), and are believed to have originated from a depth of approximately 64$\pm$ x km (Sial,1974).The nodules are similar to those that constitute xenoliths in nepheline-basanites in the Fernando de Noronha Islands (Almeida, 1958).

Three sedimentary formations named as the Gangorra, Açu and Jandaira compose the Mesozoic stratigraphic sequence of Potiguar Basin. The Gangorra formation is dated as Aptian-Albian on the basis of pollen stratigraphy. The Açu formation is of Albian-Cenomanian age according also to pollen stratigraphy. Jandaira formation is abundantly fossiliferous and its age was placed between the Turonian and the Santonian (Sampaio and Schaller, 1968).

In the Miocene, basaltic lavas were extruded in the vicinity of the city of Macau, close to the littoral zone (Fig.57;Kegel, 1957). The lavas of Macau are very similar in composition to the Miocene volcanic rocks of the Cabugi lineament (Sial,1974).This basaltic volcanism of such recent age together with the volcanism of the Cabugi lineament is unique in the Brazilian shield.

The Potiguar Basin extends offshore and is limited to the east in the continental shelf by the seaward extension of the Macau Fault. The Natal platform to the east of the Macau Fault in the continental shelf contrasts with the basinal region to the west because of its relative thin cover of sediments (less than 3 km) (Fig.57; Miura and Barboza,1972; Ponte and Asmus,1976). To the west the basin extends as far north as the Fortaleza High. The offshore-Potiguar Basin is partially barred in the upper continental rise by the Fernando de Noronha Ridge and by seamounts associated with the Ceará Plateau (Fig. 7). Sediments that bypassed the offshore-Potiguar Basin in the continental shelf either went eastward, to the south of Fernando de Noronha Ridge, or went into the Fernando de Noronha Basin to the north.

<u>Fernando de Noronha Basin</u> - Sediments that bypassed the marginal basins in the continental shelves were deposited eastward on the ocean

floor and constructed the sedimentary wedge that now corresponds to the continental margin of the Ceará sector.The name of Fernando de Noronha Basin (Gorini and Bryan, 1974) was given to this basinal area,that includes several physiographic and tectonic boundaries with a complex geological history.

Fernando de Noronha Basin has an acoustic stratigraphy that is distinct from that of the areas to the north and to the south. This acoustic stratigraphy consists of a lower, intermediate and upper reflectors.The lower and intermediate reflectors are entirely confined to the Fernando de Noronha Basin whereas the upper reflector is widespread in area and not confined to the basin only (Figs. 66,69 and 4, Profiles 58-61).

The tectonic control of ridges and troughs for the sedimentation in the Fernando de Noronha Basin is shown by the distribution of the main reflectors in the basin and by the tectonic fabric of the area (Fig.69). The lower reflector extends laterally into trough areas and it is not present on basement ridges. The intermediate reflector extends far eastward from the easternmost extension of the lower reflector. The easternmost extension of the intermediate reflector probably follows the trough area associated with Chain Fracture Zone, to the east of 30ºW longitude. The upper reflector in the Fernando de Noronha Basin does not extend appreciably through the topographic gap of the North Brazilian Ridge - Romanche Fracture Zone and its southern areal distribution follows approximately the Equatorial Mid-Ocean Canyon (Fig. 66; Damuth and Gorini, 1976). Its probable easternmost extension follows the prominent trough associated with the Fernando de Noronha Fracture Zone.

Discussion

Although it is possible that topographic features in the North Brazilian and Fernando de Noronha ridges are quite recent, as in the case of Fernando de Noronha Islands (1-12 m.y.) and Atol das Rocas, the lineaments now represented by these marginal ridges were present since the beginning of the sedimentary deposition in the Fernando de Noronha Basin. This is demonstrated by the confinement of the lower and intermediate reflectors of the Fernando de Noronha Basin against the two marginal ridges.

The lineament of the southern east-west segment of the North Brazilian Ridge apparently continues into the Ceará Terrace and into the Atlântico High in the continental shelf off Piauí and Maranhão states. Those features have also acted as sedimentary barriers. It is not clear if these features are really continuous as correspondent features of the same fracture zone (Romanche Fracture Zone) because the area of the continental rise where these features occur is close to seamount complexes that complicate the relationship. The seamounts close to the Ceará Terrace are either displaced to the north or to the south from the lineament that characterizes the North Brazilian Ridge east of 37ºW (Fig. 66). This area of complex morphology of the North Brazilian Ridge coincides with the location of the northwest-southeast segment of the North Brazilian Ridge to the north and probably also corresponds to the region where a transform motion of a continent against a continent changed to a continent against oceanic crust (Fig. 63): Consequently, it is a complex area where tectonic adjustments may have taken place. These tectonic adjustments may have developed the tectonic features of Romanche Fracture Zone in a different position than by the time when continent was against continent.

The Fortaleza High, the Ceará Plateau and the Fernando de Noronha Ridge were probably effective sedimentary barriers, at certain times, to terrigenous sediments coming from the offshore Potiguar Basin into the Fernando de Noronha Basin. The sediments during those times were most probably diverted eastward, to the south of Fernando de Noronha Ridge. At present, however, sill depths have been reached and an effective contribution of terrigenous sediments has been made from the area of the offshore-Potiguar Basin to the Fernando de Noronha Basin (Fig. 50).

The Fortaleza High, which is on the strike of the Ceará Plateau and other seamounts to the east, is most probably constituted by magmatic intrusions (Miura and Barboza, 1972) consisting of basaltic intrusives and/or lava flows (Ponte and Asmus, 1976). To the west of the Fortaleza High, the Mecejana Phonolite (30 m.y.), in the vicinity of the city of Fortaleza, constitutes the westernmost extension of the conspicuous east-west Fernando de Noronha Ridge (Fig. 57; Almeida, 1958). The Mecejana Phonolite corresponds to the only known occurrence of alkaline rocks in the northern Brazilian shield of such recent age and its alkaline character coincides with the alkaline characteristics of the volcanic rocks of Fernando de Noronha Islands.

The folded structures of Piauí-Barreirinhas and Ceará basins have their fold axes parallel to the main trends of the structures with which they are associated. In this aspect, they also compare with the fold structures of the Benue Trough, in the African side, that we re discussed earlier.

The fact that similar complex structures are close to each other in the Brazilian margin but are oriented in different trends raises several possibilities for their origin. Because they are believed not to be contemporaneous, they may be explained by different stress systems if their folds and reverse faults are interpreted as a result of compressional forces. The folded structures could also be explained by shear motion associated with a compressive stress system (Miura and Barboza, 1972). Thus, in order to explain the northeast-southwest fold axis of the structures in the offshore continuation of the Sobral-Pedro II lineament of the Ceará Basin we have to imply that at the time of the folding, the compressive stress system was active in a direction either perpendicular to or at 45º to the Sobral-Pedro II lineament. Because there is no evidence of similar structural trends derived elsewhere in the northern Brazilian margin, it is improbable that the compressive stresses would have created folds and reverse faults only in the Ceará Basin area. Because the fold axis of the structure follows the trend of the Sobral-Pedro II lineament in the continental shelf, it is also improbable that the folding and reverse faulting associated with the lineament is the result of the reactivation of the Sobral-Pedro II fault because in this case the resulting fold axis would have probably been a nortwest-southeast trend instead of a northeast-southwest trend (Fig. 57). Moreover, the different settings of the basement blocks of the transcurrent character on both sides of the offshore continuation of the Sobral-Pedro II lineament in the Ceará Basin demonstrate that differential tilting of basement blocks have occurred to the north and to the south of the fault.Thus, vertical block faulting can be also considered as a probable explanation for the origin of the folded structures in the Ceará Basin. Consequently, these structures are better visualized as a result of differential block faulting that caused the sedimentary cover to yield to differential basement tectonics. Uplifted basement blocks caused anticlines and reverse faults, This model for the explanation of the folded structures of the Ceará Basin is essentially the same given earlier for the origin of the folds and reverse faults of Piauí-Barreirinhas Basin.

The pericratonic Potiguar Basin originated as a consequence of reactivation of old structural trends of the shield and of the development of the new tectonic trends that cut the old shield structures at very high angles. The reactivation of old structural trends originated the northeast-southwest depocenter of the basin. The new tectonic trends are the flexure zone represented by the Cabugi lineament, the trend of marginal faults in the offshore-Potiguar Basin and the direction of the continental shelf-edge bordering the basin (Fig. 57).

When the two continents are considered together, (Fig. 70), the Ceará Basin (a half-graben) faces the Togo-Dahomey Basin (a half-graben) in Africa; the Fortaleza High and the Ceará Plateau are in line with the Okitipupa Ridge-Benin Flank of the Niger Delta Basin; the Potiguar Basin and the Rio Grande do Norte Plateau overlaps the Niger Delta Basin; and the Cabugi lineament, the southern border of Potiguar Basin, coincides with the Jean Charcot High under the Niger Delta and is in line with the Abakaliki folded belt and the southern border of the Benue Trough.

The Romanche Fracture Zone represented by the continental shelf highs in offshore Ghana (Africa) and in offshore Piauí and Ceará states (Brazil) mark the northern boundary of the two basins. The Okitipupa Ridge that limits the Togo-Dahomey Basin to the south, occupies the same relative position of the Mecejana Phonolite, the Fortaleza High and the Ceará Plateau that separates the Ceará Basin from the offshore-Potiguar Basin (Fig. 70). The coincidence of the limits of the two basins suggest that they may have shared a common basinal area in the beginning of their development (Fig. 65).

The lineament represented by the Okitipupa Ridge in the African side is not followed westward into the continental rise on the basis of the present data but the ridge most probably either represents an old fracture-zone direction that did not continue appreciably into the continental margin or is represented oceanward by the Chain Fracture Zone (Figs. 63 and 7). The lineament represented by the Fortaleza High and the Ceará Plateau, in the Brazilian side, is essentially east-west and is slightly displaced to the north of the axis of the Fernando de Noronha Ridge. Apparently, the lineament is connected with the Fernando de Noronha Ridge and Fracture Zone to the east. Nevertheless, there is a coincidence in strike with the hypothetical westward prolongation of Chain Fracture Zone at 30.5ºW (Fig. 7). Because there is no evidence from available data of the presence of the Chain Fracture Zone west of 30.5ºW., it becomes a matter of semantics to say which fracture zone is associated with the Fortaleza High and Ceará Plateau. The evidence is that these two features were most probably present from the onset of the structures of Ceará and offshore-Potiguar basins in the same way that probably the Benin Flank-Okitipupa Ridge lineament was present from the onset of the Benue Trough and thus represent an old fracture zone direction.

Two fracture-zone directions are believed to border the Benue Trough (Chain (?) and Jean Charcot) (Fig. 7). These fracture-zone directions are represented in the continent by the flexure zones that characterize the northern and the southern borders of the Benue Trough. Barreirinhas-Piauí Basin, in the same way as the Benue Trough, has folded sedimentary rocks with fold axis directions essentially parallel to the axis of the basin. The folding episode (Santonian-Coniacian or earlier in age) cannot be explained by the models proposed for the folding episode of the Benue Trough as mentioned earlier. However, because of the similarities and the uniqueness of the trend of the fold axes in the Benue Trough and in the Barreirinhas-Piauí Basin, in relation to their depocenters, they probably have a similar tectonic origin and evolution.

Because Barreirinhas-Piauí Basin is a graben that is in line with a fracture zone that apparantly has its features better explained by intensive vertical motions in a horst and graben topography, as suggested earlier in this work, the folding of the Barreirinhas-Piauí Basin is interpreted as a consequence of vertical block movements where tilting and differential uplift of horsts caused local and regional compression and reverse faulting in the sedimentary cover.

The extrapolation of this hypothesis for the folding episode in the Benue Trough is further supported by the evidence of outcropping shield rocks in narrow horsts. The faults that border these horsts are essentially parallel to the fold axes of the folded sedimentary rocks close to Biũ. This parallelism suggests that intensive block faulting of positive and negative motion has taken place in the area (Fig. 44). Furthermore, the Upper Benue embayment shows the majority of its fold axes in a direction roughly parallel to the margins of the embayment that is transverse to the Benue Trough (Fig. 44). This preferred orientation of the fold axes of the embayment is difficult to be explained by compressive stresses essentially in a northwest-southeast direction according to the northeast direction of the fold axes of the trough.

If, indeed, Benue Trough was developed by tectonism associated with formation of fracture zones, then fracture-zone tectonics explain the episodic nature of the folding event(s) that affected the basin (and also the Barreirinhas-Piauī Basin). The folding event (s) may possibly have originated due to changes in the direction of transform motion that was representing the drift of Africa and Brazil at a certain time in the geological past. As it was discussed in Part I of this work, changes in the transform directions cause important tectonic events in the oceanic crust. During the time when the fracture zone is adapting to the new orientation, intense fracturing, block faulting, and compression takes place affecting a considerable area of the ocean floor that otherwise would have been tectonically passive. We can easily imagine what would happen in areas affected by thick sedimentary cover during times when the continents were very close to each other. Changes in fracture zone directions could reactivate the original trends of fracture zones and thus provoke an extensive tectonic episode of deformation.

When the two continents are considered together, the location of Fernando de Noronha Ridge to the east of the Fortaleza High-Cearã Plateau lineament fits between the Benin Flank and the Natal Platform (Figs. 63 and 7). In the African side, that region corresponds to part of the Anambra Platform early Cretaceous) that later became the Anambra Basin. If the line of rupture of the two continents was between the Benin Flank and the continental slope seaward of the Natal platform, a transform fault is implied in that boundary and thus it would justify theoretically the presence of the Fernando de Noronha lineament as representative of an early fracture zone direction. However, the absence of such a lineament in the African side (if not deeply buried under the Niger Delta continental margin) is a discrepancy that probably indicated important geological asymmetries on both sides.

In the vicinity of 31ºW of longitude, the strike of Fernando de Noronha Ridge changes appreciably and at 30.5ºW corresponds to the first major lineament to the south of the Chain Fracture Zone (Fig.7). Jean Charcot Ridge has the same relative position in relation to Chain Fracture Zone at 2.5ºW in the African side (Fig. 7). However, when the two continents are considered together (Fig. 70), the probable original positions of the Fernando de Noronha and Jean Charcot ridges differ considerably and point to the conclusion that the two lineaments were not aligned with respect to each other, during the earliest phase of opening of the Atlantic Ocean. The Abakaliki folded belt and the Jean Charcot High are a continuation of/or coincide with the Cabugi lineament, to the south of Potiguar Basin, in Brazil (Fig. 70). Although Jean Charcot High and the southern border of Benue Trough continue into the continental margin as a prominent ridge, the Cabugi lineament apparently did not give rise to a prominent ridge in the continental margin of the northeastern Brazil. The only indirect evidence for any tectonic trend in the eastward continuation of the Cabugi lineament in the continental margin is the coincidence of the southern border of the Rio Grande do Norte Plateau with the hypothetical prolongation of the trend in the continental shelf (Fig. 50). Again, as in the case of Fernando de Noronha Ridge and Chain Fracture Zone, this may indicate important asymmetry in the development of marginal ridges on the two sides of the Atlantic, in contrast with the alignments associated with Romanche and Saint Paul's fracture zones.

The distribution of reflectors in the Fernando de Noronha Basin demonstrates that the sedimentation in the basin was controlled by the morphology of the basement. Troughs were filled first and with thicker sediments than the surrounding areas (Fig. 69). The location of the Equatorial Mid-Ocean Canyon was tectonically controlled by the Fernando de Noronha Ridge.

The youthfulness (1-12 m.y.) of the volcanic islands of the Fernando de Noronha Ridge demonstrates that reactivation of fracture zones close to the continental margins is not uncommon. This implies that fracture zones continue as deep crustal boundaries far from the offset of the Mid-Atlantic Ridge. A seismic profile in a north-south direction to the west of the westernmost seamount of the Fernando de Noronha Ridhe revealed a deeply buried basement block with stratified sediments lying underneath continental rise sediments (Kazumi Miura, personal communication, 1974). This suggests that Fernando de Noronha Ridge is composed of uplifted basement blocks, not just volcanic seamounts. The nature of the stratified material is uncertain and may represent the earliest deposition in the area, during the incipient separation of Africa and South America that was subsequently isolated by later tectonism.

Rio Grande do Norte - Pernambuco Continental Margin Sector

The Rio Grande do Norte-Pernambuco continental margin sector was not studied detail in the present work but is included here as a comparison with the correspondent Fernando Poo-Cameroon continental margin sector in the African side.

Physiography

The Rio Grande do Norte-Pernambuco continental margin sector has its northern boundary in the Fernando de Noronha Ridge. The southern boundary of the sector is not obvious from the present geomorphic data. However, (1) the presence of the Recife Plateau (Fig. 116), an area of overlap in the Bullard's (1965) reconstruction; (2) the occurrence of an offshore high off the city of Recife which structurally separates two adjacent basins; (3) the presence of east-west aligned seamounts at about 32ºW in the latitude of the city of Recife; and (4) other geological considerations discussed later indicate that the area of the Recife Plateau and associated east-west aligned seamounts is the southern boundary of the Rio Grande do Norte-Pernambuco continental margin sector (Fig. 71).

The continental shelf edge is generally parallel to the littoral zone from the city of Macau to Recife. Around the bight of northeastern Brazil, the shelf-edge protrudes somewhat in the area of the Rio Grande do Norte Plateau (Fig. 71; Ealey, 1969). The continental shelf is generally narrow and reaches its narrowest region between the cities of João Pessoa and Natal.

The continental slope as outlined by the 2500m bathymetric contour is generally parallel to the shelf-break except in the areas of the Rio Grande do Norte and of the Recife marginal plateaus. Between these two areas the continental slope is generally north-south oriented and is indented by several canyons in the region between Natal and João Pessoa where the continental slope is the steepest in the sector (Fig. 71; Boyer, 1969). East of the city of Macau the continental rise has a northeast topographic gradient, that gradatively changes to an eastward direction in the vicinity of the Rio Grande do Norte Plateau (Fig. 71). The general bathymetric gradient to the south of the aforementioned plateau is essentially eastward and grades laterally into the Pernambuco abyssal plain. The Pernambuco abyssal plain extends into the flanks of the Mid-Atlantic Ridge and transcends the hypothetical southern boundary of this sector of the continental margin.

Several seamounts are present in the continental margin off the cities of Cabedelo and João Pessoa (Fig. 71). Whether these seamounts belong to a mid-oceanic lineament could not be determined with the present data. To the east of 30ºW longitude, there are very few ship crossings and consequently the area is poorly known, with the exception of the mid-oceanic region, in the vicinity of Ascension Island (Van Andel et al., 1973).

The Rio Rio Grande do Norte Plateau is bounded approximately by the 500m and 1000m isobaths, and, where best developed, ranges in width from 20 km to 30 km (Ealey, 1969). A slope separates the continental shelf from the plateau and an even steeper slope (7º to 10º)separates the plateau from the continental rise (Ealey, 1969). The seaward boundary of the Rio Grande do Norte Plateau thus corresponds to a slope from 1000m to the 3500m isobaths (Fig. 67, profiles 9 and 10). The uppermost part of the plateau containsflat-lying reflectors and also complex structures (Ealey, 1969; Gorini and Bryan, 1974). The complex structures form sedimentary peaks in the seaward margin of the plateau and probably were derived from mass movements as figure 72 suggests.

The Recife or Pernambuco plateau lies to the east of the city of Recife. A steep slope separates the continental shelf from the plateau itself (Fig. 71). The area of the plateau is enveloped by depths slightly shallower than 1000m and by the 2000m isobath. The plateau is separated from the lower continental rise eastward by a steep slope from 2000m to 4250m in its northern region, and by a slope from 3000m to 4500m in the southern part of the plateau (Fig. 71).

The area of geomorphologic influence of the Recife plateau in the continental margin is better seen by the 3000m isobath. The continental rise, that is well developed in the areas to the north and to the south of the plateau, is considerably narrowed immediately to the east of the Recife Plateau (Fig. 71). In the area of the plateau itself and in its close vicinity there are several submarine mountains.

General Geology and Tectonic Framework

The Rio Grande do Norte-Pernambuco continental margin sector is bounded to the west by the northeastern Brazilian shield and from north to south, by the Natal Platform, the Recife-João Pessoa Basin and by the high area in the vicinity of Recife that separates the Recife-João Pessoa Basin from the Sergipe-Alagoas Basin to the south (Fig. 57).

Recife-João Pessoa Basin - The onshore extent of a narrow and shallow Cretaceous basin between the cities of Recife and João Pessoa constitutes the Recife-João Pessoa Basin which has been described as a homocline that dips gently to the east (Fig. 73; Asmus and Ponte, 1973). A well close to the littoral zone, at Itamaracá Island, reached basement at 388m (Fig. 73; Asmus and Ponte, 1973).

The stratigraphic sequence of the basin starts with littoral, lagoonal, estuarine and fluviatile sediments (Beberibe formation) of Santonian-Campanian age (Asmus and Ponte, 1973). The Beberibe formation is followed by continental shelf limestones of the Gramame formation of Maestrichtian age. The Gramame formation is overlain by continental shelf limestones of Paleocene-Eocene age that corresponds to the Maria Farinha formation. This latter formation is covered by clastics of continental origin of Pliocene-Holocene age.

Little is known about the tectonic framework of the Recife-

João Pessoa Basin in its onshore and offshore portions. Indirectly, by magnetic studies, Rand (1967 in Asmus and Ponte, 1973) inferred fault blocks with a similar orientation to those of the Sergipe-Alagoas Basin to the south. Ponte and Asmus (1976) pointed out that despite the tectonic framework similarities of the basin, the offshore-Recife-João Pessoa Basin has subsided less than the Sergipe-Alagoas Basin. A sedimentary cover of less than 2000m of sediments was revealed by a seismic survey in the offshore Recife-João Pessoa Basin (Souza, 1972 in Ponte and Asmus, 1976).

In contrast with the marginal basins situated to the south in the eastern Brazilian margin, no interval of salt deposition (Aptian) in the Recife-João Pessoa Basin has been found. The northernmost marginal basin of the eastern Brazilian margin that contains the widespread interval of Aptian salts is the Sergipe-Alagoas Basin situated to the south of the city of Recife.

The tectonic high off the city of Recife that separates the Recife-João Pessoa Basin from the Sergipe-Alagoas Basin to the south contains in its onshore portion the so-called Cabo Magmatic Province.

Cabo Magmatic Province - The magmatic province of the Cabo is unique in Brazil. Situated a few miles south of the city of Recife in the close vicinity of the "Cabo de Santo Agostinho", the Cabo province is characterized by volcanic, hypabissal and plutonic rocks of strikingly similar radiometric ages, that were intruded in sedimentary rocks (Fig. 74).

The sedimentary rocks in the Cabo region correspond to a very coarse conglomerate with intercalations of arkosic sandstones

(Cabo formation), a limestone, the Barreiras Series and Quaternary sands and muds (Fig. 74; Cobra, 1967; Mello and Siqueira, 1972). The Cabo formation has conglomerates that contain small boulders and cobbles composed of shield rocks. The limestone has poorly preserved fossils but Beurlen and Cobra (1960) on the basis of a fossil found in the limestone that is identical to fossils belonging to an Albian-age formation in the Sergipe-Alagoas Basin to the south, considered the limestone to be tentatively of Albian age. A suite of magmatic rocks of several lithologic types are intruded into the limestone and the conglomerates in the area of the "Cabo de Santo Agostinho" (Fig. 74). Trachytes, andesites, rhyolites, granites and basalts constitute the main type of rocks of the Cabo magmatic province (Cobra, 1967; Mello and Siqueira, 1972). The so-called Cabo granite at the "Cabo de Santo Agostinho" has been dated as 89 and 90 m.y. by the Rb-Sr method (Vandoros et al., 1966). The fact that several lithologic types have been dated either by K-Ar or Rb-Sr methods and all the samples fall very close to 90 m.y. is indicative of similar magmatic derivation (Cordani, 1970).

The geographic location of the Cabo Province coincides with the prolongation of the strike of the Pernambuco Lineament toward the littoral zone (Fig. 57).

The magmatic association of the Cabo province is very similar to the Younger granites of Nigeria, including the Tertiary granites of Cameroon.

Sergipe-Alagoas Basin - Although this basin is to the south of the Rio Grande do Norte-Pernambuco continental margin sector, it is included here because the basin is the counterpart of the Douala Basin in the African side and its description provides basic geological arguments for the time of the origin of the Cameroon Trend and for the paleophysiography of the equatorial Atlantic.

The Sergipe-Alagoas basin is situated to the south of a shallow covered basement in the continental shelf adjacent to the Recife Plateau and corresponds to a series of horsts and grabens generally aligned in a NE-SW direction (Fig. 75; Asmus and Ponte, 1973). The complex onshore portion of the basin gives way eastward to a much thicker sedimentary sequence under the continental shelf (Fig. 75). Some parts of the basin, like the Alagoas graben that is very close to the city of Maceió, are very thick and contain more than 6 km of sediments. The thickening appears to be very rapid towards the continental shelf where it may reach accumulations higher than 9 km (Fig.75; Asmus and Ponte, 1973).

The sedimentation in the Sergipe-Alagoas Basin probably started in the late Jurassic when continental deposits were laid down in the basin. This sedimentation together with a lowermost Cretaceous (Neocomian) deposition was widespread in the northeastern and eastern Brazil, and in Africa (Ponte and Asmus, 1976).

The upper Jurassic-lowermost Cretaceous sedimentation in the Sergipe-Alagoas Basin was followed by Aptian-Albian sediments consisting of basal, coarse, clastics with localized evaporitic deposition in several grabens of the basin (Fig. 76). The evaporitic deposition was contemporaneous with the widespread Aptian salt accumulation in the Brazilian and African marginal basins (Asmus and Ponte, 1973;

Pautot et al., 1973; Leyden et al., 1976). The first marine transgression in the basin took place in the Aptian, and after the Albian, several periods of marine regression and transgressions followed (Asmus and Ponte, 1973).

Sergipe-Alagoas Basin in its onshore portion structurally represents a transition from the exposed shield to the highly developed basin in the littoral and in the narrow continental shelf. The very fast thickening of the basin seaward from the coastal zone indicates that the basin is open seaward, although offshore highs have been mapped in the northern part of the offshore Sergipe-Alagoas Basin (Asmus and Ponte, 1973; Ponte and Asmus, 1976).

The absence of evaporites in the Recife-João Pessoa Basin to the north further emphasizes the tectonic as well as the geomorphological barrier that, at least in the Aptian time, the continental shelf of Recife area represented (Ponte and Asmus, 1976).

Discussion

When the two continents are considered together (Fig. 77)the Recife-João Pessoa Basin faces the Calabar Flank and the southeastern part of the Niger Delta Basin in Africa. In the large overlap of the Niger Delta and northeastern Brazil, smaller and very significant overlaps are caused by the protuberances of the Rio Grande do Norte and the Recife Plateau (Fig. 77). The Cameroon Trend represented by Mount Cameroon volcanics overlaps the Recife Plateau. The continental shelf high between Fernando Poo and Mount Cameroon overlaps the shallow covered continental shelf off Recife. It is remarkable that the Cabo

Magmatic Province, which is very similar to the magmatic provinces associated with the Cameroon Trend, overlaps the Fernando Poo-Mount Cameroon area (Fig. 57). The Pernambuco Lineament is in line with the Cameroon Trend volcanics. These volcanics close to the coastline may have erupted in the southern extension of the Ngaoundērē fault zone of the Cameroon.

The coincidence of the northern limits of the Aptian salt deposition in Africa and Brazil with the Cameroon Volcanic Line in the African side (Douala Basin), and the area of Recife continental shelf on the Brazilian side (Sergipe-Alagoas Basin) suggests that the Cameroon Trend and the continental shelf of Recife represented a physiographic barrier that separated the equatorial region from the South Atlantic in the early opening history of the South Atlantic. The Cabo Magmatic Province, the tectonic high of the continental shelf of Recife and the Recife Plateau represent a region of anomalous volcanism associated with the Cameroon Trend and with the Pernambuco Lineament during the beginning of the tectonic reactivation of the Ngaoundērē-Pernambuco fault zones (Almeida and Black, 1968) that coincided with the early opening of the South Atlantic. This anomalous volcanism was probably responsible for the development of several volcanoes that acted as barriers for the seaward sedimentation in the Recife continental margin. Consequently, low gradient areas were built by confined sedimentation among several seamounts and these constituted the Recife Marginal Plateau.

Sediment Dispersal in the Northern Brazilian Margin

As in the Gulf of Guinea, the manner of terrigenous disper sal in the northern Brazilian continental margin was highly dependent on the tectonic framework of the bordering continent and on the featu res associated with fracture zones. The bulk of the terrigenous sediments that constructed the continental-rise prism prograded from half graben basins towards the flanks of the Mid-Atlantic Ridge. In the case of the Pará-Maranhão continental margin sector this progradation was interrupted by the presence of the NW-SE trending segment of the North Brazilian Ridge.

Sediments prograded longitudinally along the axis of the ba sins in case of graben-type basins. Because these graben-type basins were open towards their eastern or western extremities in the continental rise, and because these basins were generally in strike with marginal ridges, the sediment progradation developed marginal plateaus in the continental rise behind the marginal ridges. The continental-rise prism has been built along the trend of the boundaries of the various crustal segments and has graded laterally into tongue-like abyssal plains associated with the troughs of the fracture zones. Such is the case in the Ceará continental margin sector: the continental rise extends eastward and the sediment prism branches into two lower continental rise-abyssal plain areas associated with the troughs of the Romanche Fracture Zone to the north, and with a trough of the Fer nando de Noronha - Jean Charcot Fracture Zone to the south. The eastward progradation of the continental rise of the Ceará sector is best documented by the relict Equatorial Mid-Ocean Canyon (Damuth and Gori ni, 1976) that was tectonically controlled by the Fernando de Noronha Ridge. The abandonment of the Equatorial Mid-Ocean Canyon was associat ed with the filling of the Fernando de Noronha Basin that caused the spilling of sediments through a topographic low between the North Brazilian Ridge and the Romanche Fracture Zone. From post-early Miocene time (on the basis of a prominent early Miocene reflector that truncates to the south of Romanche Fracture Zone) the pattern of terrigenous sediment dispersal changed from an eastward to a northeastward direction and started to influence the abyssal sedimentation to the north (Fig. 78).

In contrast with the Niger deep-sea fan, the Amazon cone or deep-sea fan (Damuth and Kumar, 1975) resulted from a continental margin progradation that was essentially transverse to the east-west fabric of the crustal segments of the equatorial Atlantic. Nevertheless, although the main sedimentary accumulation was northward, the terrigenous sediments coming from the Amazon River prograded also eastward and developed the southern extension of the Ceará abyssal plain (Fig. 78).

COMMENTS ON THE CONTRACTION HYPOTHESIS OF FRACTURE ZONE DEVELOPMENT

The large-offset fracture zones of the equatorial Atlantic extended far into the continents and played a fundamental tectonic role during the evolution of the rifting between South America and Africa. Individual basins of distinct character were thus formed along the sites of the future offsets within the continents. The horst and graben tectonics that dominated the evolution of these basins are correlated laterally into the transverse ridge-and-trough morphology of the mid-ocean fracture zones. Consequently, it is very probable that the same tectonic style of fracture zones that imprinted a horst-and-graben tectonic setting in the marginal basins along offset regions of the two continents, has existed to the present and contributed to the development of transverse ridges (horsts) and troughs (grabens)in mid ocean. Thus, the mechanics associated with fracture zone tectonics apparently did not change from the continental to oceanic crust.

Formation of transform faults have been observed during studies of lava crusts (Duffield, 1972) and provide an excellent model in convective systems for their origin. Menard and Atwater (1969) explained the topography associated with the large eastern Pacific fracture zones as being simply related to the present distance from the ridge crest, to the length of the offset and to the length and direction of the transform fault. Van Andel et al., (1969) considered the possibility of a rifting origin for the Vema Fracture Zone in the North Atlantic and tentatively explained the tectonics and the topography associated with the fracture zone on this basis. Girdler (1968) presented a map of the Red Sea and the Gulf of Aden showing displacement of marginal structure lines associated with small offsets of the central belt of large magnetic anomalies. Barberi et al. (1974) also invoked the possible presence of transform faults associated with offsets in geological boundaries in the Afar Triangle, and showed that the possible transform fault-zone is delineated by a linear occurrence of alkaline volcanoes or centers of eruption. The present work tentatively demonstrated that the equatorial Atlantic fracture zones were originated during the early rifting between Africa and Brazil and that they were not necessarily restricted between the rifting axes , future sites of spreading. They in some cases penetrated far into the continents (Benue Trough) and were originated either along the locations of old weakness zones (Ascension Fracture Zone (?)) or at new structural directions, cutting at high angles old structural directions of the shield, that were developed as a consequence of the break-up of Africa and Brazil (Romanche and Saint Paul's fracture zones).

In consequence, the equatorial Atlantic fracture zones and possibly the majority of the linearly continuous fracture zones must be regarded as very deep boundaries of the lithosphere not only underlying oceanic but also continental crust. Thus, any theory to explain the linearly continuous fracture zones that are traceable on land must consider the above seemingly fundamental properties of the equatorial Atlantic fracture zones.

Turcotte (1974) has commented that transform faults may be thermal contraction cracks and has based his argument on the following facts: a) the oceanic lithosphere cools away from the ridge axis where it was formed; b) the cooling of the oceanic crust develops thermal stresses sufficiently large to fracture the upper part of the lithosphere (25 km); and c) fracture zones are universal features of the ocean ridge system and exhibit the graben structure associated with crustal extension. However, Turcotte did not explain the origin of offsets and related the transform faults as occupying tension cracks resulting from the extension associated with the cooling of the oceanic crust. He mentioned that his calculations showed that a 150 km ridge segment and a temperature difference of 200ºC would cause an extension of about 300m across the transform fault. The tensional setting of the fracture zone would explain the rift-valley character of the fracture zones (Collette and Rutten, 1972) as if their sides had appeared to have rifted apart (Turcotte, 1974).

The contraction hypothesis although attractively simple does not explain the very complex morphology and petrology of the equatorial Atlantic fracture zones and the fact that fracture zones were present since the onset of rifting structures of Africa and Brazil. Further implication rises in the fact that fracture zones were not restricted between the sites of rifting of the two continents. Although the transverse ridges and transverse valleys could be explained by a horst and graben tectonics, a simple opening due to a lateral extension of hundreds of meters cannot explain the $\geq$ 50 km wide complex morphological zone of the equatorial Atlantic fracture zones and neither can explain the maintenance of the transverse ridges as either prominent topographic features or as prominent buried highs within and outside of the offset region. Furthermore the depths of the transverse ridges do not decrease with distance from the ridge axis in the same way as a cooling oceanic crust deepens with distance from the ridge axis, and rates of subsidence in fracture zones can be far greater than the one only associated with cooling as mentioned before (Bonatti et al., in press). It was also tentatively demonstrated in this work that, because of a change in transform direction, Romanche Fracture Zone was probably developed in part at expense of an "old" crustal segment far away from the axis of spreading where it had been originated and this certainly cannot be explained by contraction only. In consequence, contraction, if not overshadowed by extension due to intensive diapirism and vertical motions as components of the horizontal transform displacement of the two adjacent crustal segments,has a secondary role in developing the fracture zone morphology.

SUMMARY

In this work I have dealt with the basic problem of recognizing and tracing fracture zones across the entire equatorial Atlantic (4ºS to 3ºN) between Africa and Brazil. The Saint Paul's, Romanche and Chain fracture zones are the three most prominent fracture zones in the area. These have previously been described primarily on the basis of bathymetric data. In addition to using the bathymetric data, I have defined and traced these and other fracture zones by mapping basement structures observed on seismic-reflection profiles. Mapping of these basement structures has allowed me to produce a tectonic map for the entire equatorial Atlantic from Africa to Brazil. Earthquake epicenters in the mid-ocean region were plotted on this tectonic map. In my attempt to understand the geological role of fracture zones, I have also compiled a bathymetric map, all the dredge results in the mid-ocean and have carried out a bibliographic study of the geology of Brazil, Northwest Africa and of all the oceanic islands in the equatorial Atlantic.

The seismic mapping of the basement structures shows that fracture zones in the equatorial Atlantic Ocean are very continuous features that can be continuously traced from the mid-ocean ridge axis to the upper continental rises of Africa and Brazil. The structures on the continent can, in turn, be related to the individual features of fracture zones; this establishes the continuity of fracture zones from one continent (South America) to the other (Africa) across the entire equatorial Atlantic.

I have defined the fracture zones as very prominent and ge-

nerally linear basement features that bound segments of oceanic crust and offset the mid-ocean ridge. Because the equatorial Atlantic fracture zones cause large-offsets in the Mid-Atlantic Ridge, the prominent and linear basement features separate crustal segments which have various depths and differing thicknesses of sedimentary cover. Morphologically, fracture zones consist of basement ridges and intervening troughs.

A large number of earthquake epicenters occur within the main trough of the Romanche Fracture Zone and a fault-plane solution for one of these earthquakes shows a sense of motion that agrees with Wilson's (1965) proposed sense of motion for transform faults in the area (Sykes, 1967; Fig. 7). In contrast to Romanche Fracture Zone,epicenters in the Saint Paul's Fracture Zone show much scattering which implies that the present transform motion is accomodated in a narrower zone in the Romanche Fracture Zone than in the Saint Paul's Fracture Zone.

Serpentinized peridotites and peridotites are abundant among the rocks recovered in all depth-ranges from the walls of the transverse ridges of fracture zones. Sedimentary rocks were a minor constituent of dredged rocks and they were recovered from steep slopes of transverse ridges. Graywackes with quartz and mica, originally deposited in deep troughs, now crop out on the steep slopes of the transverse ridges. This suggests that active tectonism has raised them to their present level.

Subsidence rates for a segment of one of the transverse ridges of the Romanche Fracture Zone (0.2 mm/yr) were found to have far exceeded the subsidence rates of a "normal" cooling oceanic crust (0.01 mm/yr). The summits of transverse ridges of fracture zones between the offset portions of the Mid-Atlantic Ridge axis maintain their depths along the extent of the offset. Consequently, the topographic features associated with fracture zones are thought to be decoupled from the "normal" cooling oceanic crust between the offset portions of the ridge axis.

A discussion on the tectonics of fracture zones invokes vertical movements as a dominant cause for the development of frature-zone morphology.Vertical tectonics is thought to be caused essentially by the vertical component of the horizontal strike-slip motion and by upward intrusions of relatively low-density ultramafics. A deep fault that accomodates the transform motion of the adjacent crustal segments lies underneath the trough region and facilitates the upward migration of diapiric bodies of relatively low-density ultramafics.The diapiric masses probably originate in linear deep mantle disturbances that underlie the deep fault of the fracture zone and are probably not related to the upper mantle disturbances underlying the mid-oceanic ridge axis. The ridge-and-trough morphology of the fracture zone is thought to be a manifestation of their horst-and-graben structural setting.

Changes in the direction of the transform motion cause adjustments of the fracture zone to the new direction. These adjustments may involve inversion of relief of features associated with the previous transform direction and intense tectonism in portions of the oceanic crust far away from the original ridge axis.

The tectonic map of the equatorial Atlantic displays a mark ed east-west fabric for the entire ocean floor and shows a marked asymmetry of the position of the Mid-Atlantic Ridge axis towards the west. This asymmetry indicates that either the ridge axis "jumped"one or several times toward the west; or that asymmetric spreading with faster rates in the eastern portion has taken place.

The traces of the equatorial Atlantic fracture zones display a considerable variation in strike, width (> 50 km) and morphology. When the traces of the fracture zones are compared on both sides of the ocean a marked asymmetry in strike is apparent,an east-northeast strike of fracture zones in the eastern side contrasts with an east-west trend of fracture zones in the western side of the equatorial Atlantic. In the offset region of the Mid-Atlantic Ridge, the trace of Romanche Fracture Zone displays a similar asymmetry in strike.This may have relevance for explaining the asymmetry of fracture zone traces on both sides of the Atlantic.

Fracture zones divide the entire equatorial Atlantic ocean floor in distinct crustal segments which vary in width laterally along the strikes of bordering fracture zones. These crustal segments continue into the continental margins where each segment is bounded by marginal fracture ridges. These marginal fracture ridges are very prominent physiographic features which are the continuation of transverse ridges of fracture zones. The very high relief and the youthfulness of volcanism in some of these marginal ridges suggest that important tectonic adjustments have been taking place along fracture zones at distances far away from the offset region of the Mid-Atlantic Ridge. The prominent relief of the marginal ridges may not be related only to volcanism. I suggest that the foundations of marginal ridges may also be associated with relatively low-density ultramafics, in the manner proposed by Bonatti and Honnorez (1976). Landward, the marginal fracture ridges are in line with structural highs at the edge of the continental shelf. Troughs associated with fracture zones are in line with narrow offshore coastal basins; these basins are termed here as "graben basins". Segments of shelf which are bounded by two fracture zones, are underlain by basins which display a "half-graben" structure.

In general, sediments prograded seaward from the continents to the ocean floor along the half-graben type basins. Sediments in the graben basins prograded longitudinally (east-west) along the axes of these basins. In each case the marginal ridge acted as barriers or dams which prevented sediment progradation northward or southward into adjacent crustal segments.

When Brazil and Africa are fitted together (Fig.79), the similarities in tectonic settings of correspondent parts on each continent are apparent. Portions of the adjacent continents and continental shelves which contain marginal basins of half-graben type line up opposite each other as do portions with marginal basins of graben type.

In the early rifting history of Africa and Brazil, a horst-and-graben tectonic setting developed along trends that later developed into fracture zones. In contrast, the areas between these linear horst and grabens were probably sites of domal uplift and subsequent subsidence.

Other important structural features which are intimately related to rifting in the equatorial Atlantic and which are examined in detail are the Marajõ system of grabens, the Benue Trough and the Cameroon Trend. The Marajõ system of grabens in the Amazon area of Brazil is believed to have originated from the southward propagation of the rifting direction of the North Atlantic Ocean. Because the rifting in Liberia was probably associated with the opening of the North Atlantic Ocean, the Marajõ graben system may have predated the other marginal basins to the east in Brazil. The Benue Trough in Africa is a basin flanked by fracture-zone directions that imprinted a horst-and-graben setting on the basin. Both the Marajõ system of grabens and the Benue Trough were avenues for the progradation of sediments that developed huge sedimentary accumulations in the continental margins that today constitute the Amazon and Niger deep-sea fans. Folding in the Benue Trough is explained as vertical tectonism associated with the uplift and differential tilting of buried basement horsts. The vertical tectonism that caused reactivation of the horst and graben setting of the Benue Trough may be explained by adjustments to changes in transform motion between the African and South American plates. I have found no evidence which supports previous arguments for an episode of consumption of oceanic crust in the Benue Trough.

The Cameroon Trend is interpreted as an early-rifting structure which developed in response to the tectonic reactivation of the Precambrian Pernambuco (Brazil)-Ngaoundéré (Africa) lineament. The tectonic reactivation of this lineament caused extensive magmatism and the development of horst-like features that acted as a physiographic barrier to salt deposition in the South Atlantic during the Aptian; to the north of the lineament no salts were deposited. The Cabo Magmatic Province, a tectonic high off the city of Recife, and the seamounts and the basement highs of the Recife Plateau are thought to represent this lineament on the Brazilian side. The Fernando Poo-Annobon Ridge, the tectonic high on the continental shelf between the island of Fernando Poo and Mount Cameroon, the basaltic and alkaline volcanics, and the Tertiary granites of the Cameroon are the representatives of the Cameroon Trend in Africa. One of the most remarkable geological inferences about the Cameroon Trend is that it does not change its tectonic character from the ocean into the continent. The trend is apparently associated with the Ascension Fracture Zone in mid-ocean.

During the early-rifting period of eastern South America from Africa, the shield and platform areas of the two continents experienced a widespread tectonic and magmatic reactivation. Large areas of Brazil and Africa were affected by basaltic volcanism and hypabyssal intrusions. Predating and post-dating this basaltic volcanism, acidic volcanism was widespread in Africa but did not take place in Brazil (the single exception is the Cabo Magmatic Province). Tertiary and Quaternary alkaline and basic magmatism along the Cameroon Trend also has no counterpart in volume or in area in northern Brazil.The acidic magmatism in Africa may represent widespread crustal melting which implies that Africa was underlain by mantle disturbances that did not underlie the Brazilian shield and platform.

Fracture zones in the equatorial Atlantic were as important as the zones of rifting during the initial separation of South America from Africa. They were established at the onset of the rifting and did not necessarily form along old weakness zones in the Precambrian shields and platforms. The fracture zones developed, in some cases, along tectonic trends that are at large oblique angles to the structural trends and old weakness zones of the shields. In this respect, I think that fracture zones originate in response to deep-seated structures that are the result of a dominant stress system. Wherever the deep seated structures coincide in depth with old weakness zones, the fault zones are reactivated and may create fracture zones. Wherever the structures do not coincide with old shield lineaments, fracture zones form in "new" tectonic directions.

The geological fit of Africa and Brazil in the equatorial Atlantic shows a remarkable coincidence of several independent structures and features. The overlaps can be easily explained by sedimentary accumulations on oceanic crust or by distortions due to the projection used in Figure 79. Such coincidence of features suggests that the rifting and subsequent drifting of the two continents did not involve appreciable crustal distortion of the two continents.

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FIGURE CAPTIONS

Fig. 1→ The equatorial Atlantic and the fracture zones studied in this work.

Fig. 2→ Schematic crossings of fracture zones. Double bars are Ridge axis and A-B the direction of crossing and respective seismic profile. (a) When the crossing is parallel to the ridge axis, a basement level change is quite apparent except (b) when the crossing is at mid-distance from the axes of the two ridge segments; (c) when the ship's track is at an angle to the strike of the ridge axis in the offset region, an inversion of the apparent topographic gradient of the basement on the two crustal segments is seen in the profile; (d) close to one of the axes of two crustal segments, a pronounced change of level in the basement is apparent due to the high relief of the ridge in one side and the low level of the crust in the other. Sedimentary cover testifies to difference in age of the two adjacent crustal segments.

Fig. 3→ The equatorial Atlantic fracture zones and the continental margin sectors as discussed in the text. Numbers indicate seismic profiles in Figure 4. Mid-Atlantic Ridge axis is represented by double bars.

Fig. 4→ Seismic profile tracings. For location see Figure 3. Profiles are always north (to the left) - south (to the right) oriented. FP-AR-Fernando Poo-Annobon Ridge; JCR-Jean Charcot Ridge; JCFZ-Jean Charcot Fracture Zone; CFZ-Chain Fracture Zone; IC-GR-Ivory Coast-Ghana Ridge; RFZ-Romanche Fracture Zone; AFZ-

Ascension Fracture Zone; SPFZ-Saint Paul's Fracture Zone;GCR-Grand Cess Ridge; FNFZ-Fernando de Noronha Fracture Zone;FNR-Fernando de Noronha Ridge; NBR-North Brazilian Ridge; CT- Ceará Terrace; NW-SE/NBR-Northwest-Southeast segment of the North Brazilian Ridge. Profiles no. 2,4,6,9,12,17,21 and 37 are from Woods Hole Oceanographic Institution (Uchupi and Emery, 1974). All the other seismic profiles are from Lamont-Doherty Geological Observatory.

Fig. 5→ Tectonic map of the equatorial Atlantic in the region between the offset portions of the Mid-Atlantic Ridge axis, Romanche Fracture Zone is shown between dashed lines. Ridges are shown in diagonal hatching: troughs are stippled and buried ridges are shown in vertical hatching. Black dots are earthquake epicenters. Black square with arrows corresponds to the epicenters that Sykes (1967) used for fault-plane solutions for the Romanche Fracture Zone.

Fig. 6→ Bathymetric map of the equatorial Atlantic in the region between the offset portions of the Mid-Atlantic Ridge axis of the Romanche Fracture Zone. Bathymetry is in uncorrected meters using 1,500m/sec as the speed of sound in water. Contour interval is 500m except below 5000m where the interval is 1000m. Dotted lines represent the tracks for bathymetric data used in this work.

Fig. 7→ Tectonic map of the equatorial Atlantic. Topographic ridges are shown in diagonal hatching, topographic valleys are in stippled and buried ridges are shown in vertical hatching.The tectonic map was made primarily by correlation of features in seismic profiles of Figure 4. All available seismic profiles with the exception of those of Woods Hole Oceanographic Institution used in this work were taken from ship's crossings depicted as dotted lines. Mid-Atlantic Ridge axial segments are horizontally hatched. The tectonic features on Brazil and Africa were largely taken from published information (DNPM , 1971; ASGA-UNESCO, 1968; and others). Black dots delineate earthquake epicenters. Black square with arrows corresponds to the epicenter that Sykes (1967) used for fault-plane solutions in the Romanche Fracture Zone, Solid lines encircle fracture zones as shown in Fig. 3.

Fig. 8→ Bathymetric map of the equatorial Atlantic. Bathymetry is in uncorrected meters using 1500m/sec as the speed of sound in the water. Contour interval is 500m except in the African and Brazilian continental margins where the interval is 250m and in regions deeperthan 5000m where the contours are spaced at 1000m. Track control is seen by dotted lines. The soundings were provided by Lamont-Doherty Geological Observatory and the U.S. Hydrographic Office. Additional bathymetric information came from Allen (1964), Martin (1971); Delteil et al. (1974), and Emery et al. (1975) in the Gulf of Guinea; from Jacobi(in preparation) for the region off Liberia to the north of 3ºN and as far west as 20ºW; from Heezen et al.(1964) between 10º W and 20ºW; from Miami University sounding data between 13ºW and 25ºW; from Tomczak and Annutsch (1970) for the region between 19ºW and 25ºW; from Zembruscki et al. (1973) and Connary and Moody (unpublished map) for the northern Brazilian continental margin as far east as 30ºW longitude.

Fig. 9→ Isopach map of a portion of the northern Brazilian continental margin. Contour interval is 0.5 sec (two-way travel time) unless otherwise noticed. Ridges are shown in black; buried ridges in vertical hatching and troughs are stippled. The thickest accumulations are located on troughs. Seismic-profile crossings that were used as basic data, can be seen in Figure 7.

Fig.10→ Dredge locations in the Saint Paul's and Romanche Fracture Zones. The base map is a portion of Figure 7. Dredges are numbered and represented by triangles. Profile locations for figure 12 are marked A, B, C and D.

Fig.11→ Frequency of rock types in dredge hauls from the northern and southern walls of the Romanche Fracture Zone, east of 22ºW (after Bonatti and Honnorez, 1976).

Fig.12→ North-south topographic profiles across the Romanche Fracture Zone. Location of profiles is indicated in Figure 10. Vertical exaggeration is 20 times. The main rock types recovered at various levels along these profiles are indicated with percent estimates of relative quantities (by weight), except where the haul is too small (after Bonatti and Honnorez,1976).

Fig.13→ Scheme showing a possible way to explain the sedimentation of graywackes prior to their consolidation and incorporation in steep slopes of transverse ridges of Romanche Fracture Zone.

Fig.14→ Tectonic map of the equatorial Atlantic in the region between the offset portions of the Mid-Atlantic Ridge axis. Saint Paul's Fracture Zone is shown between dashed lines. Various shading patterns same as in Figure 7.

Fig.15→ Bathymetric map of the equatorial Atlantic in the region between the offset portions of the Mid-Atlantic Ridge axis of the Saint Paul's Fracture Zone. Conventions are the same as in Figure 8.

Fig.16→ Map of Saint Peter and Saint Paul's Rocks after Soares(1968). Numbers indicate highest points in meters. Foliation trends with corresponding dips are from Melson et al. (1972).

Fig.17→ Picture showing the highly foliated nature of rocks of Saint Paul's Rocks (photograph from Neil Opdyke, Lamont-Doherty Geological Observatory).

Fig.18→ Bathymetric map of the equatorial Atlantic in the region between the offset portions of the Mid-Atlantic Ridge axis of the Chain Fracture Zone. Conventions are the same as in Figure 8.

Fig.19→ Tectonic map of the equatorial Atlantic in the region between the offset portions of the Mid-Atlantic Ridge axis. Chain Fracture Zone is shown between dashed lines. For key refer to Figure 7.

Fig.20→ Ascension Fracture Zone and seismic profile crossings after Van Andel et al. (1973).

Fig.21→ Bathymetric map of the Mid-Atlantic Ridge in the region of

Ascension Fracture Zone (after Van Andel et al., 1973). Contours in hundreds of meters.

Fig.22→ Distribution of earthquake epicenters in the region of the Ascension Fracture Zone in the Mid-Atlantic Ridge (after Van Andel et al. 1973).

Fig.23→ Schematic tectonic section of a fracture zone within the offset region of a mid-ocean ridge. The first block diagram represents the site of future crustal rupture with underlying mantle disturbances; arrows denote motions of diapiric bodies of low-density ultramafics in the rigid portion of the lithosphere (the upper 25 km - Turcotte (1974)).Dashed lines represent fault surfaces.

Fig.24→ Scheme to explain the development of fracture-zone morphology. Arrows represent the relative motion of diapiric bodies. Curved arrows denote the tendency of the blocks to rotate away from the site of the main transform motion. Dashed lines represent fractures.

Fig.25→ Schematic block diagram of a portion of a transverse ridge of Romanche Fracture Zone based on dredge results.

Fig.26→ Geologic consequences of a change in the direction of the transform motion.

Fig.27→ Relative motions accomodated by the fracture zone due to differential subsidence of the adjacent crustal segments in regions away from the offset portions of the Mid-Atlantic Ridge axis.

Fig.28→ Fracture zone bounding adjacent crustal segments with a) similar spreading rates; and b) distinct spreading rates. Length of arrows represents magnitude of the spreading rates.

Fig.29→ Reconstruction of the equatorial Atlantic at approximately 100 m.y. The continental shelves of Brazil and Africa as we see them today are represented by the 200m isobath. The present-day marginal ridges are also shown; those on the African side are represented as dashed lines. GCR-Grand Cess Ridge; NBR-North Brazilian Ridge; FNR-Fernando de Noronha Ridge; FP-AR-Fernando Poo-Annobon Ridge.

Fig.30→ The asymmetry of Mid-Atlantic Ridge axis in the equatorial Atlantic and the fracture zones. Dashed line represents a flow line following a rotation of 14.95º around the early poles of Le Pichon and Hayes (1971) with respect to Africa and Brazil; and a remaining average rotation of 32.8º around the average pole (67.3ºN, 39.5ºW) with respect to Africa. The flow line was originated from a point in the South American continent of known counterpart in Africa, following Bullard et al's fit (1965). The continuous line represents same rotations as before but instead of starting from a point of known counterpart in the two continents, the rotations started from points in the Mid-Atlantic Ridge axis at its intersection with Romanche and Saint Paul's fracture zones. 16.4º rotations around the average pole of Le Pichon and Hayes (1971) to each side were made. The hypothetical axes of symmetry are also represented and are in dicated by arrows.

Fig.31→ Various bathymetric and tectonic features in the Gulf of Guinea. Bathymetry in kilometers (uncorrected). Contour interval is 0.25 km in the continental margin, and 0.5 km in the area of the Mid-Atlantic Ridge. Bathymetry same as in Figure 8.

Fig.32→ Tectonic map of Liberia and Ivory Coast. Modified from the tectonic map of Africa (ASGA-UNESCO, 1968). Dashed lines re present foliation trends. L, D are different radiometric provinces. Divergent arrows and stippled areas mark flexure zones or tectonic highs and platform areas. Faults are shown by continuous or dashed lines. Pattern in squares represent areas of Tertiary and Mesozoic sedimentation.

Fig.33→ Structural trends and sedimentary embayments of Liberia main ly from magnetic data (after Behrendt et al. (1974)). Solid lines represent diabase dikes; wiggly lines represent basement trends. Contours representing the depth of magnetic ba sement are in kilometers. The 200m isobath delineates the continental shelf off Liberia. The inferred boundary separa ting different age provinces is shown as dotted lines.

Fig.34→ Sedimentary accumulation and structure of the continental margin off Liberia (after Schlee et al., 1974).

Fig.35→ Tectonic map of Ivory Coast, Ghana, Togo and Dahomey, Key to various patterns same as in Figure 32.

Fig.36→ Ivory Coast Basin (after Spengler and Delteil, 1966).

Fig.37→ Geologic cross section of the Ivory Coast Basin (after Splenger and Delteil, 1966).

Fig.38→ Tectonic map of Ghana, Togo, Dahomey, Nigeria and Cameroon. Key to various patterns same as in Fig. 32. Black-filled a reas are volcanics of basaltic and alkaline composition. Dotted-filled areas in the Jos Plateau are granites. Fold axes are represented in the Benue Trough.

Fig.39→ Mesozoic magmatism and the "younger" granites of Western Africa (after Black and Girod, 1970). The "younger" grani- tes of the Jos Plateau, Aïr and Hoggar are encircled.

Fig.40→ Tertiary granites of the Cameroon. Key is the same as of figures 32 and 38. Radiometric ages are from Lasserre(1966).

Fig.41→ Tertiary and Quaternary volcanic districts of Western Afri- ca (after Black and Girod, 1970).

Fig.42→ Inferred structures of Togo-Dahomey, Niger Delta and Douala basins. Figure key is the same as of figures 32 and 38.

Fig.43→ Tectonic framework of the Niger Delta Basin and Benue Trough after Murat (1970).

Fig.44→ Geological map of the Upper Benue Depression (after Burke et al, 1970). Note fold axes parallel to elongated outcrops of shield bounded by faults.

Fig.45→ Sediment progradation of the Niger Delta Basin (after Hospers, 1965).

Fig.46→ Bouguer Anomaly Compilation Map of Niger Delta and Cameroon (after Hedberg, 1968).

Fig.56→ Inferred structures of the Amazon continental shelf. Figure key is the same as of Figure 52. NBR - North Brazilian Ridge.

Fig.57→ Tectonic map of N-NE Brazil with key for names used in the text. Structures are represented in the same way as of Figure 52.

Fig.58→ Tectonic features of the onshore Barreirinhas Basin (simplified from Pamplona, 1969).

Fig.59→ Geologic section across the Ferrer-Urbano Santos Arch and onshore-Barreirinhas Basin (from Rezende and Pamplona, 1970). For location, refer to Fig.58.

Fig.60→ Geologic sections of the São Luís, Barreirinhas, Piauí, Cearā and Potiguar basins (after Miura and Barboza (1972)). For location refer to Figure 57.

Fig.61→ Bouguer gravity anomaly map of part of Maranhão and Pará states (after de Oliveira and de Castro, 1969).

Fig.62→ Geological fit of Brazil and Africa. Figure key is the same as of Fig.55. Overlaps (black areas) are in southern Ghana and Volta Delta areas.

Fig.63→ Reconstruction of the equatorial Atlantic at about 80 m.y. (Le Pichon and Hayes, 1971). The continental shelves of Brazil and Africa as we see them today are represented by the 200m isobath. The present-day marginal ridges are also shown and those of the African side are represented as dashed lines. GCR-Grand Cess Ridge; NBR-North Brazilian Ridge;

NW-SE NBR-Northwest-Southeast segment of the North Brazilian Ridge; CRF-Continental Rise Fault (Arens et al., 1971);IGRR-Ivory Coast-Ghana Ridge; FNR-Fernando de Noronha Ridge;FNFZ-Fernando de Noronha Fracture Zone; CFZ-Chain Fracture Zone ; JCR-Jean Charcot Ridge; FP-AR-Fernando Poo-Annobon ridge.

Fig.64→ Hypothetical development of graben type of basin.

Fig.65→ Schematic diagram showing development of a half-graben type of basin (Falvey, 1974).

Fig.66→ Structural setting of the Equatorial Mid-Ocean canyon.Dashed lines indicate the approximate eastward limit of terrigenous sedimentation during late Quaternary time. Arrows indicate the predominant paths of terrigenous sediment dispersal during the late Quaternary.

Fig.67→ Bathymetric profiles in the northeastern Brazilian upper continental rise (after Ealey, 1969).

Fig.68→ Dolerite dykes of the Cabugi lineament in the Rio Grande do Norte State (Brazil). Sedimentary formations belong to the Potiguar Basin.

Fig.69→ Surface distribution of acoustic reflectors in the Fernando de Noronha Basin. Ridges are represented in black. Dotted areas show the areal distribution of the lower reflector; oblique hatching represents the surface distribution of the intermediate reflector. Other patterns are the same as in Fig.7. Figure 7 shows the track coverage for the area.

Fig.70→ Geological fit of Brazil and Africa. Figure key is the same as of Fig.55. Overlaps are in the areas of the Cearā Plateau and of the Niger Delta.

Fig.71→ Bathymetric map of the Rio Grande do Norte Pernambuco continental margin sector (after Connary and Moody; unpublished map). Isobaths are in corrected meters and are contoured in 250m interval.

Fig.72→ Seismic profile section across the Rio Grande do Norte Plateau (after Ealey, 1969). For location see profile 10 in the insert of Figure 67.

Fig.73→ Recife-João Pessoa Basin (after Asmus and Ponte, 1973).

Fig.74→ Geological Map of the Cabo de Santo Agostinho area, Pernambuco State (after Cobra, 1967).

Fig.75→ Basement structural framework of Sergipe-Alagoas Basin(after Asmus and Ponte, 1973).

Fig.76→ Distribution of evaporites in the Sergipe-Alagoas Basin (after Asmus and Ponte,1973).

Fig.77→ Geological fit of Brazil and Africa. Figure key is the same as of figure 55.

Fig.78→ Sediment dispersal in the western equatorial Atlantic. Base map is from Damuth 1975. Arrows denote main paths of terrigenous sediment dispersal.

Fig.79→ Fit of Brazil and Africa following Bullard et al.(1965) and adjusting to the apparently best tectonic fit. The continents were considered extended up to the 1500m bathymetric curve.

Figure key is the same for other similar figures used in the text. Solid heavy line delineates the 1500m contour on each continent.

APPENDIX

SHIELD GEOLOGY OF NORTHERN BRAZIL

The shield rocks in northern Brazil outcrop in extensive areas but, nevertheless, are also extensively covered by sedimentary rocks that geographically divide the craton (DNPM, 1971). Several nucleii of exposed craton or thinly covered basement areas constitute the northeastern Brazilian shield, the Guaporé platform, the Guyana shield and the São Luiz cratonic area (DNPM, 1971; Almeida et al., 1973). The sedimentary cover of the shield constitutes the intracratonic Paleozoic basins of Parnaíba and Amazonas (Figure A1). Mesozoic basins either constitute minor sedimentary patches inside of the craton or are concentrated along the maritime border of the shield(DNPM, 1971). Some of these latter basins extend appreciably inland (Potiguar and São Luiz basins) and others are restricted to the offshore regions, in the continental shelf (Ceará Basin) (Asmus and Ponte, 1973; Miura and Barboza, 1972).

Radiometrically, the Guyana shield, the Guaporé platform and the São Luiz cratonic area yielded similar ages of 1800 to 2200 m.y. Consequently, they belong to the Trans-Amazonian orogenic cycle of Almeida et al.(1973). The Guaporé platform and the Guyana shield are separated by the Paleozoic Amazon Basin. The Guaporé platform is separated from the São Luiz cratonic area by the Paleozoic Parnaiba Basin.The basement rocks of the Parnaiba Basin have yielded radiometric ages of 450-700 m.y. and consequently are within the time span of the Brasiliano orogenic cycle of Almeida et al. (1973). These Brasiliano-age rocks extend eastward into the exposed northeastern Brazilian shield rocks that have also given radiometric ages in the 450-700 m.y. range

(Fig. A1, Almeida et al., 1973). The Brasiliano orogenic cycle in Brazil corresponds to the Pan-African thermotectonic event of Kennedy (1965) in Africa.

The belt of Brasiliano-age rocks between the São Luiz cratonic area and the Guaporé platform narrows toward the Marajó Island and probably was connected with the narrow belt of Pan-African age rocks that parallel the littoral of Liberia (Figs. A1 and A2).

The radiometric boundary of the São Luiz cratonic area coincides with a prominent gravity and tectonic high (Ferrer-Urbano Santos Arch), and mylonitization zones may characterize the southern border of the cratonic area (R.Sadowski, personal communication, 1975). Thus, the border of the São Luiz cratonic area is similar in some aspects to the Accra Fault in Ghana, that was discussed earlier.

The northeastern Brazilian shield is reasonably well studied and contains important tectonic lineaments. The Patos-Paraiba and Pernambuco lineaments (Kegel, 1965; Ebert, 1966 and 1967; DNPM,1974), are long and narrow east-west zones of brecciation, mylonitization, metamorphism, magmatic emplacement and mineralization of Precambrian age, with a complex geological history of several reactivation periods. These lineaments generally delineate shield regions of distinct orientation of the foliations (Fig. A3).

The Patos lineament was reactivated during the Mesozoic as seen by small Cretaceous sedimentary basins preserved with down-faulted blocks that are either associated with the main fault zone or with subsidiary faults of the lineament (Fig. A3). The possibility that the Patos lineament may reach the littoral zone close to the city of

João Pessoa as an east-west fault zone is improbable because the structures of the shield curve northeastward to the east of the easternmost point from where the lineament is clearly present (Figure A3).Recently, Brito Neves (1975) in his discussion of the geotectonic framework of northeastern Brazil showed that the Patos lineament (a dextral strike-slip fault) bifurcated in two closely spaced northeast-southwest dextral strike-slip faults that reach the region close to the littoral,to the south of the city of Natal (Fig-A3). On the African side, no geographic counterpart of the Patos lineament has been reported in the literature indicating that most probably the lineament has its greatest representation confined to the northeastern Brazilian shield.

The Pernambuco lineament, with a mapped extension of about 600 km in the northeastern Brazilian shield and believed to represent a dextral strike-slip fault (Brito Neves, 1975), was also reactivated in the Mesozoic. This reactivation is demonstrated by the tectonically controlled northern border of the Mesozoic Tucano-Jatobá Basin (Fig. A1). The lineament is essentially east-west and its outcrop area ends up in the vicinity of the city of Recife. In the littoral zone, along the continuation of the strike of the Pernambuco lineament there is a unique magmatic province in Brazil. This corresponds to the Cabo de Santo Agostinho magmatic province (Cobra, 1967; Mello and Siqueira, 1972) that contains the socalled Cabo granite dated as old as 90 m.y. (Cordani, 1970).

Other important Precambrian lineaments in the northeastern Brazilian shield are the Sobral-Pedro II and the Jaguaribe (Fig.A3;Kegel, 1965). The Sobral Pedro II lineament has a northeast-southwest strike and extends from the Paleozoic Parnaiba Basin, where important normal faults are present (DNPM, 1974), into the exposed shield, where it separates two distinct geotectonic areas (Costa et al., 1973, in DNPM, 1974). Folded structures in Cretaceous sediments are in the strike of the Sobral-Pedro II lineament in the continental shelf (Ojeda and Ojeda in Ponte and Asmus, 1976).

The Jaguaribe lineament (Kegel, 1965) is a north-northeast-south-southwest structures that curves pronouncedly towards the west, in its southern portion. Small Cretaceous sedimentary basins are preserved in down-faulted blocks associated with this lineament. The Jaguaribe lineament is a geotectonic boundary between distinct shield blocks (Kegel, 1965) and extends northeastward towards the littoral where it becomes the western border of the Potiguar Basin (Fig.A3).

A Mesozoic lineament of east-west direction and cutting almost at right angles the foliation of the shield is present immediately to the south of Potiguar Basin. This lineament, represented by a set of parallel dolerite dikes, was called Cabugi lineament by Santos (1968).

Intracratonic Basins - The Amazon Basin contains Paleozoic sediments from the Silurian to the Permian that are generally covered by Cenozoic sediments (De Loczy, 1966). The Mesozoic is only represented in the basin by Cretaceous rocks (Bigarella, 1973). The Amazon Basin extends from the Atlantic Ocean to the foothills of the Andean belt and is structurally divided by arches in three major basinal areas: Upper Amazon; Middle Amazon; and Marajó. The Marajó Basin is separated from the Mid-Amazon Basin by the Gurupá Horst (Fig.A1;Aguiar and others, 1969; Rezende and Ferradaes, 1971). The Mid-Amazon Basin contains 7000m of sediments whereas the Upper Amazon Basin has only 3000m (Melo, 1960, in Bigarella, 1973). The Amazon Basin contains intrusives and volcanic flows of basaltic rocks that range in age from 170 to 293 m.y. according with 4 dated well samples (Bigarella,1973).

Structurally, the Amazon Basin is dominated by normal faults dipping towards the center of the basin (DeLoczy, 1966). Because no folding episode has been reported in the Amazon Basin, the basin has been interpreted as an extensive and complex graben (DeLoczy,1966),an autogeosyncline (Morales, 1960, in Bigarella, 1973) and a rift basin (De Boer, 1966, in Bigarella, 1973).

The Amazon Basin is separated from the Parnaiba Basin by the Tocantins Arch (Fig.A1; Rezende and Ferradaes, 1971; Bigarella, 1973). The oldest sediments in the Parnaîba Basin are Silurian (or? Ordovician) rocks that lie unconformably on Precambrian rocks (Mesner and Wooldridge, 1964). The Paleozoic is generally well represented.In contrast with the Amazon Basin, the Mesozoic of the Parnaîba Basin is well represented by Triassic, Jurassic and Cretaceous sediments (Bigarella, 1973).The Paleozoic sequence (2500m) is considerably thicker than the Mesozoic (500m) (Mesner and Wooldridge, 1964). The Parnaiba Basin underwent erosion during most of Cenozoic.

Basaltic flows and diabase intrusives of an early Cretaceous age (127 m.y.) are widespread in the Parnaiba Basin. The main centers of volcanism and diabase intrusions are apart from each other (Mesner and Wooldridge, 1964). The basaltic flows unconformably overlie a probably Jurassic continental formation and are overlain by a Cretaceous sedimentary sequence (Bigarella, 1973). The maximum thickness of the basaltic flows is 150m. The total thickness of diabase intrusions may reach 400m (Bigarella, 1973).

The Parnaība Basin is separated from the marginal basins of São Luiz and Barreirinhas-Piauī by the Mesozoic (Jurassic-Cretaceous) Ferrer-Urbano Santos Arch (Fig.A3; Mesner and Wooldridge, 1964; Rezende and Pamplona, 1970).

OCEANIC ISLANDS OF THE EQUATORIAL ATLANTIC OCEAN

Oceanic Islands of the Gulf of Guinea

Fernando Poo

Fernando Poo Island is situated in the continental shelf off Mount Cameroon highlands. The continental shelf edge off Cameroon curves around the island.

Fernando Poo Island is elongated in a northeast-southwest direction and it has a polygonal shape (Fig. A4). It is 70 km in length and is as wide as 37 km. Santa Isabel (3008m), San Carlos (2260m) and Biao(2009m)volcanoes constitute the bulk of the topography of the island (Hedberg,1968).Santa Isabel is a well-shaped volcanic cone.The other two peaks are volcanic calderas whose summits are aligned in an east-west direction transverse to the Cameroon Trend (Hedberg, 1968).

Fernando Poo Island is entirely constituted by volcanic rocks represented by cones, flows, pyroclastics and agglomerates. In contrast with other islands of the Cameroon Trend, no trachytes or phonolites have been reported and the rocks of the island are essentially basaltic in composition (olivine basalts - the most abundant; basalts;picritic basalts, oceanites and ankaramites)(Hedberg, 1968; Mitchell-Thomes, 1970; Baker, 1973).

The age of volcanism in the Fernando Poo Island is very recent. A radiometric age of approximately 1.1 m.y. is reported by Hedberg (1968) from a sample probably associated with the Santa Isabel volcanics. According to Fuster Casas (1954) (in Hedberg, 1968) the Santa Isabel volcanics are older than those of San Carlos and Biao. Consequently, the volcanics that now outcrop in the island may not be older than Pliocene.

The only evidence of pre-volcanic sediments in the island of Fernando Poo comes from a single boulder of limestone with an ammonite fossil that was found in a road construction south of the city of Santa Isabel. Hedberg (1968) thinks that most probably the block is a piece of pre-volcanic (Cretaceous?) sediment that was incorporated the ascending magma.

Principe Island

Raising from a continental rise 2500-3000m deep, Principe Island constitutes a small outcropping (19 x 15 kms) area of a much larger seamount (Fig.A5). Indeed, Principe Island has a large continental shelf in contrast with the other islands in the Gulf of Guinea.The island, if considered extended to the 200m isobath, is roughly elongated in a northeast-southwest direction.

Topographically, Principe Island is constituted by a low plateau areain the north (100-150m) and by a mountainous region in the south. This mountainous region is formed by volcanic necks and flows which reach a height of 948m at Pico do Principe. Pico do Principe is a volcano and occupies a central position along an east-west trending range formed by other peaks such as the Mencorne (937m) to the east and Carriote (840m) to the west (Hedberg, 1968; Fig.A6).

Principe Island is a volcanic island that presents two basic petrologic series. An alkalic-calcic series, consisting of olivine-trachy-andesites and olivine-basalts, is mostly present in the northern half of the island. An alkaline series consisting of a larger variety of types, such as nepheline and sodalite-phonolites, tephrites and several varieties of nephelinites dominates the southern half of the island (Hedberg, 1968; Mitchell-Thomes, 1970; Fig.A6). The basaltic series predated the phonolitic outpourings.

Fractures are generally oriented in an E-W, N-S and NW-SE trend. Less often the fractures have a NE-SW orientation. Several dykes of basalts and trachytes showing a NE-SW direction have been reported (Mitchell-Thomes, 1970).

Two separate occurrences of shallow-water limestones (30-70m) have been registered in Príncipe Island. They all occur as loose blocks on streams and no outcrop "in situ" has been found (Hedberg , 1968). The limestones contain micro and macro-fossils indicative of an early Miocene age and their fossil assemblage is very similar to the Miocene in the Lisbon and Algarve regions of Portugal (Teixeira,1955; in Mitchell-Thomes, 1970). The limestone blocks were found in elevations of 70m, 130m and 900m in three different sites and at the waterfront in Santo Antonio (Mitchell-Thomes, 1970; Hedberg, 1968). Because of the loose character of the limestone boulders, the similarity of the fauna and flora with the Miocene of Lisbon and Algarve,the absence of contact metamorphism in the blocks, the occurrence of limestone blocks in Santo Antonio waterfront with boulders of granites, phyllites, schists, etc... that were clearly used as ballasts by old trading vessels, Hedberg (1968) interpreted these limestone blocks as not indigenous to the island.

Mitchell-Thomes (1970), although presenting roughly the same evidence as Hedberg (1968), considered the limestones "in situ" resting on basalts and because of that he considered the volcanics of the island as of a pre-Miocene in age (Cretaceous?). The several altitudes of the outcrops were considered to be further evidence for differential uplift in the island after the Miocene. The present author considers Hedberg's arguments very strong, especially because he visited the outcrop areas of limestone and did not find any contact with the volcanics or conclusive evidence of "in situ" occurrence. Consequently, the post-Miocene uplift of some 160 to 200m for the Principe Island on the basis of the Miocene limestone evidence, as proposed by Mitchell-Thomes (1970) is doubtful. Nevertheless, recent vertical movements are suggested by the strong dip (40ºS) of a thin lignite bed intercalated with clays, as reported by Neiva (in Mitchell-Thomes, 1970) between Praia Grande do Norte and Santana; and a raised beach deposit dipping 9º to SSW.

Oil seepages in fractured volcanics have been reported in two localities in Principe Island (Hedberg, 1968; Mitchell-Thomes, 1970; Baker, 1973) and may be evidence that the island pedestal may be not only constituted by volcanics but also by sediments.

No radiometric dates are reported for the rocks of Principe Island.

São Thomé Island

São Thomé Island is located to the southwest of Principe Island and belongs to a complex seamount that rises from the continental rise of Gabon from approximately 3000 meters (Fig.A5). At least 3 peaks that do not form islands are outside of the shallow platform of São Thomé and share the foundations of the island. Between São Thomé and Annobon Islands, another seamount is present and probably corresponds to an extension of the high basement of São Thomé Island (Fig. A5).

São Thomé Island is the second largest island of the Gulf of Guinea (47 x 28 km). It has a rugged topography in which the "pico" São Thomé is the highest elevation (2,024m).

São Thomé is a volcanic island as seen by volcanoes present in the northern, central and southern parts of the island. Volcanic cones, bombs, lapilli, volcanic sands, ropy lavas and tuffs are widespread.

Phonolite necks and plugs, partially eroded, form prominent peaks (Mitchell-Thomes, 1970). The island is mostly constituted by olivine-basalts which predominate in the northern part (Fig.A7). Phonolites are most common in the central and southern part. Subordinate trachytes and andesites were mapped by Cotello Neiva (1956),(in Hedberg, 1968). Assunção (in Mitchell-Thomes, 1970) concluded that the volcanics of São Thomé are principally alkaline (sodic) but also include calc-alkaline types similar to rocks studied in western Cameroon by Jérémine (1943). These rocks have been dated radiometrically by Hedberg (1968). The K-A_r age determinations in a total of 8 samples yielded an age range of 0.1 to 3.0 m.y. However, data from Grunau and others (1975) using whole-rock K-A_r showed a 15.7 $\pm$ 0.8 m.y. age for an albitized trachyte that constitutes the oldest recorded magmatic rock in São Thomé to date. There is no evidence of volcanic activity in historic times at São Thomé Island. Raised beaches are found in many localities. In one locality, beach pebbles and sands were found 10m above sea level (Mitchell-Thomé). These raised beaches represent the most recent vertical uplift of São Thomé and, consequently, demonstrate that the island has been tectonically active in very recent times.

An occurrence of quartziferous sandstones and shales has been mapped by Hedberg (1968) in the island of São Thomé (Fig.A7).The sandstones outcrop in less than 1 km^2 and are overlain by a trachytic flow (Fig. A8). These sandstones were named Ubabudo formation by Hedberg (1968). The Ubabudo formation consists of a lower unit 160m thick of coarse basal conglomeratic sandstones and sandstones. These sandstones are overlain, presumably conformably, by an upper unit consisting of intercalations of shales and sandstones (115m). The lower unit lacks fossils with the exception of poorly preserved radiolarians and external molds of small brachiopods. These fossils, however, were not diagnostic enough to be dated. The heavy-mineral assemblage of the lower sedimentary unit indicates a granitic-gneissic source area. Hedberg (1968) pointed out the similarities of these lower units, with respect to the heavy-mineral assemblage and primary structures, with a basal sandstone formation of Cretaceous age in Gabon and with massive sandstones of the Cameroon that are also early Cretaceous in age, Because of that, Hedberg tentatively considered the lower unit as early Cretaceous in age.

The upper unit is a sequence of alternating fine to very fine-grained, quartz-sandstones and shales or claystones, The sandstones are somewhat finer grained than the lower unit. The claystones are generally very thin bedded and siltic in part. Fossil from this unit included radiolarians and arenaceous foraminifera. No pollen has been found. Based on this fauna, the tentative age of this unit was considered to be Cretaceous to Tertiary (Hedberg, 1968).

Grunau and others (1975) interpreted somewhat differently the Ubabudo formation of Hedberg (1968). The lower unit was thought to be possibly of Cretaceous age. The upper unit was interpreted as bedded tuffaceous sediments of Tertiary and presumably post-Paleocene age, and of deep marine (abyssal) origin.

Hedberg (1968) concluded from the primary structures and the textures of the Ubabudo formation that its lower unit most probably was of a very shallow-water origin. Houbolt (1973) described the Principe channel with respect to its morphology and sediments, and, after visiting São Thomé Island, pointed out the similarities in primary sedimentary structures of the sandstones and shales of the São Thomé Island with the sands and muds associated with the floor of the Principe channel and other canyons in the Gulf of Guinea (J.J.H.C. Houbolt, personal communication 1975).

The continental provenance of the quartziferous sandstones of the São Thomé Island cannot be questioned. The presence of radiolarians and virtually no pollen in the lower section of the Ubabudo formation is difficult to correlate with the fluvio-lacustrine deposits of early Cretaceous age in Gabon and Cameroon. Cores in the Principe channel were mostly constituted by sands. In other areas of the continental rise, the occurrence of abundant sandy bottoms, such as in the Amazon Cone in Brazil, is common and suggest the possibility that the São Thomé sandstones represent continental rise deposits that were later uplifted by tectonism and magmatism. The tectonism is exemplified by the average high dips of the formation (50º) and by its highly fractured character as shown by the geological mapping of Hedberg (1968) (Fig. A8).

Oil seepages have been known in the island and this stimulated the interest of oil companies in the geological study of the island (Fig.A8). Considering the geographical setting of the island and that most of the São Thomé island is constituted by volcanic rocks, the oil seepages and the occurrence of quartziferous sandstones are a surprise. Therefore, São Thomé Island may represent not only a volcanic pile but rather a tectonic block that was uplifted at a time when it was partially covered by continental rise or even abyssal sediments. Magmatism was widespread along fractures in this block and eventually covered and constituted the bulk of the surface of the island.

Annobon Island

Annobon is the southernmost island of the Cameroon Trend and is 700 km southwest of Cameroon. It measures 6 km in a N-S direction and 3 km in an E-W direction. The island rises from an ocean floor 3500m deep and is entirely volcanic in origin (Fig.A5). Pico de Santa mina is the highest elevation of the island with 700m according to Mitchell-Tomes (1970). However, Hedberg (1968) claims that the Pico del Centro with 831m is the maximum elevation of the island (Fig.A9). Pico de Fogo (450m) is a bare volcanic cone with a small crater lake and corresponds to a trachytic plug. Quiveo is a crater (640m) and is located in the central part of the island (Mitchell-Tomes, 1970).

Annobon Island is entirely constituted by basaltic rocks that contain abundant olivine phenocrysts and rare plagioclases (Hedberg, 1968). Oceanites and ankaramites are present in minor amounts. A biotite-trachyte flow post-dating the basaltic series was reported by Tyrrel (1934) (in Hedberg, 1968). The oceanites and picritic basalts, although showing extreme basicity, have their content of alkalis sufficiently high so that the rocks can be classified as alkali-basalts of extremely basic type. Fuster Casas (1950, 1954) (in Mitchell-Tomes, 1970) stressed the strongly basic character of the Annobon rocks that are more ultrabasic than the rocks of the other islands of the Gulf of Guinea. Olivine nodules are common in the island and they consist of pure olivine or olivine, augite and magnetite.

No radiometric datings are available from the Annobon island. No historic volcanic flows have been reported and today the only evidence of volcanic activity is represented by springs of carbonated water (Hedberg, 1968). Judging from the volcanic landforms and structures still preserved in the island, Annobon may not be older than Miocene in age (Hedberg, 1968).

Ascension Island

Ascension Island is at 7°57'S and 14°22'W in a region about 100 km from the median valley of the Mid-Atlantic Ridge and 50 km south of the Ascension Fracture Zone (Fig.A10). The island represents the highermost portion of a pedestal roughly circular (50 km of diameter) that is based at 3000m depth and reaches an altitude of 1000m above sea level (van Andel and others, 1973; Fig. A10).

The Ascension Island is essentially volcanic in character with widespread basic to intermediate lava flows, trachytes and pyro clastic cones and deposits (Fig. A11). The petrographic types in the island are diversified and include basalts, trachydolerites, andesites (?), trachyandesites, trachytes and quart-trachytes (rhyolites) (Mitchell-Thomes, 1970). In general, basalts and rocks of allied types comprise about 85% of the island and rocks of trachytic composition constitute the remainder (Fig. A11; Mitchell-Thomes, 1970;Baker, 1973).

Plutonic rocks occur in Ascension Island as xenoliths.Blocks of dunites, peridotites and gabbros are confined to the Dark Slope Crater (Fig. A11). These blocks have the characteristics of cumulates and are considered to have tholeiitic affinities in contrast to the more alkaline nature of the lava flows (Atkins and Bell, 1967, in Baker, 1973). Xenoliths of intermediate composition are also present and correspond to pyroxene-hornblende-quartz-syenites, hornblende-syenites and monzonitic diorites. Xenoliths of alkaligranites were found in agglomerates around Green Mountain (Fig. A11). These granites that include alkali-amphibole-granite, alkali-hornblende-biotite-granite and biotite-granite, constitute isolated blocks of several sizes that occur on stream valleys (Mitchell-Thomes, 1970; Baker, 1973). Because the granites ans syenites greatly differed in chemical constitution and mineralogical content from the rhyolites and trachytes of the island, Daly (1925 in Mitchell-Thomes, 1970) was led to conclude that the granitic and syenitic fragments were representatives of an older basement on which the island cone probably rested. The obvious implication that the basement rocks were from a sialic basement in the middle of the ocean followed. However, Tilley (1950 in Baker, 1973) and Roedder and Coombs (1967 in Baker, 1973) explain the granitic blocks as subvolcanic equivalents of the pantelleritic lavas of the island. ?

Oceanic Islands of Northern Brazilian Margin

Fernando de Noronha

The Fernando de Noronha archipelago consists of a group of small islands in the close vicinity of the main Fernando de Noronha Island. They are situated 345 km from the littoral of northeastern Brazil and were thoroughly studied geologically by Almeida (1958). His work describes the geology of the island with respect to the petrography, geomorphology and stratigraphy in detail. Because of this, Fernando de Noronha Islands will not be discussed in detail here and the reader is referred to Almeida's work (1958).

Fernando de Noronha Islands constitute the summits of a much larger submerged mountain that raises from an ocean floor 4000m deep. The top of this mountain is essentially flat at depths shallower than 100m except for the islands themselves (Fig. A12). The seamount as viewed by the 4000m contour (Fig. A13) is roughly circular and its diameter is about 60 km. In depths shallower than 2500m the seamount is pronouncedly east-west oriented. Geomorphologically, the seamount is isolated from other seamounts of the Fernando de Noronha Ridge. The island of Fernando de Noronha has a total area of 18.4km^2.

The morphology of the archipelago is a mixture of low-altitude plateau areas, isolated peaks and a concentration of mountainous areas. Steep falesias, sometimes constituting flanks of mountains, and other times interrupting the low-altitude plateau areas, contrast with strings of sandy and gravel beaches that occasionally are in the landward side of fringing-reefs (Fig. A14). The highest elevation of the island is the "Morro do Pico" which rises to 321m and is the most spectacular landmark of the island (Fig. A15); Almeida, 1958). Fernan do de Noronha Islands are essentially volcanic in nature. A mixture of lava flows,tuffs and pyroclasts, hypabissal and superficial intrusives, plugs, domes and volcanic necks compose the structure of the islands (Almeida, 1958).

The volcanic phenomena that occurred on the island as described by Almeida (1958) started with a phonolitic-trachytic volcanism with ultrabasic intrusives (Remédios formation). This volcanism was followed by a widespread erosion that destroyed the early volcanic centers. The Quixaba formation followed and rested on the erosional surface of the Remédios formation.The Quixaba formation constitutes volcanic deposits from explosive volcanism, and from ankaratritic lava-flows; and intrusions of nephelinites. The Quixaba formation was followed by the lava-flows of nepheline-basanites of São José formation (Fig. A14).

Plutonic rocks do not outcrop in the Fernando de Noronha Islands but are only present as xenoliths in lava flows, in intrusives or in pyroclastic accumulations. In the Remédios formation xenoliths and ejecta of nepheline syenite, hornblende gabbro, hornblende pyroxenite and of other plutonic rocks were described by Almeida (1958). In the São José formation that is restricted to the islands of São José, Cuscús and de Fora, there are abundant xenoliths of peridotites (Figs. A14).

Besides volcanic formations, the Fernando de Noronha Islands have eolianites constituted by $CaCO_3$ sands, raised beach-terraces, mo dern dune fields, present-day sand and gravel beaches and present-day algal reefs that fringe a considerable part of the island.

There are no historical reports of either volcanic activity or earthquakes in the islands and there are no preserved volcanic co nes in the island.

Almeida (1958) tentatively considered Fernando de Noronha Is lands not older than Senonian. This dating was the basis for the age of 80 m.y. considered by Wilson (1965) for the islands. The first suc cessful attempt in dating Fernando de Noronha Islands produced an age of Miocene or younger on the basis of paleomagnetic data (Richardson and Watkins, 1967). Cordani (1970) in his work about the volcanism of the South Atlantic dated several rocks from Fernando de Noronha Islands. His studies in general supported Almeida's stratigraphic sequence but found that the rocks were much younger than previously supposed and that they varied in age from 1.8 to 12.3 m.y. The oldest date was 21 m.y. in a sample of nepheline-basanite from São José formation. Cordani (1970) did not consider the dating on this sample to be valid because of possible contamination from the xenoliths of peri dotite that are abundant in the formation and because other datings in the nepheline-basanites of São José Island yielded a consistent 8.1 to 9.3 m.y. age.

Although there are numerous dikes of intrusive rocks in Fer nando de Noronha Islands with different widths, shapes, petrographic constitutions and structures, Almeida (1958) pointed out the curious northeast-southwest preferred orientation of 70 dikes that intruded the Remédios formation. This northeast-southwest strike of extensional fracturing can be explained by localized stress systems originated from the magmatic intrusions and extrusions, at the time of Remédios formation. However, it can be also explained by a stress system in which the maximum compressive stress direction is given by the extensional fractures represented by the dykes. The least compressive stress is given by a direction perpendicular to it in the same plane. This stress system predicts a shear failure either at an east-west or at a north-south direction or both directions. The presence of the greatest axis of orientation of the pedestal of Fernando de Noronha Islands in depths shallower than 2500m and the overall orientation of the Fernando de Noronha Ridge in the east-west axis seem to support a shear direction in a left-lateral direction in an east-west orientation.

Because Fernando de Noronha Islands are an integral part of the ridge and fracture zone of the same name, it demonstrates that magmatic activity may take place in fracture zones in regions far away from the offset region of the axial segments of the Mid-Atlantic Ridge and shows that fracture zones are the most probable places to accommodate further tectonic and magmatic events.

Atol das Rocas

Atol das Rocas corresponds to two crescent-shaped islets constituted by algal carbonate banks that are crowned by calcareous eolianites (Ottmann, 1963). It is situated at 3º52'S and 33º49'W, off the coast of the Rio Grande do Norte state. The two islets ("Ilha do Farol" and "Ilha do Cemitério") together with a circular-shaped algal reef front enclose a very shallow lagoon (Fig. A16). As pointed out by Ottmann (1963), Atol das Rocas is not a coralline atoll of the Pacific Ocean type, but it is really an algal reef built on an abrasion platform.

The two small islands of the Atol das Rocas correspond to mounds of detrital carbonates that do not even reach 3m of altitude (Mabesoone and Coutinho, 1970). The detrital carbonates are constituted by fragments of calcareous algae with only a few other organisms (coral, bryozoan, plates and spines of echinoids and foraminifera)(Mabesoone and Coutinho, 1970). The algal species are chiefly of the genera Halimeda, Jania and Amphiroa.

A beach-rock in the Cemitério Island and pinnacles of dead algae lying 3-4m above the sea level (Fig.A16) suggest a former high sea-level stand at +2.5m (Mabesoone and Coutinho, 1970).

There are no occurrences of magmatic rocks in the Atol das Rocas. The continental shelf that surrounds the Cemitério and Farol Islands is relatively smooth and is generally covered by algal carbonates. These algal carbonates consist of algal crusts, algal sand with pebbles, fine algal sand and Halimeda sand (Mabesoone and Coutinho, 1970).

Atol das Rocas constitutes the highermost portion of a complex seamount that rises from a sea floor 4000m deep (Fig. A13). This complex seamount actually corresponds to two apparently isolated peaks that are joined by the 2500m bathymetric curve. Atol das Rocas is situated in the westernmost peak of the complex seamount. The entire submarine mountain is essentially aligned along an east-west direction and is an integral part of the Fernando de Noronha Ridge. A third seamount that is as shallow as 200m is present in the southern vicinity of the eastern peak of the complex seamount.

The absolute age of the Atol das Rocas has not been determined yet. Its relative age however can be inferred from geomorphological comparisons between Fernando de Noronha Islands and other seamounts associated with the Fernando de Noronha Ridge. Atol das Rocas is probably older than the Fernando de Noronha Islands and probably younger than the flat-topped seamounts of the Fernando de Noronha Ridge, mentioned by Ealey(1969). It is probably younger that the flat-topped seamounts to the west (seamounts IHGF of Ealey, 1969) because their depths are generally below the depth of effective wave action and the lower limits of glacial eustatic lowerings of sea level. Consequently, as Ealey (1969) concluded, the seamounts must have subsided to their present depths and this contrasts with the present-day sea level situation of the Atol das Rocas. It is probably older than Fernando de Noronha Islands because,by considering the wave-cut and algal-reef paved insular platform of Fernando de Noronha that is interrupted by the islands themselves, it is not difficult to visualize Fernando de Noronha Islands as a wave-cut platform and algal-reef covered in the geological future, when no more magmatic activity will take place. The concomitant and follow-up stage is the subsidence of the entire pedestal that places the wave-cut and reef platforms be- low the effective wave-action level and algal-growth depth. In this situation, the flat-topped seamount is the receptacle for pelagic sediments that tend to further enhance the flatness of the guyot and mask the subsidence of the seamount by progressive upward accumulation.

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FIGURE CAPTIONS FOR THE APPENDIX

Fig. A1→ Tectonic map of N-NE Brazil. Structures from DNPM (1971) and others.

Fig. A2→ Fit of Brazil and Africa following Bullard et al. (1965) and adjusting to the apparently best tectonic fit. The continents were considered extended up to the 1500m bathymetric curve. The overlaps (black areas) are mainly seen in southern Liberia (Amazon Cone) and Southern Ghana. Heavy dashed lines are boundaries between different radiometric provinces.

Fig. A3→ Tectonic map of N-NE Brazil with key for names used in the text. Structures are represented in the same way as of Figure A1.

Fig. A4→ Geographic map of Fernando Poo Island (after Mitchell-Thomes, 1970).

Fig. A5→ Various bathymetric and tectonic features in the Gulf of Guinea. Bathymetry in kilometers (uncorrected). Contour interval is 0.25 km in the continental margin, and 0.5 km in the area of the Mid-Atlantic Ridge.

Fig. A6→ Geologic map of Principe Island.

Fig. A7→ Geologic map of São Thomé Island.

Fig. A8→ Geologic map of the Ubabudo Area, São Thomé Island (after Hedberg, 1968).

Fig. A9→ Geographic map of Annobon Island (after Mitchell-Thomes (1970)).

Fig. A10→ Bathymetric map of the Mid-Atlantic Ridge in the region of Ascension Fracture Zone (after Van Andel et al., 1973). Contours in hundreds of meters.

Fig. A11→ Geologic map of Ascension Island (after Atkins et al., 1964, in Baker, 1973).

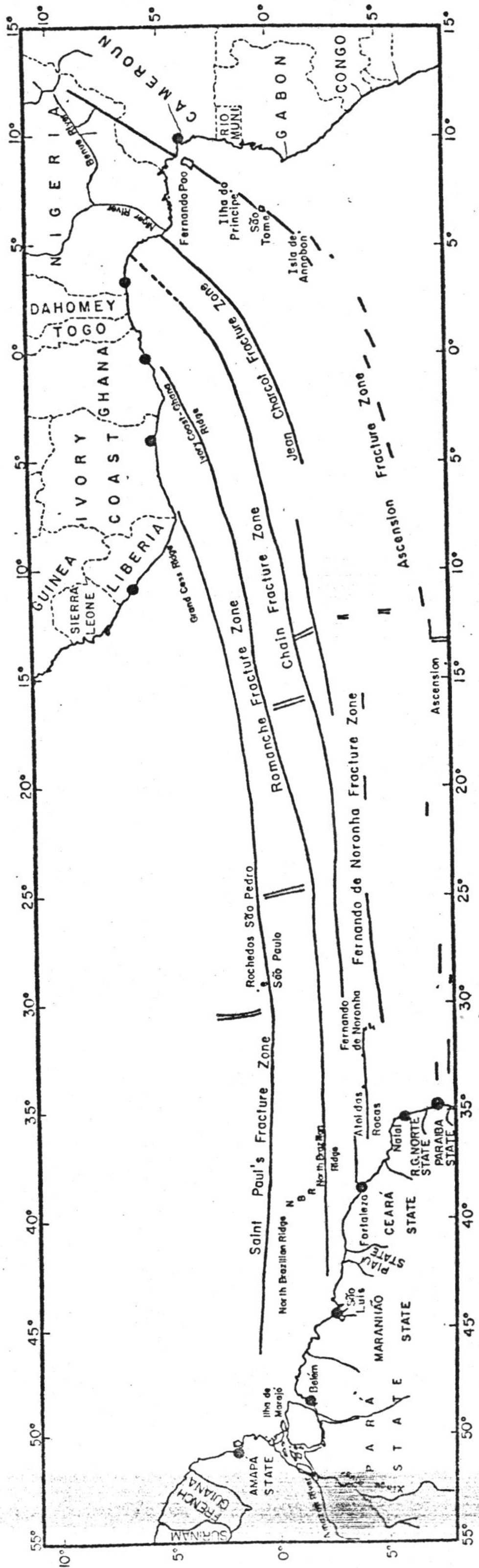
Fig. A12→ Topographic profiles of the pedestal of Fernando de Noronha Island (after Almeida, 1958).

Fig. A13→ Bathymetric and tectonic features off Northern Brazil. Bathymetry in kilometers (uncorrected). Contour interval is 0.25 km in the continental margin and 0.5 km in the area of the Mid-Atlantic Ridge.

Fig. A14→ Geologic map of Fernando de Noronha Islands (simplified after Almeida, 1958).

Fig. A15→ W-E aerial view of Fernando de Noronha Island. "Morro do Pico" is the peak in the foreground to the left. In the background "Morro da Boa Vista" is seen. (Photograph from the author).

Fig. A16→ Atol das Rocas seen in planar view and in cross-section (after Soares, 1968 and Ottmann, 1963).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Isla de Annobon
DAHOMEY
TOGO
GHANA
IVORY COAST
LIBERIA
SIERRA LEONE
GUINEA
Ivory Coast-Ghana Ridge
Grand Cess Ridge
Charcot Fracture Zone
Jean Charcot Fracture Zone
Romanche Fracture Zone
Chain Fracture Zone
Ascension Fracture Zone
Ascension
Fernando de Noronha Fracture Zone
Rochedos São Pedro São Paulo
Saint Paul's Fracture Zone
Fernando de Noronha
Atol das Rocas
North Brazilian Ridge
NBR
Natal
R.G. NORTE STATE
PARAIBA STATE
Fortaleza
CEARÁ STATE
PIAUI STATE
São Luis
MARANHÃO STATE
Belém
PARÁ STATE
Ilha de Marajó
AMAPÁ STATE
Amazon River
FRENCH GUIANA
SURINAM

Figure 1

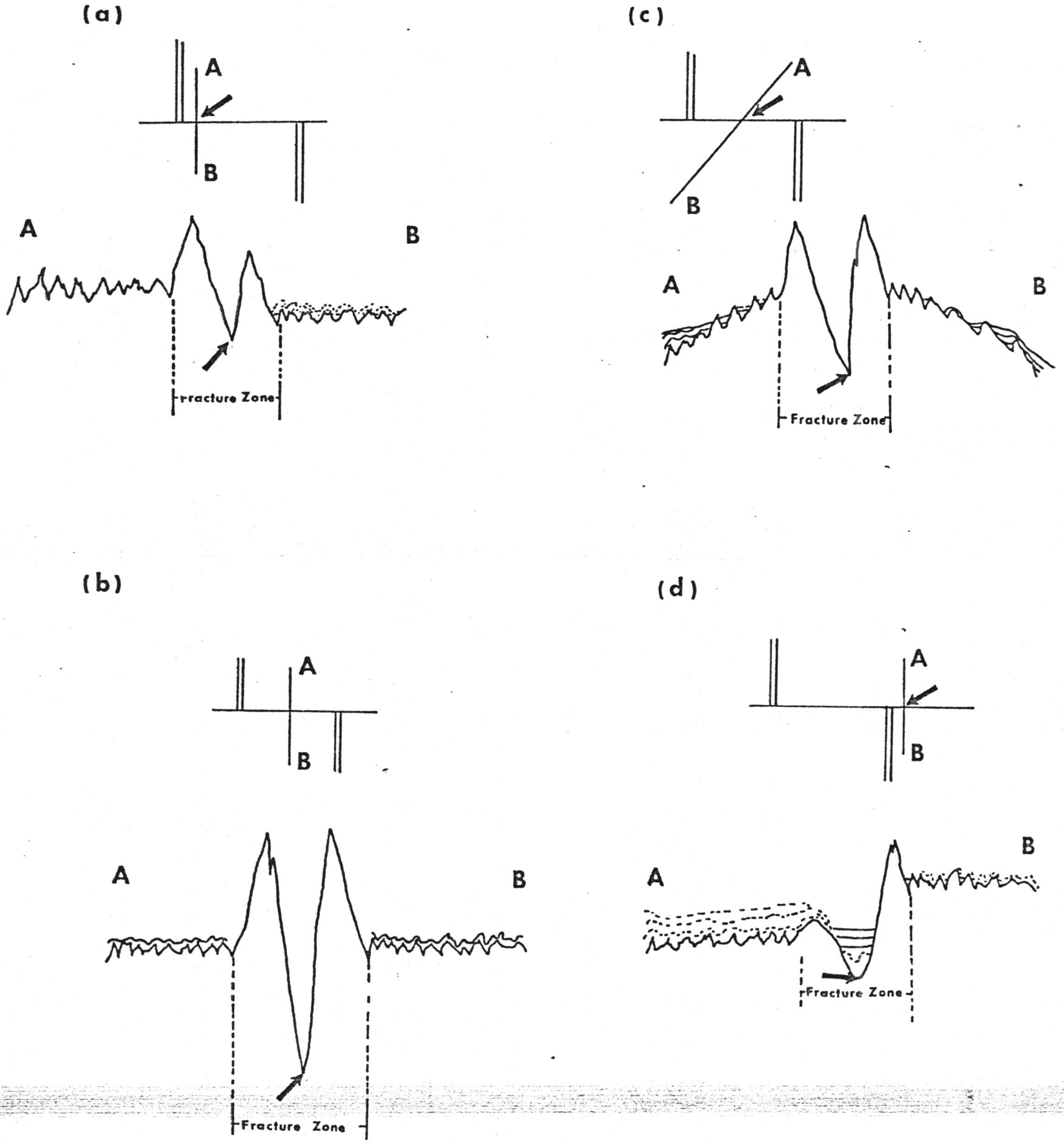

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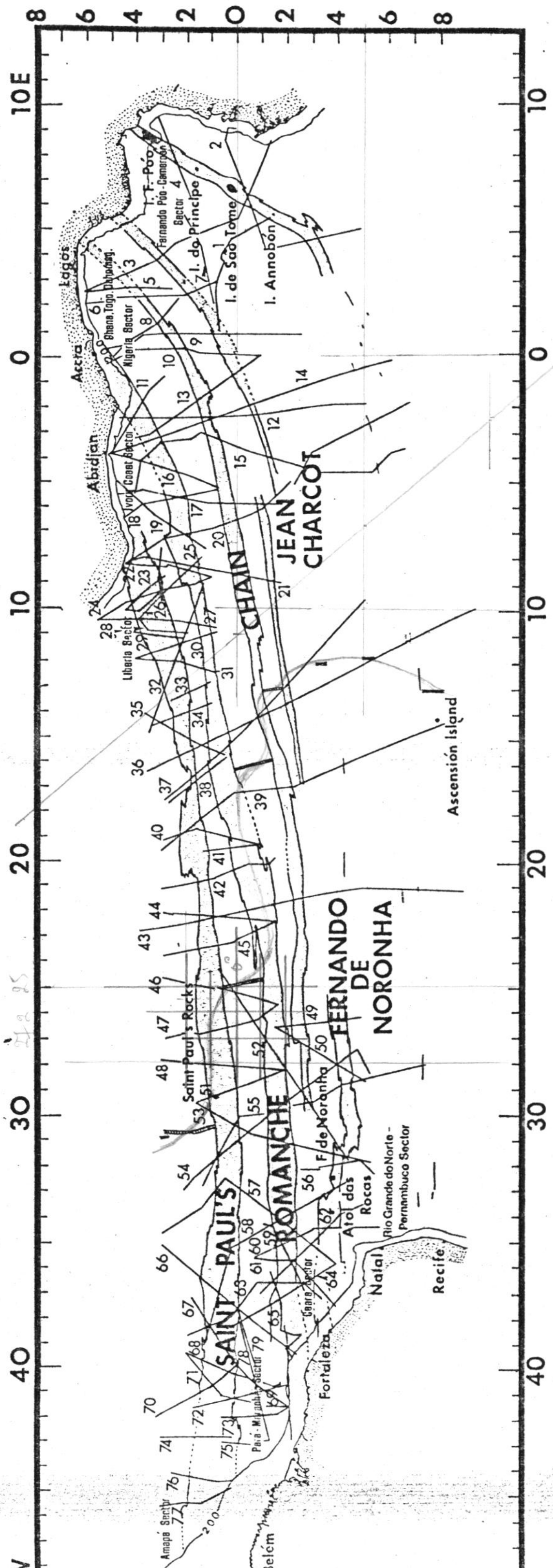
Figure 2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FERNANDO DE NORONHA
Ascension Island
Saint Paul's Rocks
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I. Annobón
I. do Principe
Atol das Rocas
F. de Noronha

Figure 3

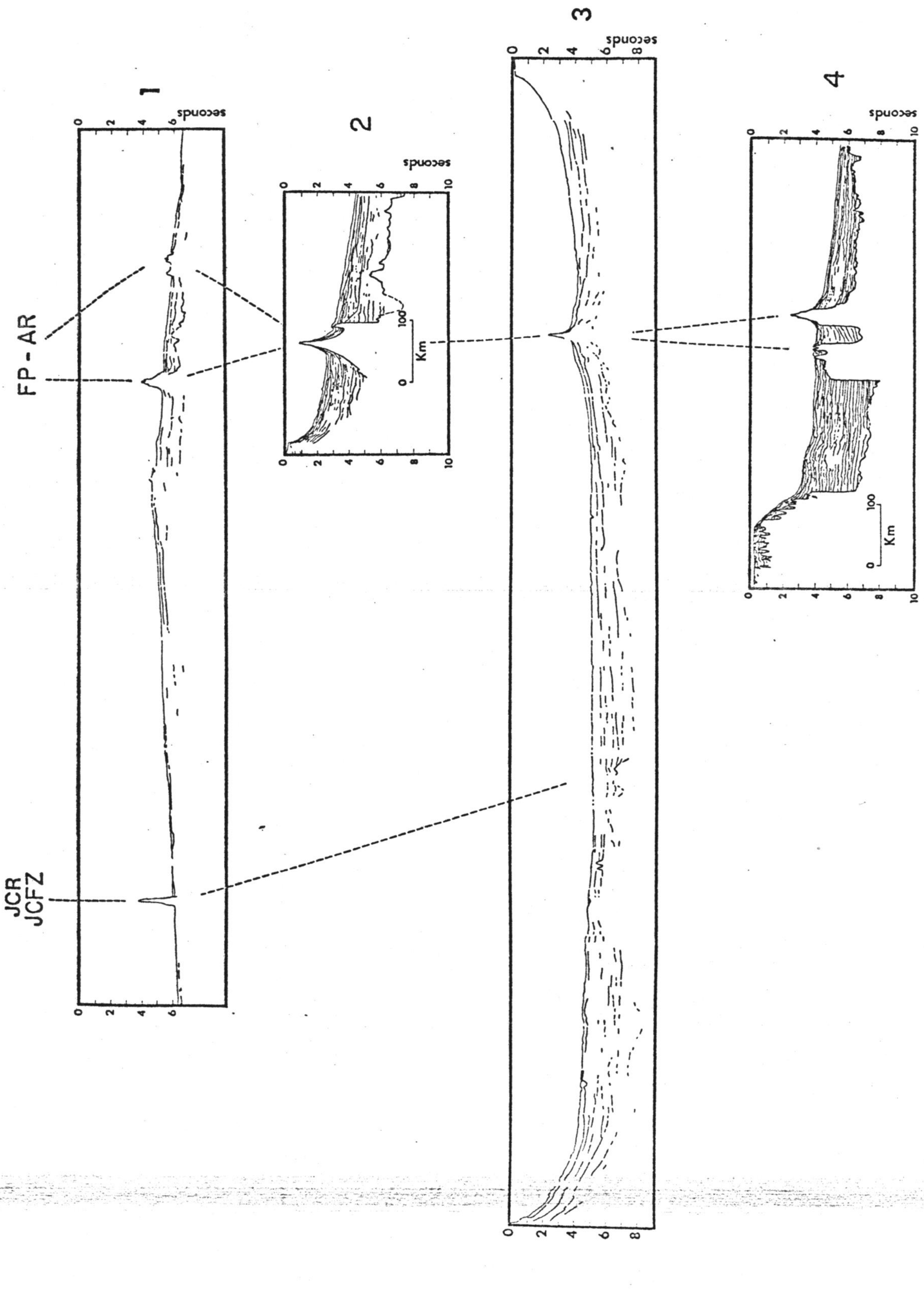
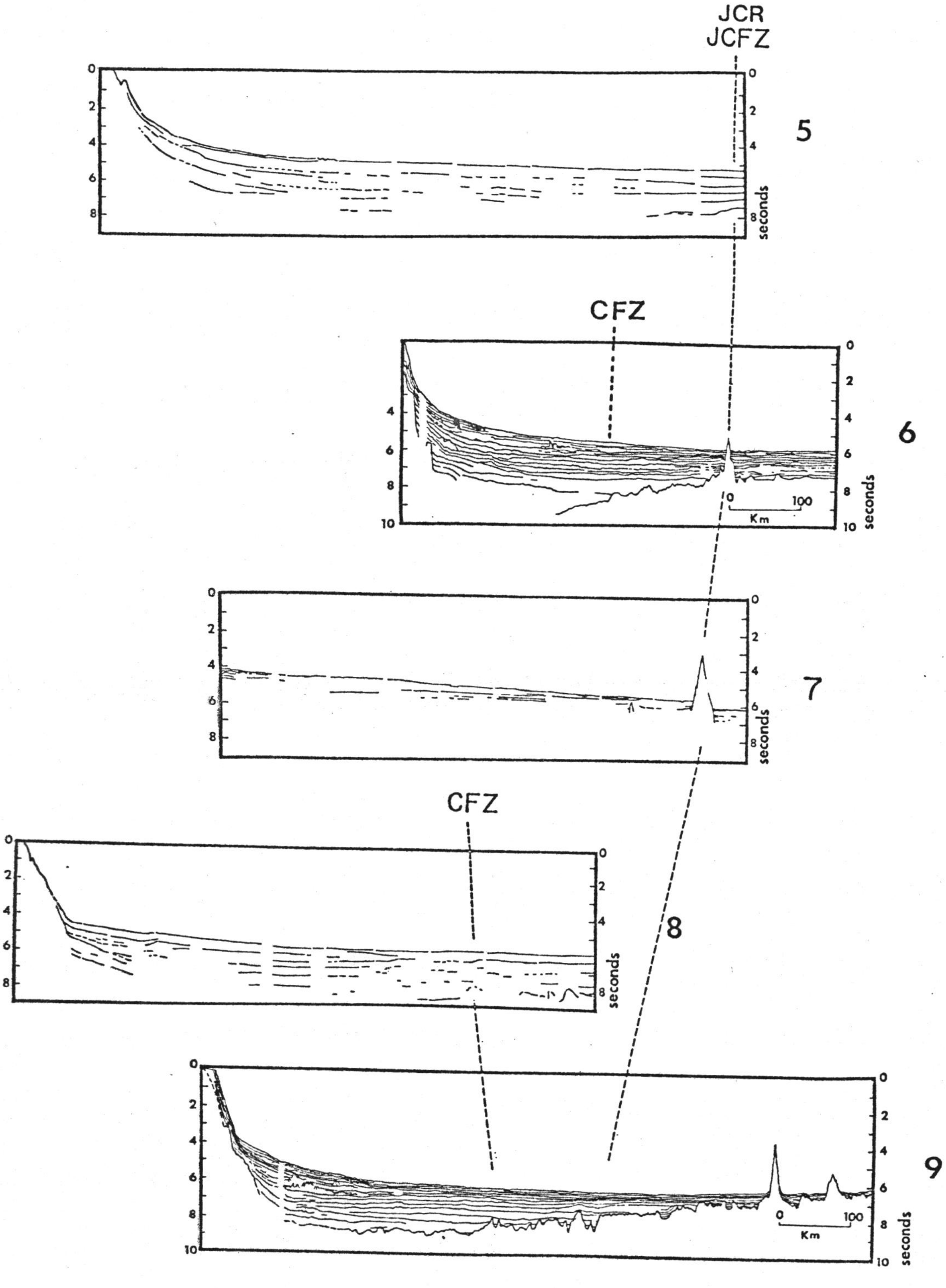


Figure 4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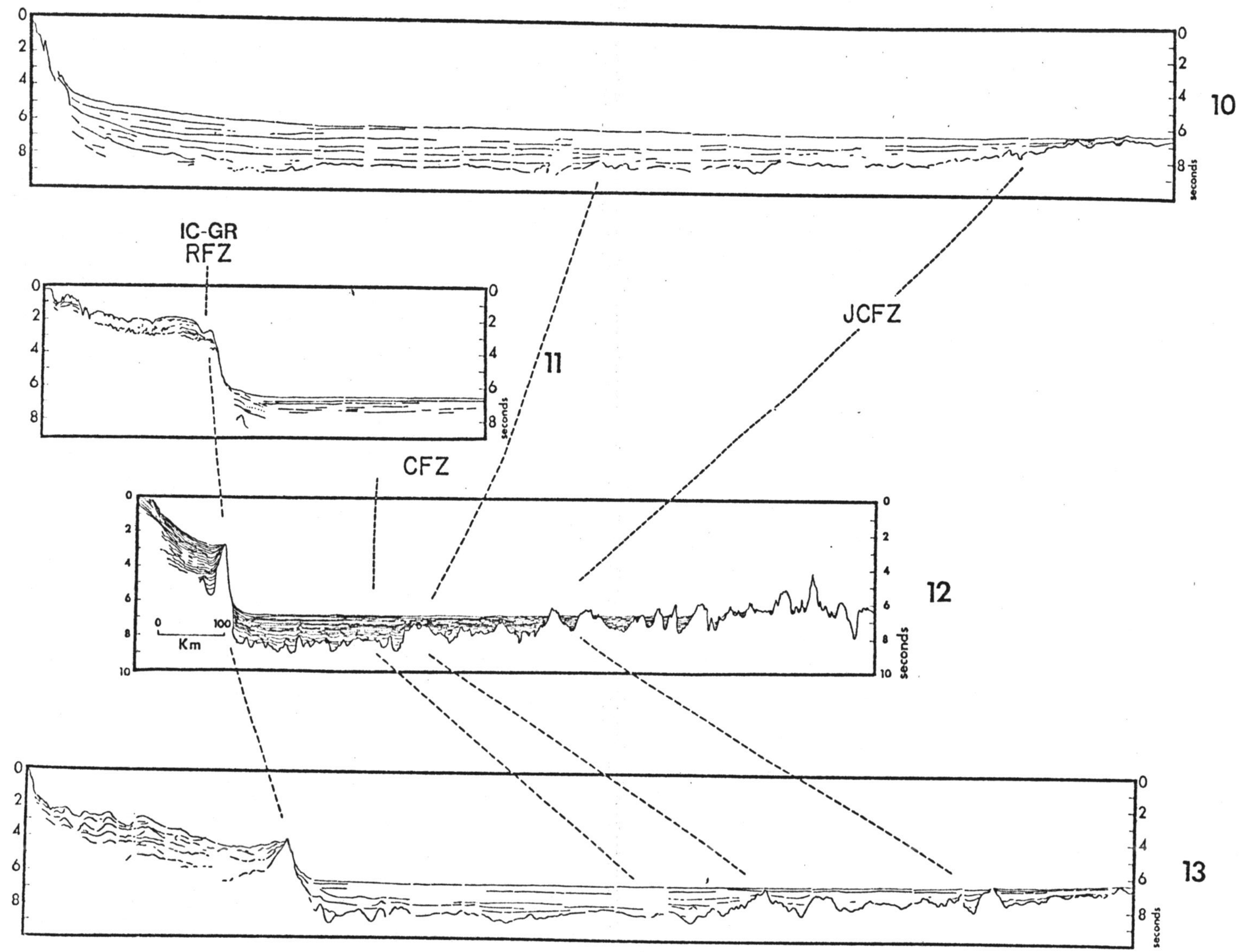
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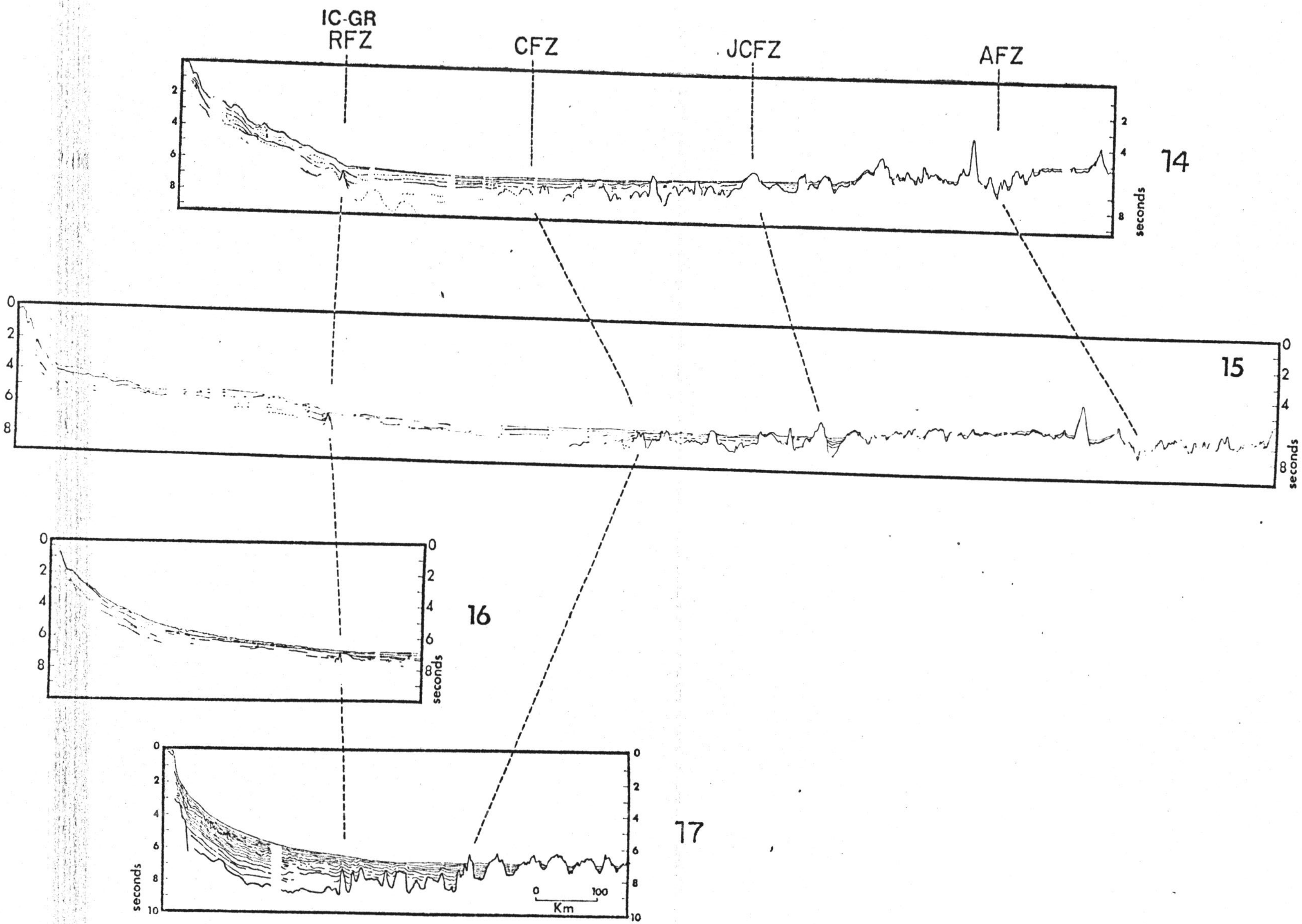
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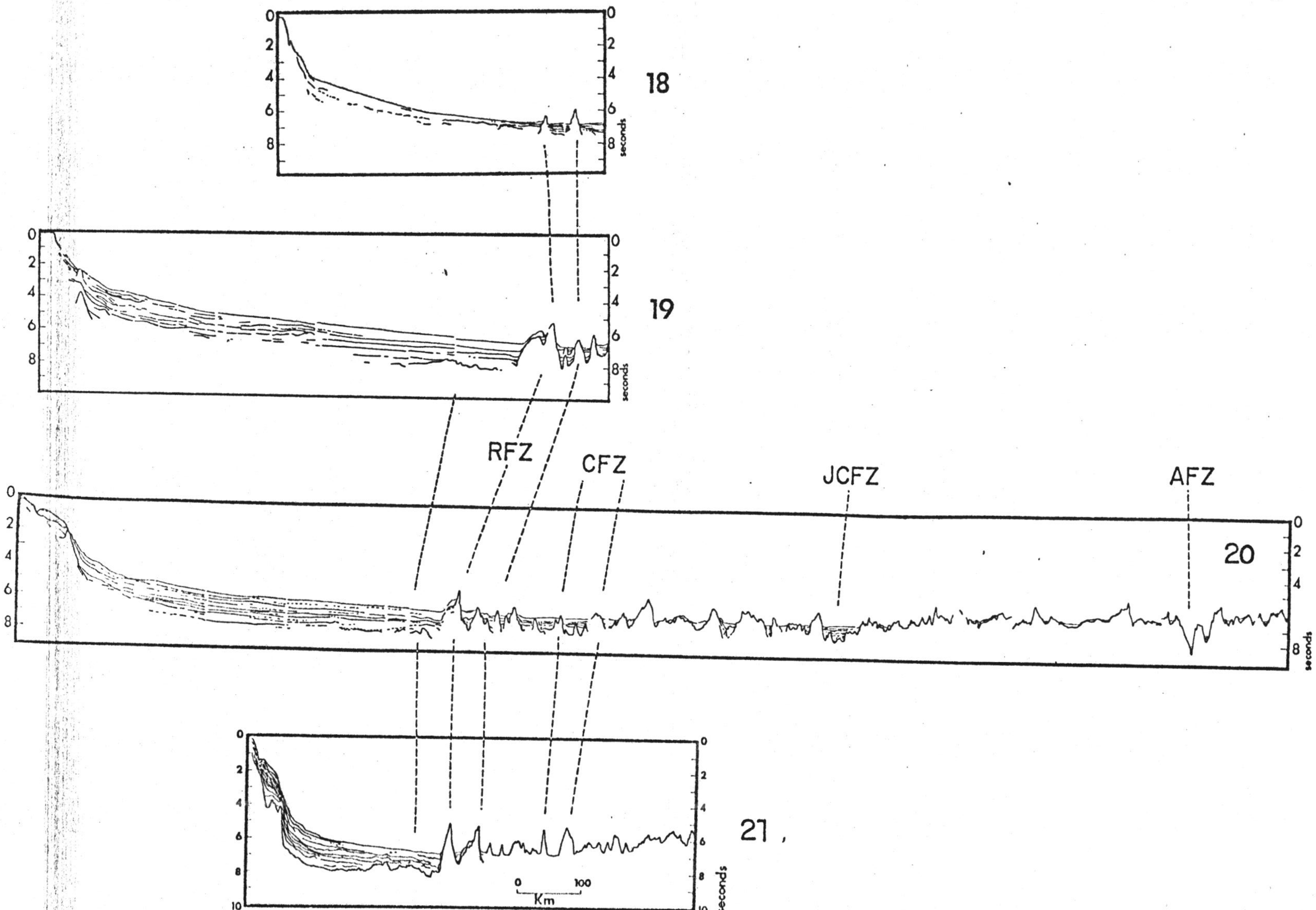
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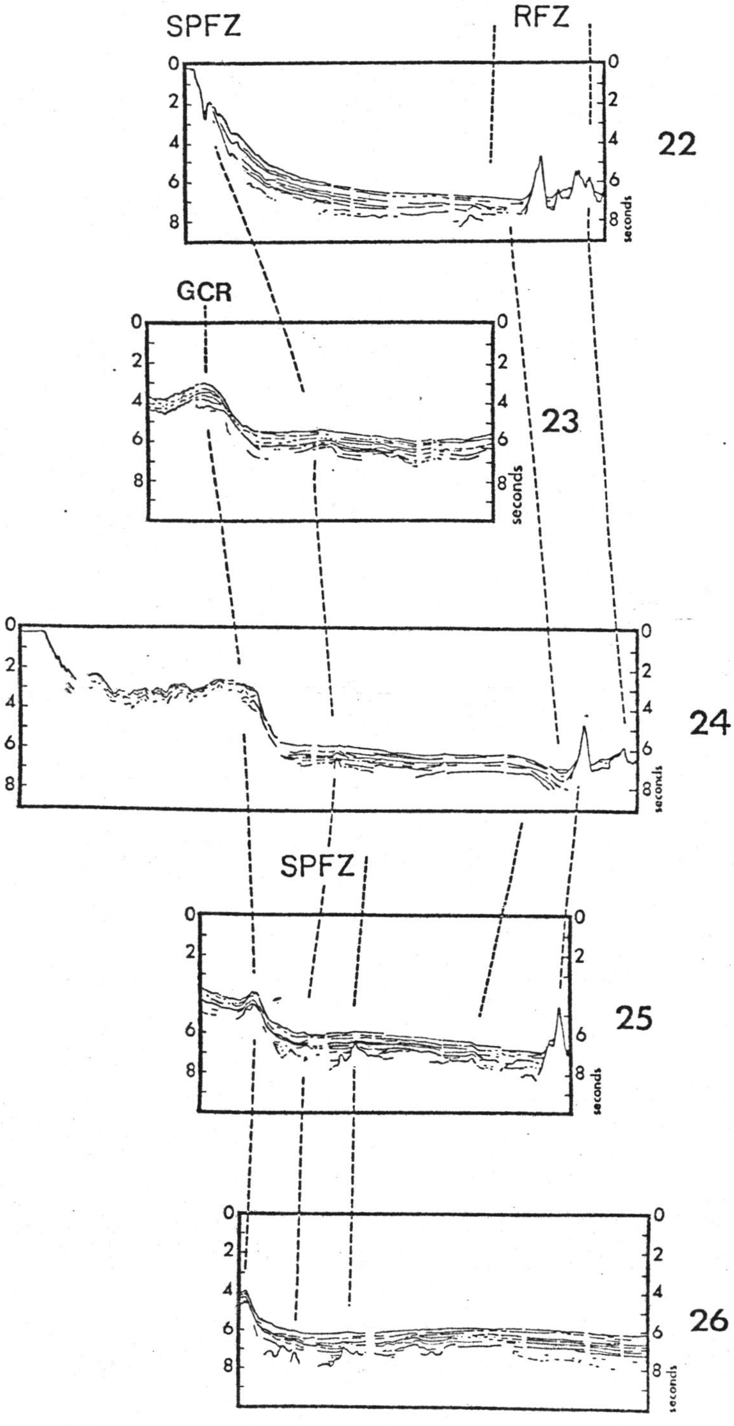
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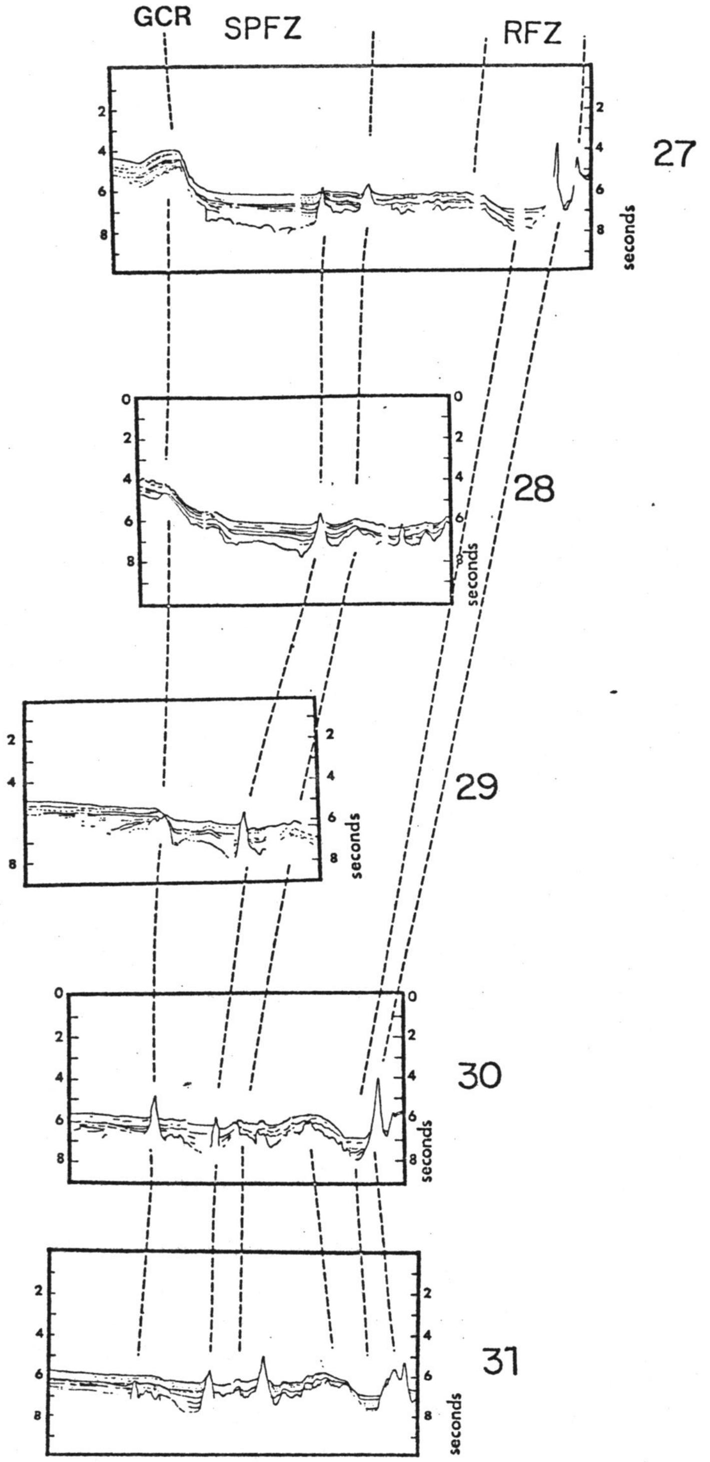

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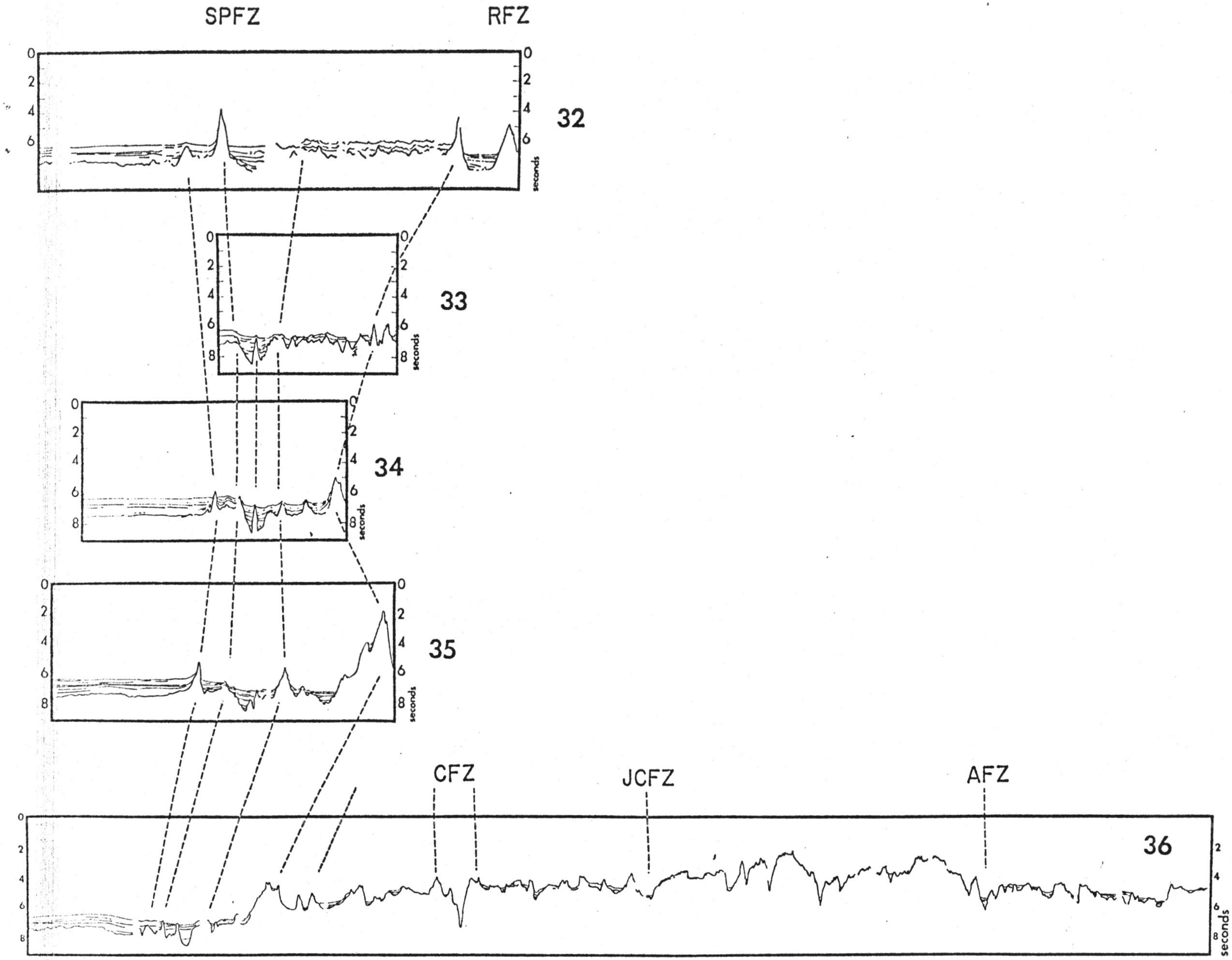

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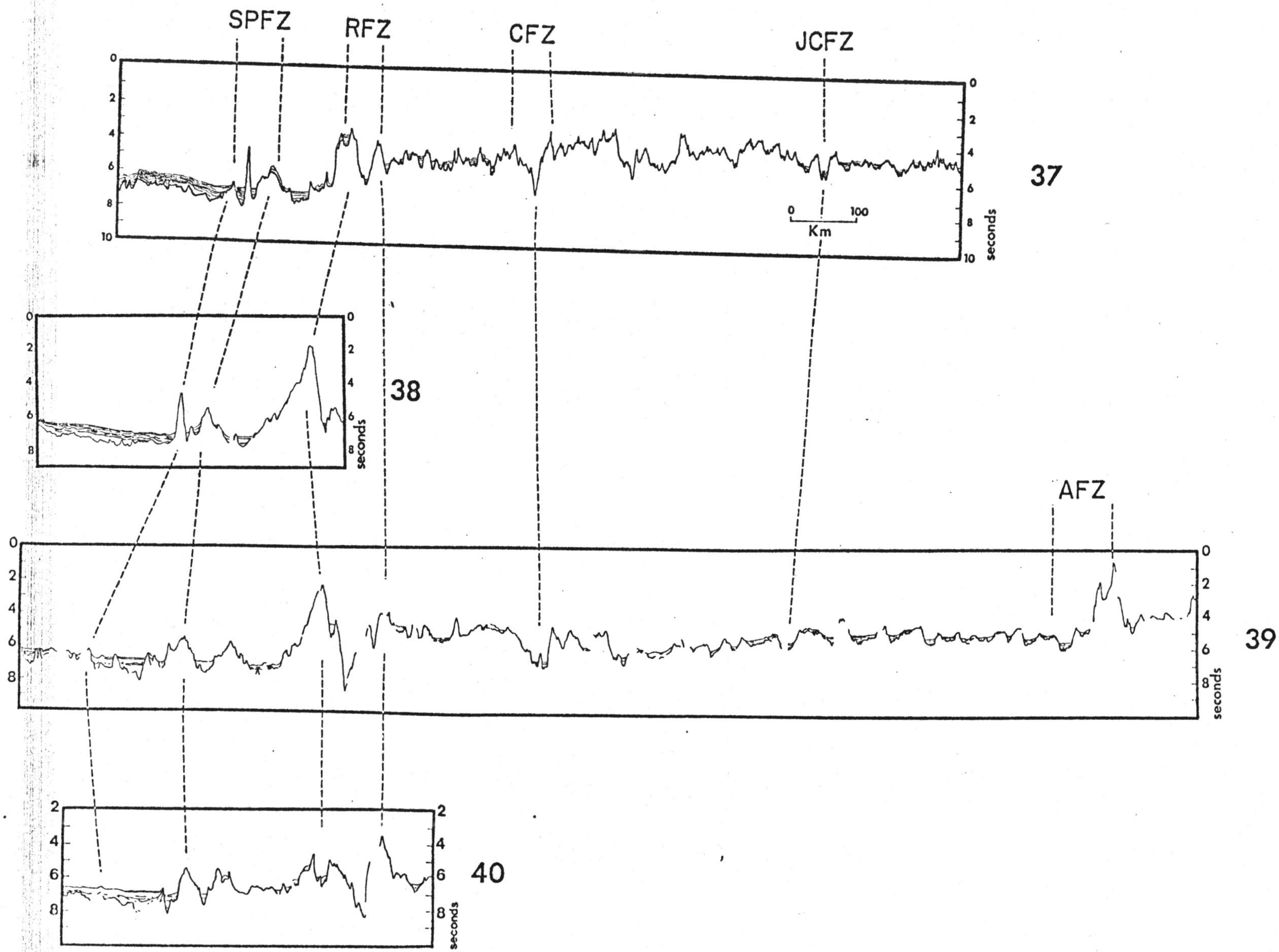
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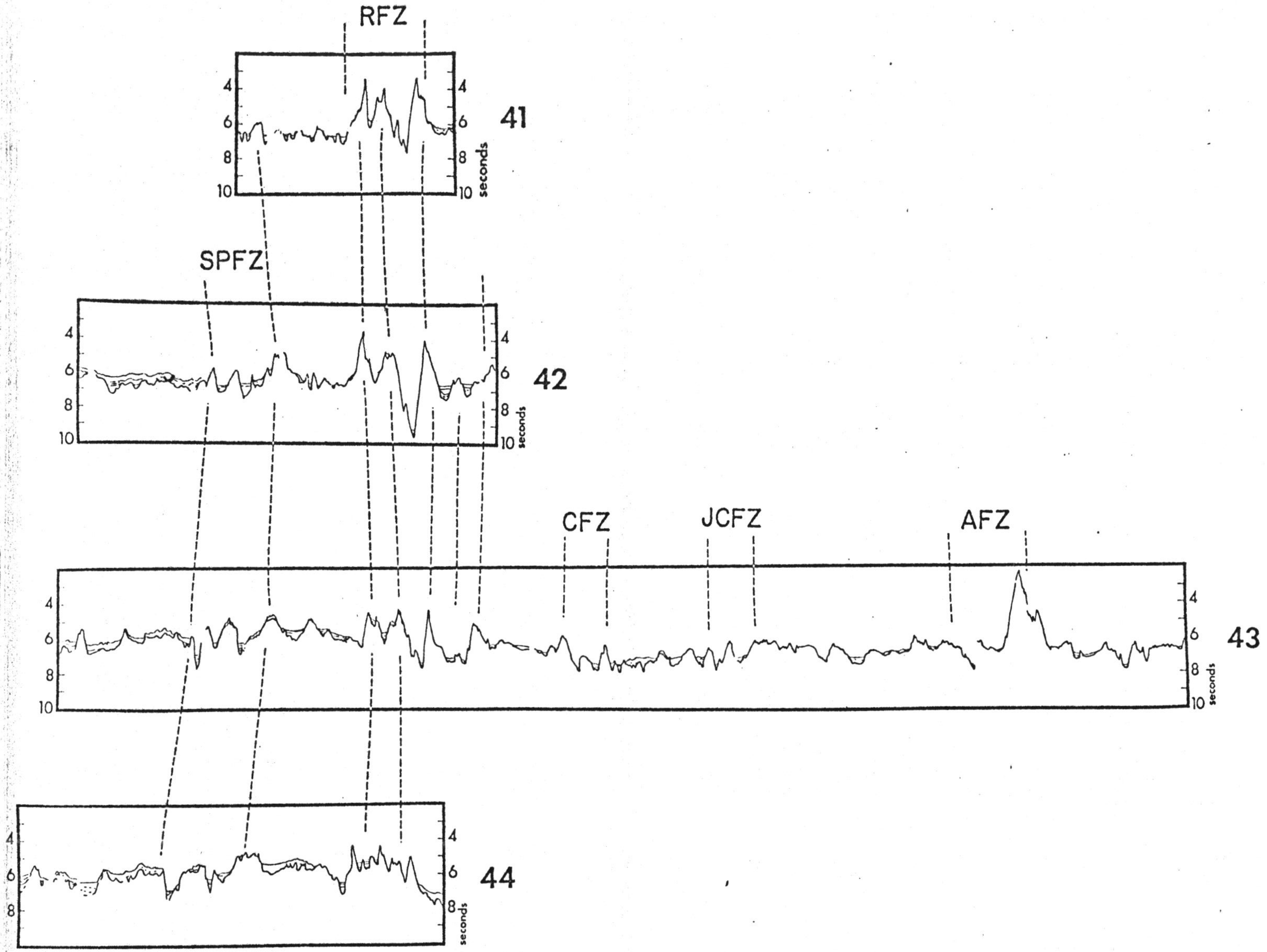

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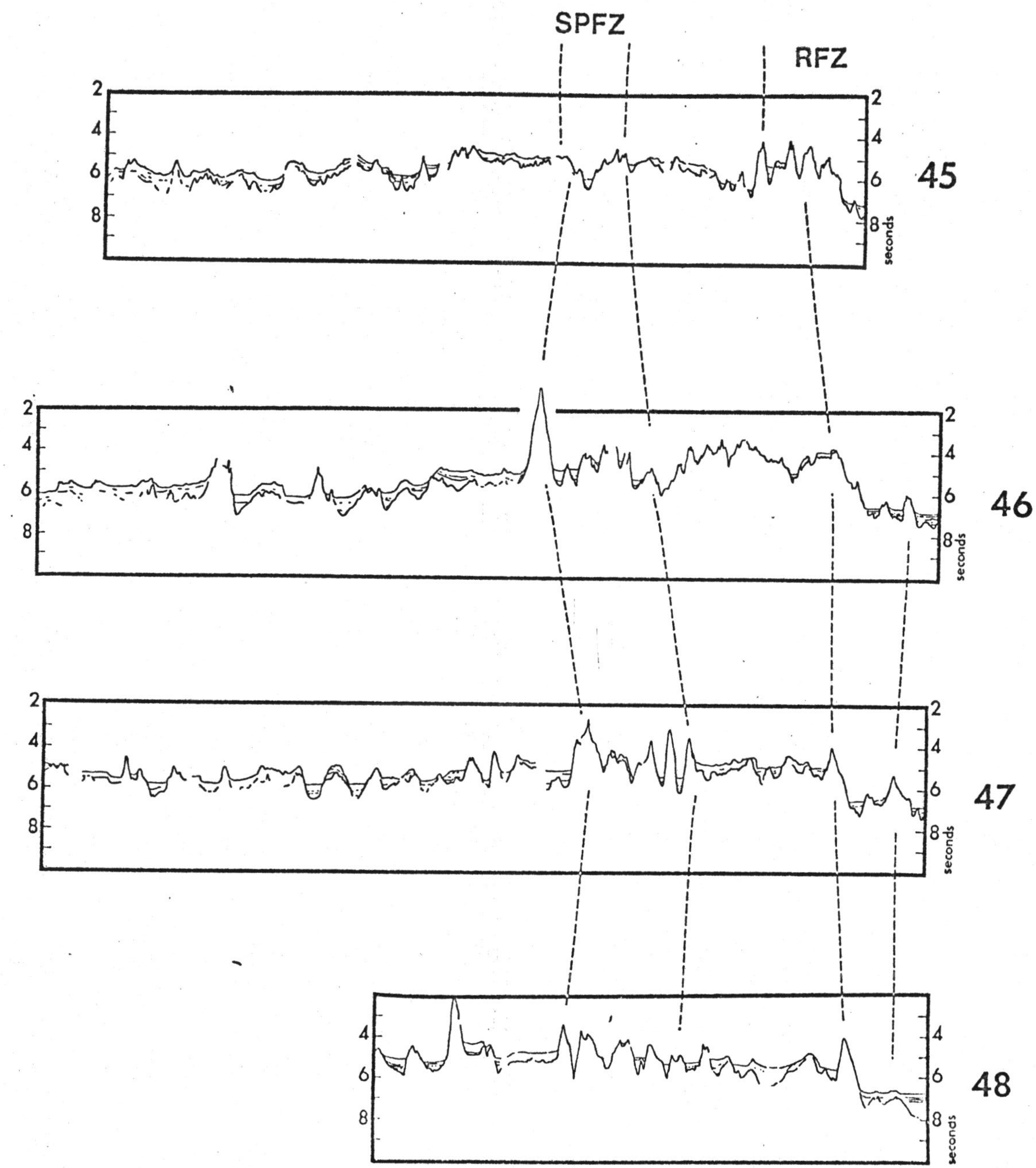
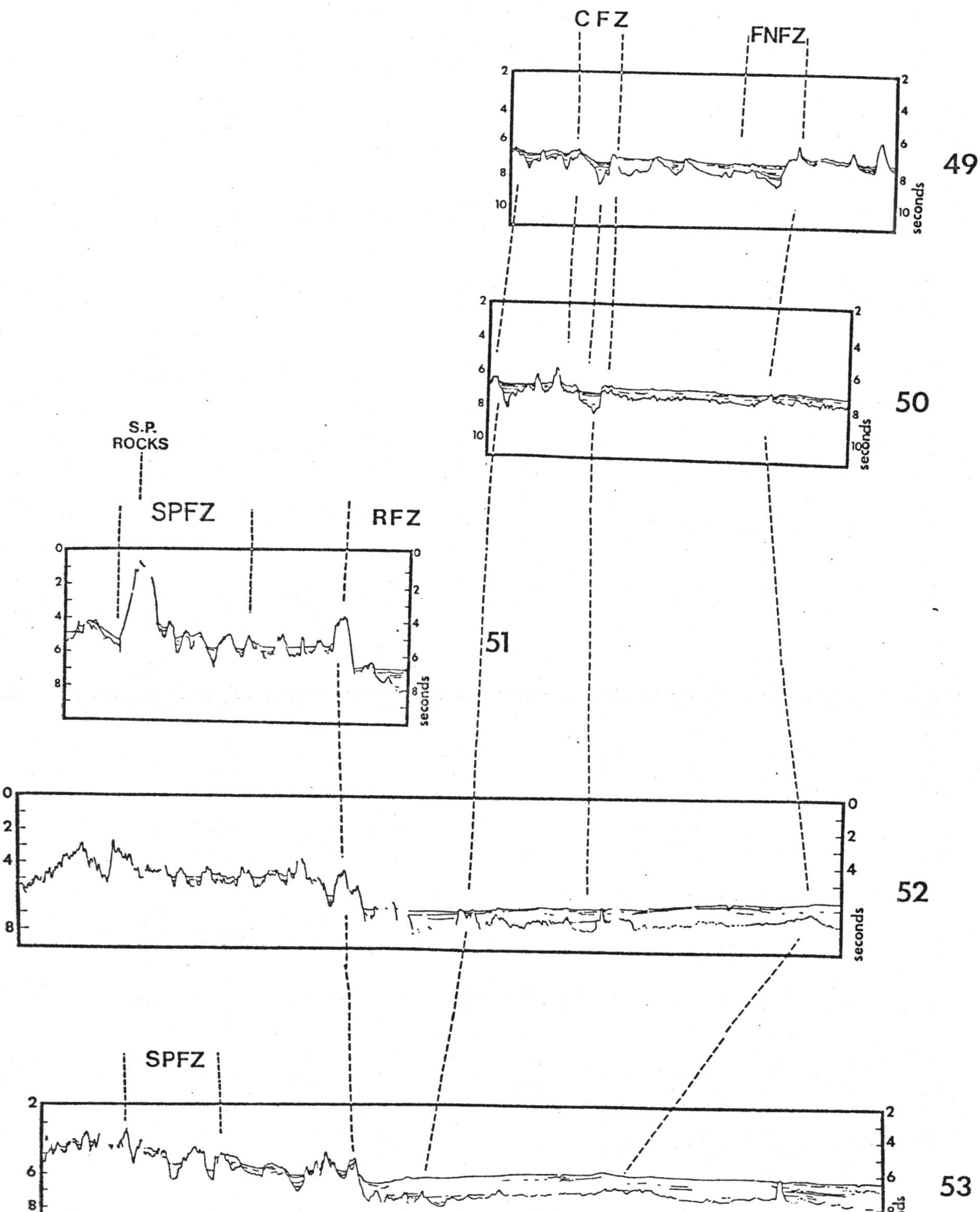
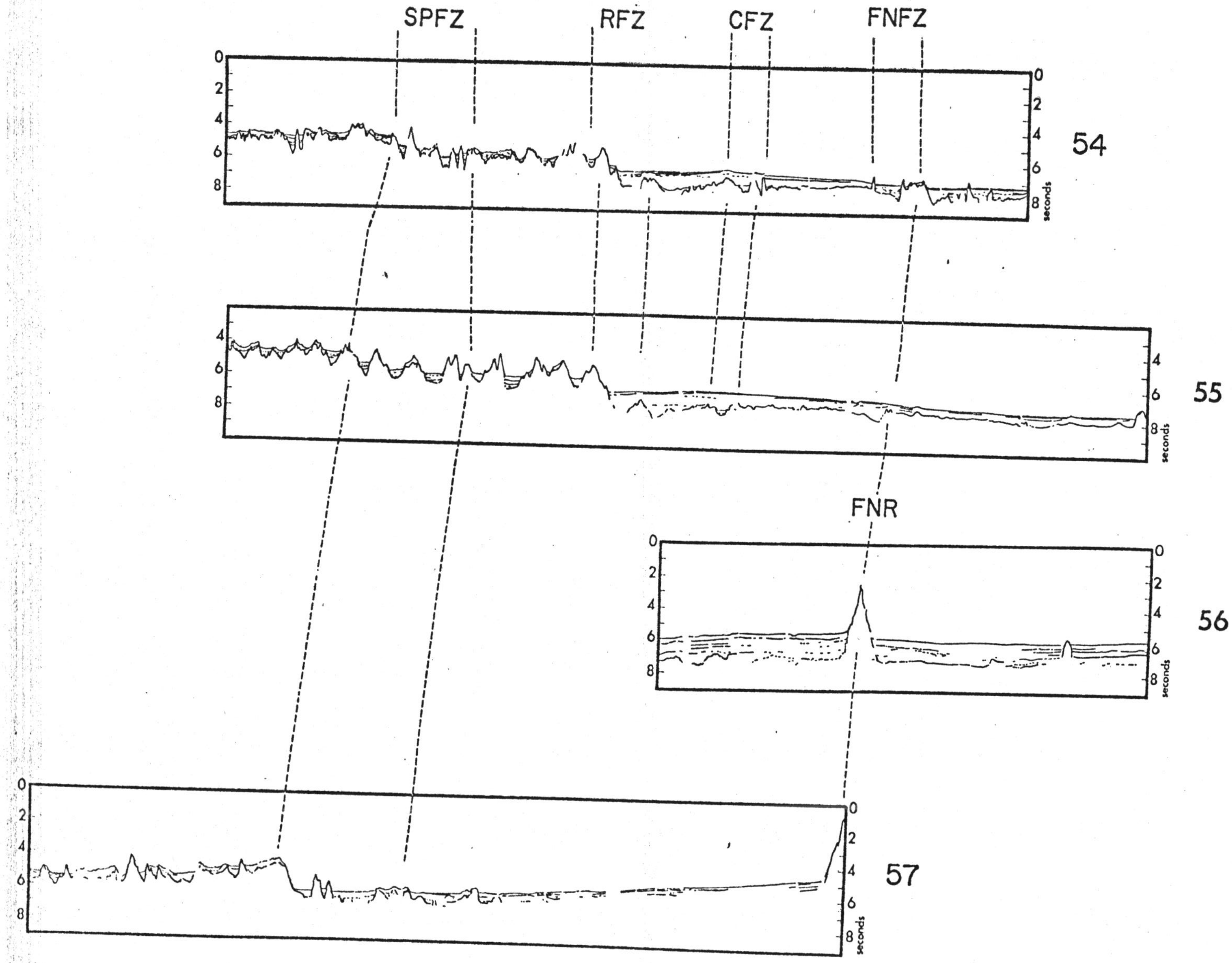
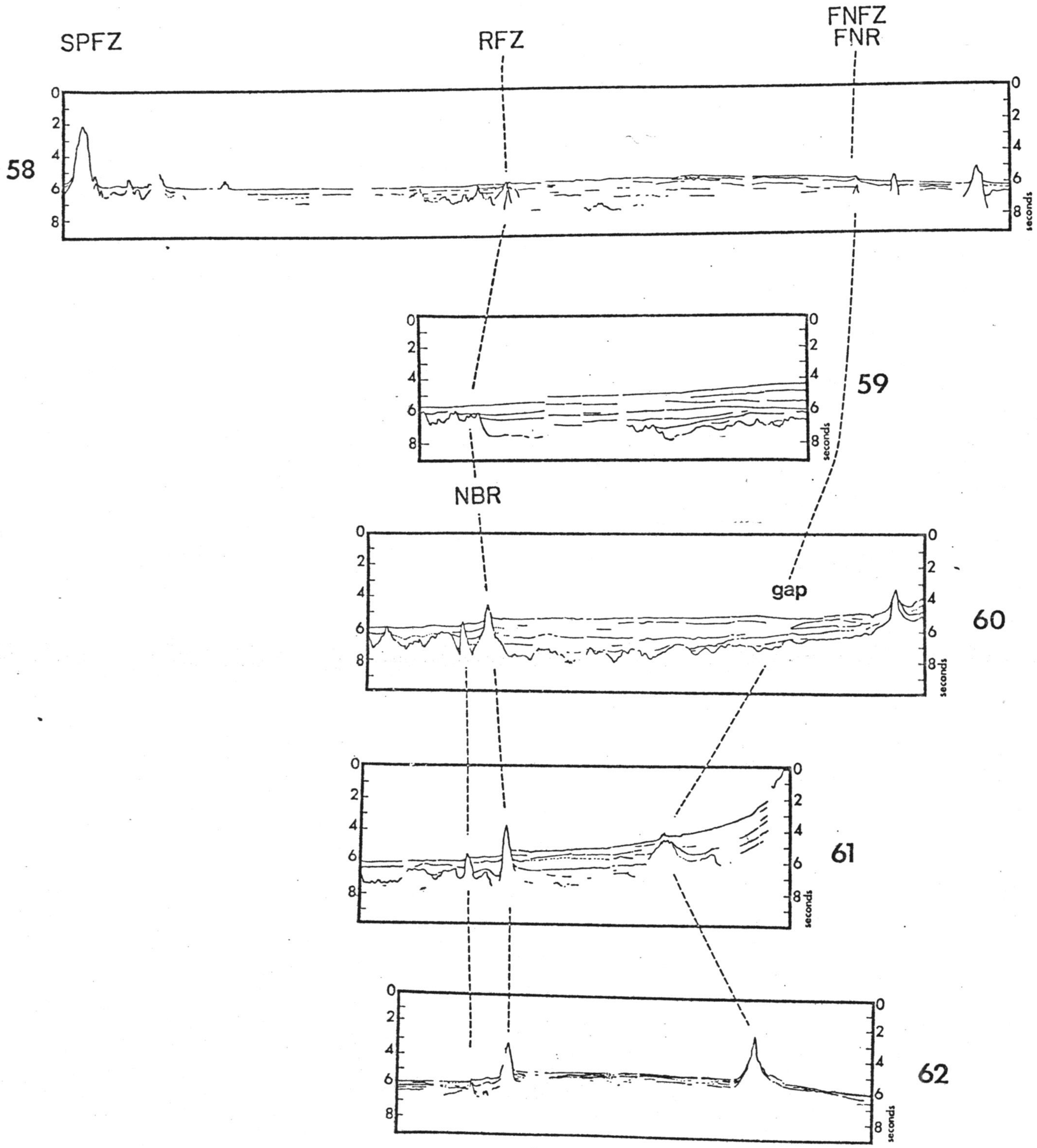
JCFZ
AFZ
43
44
seconds

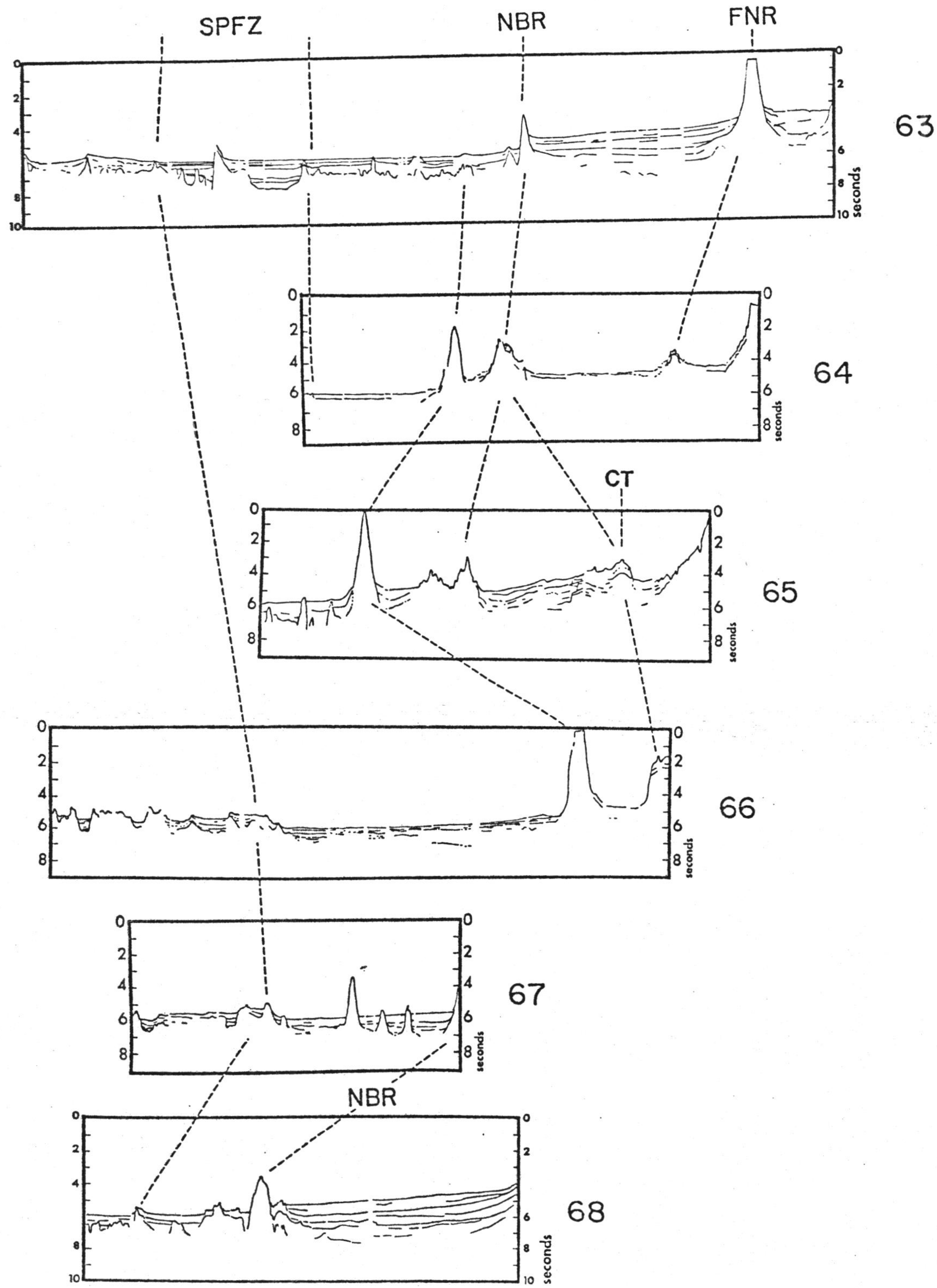

SPFZ
RFZ
45
46
47
48
seconds

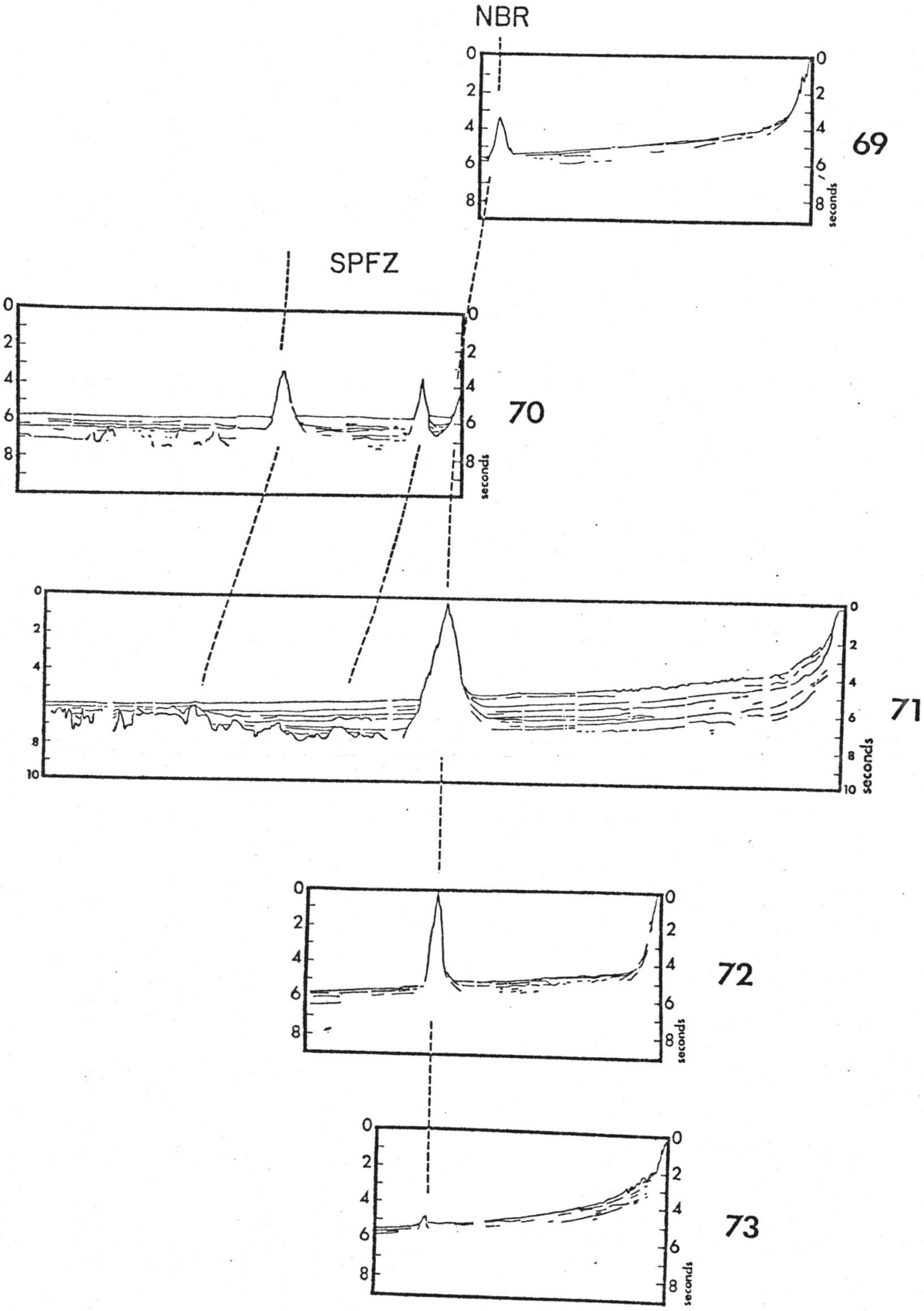


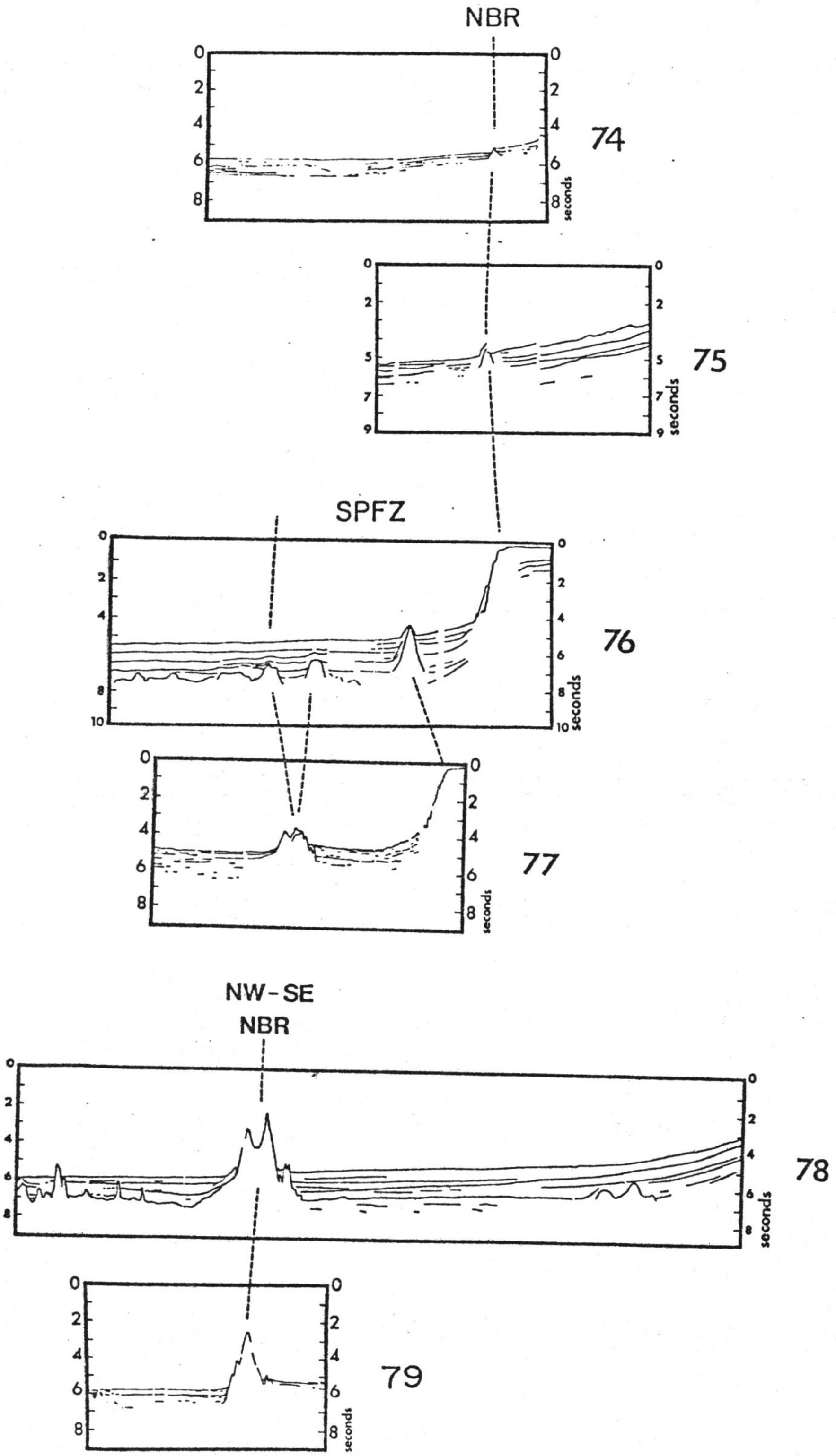
C F Z
FNFZ
49
50
S.P. ROCKS
SPFZ
RFZ
51
52
SPFZ
53
seconds


SPFZ
RFZ
CFZ
FNFZ
FNR
54
55
56
57
seconds


SPFZ
RFZ
FNFZ
FNR
58
59
NBR
gap
60
61
62
seconds
0
2
4
6
8


SPFZ
NBR
FNR
63
64
CT
65
66
67
NBR
68
seconds


NBR
SPFZ
69
70
71
72
73
seconds


NBR
74
75
SPFZ
76
77
NW-SE
NBR
78
79
seconds

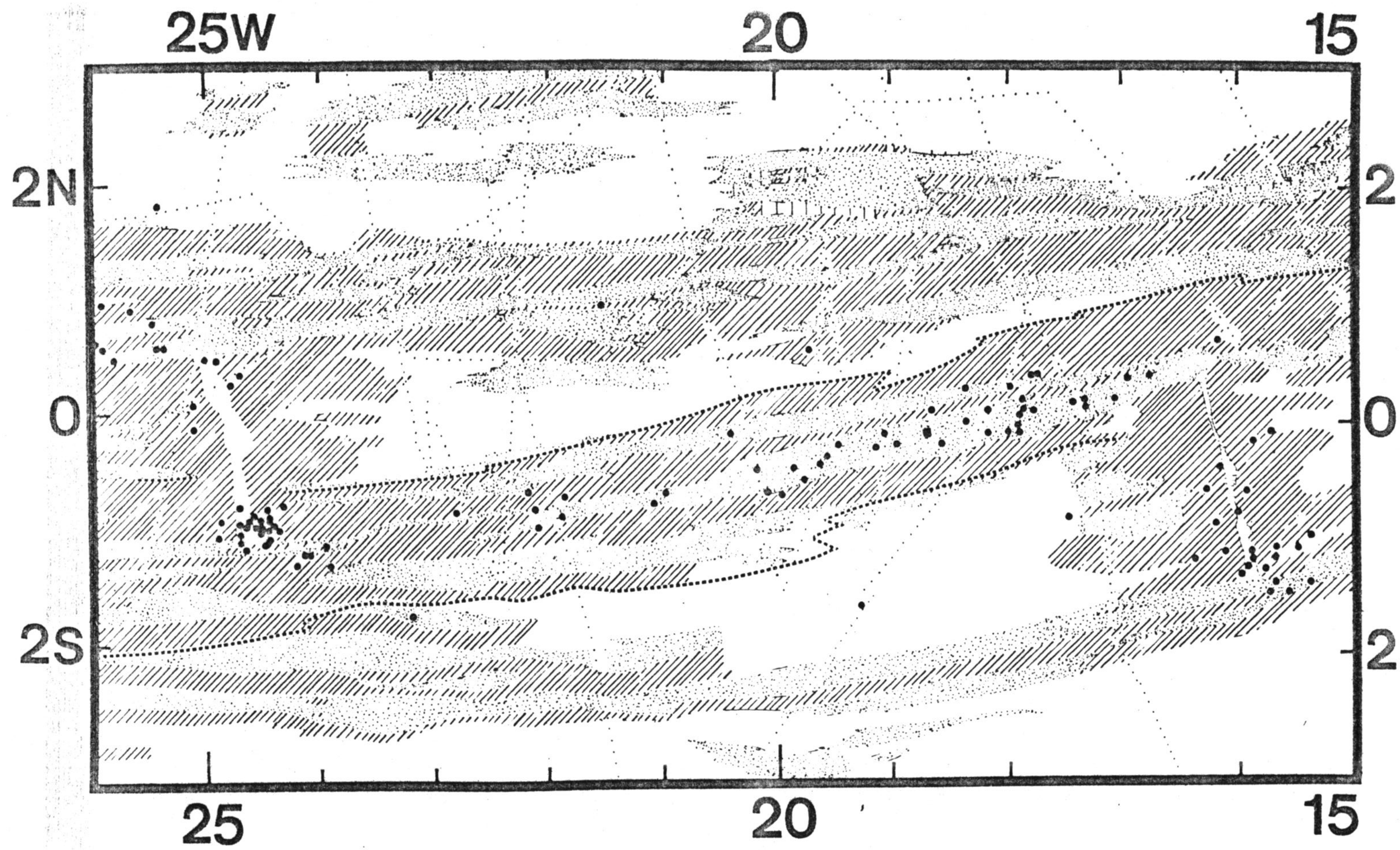


Figure 5

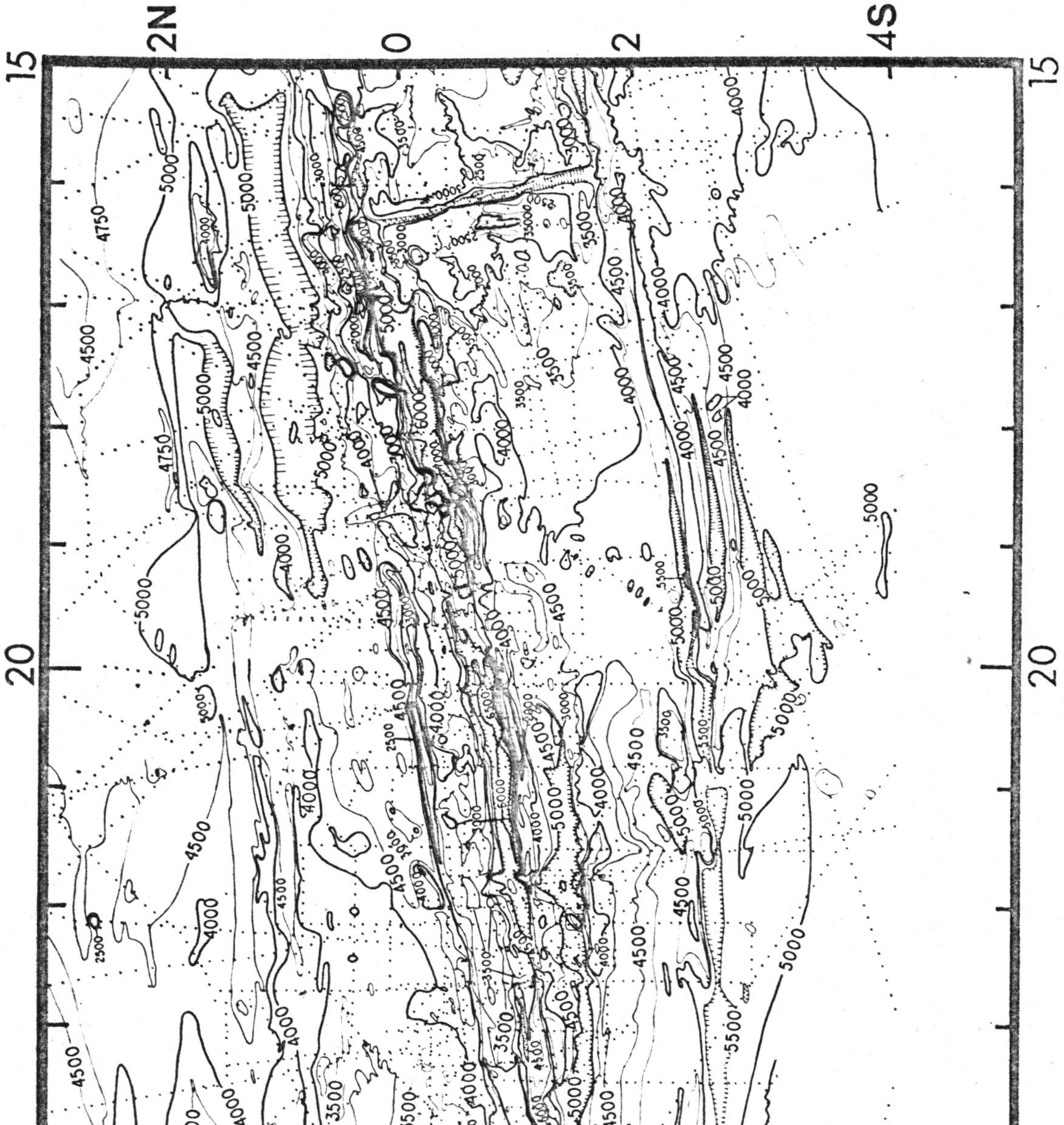


Figure 6

A F R I C A
B R A S I L
LAGOS
ACCRA
ABIDJAN
Cape Palmas
Cape Three Points
Cape Formoso
DOUALA
I. F. POO
I. DO PRINCIPE
I. DE SÃO TOMÉ
I. ANNOBÓN
ST. PAUL'S ROCKS
BELÉM
SÃO LUIS
FORTALEZA
I. FERNANDO DE NORONHA
ATOL DAS ROCAS
NATAL
50W 45 40 35 30 25 20 15 10 5 0 5 10E
10N 8 6 4 2 0 2 4 6S

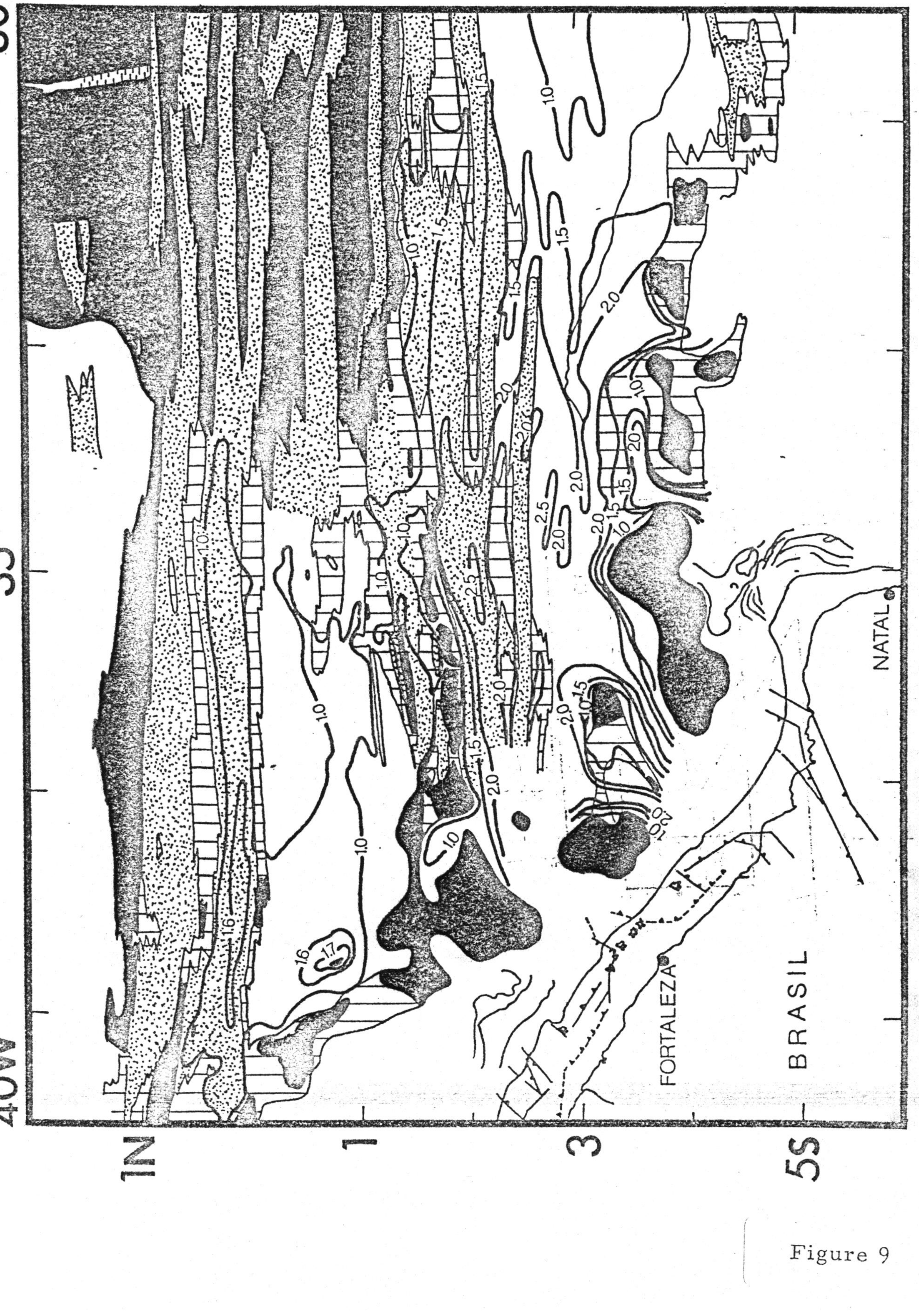


Figure 9

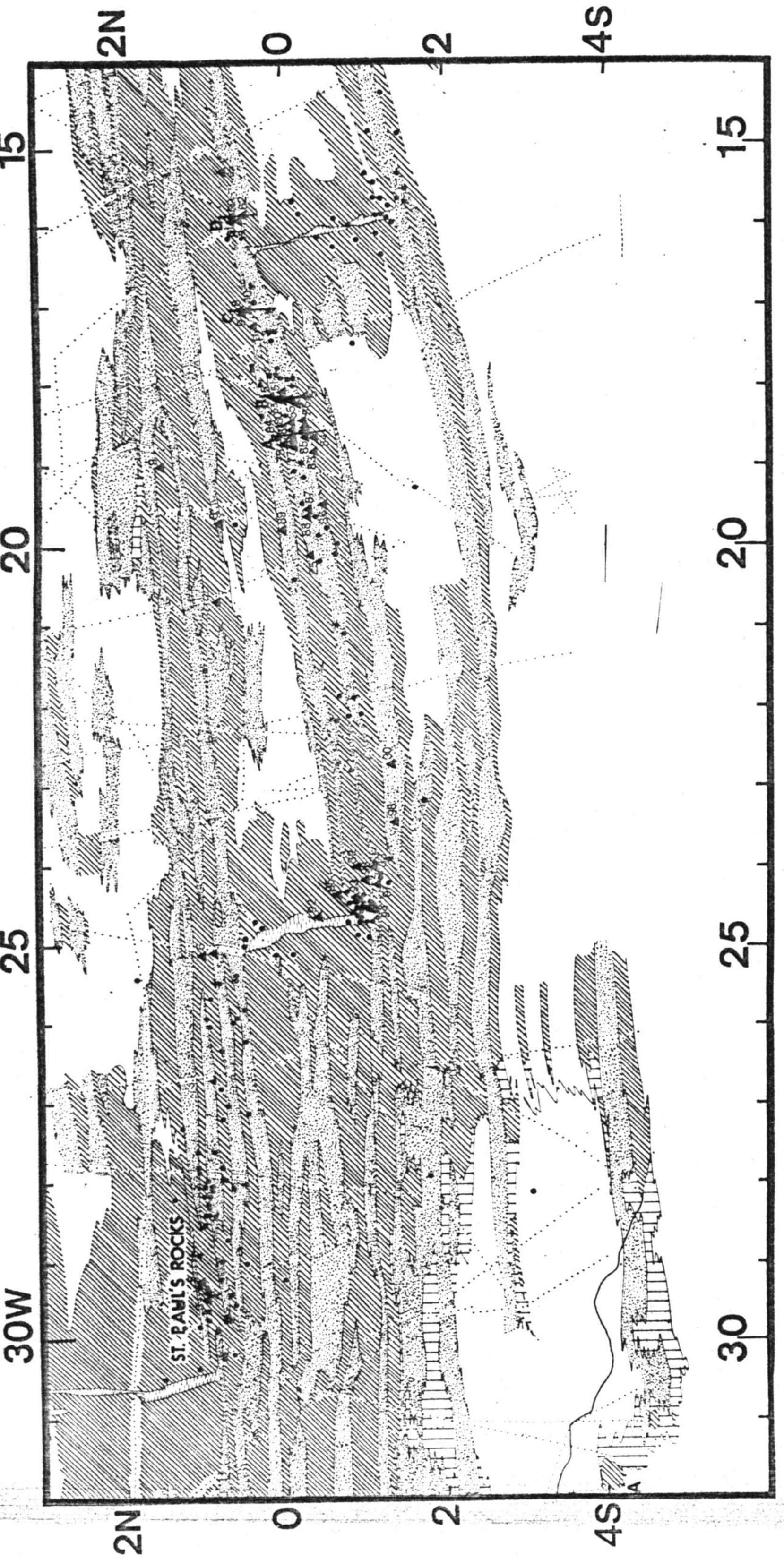


Figure 10

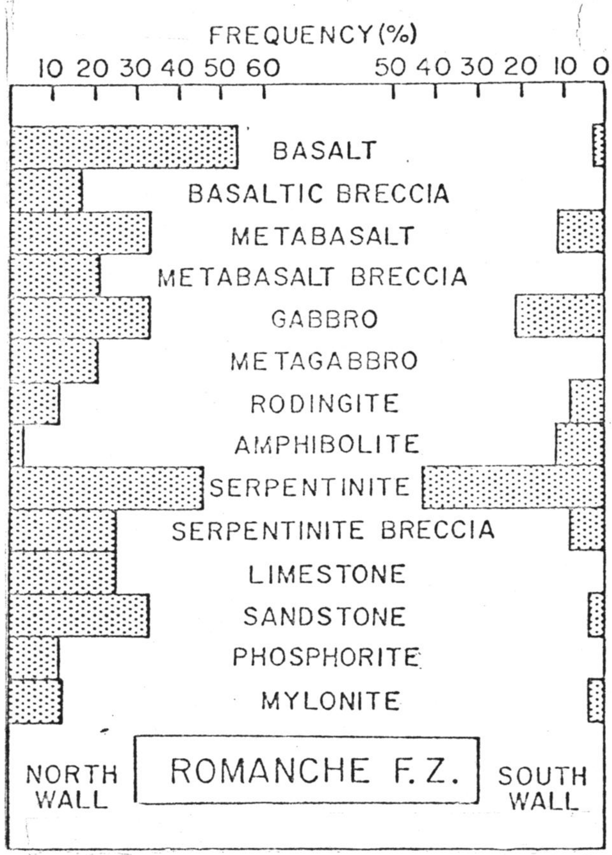
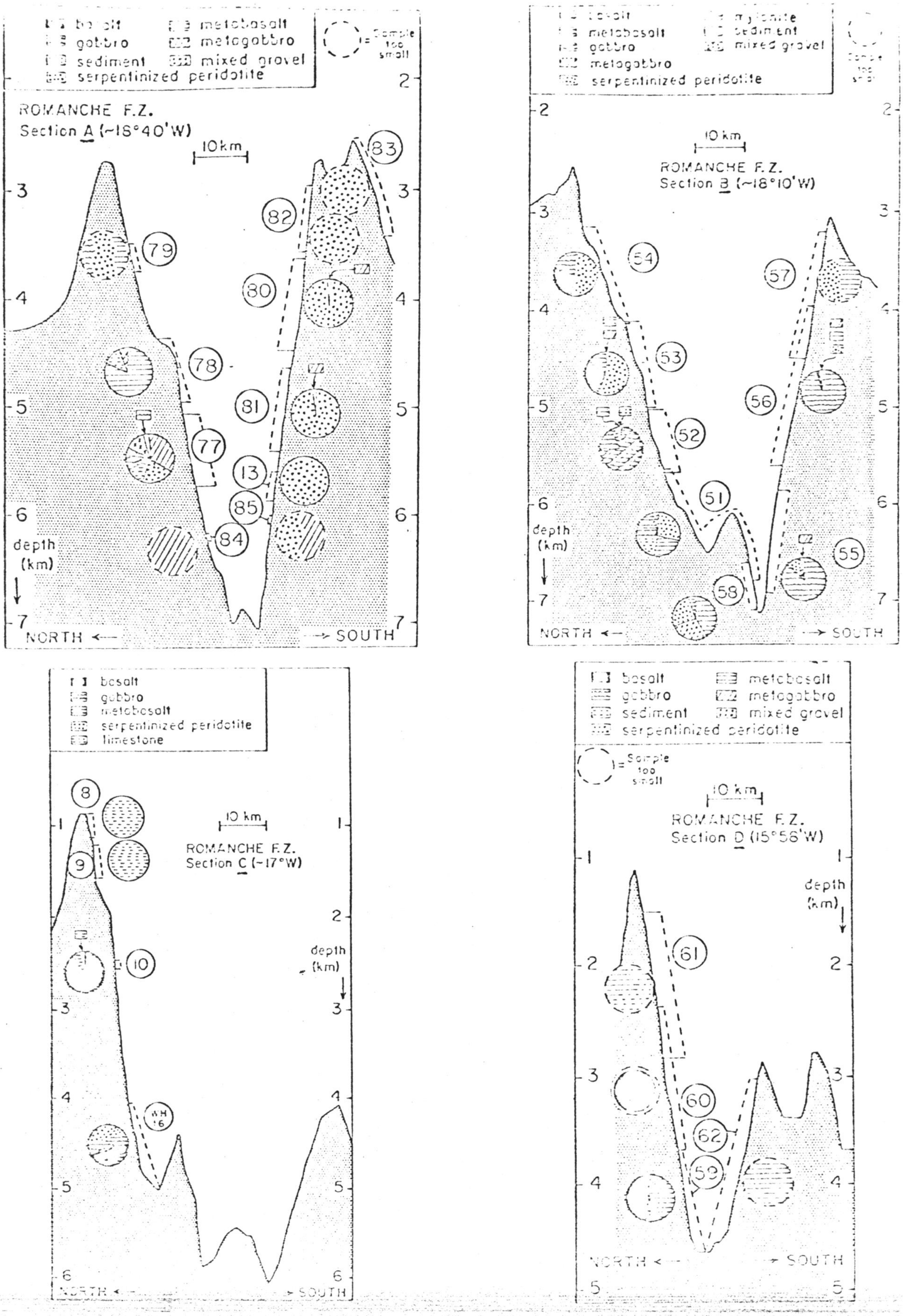

FREQUENCY (%)
10 20 30 40 50 60
50 40 30 20 10 0
BASALT
BASALTIC BRECCIA
METABASALT
METABASALT BRECCIA
GABBRO
METAGABBRO
RODINGITE
AMPHIBOLITE
SERPENTINITE
SERPENTINITE BRECCIA
LIMESTONE
SANDSTONE
PHOSPHORITE
MYLONITE
NORTH WALL
ROMANCHE F. Z.
SOUTH WALL

Figure 11



basalt
gabbro
sediment
serpentinized peridotite
metabasalt
metagabbro
mixed gravel
= Sample too small
ROMANCHE F.Z.
Section A (~18°40'W)
10 km
depth (km)
NORTH
SOUTH
metabasalt
mylonite
ROMANCHE F.Z.
Section B (~18°10'W)
limestone
ROMANCHE F.Z.
Section C (~17°W)
ROMANCHE F.Z.
Section D (15°56'W)

Figure 12

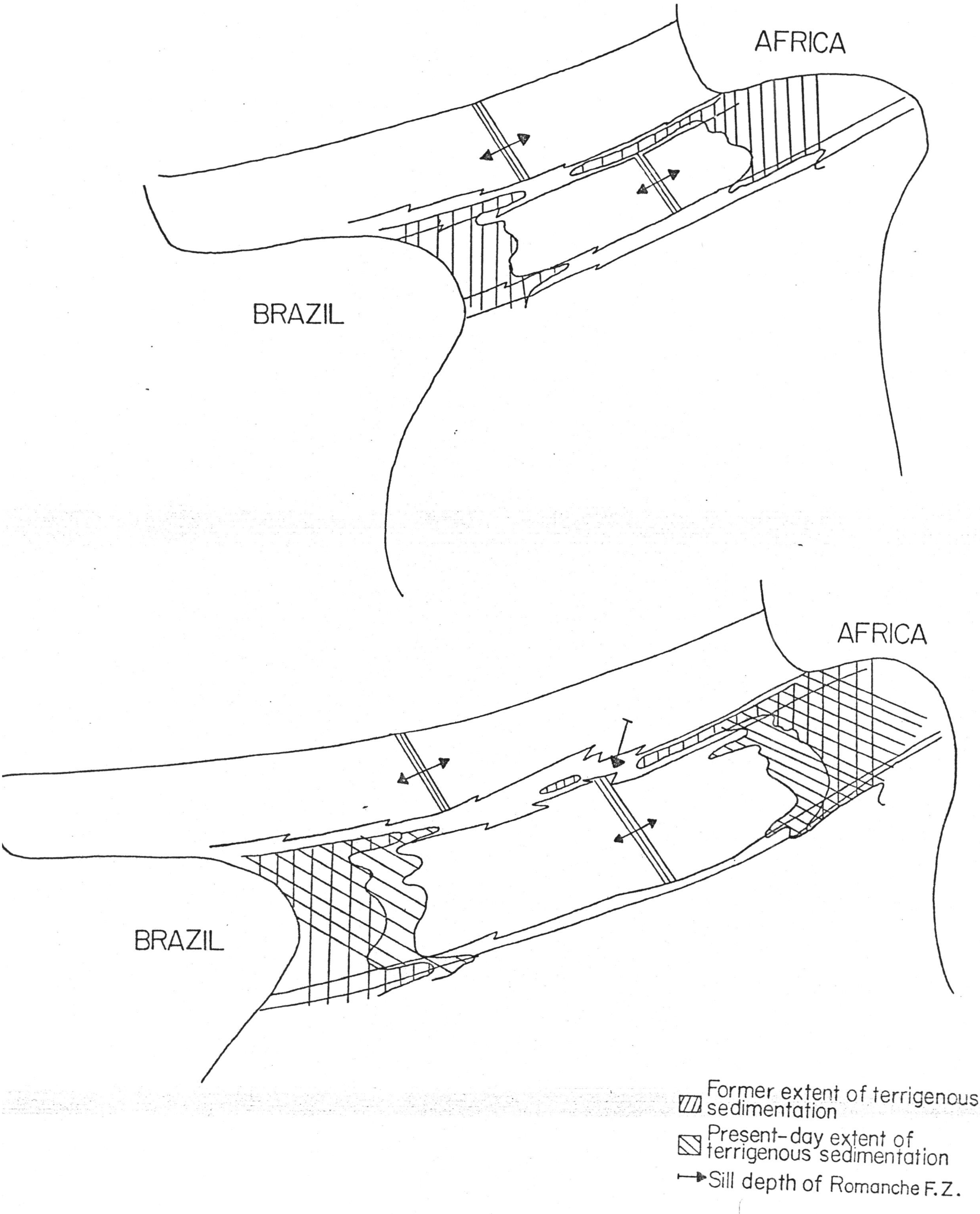

AFRICA
BRAZIL
AFRICA
BRAZIL
Former extent of terrigenous sedimentation
Present-day extent of terrigenous sedimentation
Sill depth of Romanche F.Z.

Figure 13

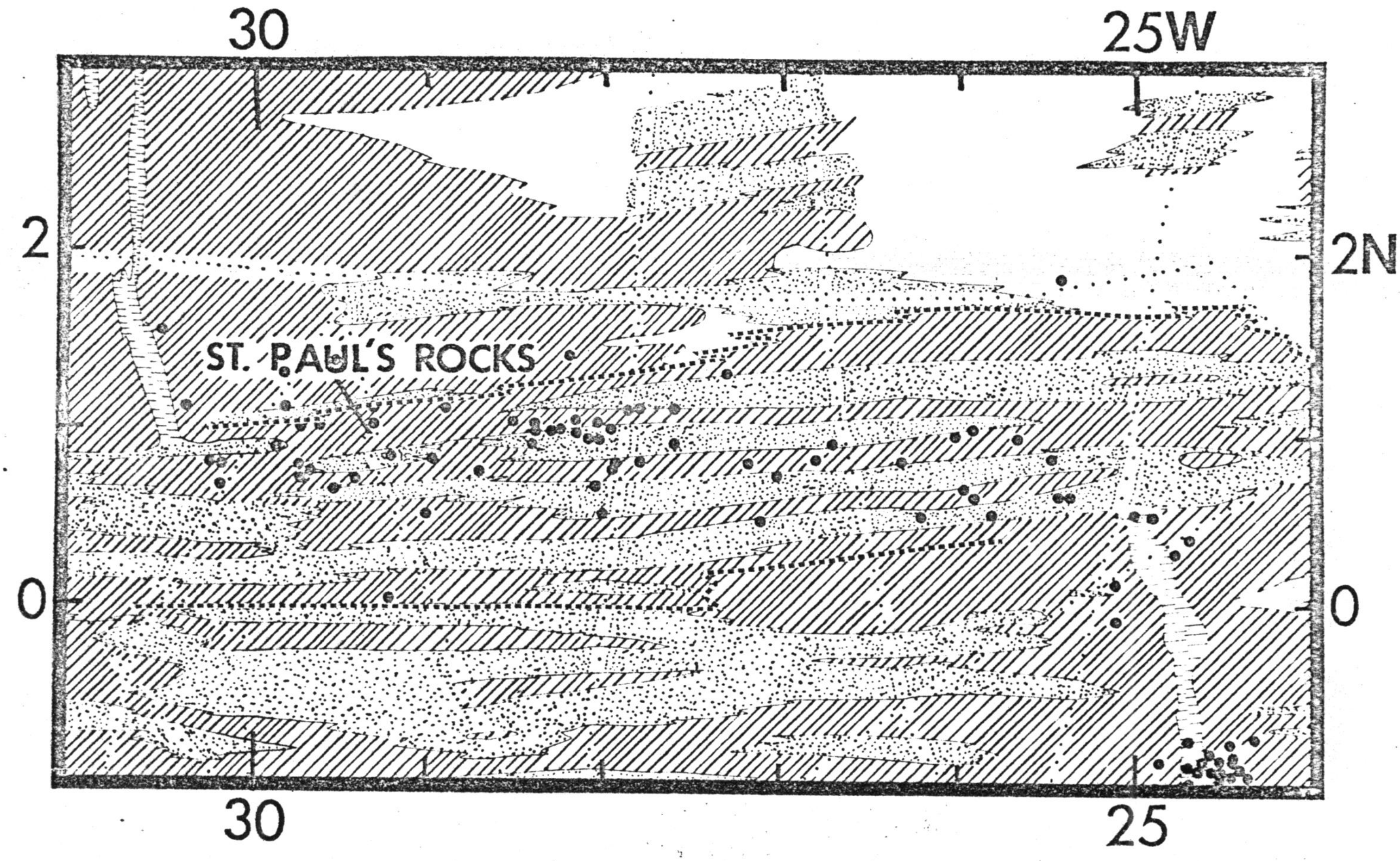


Figure 14

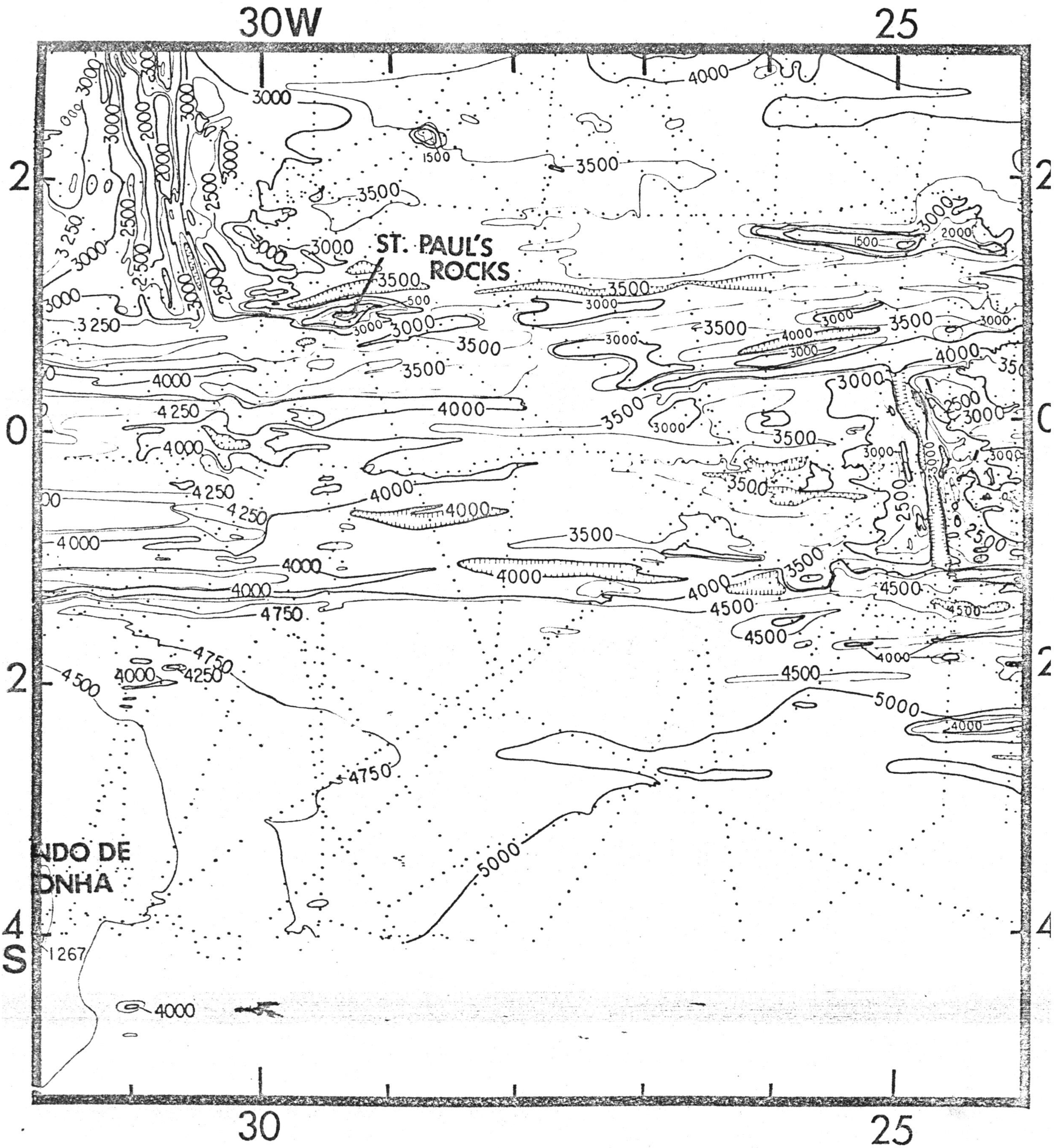

30W
25
ST. PAUL'S
ROCKS
NDO DE
ONHA
1267
30
25
2
0
2
4
S
3000
3500
4000
4250
4500
4750
5000
1500
2000
2500

Figure 15

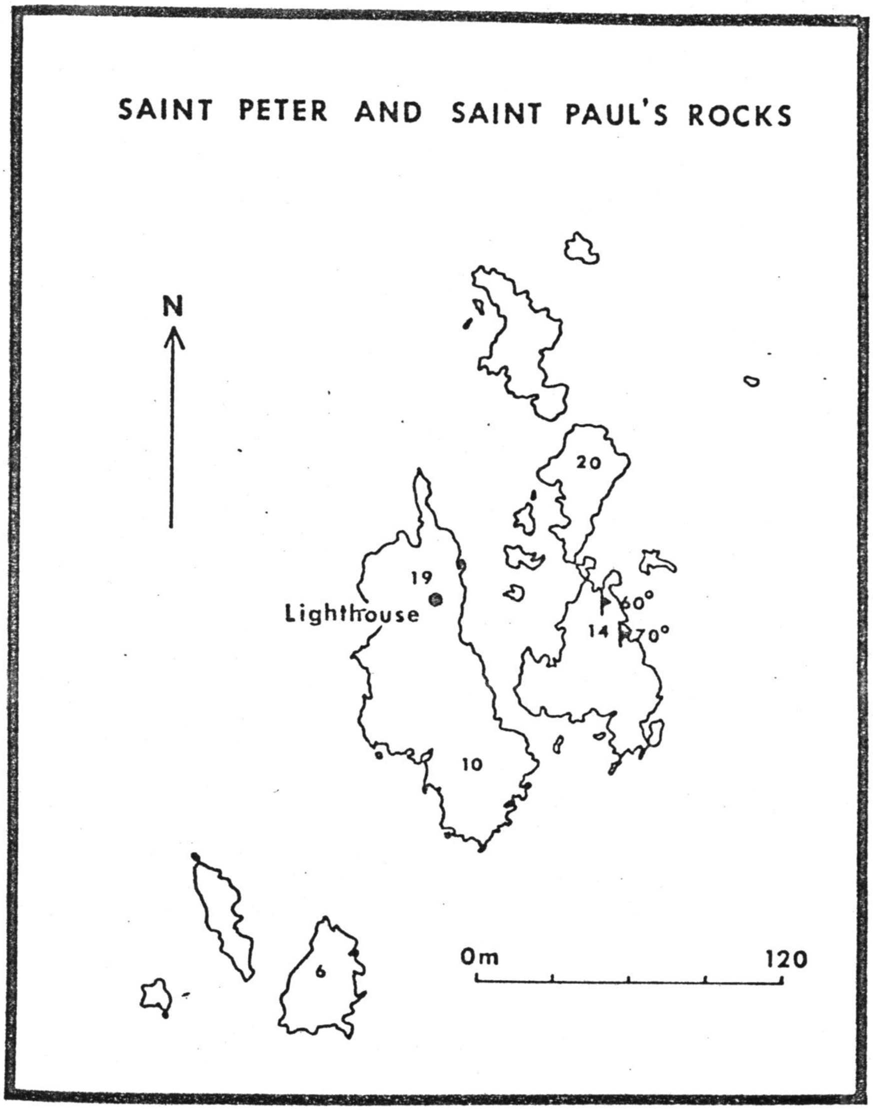

SAINT PETER AND SAINT PAUL'S ROCKS
N
20
19
Lighthouse
60°
14
70°
10
6
0m
120

Figure 16

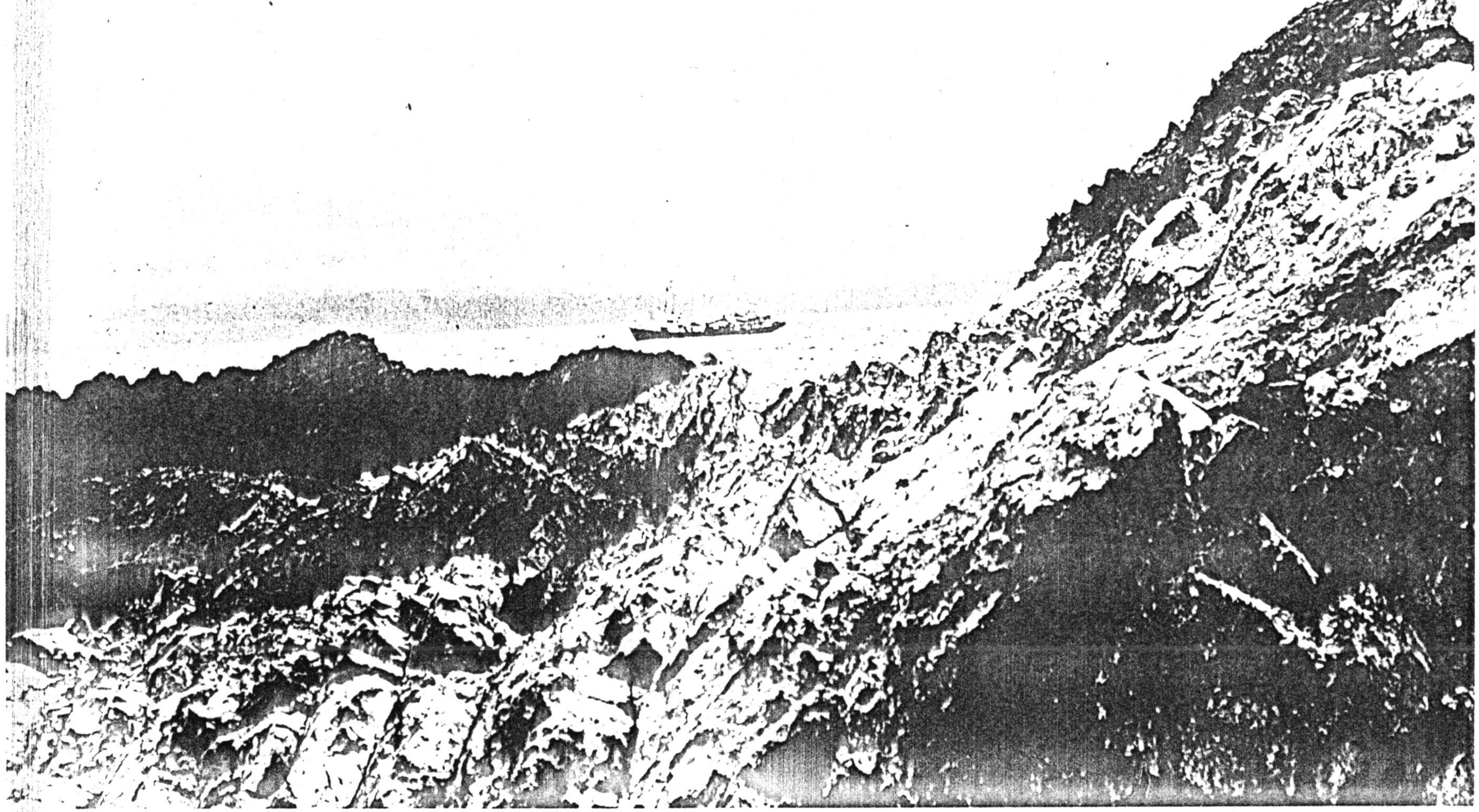
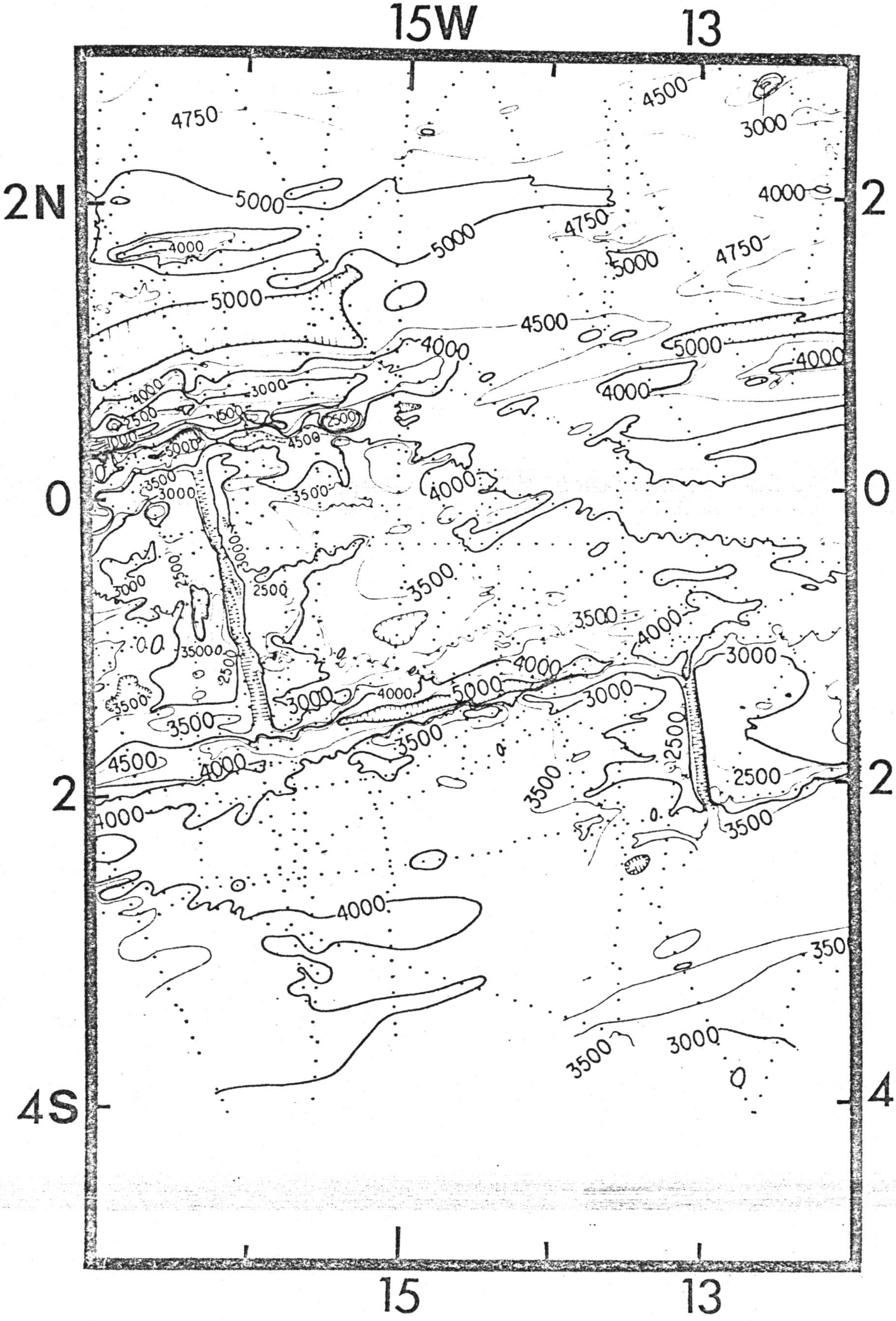


Figure 17



15W
13
2N
0
2
4S
15
4750
4500
3000
5000
4000
4750
5000
4000
4750
5000
4500
4000
5000
4000
4000
3000
2500
1500
2500
4500
3500
3000
4000
3500
3000
2500
3000
2500
3500
3500
4000
3000
4000
4000
3000
5000
3500
3500
3500
2500
2500
3500
3500
4500
4000
2
4000
4000
350
3500
3000
4

Figure 18

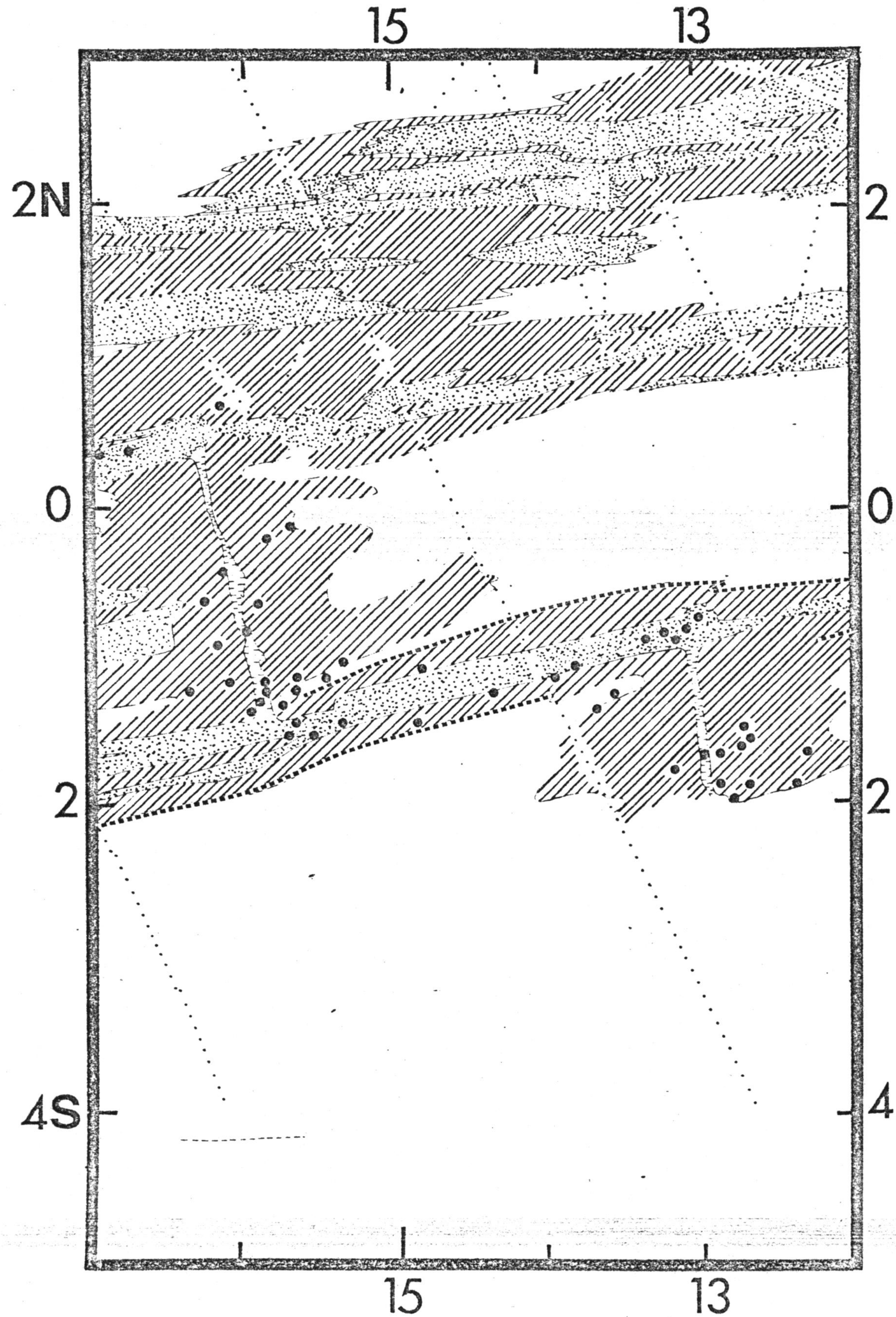


Figure 19

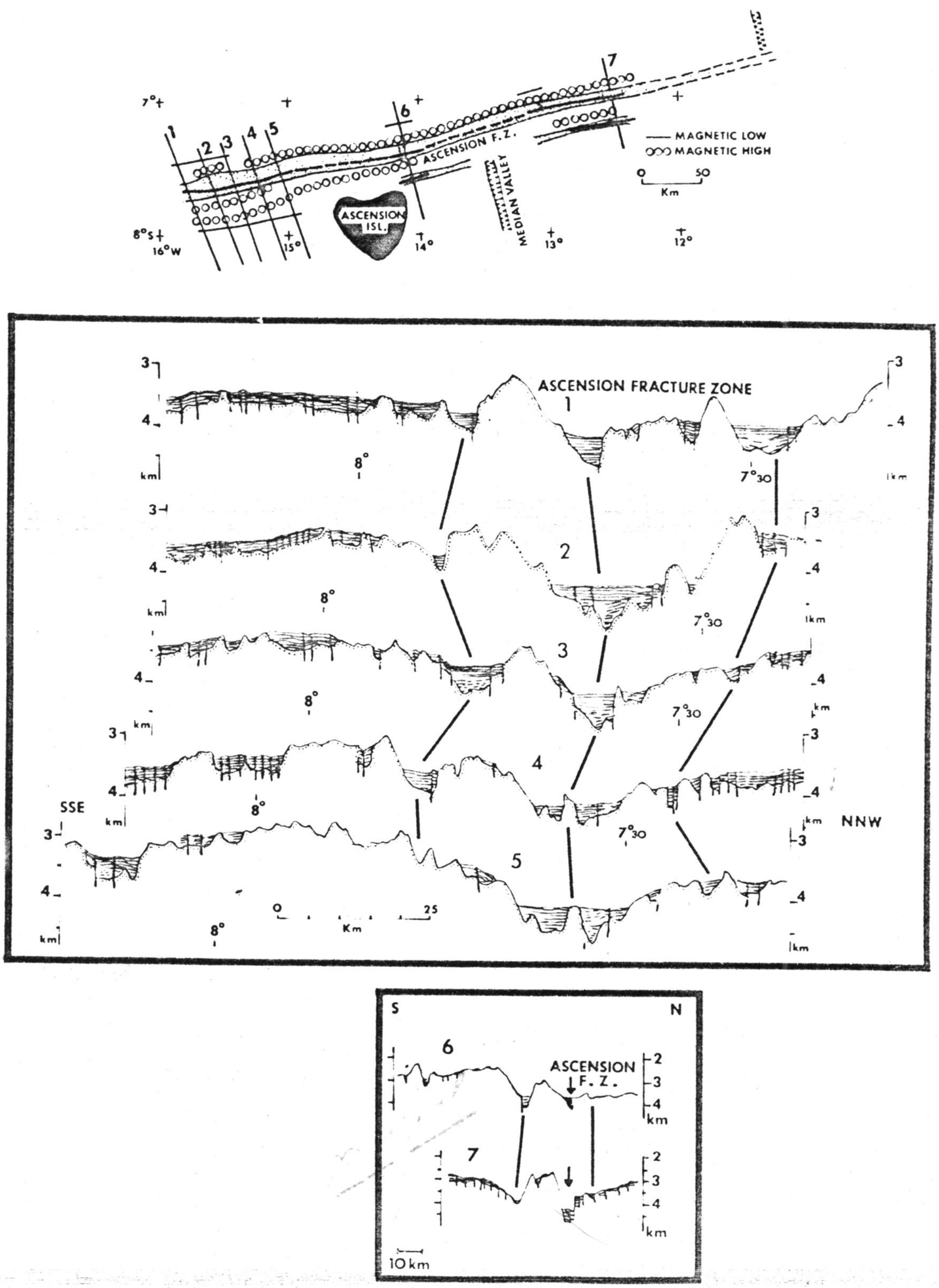

ASCENSION F.Z.
ASCENSION ISL.
MEDIAN VALLEY
MAGNETIC LOW
MAGNETIC HIGH
Km
ASCENSION FRACTURE ZONE
SSE
NNW
S
N
ASCENSION F.Z.
10 km

Figure 20

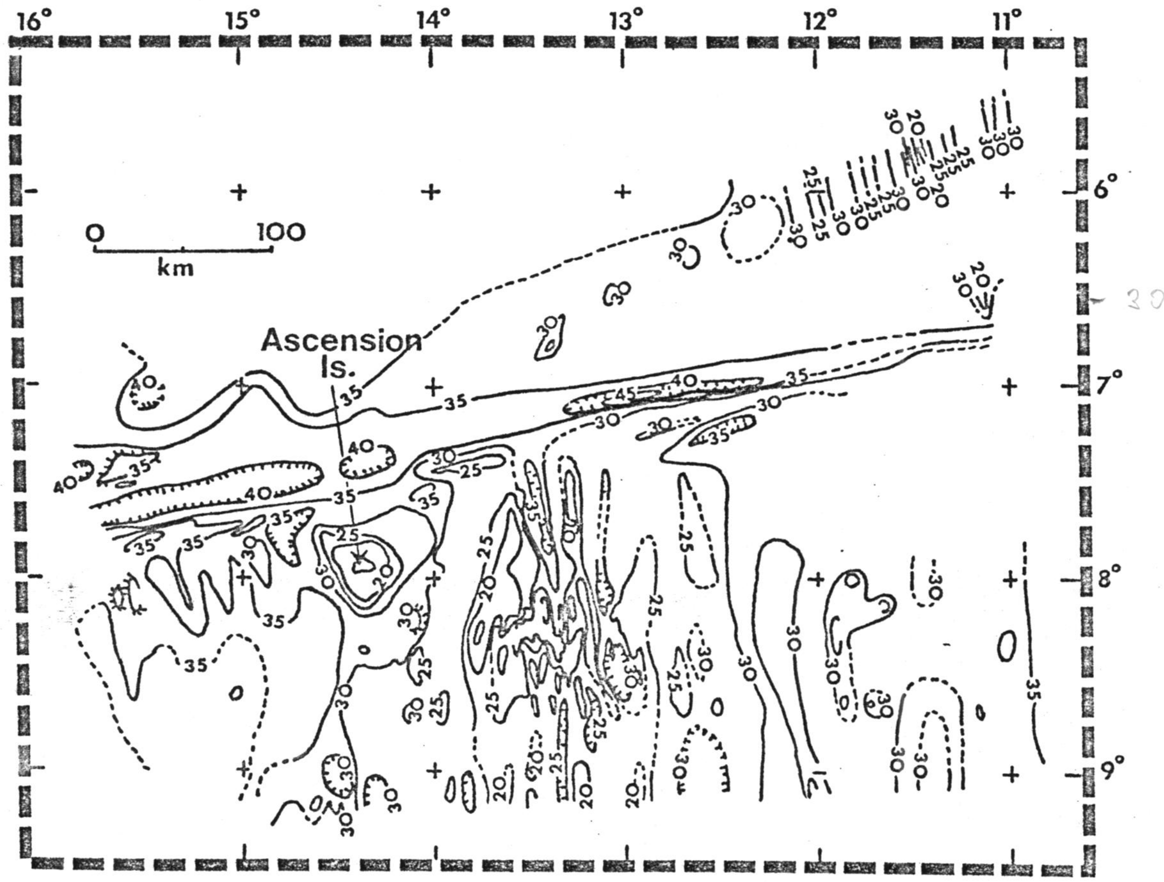


Figure 21

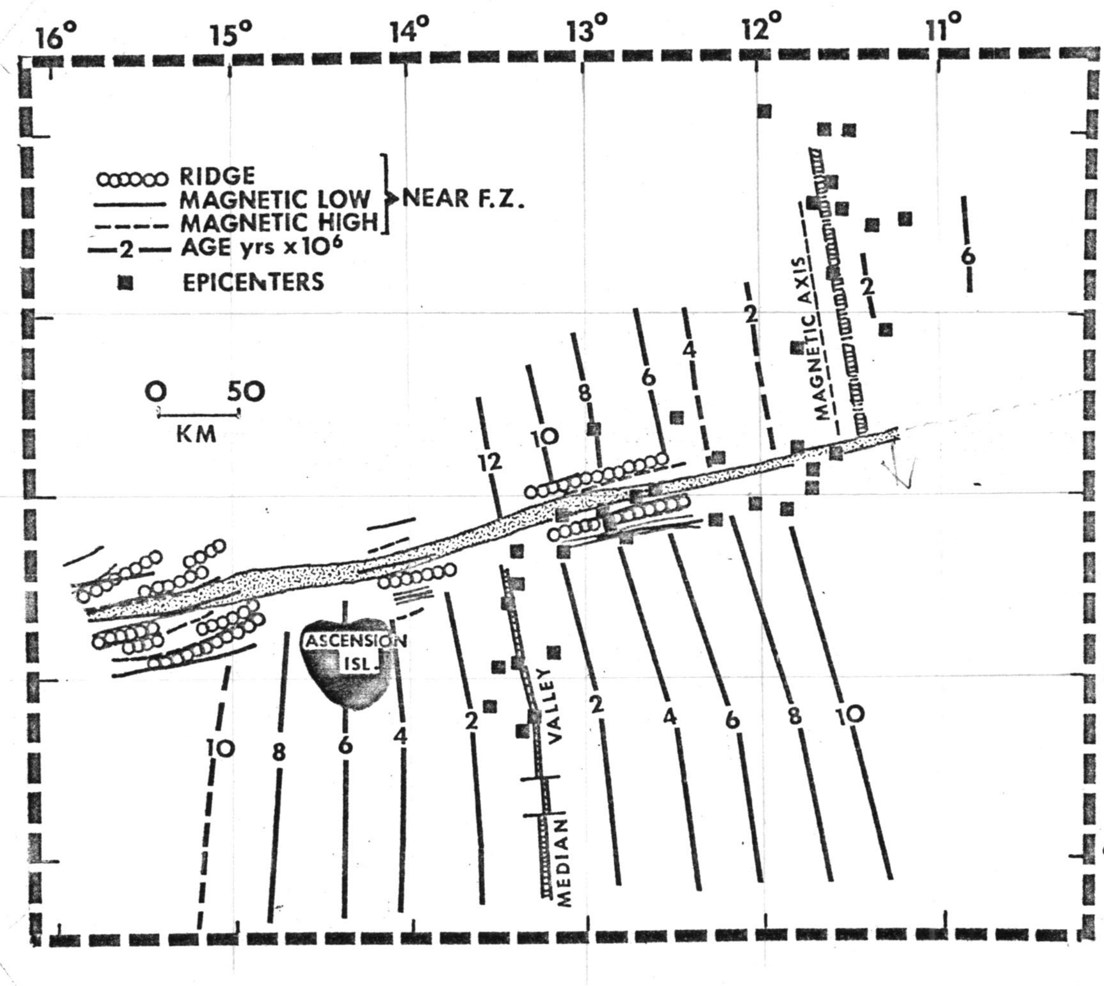

16°
15°
14°
13°
12°
11°
5°
6°
7°
8°
9°
RIDGE
MAGNETIC LOW
MAGNETIC HIGH
NEAR F.Z.
2 AGE yrs x10⁶
EPICENTERS
0 50
KM
MAGNETIC AXIS
ASCENSION ISL.
MEDIAN VALLEY
12
10
8
6
4
2

Figure 22

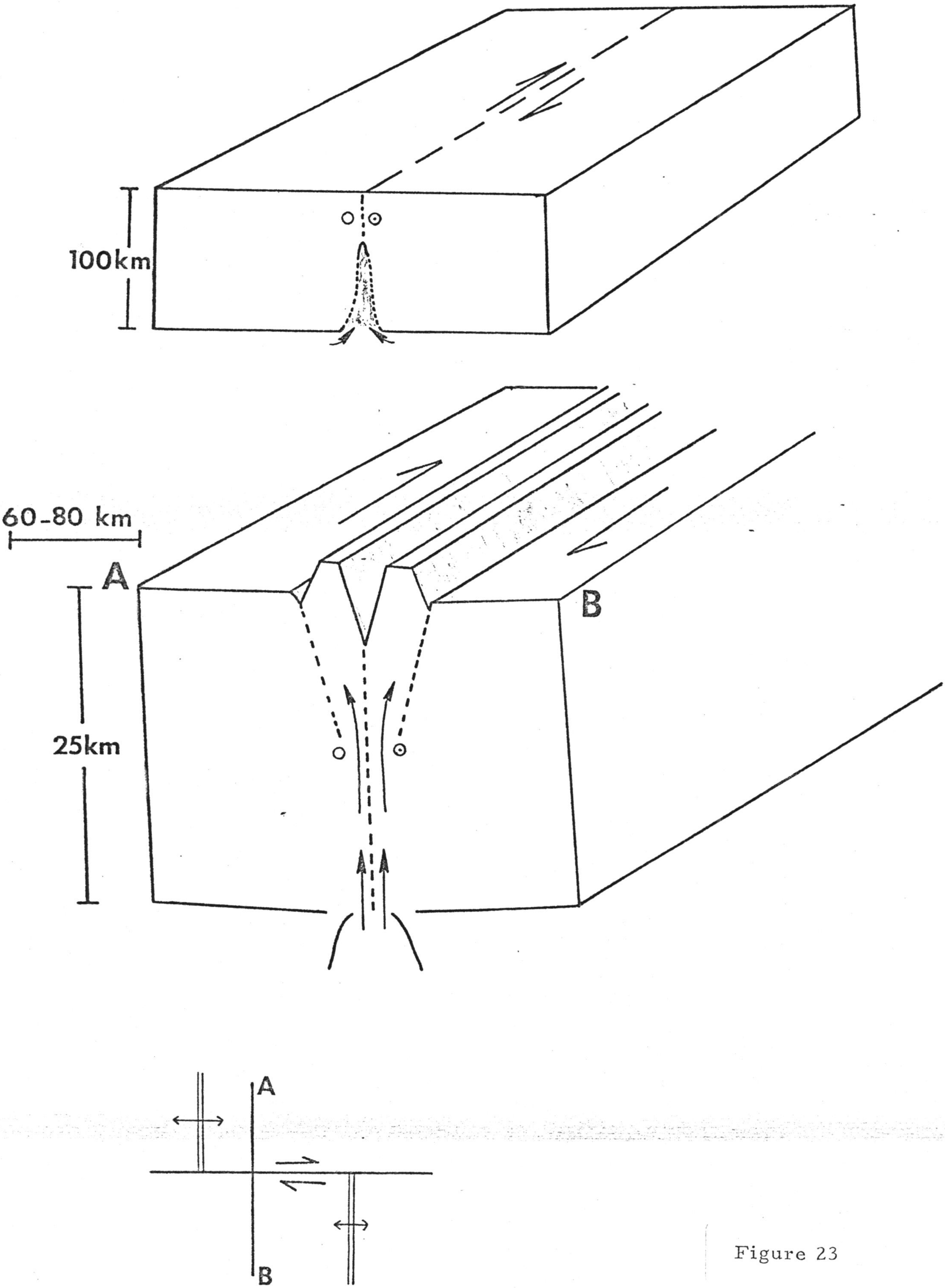


Figure 23

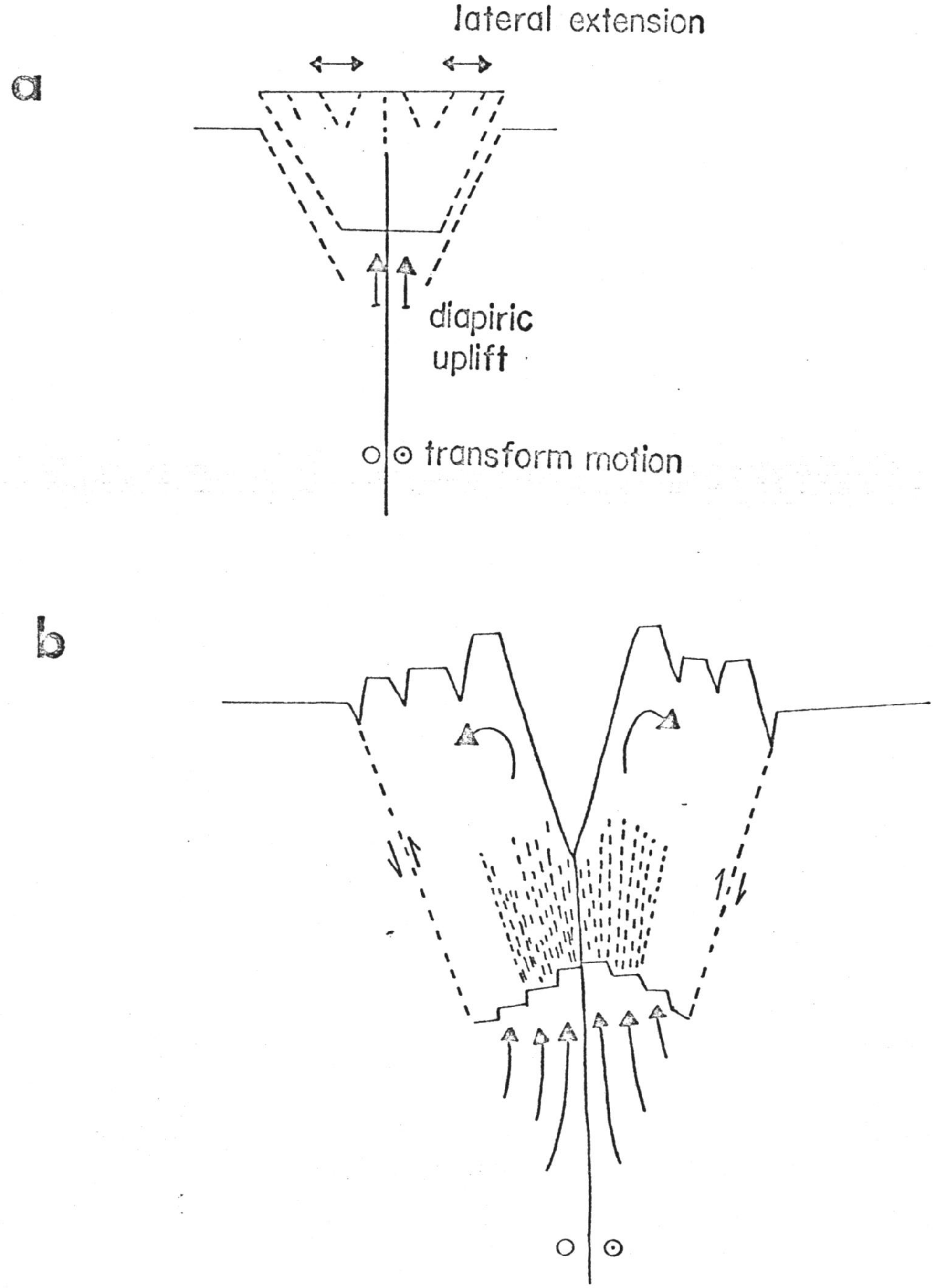


Figure 24

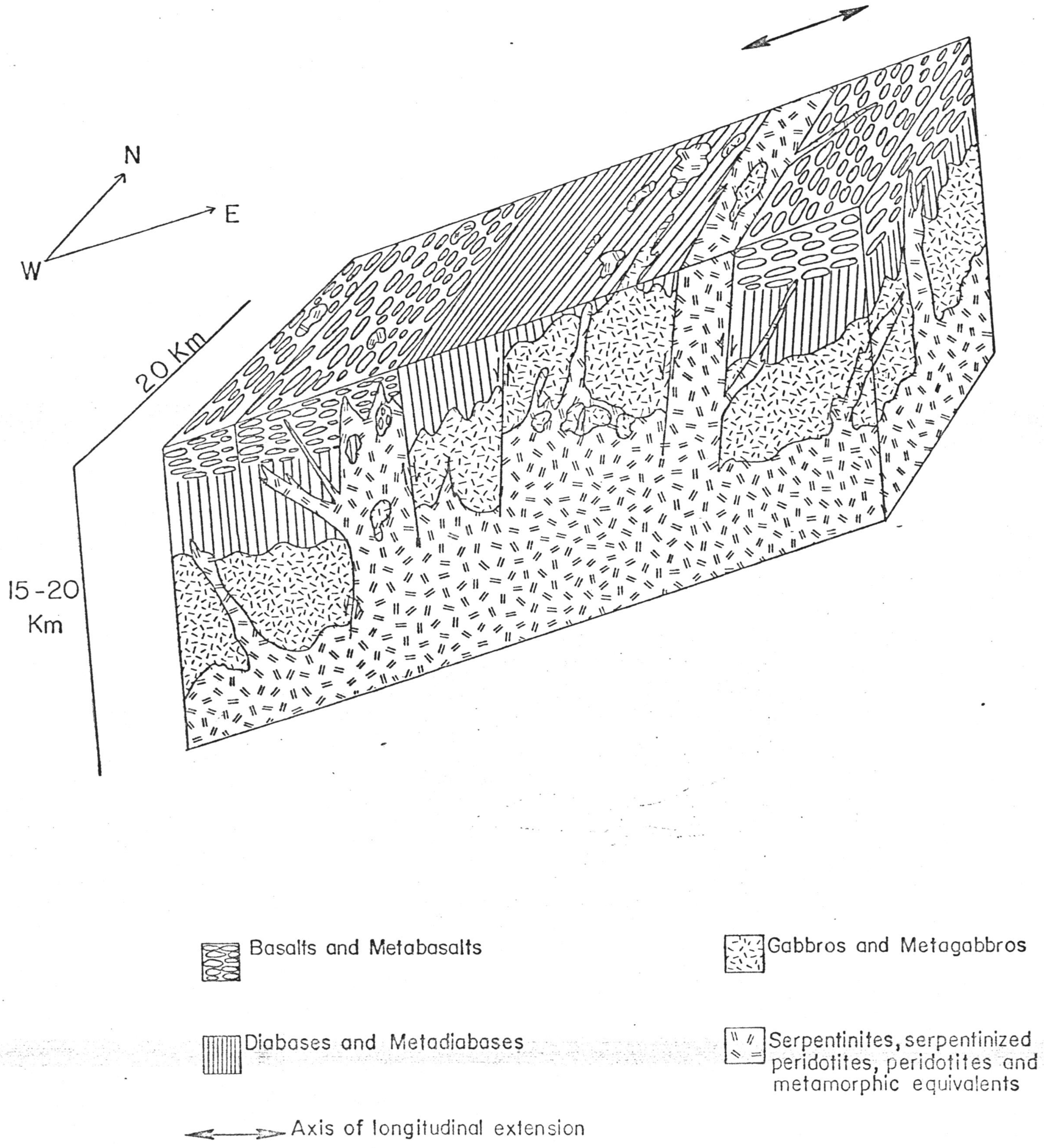
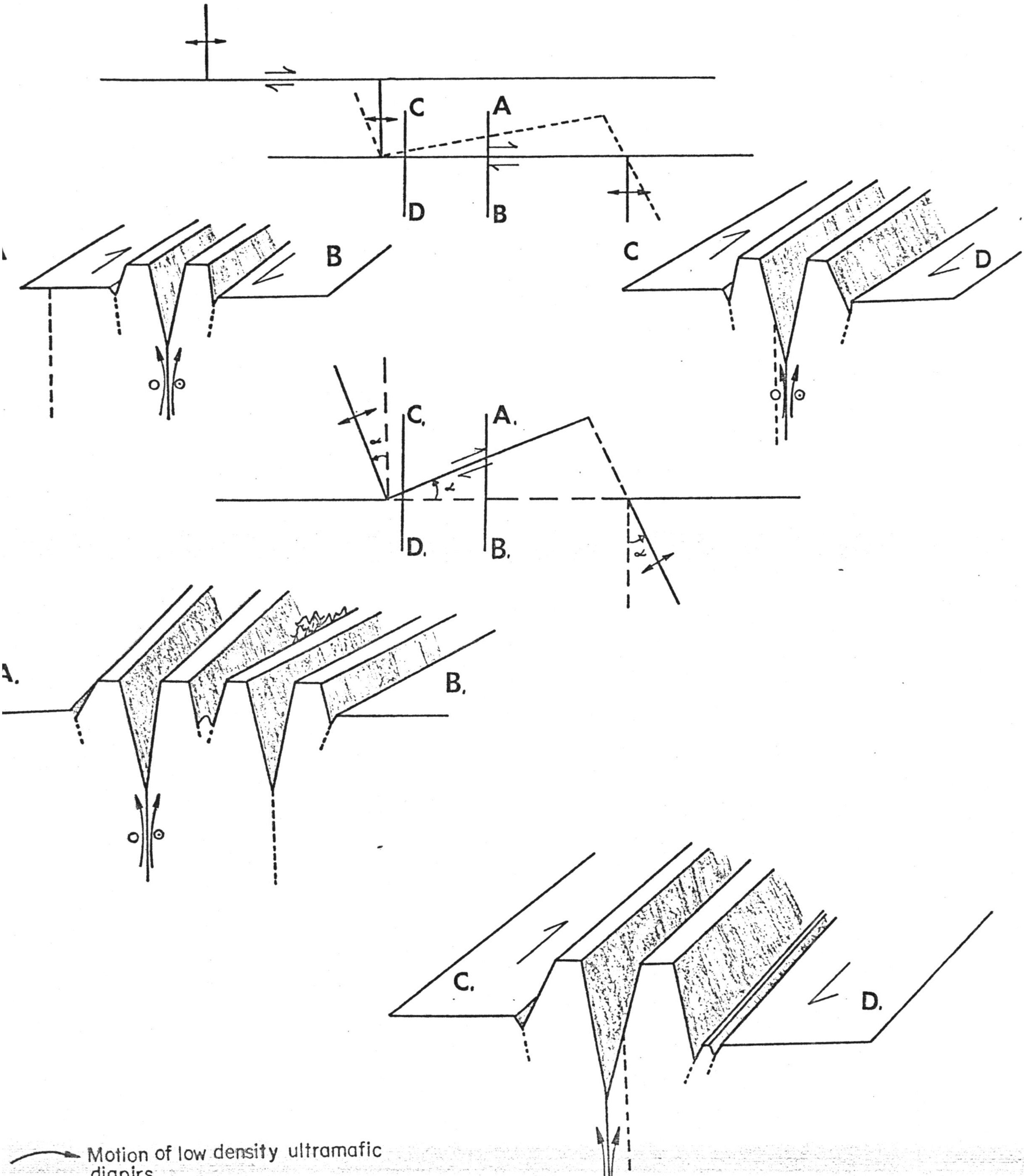

N
E
W
20 Km
15-20
Km
Basalts and Metabasalts
Diabases and Metadiabases
Axis of longitudinal extension
Gabbros and Metagabbros
Serpentinites, serpentinized peridotites, peridotites and metamorphic equivalents

Figure 25

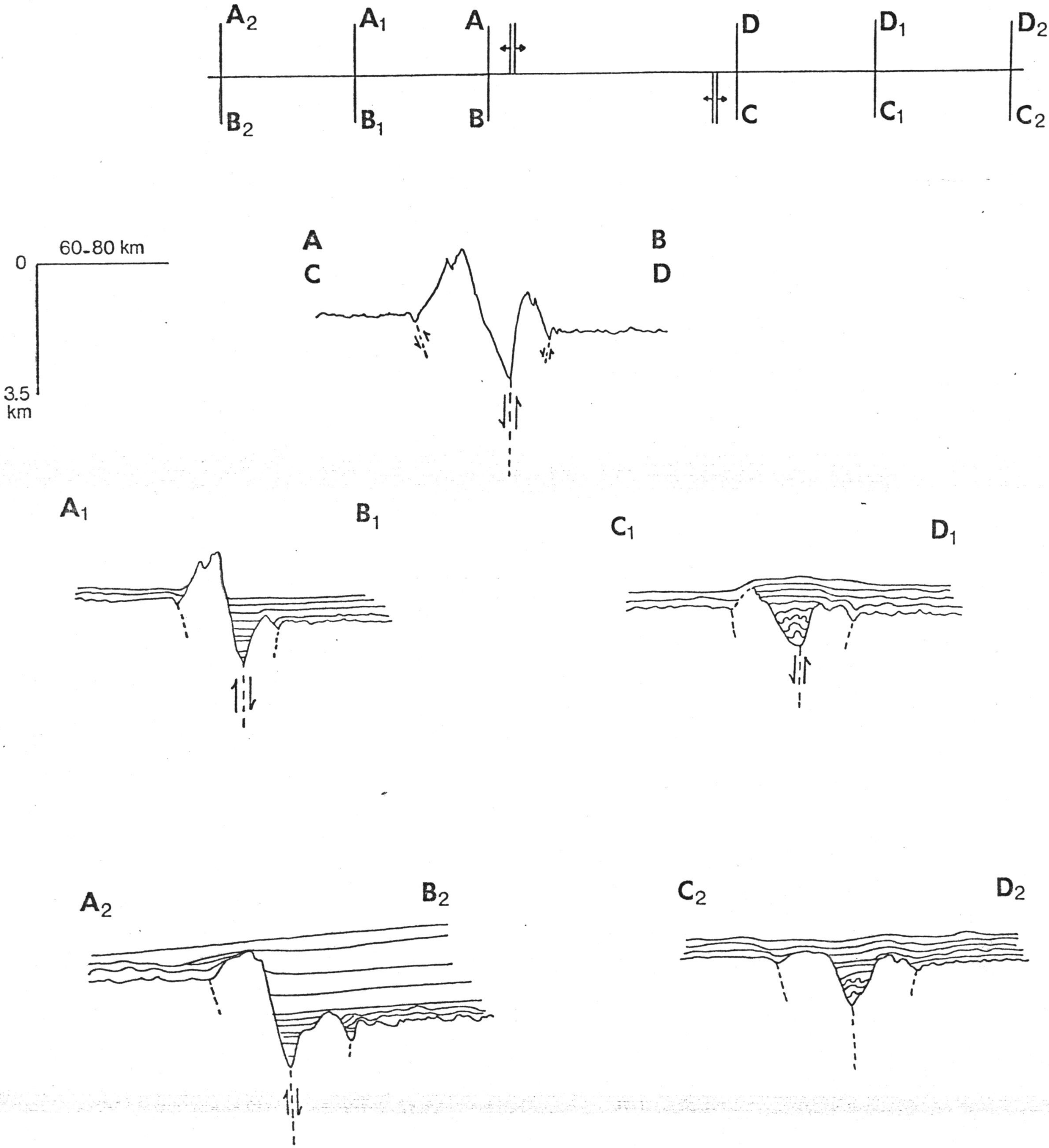


Motion of low density ultramafic diapirs

oo Denotes the shear motion

---- In planar views denotes old transform trends; faults in the sections

Figure 26



A2
A1
A
D
D1
D2
B2
B1
B
C
C1
C2
0
60-80 km
3.5 km
A
B
C
D
A1
B1
C1
D1
A2
B2
C2
D2

Figure 27

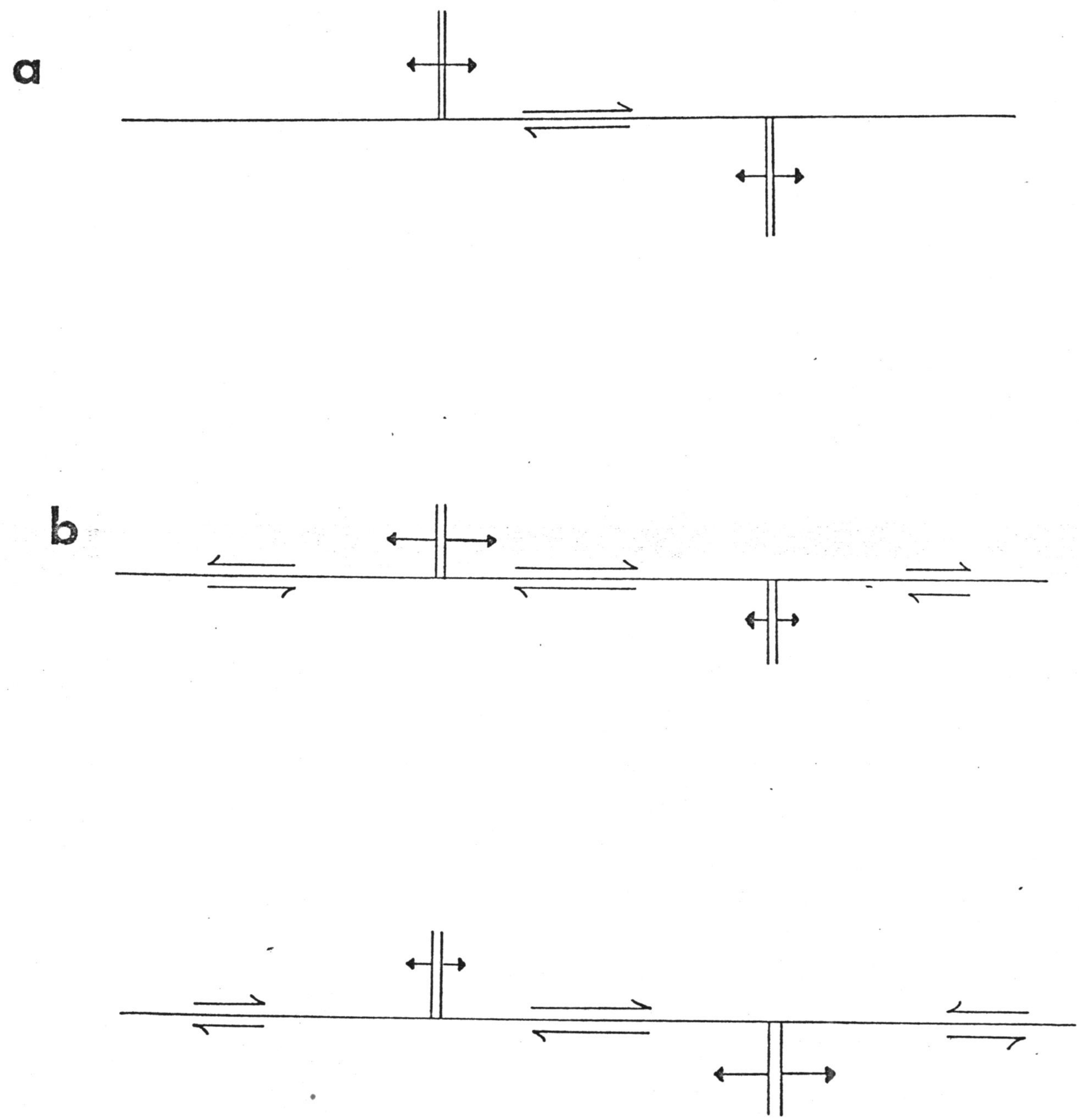


Figure 28

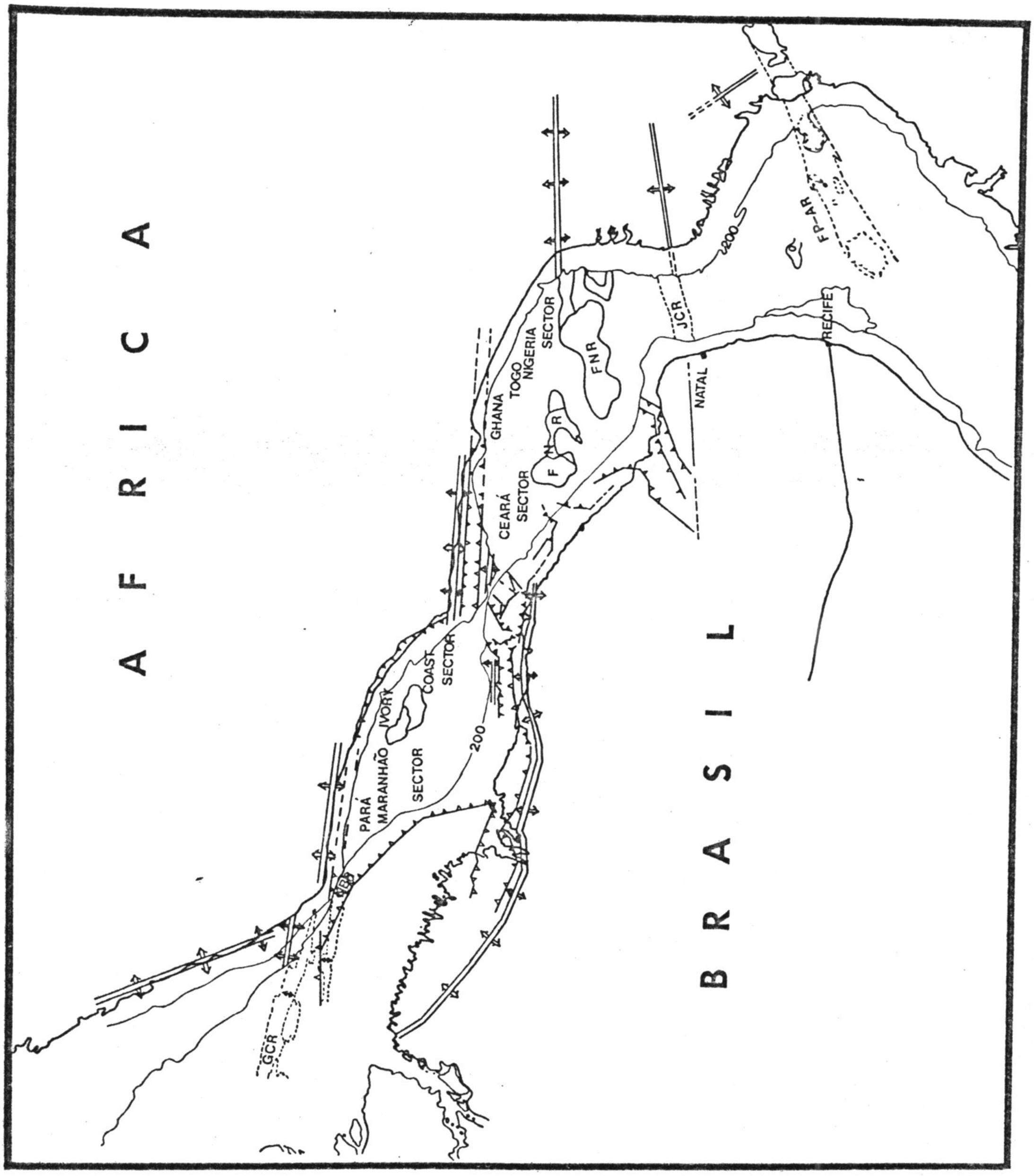

AFRICA
BRASIL
PARÁ
MARANHÃO
SECTOR
IVORY
COAST
SECTOR
CEARÁ
SECTOR
GHANA
TOGO
NIGERIA
SECTOR
FNR
JCR
NATAL
RECIFE
FP-AR
GCR
200

Figure 29

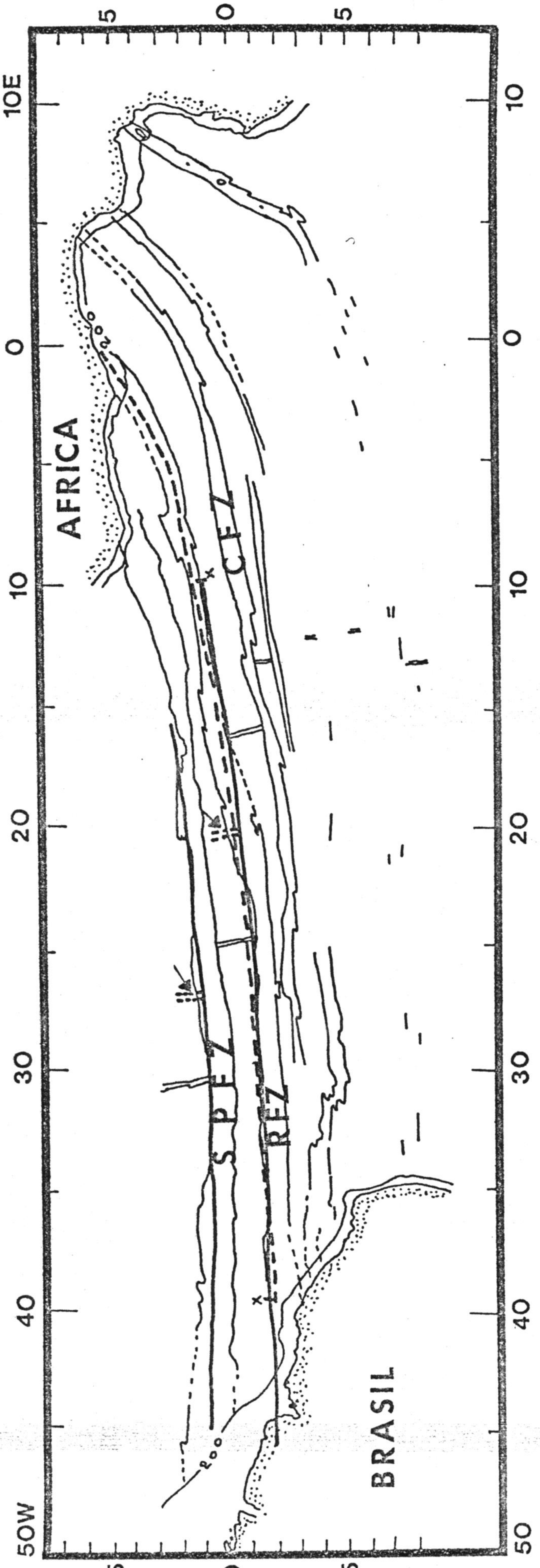


Figure 30

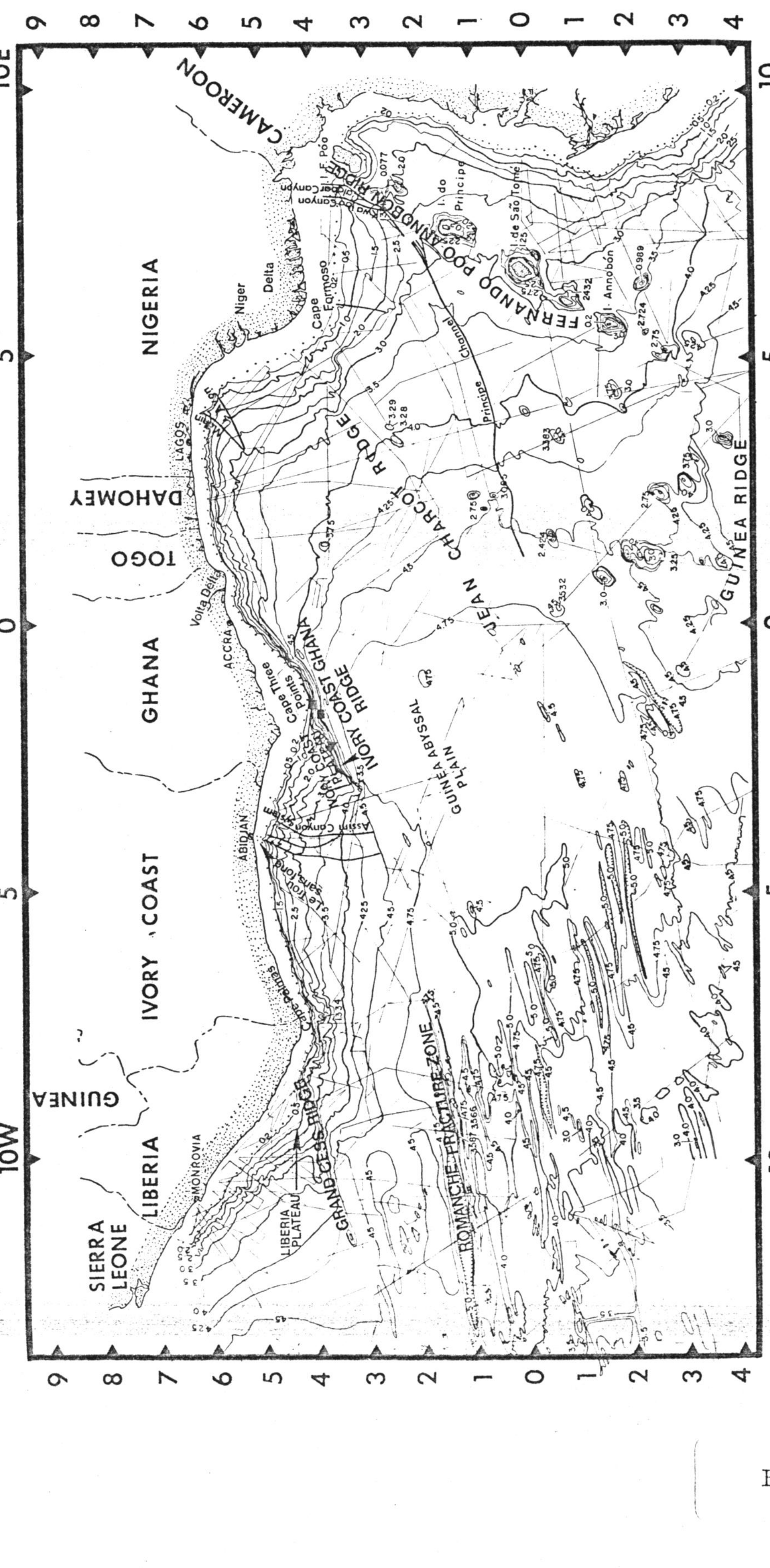

SIERRA LEONE
LIBERIA
GUINEA
IVORY COAST
GHANA
TOGO
DAHOMEY
NIGERIA
CAMEROON
MONROVIA
ABIDJAN
ACCRA
LAGOS
Volta Delta
Niger Delta
Cape Three Points
Cape Formoso
LIBERIA PLATEAU
GRAND CESS RIDGE
ROMANCHE FRACTURE ZONE
IVORY COAST GHANA RIDGE
GUINEA ABYSSAL PLAIN
CHARCOT RIDGE
JEAN CHARCOT RIDGE
GUINEA RIDGE
FERNANDO POO ANNOBON RIDGE
Principe Channel
I. do Principe
I. de São Tomé
I. Annobon
10W
5
0
10E
10

Figure 31

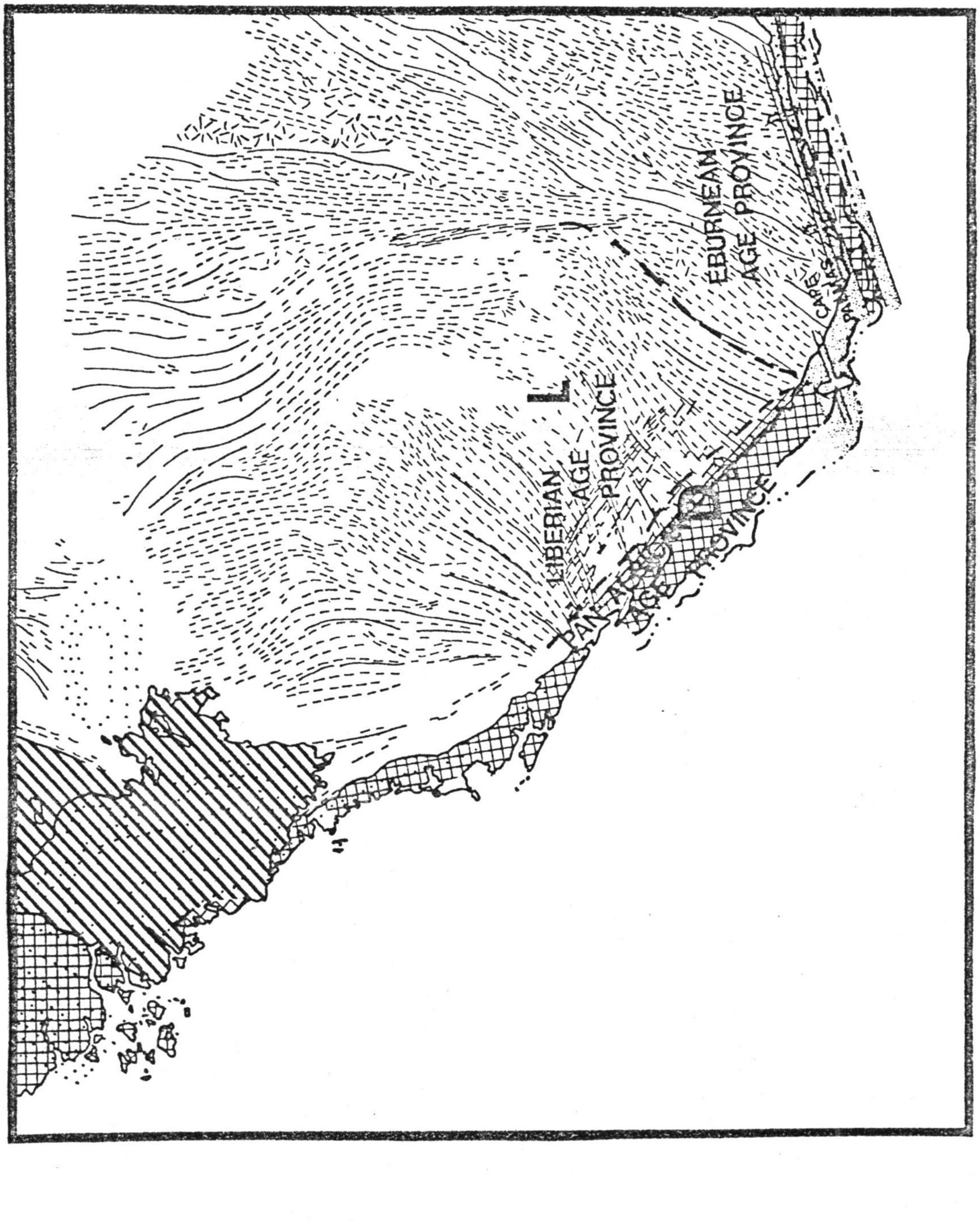


Figure 32

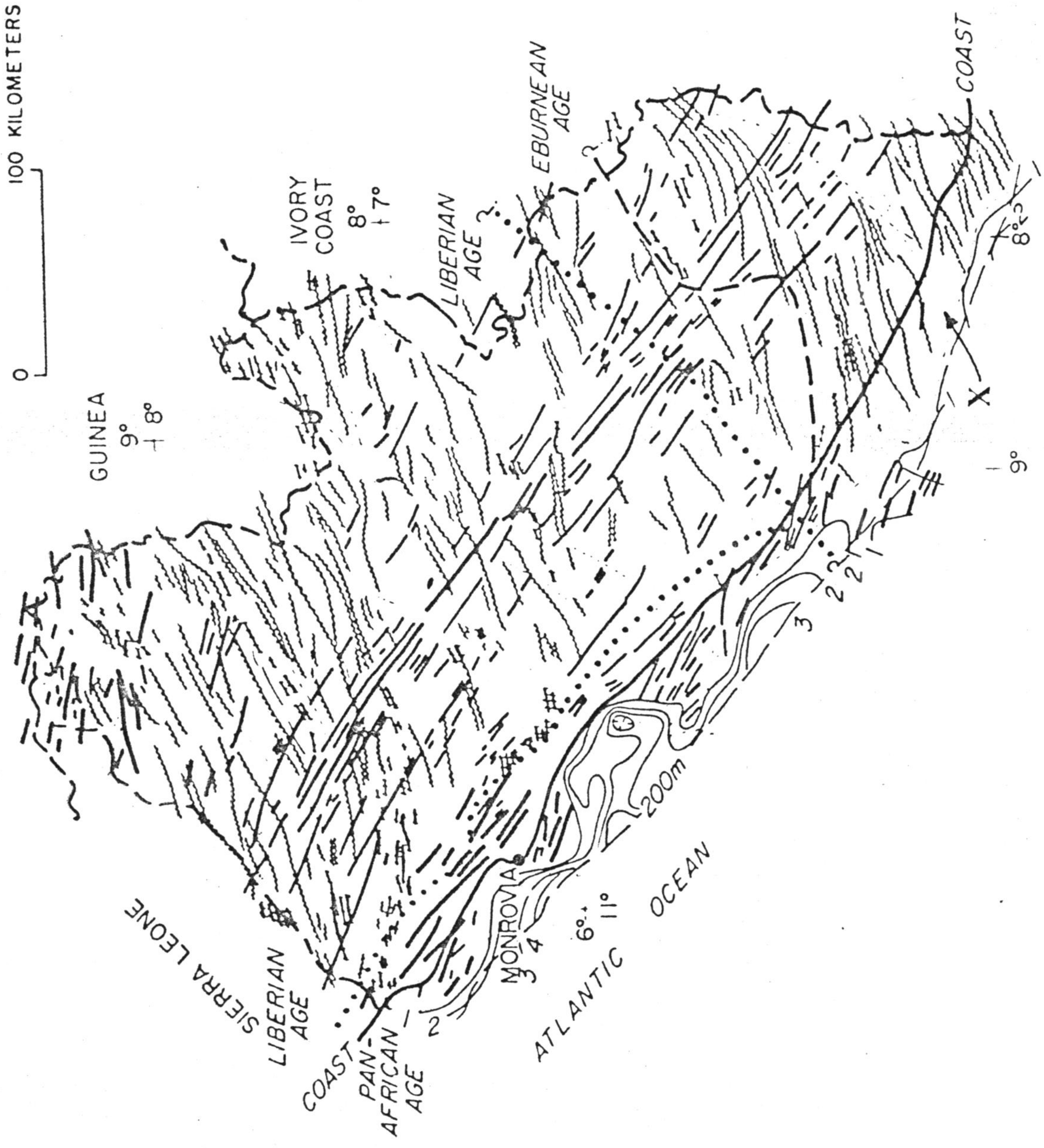

0
100 KILOMETERS
GUINEA
IVORY COAST
LIBERIAN AGE
EBURNEAN AGE
COAST
SIERRA LEONE
LIBERIAN AGE
COAST
PAN-AFRICAN AGE
MONROVIA
200m
ATLANTIC OCEAN

Figure 33

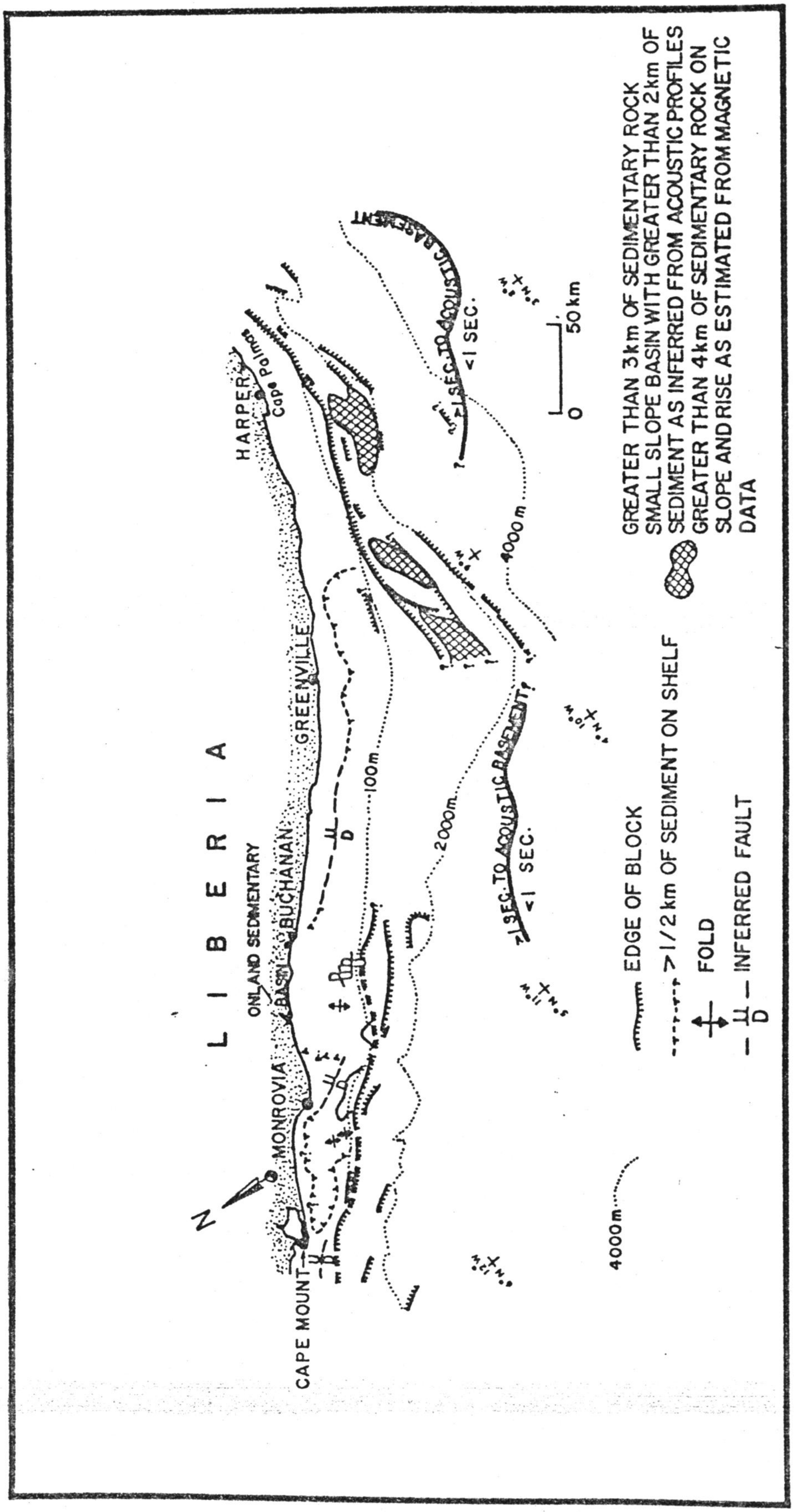

L I B E R I A
N
CAPE MOUNT
MONROVIA
ONLAND SEDIMENTARY BASIN
BUCHANAN
GREENVILLE
HARPER
Cape Palmas
100m
2000m
4000m
4000m
≥1 SEC. TO ACOUSTIC BASEMENT
<1 SEC.
≥1 SEC. TO ACOUSTIC BASEMENT?
≥1 SEC. TO ACOUSTIC BASEMENT
<1 SEC.
0
50 km
EDGE OF BLOCK
>1/2 km OF SEDIMENT ON SHELF
FOLD
U D — INFERRED FAULT
GREATER THAN 3 km OF SEDIMENTARY ROCK
SMALL SLOPE BASIN WITH GREATER THAN 2 km OF SEDIMENT AS INFERRED FROM ACOUSTIC PROFILES
GREATER THAN 4 km OF SEDIMENTARY ROCK ON SLOPE AND RISE AS ESTIMATED FROM MAGNETIC DATA

Figure 34

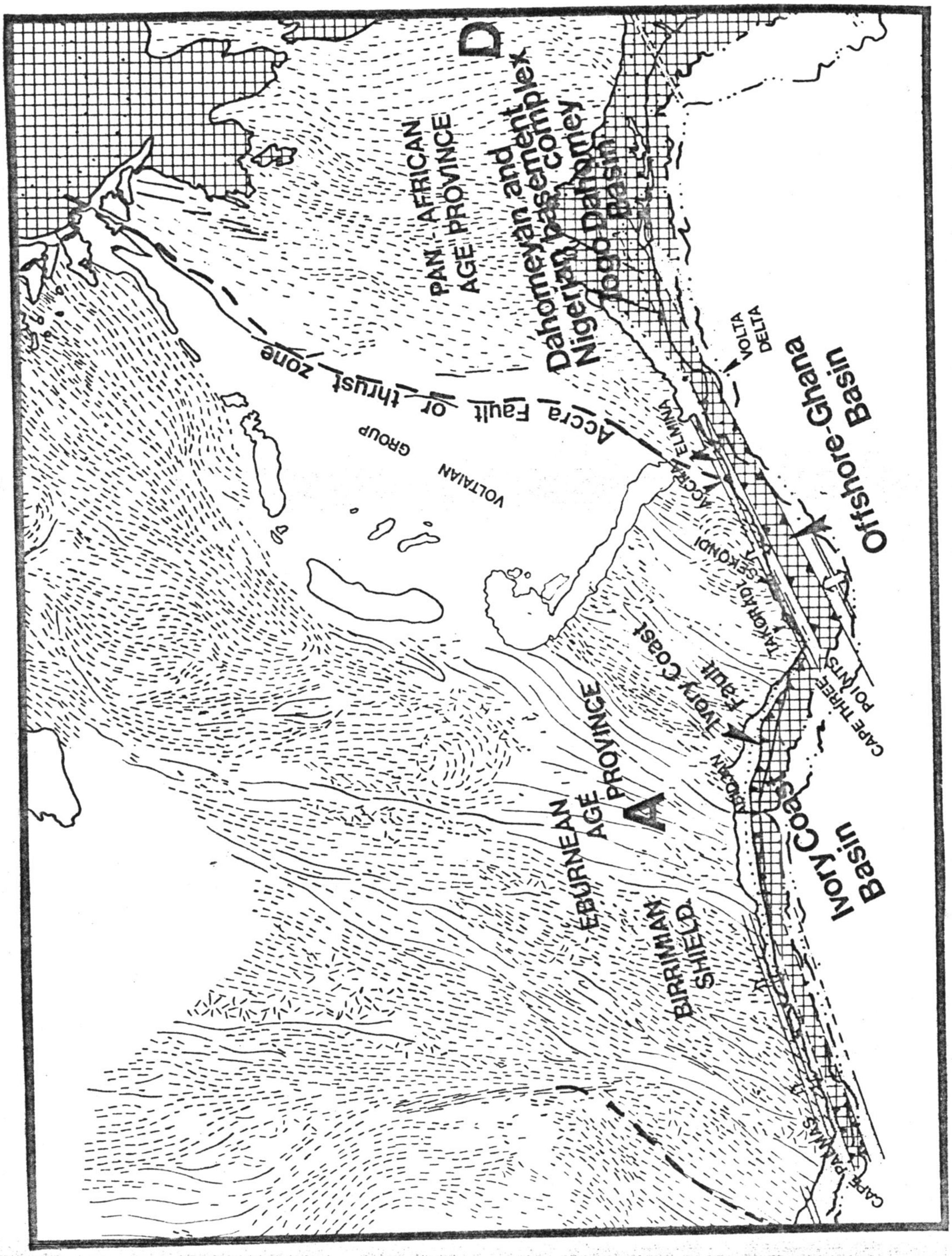

D
PAN - AFRICAN
AGE PROVINCE
Dahomeyan and
Nigerian basement complex
Togo Dahomey
Basin
VOLTA
DELTA
Offshore-Ghana
Basin
Accra Fault or thrust zone
VOLTAIAN
GROUP
ELMINA
ACCRA
SEKONDI
TAKORADI
CAPE THREE
POINTS
Ivory Coast
Fault
EBURNEAN
AGE
PROVINCE
A
ABIDJAN
Ivory Coast
Basin
BIRRIMIAN
SHIELD
CAPE PALMAS

Figure 35

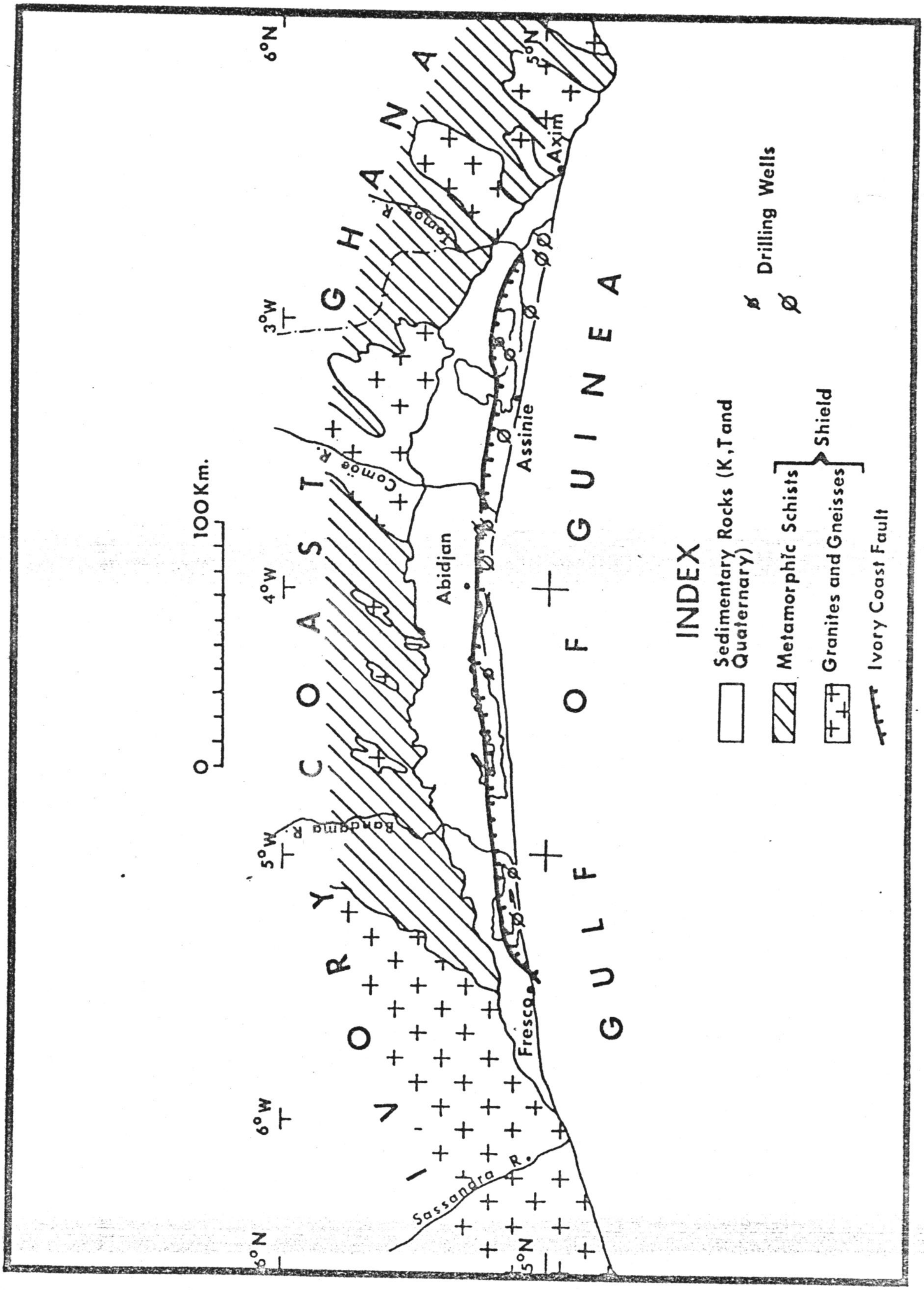

IVORY COAST
GHANA
GULF OF GUINEA
Abidjan
Assinie
Axim
Fresco
Sassandra R.
Bandama R.
Comoe R.
0
100 Km.
6°W
5°W
4°W
3°W
5°N
6°N
INDEX
Sedimentary Rocks (K,T and Quaternary)
Metamorphic Schists
Granites and Gneisses
Shield
Ivory Coast Fault
Drilling Wells

Figure 36

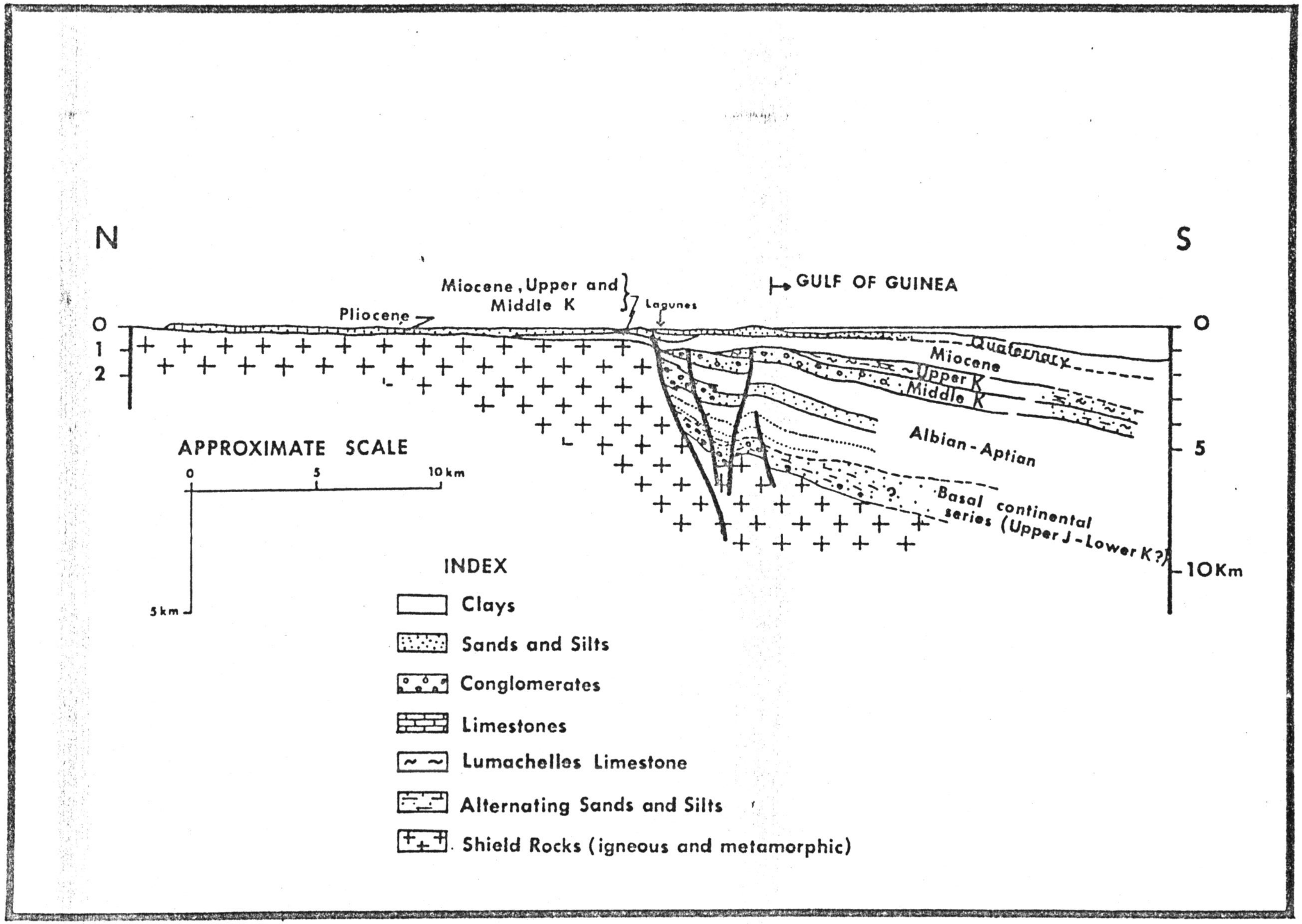

N
S
Miocene, Upper and Middle K
Pliocene
Lagunes
GULF OF GUINEA
Quaternary
Miocene
Upper K
Middle K
Albian - Aptian
Basal continental series (Upper J - Lower K?)
0
1
2
0
5
10Km
APPROXIMATE SCALE
0
5
10 km
5 km
INDEX
Clays
Sands and Silts
Conglomerates
Limestones
Lumachelles Limestone
Alternating Sands and Silts
Shield Rocks (igneous and metamorphic)

Figure 37

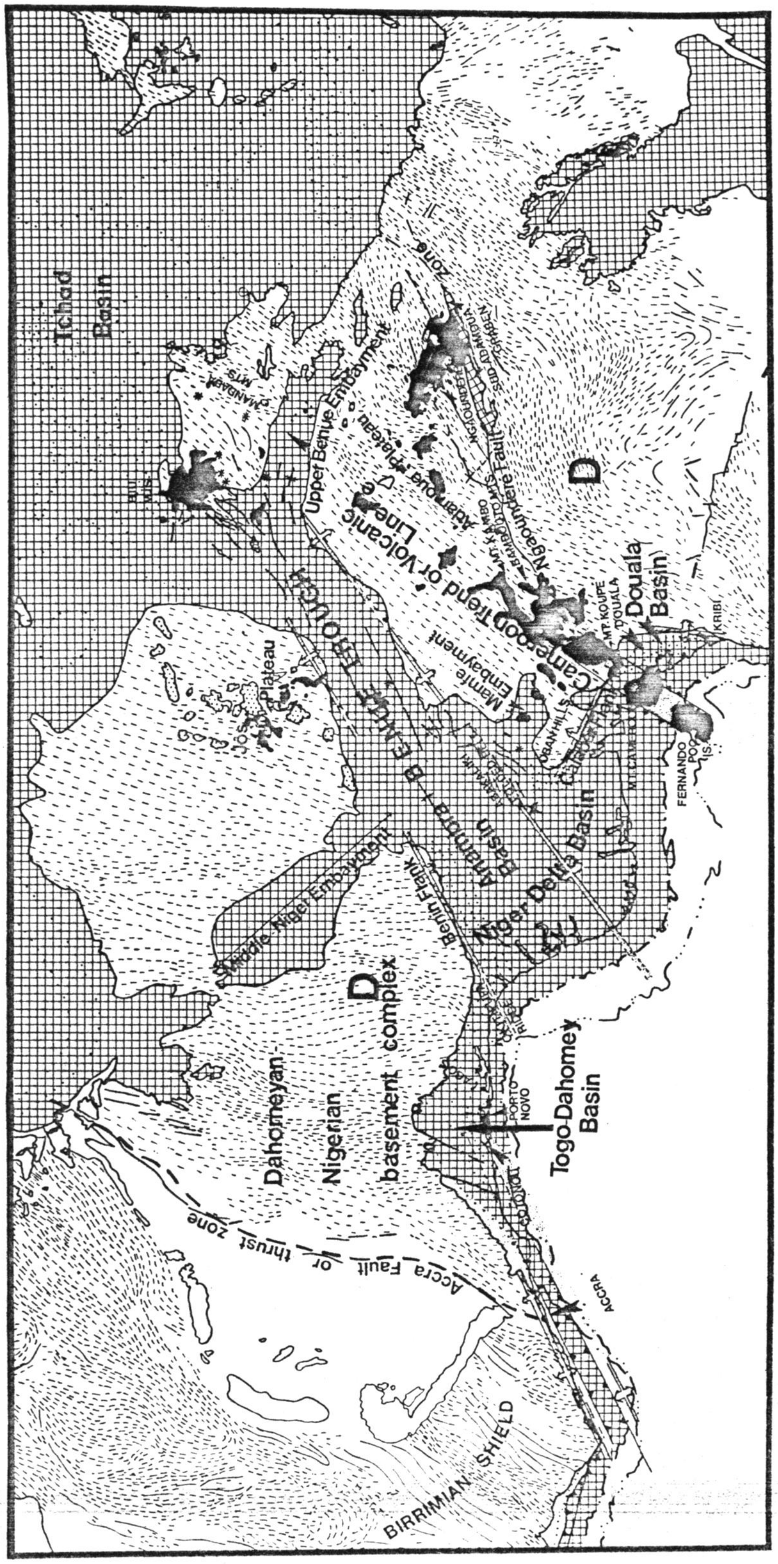

Tchad Basin
Upper Benue Embayment
Adamoua Plateau
Cameroon Trend or Volcanic Line
Ngaoundere Fault
D
Douala Basin
MT. KOUPE
DOUALA
KRIBI
Mamfe Embayment
BENUE TROUGH
Jos Plateau
OBAN HILLS
Calabar Flank
MT. CAMEROON
FERNANDO PO IS.
Niger Delta Basin
Anambra Basin
Benin Flank
Middle Niger Embayment
D
Dahomeyan
Nigerian
basement complex
Togo-Dahomey Basin
PORTO NOVO
Accra Fault or thrust zone
ACCRA
BIRRIMIAN SHIELD

Figure 38

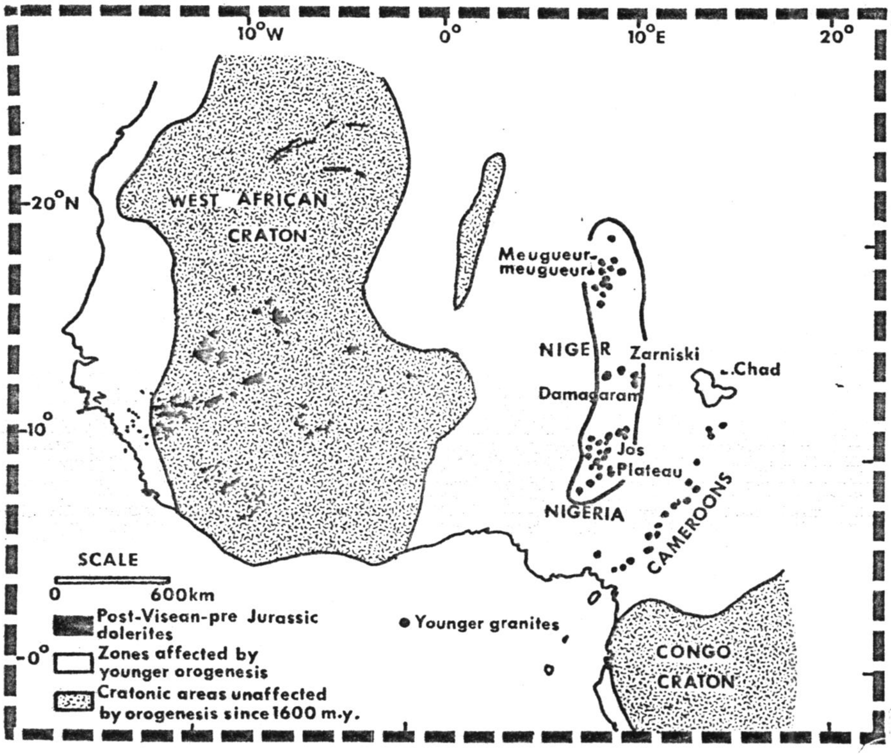
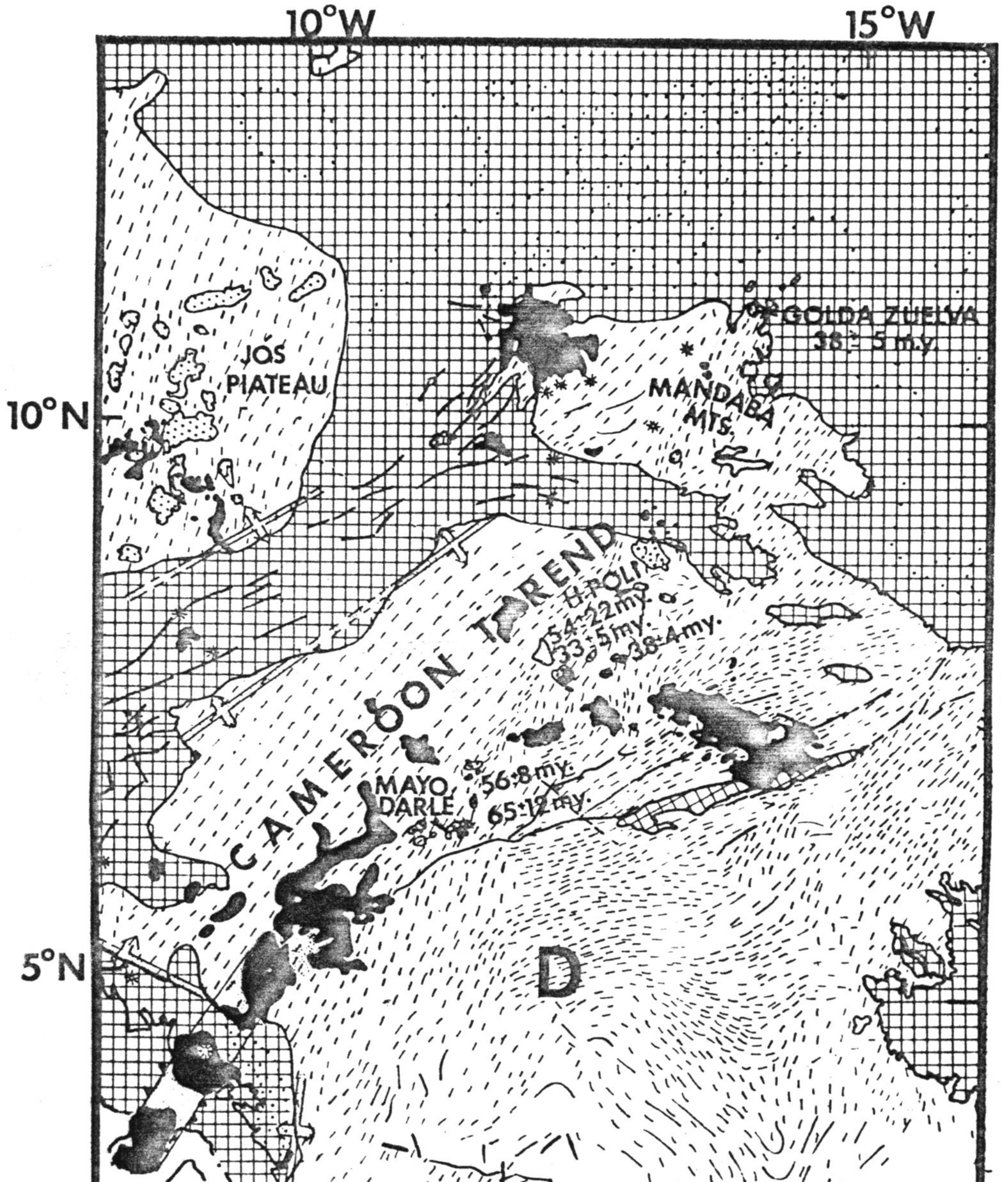

10°W
0°
10°E
20°
20°N
10°
0°
WEST AFRICAN CRATON
Meugueur-meugueur
NIGER
Zarniski
Chad
Damagaram
Jos Plateau
NIGERIA
CAMEROONS
CONGO CRATON
SCALE
0
600km
Post-Visean-pre Jurassic dolerites
Zones affected by younger orogenesis
Cratonic areas unaffected by orogenesis since 1600 m.y.
Younger granites

Figure 39



10°W
15°W
10°N
5°N
JOS
PLATEAU
GOLDA ZUELVA
38 ± 5 m.y.
MANDABA
MTS.
CAMEROON TREND
H. POLI
54 ± 22 my.
33 ± 5 my.
38 ± 4 my.
MAYO
DARLE
56 ± 8 my.
65 ± 12 my.
D

Figure 40

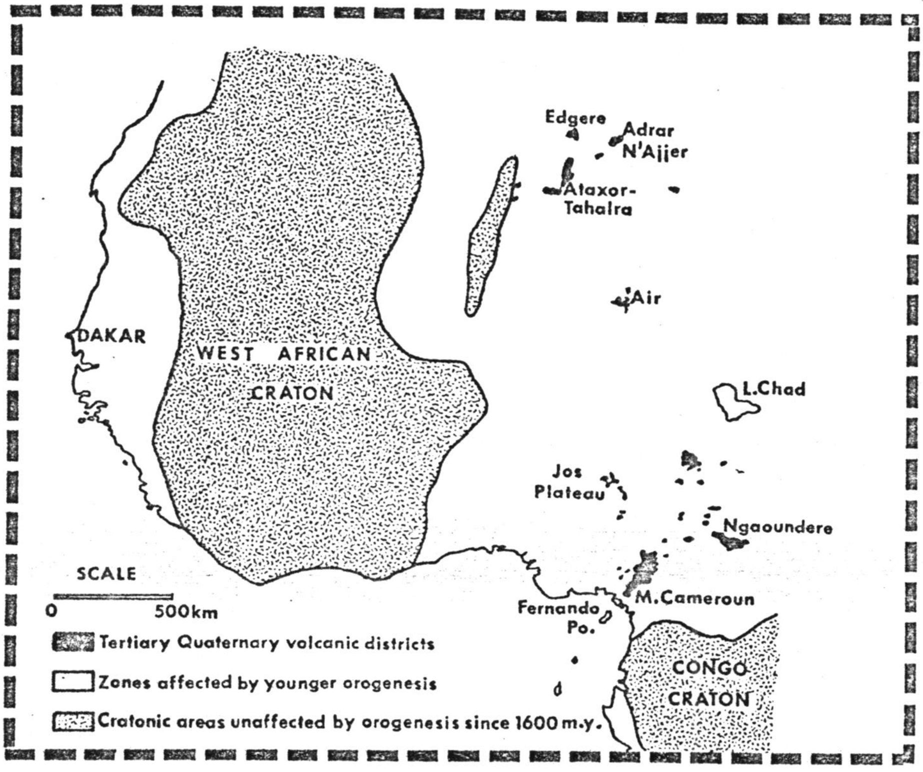

Edgere
Adrar
N'Ajjer
Ataxor-
Tahalra
Air
DAKAR
WEST AFRICAN
CRATON
L.Chad
Jos
Plateau
Ngaoundere
SCALE
0
500km
Fernando
Po.
M.Cameroun
CONGO
CRATON
Tertiary Quaternary volcanic districts
Zones affected by younger orogenesis
Cratonic areas unaffected by orogenesis since 1600 m.y.

Figure 41

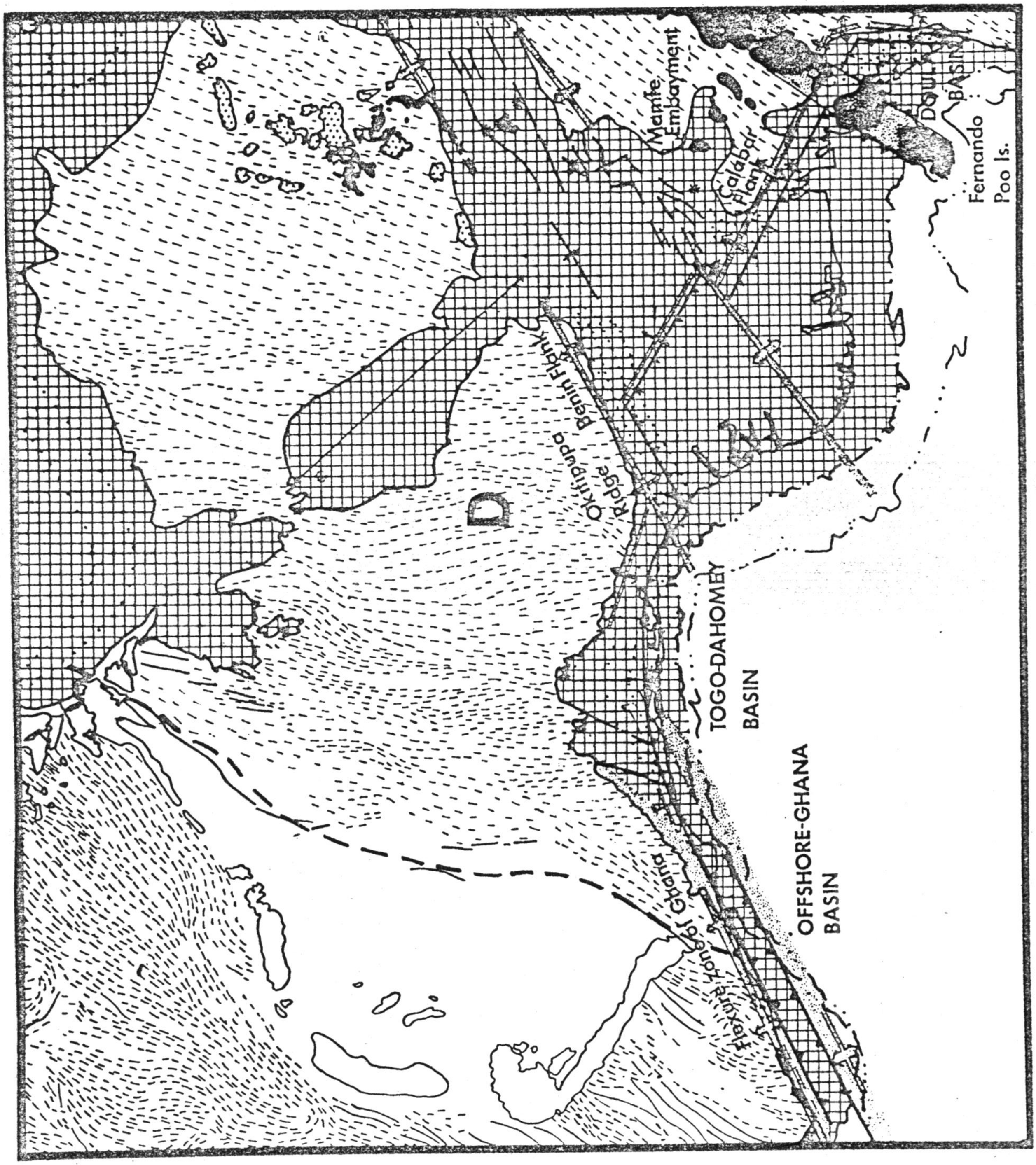
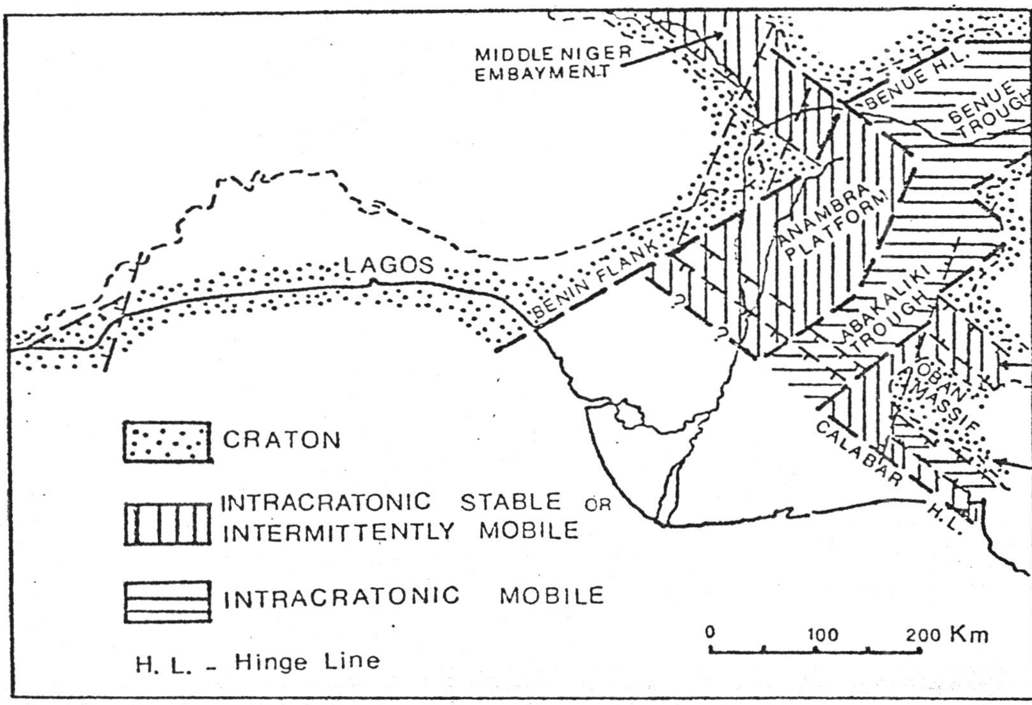
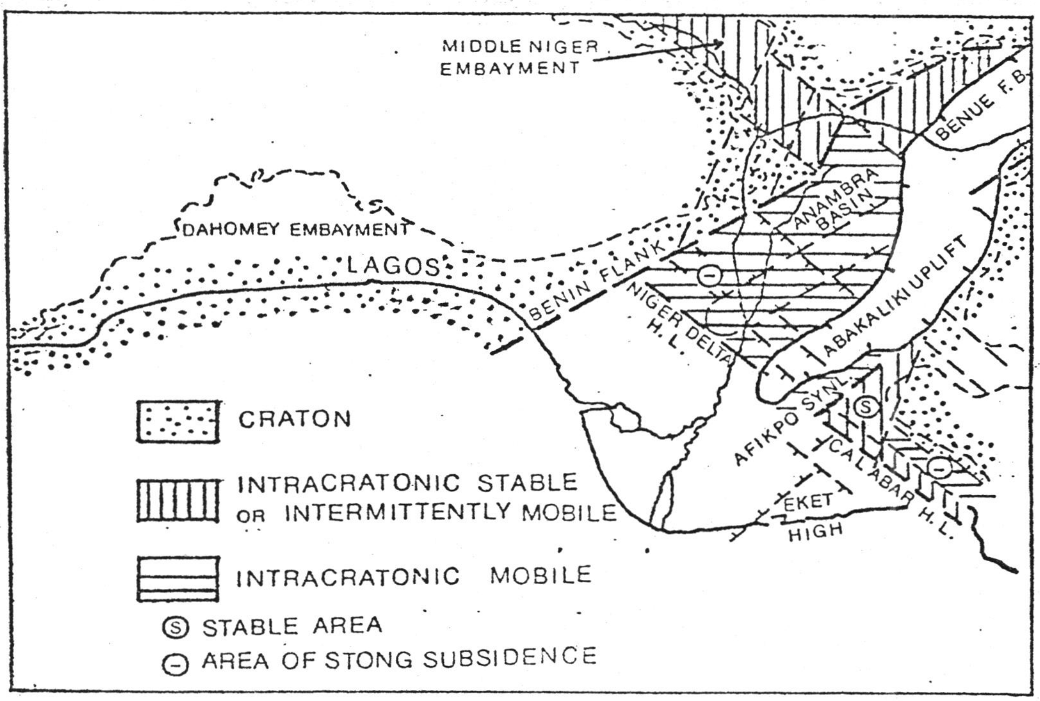

Mamfe Embayment
Calabar Flank
BASIN
Fernando Poo Is.
Benin Flank
Okitipupa Ridge
D
TOGO-DAHOMEY BASIN
OFFSHORE-GHANA BASIN
Flexure Zone of Ghana

Figure 42



Megatectonic frame - MIDDLE ALBIAN to SANTONIAN



Megatectonic frame- CAMPANIAN to EOCENE

Figure 43

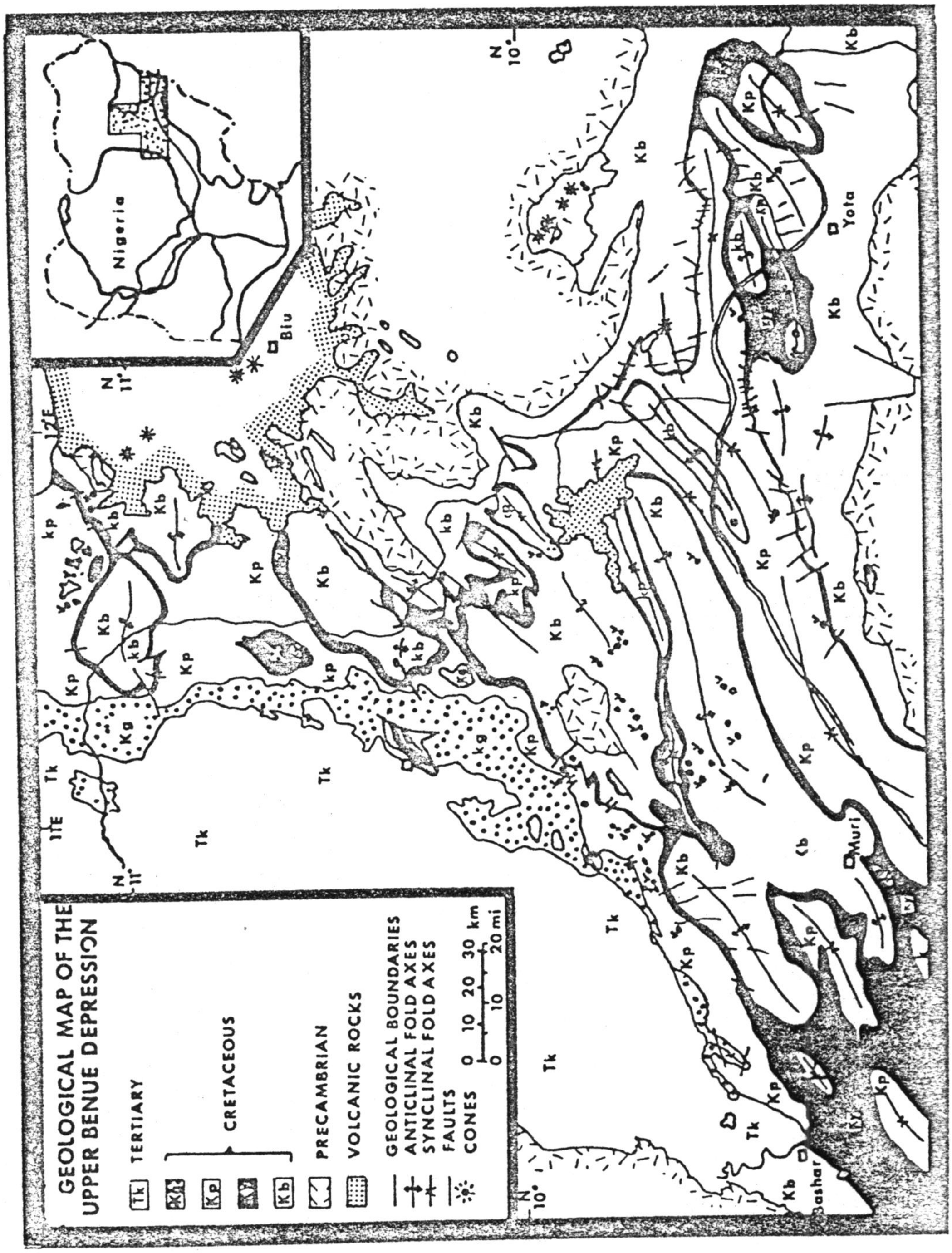
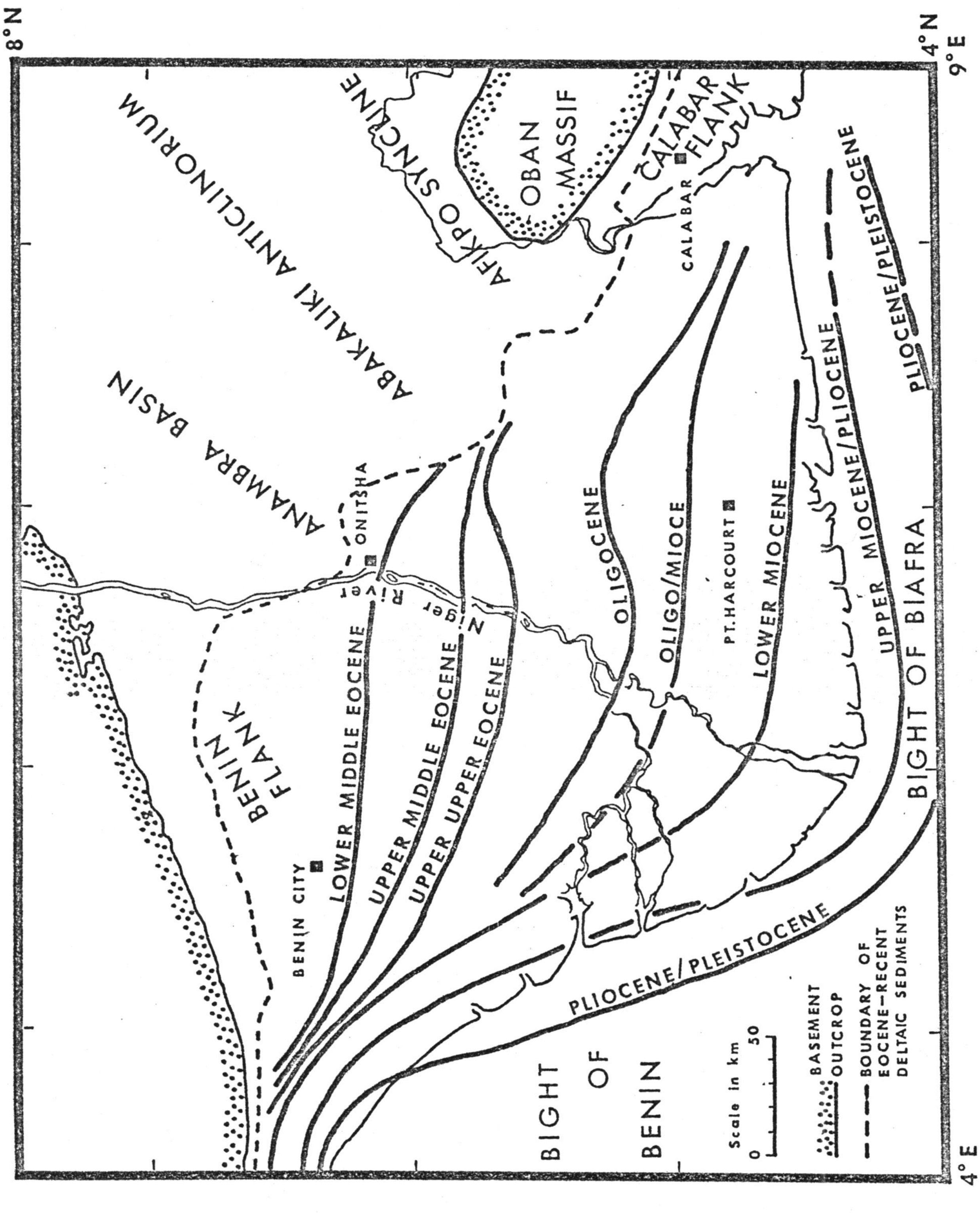
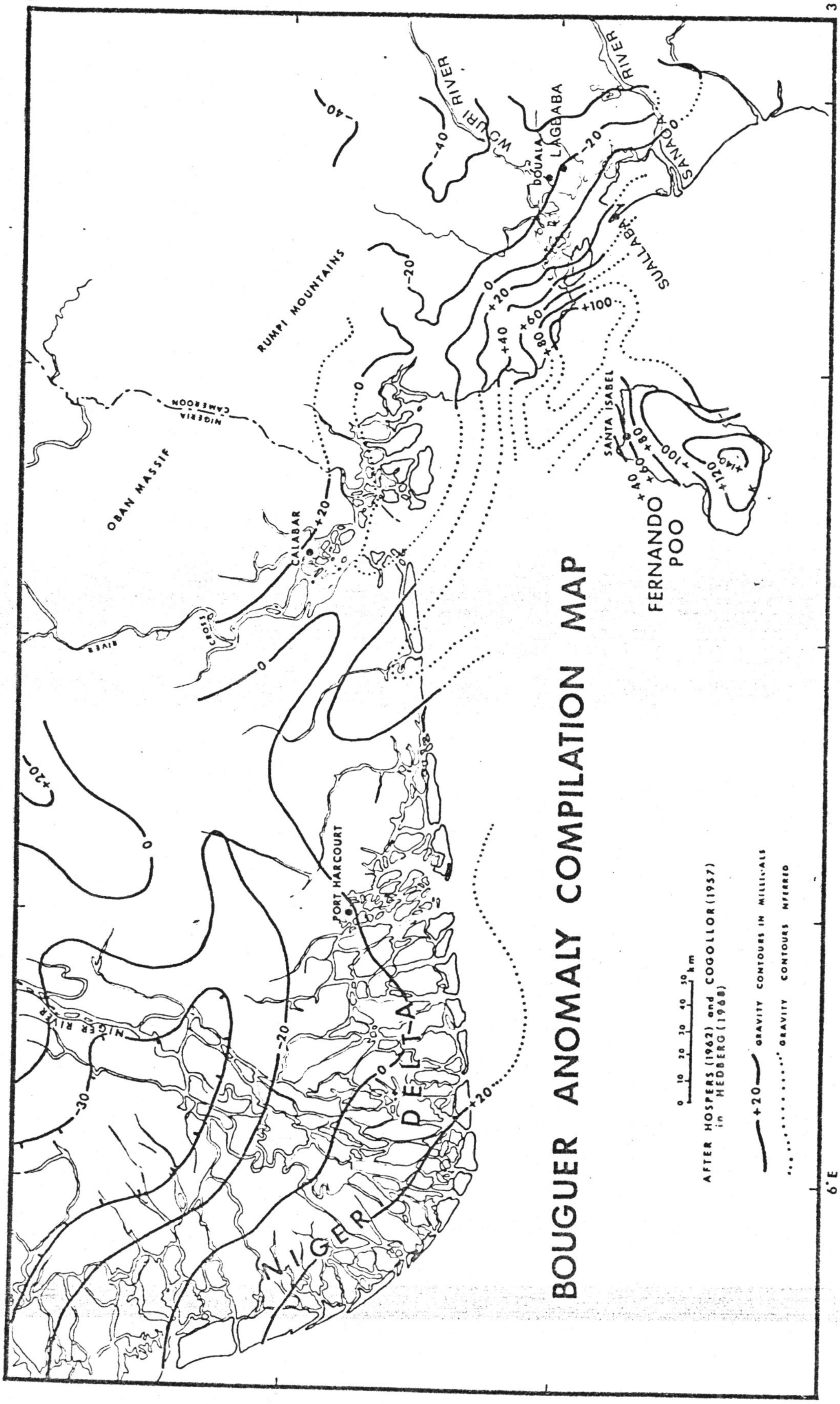

GEOLOGICAL MAP OF THE
UPPER BENUE DEPRESSION
Tk TERTIARY
Kg
Kp
Kb
CRETACEOUS
PRECAMBRIAN
VOLCANIC ROCKS
GEOLOGICAL BOUNDARIES
ANTICLINAL FOLD AXES
SYNCLINAL FOLD AXES
FAULTS
CONES
0 10 20 30 km
0 10 20 mi
Nigeria
Biu
Yola
Muri
Bashar
Tk
Kg
Kp
Kb

Figure 44



8°N
4°N 9°E
4°E
ABAKALIKI ANTICLINORIUM
ANAMBRA BASIN
AFIKPO SYNCLINE
OBAN MASSIF
CALABAR FLANK
CALABAR
ONITSHA
Niger River
BENIN FLANK
BENIN CITY
LOWER MIDDLE EOCENE
UPPER MIDDLE EOCENE
UPPER UPPER EOCENE
OLIGOCENE
OLIGO/MIOCE
PT. HARCOURT
LOWER MIOCENE
UPPER MIOCENE/PLIOCENE
PLIOCENE/PLEISTOCENE
PLIOCENE/PLEISTOCENE
BIGHT OF BIAFRA
BIGHT OF BENIN
Scale in km
0
50
BASEMENT OUTCROP
BOUNDARY OF EOCENE–RECENT DELTAIC SEDIMENTS

Figure 45



BOUGUER ANOMALY COMPILATION MAP
AFTER HOSPERS (1962) and COGOLLOR (1957) in HEDBERG (1968)
GRAVITY CONTOURS IN MILLIGALS
GRAVITY CONTOURS INFERRED
NIGER DELTA
NIGER RIVER
PORT HARCOURT
CALABAR
CROSS RIVER
OBAN MASSIF
NIGERIA
CAMEROON
RUMPI MOUNTAINS
DOUALA
WOURI RIVER
SANAGA RIVER
LAGBABA
SUALLABA
SANTA ISABEL
FERNANDO POO
6°N
3°N
10°E
6°E

Figure 46

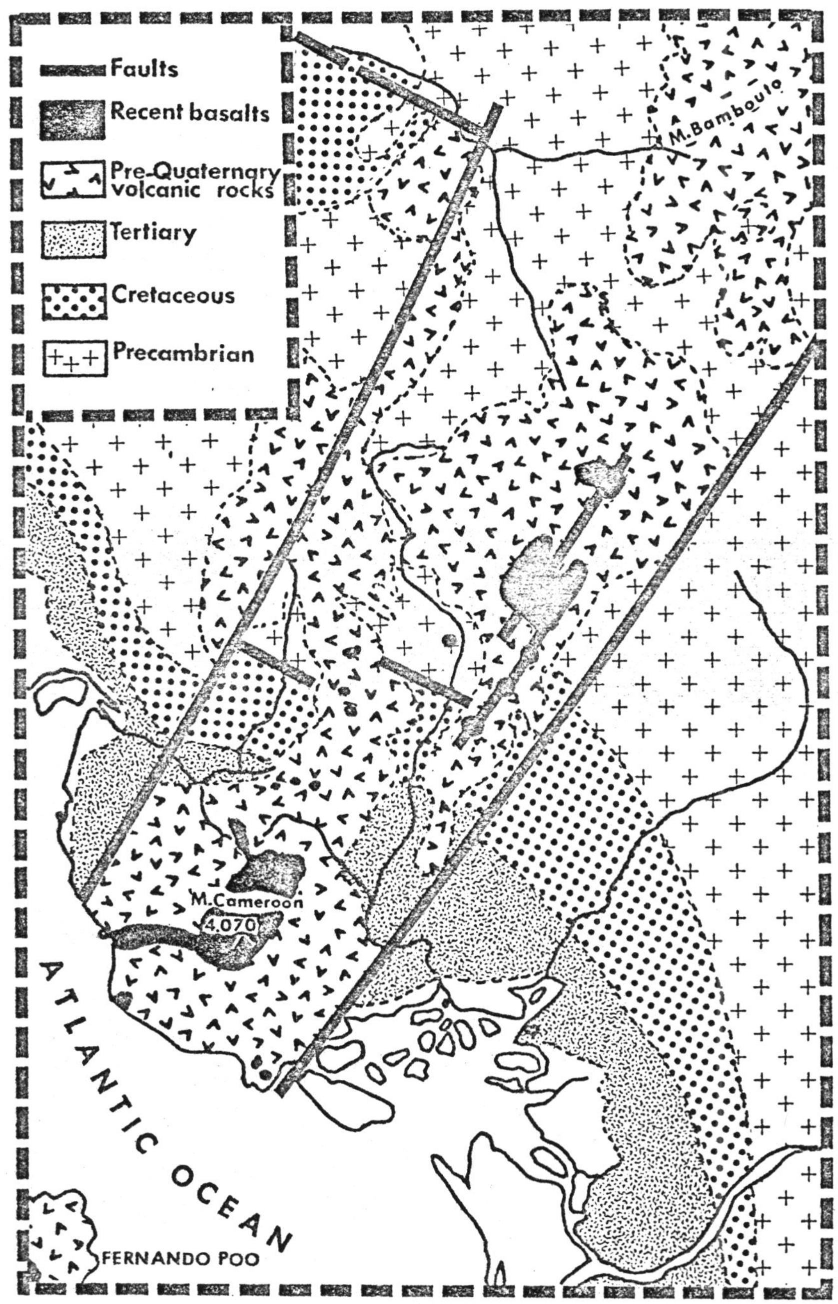

Faults
Recent basalts
Pre-Quaternary volcanic rocks
Tertiary
Cretaceous
Precambrian
M.Bamboulo
M.Cameroon
4.070
ATLANTIC OCEAN
FERNANDO POO

Figure 47

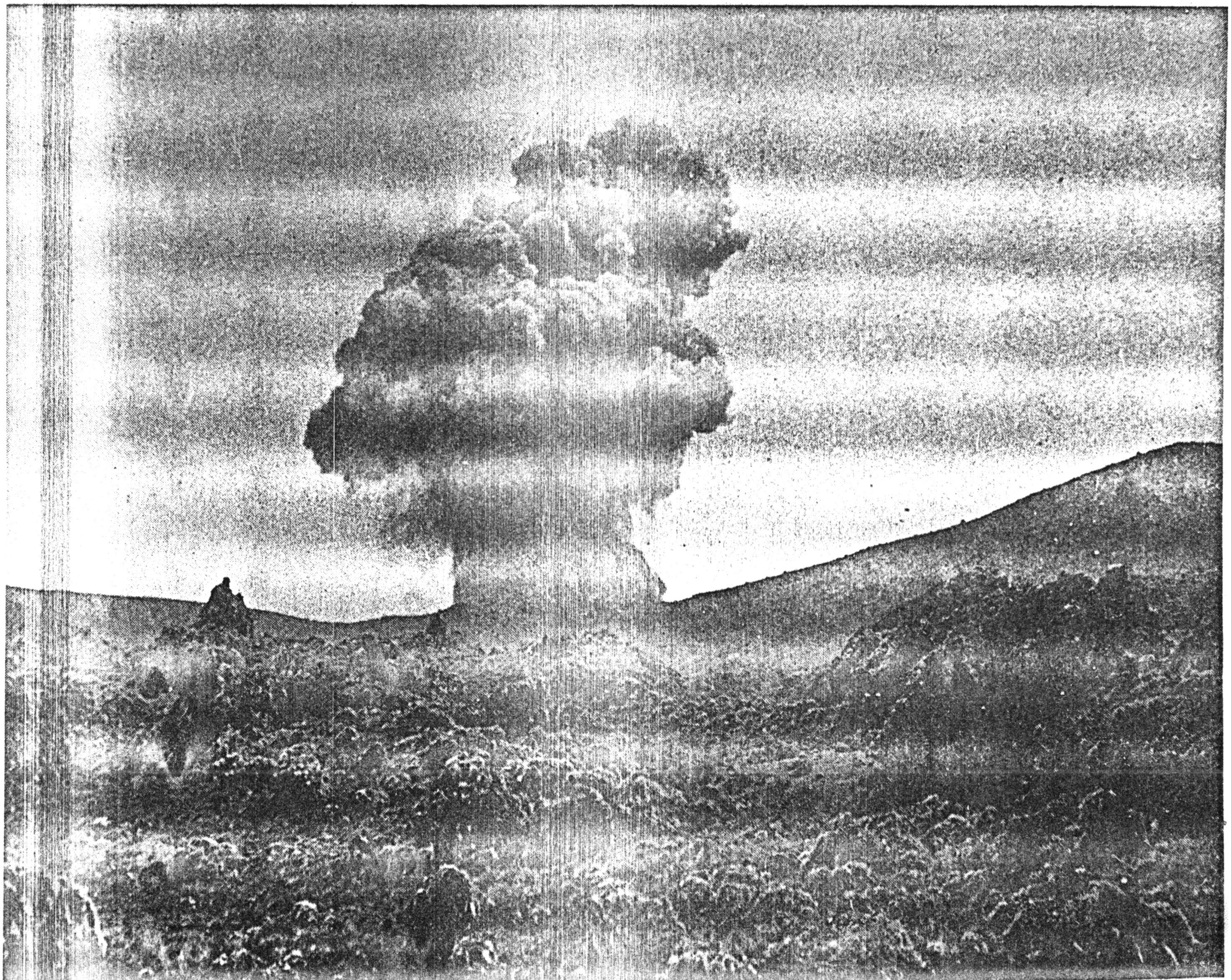
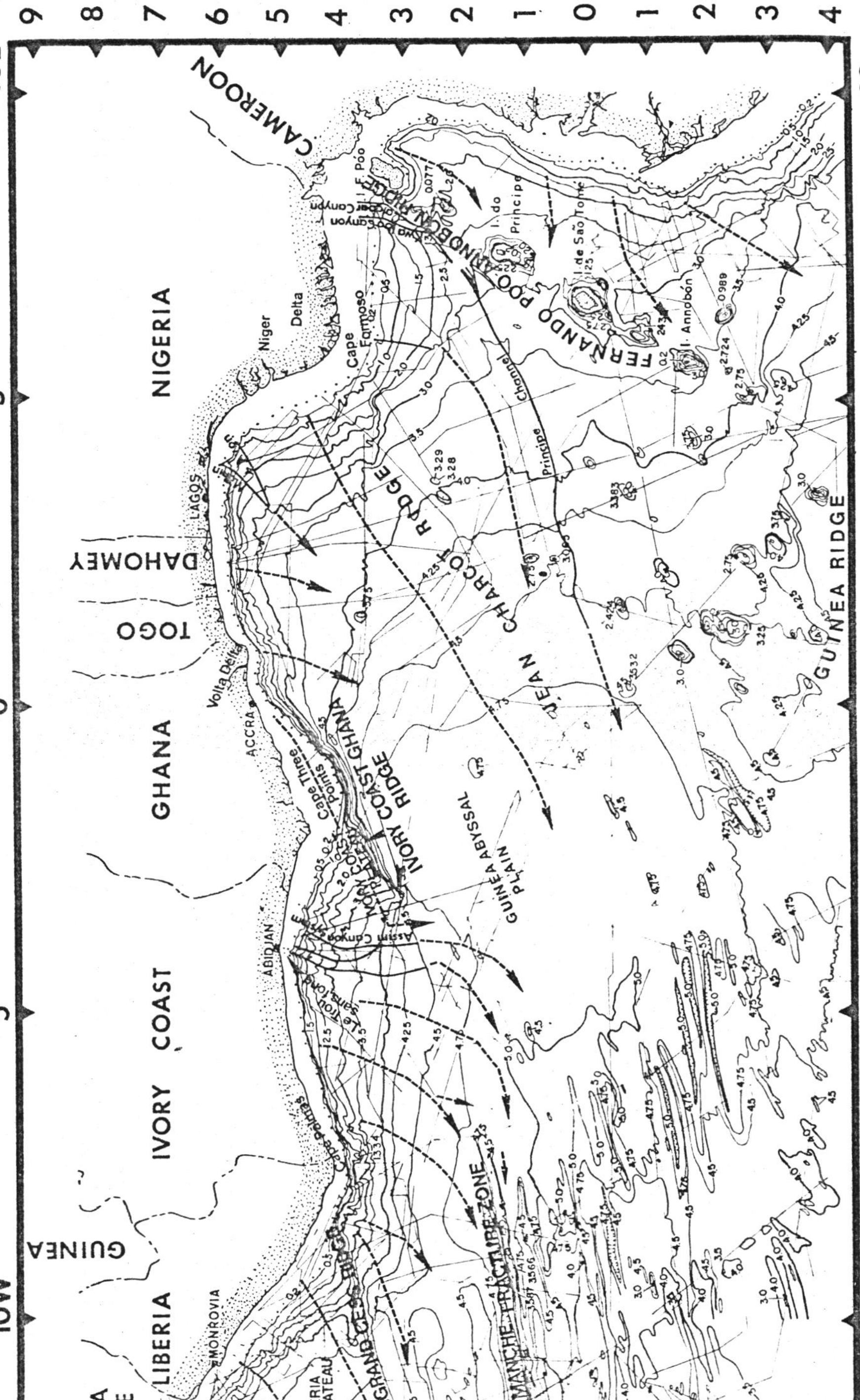


Figure 48



10E
5
0
5
10W
CAMEROON
NIGERIA
DAHOMEY
TOGO
GHANA
IVORY COAST
GUINEA
LIBERIA
SIERRA LEONE
MONROVIA
ABIDJAN
ACCRA
LAGOS
Niger Delta
Cape Formoso
Volta Delta
Cape Three Points
Cape Palmas
Assini Canyon
Trou sans Fond
LIBERIA PLATEAU
GRAND CESS RIDGE
ROMANCHE FRACTURE ZONE
IVORY COAST RIDGE
IVORY COAST GHANA RIDGE
GUINEA ABYSSAL PLAIN
JEAN CHARCOT RIDGE
Principe Channel
FERNANDO POO-ANNOBON RIDGE
I. do Principe
I. de São Tomé
I. Annobón
GUINEA RIDGE

Figure 49

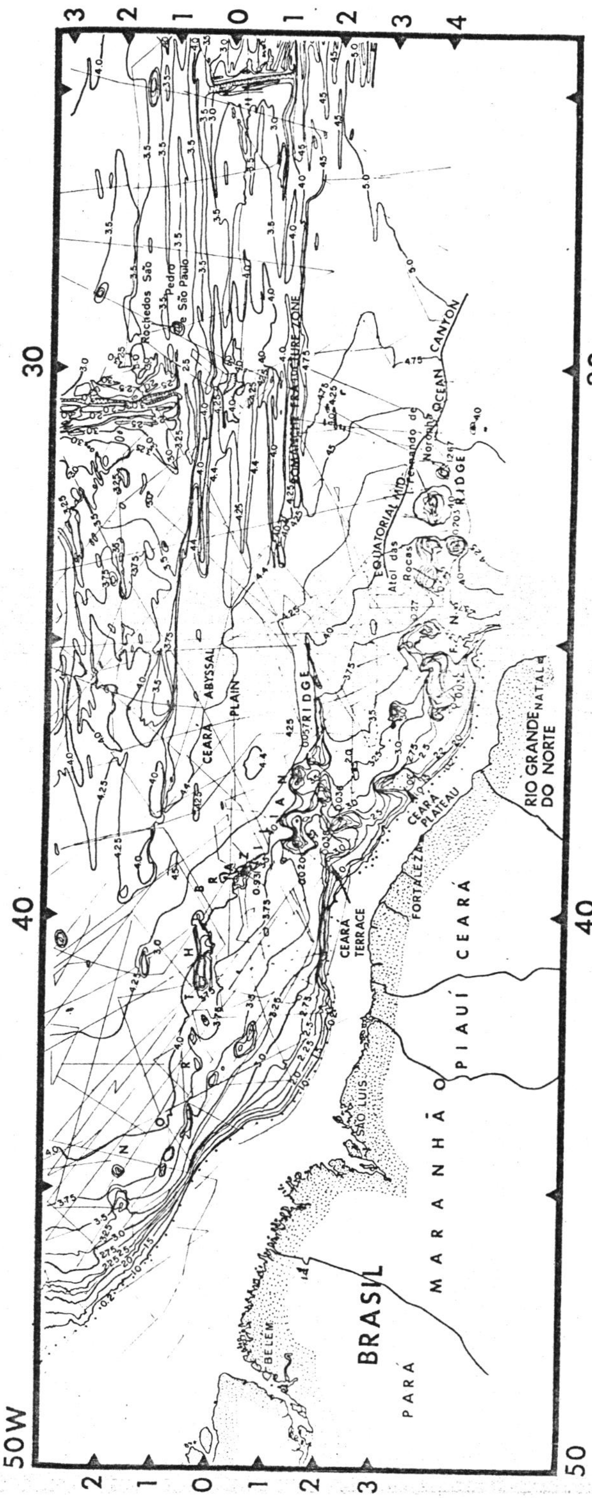

50W
30
40
50
BRASIL
PARÁ
MARANHÃO
PIAUÍ
CEARÁ
RIO GRANDE DO NORTE
NATAL
FORTALEZA
SÃO LUIS
BELEM
CEARÁ TERRACE
CEARÁ PLATEAU
CEARÁ ABYSSAL PLAIN
EQUATORIAL MID-OCEAN CANYON
ROMANCHE FRACTURE ZONE
Rochedos São Pedro e São Paulo
Atol das Rocas
Fernando de Noronha
RIDGE

Figure 50

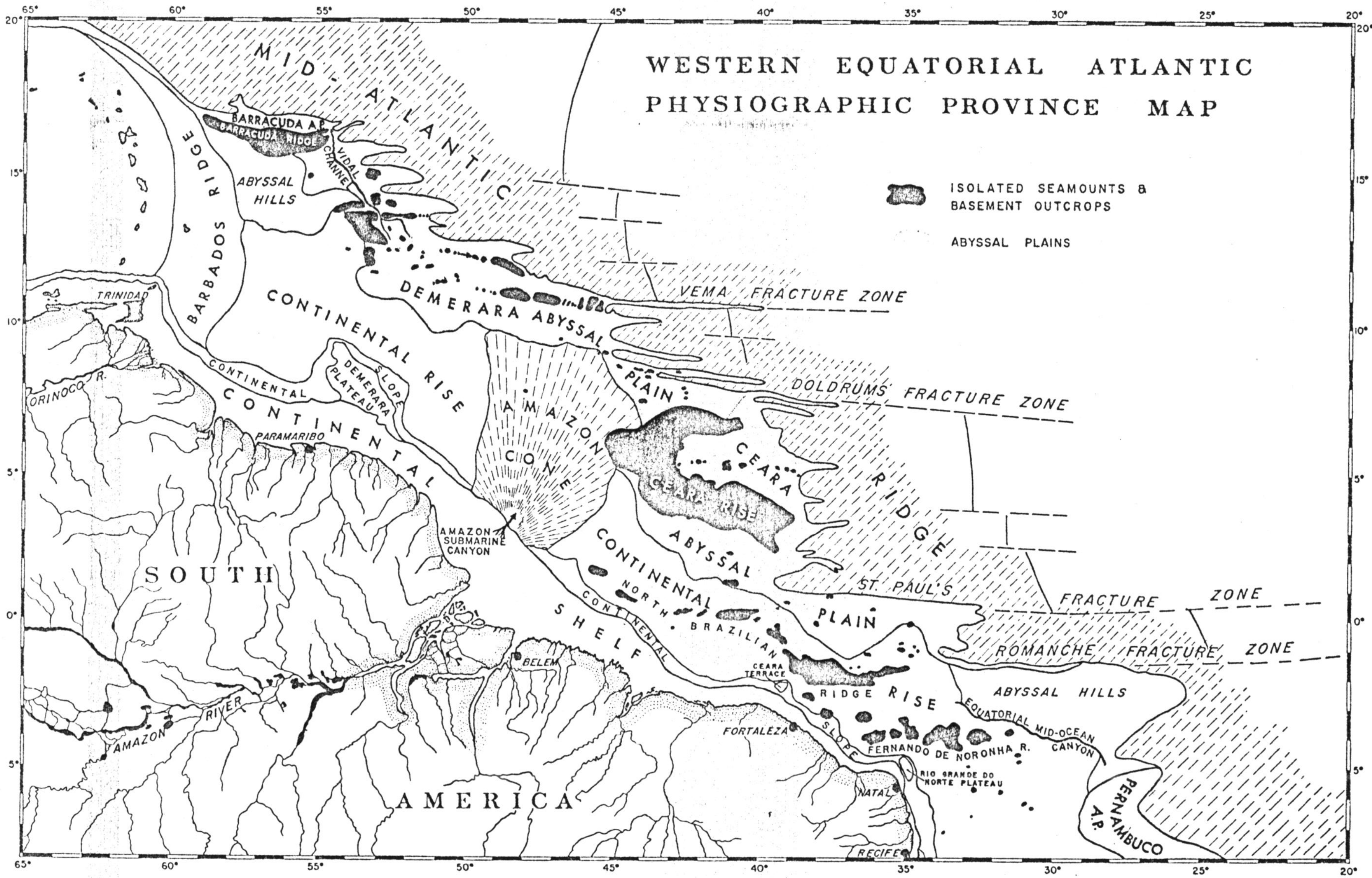

WESTERN EQUATORIAL ATLANTIC
PHYSIOGRAPHIC PROVINCE MAP
ISOLATED SEAMOUNTS & BASEMENT OUTCROPS
ABYSSAL PLAINS
MID-ATLANTIC
BARRACUDA A.P.
BARRACUDA RIDGE
VIDAL CHANNEL
ABYSSAL HILLS
BARBADOS RIDGE
TRINIDAD
ORINOCO R.
DEMERARA ABYSSAL PLAIN
CONTINENTAL RISE
CONTINENTAL SLOPE
DEMERARA PLATEAU
SLOPE
CONTINENTAL
PARAMARIBO
VEMA FRACTURE ZONE
DOLDRUMS FRACTURE ZONE
AMAZON CONE
AMAZON SUBMARINE CANYON
CEARA RIDGE
CEARA RISE
CEARA ABYSSAL PLAIN
ST. PAUL'S FRACTURE ZONE
ROMANCHE FRACTURE ZONE
CONTINENTAL
CONTINENTAL SHELF
NORTH BRAZILIAN RIDGE
CEARA TERRACE
RISE
SLOPE
ABYSSAL HILLS
EQUATORIAL MID-OCEAN CANYON
FERNANDO DE NORONHA R.
RIO GRANDE DO NORTE PLATEAU
PERNAMBUCO A.P.
SOUTH AMERICA
AMAZON RIVER
BELEM
FORTALEZA
NATAL
RECIFE
65° 60° 55° 50° 45° 40° 35° 30° 25° 20°
20° 15° 10° 5° 0° 5°

Figure 51

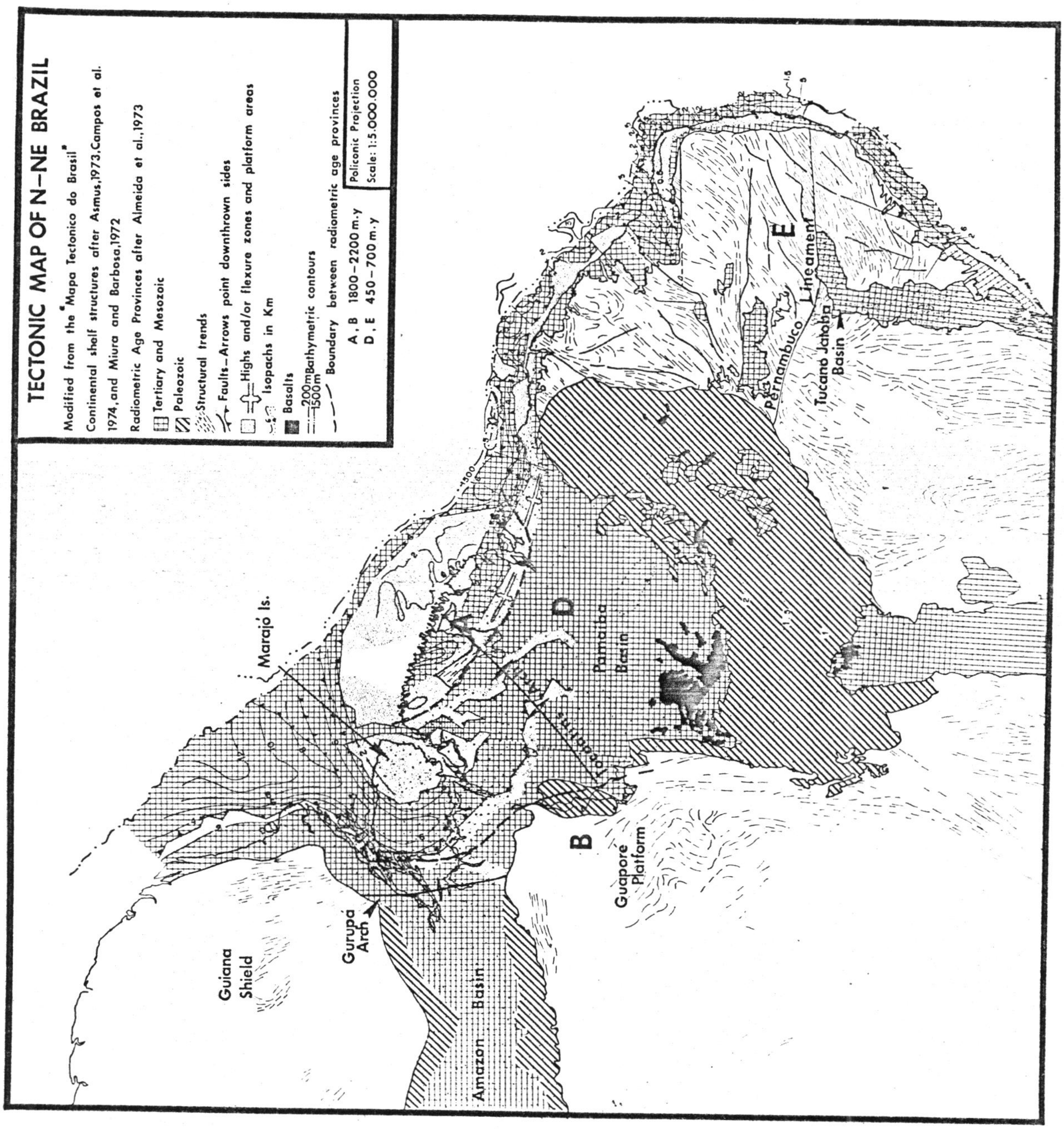

TECTONIC MAP OF N-NE BRAZIL
Modified from the "Mapa Tectonico do Brasil"
Continental shelf structures after Asmus,1973,Campos et al.
1974,and Miura and Barbosa,1972
Radiometric Age Provinces after Almeida et al.,1973
Tertiary and Mesozoic
Paleozoic
Structural trends
Faults_Arrows point downthrown sides
Highs and/or flexure zones and platform areas
Isopachs in Km
Basalts
200m 1500m Bathymetric contours
Boundary between radiometric age provinces
A, B 1800-2200 m.y
D, E 450-700 m.y
Policonic Projection
Scale: 1:5.000.000
Marajó Is.
Guiana Shield
Gurupá Arch
Amazon Basin
B
Guaporé Platform
D
Parnaíba Basin
Tocantins
E
Pernambuco Lineament
Tucano Jatobá Basin

Figure 52

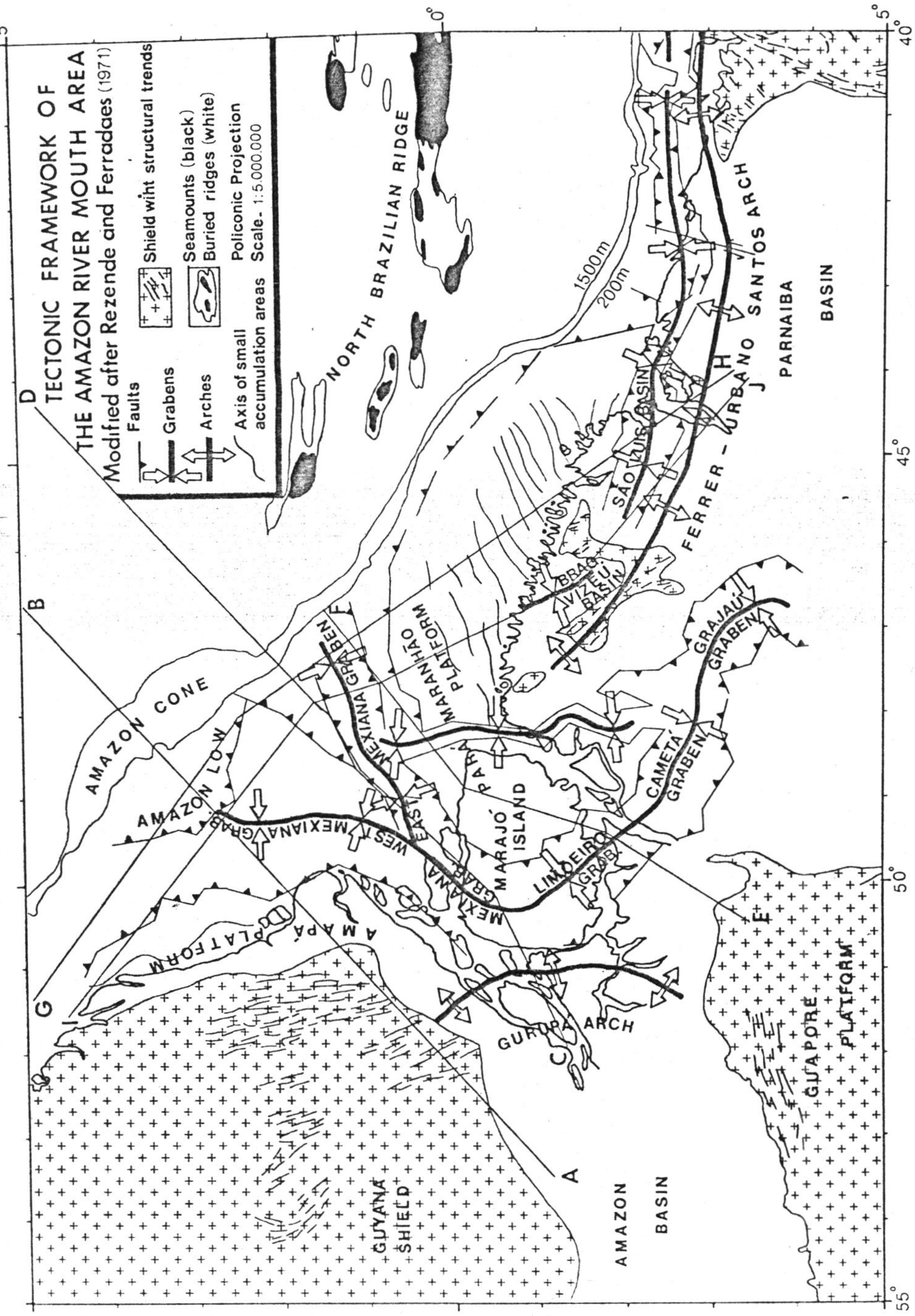

TECTONIC FRAMEWORK OF
THE AMAZON RIVER MOUTH AREA
Modified after Rezende and Ferradaes (1971)
Faults
Grabens
Arches
Axis of small accumulation areas
Shield wiht structural trends
Seamounts (black)
Buried ridges (white)
Policonic Projection
Scale.- 1:5.000.000
NORTH BRAZILIAN RIDGE
AMAZON CONE
AMAZON LOW
1500m
200m
MARANHÃO PLATFORM
MARAJÓ ISLAND
LIMOEIRO GRAB.
CAMETÁ GRABEN
GRAJAÚ GRABEN
SÃO LUIS BASIN
BRAG. VIZEU BASIN
FERRER - URBANO SANTOS ARCH
PARNAIBA BASIN
GUAPORÉ PLATFORM
GURUPÁ ARCH
AMAPÁ PLATFORM
GUYANA SHIELD
AMAZON BASIN
WEST MEXIANA GRAB.
EAST MEXIANA GRABEN
MEXIANA GRAB.
55°
50°
45°
40°
5°
0°

Figure 53

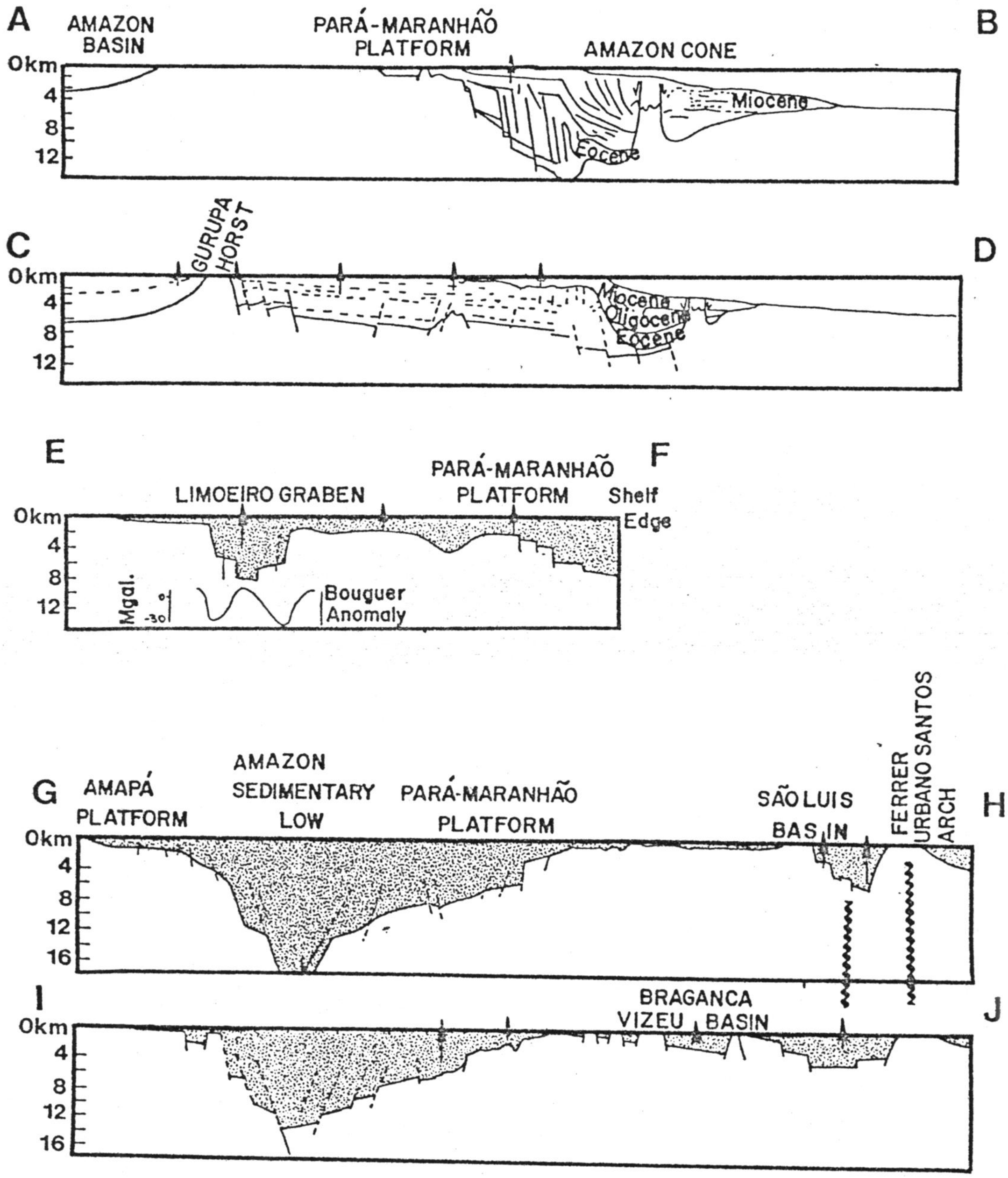

A
AMAZON BASIN
PARÁ-MARANHÃO PLATFORM
AMAZON CONE
B
0km
4
8
12
Miocene
Eocene
C
GURUPA HORST
D
Miocene
Oligocene
Eocene
E
LIMOEIRO GRABEN
PARÁ-MARANHÃO PLATFORM
F
Shelf Edge
Mgal.
0
-30
Bouguer Anomaly
G
AMAPÁ PLATFORM
AMAZON SEDIMENTARY LOW
PARÁ-MARANHÃO PLATFORM
SÃO LUIS BASIN
FERRER URBANO SANTOS ARCH
H
16
I
BRAGANCA VIZEU BASIN
J

Figure 54

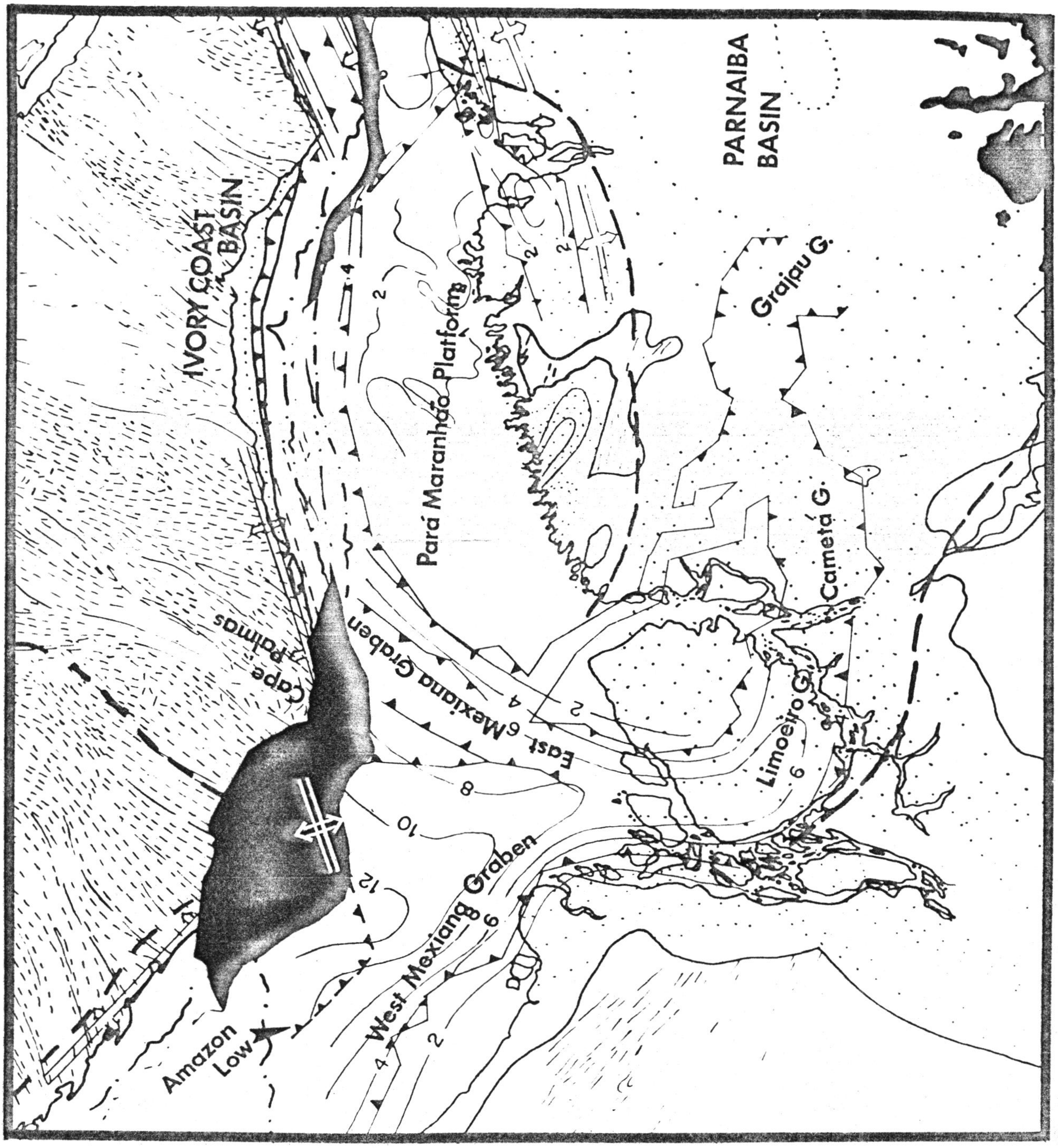

IVORY COAST BASIN
Cape Palmas
Amazon Low
West Mexiana Graben
East Mexiana Graben
Pará Maranhão Platform
PARNAIBA BASIN
Grajaú G.
Cametá G.
Limoeiro G.

Figure 55

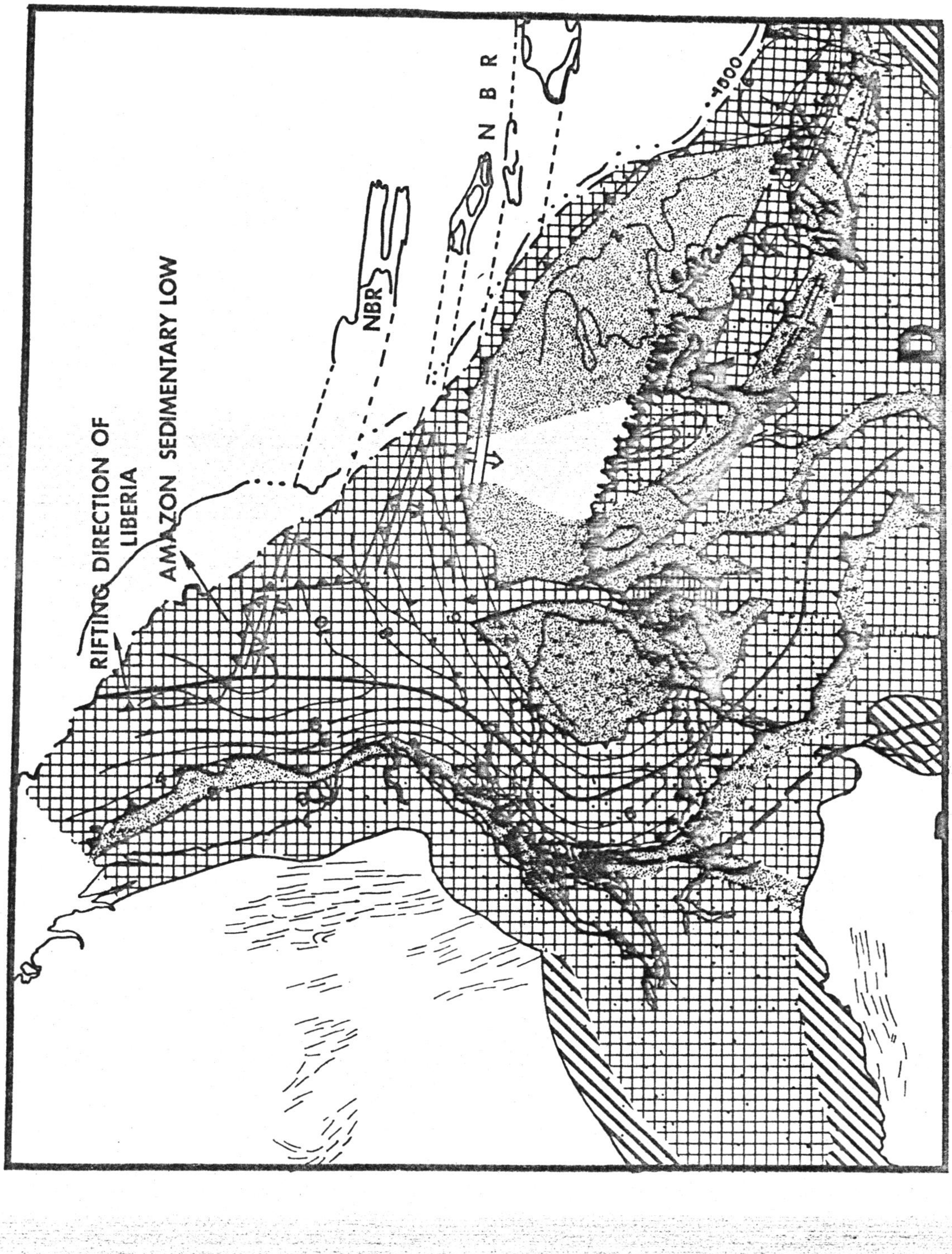

RIFTING DIRECTION OF
LIBERIA
AMAZON SEDIMENTARY LOW
NBR
N B R
D

Figure 56

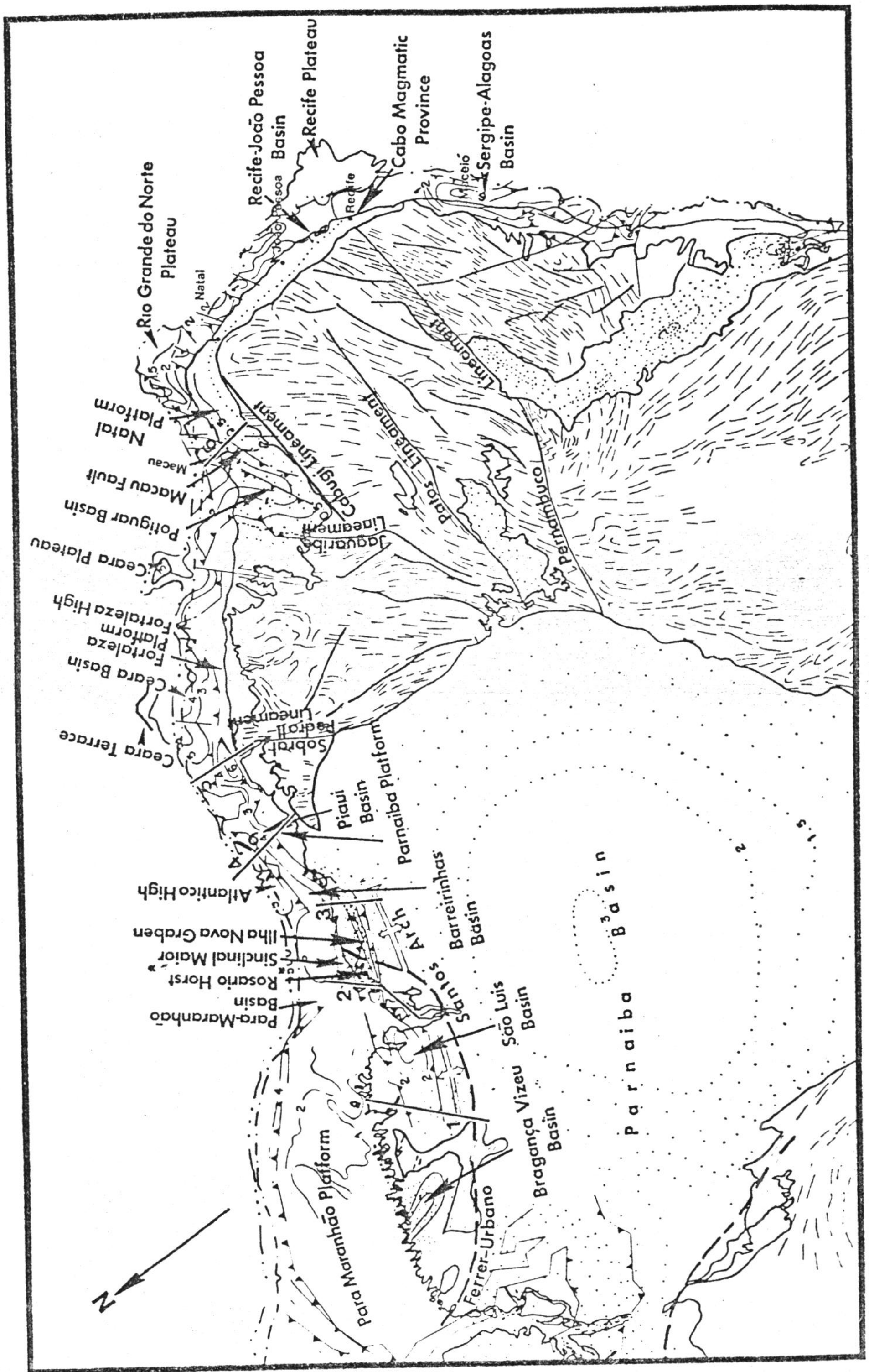

Rio Grande do Norte Plateau
Natal
Recife-João Pessoa Basin
João Pessoa
Recife Plateau
Recife
Cabo Magmatic Province
Maceió
Sergipe-Alagoas Basin
Natal Platform
Macau
Macau Fault
Potiguar Basin
Cabugi Lineament
Jaguaribe Lineament
Patos Lineament
Pernambuco Lineament
Ceara Plateau
Fortaleza High
Fortaleza Platform
Ceara Basin
Ceara Terrace
Sobral-Pedro II Lineament
Piaui Basin
Parnaiba Platform
Atlantico High
Ilha Nova Graben
Sinclinal Maior
Rosario Horst
Para-Maranhão Basin
Santos Arch
Barreirinhas Basin
São Luis Basin
Bragança Vizeu Basin
Ferrer-Urbano
Para Maranhão Platform
Parnaiba Basin
N

Figure 57

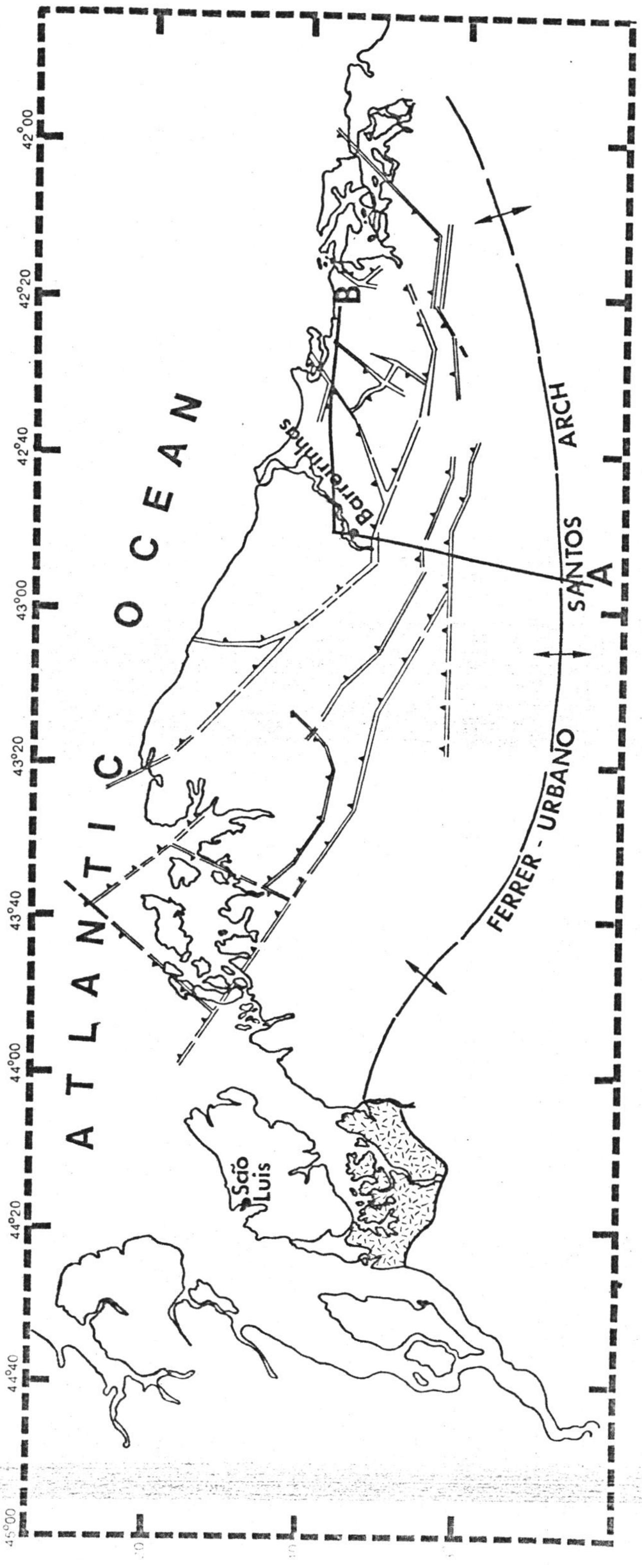

ATLANTIC OCEAN
São Luis
Barreirinhas
SANTOS
FERRER - URBANO ARCH
A
B
42°00
42°20
42°40
43°00
43°20
43°40
44°00
44°20
44°40

Figure 58

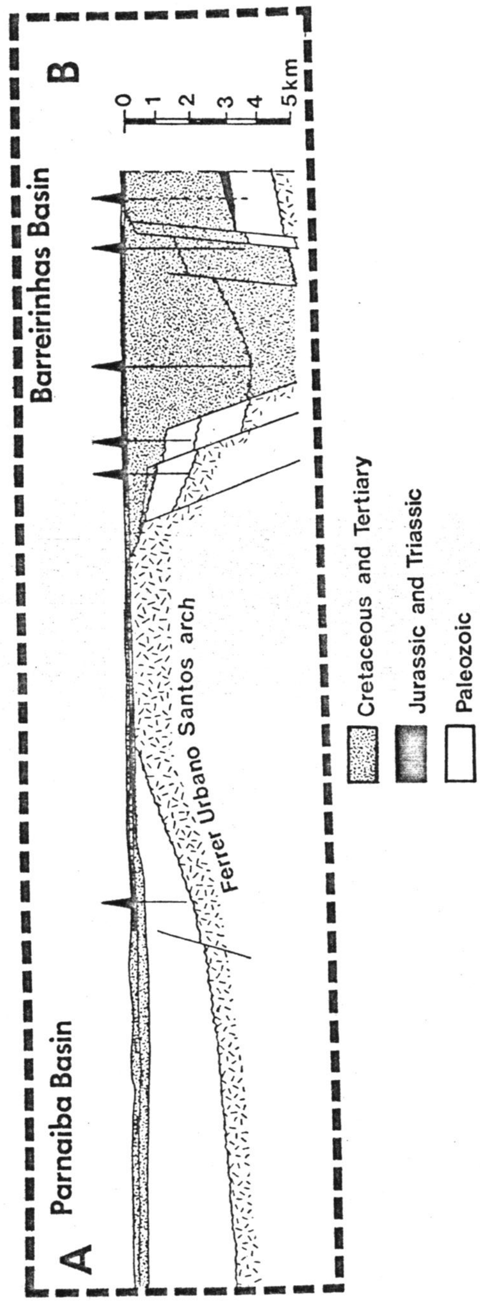

A
B
Parnaiba Basin
Barreirinhas Basin
Ferrer Urbano Santos arch
0
1
2
3
4
5km
Cretaceous and Tertiary
Jurassic and Triassic
Paleozoic

Figure 59

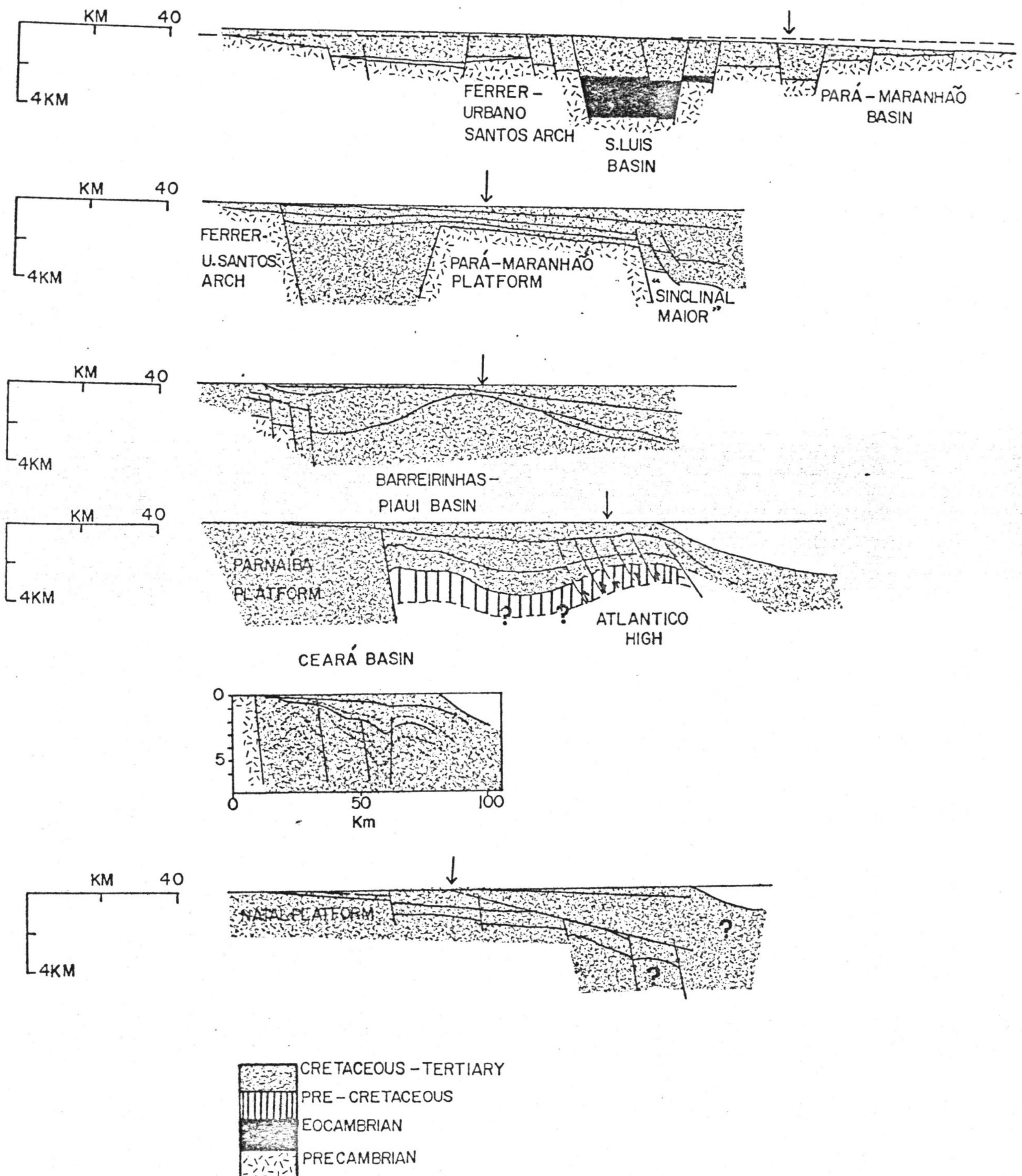

KM 40
4KM
FERRER-
URBANO
SANTOS ARCH
S.LUIS
BASIN
PARÁ-MARANHÃO
BASIN
KM 40
4KM
FERRER-
U.SANTOS
ARCH
PARÁ-MARANHÃO
PLATFORM
"SINCLINAL
MAIOR"
KM 40
4KM
BARREIRINHAS-
PIAUI BASIN
KM 40
4KM
PARNAIBA
PLATFORM
ATLANTICO
HIGH
CEARÁ BASIN
0
5
0
50
100
Km
KM 40
4KM
NATAL PLATFORM
CRETACEOUS-TERTIARY
PRE-CRETACEOUS
EOCAMBRIAN
PRECAMBRIAN

Figure 60

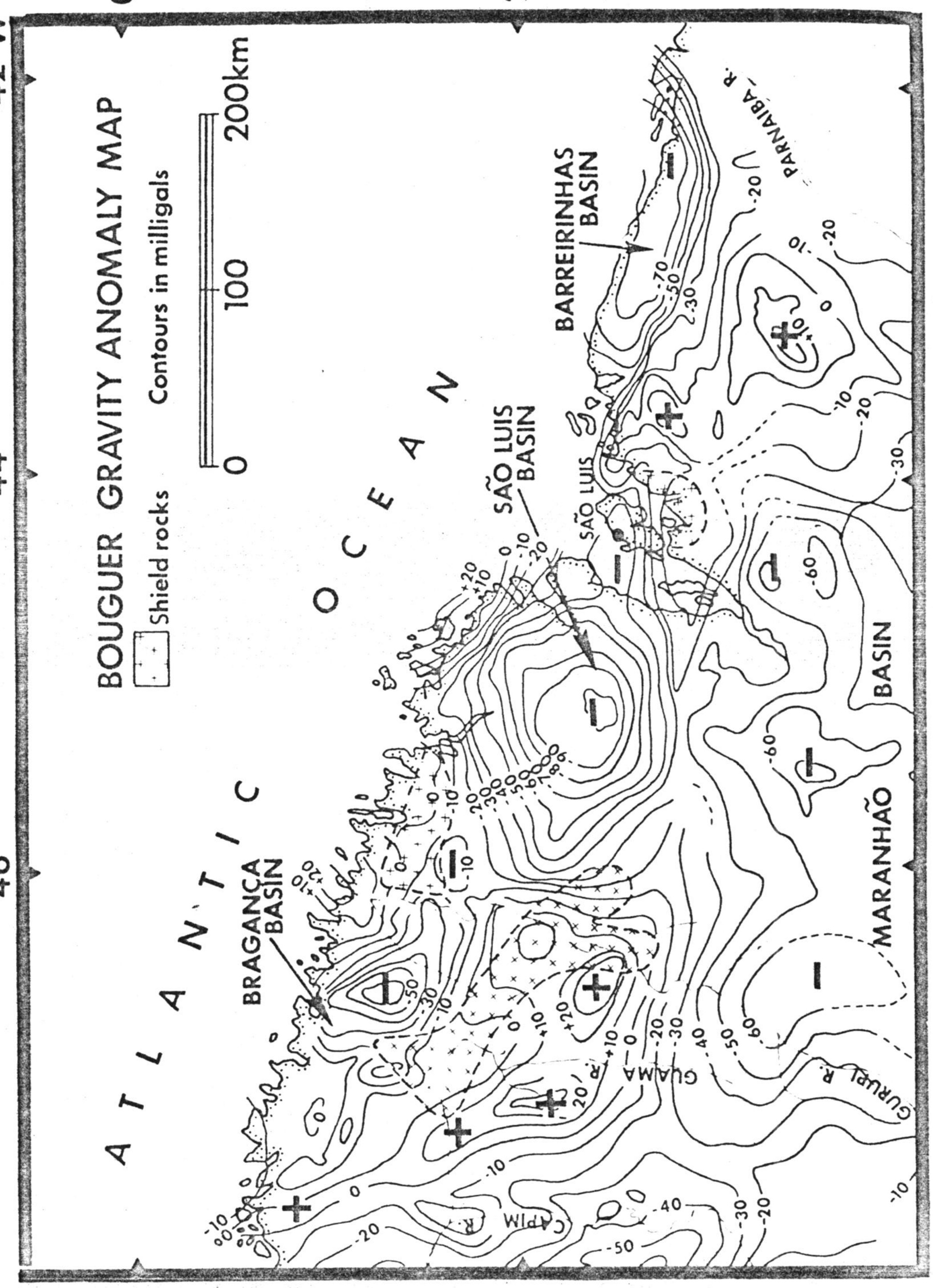

BOUGUER GRAVITY ANOMALY MAP
Shield rocks
Contours in milligals
0
100
200km
42°W
44°
46°
0°
2°S
ATLANTIC
OCEAN
BRAGANÇA BASIN
SÃO LUIS BASIN
BARREIRINHAS BASIN
SÃO LUIS
MARANHÃO
BASIN
GUAMA R.
GURUPI R.
CAPIM R.
PARNAIBA R.

Figure 61

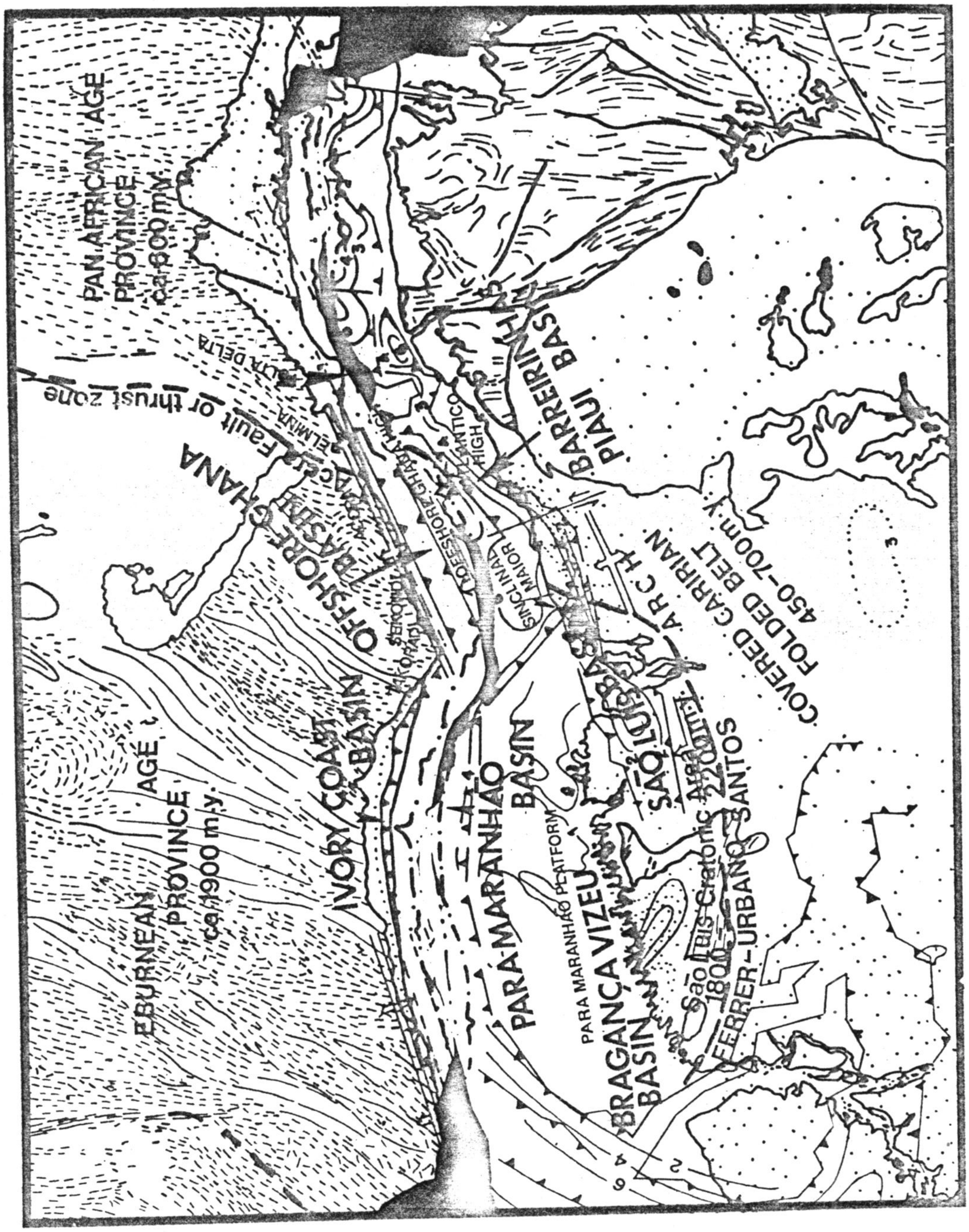

PAN AFRICAN AGE PROVINCE ca 600 m.y.
EBURNEAN AGE PROVINCE ca 1900 m.y.
Fault or thrust zone
IVORY COAST BASIN
OFFSHORE GHANA BASIN
PARA-MARANHAO BASIN
PARA MARANHAO PLATFORM
BRAGANÇA VIZEU BASIN
SÃO LUIS
FERRER-URBANO SANTOS
ARCH
PIAUI
COVERED CARIRIAN FOLDED BELT 450-700 m.y.

Figure 62

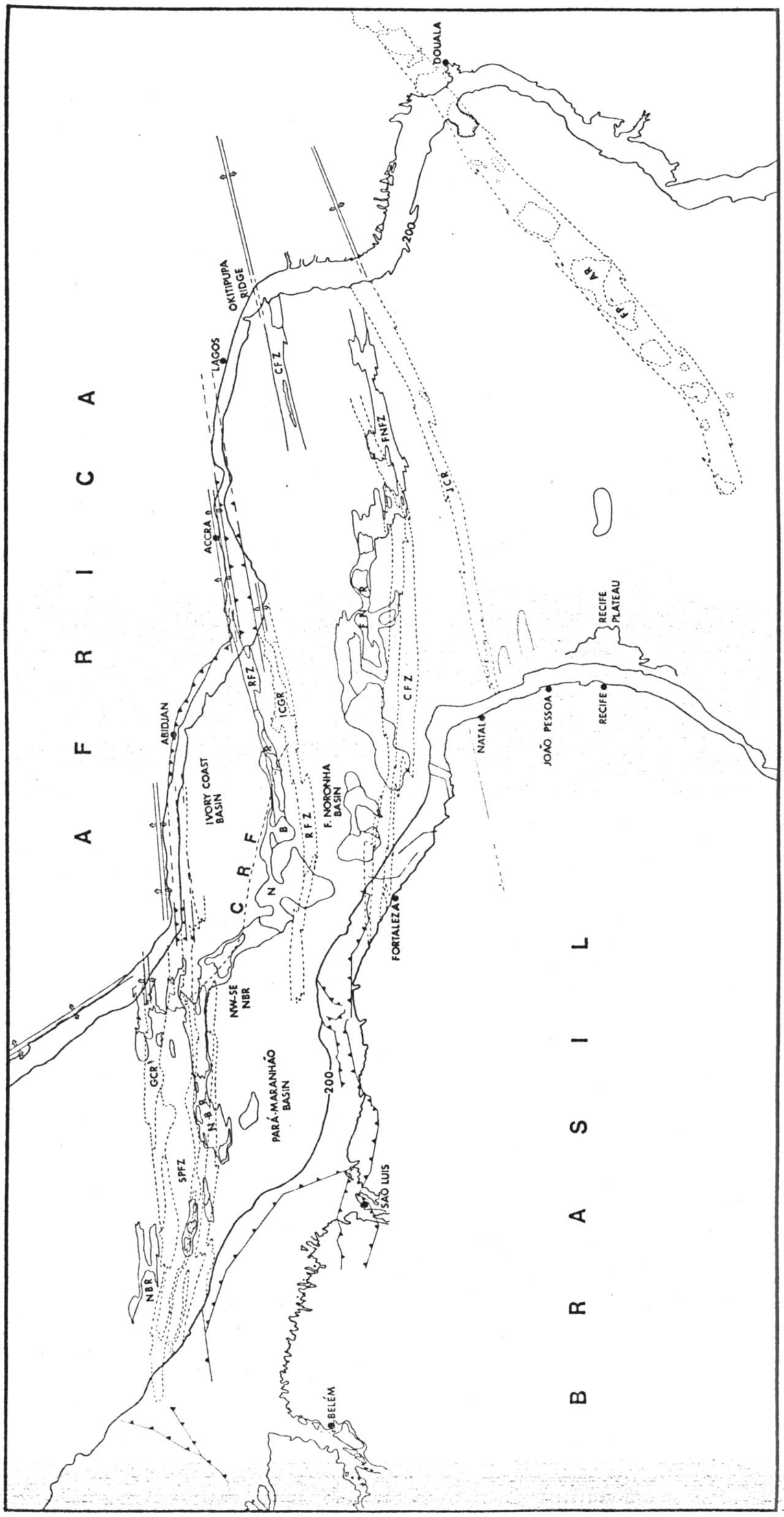
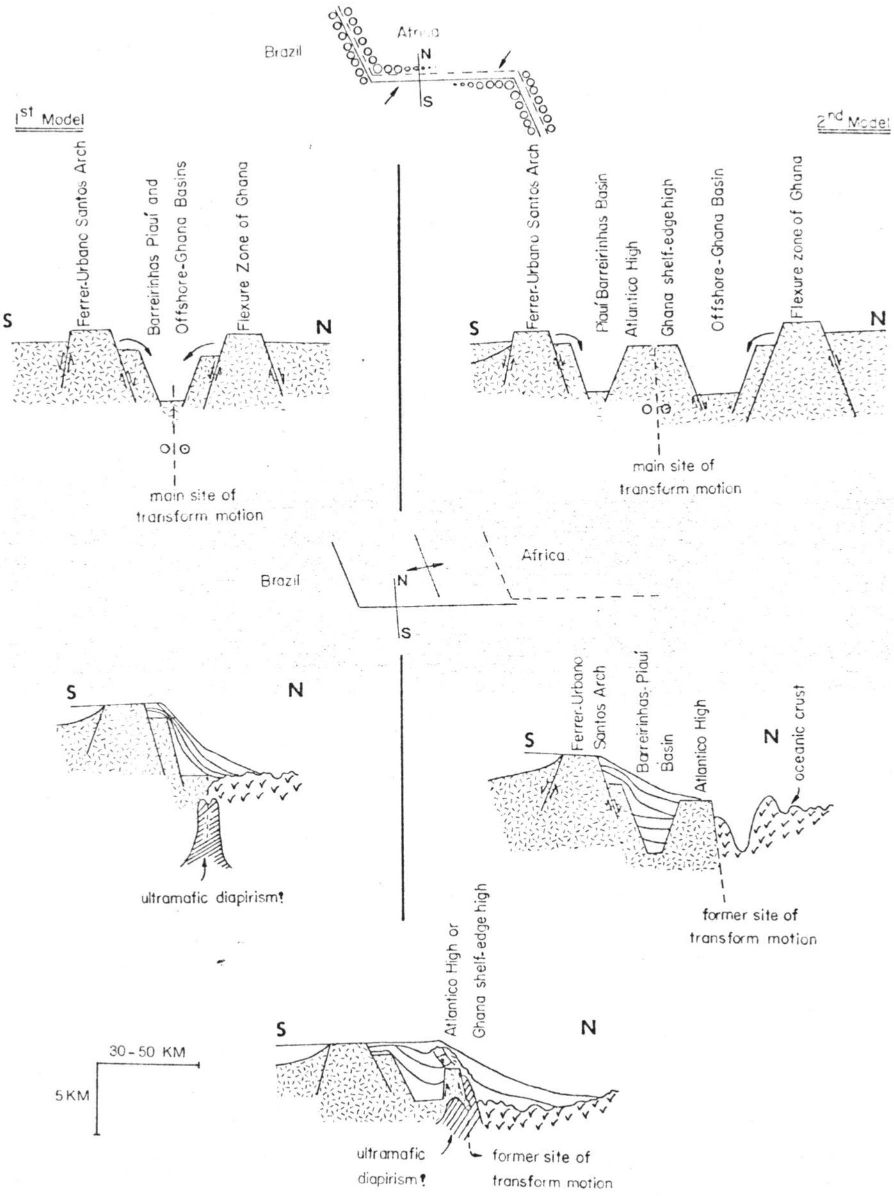

AFRICA
BRASIL
DOUALA
OKITIPUPA RIDGE
LAGOS
ACCRA
ABIDJAN
IVORY COAST BASIN
F. NORONHA BASIN
PARÁ-MARANHÃO BASIN
NW-SE NBR
CRF
RFZ
ICGR
CFZ
FNFZ
ICR
GCR
SPFZ
NBR
200
NATAL
JOÃO PESSOA
RECIFE
RECIFE PLATEAU
FORTALEZA
SÃO LUIS
BELÉM

Figure 63



Brazil
Africa
N
S
1st Model
2nd Model
Ferrer-Urbano Santos Arch
Barreirinhas Piauí and
Offshore-Ghana Basins
Flexure Zone of Ghana
main site of
transform motion
Piauí Barreirinhas Basin
Atlantico High
Ghana shelf-edge high
Offshore-Ghana Basin
Flexure zone of Ghana
Africa
Brazil
ultramafic diapirism?
Ferrer-Urbano
Santos Arch
Barreirinhas-Piauí
Basin
Atlantico High
oceanic crust
former site of
transform motion
Atlantico High or
Ghana shelf-edge high
30-50 KM
5 KM
ultramafic
diapirism?
former site of
transform motion

Figure 64

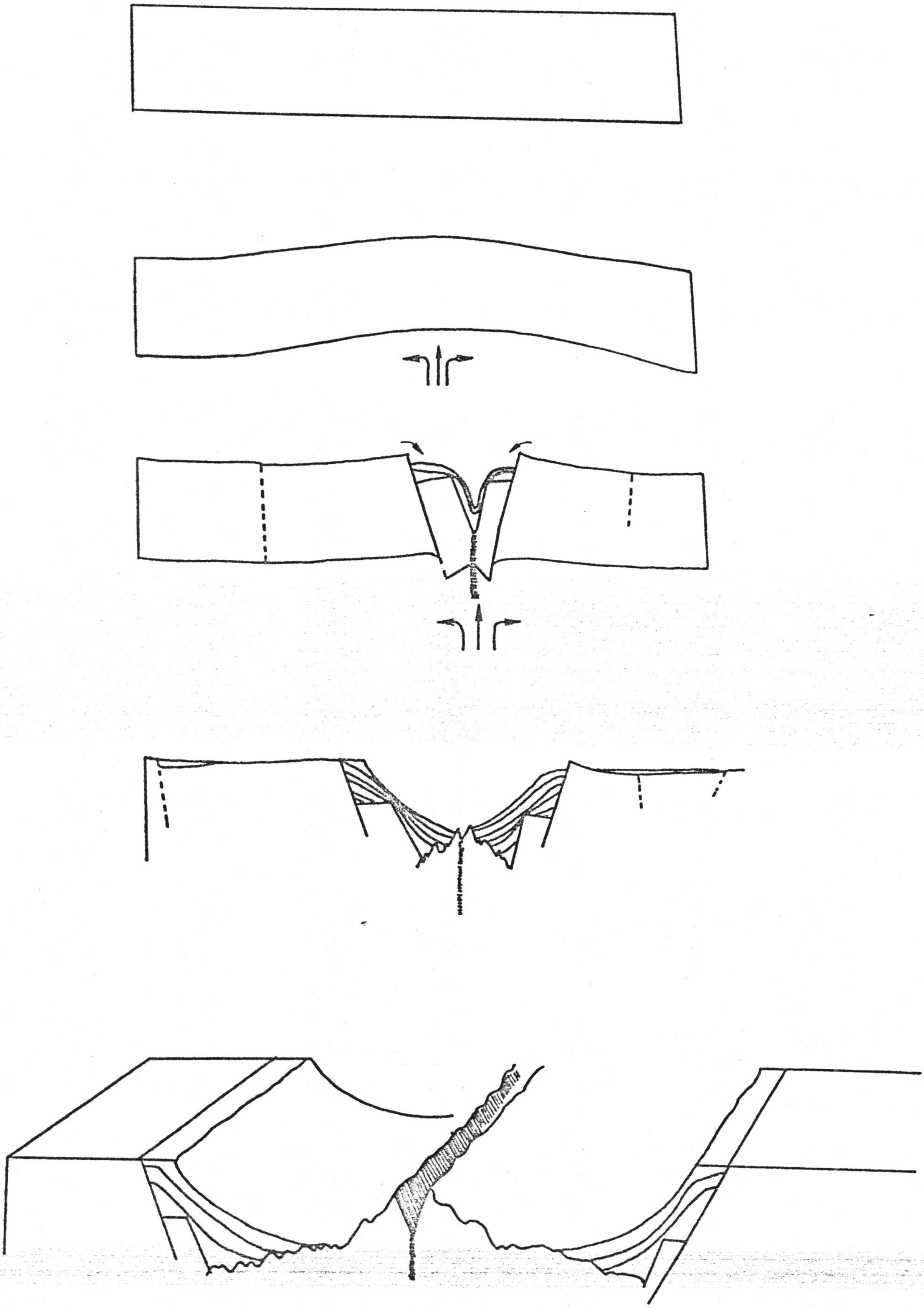


Figure 65

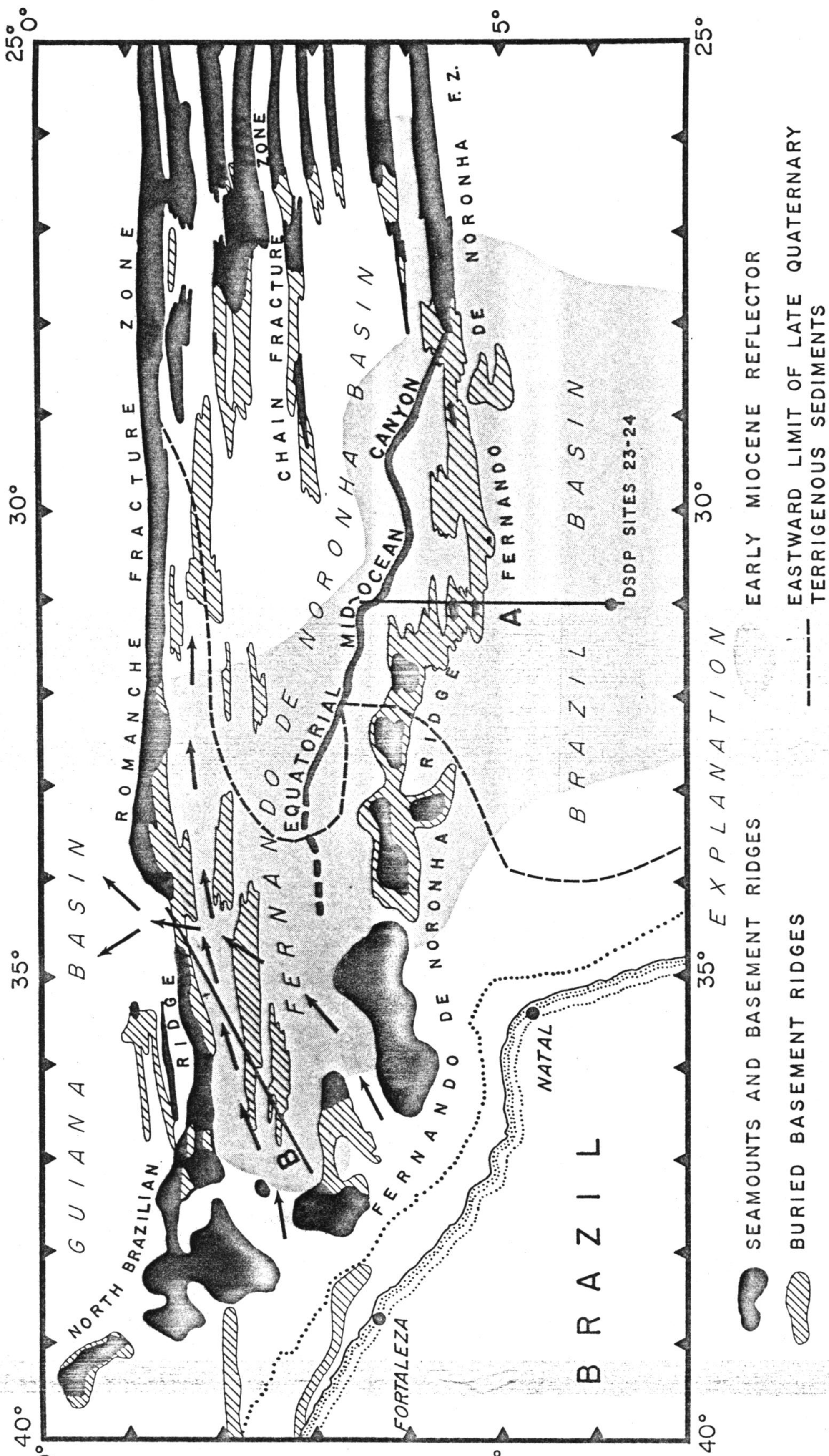

GUIANA BASIN
NORTH BRAZILIAN RIDGE
ROMANCHE FRACTURE ZONE
CHAIN FRACTURE ZONE
FERNANDO DE NORONHA BASIN
EQUATORIAL MID-OCEAN CANYON
FERNANDO DE NORONHA RIDGE
FERNANDO DE NORONHA F. Z.
BRAZIL BASIN
DSDP SITES 23-24
A
B
BRAZIL
FORTALEZA
NATAL
25°
30°
35°
40°
0°
5°
EXPLANATION
SEAMOUNTS AND BASEMENT RIDGES
BURIED BASEMENT RIDGES
EARLY MIOCENE REFLECTOR
EASTWARD LIMIT OF LATE QUATERNARY TERRIGENOUS SEDIMENTS

Figure 66

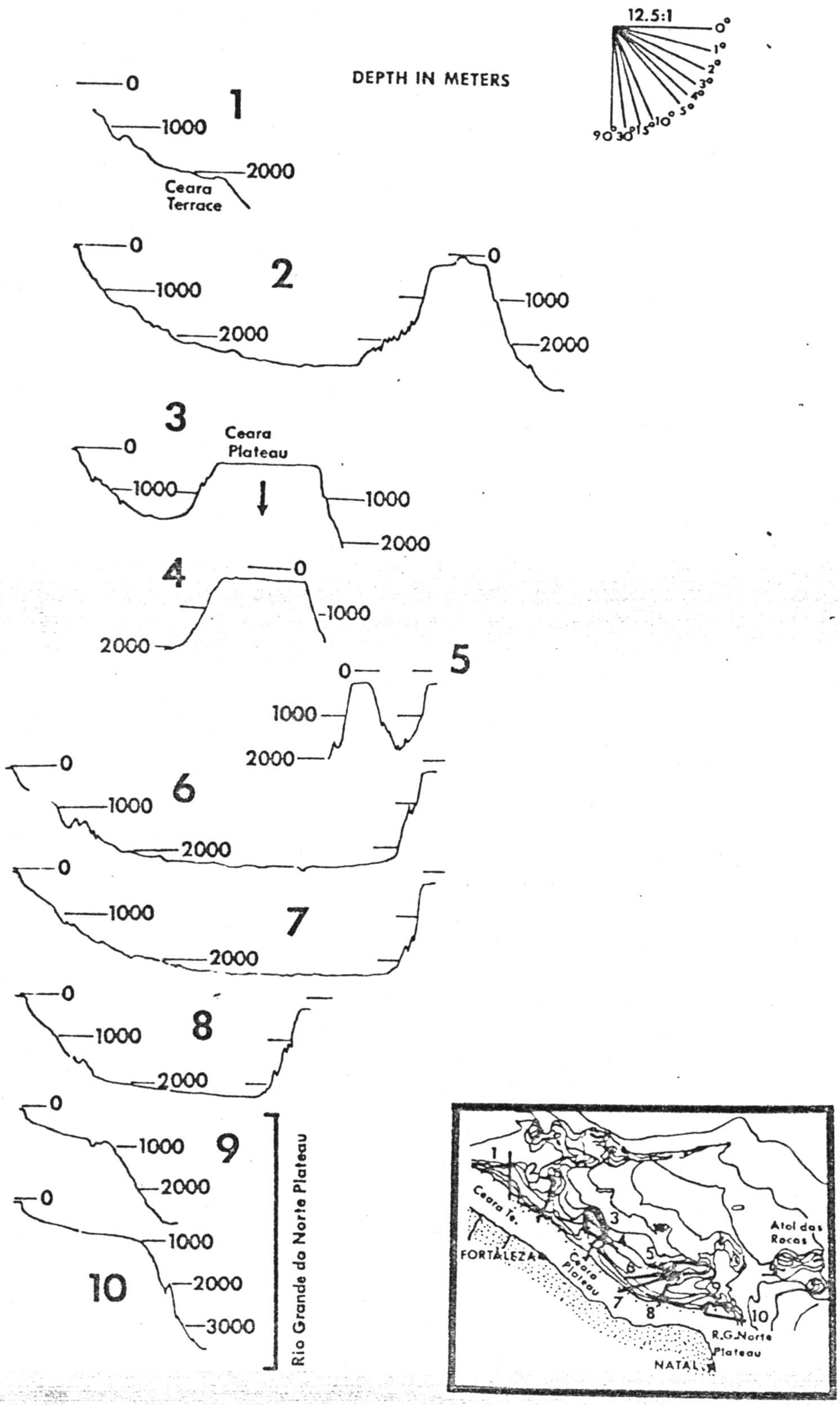
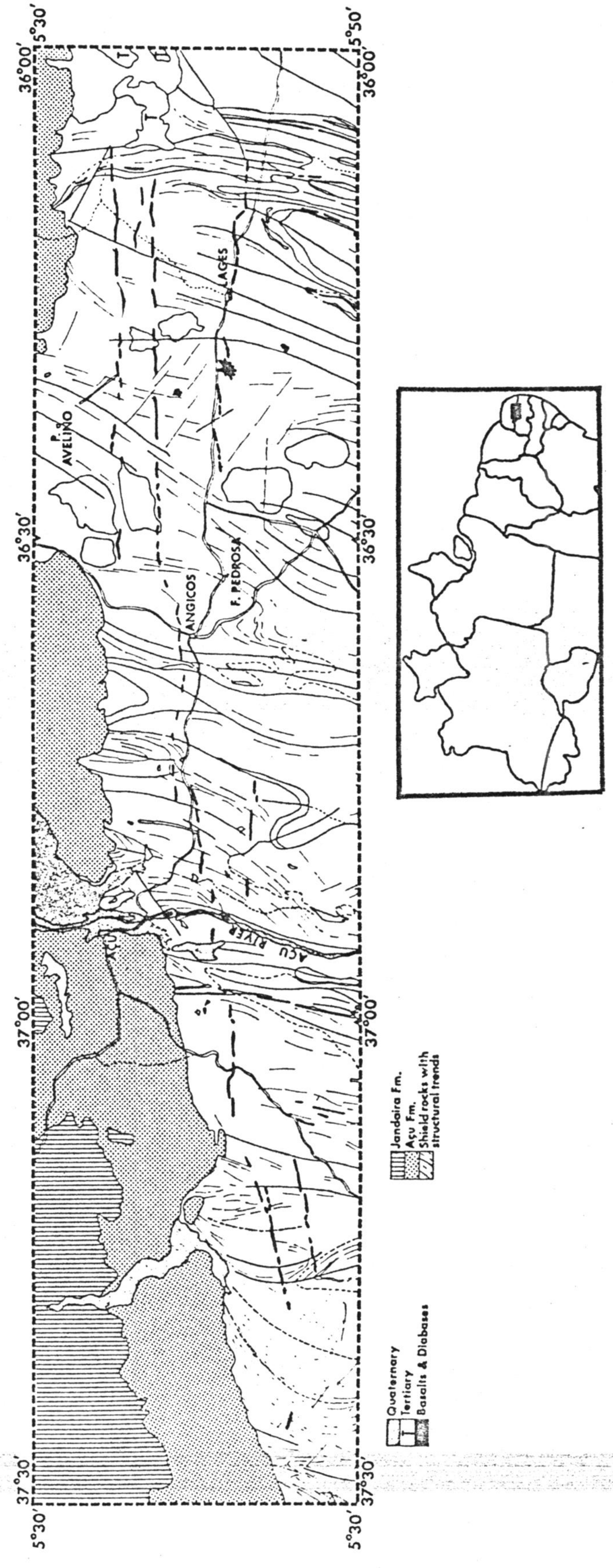

DEPTH IN METERS
12.5:1
Ceara Terrace
Ceara Plateau
Rio Grande do Norte Plateau
FORTALEZA
Ceara Te.
Ceara Plateau
Atol das Rocas
R.G.Norte Plateau
NATAL

Figure 67



5°30'
36°00'
5°50'
36°00'
36°30'
37°00'
37°30'
5°30'
P. AVELINO
ANGICOS
LAGES
F. PEDROSA
AÇU RIVER
Quaternary
Tertiary
Basalts & Diabases
Jandaira Fm.
Açu Fm.
Shield rocks with structural trends

Figure 68

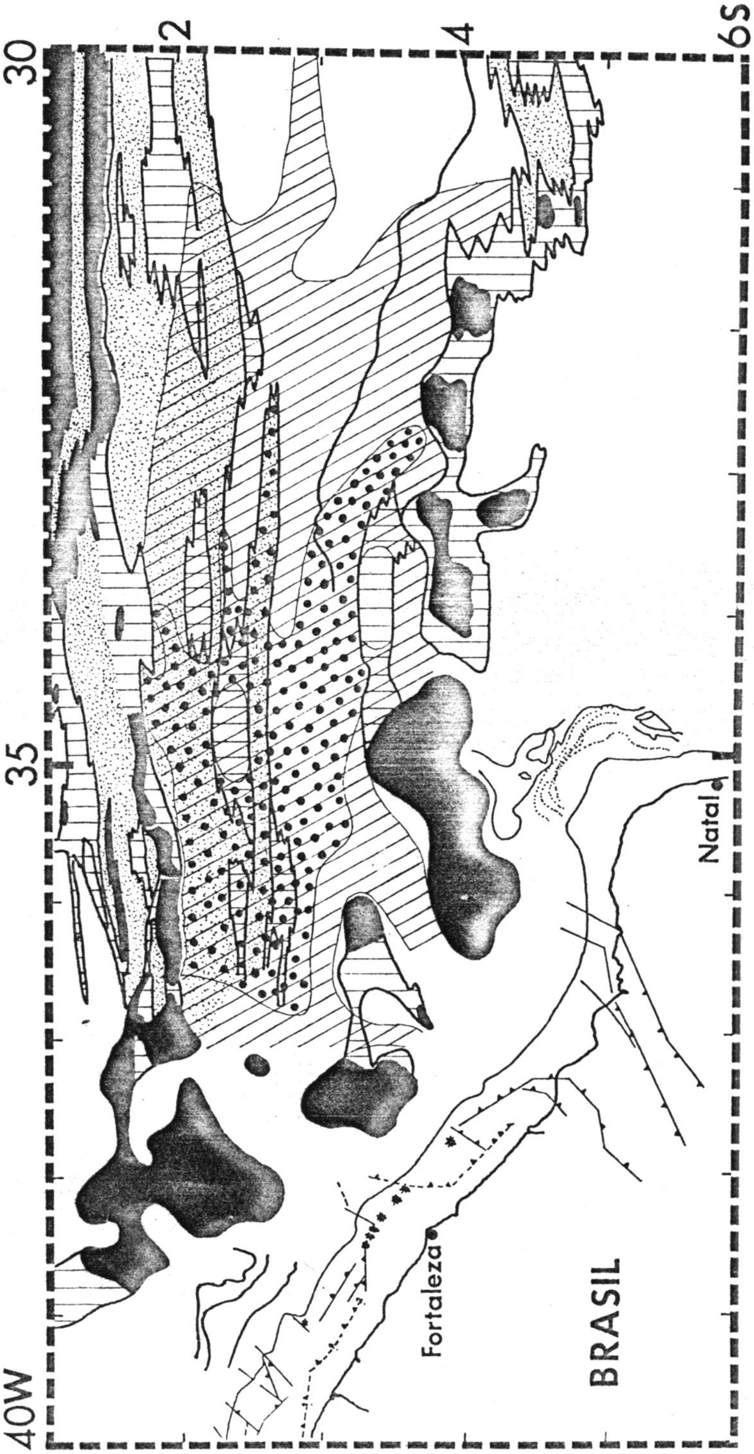


Figure 69

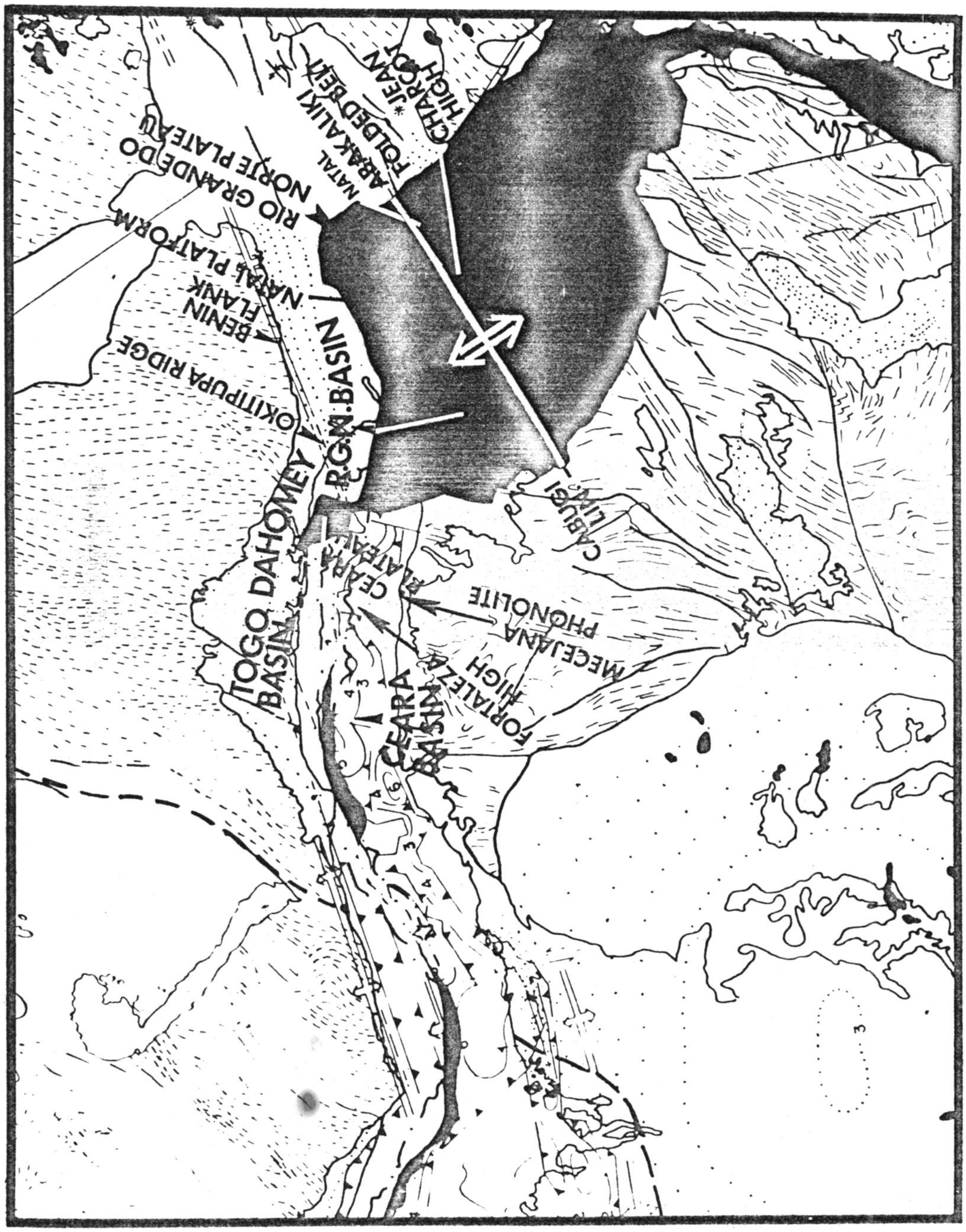

TOGO DAHOMEY BASIN
R.G.N. BASIN
OKITIPUPA RIDGE
BENIN FLANK
NATAL PLATFORM
RIO GRANDE DO NORTE PLATEAU
NATAL
ABAKALIKI FOLDED BELT
JEAN CHARCOT HIGH
CEARA PLATEAU
CEARA BASIN
MECEJANA PHONOLITE
FORTALEZA HIGH

Figure 70

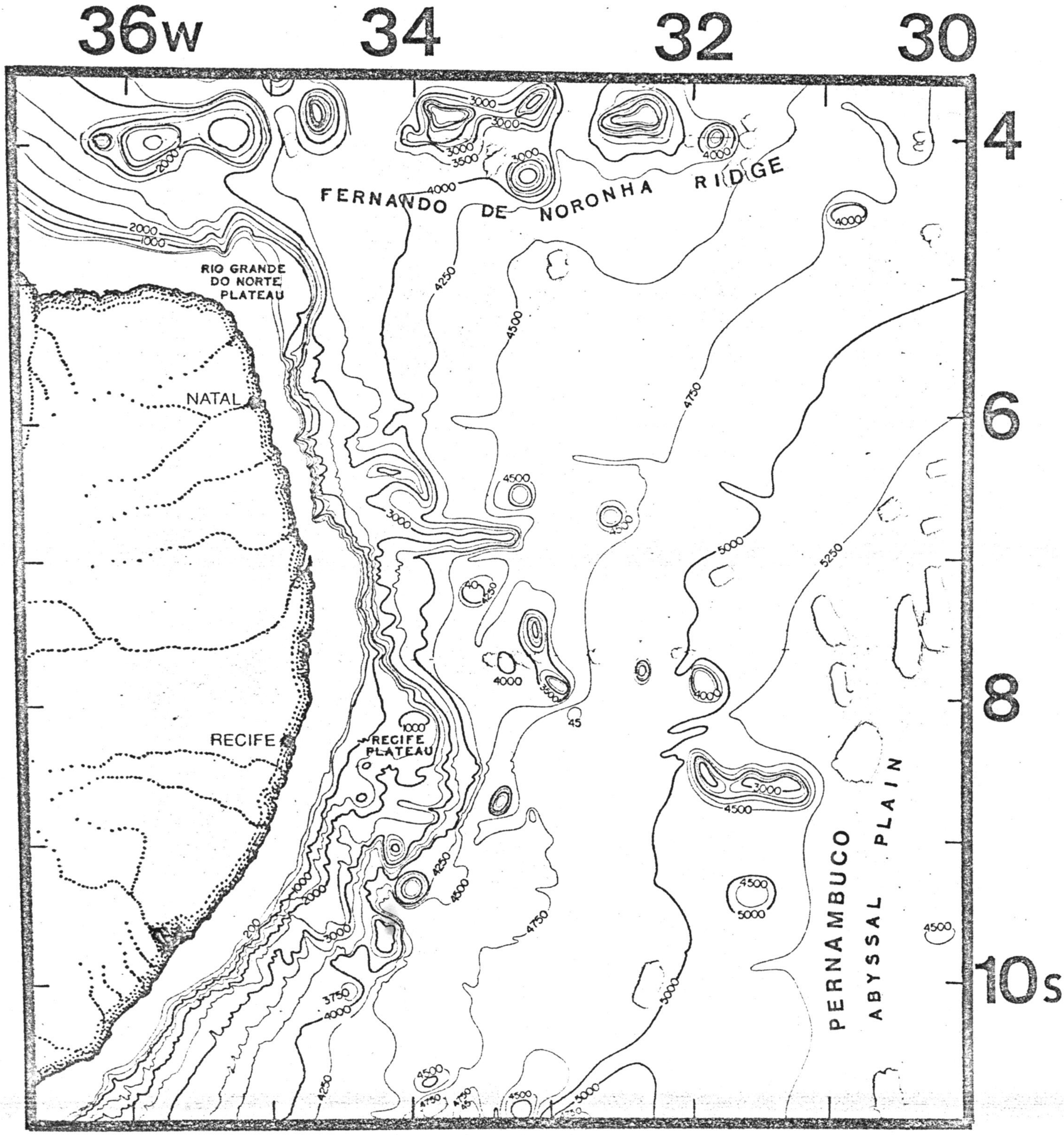

36w
34
32
30
4
6
8
10s
FERNANDO DE NORONHA RIDGE
RIO GRANDE DO NORTE PLATEAU
NATAL
RECIFE
RECIFE PLATEAU
PERNAMBUCO ABYSSAL PLAIN

Figure 71

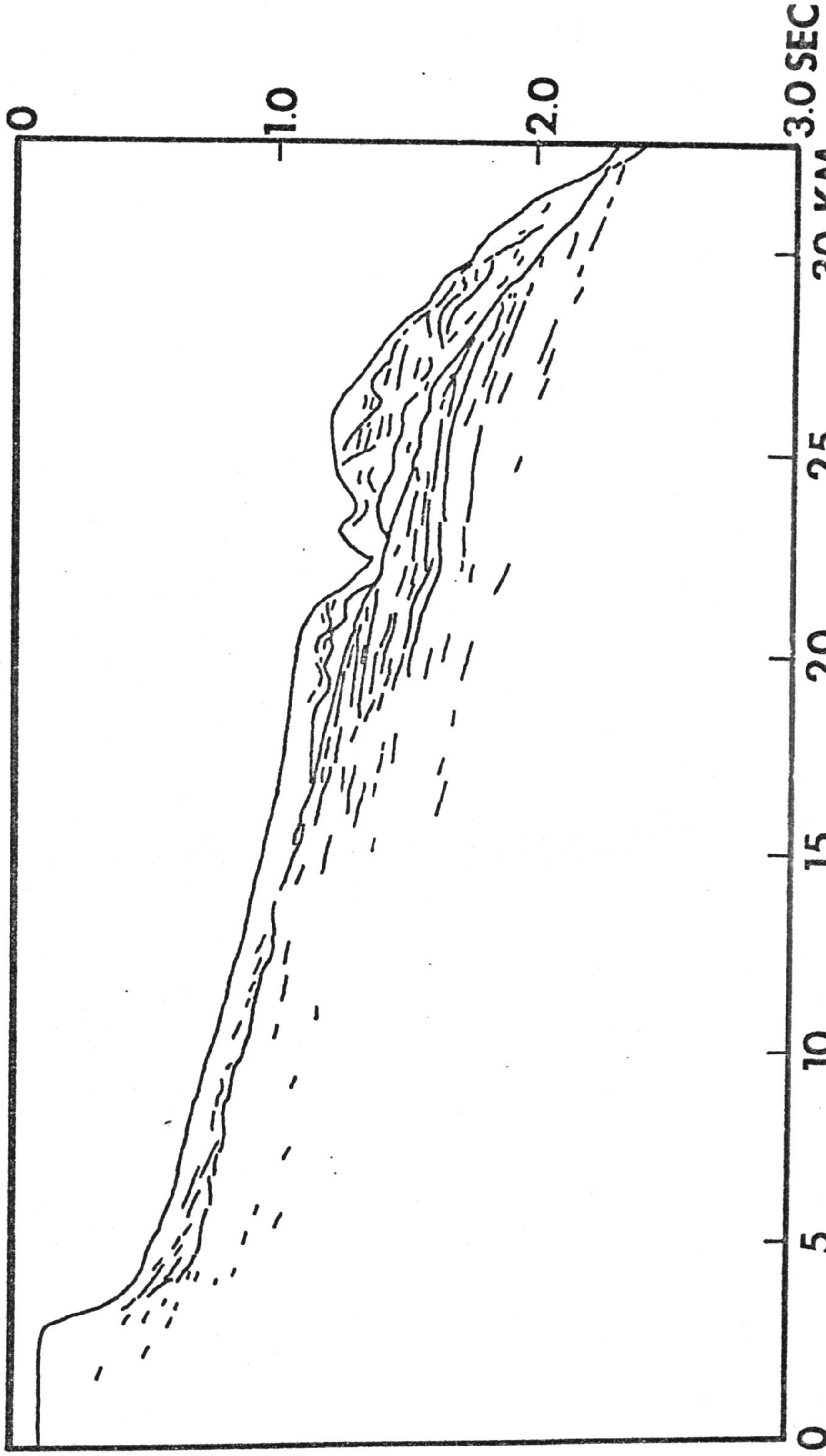
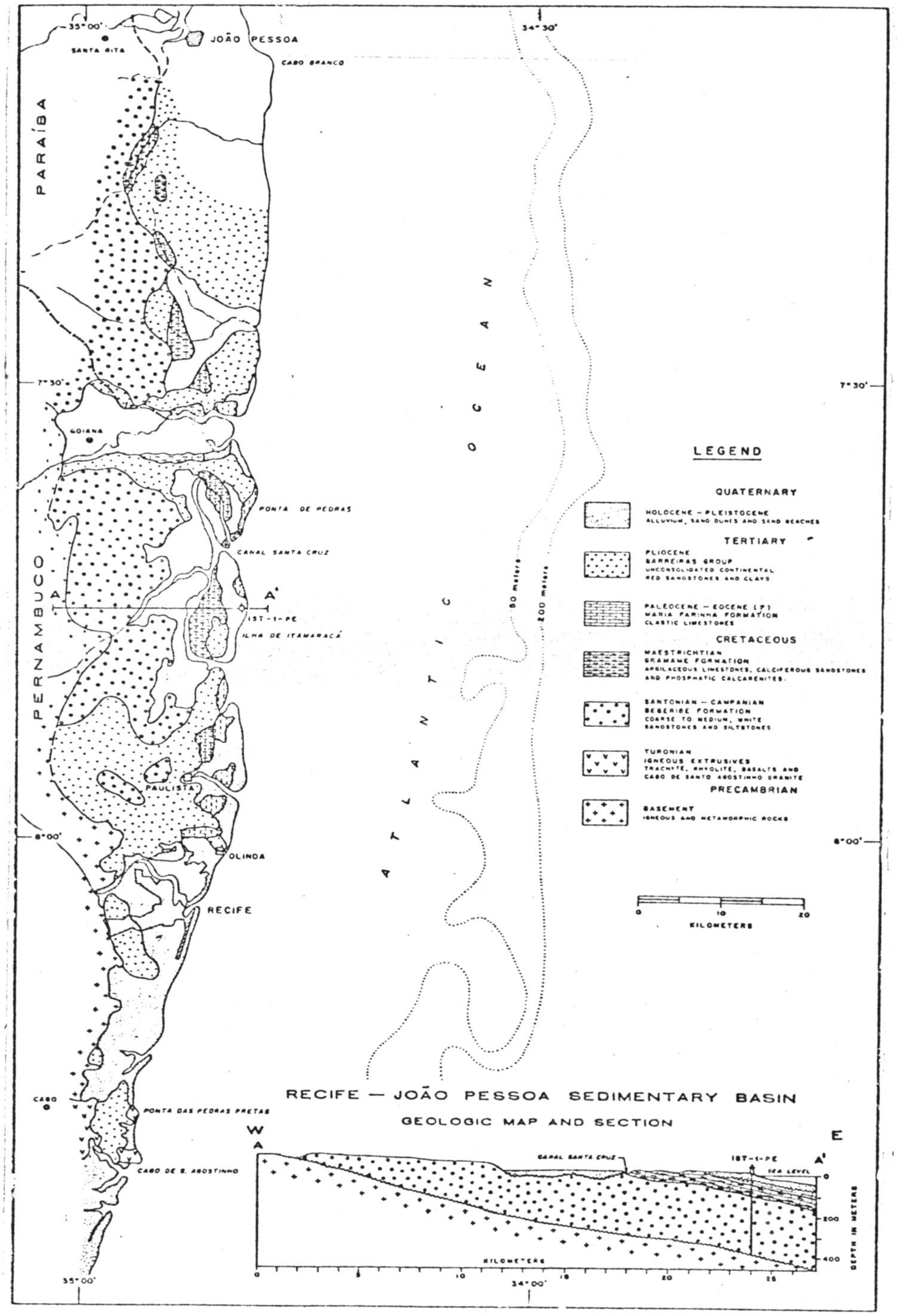


Figure 72



RECIFE — JOÃO PESSOA SEDIMENTARY BASIN
GEOLOGIC MAP AND SECTION
LEGEND
QUATERNARY
HOLOCENE – PLEISTOCENE
ALLUVIUM, SAND DUNES AND SAND BEACHES
TERTIARY
PLIOCENE
BARREIRAS GROUP
UNCONSOLIDATED CONTINENTAL
RED SANDSTONES AND CLAYS
PALEOCENE – EOCENE (?)
MARIA FARINHA FORMATION
CLASTIC LIMESTONES
CRETACEOUS
MAESTRICHTIAN
GRAMAME FORMATION
ARGILACEOUS LIMESTONES, CALCIFEROUS SANDSTONES
AND PHOSPHATIC CALCARENITES.
SANTONIAN – CAMPANIAN
BEBERIBE FORMATION
COARSE TO MEDIUM, WHITE
SANDSTONES AND SILTSTONES
TURONIAN
IGNEOUS EXTRUSIVES
TRACHYTE, RHYOLITE, BASALTS AND
CABO DE SANTO AGOSTINHO GRANITE
PRECAMBRIAN
BASEMENT
IGNEOUS AND METAMORPHIC ROCKS
KILOMETERS
ATLANTIC OCEAN
50 meters
200 meters
PARAÍBA
PERNAMBUCO
JOÃO PESSOA
SANTA RITA
CABO BRANCO
GOIANA
PONTA DE PEDRAS
CANAL SANTA CRUZ
1ST-1-PE
ILHA DE ITAMARACÁ
PAULISTA
OLINDA
RECIFE
CABO
PONTA DAS PEDRAS PRETAS
CABO DE S. AGOSTINHO
35°00'
34°30'
34°00'
7°30'
8°00'
A
A'
W
E
SEA LEVEL
DEPTH IN METERS

Figure 73

Geological Map of the Cabo of Santo Agostinho area, Pernambuco

by Rubem Queiroz Cobra, 1960

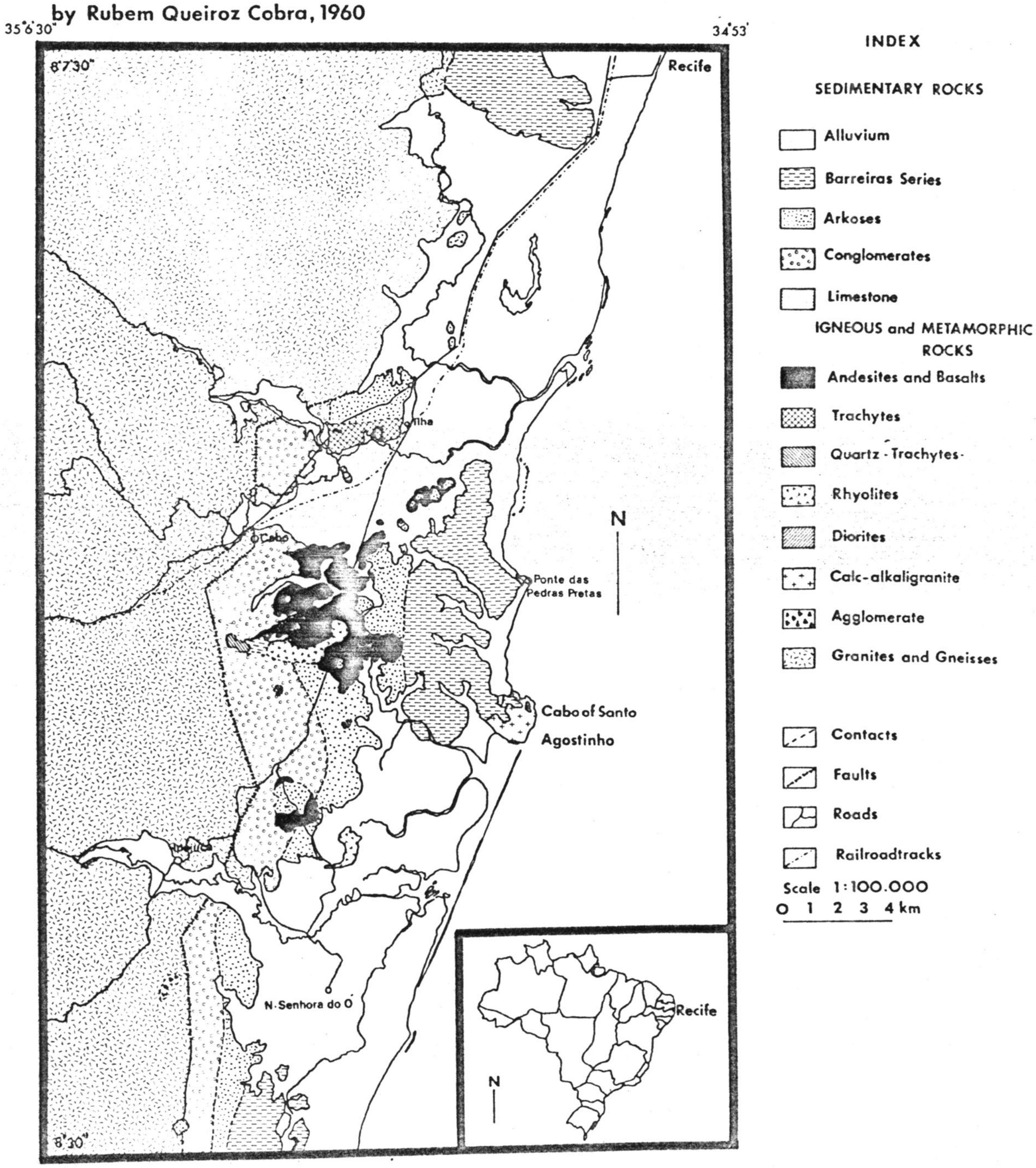
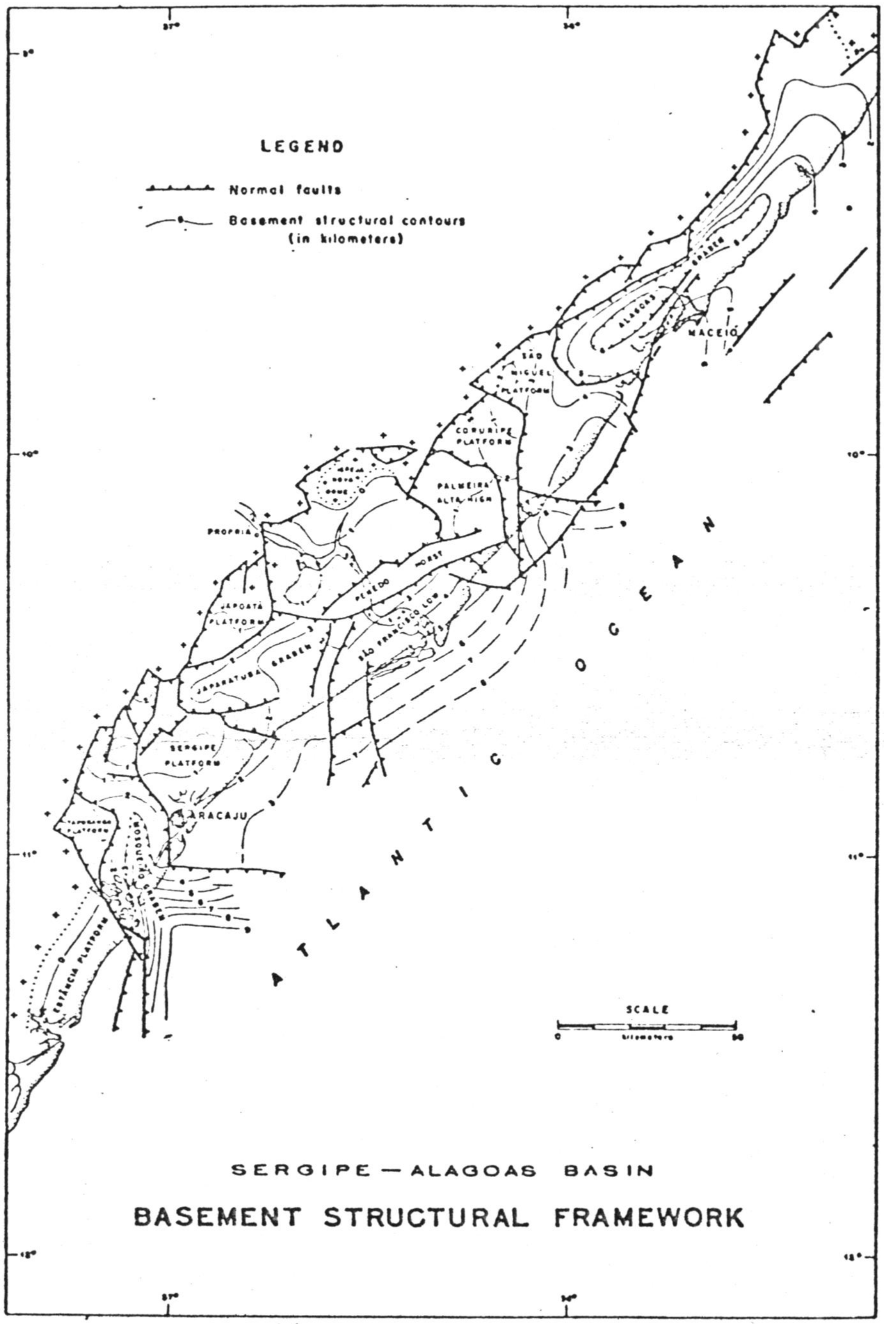


Figure 74


LEGEND
Normal faults
Basement structural contours
(in kilometers)
MACEIÓ
SÃO MIGUEL PLATFORM
CORURIPE PLATFORM
PALMEIRA ALTA HIGH
PROPRIÁ
PENEDO HORST
JAPOATÃ PLATFORM
SÃO FRANCISCO LOW
SERGIPE PLATFORM
ARACAJU
ESTÂNCIA PLATFORM
ATLANTIC OCEAN
SCALE
SERGIPE — ALAGOAS BASIN
BASEMENT STRUCTURAL FRAMEWORK

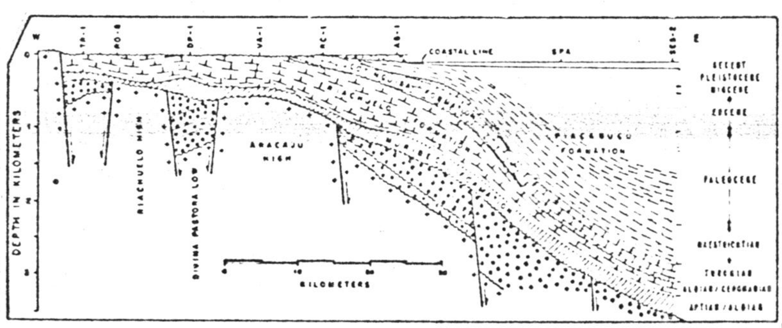

W
E
COASTAL LINE
DEPTH IN KILOMETERS
RIACHUELO HIGH
DIVINA PASTORA LOW
ARACAJU HIGH
PIAÇABUÇU FORMATION
KILOMETERS
EOCENE
PALEOCENE

Figure 75

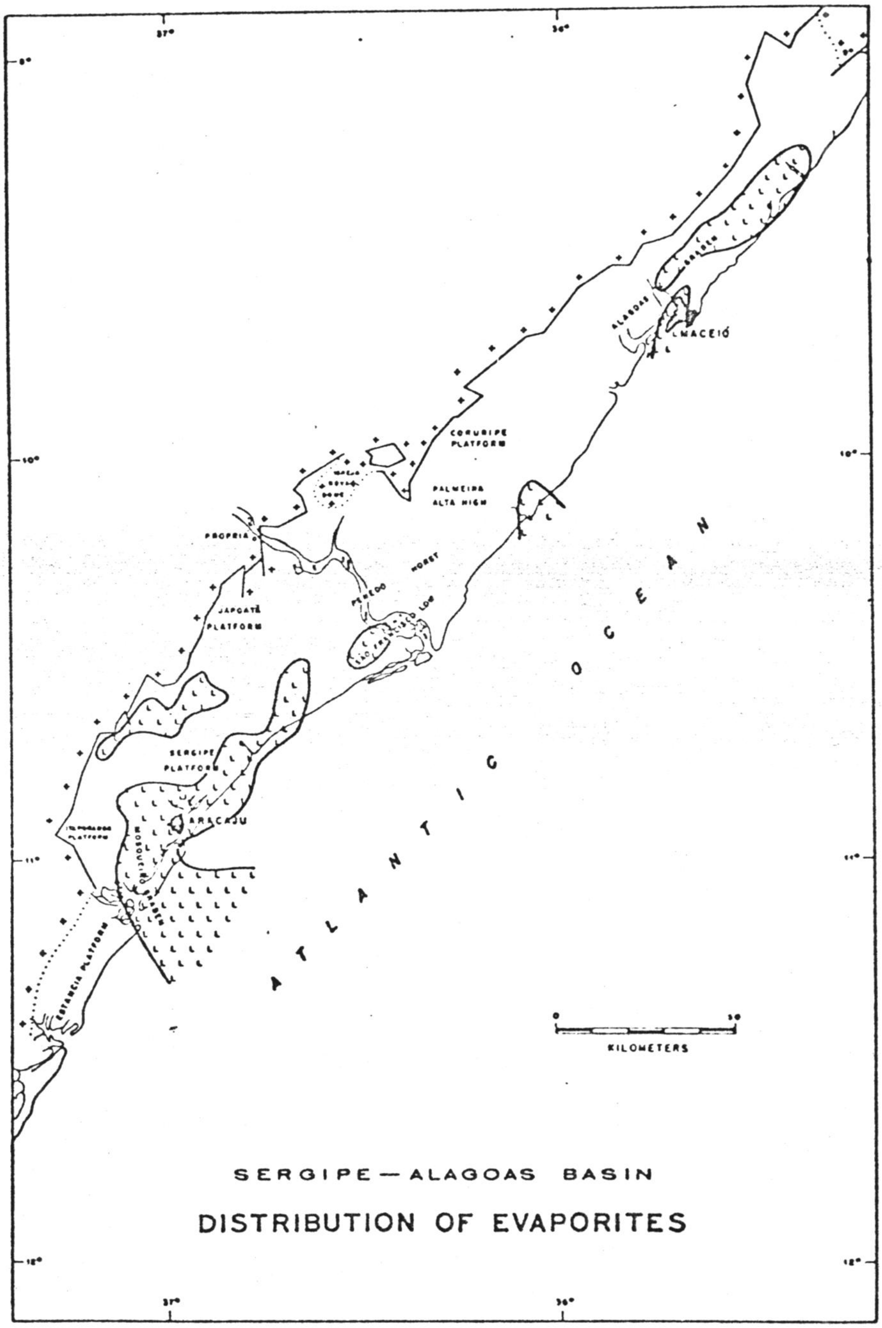

37°
36°
9°
10°
11°
12°
CORURIPE PLATFORM
PALMEIRA ALTA HIGH
PROPRIA
JAPOATÃ PLATFORM
PENEDO HORST
SERGIPE PLATFORM
ARACAJU
MACEIÓ
ATLANTIC OCEAN
0
50
KILOMETERS
SERGIPE — ALAGOAS BASIN
DISTRIBUTION OF EVAPORITES

Figure 76

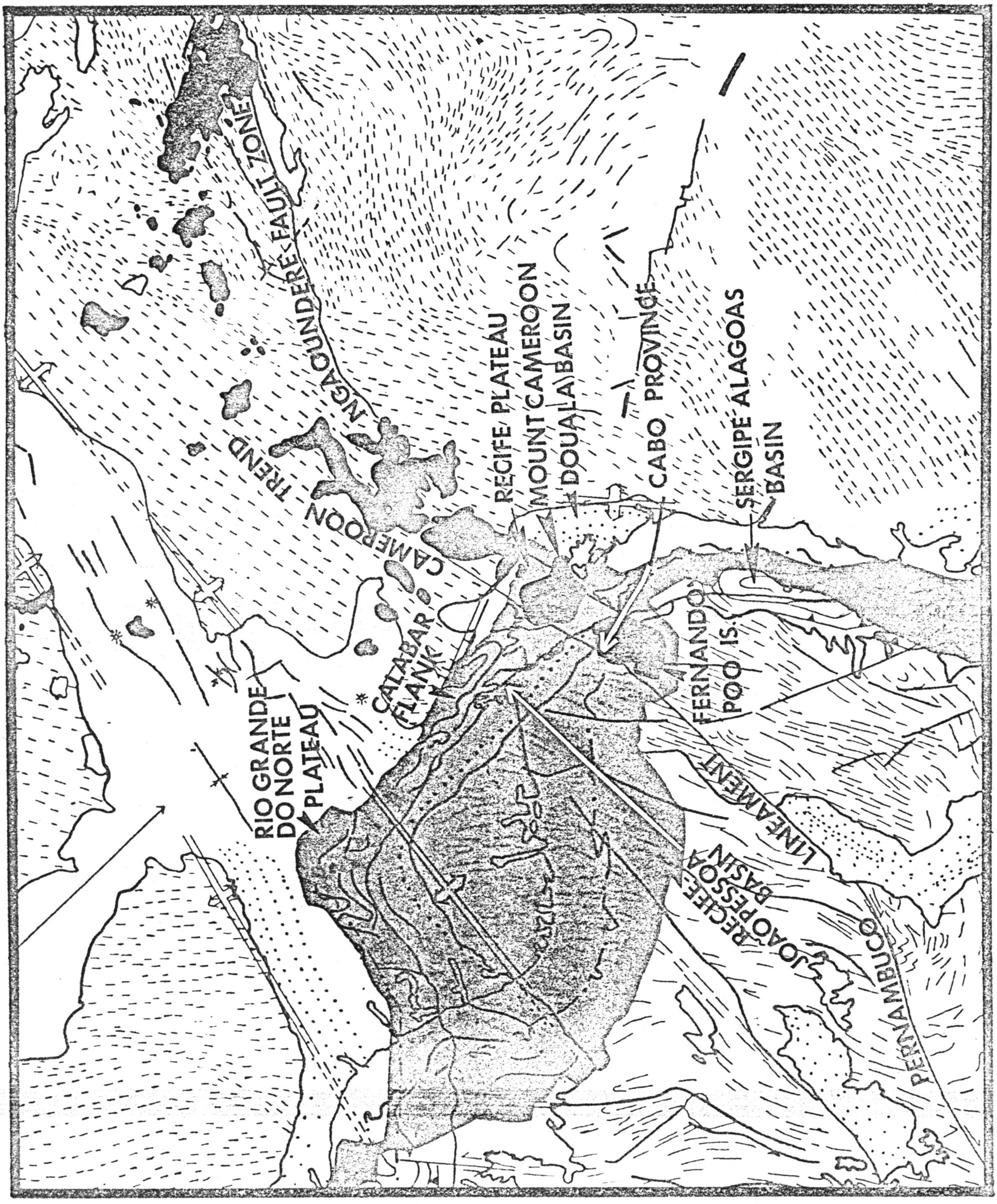

NGAOUNDERE FAULT ZONE
CAMEROON TREND
RECIFE PLATEAU
MOUNT CAMEROON
DOUALA BASIN
CABO PROVINCE
SERGIPE ALAGOAS BASIN
FERNANDO POO IS.
CALABAR FLANK
RIO GRANDE DO NORTE PLATEAU
PERNAMBUCO LINEAMENT
JOAO PESSOA RECIFE BASIN

Figure 77

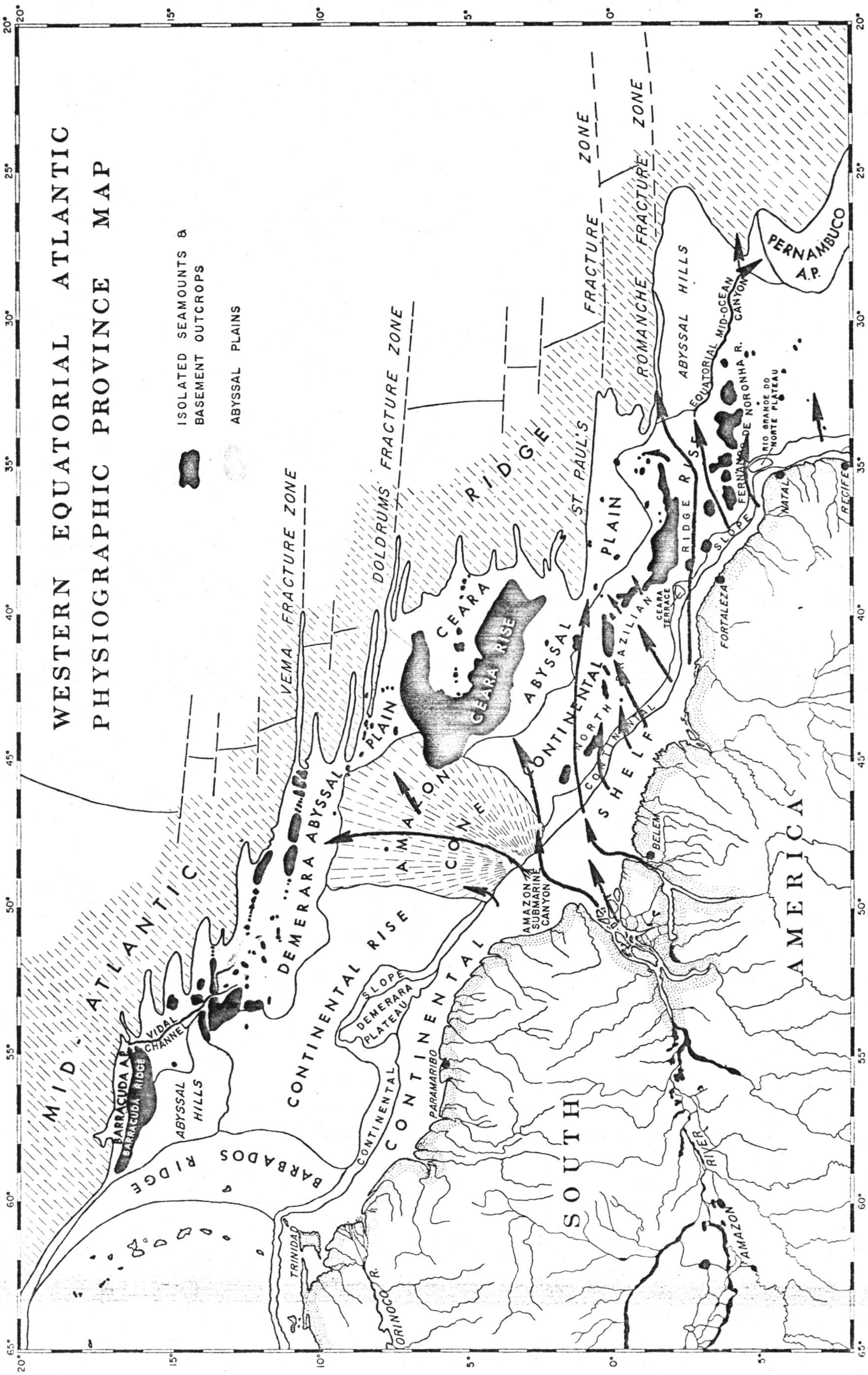
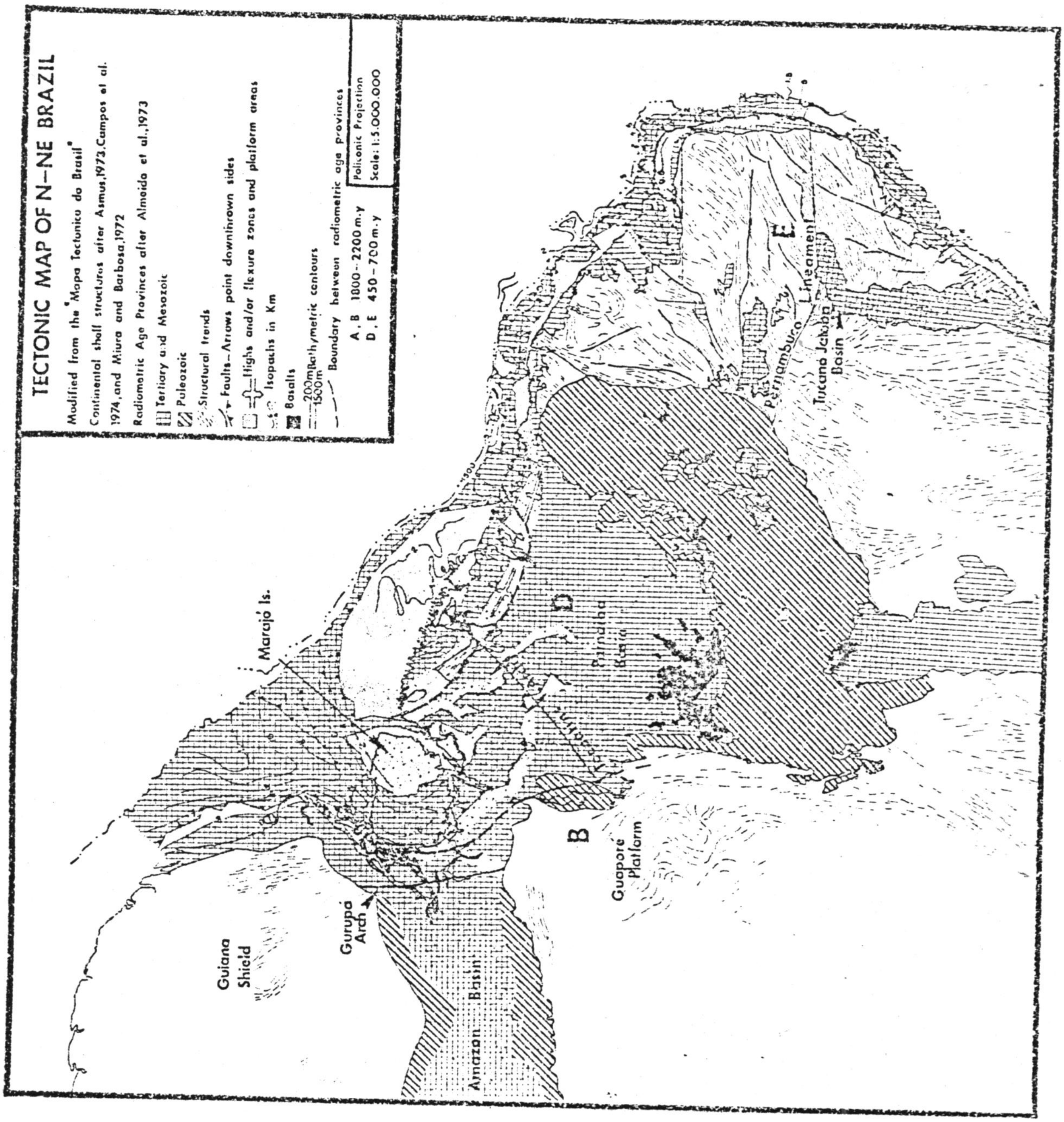

WESTERN EQUATORIAL ATLANTIC
PHYSIOGRAPHIC PROVINCE MAP
ISOLATED SEAMOUNTS & BASEMENT OUTCROPS
ABYSSAL PLAINS
MID-ATLANTIC
RIDGE
VEMA FRACTURE ZONE
DOLDRUMS FRACTURE ZONE
ST. PAUL'S FRACTURE ZONE
ROMANCHE FRACTURE ZONE
BARRACUDA A.P.
BARRACUDA RIDGE
VIDAL CHANNEL
ABYSSAL HILLS
BARBADOS RIDGE
CONTINENTAL RISE
DEMERARA ABYSSAL PLAIN
DEMERARA PLATEAU
SLOPE
CONTINENTAL
AMAZON CONE
AMAZON SUBMARINE CANYON
CEARA ABYSSAL PLAIN
CEARA RISE
CONTINENTAL RISE
NORTH BRAZILIAN RIDGE
CEARA TERRACE
CONTINENTAL SHELF
SLOPE
EQUATORIAL MID-OCEAN CANYON
FERNANDO DE NORONHA R.
RIO GRANDE DO NORTE PLATEAU
PERNAMBUCO A.P.
SOUTH AMERICA
TRINIDAD
ORINOCO R.
PARAMARIBO
BELEM
FORTALEZA
NATAL
RECIFE
AMAZON RIVER
20°
15°
10°
5°
0°
5°
20°
65°
60°
55°
50°
45°
40°
35°
30°
25°
20°

Figure 78



TECTONIC MAP OF N–NE BRAZIL
Modified from the "Mapa Tectonico do Brasil"
Continental shelf structures after Asmus,1973,Campos et al. 1974,and Miura and Barbosa,1972
Radiometric Age Provinces after Almeida et al.,1973
Tertiary and Mesozoic
Paleozoic
Structural trends
Faults–Arrows point downthrown sides
Highs and/or flexure zones and platform areas
Isopachs in Km
Basalts
200m
1500m
Bathymetric contours
Boundary between radiometric age provinces
A, B 1800–2200 m.y
D, E 450–700 m.y
Policonic Projection
Scale: 1:5.000.000
Marajó Is.
Guiana Shield
Gurupá Arch
Amazon Basin
B
Guaporé Platform
D
Parnaiba Basin
E
Pernambuco Lineament
Tucano Jatoba Basin

Figure A1

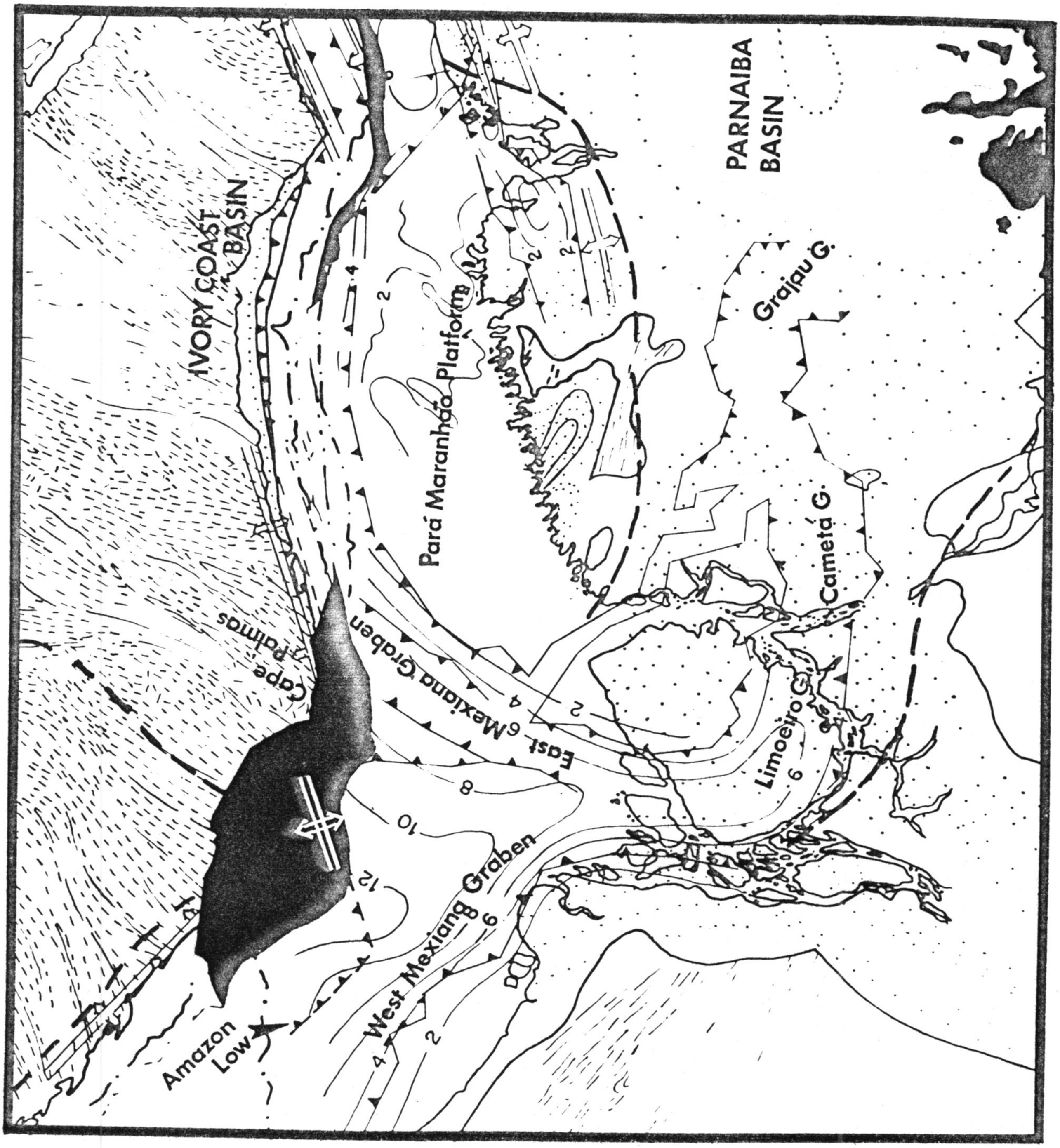
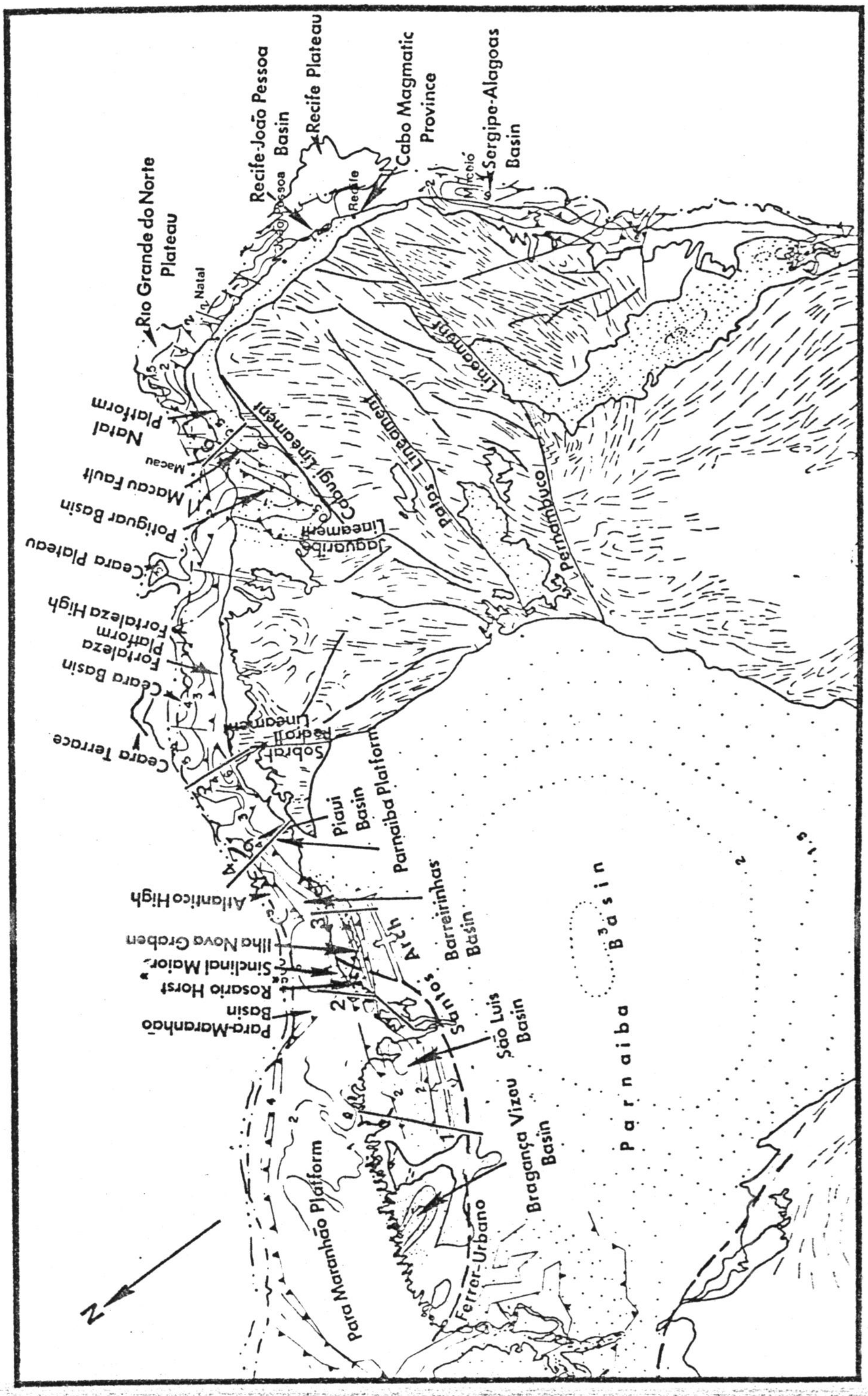

IVORY COAST BASIN
Cape Palmas
Amazon Low
West Mexiana Graben
East Mexiana Graben
Pará Maranhão Platform
PARNAIBA BASIN
Grajau G.
Cametá G.
Limoeiro G.

Figure A2



Rio Grande do Norte Plateau
Natal
Recife-João Pessoa Basin
João Pessoa
Recife Plateau
Recife
Cabo Magmatic Province
Maceió
Sergipe-Alagoas Basin
Natal Platform
Macau
Macau Fault
Potiguar Basin
Ceara Plateau
Fortaleza High
Fortaleza Platform
Ceara Basin
Ceara Terrace
Cabugi Lineament
Jaguaribe Lineament
Patos Lineament
Pernambuco Lineament
Sobral Pedro II Lineament
Piaui Basin
Parnaiba Platform
Atlantico High
Ilha Nova Graben
Sinclinal Maior
Rosario Horst
Para-Maranhão Basin
Santos Arch
Barreirinhas Basin
São Luis Basin
Parnaiba Basin
Bragança Vizeu Basin
Ferrer-Urbano
Para Maranhão Platform
N

Figure A3

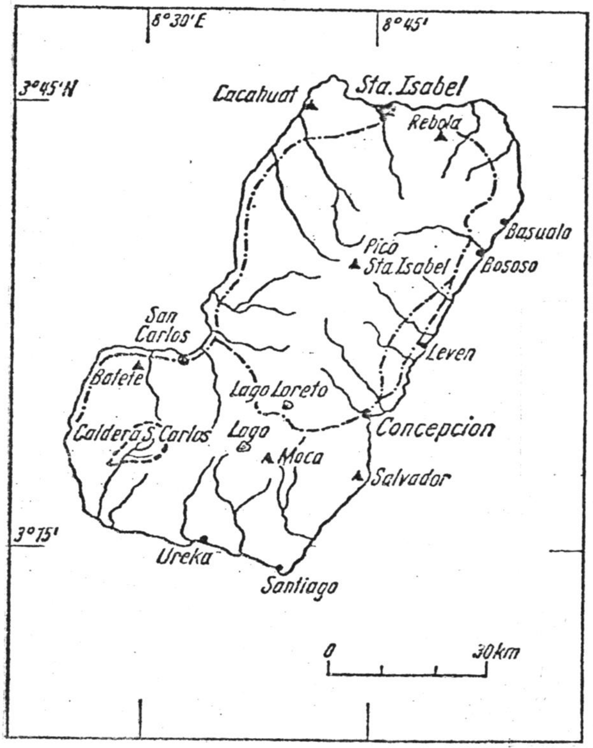

8°30'E
8°45'
3°45'N
Cacahuat
Sta. Isabel
Rebola
Basuala
Pico
Sta. Isabel
Bososo
San
Carlos
Leven
Batete
Lago Loreto
Concepcion
Caldera S. Carlos
Lago
Moca
Salvador
3°75'
Ureka
Santiago
0
30km

Figure A4

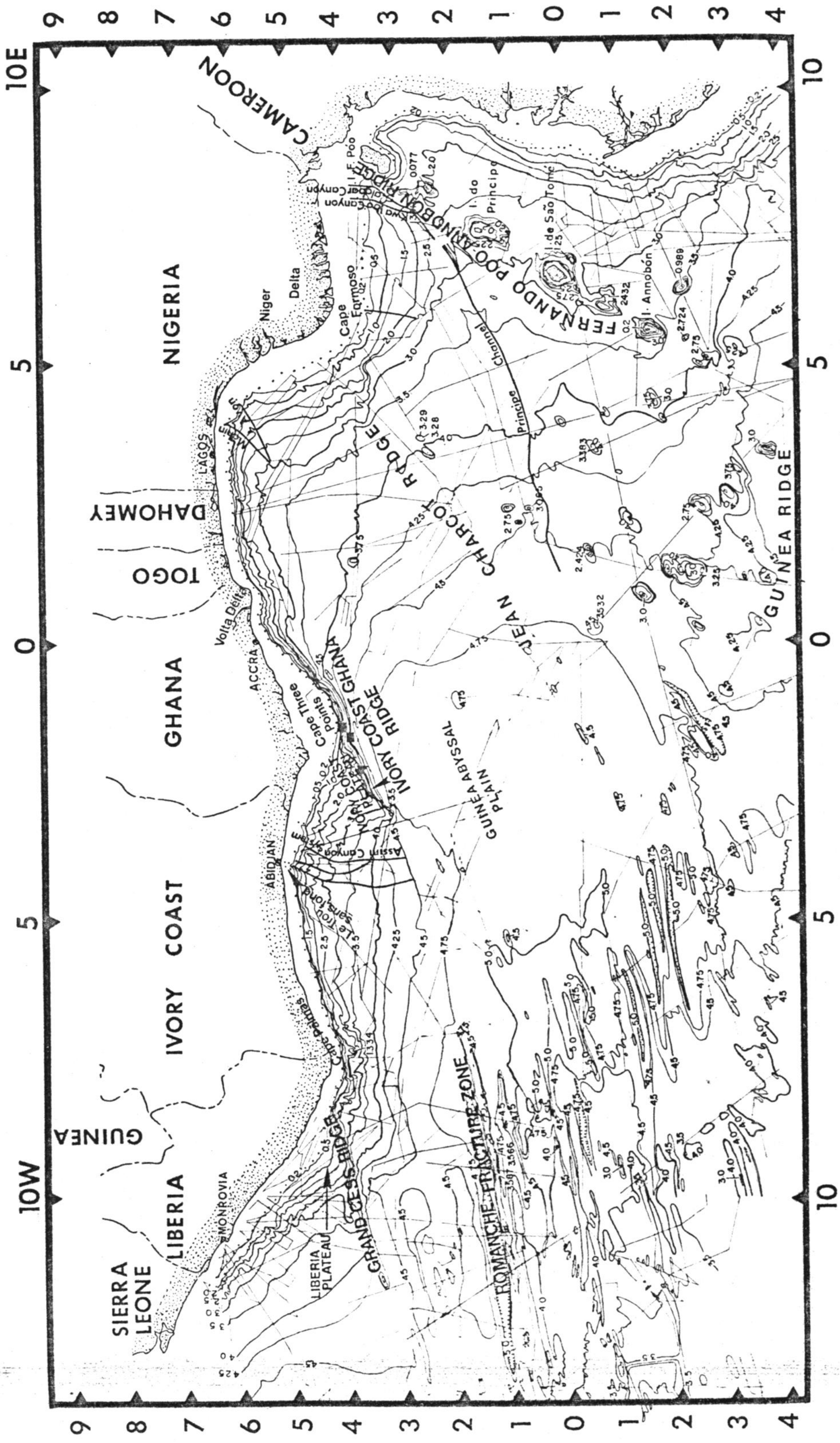

10W
5
0
5
10E
SIERRA LEONE
LIBERIA
GUINEA
IVORY COAST
GHANA
TOGO
DAHOMEY
NIGERIA
CAMEROON
MONROVIA
ABIDJAN
ACCRA
LAGOS
Volta Delta
Niger Delta
Cape Formoso
Cape Three Points
Cape Palmas
LIBERIA PLATEAU
GRAND CESS RIDGE
ROMANCHE FRACTURE ZONE
IVORY COAST RIDGE
IVORY COAST GHANA
GUINEA ABYSSAL PLAIN
CHARCOT RIDGE
JEAN
Principe Channel
FERNANDO POO ANNOBON RIDGE
I. do Principe
I. de São Tomé
I. Annobon
GUINEA RIDGE
Assini Canyon

Figure A5

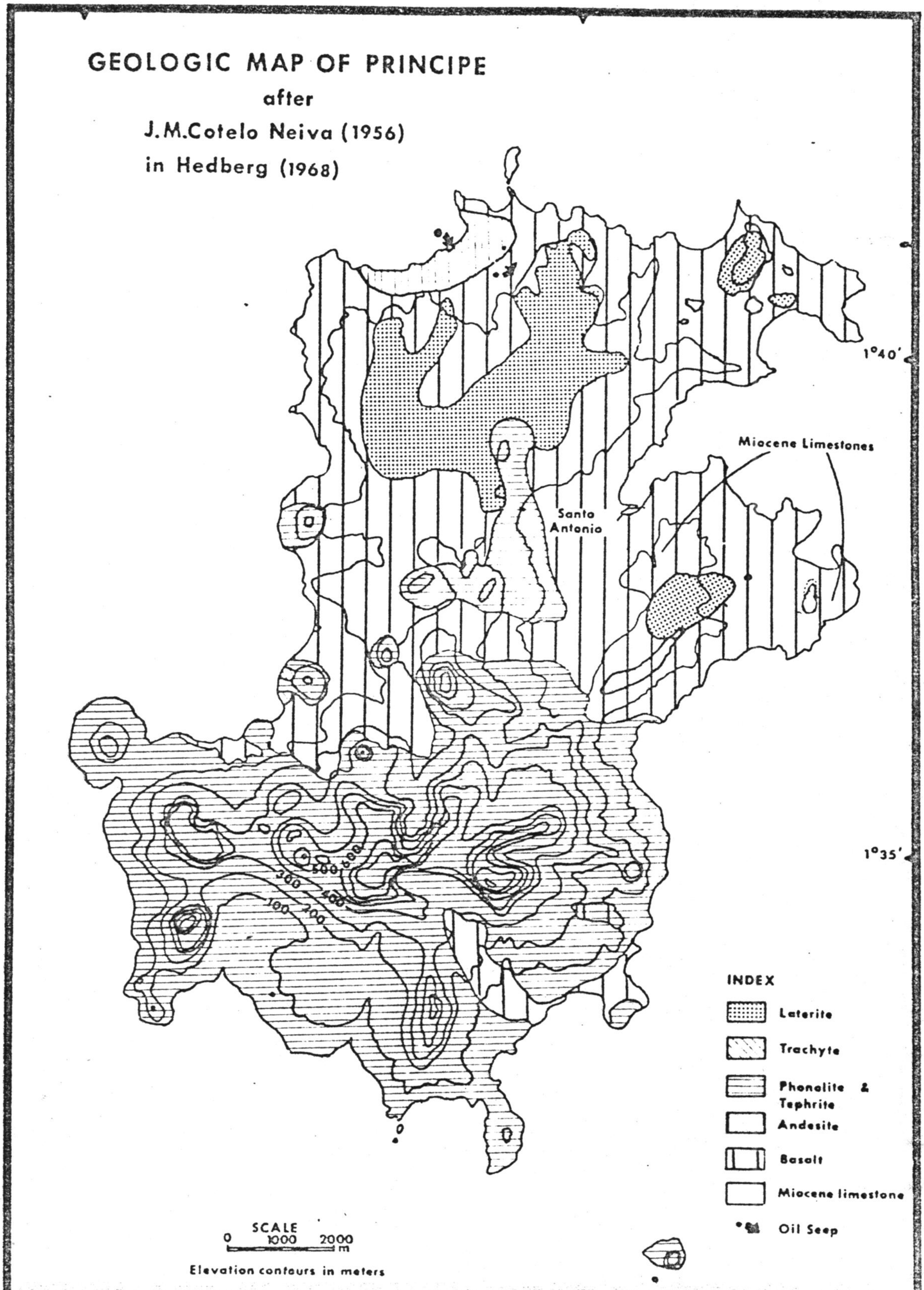

GEOLOGIC MAP OF PRINCIPE
after
J.M.Cotelo Neiva (1956)
in Hedberg (1968)
1°40'
Miocene Limestones
Santo Antonio
1°35'
500
600
300
400
100
200
INDEX
Laterite
Trachyte
Phonolite & Tephrite
Andesite
Basalt
Miocene limestone
Oil Seep
SCALE
0 1000 2000 m
Elevation contours in meters
7°20'
7°25'

Figure A6

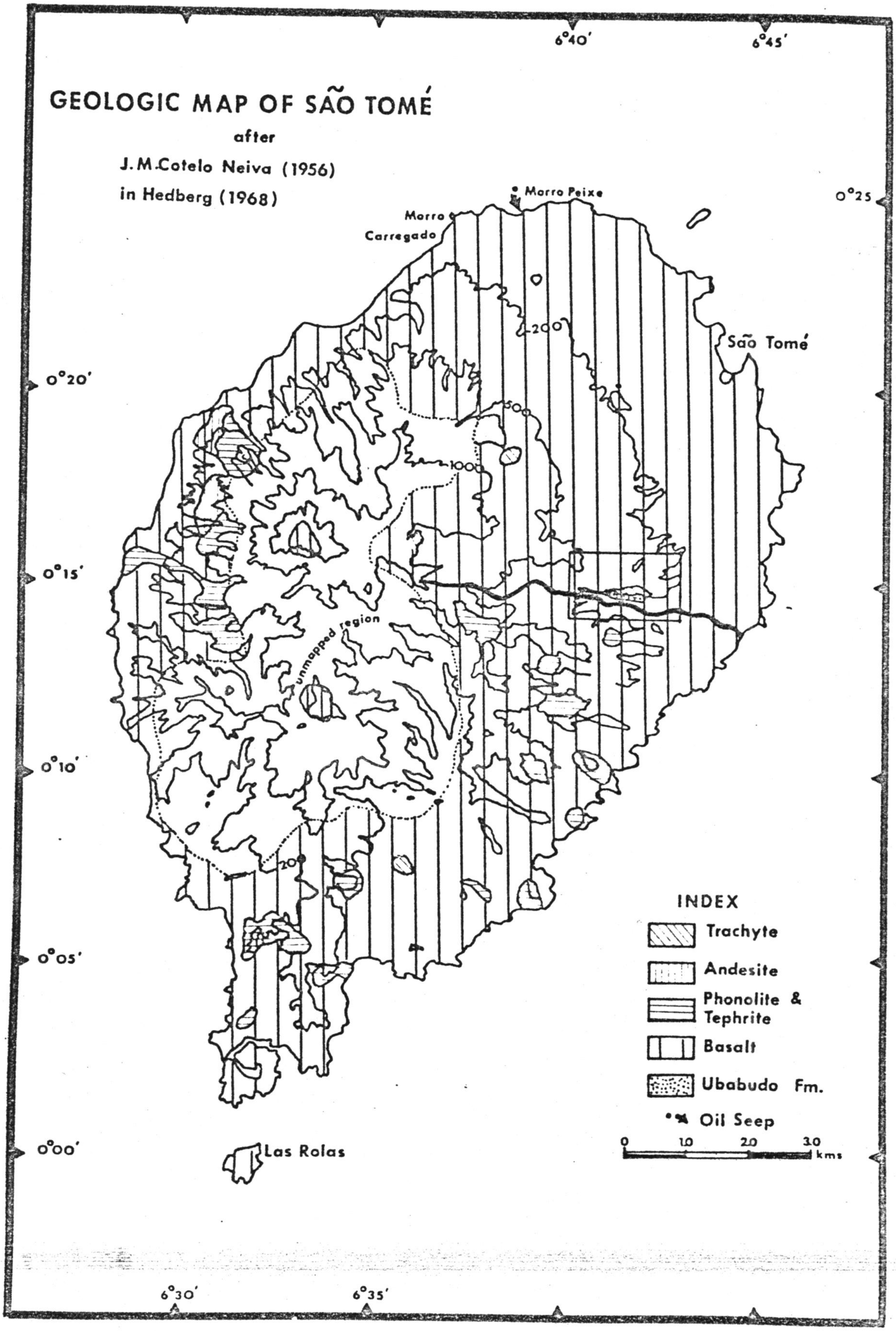

GEOLOGIC MAP OF SÃO TOMÉ
after
J. M. Cotelo Neiva (1956)
in Hedberg (1968)
6°40'
6°45'
0°25
Morro Peixe
Morro
Carregado
São Tomé
200
500
1000
0°20'
0°15'
0°10'
0°05'
0°00'
unmapped region
INDEX
Trachyte
Andesite
Phonolite & Tephrite
Basalt
Ubabudo Fm.
Oil Seep
0
1.0
2.0
3.0
kms
Las Rolas
6°30'
6°35'

Figure A7

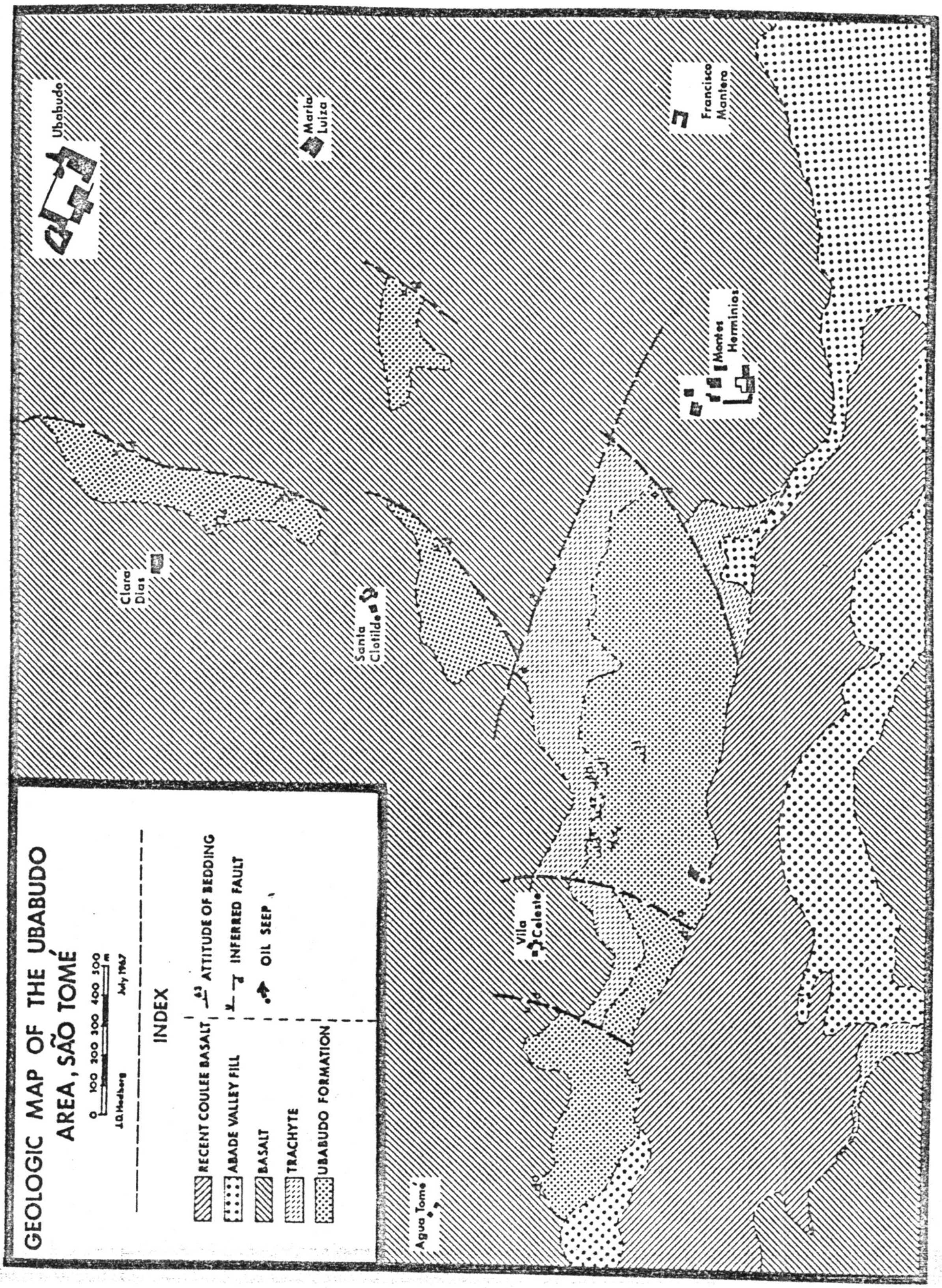

GEOLOGIC MAP OF THE UBABUDO AREA, SÃO TOMÉ
0 100 200 300 400 500 m
J.D. Hedberg
July 1967
INDEX
RECENT COULEE BASALT
ABADE VALLEY FILL
BASALT
TRACHYTE
UBABUDO FORMATION
ATTITUDE OF BEDDING
INFERRED FAULT
OIL SEEP
Ubabudo
Maria Luiza
Francisco Mantero
Montes Herminios
Clara Dias
Santa Clotilde
Vila Celeste
Agua Tomé

Figure A8

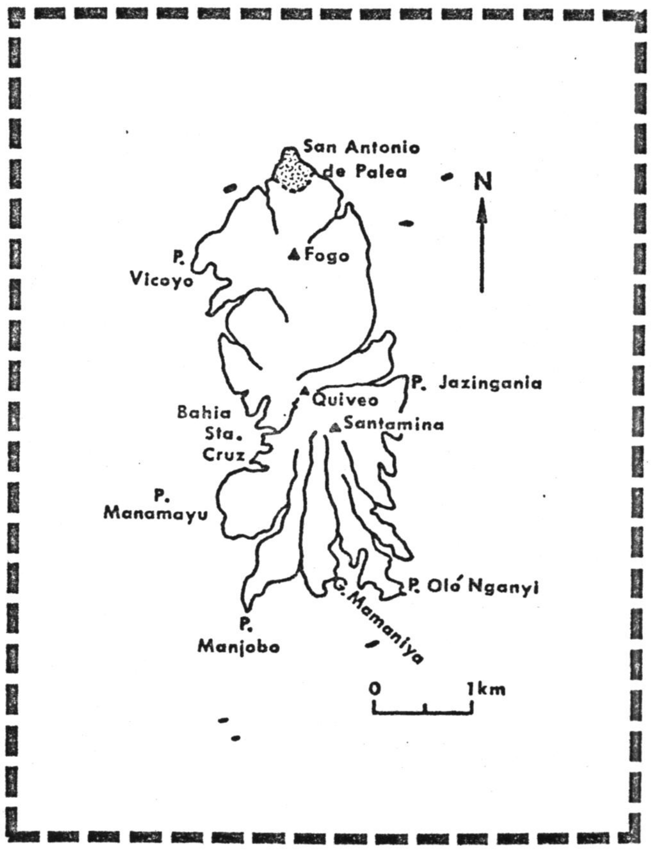

San Antonio de Palea
N
Fogo
P. Vicoyo
P. Jazingania
Quiveo
Santamina
Bahia Sta. Cruz
P. Manamayu
P. Olo Nganyi
P. Manjobo
C. Mamaniya
0
1km

Figure A9

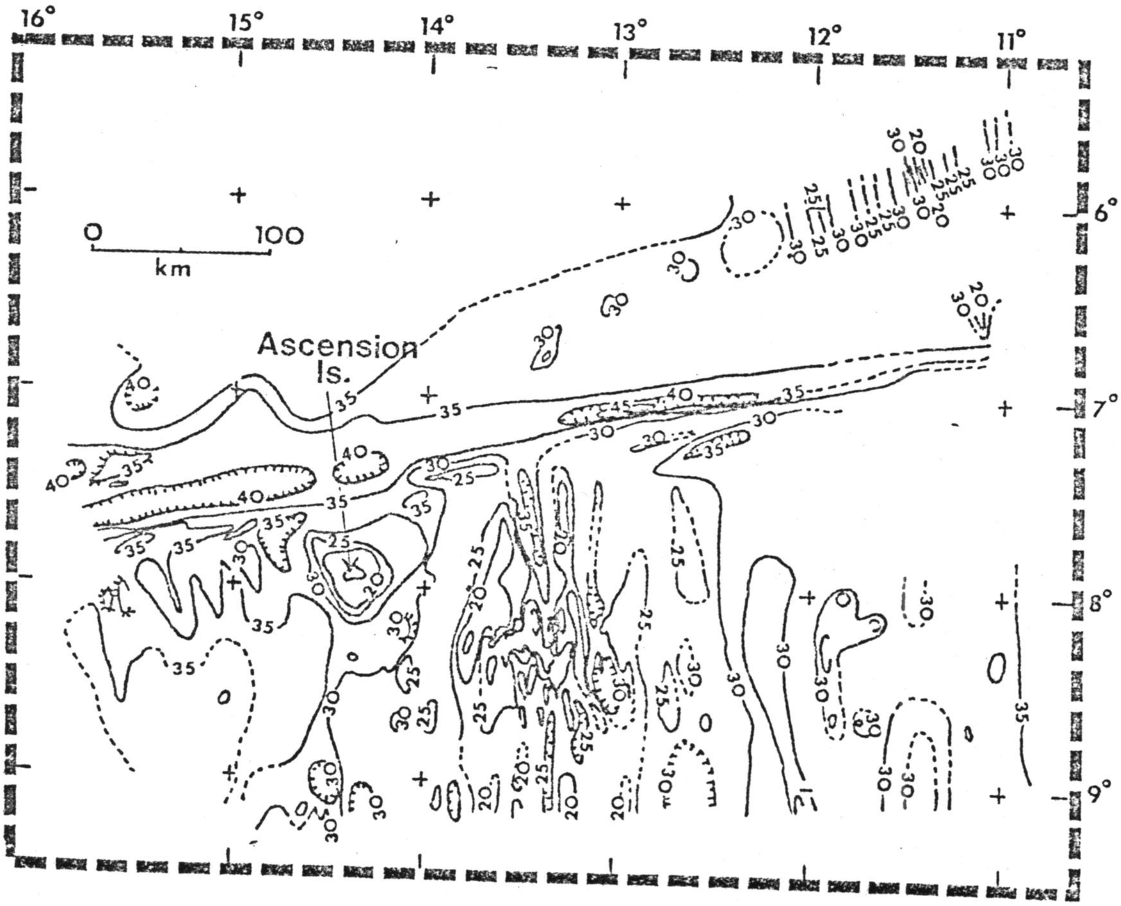
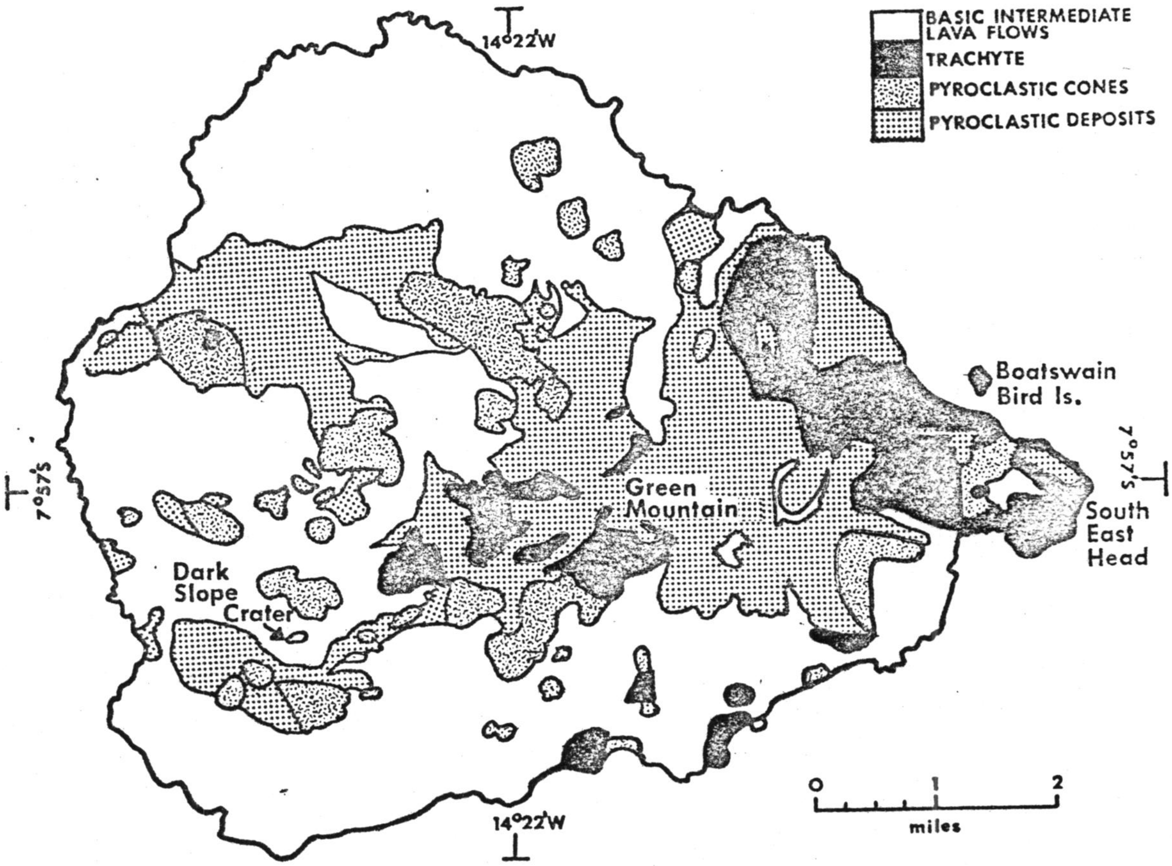


Figure A10



BASIC INTERMEDIATE LAVA FLOWS
TRACHYTE
PYROCLASTIC CONES
PYROCLASTIC DEPOSITS
14°22'W
7°57'S
Boatswain Bird Is.
Green Mountain
South East Head
Dark Slope Crater
0
1
2
miles

Figure A11

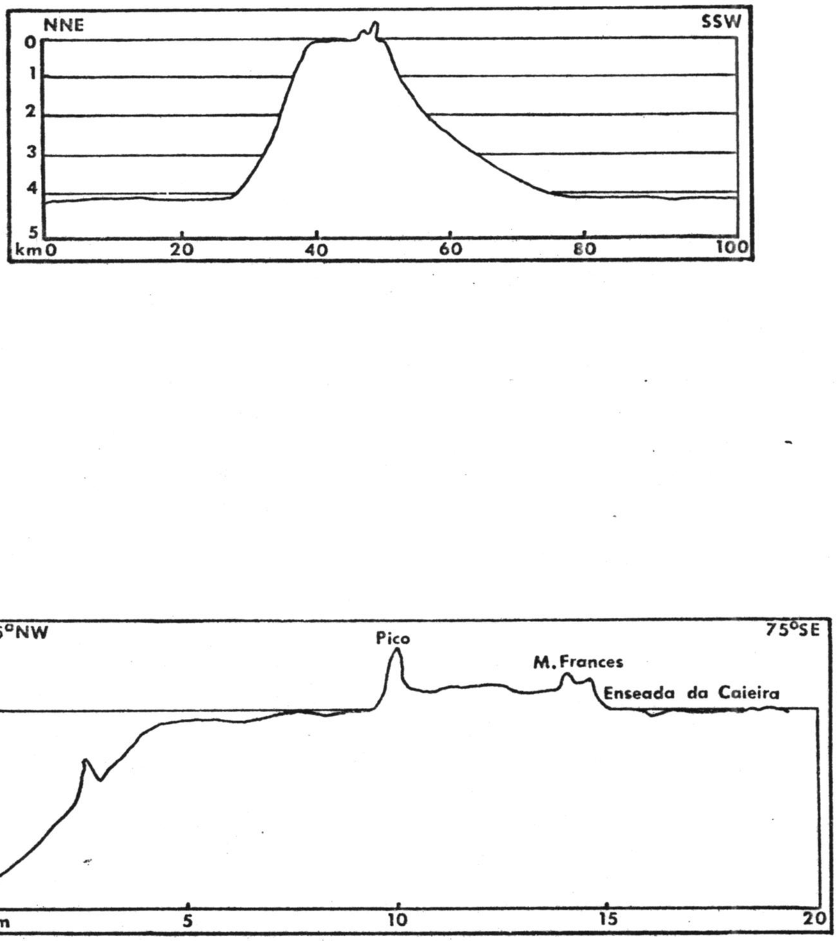
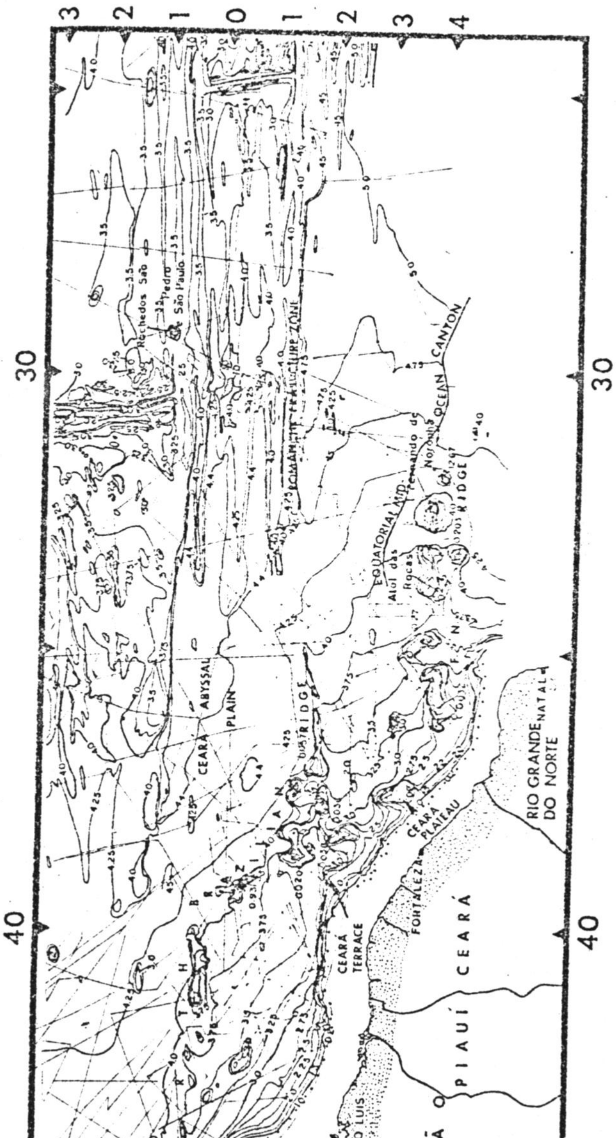

NNE
SSW
0
1
2
3
4
5
km 0
20
40
60
80
100
400
75°NW
75°SE
Pico
M. Frances
Enseada da Caieira
200
0 m
200
400
600
800
0 km
5
10
15
20

Figure A12



50W
40
30
CEARÁ ABYSSAL PLAIN
ROMANCHE FRACTURE ZONE
EQUATORIAL MID OCEAN CANYON
Atol das Rocas
RIO GRANDE DO NORTE
NATAL
CEARÁ PLATEAU
FORTALEZA
CEARÁ TERRACE
CEARÁ
PIAUÍ
MARANHÃO
SÃO LUIS
BRASIL
PARÁ
BELÉM

Figure A13

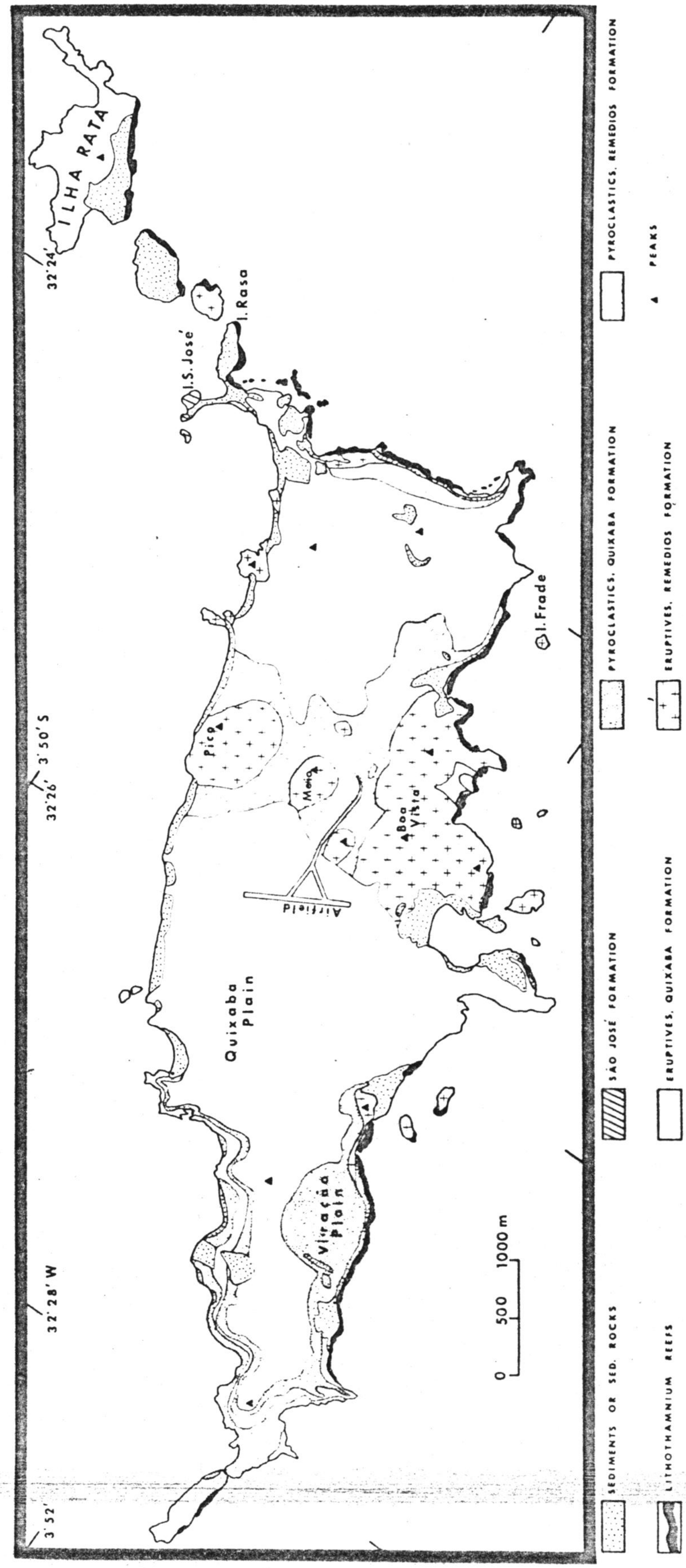

ILHA RATA
32°24'
I. Rasa
I.S. José
I. Frade
32°26'
3°50'S
Pico
Boa Vista
Airfield
Quixaba Plain
Viração Plain
32°28'W
3°52'
0 500 1000 m
SEDIMENTS OR SED. ROCKS
LITHOTHAMNIUM REEFS
SÃO JOSÉ FORMATION
ERUPTIVES, QUIXABA FORMATION
PYROCLASTICS, QUIXABA FORMATION
ERUPTIVES, REMEDIOS FORMATION
PYROCLASTICS, REMEDIOS FORMATION
PEAKS

Figure A14

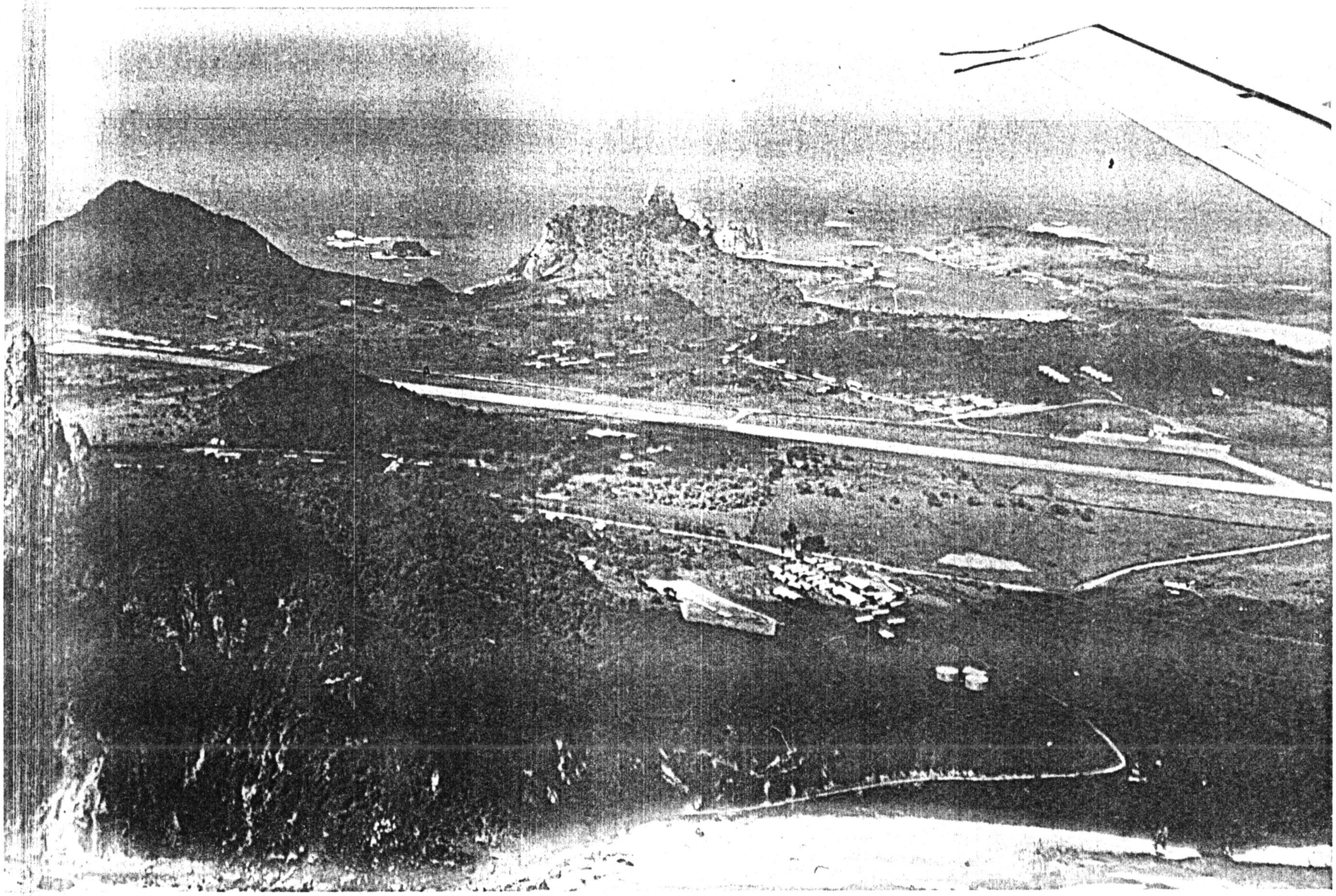
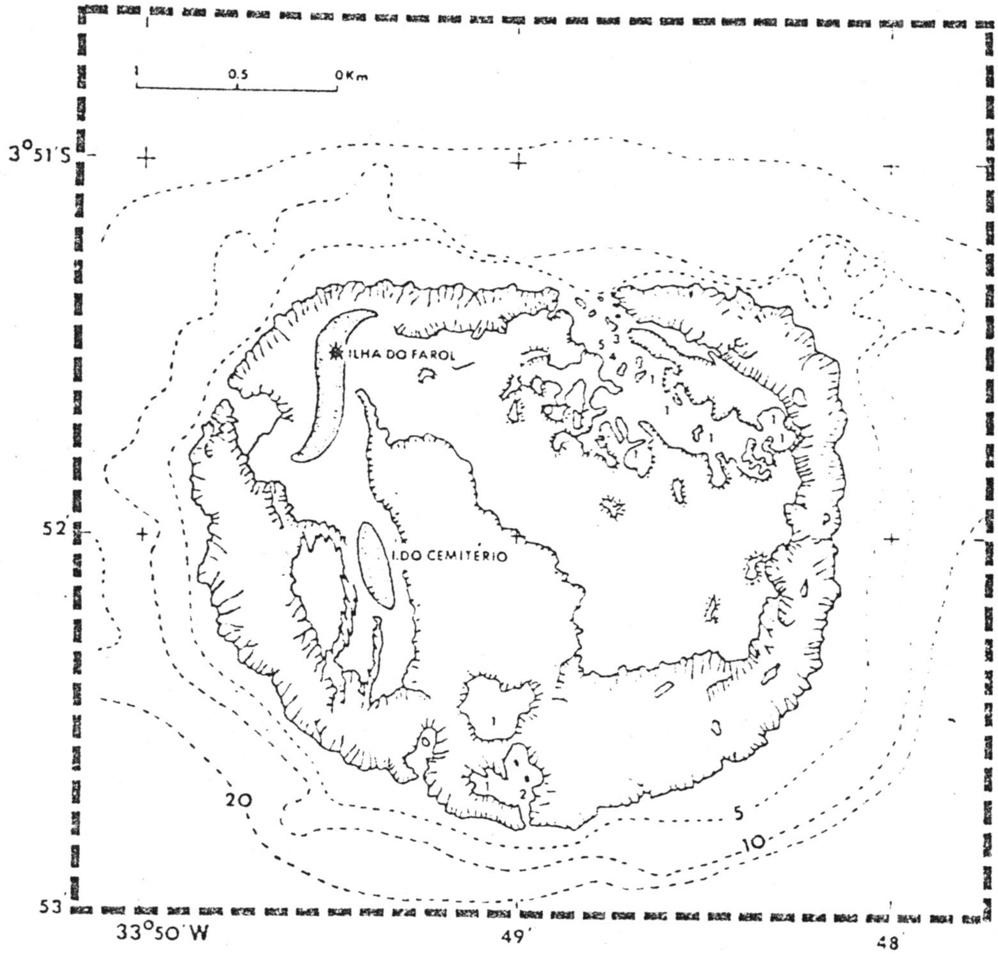
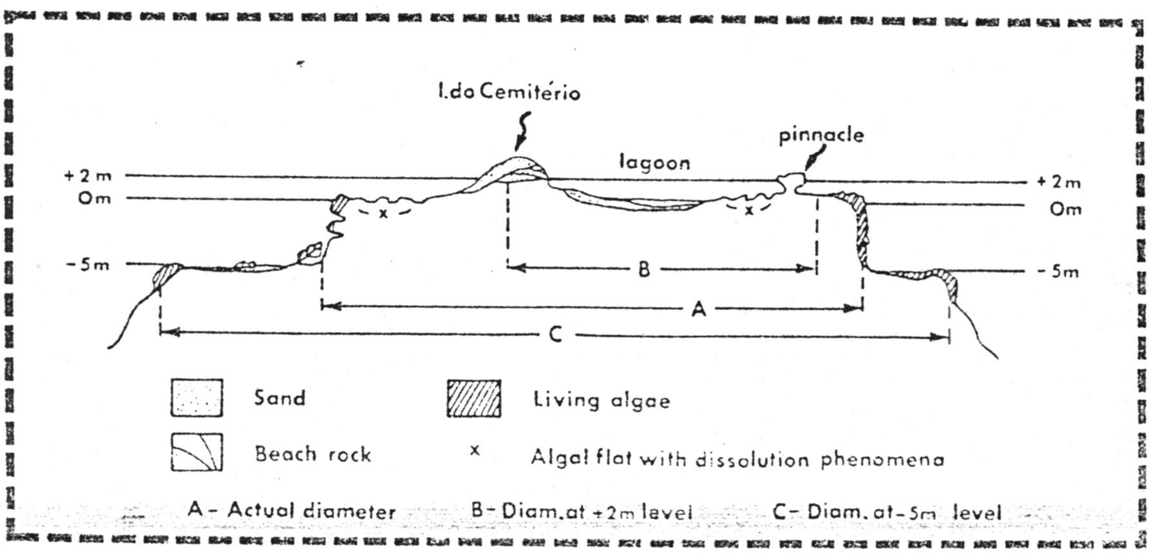


Figure A15



1
0.5
0 Km
3°51'S
ILHA DO FAROL
52'
I.DO CEMITÉRIO
20
5
10
53'
33°50'W
49'
48'



I.do Cemitério
pinnacle
lagoon
+2 m
0m
-5m
B
A
C
Sand
Living algae
Beach rock
x Algal flat with dissolution phenomena
A - Actual diameter
B - Diam. at +2m level
C - Diam. at -5m level

Figure A16