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Equatorial Segment of the Mid-Atlantic Ridge

Initial Results of the Geological
and Geophysical Investigations
under the EQUARIDGE Program,
Cruises of r/v 'Akademik Nikolaj Strakhov'
in 1987, 1990, 1991

Editor: G.B. Udintsev

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ABSTRACT

The Equatorial Segment of the Mid-Atlantic Ridge is a part of this mid-oceanic ridge limited by a cluster of fracture zones - Cape Verde, Marathon, Mercury, Vema, Doldrums, Vemadsky and Sierra Leone - in the North, and a similar cluser of fracture zones - St. Paul, Romanche and Chain - in the South. During recent decades, following the publication of the 5th edition of the General Bathymetric Chart of the Oceans (GEBCO), there has been a great deal of geological-geophysical research and mapping of the World Ocean. The results have led to the development of a number of theories concerning the essential heterogeneity of the structure of the ocean floor and, in particular, the heterogeneity of the structure and segmentation of mid-oceanic ridges. Research on the nature of such segmentation is of great importance for an understanding of the processes of development of such ridges and oceanic basins as a whole.

Detailed surveys of the Equatorial Segment of the Mid-Atlantic Ridge were carried out under an international programme, EQUARIDGE, supported by the Intergovernmental Oceanographic Commission (UNESCO), during three cruises (7, 11 and 12) of the Russian R/V “Akademik Nikolaj Strakhov” (Geological Institute of the Academy of Sciences, USSR) between 1988 and 1991.

On the basis of the results obtained, the participants in the expeditions concluded that the numerous faults which intersect the equatorial part of the Mid-Atlantic ridge are of varying ages and relate differently to the structures of the rift zone in time and space. The data indicate that the structure of the Equatorial Segment is anomalous and differs substantively from the adjacent north and south parts of the ridge. It was also observed that spreading in this area is clearly evident only in the narrow axial zone of the Segment. The structure of the Segment’s flanks is mosaic in appearance, reassembling a plateau rather than a spreading system.

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Introduction.

G.B. Udintsev, J. Makris

The rapid progress made by the earth sciences over the last decade would not have been possible without considerable advances in the microelectronic and computer sciences. As a result of these developments, the geophysicist can now choose from a wide variety of techniques and instrumentation, including multibeam echosounders and powerful side-scan sonar systems for highly acurate and well resolved mapping of the sea-floor, multi-channel streamers, ocean bottom seismographs (OBS), gravity meters, manometers, deep sea drilling, and dredging and coring of the sediments. The new generation of bathymetric charts, e.g. the GEBCO maps, reveal the complex topography of the oceans with astounding accuracy, while thanks to the Global Positioning System (GPS), positioning at sea has attained a precision undreamt of only a few years ago.

These technological innovations have resulted in the collection of an abundance of good quality data which in turn have stimulated the development of new ideas and concepts, The complexity of the sea-floor and the inhomogeneity of the crust and mantle in both oceanic and continental areas, which is becoming more and more evident, is the subject of heated discussion and debate in the scientific community. In this context, the study of mid-ocean ridges is of particular importance, since it is here that oceanic crust is generated.

Clear examples of segmentation of the mid-oceanic ridges have been observed in a number of highly detailed studies of the Mid-Atlantic Ridge (MAR). One of the best examples is the equatorial segment of the ridge, limited to the north by a series of fracture zones (Vema, Doldrums, Vemadsky), and in the south by a further series (Sao Paolo, Romanche, Ascension).

The assumption regarding the the anomalous character of the ES MAR arose a long enough time ago in relation to the anomalously old (for the area close to modem rift zone) age of the igneous rocks of the Island St. Peter and Paul (Sao Paulu) (Hart, 1964), This was also a result of the finds (mainly by E. Bonatti) on the ocean floor in this area of igneous and sedimentary rocks with anomalously old age and with evidence of shallow-water or continental origin (Cifelli et al., 1968, Frey, 1970, Melson, Thompson, 1970,1971, 1973, Bonatti, Honnorez, 1971, Bonatti et al., 1973, Bonatti, 1973, Bonatti, Honnorez, 1976, Barash, Lavrov, 1986, Bonatti, 1978, Bonatti et al,, 1979, Bonatti, Chermak, 1981, Bonatti, Crane, 1982, Melson et al., 1982, Bonatti et al., 1983, Bonatti, Crane, 1984, Emery, Uchupi, 1984, Vogt, Tucholke, 1986).

The anomality of the ES MAR is displayed in its morphology with the unusually gentle slopes of the flanks, well observed on bathymetric, physiographic and gravity maps (Heezen, Tharp, 1977, GEBCO, 1984, Haxby, 1988, Sandwell, Smith, 1996). It is also displayed in specific features of the deep structure, in structural features of the long-period anomalies of the magnetic field, observed with Project MAGSAT (Langel et al., 1982), in the structure of the short-period magnetic anomalies, observed at the eastern flank (Gordin, 1986), in the anomalously cold upper mantle (Dziewonski, Woodhouse, 1986), and in the mosaic structure of the Earth's crust (Kogan, Malovitsky, Udintsev, 1977, Kogan, 1988). We find most probable the hypothesis proposed by E. Bonatti on the nonspreading origin of the ES and, alternatively, regarding the spreading origin in conditions of jumping of the rift and transform faults, which resulted in the existence of blocks of the crust with anomalous age and structure.

The anomality of the ES MAR is also manifested in the positioning on its eastern flank of the large submarine Sierra Leone Rise. One of the efforts at beginning systematic research on the eastern part of the ES MAR were the expeditions organized in 1981 by the Institute of the Physics of the Earth of the USSR Academy of Sciences with r/v "Ivan Kireev" (Gordin, 1986, Efimov, Beresnev, Agapova, Zenkevitch, 1989-1990), and expeditions of the Ministry of Geology of the USSR in 1978-1980 (Junov, 1989-1990).

The first expedition for systematic research of ES MAR was the 7th cruise of r/v “Akademik Nikolaj Strakhov” of the Geological Institute of the USSR Ac.Sci. (1.04.4.08. 1988). From the very beginning, this expedition was arranged as an international project, Cooperation was arranged with scientists of Brazil, USA and the United Kingdom. According to the usual procedure for the Geological Institute, the adviser to the project was one of-the leading scientists of the Institute, Prof. P. Timofeev. G. Udintsev was the Chief scientist for the 7th cruise of the r/v “Akademik Nikolaj Strakhov.”

The research of this expedition covered the part of ES from the 4°North Fracture Zone (later renamed the Strakhov Fracture Zone) in the North to the Saint Paul fracture zone in the South, and from 26° W in the East (i.e. from the eastern margin of the Sierra Leone Rise) to 36°W, and a separate sections up to the submarine continental margin of NE Brazil. The research included multibeam echosounding, continuous single channel seismic profiling, magnetic survey, core sampling and dredging. Geophysical surveys were conducted along the system of W-E profiles with intervals of between them about 5 miles. Two areas were surveyed in detail, along the systems of N-S profiles positioned with intervals of about 2 miles the active part of the 4°North f.z. and the eastern rift-transform part of the Saint Paul f.z.

In the light of the data obtained by the expedition, it became clear, that ES in fact has essential features distinct from the adjacent segments of the MAR, usually considered as typical for the ridge. In accordance with the data obtained, the hypothesis was proposed that this segment is a kind of structural barrier between the North and the South Atlantic (Timofeev, Udintsev, et al, 1990).

The cruise participants considered that their assumptions were based on data insufficient for definite conclusions, and served solely as a working hypothesis, an alternative to the popular idea of the spreading origin of the structure of the ES. However, the reliable further development of this or any other alternative hypothesis required more detailed and complex investigation. Nevertheless, the solution of the problem of the nature of the ES seemed important to us for an understanding of the causes of the segmentation of the MAR of its areal inhomogeneity, and of the active factors of the ridge's development.

We therefore proposed an international program of further studies of the ES. To carry out this program the Geological Institute organized two expeditions of the r/v "Akademik Nikolaj Strakhov": the 11th cruise (scientific adviser Ju.Pucharovsky), and the 12th cruise (scientific adviser P. Timofeev). The results of the research of these two cruises, together with those of the 7th cruise, were designed as the basis for the International Program "Equaridge" (Scientific Adviser G. Udintsev), under the auspices of the Division of Geology, Geophysics, Geochemistry and Mining Science of the Academy of Sciences of the USSR, and of the Intergovernmental Oceanographic Commission (IOC) UNESCO. "Equaridge" is one of the mid-oceanic ridge programs such as "Ridge" (USA), "Bridge" (UK) and the French,. Japanese and other projects under "Interridge," the International cooperative program for studies of mid-oceanic ridges.

Research under the "Equaridge" program was carried out as follows the 11th cruise was completed on 10.08-9.12.1990, and the 12th cruise on 31.01-5.05.1991. As distinguished from the the 7th cruise, they were based on a detailed polygon survey. The research covered the axial rift zone and flank plateaus from 5°N in the North to 0°50'N in the South, and from 28°00'W in the East to 34°30W in the West. The tracks at the polygon surveys were spaced with an average interval of 1,5 miles and this provided about 70% coverage of the area with multibeam sounding. Reconnaissance surveys were continued on a number of tracks between the main area of the survey to ports of fuel supply in northeastern Brazil: Recife and Fortaleza. G. Udintsev continued to act as Chief Scientist for both cruises.

Given the international nature of the program, scientists from Brazil, Germany, USA, UK, Israel, Italy, and Sweden participated in these three cruises, and their contributions were most valuable. The , German group from the Institute of Geophysics of the University of Hamburg (Director Professor Jannis Makris) provided substantive financial support, a gravitymeter, magnetometer, ocean bottom seismographs, a large airgun and long seismic array of multi-channel configuration, and worked actively with this equipment. Professor M. Gorini (Niteroy University, Brazil) shared the valuable results of his investigations in this area and assisted in obtaining permission from the Brazilian government for work in the 200-mile exclusive economic zone surrounding the Brazilian St. Paul Islands. The Brazilian group from Niteroy University, Petrobras University, and the Hygrographic Survey of Brazil was active in echosounding and seismic profiling. Professor E. Bonatti (Institute of Marine Geology, Bologna University, Italy) shared his experience studying this area and took an active part in the dredging. Professor John Hall (Geological Survey of Israel, Jerusalem) supplied contemporary computer technology and methods of map compilation, and was active in work on echosounding. The scientists from the United Kingdom (Scientific Adviser: Dr. Peter Hunter, Southampton Oceanography Centre) provided a considerable amount of bathymetric maps and data,and were active in work on echosounding and seismic profiling.

In addition to multibeam echosounding, continuous single channel seismic profiling, the research of the 11th and 12th cruises of r/v "Akademik Nikolaj Strakhov" included multi-channel seismic profiling, seismic refraction experiments and seismological observation using ocean bottom seismographs, gravity and magnetic survey, dredging and core sampling. The results of these two cruises yielded both a large amount of additional information on the structure of the ES and confirmed assumptions regarding its anomalous nature. Such information was highly useful in the development of ideas on the geological structure and history of the Equatorial Segment.

The following data was obtained as a result of the three cruises

1/ Complex geophysical survey, mostly on polygons and closely spaced profiles -39.910 miles,
2/ multi-channel seismics and seismic refraction - 5 profiles,
3/ seismicity observations - 5 points,
4/ basement rock samples - 81,
5/ core samples - 42

The results of the studies of all three cruises were worked on by a large group of scientists from many institutions, including some who did not participate in the expeditions. This book presents a summary of the results of that work and of data obtained within the ES in 1995 by the 21st cruise of r/v "Akademik Boris Petrov." Several chapters describing the results of other expeditions in adjacent regions are included to provide a better understanding of "the nature of the Equatorial Segment. We consider all the data important for grasping the correlation of the ES with the structural system of the Mid-Atlantic Ridge and Equatorial Atlantic in whole.

The participants of the "Equaridge" Program are most grateful to the administration of the Geological Institute and its scientific advisers for the opportunity for intensive and complex research on the three cruises of r/v "Akademik Nikolaj Strakhov", as such an opportunity for research on the same subject of investigation is extremely rare in the programs of today's oceanographic cruises, The participants in this Program hope that their efforts will lead to a better understanding of the nature of segmentation of the mid-oceanic ridges and of the inhomogeneity of the ocean floor.

This research was made possible by support of IOC UNESCO through its Programme on the Promotion of Marine Science (PROMAR), the RFFI (Russian Foundation for Basic Research) and the Ministry of Science and Technology of the Russian Federation through its program " World Ocean".

Chapter **1.**

Features of the Topography of the Ocean Floor in the Equatorial Atlantic

G.B.Udintsev

Since we single out the Equatorial segment (ES) of the Mid-Atlantic Ridge (MAR) as a massif, bounded in the North by a cluster of fracture zones of Cape Verde (15°20'N), Marathon, Mercury, Vema, Doldrums, Vernadsky and Sierra Leone, and in the South by a similar cluster of fracture zones of Saint Paul (Sao Paulo), Romanche, Chain, Charco and Bode Verde, we also consider as the Equatorial Atlantic a part of the Atlantic Ocean limited by these fracture zones. All these fracture zones belong to the category of trans-oceanic faults, crossing not only the axial zone of the MAR and its flanks, but also the floor of the oceanic basins adjacent to the ridge and submarine continental margins of Africa and South America. (Plates 1, 3).

It is noteworthy that the clusters of fracture zones on the margins of the ES are very complex and can be considered as transitional zones to neighboring segments of the MAR In the northern transition zone, besides the 6 transform faults mentioned above, marked by considerable lateral displacement of the axial rift, there are at least 7 untitled fractures with much smaller lateral displacements (Muller, Roest, 1992, Sandwell, Smith, 1996). Thus, the northern cluster consists of at least 12-14 fractures spaced at an interval of about 20-30 miles. The width of the fractures is about 10-15 miles, and narrow segments of the ridge showing a similar width between them most probably appear as transversive ridges rather than as segments of the midoceanic ridge. The southern cluster consists of a smaller number of fractures, but is also complex. The structure of the Saint Paul fracture zone, which consists of 4-5 troughs, separated by transversive crests, (Udintsev, Agapova, 1991) is also very complex, and the Romanche fracture zone, along with the Chain and Charco fracture zones, separated by narrow transversive ridges are of similar complexity. The largest lateral displacement occurs at the northern side of the Sao Paulo f. z. and along the southern side of the Romansh f.z..

Due to the system of these fracture zones the MAR is complexly fragmented in the Equatorial Atlantic, and its axial rift is exposed to a remarkable summary lateral displacement, occurring in the South Atlantic 2000 miles to the east of its position in the North Atlantic. This is the greatest lateral displacement of the axis of the midoceanic ridge observed in the Atlantic Ocean, and also in the World Ocean Rift System. Close to this are the lateral displacements along the Eltanin f.z. in the East Pacific Rise - about 500 miles - and along the Charlie Gibbs f.z. in the North-Atlantic Ridge - about 320 miles.

Within the limits between the northern and the southern clusters of fractures the ES of the MAR is about 400 miles long. The hypsometric position of this segment of the MAR is substantively higher than its neighboring segments in the north and south. The prevailing depths within its limits are less than 4500 m, and in the axial zone - less than 1000-2000 m. The width of the ES within the limits of the isobath 4500 m reaches about 1200 miles, with the full width of the ocean basin between submarine continental margins of Africa and South America in this area being about 1500 miles.

The relatively high position of the ES is well shown both on the bathymetric maps (GEBCO, 1984) and on the map of the gravity field, based on satellite altimetry (Haxby, 1988, Sandvell, Smith, 1996). To its northeast in the Cape Verde basin the depths are over 5000 m, and to the west in the Guiana basin - over 4800 m. Similarly, to the east and southeast of the ES in the Sierra Leone and Guinea basins, and in the Brazil basin to the west and southwest, the depths are over 5000 m. The ES, with its depths under 4500 m, therefore thanks to its relatively upraised location and large width in relation to the whole width of that part of the ocean, acquires the significance of a massive but not so high morphostructural barrier or intercontinental threshold, which separates the North and the South Atlantic. Here it is possible to compare the ES with the Icelandic segment of the North-Atlantic Ridge, which almost completely partitions off the ocean and separates the North Atlantic from the Norway-Greenland basin.

In its morphology the ES, even in the most general features differs notably from adjacent segments of the MAR. There the ridge represents a wide rise with a well-expressed crest, with a more or less symmetrical V-shaped profile, and with certain slopes of flanks from the axial rift to the floor of the wide oceanic basins on both sides of the ridge. The total fall of slopes there is about 3000-3500 m., while the flanks of the ES have a very small fall and are close to being almost horizontal, with a total fall to the floor of adjacent basins of only a few hundred meters. The flank plateaus of the ES spread to the east and west almost to the foot of the continental slopes of Africa and South America, from which they are separated by a narrow tongue of the basins of Sierra Leone in the east and Guyana in the west (Plates 1, 3,4, 5).

The transition zone between the ES and the submarine continental margin of the West Africa is represented by the rather narrow Sierra Leone basin. The floor of the Sierra Leone basin to the east of the ES is contoured with isobath 4500 m. It is smoothed with sediments about 1000 m thick, and represents the Sierra Leone abyssal plain with a depth of up to 5000 m. Drill hole 13 of DSDP penetrates only into the upper part of this sedimentary layer. The character of change in the sediments discovered with depth testifies to the deepening of the Sierra Leone basin through the postcampanian age. The stratification of the sedimentary basement in the Sierra Leone basin and low seismic velocities in the “third layer” - 6,2-6,7 km/see - correspond to velocities in the granite-metamorphic basement of the Sahara plate (Plates. 17, 18). This allows us to assume that the basin originated as a result of the submergence of the margin of the African continents along the system of stepped faults in conditions of regional extension (Volkov, Gagelgants et al, 1981), similar to progressive “oceanization” in the Gulf of Guinea (Rosendahl, Meyers, Groschel, Scott, 1992).

The submarine continental margin of Africa on the eastern side of the Equatorial Atlantic is represented by peripheral parts of the structural systems of the Sahara plate of the African-Arabian platform. To the west of Abidjan, approaching the shore, is the margin of the Precambrian Leone-Liberian massif, fringed by the southern end of the Mauritano-Senegal system of the Baikalian folded belt. The south-west branch of that system corresponds to the ledge of continent where this branch is cut by the shore at a right angle. The margin of the continent is chopped off along the shore of Cote d’Ivoire by a steep fauled scarp, associated with the Saint Paul transoceanic fracture zone, prograding deep into the continental margin (Fairhead, Green, 1989). This scarp continues further to the west along the southern edge of the triangle protrusion of the Liberian submarine marginal plateau.

The fractures of lower categories with NW-SE lineations, parallel to the shore of Guinea-Bissau, and of N-S lineation, parallel to the shore of Senegal, determine the fracturing of the west margin of the African continent into a system of step faults, highly typical of the entire Atlantic zone of the periphery of Africa. One such step is the Liberian marginal plateau, subdivided by the fault into two terraces, with depths of about 900 and 1800 m respectively. The surfaces of the terraces are inclined to the west. Due to the presence of these terraces, the south-west continental slope of Liberia is not as steep as the faulted southern slope. The continental slope along the shore of Guinea-Bissau has a similar structure, where the system of stepped faults led to a submergence to a depth of 500-1000 m of the marginal Guinean plateau, in which the basement probably continues the south-west branch of the Mauritano-Senegal folded system. The northern boundary of the Guinean plateau, similarly to the northern edge of the Sierra Leone Rise,is associated with prograding into the continental margin of the fracture zones of Cape Verde, Doldrums and others, and united into a cluster of fractures of the northern boundary of the ES.

Drilling on the continental shelf of Nigeria opened up a mighty complex of sedimentary rock from the lower Devonian, covered with basalts of the Jurassic - early Cretaceus age and on them from the Cretaceus until the recent past, with total thickness attaining 4000-5000 m. Judging by the data of seismic profiling, a similar structure is characteristic of the continental margin of the Equatorial Atlantic, also farther to the north-west (Tucholke, Uchupi, 1989- 1990). On the shelf of Sierra Leone and Guinea-Bissau the thickness of the sedimentary body reaches 2000-3000 m, and on the submarine marginal Guinean plateau 6000 m of sedimentary prism is compiled below by a dislocated complex of sediments of pre-Cretacious age and in the upper part by a nondislocated upper Cretaceus complex (Plate 18). The Cenozoic rocks are absent in the section and probably were eroded (Geology of the continental margins. 1978). In the continental apron the thickness of the sedimentary body reaches more than 3000 m here. (Tucholke, Uchupi, 1989-1990).

To the west of the ES is the relatively narrow Guiana basin, elongated in the NW-SE direction, which serves as a transition zone between the ES and the continental margin of South America. In the north-western part, to the south of Vemadsky f.z. and of the Sierra Leone f.z., but to the north of the Strakhov f.z. (the Fourth North f.z.) the depths of the basin above the abyssal plain Para are about 4600-4500 m. To the south of the Strakhov f.z. depths of about 4400-4300 m prevail and only to the south of Saint Paul f.z. do they increase to 4500 m in a small basin, bounded by the transverse rise along the western branch of this fracture zone. This transverse rise with a depth above it of less than 4000 m, is continued to the west in the form of the Parnaiba ridge. Bearing in mind that to the north of the Sierra Leone f.z. and to the south of the Sao Paulo f.z. the depths of the oceanic basins are more than 5000 m, the floor of the Guiana basin in the area of the Equatorial Atlantic, with its depth of under 4500 m, can be considered as part of a morphostructural barrier between the North American and Brazil basins.

The accumulative bodies of the abyssal plains of the Guiana basin - Demerara, Seara and Para abyssal plains - are fed by terrigenous material, transported by turbidity currents. Its thickness reaches 200-1000 m.(Tucholke, Uchupi, 1989-1990). The sedimentary cover disguises partly or completely the topography of the basement, and its features such as marginal grabens and horsts in the system of pericratonic stepped faults and transverse ridges along transoceanic fracture zones, which impact on the directions of the flow of turbidity currents. This impact was expressed very vividly by the transverse ridges Seara and Belem (Sao Paulo f.z.), Parnaiba (Romanche f.z.) and Fernando di Noronja (Chain or Charco f.z.) ..

The changes through time in the morphostructural plan of this area, e.g., the submergence of the Seara ridge in Paleocene-Miocene, led to migration of the channels of turbidity currents, such as the movement of the channel of the Equatorial mid-oceanic canyon in Miocene from the Pernambuku abyssal plain (south of the Romanche f.z.) to the Seara abyssal plain (north of that f.z.) (Gorini, 1981). The growth of the Amazon cone in the northern direction to the Demerara abyssal plain is very notable, though not in the NE direction, at a right angle to the continental slope or along transverse ridges of the Sao Paulo f.z in the eastern direction. The probable reason for this could be the higher position prior to Miocene of the Seara ridge and the ES in general.

The correspondence of the northern limit of the Aptian deposits of salt in Brazil and Africa in the South Atlantic to the south boundary of the ES, and the similar correspondence of the southern limit of the evaporates of the North Atlantic with the northern boundary of this segment, allow us to suppose that the margins of both these continents were then parts of a morphostructural barrier which limited the areas of accumulation of evaporates in shallow-water areas during the early stage of the history of the ocean.

To the west of the ES the continental boundary of the ocean is represented by the northern part of the Guyana-Brazil shield of South America. The Precambrian basement of this shield is exposed over a wide area, and is covered with sediments in small areas only, and along the shore of the ocean - with basalt traps (Khain, 1971). The shore line, shelf margin and continental slope have broken contours, reflecting a combination of NE-SW lineations of early Paleozoic fractures of the shield, and NW-SE and N-S lineations of pericratonic faults of early Cretaceus activation, which determined the present shape of the general contour of the continental margin, and W-E lineation of two clusters of transoceanic fracture zones, which limit the ES and prograde into the continent at acute angles. The complex combination of such fractures and faults, the horsts and grabens associated with them which act as sedimentary traps, and a number of volcanic mountains along the fractures determine the features of the topography of the continental margin.

The prograding of the transoceanic fracture zones into the margin of South America is marked by volcanic activity unique to the Brazilian shield. The outpouring in the Miocene of lavas of ancaranite, nefeline basanites and olivine basalts with xenolites of peridotite (spinel lercolites and harcburgites) came from a depth of about 60 km (Gorini, 1981). Thus, for example, the western continuation of the Sharco f.z. is traced in the shield along the Kabugi lineament through Paleogenic-Neogenic volcanism, the continuation of the Chain f.z. - in the submarine ridge Fernando de Noronha capped by volcanoes formed with nefeline basalts and through the outpouring of fonolites of Mesejaki on the shore near Fortaleza. The Fernando de Noronha ridge restricts the northern edge of the marginal plateau of the continental slope Rio-Grande-de-Norte. There was a coral reef at the eastern edge of this plateau on the submerging early-Paleozoic basement. In the sedimentary trap to the west of the reef a kilometer-thick pile of sediments was accumulated. To the north of Fortaleza the western branch of the Romanche f.z. is marked on the continental slope by the Parnaiba submarine ridge; to the north of this is the Belem ridge, associated with the western branch of the Sao Paulo f.z. These submarine ridges act as structural barriers, against which powerful sedimentary bodies were formed in the Fernando-de-Noronha (to the south of Parnaiba ridge) and Para-Maranjao (to the south of the Belem ridge) basins.

The protraction of fracture zones into the continent is marked also by the creation of a syneclise of the land formed on the Precambrian basement. Accumulation of sediment began there in Cenomanian and was most intensive after early Miocene. The thickness of sediments there reaches 3-6 km.

The Amazon cone is another thick sedimentary body - up to 10 km - on the continental slope in the Equatorial Atlantic. It has the shape of a rounded ledge of the slope, contoured with isobath 4800 m, protruding ahead of the nearly straight margin of the shelf into the ocean to about 400 miles. The submarine Amazon canyon deeply cuts into the upper part of the cone, and is replaced in its lower part by a branched system of channels of turbidity currents, bordered with levees and meandered (Damuth et al., 1983). The sedimentary prism of the Amazon cone and Amazon shelf serve as a continuation of the sedimentary basin of the transcontinental Amazon syneclise - graben Maraju. That graben was laid at the end of the Proterozoic or at the early Paleozoic, but was most actively developed in the Cretaceus and Cenozoic. It is filled with marine shallow water sediments from silur to carbon, overlapped with Jurassic and early Cretaceous,and overlaid with continental sediments. The association of this graben with the western end of the Romanche and Sao Paulo transoceanic fracture zones is very likely. For a long period of time the transportation of sediments from this continental basin to the ocean was limited by the existence of a structural barrier at the outer shelf In the mid-Miocene only the sedimentary trap was completely filled and after that terrigenous sediments began to move to the outer shelf and slope, forming the cone, which is only 22 mln years old (Kumar, 1978). The youth of the cone can probably also be explained by the rather recent turn of the Amazon river to the east instead of west after the rise of the Andean mountains in the early Miocene (Damuth, Kumar, 1975), or by a similar recent intensification of erosion due to the rapid drop in the oceanic level, stimulated by the beginning of glaciation of the Antarctic continent (Hays, Frakes, 1975).

The ledge of the continental slope to the north-west of the Amazon shelf and the cone in the area between 52 and 55’ W - the marginal plateau Guiana (or Demerara) - probably correspond to the northern ledge of the Guiana shield, limited in the west by the Takutu graben, associated with the transcontinental Pisco-Juruja fracture zone. The northern and southern limitations of that plateau correspond to pericratonic faults or to western ends of the transoceanic fracture zones. Holes 143 and 144 of the DSDP demonstrate that the plateau is covered with shallow water sediments of the Jurassic and early Cretaceus epicontinental sea covered with younger sediments of the open sea at a depth close to the plateau’s recent depth (Hayes et al., 1972). Judging by that, the basement of this plateau has been submerged in the post-albian time with an amplitude of about 4200-4400 m. The Guiana shelf basin, filled with 2000 m of sediments, corresponds to the offshore continuation of the graben Takutu. The shallow water sediments here cover Jurassic basalts and are interlayered with evaporates.

The correspondence of the limits of distribution of the Aptian deposits of salt on the continental margins of Brazil - to the south and to the north of the location of transoceanic fracture zones, bordering on the ES -to the similar limitation of evaporates at the continental margin of Africa, allow us to assume the former connection of the margins of both continents or its association with the morphostructural continental bridge, including the ES, which, like a barrier, separated the shallow-water epicontinental seas of the young ocean on the south and north sides of that barrier in the Aptian time, and thus determined the deposition of evaporates only to the north and to the south of the equatorial Atlantic.

The main features of the morphology of the ocean floor of the Equatorial Atlantic create an impression of an area which formerly united the African and South American continents, and served as a barrier between the North and South Atlantic. Their connection was made possible by the further fracturisation of that barrier in conditions of a general stretching (or extension), and by its submergence and the beginning of rifting, localized in the recent axial rift zone of the ES.An alternative explanation of the history of the connection between the North and South Atlantic in the framework of the popular theory of plate tectonics was given by J. Jones (1987) and in several versions by E, Bonatti (1990). We prefer our own explanation as more appropriate to the data observed during our expeditions, and we shall try to base this explanation on the following descriptions of our data. Our explanation leads us to the hypothesis of the development of the ES and Equatorial Atlantic as a whole as a result of mantle diapirism in conditions of a very moderate expansion of the Globe (Plate 3).

Chapter 2.

Topography of the Equatorial Segment of the Mid-Atlantic Ridge taller Multibeam Echosounding.

G. B. Udintsev, J.K. Hall, V. G. Udintsev, A.B. Knjazev

The introduction of a new tool - multibeam echosounders (Renard, Allenou, 1979, Udintsev, Odinokov, Golod 1987) into the practice of marine geological investigations proved highly successful in studying the topography and structure of the ocean floor. Multibeam echosounders were especially efficient in studies of the complex morphostructure of the midoceanic ridges, which were previously imaged on bathymetric charts in a rather simplified and general manner due to the lack of information accessible to ordinary singlebeam echosounders.

In the last two decades investigations of the mid-oceanic ridges with the use of multibeam echosounders, in combination with the entire complex of marine geological-geophysical research, has yielded a plethora of information on the image and structure of these significant geostructures of the oceanic areas of the Earth. One of the most important features of their structure turns out to be segmentation, i.e. subdivision into block-segments separated to a greater or lesser degree.

After confirmation of the planetary unity of the system of midoceanic ridges (Ewing, Heezen, 1960), the first years of studies already shed some light on the existence of segmentation of the ridges due to their dissection with numerous cross-setting fractures. Tuso Wilson applied to such fractures the genetic determination: transformation faults, i.e. providing transformation-adjustment through the displacement along them of segments whose spreading axis is scattered in space according to the contours of the drifting lithospheric plates. During the process of detailed studies the discovery was made of the multiformity of the morphostructures of the cross-setting fractures, both in their length and configuration and in their orientation in relation to the axis of the rift. Consequently a certain hierarchy became visible of both the fractures and the segments which they separated (Searle, 1986, Macdonald et al., 1988, Udintsev, 1987). Another step toward a more accurate image of the segmentation of the mid-oceanic ridges appeared in the form of results of the very detailed survey using multibeam echosounding. One very important feature of the structure of the rift zones proves to be the existence of discontinuity without lateral displacements, without any transformation faults. The explanation of segmentation caused for this reason was given in space-time variations in interrelations between magmatic and tectonic activity in the rift zone (e. g.. Pockalny, Detrick, Fox, 1988, Fox, Grindlay, Macdonald, 1991, Zonenshain, Kuzmin, Lisitsin, Bogdanov, Baranov, 1989, Udintsev, 1989). Such variations can be related to a difference in the thermal and geochemical regime of the Earth’s upper mantle, whose upwelling (or mantle diapirisrn) determines the character of the tectonic-magmatic activity in the crust of the rift zone.

Thanks to the new method of research, the approach to studies of previously discovered anomalies in age and genesis among segments of the mid-oceanic ridges became easier (Bonatti, Honnorez, 1971). The possibility appeared for the discovery of evidence of kinematic inertia of some blocks (Bonatti, Chermack, 1981,), and of signs of migration in space and time of the axis of spreading and corresponding transformation faults (Bonatti, Crain, 1984, Bonatti, Ligi, Gasperini et al., 1994). From our point of view, in the segmentation of the mid-oceanic ridges signs of age and generic anomalies deserve special attention. The studies of such anomalies can throw some light on the more complex genesis of these ridges in comparison with the idealized scheme of their development in the process of conveyer type spreading. It in turn turn will help in the understanding of the reasons for the heterogeneity of the floor observed in some areas of the World Ocean.

The interest in anomalies in the segmentation of the mid-oceanic ridges stimulated the organization by the Geological Institute of the USSR Academy of Sciences (now Russian Academy of Sciences) from 1987-1991 of three expeditions with r/v “Akademik Nikolaj Strakhov” for complex geological-geophysical research of the Equatorial segment (ES) of the Mid-Atlantic Ridge (MAR) under a joint international program of the UNESCO Intergovernmental Oceanographic Commission (IOC), Program “EQUARIDGE” (an abbreviation of the words Equatorial Ridge). Some additional surveys of the rift zone of ES were made during the 2l th cruise of r/v “Akademik Boris Petrov.”

The ES, bordering in the North on the cluster of fracture zones - Vema, Cape Verde, Doldrums, Vernadsky and Sierra Leone, and in the South on a similar cluster of fracture zones - Saint Paul, Romanche and Chain, attracts the attention of many scientists due to the position thereof the tiny island the Rocks of Saint Peter and Paul, where ultrabasic rocks of the Earth's mantle of an unusually ancient age for the axial zone of the MAR age (Hart, 1964) were brought to the Earth's surface, and a petrography type analogous to the continental type (Bonatti, 1990). Our interest in the ES was also stimulated by the results of previous investigations of the Sierra Leone Rise (1981) and by the results of research at cros-sections of the ES at the 2nd cruiese of r/v "Akademik Boris Petrov" of the Vemadsky Institute of Geochemistry and Analytical Chemistry (1985). These results made it possible to assume the presence on the ES of an anomalously high thickness of sedimentary cover and for the first time revealed the overlapping of the transformation faults on the riftogenous structures of the ridge, unknown to us at that time for other segments of the MAR (Udintsev, Odinokov, Golod, 1987).,

The first of these above mentioned expeditions - the 7th cruise of the r/v "Akademik Nikolaj Strakhov" - took place in 1987, the second -11th cruise - in 1990, and the third - 12th cruise - in 1991. The the first author of this paper was the chief scientist for all three cruises, and at the Geological Institute P. P. Timofeev was the scientific head for the 7th and 12th cruises, and Ju. M. Pusharovsky - for the 11th cruise. A rather limited but important survey of the rift zone was done in 1995 during the 21st cruise of r/v''Akademik Boris Petrov"

During these expeditions detailed geophysical surveying was carried out including multibeam echosounding. The "Hollming-625" 15-beam echosounder was used. During the 7th cruise the main part of the survey of the ES was conducted with the use of the system of latitudinally lineated recognition tracks for the space of 300-400 miles and only partly for three polygons limited in space to about 30x60 miles. During the 11th and 12th cruises, on the other hand, the surveys were mostly in the form of successive polygon surveys with a total coverage of 18 of such polygons in the area between 1'N to 61N and between 25°W to 35°W. The parallel tracks of detail survey were positioned with spacing at an interval of about 1,5-2,0 miles at 900 fan of sounding beams, and at depths of about 1500-3000 m. It supplied practically 70% coverage of the floor. The area of such complete covering with echosounding was connected to the east of 25°W to the area of the rather detailed survey with singlebeam echosounding carried out by our group in 1980 with r/v "Ivan Kireev" at the Sierra Leone Rise and connected to the western margin of the African continent with through-passage tracks (Gordin, 1985, Efimov, Beresnev, 1989-1990). On the west the area of our detailed survey at the ES was connected with 12 through-passage tracks to the Brazilian harbors of Fortaleza and Recife to the continental margin of South America. In this way a rather full coverage was achieved providing information on the topography of the whole space of the ES MAR and adjacent areas of the Guinea-Brazil geotraverse.

In comparison to surveys of other segments of the MAR, carried out in recent years with the use of multibeam echosounders and covering mainly the axial part of the MAR without any large parts of its flanks, the detailed survey under "EQUARIDGE" Program covered both the axial zone and the flanks of the ES MAR (Plates 1, 2, 3, 19).

The navigation for the survey during the 7th cruise was arranged by the satellite system "Transit" and errors in position were evaluated at plus-minus 150 m, and during the 11th and 12th cruises - with the satellite navigation "NAVSTAR" Global Positioning System. The errors in positions were evaluated at plus-minus 50m.

Bathymetric survey in all three cruises was accompanied by continuous single channel seismic profiling, on the 7th and 12th cruises - by a magnetic survey, and on the 11th cruise - by a gravity survey.

Sufficiently dense coverage of the aquatoria of the ES MAR made it possible for the first time in the practice of our group to make a serious effort towards computer compilation of the bathymetric maps on the basis of a digital terrain model. In the course of this work active computer compilation was accompanied by following manual compilation when the computer was used only for compilation of the mosaic of tracks with bathymetry compiled in the limits of the swath of the 15-beam sounding, and interpolation of the isobaths between the stripes with computer compiled bathymetry were based on the manual drawing of isolines, bearing in mind the regularity of topographic images on profiles and certain ideas about the origin of the topography of the Ocean Floor. This was considered interesting for evaluation of the accuracy of computer compilation with manual compilation using a method, introduced by our group in the past for compilation of the bathymetric map of the Pacific Ocean (Udintsev et al, 1963, Udintsev, 1972) and maps of some other regions. We called the method a method of geomorphologic interpolation. Virtually the same method was developed independently by several groups of marine geologists in Britain, USA, France, Germany and Japan. These groups united to compile the 5th edition of the international "General Bathymetric Chart of the Oceans" (GEBCO, 1984).

The results of a comparison of the maps compiled in different ways lead the authors of this paper to evaluate the difference as rather insignificant for such maps on a scale of 1:250 000, and particularly insignificant on scales of 1:1 mln and smaller. After manual editing, the maps definitely look more impressive in the geomorphological sense, giving more details of topography, available to geomorphologists from profiles, and synthesized with all other geological information for images of the bottom topography, contoured with manually drawn isolines. But such manual drawing takes much more time. Computer compilation of a map for one polygon of 30x60 miles takes only a few hours and is very helpful in active planning of the geological and geophysical experiments based on bathymetry. The manual compilation of the bathymetry for the same polygon takes several days, and therefore the map cannot be used as actively. With maps of 1:500 000 (e.g. Plate 6, 7) and smaller scales, the difference between maps compiled by different methods is negligible and does not merit any serious attention. It is clear that all this is true only for detail surveys with coverage of about 70% and higher.

Computer compilation of the bathymetric maps of the ES was done by V.G. Udintsev, J.Hall and A, M. Sazonov with software prepared by them jointly and later improved on by V.G. Udintsev. In this paper we apply to the description of the topography of the ES MAR a map, compiled on a scale 1:1 mln and slightly reduced during photocopying and printing (Plates 1, 4).

In the limits of the studied area three main categories of the morphostructures (or provinces) can be distinguished the axial rifl zone of the ES, its flanks, and the fracture zones (f.z.) crossing the first two, The morphostructure of the rift zone is formed by the axial rift valley and the rift crests or ridges framing it. The flanks of the segment are represented by relatively gentle slopes, complicated by a number of crests, smaller in height and length than the crests of the rift zone. The transverse fractures, cross-setting the ES, are the number of troughs with different depth and width and are at different angles to the structures of the rift zone (Plates 5, 10).

<u>The axial rift zone</u> of the ES to the north of the Strakhov f.z., crossing it approximately at 4°N, is lineated in the direction close to the meridian and lies in a stripe about 50 miles wide from 32°20'W to 33°10'W. The axis of the rift valley on the northern margin of the studied area lies at 32°43'W and on the trans-section with the Strakhov f.z. - at 32°35'W. The floor of the rift valley is contoured with isobath 3400 m and is 5-6 miles wide. The floor of the rift valley has a cell structure resulting from the complexity of configuration of slopes of the valley and is subdivided into three tiny basins with depths of up to 4000 m. The ledges of slopes create between these basins some thresholds with depths of less than 3600 m. At the trans-section of the rift zone with the Strakhov f.z. the depth in the so-called nodal basin rapidly increases to 5348 m.

The ridges bordering on the rift valley to the north of the Strakhov f.z. are composed of three pairs of blocks, positioned in echelon with a gradual displacement to the east. These blocks are contoured with isobath 3000 m, and their width inside of this isobath is about 20-25 miles. The summits of blocks arise to depths less than 2000 m, but in the saddles between blocks the depths increase to 2800-2900 m. At the margin of the Strakhov f.z. the crests of the rift ridges are the highest to the depth of 852m at the western block, titled Nadejda (Hope) seamount and to the depth of 1750 m at the eastern block for which we propose the name Muratov seamount. The summits of the rift blocks rise above the floor of the valley to 2000-3000 m and above the flanks adjacent at the west and east - about 1000-1800 m.

To the south of the Strakhov f.z. the rift zones are displaced to the east to 60 miles, and its general lineation is close to 340°-160° at the section before discontinuity marked approximately at 2°40'N and close to 345°-165° to the south of that discontinuity, and until the Saint Paul f.z., which transects the rift zone approximately at 0°45'N.

The rift zone in these parts of the ES has a width of about 50 miles in the north and 60 miles in the south. Near the Strakhov f.z, it occupies the stripe from 31°10'W to 32°00'W and near the Sao

Paulo f.z. - the stripe from 30°00'W to 30°55’W, The axis of the rift valley at the transection with Strakhov f.z. lies at 31°35’W and at transection with Saint Paul f.z. - at 30°20'W.

The floor of the rift valley in these parts of the ES is contoured with isobaths 3000-3200 m, and its width within these limits is 5-7 miles. Thanks to complex configuration of the slopes of the valley, determined mainly by en echelon position of ridges which border on the valley, the floor of the valley has a cell morphology and is subdivided into a series of tiny basins with depths of up to 3000 m at thresholds, and the depth between them is less than 3200 m. The greatest depth in the sedimentaryless nodal basins at intersections with Strakhov f.z. in the north is 5113 m, and at discontinuity at 2°40'N - more than 4000 m. To the south of that discontinuity in a number of places inside the valley narrow interior ridges are observed, supposedly neovolcanic rises, lineated along the axis and raised to 200-300 above the adjacent floor.

The rift ridges on both sides of the valley have block structure. The size of the blocks in comparison with those observed to the north of the Strakhov f.z. is markedly prominent their width is up to 30 miles and their length is up to 45 miles. The summits of blocks rise to depths less than 2000m and in places - to 1800m. The tallest summits of the so-called corner rise on the western ridge near the margin of the Strakhov f.z. at the area of rift-transform intersection - 1800 m - and the summits of blocks of the western, and eastern ridges near the discontinuity at 2°40'N (about 1100 m) and the summit of the western ridge at 1°50'N (less than 1800 m).In the saddles of crests between the raised blocks the depth increased to 2800 m. The elevation of crests above the floor of the valley varied from 600 to 2000 m., and above adjacent flanks in the west and east - from 400 to 1800 m. Their en echelon positioning with gradual displacement to the east, accompanied by subdivision of the rift valley into a number of tiny basins, is well observed in the location of the blocks (Plate 1,3,6,7, 10).

The discontinuity of the elementary structures, their en echelon location and accompanying segmentation of the rift zone, observed in the ES, are characteristic in general of the global system of the rifts of the midoceanic ridges (Macdonald et al., 1988). Here we observe forms of such discontinuity of all four categories of hierarchy, distinguished by Macdonald and his co-authors: from the first - discontinuity with lateral displacements in hundreds and tenths of km along transform faults (Saint Paul and Strakhov f.z.), the second - with displacements of the order of 3-5 km without direct morphological connection with transform faults (in the area 2°40'N) and, finally, to the third and the fourth category - with tiny (less than 3-5 km) and negligible (less than 0,5 km) displacements, often in the form of en echelon overlapping of the ends of elementary structure (e.g. at 4°30'N, 3°30'N, 3°10'N, 1°40'N and 1°20'N). Here, the morphostructure of the rift zone of the ES corresponds to the general regularity of the structure of the oceanic rifts as a whole and to the rifts of the Atlantic Ocean, and is not anomalous. Judging by the weak expression in the topography of the floor of the rift valley of the interior axial ridges, interpreted usually as neovolcanic ridges, the dominant role in the recent development of the rift zone of the ES belongs to vertical tectonic movements, while lateral displacements (or extensional displacements) play a smaller and subordinated role

Passing from the axial rift zone to the flanks of the ES, we observe essential changes in the general character of the morphostructure. First of all the extremely gentle general sloping of their surface is noticeable. If the slope of the slopes of the rift ridges, bordering on the rift valley, have a steepness of up to 20III and more, the flanks of the ES on the whole are characterized by a general slope only 1III and can be called flank plateaus (Plate 5). We make a reservation here because within the limits of these generally gently sloping surfaces exist numerous raised blocks, whose slopes are quite steep. However these rises do not determine the general image of the morphology of the flank of the ES, and only slightly, in terms of their details, complicate the almost plain and very gentle sloping surfaces which make up most of the space of the flanks.

Judging by their morphology, the above-mentioned rises differ substantively from the ridges of the axial rift zone: they are rather narrow horst crests with a width at the foot of 3 to 7 miles and a length of up to 7-15 miles, heights about 600-800 m, rarely - to 1000 m. Their lineations in many places comese close to NW-SE, and differ from that typical of rift structures, which are close to meridional. The predomination in the space of the flanks of wide plain surfaces over narrow horst-crests creates an image of an alternation of the wide grabens with narrow horsts. However, between horst structures at flank plateaus volcanic features are also observed. One the most remarkable is positioned at 2°17'N and 28°40'W. Its flat top attains a depth of 914 m, and its height is about 2000 m. For this seamount we propose the name Vinogradov seamount (in the memory of the outstanding Russian specialist in marine geochemistry A.P.Vinogradov).

Judging by the character of deformation of the sedimentary cover, observed in the records of continuous seismic profiling, the development of the morphostructural complex of the flank plateaus occurred in conditions of regional extension under which narrow horsts happened to find themselves as the remains of subsidence. Undoubtedly, the plain character of the surface between the horsts is aggravated by the presence of a rather thick sedimentary cover, but it reflects nevertheless the main structural peculiarities of the flanks basement (Plates 8,9, 10).

The structure of the flank plateaus is significantly complicated by the presence along the margins of the axial rift zone of a number of huge block rises and separating troughs. We have called the system of such structures a zone of marginal dislocation. For the first time we focused attention on the existence of dislocations of such type at the Angola-Brazil geotraverse, and associated their origin with the action of the force of compression directed from the side of the South-Atlantic Ridge to the floor of adjacent basins - the Angola and Brazil basins (Udintsev, Beresnev, Gordin, 1980). In the case of the ES MAR it is logical also to associate the zones of such types of dislocation with the results of the action of the force of compression directed from the axial rift zone to the margins of the flank plateaus. We see these marginal dislocations as a combination of horsts and grabens, which originated in conditions of such compression. This is confirmed by the character of deformations of the sedimentary cover, observed on the records of the continuous seismic profiling and with the samples of tectonic breccia dredged at these zones.

The transform faults or fracture zones are displayed in the topography of the ES MAR in different degrees of their morphostructural expression. The largest and most vividly expressed in the topography Saint Paul f.z. is at the same time one of the largest in the Equatorial Atlantic as a whole. Its length reaches 2700-2800 miles and the ends of it prograde into continental margins: in the east to the margin of Africa, forming the faulted boundary of the marginal Guinean Plateau and northern boundary of the western part of the Guinean Bay, and in the west - to the margin of South America, and there it probably is associated with the lineament of the Amazon river.

The Saint Paul f.z. was studied by us in the west out-of-rift branch only on several single tracks, crossing it on the way to Fortdeza and Recife harbors, in more detail in its northern part near the islet of Saint Peter and Paul Rocks (Is1. Sao Paulo) and in more or less substantive detail near the eastern rift-transform intersection (Plate 8). It is there that the complex character of the structure of this fracture zone was recognized, which is represented as a combination of four or perhaps five parallel graben-trenches with a width of each about 5 miles, separated by narrow transverse ridges and filled with sediments in different stages. The depths above the crests of the transverse ridges are less than 2000 m. The floor of the trenches descend from the north to the south like the steps of a staircase and their depths increase consequently from 3700 to 4200 m. However such step-like deepening of the floor of these grabens in the topography of their basement is not so acutely expressed - these depths increase from north to south only from 4600 to 4700 m. The step-like deepening of the floors of the trenches in the topography of the ocean floor is more contrasted due to a rather sharp decrease of the thickness of the sedimentary filling of trenches - from 1000 m at the northern graben to 400 m at the southern, On the northern margin of the St.Paul f.z. in the area studied in detail is situated the large seamount we named the Belousov seamount. Its top (1°28'N, 24°58'W),which is surrounded with abrasional terraces, attains a depth of 623 m. (Agapova, 1994),

The greatest lateral displacement of the axis of the rift of the ES occurs along the most northern graben-fractures of the Saint Paul f.z. - from 30°25'W to 26°32'W, that is 230 miles to the east. However this displacement to the east continues for 95 miles more in the area between 26°32'W and 25°00'W, but in step-like form, through a complex of three-four graben-fractures separated by transversive blocks with short - less than 20 miles - rifts inside them. As a result of this, an axial rift to the south of Saint Paul f.z. occurs, positioned relatively to the axial rift of the ES at 25°0'W, that is displaced 325 miles to the east. It is worth noting the existence, at intersections of short rifts with trenches of graben-fractures, of sharply deepened nodal basins with depths of 5224,5222, 5120 and 4920 m, having no sedimentary tilling.

The second in size in the studied area of the ES is the Strakhov f.z. It crosses the axial rift zone at 3°52'N at the azimuth 84° when the general azimuth of the axis of the rift is 350° to the north of it and 340° to the south of it. The total length of the Strakhov f.z. is about 1000 miles. In the east it inprods into the Sierra Leone submarine rise, and in the west to the Seara submarine rise, without displacing the prograding into the submarine continental margins of Africa and South America.

Unlike the Saint Paul f.z., the Strakhov f.z. is represented in the topography as a single narrow graben-trough with V-shaped profiles, contoured with isobath 3000 m. Its width inside of this isobath is about 5-6 miles, and the depth of the floor is predominantly of the order of 4500 m in the between rift part and 400-4100 m in the out-of-rift branches (Udintsev, Agapova, Beresnev et al., 1995).

The axis of the rift to the south of Strakhov f.z. is located 60 miles to the east from its axis to the north of this f.z. The slopes of the trough in the between-rift part have a height of 2500-3500m and a steepness of 20°-25°. The topography of slopes is complex: in some places they are cut with deep hollows, crossing the floor of the trough and traced on the opposite slope (e.g.at 32°07’W). According to this the floor of the trough has a cell structure,being divided into a number of tiny basins, which to a very small extent are tilled with sediments. Their depth is about 4500 m. The greatest depths of the between-rift part - 5000 m and 5100 m - is associated with the above-mentioned nodal basins, deprived not only of sediments but also, according to seismic refraction data, having no so-called second and third seismic layers of the Earth’s crust. It is known that the nodal basins are the characteristic detail of the structure of all transform faults of the riftogenal mid-oceanic ridges of the World Ocean (Detrick, Purdy, 1980, Karson, Dick, 1983, Fox, Gallo, 1984, 1986, Macdonald et al., 1986, Pockalny, Detrick, Fox, 1988). These authors explain the origin of the nodal basins through variations through time in the volcanic activity in the rift. However an alternative explanation seems to us more probable, linked to the summing up of the vector of lateral spreading across the rift, and of the vector of extension along the axis of the mid-oceanic ridge, directed across the transform fault,

Along the between-rift and out-of-rift parts of the Strakhov f.z. there are no transverse ridges, observed along some other fracture zones. However along the margin of this fracture zone there are well expressed elevations of the rifted ridges transsected with it. In a less detailed survey such elevations could be interpreted as a transverse ridge associated with transform fault. The high comer rises at rift-transform intersections with rifted ridges of the axial rift zone are very notable, such as the above mentioned Nadejda and Muratov seamounts.

The out-of-rift western and eastern branches of the Strakhov f.z. are separated from the part between the rifts by peculiar structural thresholds. In the area 30°50'-31°30’W to the east of the eastern nodal basin and in the area 32°55’-33°25’W to the west of the western nodal basin the trough is narrowed to 2-3 miles and its floor is crossed with the ridges, which descend into the trough from the out-of-transform parts of the ES. The heights of these ridges are about 200-300m and they correspond to the structures of the above mentioned zones of the marginal dislocations, which lie along the joint of the axial rift zone with the flank plateaus. The depths of the trough at these places decrease to 3900-3800m, but then increase to the east and west of the axial rift zone to 4000-4100m.

With the distance from the between-rift part, the west and the east branches of the Strakhov f.z. gradually widened to 8-12 miles and in the westernmost part of the west branch -15 miles in the area 33°30'-34°10'W, where there is local lowering of the southern slope and the flank plateau itself The slopes of the trough in both out-of-rift branches are asymmetric: the northern are higher and steeper, have a more complex configuration with crests and hollows, and the southern are more gently sloped, lower and are more smoothed with sediment . The thickness of sediments, tilling and smoothing the floor of the trough, increased just after the coming from the thresholds, separating the out-of-rift branches from the between-rift part, to 600-800m. The trough has the form of almost U-shaped profiles. The depths reach the order of 4000-4100m. In spite of certain masking by sedimentary cover, in the topography of the slopes and the floor of trough the continuation from slope to slope via the floor of the number of crests and hollows, peculiar to the flank plateaus, is clearly traced.

To the south of the Strakhov f.z. lay one more fault notable in the topography of the ES, called the St. Peter f.z. It lies parallel to the Strakhov f.z. at azimuth 84°. The weell- studied western out-of-rift branch of the St. Peter f.z. is positioned approximately at 2°40’N, contoured with isobath 3800 m, and has a width of the trough about 10 miles at flat, sediment-filled floor. The depths of this floor increased from east to west from 3800 to 4000m. On approach to the axial rift zone at 2°50'N the trough is crossed with structures of the zone of marginal dislocations, but its continuation nevertheless is traced to the east until 31°40'W, where at 3°05'N it blindly prods to the west ridge of the axial rift zone. A special detailed survey of the 21th cruise of r/v "Akademik Boris Petrov" provided us with rather clear evidences that the St.Peter f.z. does not cross the rift zone. However it is possible to suppose that the eastern out-of-rift branch of the St.Peter f.z. corresponds to the area of depths of about 3700 m, observed on the eastern flank plateau in the strip between 3°00'N and 3°10'N. Scarcity of information for that part of the ES does not allow us to state this with confidence. The supposition about sharp displacement of the St. Peter f.z, to the south and crossing the rift zone at 2°40'N., leading to discontinuity of the rift structures with displacement of the axis of the rift to the east by 30 miles seems to us unrealistic. Discontinuity of the St. Peter f.z. probably is the result of the juxtaposition of the rift zone on the previously existed fracture zone,

To the south of the western out-of-rift branch of the St. Peter f.z, the topographically well expressed trough with a depth of over 3600m, lineated at azimuth 60,° attracts attention. It blindly prods to the western ridge of the axial rift zone approximately at 2°20'N. More to the south just near the Sao Paulo f.z. on the east flank plateau it is possible to suppose the existence of a fracture with lineation at azimuth also 60°, expressed in the topography by a depression with depths of 3400-3600 m. The existence of the fracture zone without expression in the topography can be assumed in the northern part of the western flank plateau at 4°30'N., on the base of expression only in anomalies of gravity field in Bouge reduction, reflecting the structure of the basement. This hypothetic fracture zone is lineated parallel to the Strakhov f.z. and is marked in the structure of the axial rift zone with some submergence of crests of the ridges. Its continuation can probably be traced, though not with such certainty on the east flank plateau.

In the general image describing the morphostructure of the ES the following features seemed to us important for the understanding of its origin: the obvious discontinuity of the rift structure, reflecting the phenomena of segmentation, associated with irregularities or heterogeneity of the mantle diapirism (Macdonald et al., 1988). Equally obvious is the overlapping of both rift structures on the structures of transform faults (e.g. St.Peter f.z.,, 4°30'N f.z., 2°40'N f.z., 1°20'N f.z.) and the overlapping of the transform fault structures on the rift structures on other occasions (e.g. The Strakhov f.z., The Saint Paul f.z.). That is considered by us as evidence of their genetic independence and creation in different space-time interrelations.

It seems to us possible to explain the origin of the axial rift zone structures mainly by extension (spreading) across the MAR, accompanied by vertical movements, and the origin of the structures of the transform faults with extensions along the axis of the MAR, accompanied by both the displacement and - to a smaller degree - by vertical movements along their margins. The ideas regarding such type of movements in the rift zones had been proposed previously (Francis, 1977, Lillwall, 1982, Zonenshain et al., 1989) as well as the ideas about the transform faults being extension rather than displacement structures (Larin, Solovyova, 1979, Udintsev, 1987, 1989, Udintsev, 1990).

Considering the interrelations between the morphostructures typical of the axial rift zone, the flank plateaus and the transform faults, and bearing in mind the difference in the age of the rocks of their basements and sedimentary cover (Udintsev, et al 1990), we hold to the following opinion concerning the development of the morphostructure of the ES and the adjacent basins of the Equatorial Atlantic through the following stages (Plate 3):

a) the general extension of the ancient continent with the creation of the net of latitudely lineated transoceanic fracture zones and more local fractures of NW-SE lineation, parallel to newly formed margins of separating continents, accompanied by the outpouring of a wide magmatic field - a large igneous province on the floor of the intercontinental sea and the beginning of the submergence of the floor of that sea,

b) The localization of tectonic-magmatic activity in the axial zone of the ES and the formation of its rifted structures with continuation parallel to it of the regional extension, and rather moderate submergence at the flank plateaus, and more progressive submergence in the oceanic basins. All of this is in line with today's image of the ocean floor.

Chapter 3.

Topography of the Rift Zone and Some Parts of the Fracture Zones in the Equatorial Segment

G.V. Agapova

Geomorphology of the Rift Zzone between the Strakhov and St. Paul fracture zones.

In the explored area the Mid-Atlantic Ridge has the outlines of a trapeze, the wide south base of which leans on the St. Paul fracture zone. The area has not been sufficiently studied. The existing bathymetric maps based on the same data - sheets 1:1 mln GEBCO and few tracks - represent the morphology in different ways. On the GEBCO sheet 5.12 (editor B. Heezen) the axial part of the Mid-Atlantic Ridge is presented as a system of five big angular blocks, separated by Sublatitude fracture zones. Along these zones the axis of the Ridge is displaced near 4°N and 2°45'N about 50 miles and near 2°45'- somewhat less. Some blocks have two rift valleys with depths near 3500 m.

On the GEBCO sheet 5.08 (editor A. Laugthon) the axial zone of the Mid-Atlantic Ridge is represented by an echelon series of ridges and depressions with sublongitude orientation. Depths of these depressions is not more than 4000 m. Tops of rift ridges are placed at depths of near 2000 m. The axis of the Ridge is almost not broken into segments. Only the configurations of isobaths provide a possibility for assuming the existence of tectonic dislocation near 2°40' and 1°50'N. On the English bathymetric chart scale 1:2400000 (editor P. Hunter) bottom topography is represented in greater detail than on 5.08 GEBCO. Extended rifts and depressions disposed by echelon are represented on this chart. A fracture zone is marked near 2°45'N, where the rift valley and ridges break off. All these maps are composed on the basis of ordinary echo sounding and may be inaccurate in the interpretation of the sizes and extensions of separate forms of topography. On the maps, based on the multibeam echosounding, interpretations of sizes, outlines and extension of separate forms are more exact, Maps of this area reveal the complex structure of the ridge part. Two main provinces are distinguished within the Ridge: that of bathymetric position, and that of the typical complex of bottom topography, One is the rift ridge province proper, and the other is the volcanic plateau province.

Rift ridge province is represented as a narrow, about 60-80 mile belt of contrast relief, consisting of ridges and depressions. The amplitude of depths reaches 1500-2000 m. The rift ridge province is placed on the volcanic plateau and is composed of big isometric blocks, the tops of which form complicated volcanic ridges. The crest of the ridge is in many cases placed at depths of 1000-1500 m. Extensions of individual ridges and depressions are from 0 to 30 miles. Crests of ridges have many peaks, slopes are steep, dissected by faults and step-like. Rift ridges are placed by echelon. Rift depressions are placed in a more complex manner. All depressions in the axial zone may be divided into two kinds axial rift depressions and inter-block depressions. There are two axial rift depressions in this segment of the Ridge. The depths of these depressions are 3500-3700 m. In the place of intersection of the rift valley and fracture zone trough (nodal deep) depths of depression increase to 5000m and more. The west rift depression is deeper than the east. Median elevations of 100 m high are placed on the bottom of the west rift valley. The bottom of this depression has a V-shaped cross profile and has no sediment covers. East depressions have smaller depths and are separated by basins with sediment cover 100-200 m thick. It is possible that the west depressions are younger and more active than the east depressions. Rift depressions on the limit of blocks are broken off or branched. On the orographic scheme (Plate 10.) one can see two zones near 2°20'-2°35'N and 3°05'-3°10'N near which are changed structural patterns. Part of sublongitude ridges and depressions near 3°05'-3°10'N are thinning and there appears a chain of sublatitude basins with a flat bottom. This chain approaches the axis of the ridge closely but does not intersect it. It is possible that this small fracture zone is buried by sediment cover. Another zone has a more complex structure. A sublatitude ridge appears along 2°25'N. This ridge extends for 20 miles, with a minimum depth of 720 m. The rift valley is divided into two branches as overlapping on the north of this ridge. This zone is represented by a small fracture zone. Similar fracture zones of small length have been marked in other places of the Mid-oceanic Ridge, where these structures are connected with the migration of active rifts and interruption of spreading processes(MacDonalds K.C. et al., 1988). Morphology of this fracture zone is connected with another rift form, allowing for the assumption that it appeared as a new zone. The name St. Peter was proposed and adopted by the GEBCO subcommittee for this zone (in memory of the ship which discovered the St.Paul and St. Peter islands).

A volcanic plateau occupies a vast area of the Mid-Atlantic Ridge and lies at a depth of 2500 to 3000 m. The modern relief of the plateau y presents an interleaving of elevations af acoustic basement with flat surface (Plate 5). Amplitudes of the topography typical of the axial zone decrease towards the foot of the ridge from 1000 to 500 m. This is connected both with the decrease in irregularity of basement and with flattening of the surface by sedimentation in depression of the relief Velocity of flattening of the surface plateau has a marked degree, so the thickness of sediments on the west flank is 600-1000 m and on the east flank 600-800 m (Plate 8). In the process of sedimentation on the flanks of the plateau basins are formed on different bathymetric levels. As a result of this stepped slopes of the plateau are formed. Flat or slightly sloping surface of steps constantly increases in size to the foot of the Ridge.

In contrast to the regularly oriented rift forms of the axial part of the Ridge, the relief of the volcanic plateau has isometric contours and irregular stretch. Typical size of elevations on the plateau are from 5 to 15 miles. Large seamounts of the plateau present a more considerable elevation of acoustic basement (Plate 9). The flat top of one such seamount situated at 2°17'N-28°40'W has a depth of 914 m. According to results of dredging the flat top of this seamount is covered by shallow water sediments.

The boundary between the axial part of the Ridge and volcanic plateau looks like a narrow zone of intensive broken relief. This zone is represented by series of blocks, which are raised or lowered on different bathymetry levels by vertical movements. Similar zones are named zones of marginal dislocations (Udintsev, Beresnev, Gordin, 1980, Odinokov, Udintsev, Beresnev, 1990).

Bottom topography of the active part of the Strakhov fracture zone.

The Strakhov fracture zone is transoceanic. On the east zone it is adjoined to Sierra-Leone Rise, on the west - to Seara Rise. The fracture zone has been first presented as a significance lineament on the GEBCO 5.08 sheet as a depression with extension for the 700 miles, and with the depth of the active part at less than 4000 meters. On the sheet 5.12 GEBCO the fracture zone has extenuation for 900 miles, The Strakhov fracture zone had not been studied in detail before the expedition of R/V "Academic N. Strakhov"(1988). In this expedition a multibeam survey was carried out in the active part the zone, and during the next expedition passive parts of the zone were studied. The survey of active parts was executed by a system of longitude tracks with each track 30 miles long and an interval between tracks of 3 miles. Bathymetric maps were prepared by G. Agapova to a scale of 1:250000 on the basis of the mosaic of a bathygram by hand. (Agapova, 1993) The depth is not correct in velocity of sound in sea water. Bathymetric maps reveal many important peculiarities of bottom topography, typical for many active parts of the fracture zone so the regional. Some of this morphologic peculiarity does not correspond to the standard scheme of spreading.

The active part of the Strakhov fracture zone has a deep linear depression kind of graben. The bottom of the depressions are 5 miles wide, more then 4000 meters deep. The extension of the active part is 60 miles, azimuth is 850. The bottom of the depression is divided by little ridges with a height of less than 300 meters into separate small basins with a length from to 0 miles. Nodal basins are placed on the intersection of the axial rifts and fracture trough, Basins have angular contours and depths of 5348 m in the west and 53 m in the east. On the bottom of nodal basins a sediment cover is absent, and also these basins lie on the lowest bathymetric level. This may be evidence of new genesis of basins or active processing of sediment layers.

A peculiarity of the morphology of the active part of the fracture zone is the superpositioned axial rift on the fracture depression. Rift ridges and valleys approach the edge of the depression from north and south, and are submerged on the slope of depression. Tops of some ridges are raised to 2000 meters depth and less. Volcanic cone has a top at 852 m. It is placed on a ridge, which limits the rift valley from the west side. The volcanic cone has been explored and named NADEJDA in honor of the ship of the Russian expedition of 1803-1806 in this area. The name has been adopted by the GEBCO Subcommittee of Geographical Names.

At a depth of approximately 3500 m, equal to the average level of the rift block,the fracture zone cuts across the crests and valleys of the upper part of the slope, Many steps and escarpments which attest to the vertical tectonic movements appear towards the lower part of the slope. The bottom of the fracture depression is narrowing in places of the submergence of crests and of the continuation of the valley. It is interesting that some depressions, which intersect the fracture zone trough, are traced for a long distance from the fracture zone, within the limits of the rift blocks. On the other side of the fracture depressions opposite the active rift another kind of valley is usually located. This valley may be regarded as a relict rift valley. The topography of active and relict rift valleys are different. In the Strakhov fracture zone the active part is deeper, with a complex of neovolcanic forms on the bottom. On the bottom of the active rift valley the sediment cover is thin and irregular, The relict rift valley is not so deep and is short, and its bottom is covered by 200-400 meters of sediments.

Large valleys cross the central part of the fracture zone along 30°08'W. These valleys are expressed clearly on the opposite side of the fracture zone. The valley cuts into the slope of fracture depression in 300-500 m. The same kind of depression, which intersects the fracture zone, was previously discovered by r/v “Akademik Boris Petrov” through the Sea-Beam survey in the Doldrams fracture zone (Udintsev et al., 1987), and later by r/v “Akademik Nikolaj Strakhov” in the Cape Verde fracture zone (Peive A.A. et al., 1989). A clear enough cross-section of valleys appeared in a detailed bathymetric map of the Kane fracture zone (Karson J.A.,Dick H.J. ,1983) which was used as a topography base for the tectonic model of fracture zone origin. It is possible to note the morphological similarity of the Kane and Strakhov fracture zones. Both zones have only one clear trough of the same extension, relict rift valleys opposite active rift valleys,and cross-section valleys in the central part of the depression. The existence of relict and cross-section valleys in this fracture zone make more difficult the interpretation of genesis of zones as a result solely of horizontal movements along fracture depressions. If the fracture depressions are formed as horizontal faulting, it is difficult to combine in space three valleys in three points. The modern position of active, relict and cross-section valleys may be explained by the spreading fracture depressions along the axis of the Mid-Atlantic Ridge. As a result of this movement all valleys are cut by depressions. The existence of gaping faults in fracture zones was noted by Pysharovsky J.M. (1989). Chebanenko I.I. (1983) gave a scheme of deformation, according to which in the latitude zone 0°-48°N and S spreading in the meridian direction prevail as well. This scheme goes along with the scheme of extensions of fracture zones in oceans (Agapova G. V., Volokitina L.P.,1991). An explanation of this morphological distinction in the position of valleys is possible on the basis of the concept of moderate spreading of the Earth. On the basis of the morphological similarity of rift valleys and fracture depressions Lain V.N. (1979) assumed that in mid-oceanic systems both lateral and longitudinal stretches of crust exist. The interaction of this process has a global character and plays a significant role in tectonic development of the system of rift-fracture zones. Lateral stretching is typical of the rift structures of mid-oceanic ridges. Longitudinal stretching is typical for the fracture zones where tension fractures increased gradually with depth and width. Joint action of lateral and longitudinal forces determine the origin of nodal basins in the intersection of rift valleys and fracture depressions (Udintsev, Agapova, 1990)..

Topogr aphy of the active part of the St. Paul fracture zone.

The St. Paul fracture zone belongs to transoceanic lineaments. This zone is limited in the south by the equatorial segment of the Mid-Atlantic Ridge and is traced from the continental margin of South America to the continental margin of Africa. On many existing maps the St. Paul fracture zone is represented schematically as a system of very long sublatitude ridges and depressions. On the maps of 1:1000000 scale, published by Gorini (1986) the complex structure of the zone is first reflected. These maps are usually baaed on echosounding with space tracks. In many cases the outlines of ridges and depressions are shown as more extended than they are in fact. On the 7th cruise “Akademik N. Strakhov” did a detailed Sea-Beam survey of the east active part of the St.Paul fracture zone between 24°40'-26°50'E and 0°20'- 1°40'N (Plate 8). On the basis of this data bathymetric map 1:250000 scale was compiled (Agapova, 1993).

The east active part of the St, Paul fracture zone represents a complex system in which short fragments of axial structure of Mid-Atlantic Ridge are separated by extension fracture depressions. The width of the zone is 80 miles. The cross-profile of the zone is asymmetric,and steeper to the north. Depths in depressions increase from north to south. Along the north flank of the zone is an extended ridge on which are located the San Pedro and St. Paul islands and separate high seamounts. One of the seamounts has been explored and named Beloussov seamount. This name was adopted by the GEBCO subcommittee in 1993, Seamounts have the form of a volcanic cone with a height of 2500 m and depth of top of 623 m. At a depth of 1300 m the slope of the seamount has a wide abrasion terrace. The formation of transverse ridges in the equatorial Atlantic (E. Bonatti, 1973) is connected with basalts rising to the surface during the early stage of spreading along the oldest fracture zone.

In the fracture zone depressions prevail at a depth of 4100-4200 m in the south and less than 4000 m in north of the system. In the north depression bottom is flat, sediment thickness is 800-1000 m. In the south of the depressions thickness of sediment is 500-600m and the bottom of depression is broken by a chain of little basins. Such a pattern maybe evidence of fault movements which were caused by spreading.

The east active part of the St.Paul fracture zone consists of short rift valleys in the submeridianal direction and sublatitude fragments of rift ridge, and the length of the rift valley in the area of detailed survey is no more than 20 miles, 3-5 miles wide, 4500 m deep. It is typical of many equatorial fracture zones that the length of rift valley is 2-3 time shorter than the length of adjoining fracture depressions (Agapova G. V., Volokitina L. P., 1991). The axial part of the Mid-Atlantic Ridge in this area is about 20 miles and consists of angular blocks. Higher blocks border the rift valleys. Volcanic massifs and separate volcanic cones are located on the tops of.

Nodal basins are location on each intersection of rift valleys and fracture depressions. Maximum depths of zone are connected with nodal depths of 5224, 5222, 5120 and 4920 m. Nodal depths have angular outlines, steep slopes and do not have a considerable cover of sediment on the bottom.

New data about the morphology of the St. Paul zone permit us to propose that the zone has many-step grabens which have been formed by a combination of vertical and horizontal fault movements. Prevailing movement in the genesis of fracture depressions may be gaping faults. The presence of gaping faults (stretching) in the fracture zone was pointed out by many scientists. Larin V.N.(1991) connects the origin in the fracture zone of mid-oceanic ridges with spreading along the axis of ridges. Stretching in the St. Paul fracture zone was started in the northern part and was then extended to the south. This proposition is confirmed by the smoother and older topography of the northern part of the fracture zone, and by the more contrasted and young topography of the southern part. Large distances between rift valleys may be connected both with lateral movements and with changes of the position of the spreading axis in the process of formation of the ritt-fracture system. The possibility of similar changes of the axis of the ridge is assumed by A.V. Il'in (1976) on the basis of morphological data. E. Bonatti (1987) considered the possibility of modification in the position of the spreading axis in the equatorial Atlantic. Some morphological evidence such as relict rift valleys and cross-section valleys in many fracture zones (Kane, Vima, Doldrams) confirm this assumption.

Chapter 4.

The Structure of Sedimentary Cover from Single-Channel Seismic Profiling Data.

A.F.Beresnev, V.N.Efimov, A.V.Koltsova, V.M.Golod, S.Yu.Sokolov, M.V.Zakharov

The goal of single channel seismic profiling (SCSP) during the 7, 11 and 12 cruises of R/V "Akademik Nikolaj Strakhov" was to carry out a detailed survey, discovering and mapping of sedimentary production. This survey (about 50,000 miles) resulted in the determination of numerous sedimentary bodies with various structures and thickness, big enough for thickness and acoustic basement map creation. Multibeam sounding bottom relief data and SCSP data on sediments and acoustic basement were processed together, which allowed for the recovery of lost sounding information in case of bad weather.

Besides the authors, SCSP actively involved in the survey were: Whitington R. (United Kingdom, 7 cr.),Poberzhin V.M.(11, 12 cr.), Erofeev S.A, (7, 11 cr.), Efimov P.N.(12 cr.), Abuev A.G. (7 cr.), Gladkikh P. A., Morozov Yu.I. (7, 11, 12 cr.). Outboard equipment and survey control were provided by Efimov V.N. Inboard electronic equipment for analog wavefield data acquisition was developed by Efremov V, N., digital acquisition and software was made by Sokolov S.Yu. in collaboration with Vanyan L.L. and Beresnev I.A.; SCSP and sounding data comparative processing software was developed by Ioffe A. I.; Koltsova A.V. compiled sediments thickness and acoustic basement relief maps and morphostructural scheme of the basement on the basis of bathymetry compuiled by hand by Agapova G.V. (7 cr.), Golod V. M. (11 cr.), Volokitina L.P. (12 cr.) to a scale of 1:250,000 and compiled by computer to a in scale of 1:250,000 and 1:1,000,000 by Udintsev V.G.

SCSP Survey Method and Technique.

SCSP has become a standard method for the investigation of large regions of the ocean bottom and was described by many authors (Ewing, Tirey, 1961, Hersey, 1963, Leenhardt, 1969). Here we present only a brief description of the technique used and survey specifications which are necessary for comparison of our data with other research.

Seismic signal radiation was provided by an array, consisting of two separate airguns (Beresnev et.al, 1988, Etimov, 1988), which were operated from both sides of the ship with synchronic shooting. These airguns are arranged as a deeply modified PAR BOLT prototype. High pressure air supply was also provided separately at an average value of 80 bars in 0.51 chambers. Operation depth was about 4-5 mat a distance of 1-2 m from the stern and 17-18 m between airguns. This performance of radiation system represents simple linear array.

The reception of seismic signals was provided by single channel seismic streamer, and its basic construction does not differ from that described by Saidov A. Yu. (Saidov, 1975). The streamer consists of main section 30 m long with 50 equally spaced receivers, head section 10 m long, which is a shock absorber, and the final stabilizing section, which was also 10 m long. The depth of operation was about 4-6 m, with the distance between the center of the main section and the source of seismic energy equal to 350 m. The acquisition of seismic data was made in the frequency band from 70 to 500 Hz. The survey had been undertaken at the speed of 9-12 knots. Firing moments were controlled by the timer, tuned to equal intervals of time. In most parts of the region they were equal to 8 s. In some cases with large values of the bottom depth and sediment thickness, it was increased to 12 s. Navigation fixing and correlation of SCSP and multibeam sounding data, containing geographical coordinates from Global Positional System receiver, was made by Greenwich Mean Time.

SCSP Data processing and Map Description.

SCSP data processing for map construction was carried out through the digitizing of seismic sections from line Sean recorder performance. The bed of sediments was outlined by the tops of the diffracted waves hyperbolas, because in most cases the acoustic basement could be considered a scattering surface. Seismic waves propagation velocity in sediments was assumed to be equal to 2 km/s. It was shown (Neprochnov, 1976) that the use of this value could lead in thickness determination to a relative mistake of no more than 15% in thickness range from 500 to 2500 m, which is quite normal for mapping. According to the general model (Semenov, 1990) in abyssal basins of the Atlantic ocean with pelagic character of deposits, sedimentary layer with 450 m thickness has a linear increase of velocity from 1.45 to 1.78 km/s. Having this value of velocity, the thickness could appear to be enlarged by 20-30%. It seems to us that having no velocity determination on each local structure and no chance of experimentally providing the model, mentioned above, the usage of 2 km/s value in the discovered region is admissible. Estimated at this value of velocity, the thickness of sediments in meters is numerically equal to the two-way travel time of seismic waves in mini-seconds. Moreover, the reestimation would be significantly easier with the addition of more precise information about velocities. Acoustic basement determination control was provided by adding of central beam relief data to the seismic section interpretation. The sediment thickness and acoustic basement relief maps are done with the other sounding beam information in the space between the seismic lines.

SCSP Results.

The region discovered (total square 240,000 sq. km) is located between the St.Paul fracture zone (f.z.) in the south and the Sierra-Leone f, z. in the north, and includes the equatorial segment of the Mid Atlantic ridge (MAR), Strakhov (4° N) and St. Peter (2°40’N) f.z. and the blocks between them (Plate 2, 23) . Our study resulted in the series of profiles (Plate 24-33), sediment thickness map (Plate 8)) and map of acoustic basement relief (Plate 9) on a scale of 1:1,000,000 with the isoline section corresponding to 100 and 200 m. The morphostructural scheme of the acoustic basement was created on this basis (Plate 10). Following the results of our survey certain corrections were inserted into the map “Thickness of sedimentary cover” (Tucholke, Uchupi, 1989-1990) included in the “International Geological-geophysical Atlas of the Atlantic Ocean”.

The quality analysis of SCSP sections wavefield has shown that sedimentary cover in the region studied could be classified in three seismic complexes upper, middle and lower. The upper complex is represented by well stratified sediments, with horizontal layering or inclination which corresponds to regional inclination of the basement surface. It is characterized by good phase correlation of observed internal reflections. According to SCSP data, this complex is universal for the equatorial segment of the Atlantic and usually has good representation on seismic records. The thickness of this complex in the basins reaches 300 m. The middle complex is also stratified, but as opposed to the upper complex, does not have long phase correlation. Numerous discontinuities, dislocations and wedging out of layers were observed in this complex. The middle complex has no legible roof. The thickens in the basins reaches 300 m. The lower complex is poorly stratified and fills numerous basins in acoustic basement relief Dislocations could also be observed in this complex.

Sedimentary cover in rift zone.

The rift zone of the ES MAR is divided by the Strakhov and St. Peter faults into three segments northern, central and southern. Sedimentary cover is very irregularly developed within their limits. It is represented by upper and middle seismic complexes and often complicated by basement intrusions up to the bottom surface. In the northern segment of MAR the sediments are virtually absent. The rift valley of the central segment sometimes contains only a few isolated sedimentary bodies, which have a thickness of up to 100m, and on the ridges of up to 250 m. The sediments are also absent in the rift valley of the southern segment. However, the quantity of sedimentary bodies on the rift adjusting ridges is gradually growing towards the equator.

The sediment bodies are mostly situated in numerous small submeridional depressions or in the transverse fault with a small shift zone, where they are sitting on its sublatitudinal depression’s southern slope. The sediments are found also to the east of the rift valley’s Northern segment, where they occupy the lower parts of wide depressions, oriented northwest-southeast.

In the areas of maximum sedimentation their thicness reaches 500-400 m, Three seismic-units are distinguished analogous to the above described units II, III and IV for the Sao-Paulo FZ polygon. This is evidence of the common development of the MAR equatorial segment from at least 3°30’N to the Equator during the formation of units II-IV. In our opinion, the similarity of the seismic sections of sediments in the Sao-Paulo FZ and riftogenic ridge sublatitudinal depressions is evidence of the absence of significant differences in the sedimentation patterns of these structurally very different forms.

The features of the deposits yielded evidence of the presence of sediment bodies on the rift valley bottom (Plate 8), which is very important for an understanding of rift valley tectonics. The sediment thickness exceeds 150 m. The whole sediment section represents seismic-unit IV and, judging by seismic record character and the absence of deformations, it is possible to speak of quiet sedimentation conditions. It is interesting that sediment bodies are related to the uplifted parts of the rift valley bottom with depths of 3200-3700m. Some bodies are situated in the near-slope part of the rift valley bottom.

The sediment distribution pattern described allows us to suppose that the rift valley had been generated before the seismic-unit IV formation. After that, in the rift valley’s most active parts a sediment reworking occurred as a result of a tectonic collapse, with deepened bottom parts formed or sediments buried by fresh rift basalt flows.

At 2°35'N along the step on the small transverse fault depression's southern slope are located a series of elongated isolated sediment bodies united by the same direction of big axes. The differences observed in their structure and deposition are the following: to the west of 31ııı25'W seismic-units III and IV are distinguished, while to the east - only unit IV; the sediments to the east of 31ııı25'W are characterized by deformation and top surface tilting, while to the west - deformations exist only in unit III and unit IV is distinctly horizontally layered. These differences may be explained by different sediment position relative to the rift valley northern segment; that is the sediment deformation zone which is situated along the nodal basin side at 31°25'W.

Most sediment bodies are related to submeridionally oriented depressions of riftogenic ridge. The sediments in them are of two types, each related to depressions having a specific position to MAR. Sediment bodies related to interlock depressions are bigger in surface, have a more complex inner structure (units II, III and IV are observed) and have greater thickness (up to 400-500 m).

In-block sediments are not so thick (up to 200 m) and represent only the upper seismic-unit, but bear traces of contemporary deformations indicating the relatively small-amplitude in-block tectonics which dominate here.

The sediments in the set of depressions oriented northwest-southeast are sharply different. These depressions are situated to the east of the rift valley north segment, and they are much wider and longer than submeridional depressions. They have anomalously thick sediments - up to 750 m, and are found only 40 miles from the rift valley. The sediment section has 3 seismic-units - II, III and IV. It is curious that traces of consedimental slope uplifts during units II and III formation are marked on seismic records. Thus we assume the northwest-southeast depressions to be rather old structures (older than the age of unit II), which had undergone activation during the formation of units II and III. This activation is expressed as a generation of positive "protruding" structures of the same orientation as the main direction of depressions.

Regional section analysis shows that sediment thickness to the west of MAR is greater than to the east and that both maximal and mean values at equal distance from the MAR axis differ about 1,5 times. The thickness increases both due to the thickening of corresponding seismic-units and due to the emergence of the most ancient seismic-unit I. Furthermore, the tectonic processes clearly express themselves more actively on the eastern flank of the ridge. This activity is seen in the edge dislocation zone, where sediment top surface distortions, numerous "protruding" structures, inner sediment deformations, acoustic basement deformations interpreted as horsts and faults..

Sedimentary cover of the Strakhov fault.

The Strakhov f.z. shifts the rift valley almost up to 110km. It was studied on the interval of 680km. The depth of the fault trough reaches 5100 m in nodal basins, and the tops of the rises, and surrounding trough have a depth of about 1600 m. Thus, the relief depth difference in the "active" zone of the fault is nearly 2000-3000m, and in the "passive" - 1000-1500 m. Sedimentary bodies of an "active" fault zone are distributed over the slopes and basins, and have an average thickness of about 100-150 m. Sediments are absent in nodal basins. "Passive" zones of the Strakhov f.z. are separated from the "active" zone at 31030' W and 33033' W by complex built fault crossing escarpments, which are located at depths of about 3800 m. One can observe a few sedimentary bodies at the trough bottom between escarpments, with a maximum thickness of 250 m, including the upper and lower seismic complex. Horizontal layering of the sediments attests to calm conditions of deposition. The bottom of the trough to the East and West of the escarpments is flat. There is, however, a V-form depression of acoustic basement, which is filled by deposits of all three seismic complexes with a total thickness from 300 to 800 m. These sediments on the map have bead-like west-east performance with the length of each chain about 20-30 km. They are divided by narrow (2-8km) isthmuses. The sediments on rift adjusting rises are located mosaically and have a thickness of no more than 100-200 m sediments. The absolute depth of the fault trough decreases from 1800 m to 1000 m in the basins. The southern and northern slopes are not symmetrical.

Sedimentary cover of the St. Peter f.z..

According to a recent detailed survey (21st cruise of r/v "Akademik Boris Petrov") the St. Peter f.z. does not intersect the rift valley . The seismic profiling of its west branch was rather detailed, but the east branch was not sufficiently studied.

The relief depth difference at the western branch of the fault is about 2500-3000 m. . In the trough bottom of the west branch we observe three steps running in sequence from west to east. Each of them is from 80 to 10 km long, and their depths are equal to 4000, 3800 3600 m. These steps are bordered by escarpments with a depth difference of 200-400 m. One could thus outline a steplike rise of oceanic crust on the approach to the rift zone.

The fault trough of the acoustic basement is V-shaped and tilled with sediments of all three seismic complexes. The maximum thickness of sedimentary cover (up to 1000 m) is observed on the central step. Average thickness here is 600-800 m. Acoustic basement surface along the fault trough direction is very complicated. Sedimentary cover is sometimes penetrated by basement intrusions. In such places the lower seismic complex is absent. It is also necessary to note that the St, Peter f.z. trough contains the biggest values of each seismic complex thickness. For instance, the thickness of the middle complex is 200-300 m greater than the same complex thickness in the Strakhov f.z.. A few mosaically located sedimentary bodies are observed on fault adjusting rises, which are represented by the upper seismic complex, with a thickness of about 100-200 m. On these rises the acoustic basement surface produces stronger reflections than other regions. The rises are crossed by depressions filled with sedimentary cover with up to 600 m thick, in which it is possible to determine all three seismic complexes.

Sedimentary cover of the Sao-Paulo fracture zone.

The Saint Paul (Sao Paulo) fracture zone (FZ) bathymetry is relatively well studied, but polygon surveys with single-channel seismics (CSP) in this region are not mentioned in existing publications, on an overview map of sediment thickness of the Atlantic (Geological-Geophysical Atlas) the Sao-Paulo FZ depression is marked as filled with sediments whose thickness regularly increases at a distance from the Mid-Atlantic ridge (MAR) axis, on the 7-th cruise of R/V “Akademik Nikolaj Strakhov” the Sao-Paulo FZ central part was surveyed with 24 meridian lines (Plate 8).

Five sublatitudinal depressions can be seen within the polygon . The maximum sediment thickness reaching 900 m is observed on the northeast of the polygon. Based on the wavefield characteristics the sediment section in the areas of maximum accumulation may be divided into four seismostratigraphic units. The lower seismic-unit I lies on the rough acoustic basement, the sediments are strongly deformed, and the wavefield is full of diffracted waves. The relatively acoustically transparent seismic-unit II with slightly inclined inner reflectors lies above. The seismic-unit III with strongly deformed inner reflectors is higher. The section is topped by the well-stratified seismic-unit IV, characterized by horizontal layering.

The different angles of intermediate layers’ inclination can be observed. This may indicate sharp differential vertical tectonic movements occurring on the sides of the depressions in which these sediment bodies lie (Plate 26). The possibility of the existence of regional sediment gap cannot be excluded, but this may be proved only through special seismic studies and deep-sea drilling. Five separate sublatitudinal depressions can be seen, with acoustic basement highs of a complicated shape. All the depressions are parallel to each other. It is interesting that mean depth values of all of the depression bottoms are nearly equal -4500-4600 m, excluding nodal basins with values of 5100-5300 m. The depth of the depression on the northeast of the polygon is also similar - 5200 m. All depressions are bead-like in shape, their wide deepened parts are separated by narrow saddles, and all of them are lying on approximately the same level -4000-4200 m.

Morphometric analysis of the acoustic basement topography map showed that the lengths of the long axes of the deepened hollows of the depressions are within the two intervals with modal values 15 and 25 miles. In the limits of a depression the deepened hollows of both types may be found. The depressions’ axis lines are wavy in shape, and the distance between the axis lines varies from 12 to 19 miles, with the most frequent value 16 miles.

Thus the two types of relief rhythmics expressed in the acoustic basement topography form fragmentation of two levels which are distinctly visible on the acoustic basement topography map.

To simplify the description we gave numbers running from south to north to all the depressions comprising Sao-Paulo FZ on the polygon. Two rift valley segments are traced on the polygon: the northern - between depressions 3 and 2, and the southern - betwveen depressions 1 and 2. A small amount of sediments is observed in both segments. In the northern the sediments up to 100 m thick are on the near-foot step of the rift valley western slope. In addition, two spots of sediments of the same thickness are observed on the rift valley slopes facing nodal basins of depressions 2 and 3 respectively. Strongly deformed sediments up to 200 m thick exist on the uplifted part of the rift valley bottom.

The sediments in the “active” part of the depression 2 are represented by 4 bodies isolated by the depression slopes uplifts. The maximum sediment thickness is 400 m. The angles of inner reflectors in some cases reach 5-8°. Based on the sediment structure and distribution features in the “active” part of the Sao-Paulo FZ depression 2 it may be asserted that a common sediment cover existed on the site of both the "active" zone and of 2 segments of the rift valley. Then it was reworked during rift valley formation. Tectonic and volcanic processes causing the sediment transformation also began in the "offset" part of the depression. We can suggest this as due to certain features of the CSP method even a relatively small sediment reworking by volcanic, magmatic or metasomatic processes will produce a wavefield on the record similar to acoustic basement, but not to sediments.

The Sao-Paulo FZ northern depression bottom in its western "passive" part is covered with sediments up to 800 m thick. The area of maximum sedimentation presents the seismic-units II, III and IV, and the top surface of all of them is practically horizontal. The units III and IV are everywhere characterized by the sticking of inner reflectors to the depression sides. At 26°17' W one can see a strong sediment deformation - on the record a prograding dome-like structure is visible which caused obvious post-sedimentary distortions of inner reflectors. This structure may have been caused by the rifting process. From the acoustic basement topography map it is seen to lie on the fault cutting the ridge which separates depressions 1 and 2, and this fault is one of the set of en echelon faults, the main one of which forms the rift valley northern segment's west side.

In the depression 2, the western "passive" part the sediments occupy 3 deepened hollows. In the section seismic-units III and IV are marked with different inner reflectors inclination. The significant feature of unit III is a connection to the sides and an enveloping of the hollow bottom which may indicate the con sedimentary relative uplift of the corresponding sides.

The reflectors of the unit IV in the eastern hollow are inclined to the south, in the central hollow - towards the north, and in the western area lie horizontally.Thus different directions of neotectonic movements of ridges bordering the depressions can be concluded. Evidence pointing to the existence of tectonic movement inversion was noted. On line 22 a distinct northern dip of unit III reflectors and their connection to the depression's southern side is seen, and on the other hand, the southern dip of unit IV reflectors may be observed.

Practically all the depression 3 western "passive" part bottom is occupied by sediments. 10 miles east from rift valley the sediment thickness already reaches 200 m, and at 25°25'W - even goes up to 800 m (only 55 miles from the rift valley!). Seismic-units II, III and IV with very different inner reflectors inclination are marked here. Near 25°45'W a strong tilt of modem sediments surface is marked indicating the contemporary relative uplift of the depression's southern side. Traces of vertical movements in the unit II, which do not influence the upper sediments, are visible on seismic sections. At 26°15'W a sediment body 200 m thick exists on the bottom of depression 3. A seismic section shows this body wavefield to differ greatly from ordinary sediments in depressions and is in fact very similar to the wavefield of the sediment "cap" situated 1000 m higher. It is possible to presume that these sediment bodies have been united in the past, and then either the uplift of the depression northern side or a deepening of the bottom occurred. In both cases, however, the presumed amplitude of the tectonic movements is about 1000 m.

<u>Blocks between the fracture zones.</u>

Taking into consideration the geographical position and morphostructural features of the region studied we have specified 6 blocks (FB) between the fracture zones (Plate 8, 9), 1, 2, 3, 4 FB are represented by north-south directed ridges and depressions. The ridges appear as intrusions of acoustic basement and have the shape of long and narrow crests with a 10-20 km base width and up to 25-30 km long. They have coulisse-like locations and the crest bottom depth ranges from 2000 to 2800 observed in small square depressions. The thickness of sediments on the slopes of these depressions does not exceed 100-150 and is from 30 to 50 km long. They are tilled by sediments of all three seismic complexes. The average thickness of sediments in 1 and 2 FB varies from 200 to 400 m, and in 3 and 4 FB from 400 to 600 m. The reflecting horizon of the middle complex often has a concave shape. In the limits of 3 and 4 FB the biggest total value of all types of deposits can be observed . At 33° W it is possible to watch north-south crest 90 km long, that is crossing 3 FB The acoustic basement relief of 5 FB is strongly dismembered. Depressions are filled by sediments of the upper complex with a thickness of about 200-300m . A big west-east depression is located in the south-eastern part of 5 FB. It is filled by sediments of all three complexes with a total thickness about 400-600 m.

The acoustic basement surface of 6 FB is also strongly dismembered. The roof of sediments has an inclination and is complicated by penetrations of the acoustic basement. An arc-form north-south crest 120km long could be observed in the limits of 6 FB, while the direction changes to eastward in the northern part of FB. We also observe separate sedimentary bodies with a thickness of less than 100 m.. This crest is associated with big depressions filled by sediments of all three complexes with thickness about 400-600 m.

SCSP Survey Result Conclusions

1. separate sedimentary bodies, having a thickness of about 100 m, were discovered at the rift valley of MAR between the Strakhov and St, Paul fracture zones. The sediments in the rift valley to the north and south of these bodies are absent in seismic resolution (30 m).

2. Passive parts of the fracture zones have a v-shaped section of acoustic basement and are filled with sediments up to 500 m thick.

3. Complex built fault trough crossing escarpments were discovered in the acoustic basement relief at the Strakhov f.z. near the transfer from the active part to the passive branches.

4. A steplike west-east rise was discovered in the acoustic basement relief of St. Peter f.z. The maximum value of sediment thickness was observed on the central step 180 km from the rift zone axis, and it reaches 1000 m.

5. The maximum average values of sediment thickness are about 600-800m. They were discovered in the limits of the passive zones of transform faults, but there is no increase of average thickness from the rift zone axis to periphery.

6. The acoustic basement relief of fracture zones troughs has bead-like shapes with alternating west-east depressions and narrow isthmuses.

7. Rare and thin (100-150 m) sedimentary bodies were discovered in the active zone of the Strakhov fracture zone.

8.The sedimentary cover is usually absent at fracture zones' adjusting rises and appears only in crossing depressions.

9. The ridge crests in blocks between the fracture zones have coulisse-like location. They alternate with north-south depressions, filled with sediments. The average thickness of sediments is about 300-500 m. The total quantity of sediments increases in the southern direction.

10. Through structural divisions the sedimentary cover, observed in the region studied, could be divided into three seismic complexes: the upper, with horizontal or weak inclined layering of reflecting horizons; middle, with numerous dislocations and penetrations of the acoustic basement and deposited under strong tectonic activity; lower, acoustically apparent, with significant dislocations and probably formed at the early stage of bedrock relief formation.

11. The sediments in Strakhov and St. Peter f.z., and also in big depressions at the blocks between the faults, are represented by three seismic complexes.

12. The basins of the acoustic basement, north-south ridges in the blocks between the faults and fault adjusting rises are represented only by sediments of the upper seismic complex.

Chapter 5

Tectonic Deformations within the Mid-Atlantic Ridge and a Model of their Genesis.

M. P. Antipov.

Seismic investigations in the World Ocean found widely-distributed dislocations of' sediments and basement of different forms. Such dislocations within the transitional zone between the Mid-Atlantic Ridge (MAR) and abyssal plain are of special interest (Udintsev, Beresnev, Gordin, 1980, Odinokov, Udintsev, Beresnev, 1990). A.V.Pejve (1975) assumed that tension and extension took place during the evolution of the MAR. These events would cause tectonic deformations. The fold and fault structure of the basement and sedimentary cover were discovered and described by M. Antipov et al.,(1990) during investigations of both MAR flanks near 12 N. The magnitudes of these dislocations were 300-500 m.

The deformation of sediments within the Equatorial Atlantic are distributed on the wide band along the Equatorial Segment (ES) MAR. There is no good evidences of the dislocation in the transitional zone from the ES to the abyssal plains. Within the ES, however, the data of seismic profiling give multiple evidence of tectonical deformations of the sedimentary cover (Plate 8, 9,10, 23, 24-33).

The relief of the western flank is smoothed from the ES axis. The amplitudes of relief irregularities are smaller and vary from 200 to 1500m. Seismic records show acoustic basement and some different reflectors correlated with second and first oceanic layers respectively.

The very irregular acoustic basement surface in depressions reaches a depth of up to 5 -5.5 km and of 3-3,5 km on a rise. There is no correlation between two parallel neighboring profiles along ES. There are no linear structures within its western flank. There are some local basement rock exposures on the ocean floor, and the basalts and Fe-Mn crust were dragged here. The sedimentary cover filled the lower parts of the basements surface, burying its irregularities. The horizontal sedimentary layers overlap to the basement surface. The width of the local basins vary from 10-20km in the east to 40-50km in the west. The thickness of sediments is up to 800-1200. There are no good distinctive characteristics on the distribution of the sediment thickness. It is possible that the thickness of sediments depends on the basin size and the quantities of terrigenous material. The composition of Holocene sediments is carbonate with a clay component. The terrigeneous materials are debris of the rocks of the surrounding mountains. The basin square and thickness sediments increase towards the south. Along the longitudinal profile (33°05' W) the ocean floor lifts slowly from 3700 m to 3200 m towards the North. The level of the floor of two neighboring basins is different. This difference is 350 m, and the thickness of the sediments within these basins is 700 m.

There are two types of local basins. The first type consists of narrow (up to 30 km) simple basins filled by horizontal sediments. There are long reflectors distributed across the whole basin in the seismic profiles. When the basement surface is very irregular, we sometimes have no correlation of reflectors of the lower part of profiles. Locally, some reflectors lift towards the flank.

The second type is the wide (up to 50 km) basin with wavey basement surface. Seismic records across these basins demonstrate the parallel horizontal reflectors in the upper part (300-400 m) and transparent lower part. There are some wave and noncorrelative reflectors in the lower part of profiles in some cases. There is a dome-shaped rise or anticline on the western flank, which are assumed to consist of older sedimentary rock. There are flexures, horsts, grabens and simple faults on the profiles across this dome . Sometimes faults deform the floor surface. The horst and grabens width are 10 km, and magnitudes of the displacement are 500 m.

The profiles along the longitude band between 34-35°W and 32-32°30'W better demonstrate this rise. The deformation structure changes towards the south. There are faults and sharp flexures of the floor surface above the basement dome structure (projection) between two meridians -34°30' and 34° W. The lower part of the sedimentary cover is exposed here. There is a different floor depth of three neighboring basins along the profile between 33°30' and 30°W at an interval of about 90 miles towards the east from 4000 m to 3300 m. The width of the eastern basin is about 60 miles. The basement surface is irregular with amplitudes of the relief up to 500 m. The profile demonstrates some short reflectors. The thickness of the sediment is up to 700 m. There are two intervals with deformed sedimentary layers between 33° W and 31° 50'W and between 33°W and 32° W. The floor surface is wavy with flexure amplitudes of up to 300 M. There are short wavy reflectors deformed by faults.

The structure of the sedimentary cover varies towards the south on the western flank of the ES MAR. Sediments bury the irregular basement. The sediments thickness increases from 200 m to 1200 m towards the west. The flexures, anticlines and faults are fixed above the basement projections. The deformation amplitude varies within the range of 300-400 M. The length of the reflectors varies considerably laterally and vertically. There are numerous reflectors across the whole vertical section locally and there are no reflectors within other places. The western edges of profile 1 and 2 (Plate 24,25) crossed the margin of the Seara rise. The horizontal sedimentary layers filling the narrow basins are up to 700 m thick. There is a dome rise of basement along profile 2. The sediment thickness decreases up to 300 m here. The dip is up to 5-7 within the limbs of this dome rise. The width of the rise is 100 km.

Some profiles crossing the eastern flank of the ES demonstrated the sedimentary cover structure and relief of the basement. There is no good correlation between two parallel neighboring profiles. The surface basements relief is sharply irregular, its amplitudes are above 2 km decreasing to 0,5- 1,5 km towards the east. The depth of the basement increases up to 5 km towards the east. The width of the basement projections varies from 30 to 60 km and more, wider than the size of the neighboring basins. The width of the basement projections and neighboring basin is equal within the eastern parts of these profiles. Sediments fill the lower part of the basement relief. The thickness of sediment ranges from 200-300 m to 700-1000 m. Sedimentary layers are parallel and horizontal. The floor depth is different within neighboring basins, with a difference of up to 200 m. There is a large complete rise between meridians 28 20 W and 29 20 W along neighboring profiles. This rise dips towards the north and south. The basement relief is very rugged with amplitudes of up to 1,5 km. The horizontal sediments filled the narrow grabens -10 km wide and 500 m thick. The sedimentary cover thickness decreases towards the west. There are short reflectors and many faults here. The basement relief becomes calmly wavy towards the south and east and is buried by sediment. The basement rock is exposed in some places. Seismic records demonstrate the complicated wave pictures to the east from longitude 290 W. There are flexures, faults of the sedimentary layers and basement. Some scarps are expressed on the floor relief with an amplitude of up to 500 m. These scarps are normal faults with eastern and southern upthrows.

Some profiles coincide with the fracture zone. The eastern part of these lines cross the smooth floor (the depth is 37004000 m). The basement relief is also smoothed out with local waves, but it is more irregular near the axis of ES and on the east between 28 and 27°W, and the height of these irregularities is up to 1,5 km. Sometimes the basement rock is exposed on the floor and forms sea mounts up to 700 m high. The parallel-horizontal sedimentary layers bury the basement relief Their thickness is up to 1000 m. The parallel- horizontal structure of the sedimentary cover is disturbed between 28-29°W and numerous short reflectors show the anticline. There is a fault on the eastern limb of this anticline. The thickness of the sedimentary cover varies within a range of 300-1000 m.

An analysis of the sedimentary structure and particularities of the basement in the flanks of the Equatorial Segment shows its compound mosaic structure. The thickness of the sedimentary cover changes within a wide range from the first dozen meters up to 1.5 km without any evidence of a regularity of distribution. The sizes of the basin increase towards the flanks from the axis of the rift zone. Flexures and faults of the sedimentary cover and basement surface show the complicated different vertical displacements and dislocations. The age of sediments is Pliocene-Quaternary on the central parts of the MAR and Oligocene-Miocene (Emery, Uchupi, 1984).

Chapter 6

The Crustal Structure and Seismicity of the Equatorial Segment in the Strakhov and St. Peter Fracture Zones area.

J.Makris, U.Vogt, M.V.Zakharov, B.N.Grinko, T.Funk.

During the 11th cruise of the R/V “Akademik Nikolaj Strakhov” a deep seismic survey using ocean bottom seismographs (OBS) and powerful airguns was conducted. German scientists from the Institute for Geophysics, Hamburg University, supplied the receiving equipment, Russian scientists - the airguns, and the observations were conducted jointly. The German side undertook responsibility for data processing, and provided copies of maps and diagrams, Russian side prepared the current very brief paper which does not claim to be a full presentation of the results.

Seismic experiment consisted of two parts: deep seismic sounding in the St. Peter fracture zone trough and seismological observations in the Strakhov fracture zone. The available visual materials of the experiment comprise Plate 13 and 14 of the Appendix.

Deep Seismic Sounding (DSS).

The DSS observation system in the St. Peter fracture zone consisted of 5 OBS situated on the sublatitudinal profile along the 2°40’ N of 100 km total length. OBSes covered a 30 km distance, and the eastern and western flanks of the profile were each 35 km long. The OBS spacing was about 7.5 km, water depth about 4000 m,and an airgun with a 60 litre capacity was fired at 300 m spacing.

All of the OBS recorded seismic waves up to maximum distances of 50-60 km on the system of reversed and overlapping traveltime curves. As an example Plate 14 presents a seismograms for OBS 1, 3, 5 reduced to 6.0 km/s. For all records channel 1 is vertical, channels 2, 3- are horizontal components.

As a result of data processing german scientists proposed the following compressional wave velocity model of the earth crust deep structure (Plate 13). Sedimentary layer thickness decreases from west to east from 1.1 to 0.5-0.6 km, velocity 2.0 km/s. The “Second” layer consists of rocks with Vp=4. 8-5.4 km/s and 5.6-6.2 km/s (sublayers 2B, 2C) and has a thickness from 1.5 km in the center to 2.3 km on the flanks of the profile. The “Main” oceanic layer has Vp=6.6-6. 8 km/s and thickness 2.0-2.5 km. The Moho boundary velocity is 8.1-8.2 km/s.

In general the wavefield recorded on all 5 OBS is distinctly separated into two types.

The first type wavefield is typical for OBS 1, 2. Here in the first arrivals the refracted waves are virtually nearly absent, but in the precritical area waves PS and PSSS are clearly expressed. PS is a basement reflected wave with had undergone exchange on the basement-sediments boundary. PSSS is the same wave with multiple reflection in the sediment layer. In the later arrivals the refracted waves P3S and PmS from the “main” oceanic layer and upper mantle are recorded, which had undergone exchange at the end of their way on the basement-sediment boundary. At greater times these waves are repeated as multiples in the sediment layer, and an amplitude increase of horizontal components related to vertical is observed.

In our opinion, records of OBS 4, 5 belong to the wavefield of the second type. Refracted waves P3 and Pm are clearly seen with apparent velocities 6.6-6.8 and 8,1-8.2 km/s on distance intervals 8-22 km and 28-60 km respectively. In the later arrivals at distances 18-50 km a postcritical Moho reflected wave is clearly traced.

We consider the OBS 3 (profile center) wavefield as transitional from the first type to the second. Together with exchange waves PS and PSSS on the eastern half of the seismogram in the first arrivals the crustal refracted waves are fragmentarily traced at distances of 10-20 km.

The following explanation may be proposed for the wavefield difference observed along the profile.

The total absence of refracted waves in the first arrivals typical of the first type of wavefield is related to the strong refracted wave attenuation in the fractured basement media covered with sediments about 1 km thick. The characteristic block size of the fractured basement is supposed to be comparable to the OBS spacing. The second type of wavefield with good tracing of refracted waves in the first arrivals is assumed to reflect the structure of unfractured oceanic basement.

Seismological observations.

In the central part of the Strakhov fracture zone at the intersection with the northern rift branch seismological observations were carried out for approximately 3 days. The observational system consisted of 5 OBSes in the form of a right angle with the top in the nodal basin center (5000 m contour), with the beams to the north and to the east (Plate 14).

On the OBS records 28 seismic events were revealed with local magnitudes 1 - 3. The main group of events (24 events) is located in the area with the central coordinates 4° N, 32° W having hypocenter depths 4.9-5.2 km and magnitudes 1-3 (18 events) and hypocenter depths 6.0-7.0 km and magnitudes 2-3 (6 events).

Thus the maximum number of local earthquakes is located on the northern rift axis,with others distributed latitudinally (3500-4500 m contours).

According to the data from the teleseismic network catalogue for 1972-1992 about 40 earthquakes with magnitudes 4.3-6.0 are located in the area with coordinates 3.64.3° N, 30-34° W. A classic pattern of earthquake distribution is observed in the rift-transform fault junction.

Chapter 7

Structure of the Sierra Leone Rise on the Eastern Flank of the Equatorial Segment and Guinea Plateau of the Continental Margin of West Africa

A.Yu.Yunov

Gravity and Magnetic Investigations of the Sierra Leone Rise on the Eastern Flank of the Equatorial Segment of the Mid-Atlantic ridge

On the basis of the data of the on - board magnetic and gravimetric investigation, the anomalous field of the Sierra Leone rise and of the basin bearing the same name can be divided into three regions: the northern, southwestern and southeastern. The northern region is characterized by high amplitude gravity anomalies. They are isometric on the map and are roughly in accordance with the North block. The predominant stretching of the anomalies is to the north - east and latitudinal, High frequencies of variable signals with an intensity of over 1000 nT1 are noted here. The most intensive gravity anomalies correspond to the positive and negative magnetic anomalies or to the zones of the latter's maximal gradients. On the southwestern region of the heightened level of gravity anomalies in free air are observed. Anomalies are distinguished on the South block high frequency, and here the negative magnetic field is predominant with an intensity not more 600 nT1.

The south-eastern region corresponds to the neighboring part of the Sierra Leone basin. Earlier publications (Russakov et al., 1987) state that the regional anomalies in free air on the region of the investigations are characterized by significance near zero, and that the range of their changes is not more $^{+}$25 mgl. This should be evidence of the isostatic equilibrium for the main part of the region. The intensive anomalous field is explained by topographic and structural features.

For the estimation of the isostatic state of the crust Yu.Pavlov's simplified method is used - (Yu.Pavlov, V.Semakin,1973), based on the correlation of the Bouger anomalies and bottom relief Applying the palette with the correlation graphics of the dependence of the gravity field to the relief of the oceanic bottom, lines of equal significance of the coefficients of overcompensation and of the thickness of the anomalous compensating layer correspond to the isostatic hypothesis of Eri Veining-Meinesz.

The application of this method resulted in the chart of the thickness of the anomalous compensation layer (K), characterized by the difference between the "theoretical" and "true" thickness of the earth crust. The analysis of the chart of the K shows that the isostatic equilibrium is upset in the northern part of the Sierra Leone Rise. Based on the thickness of the K and the relief of the rise, the deficiency of the earth crust's thickness is characterized for that part of the rise which spreads the volcanic sub-bottom sea mountains. The parts of the rise with the quiet relief are in fact characterized by isostatic equilibrium.

The density model of the section of the earth's crust was made on the 8701 and 8702 profiles. This made use of the top and bottom of the sediments on CDP data, and also of the value of the redundancy densities 0,89 g/cm for the boundaries water - sediments and sediments - basement. The gravity effects of the water thickness and thickness of sediments are subtracted from the observed significance of the gravity field. The obtained residue significance of the gravity field on the Sierra Leone Rise and Basin depend to a small extent on the variations of the redundancy density in the bottom of the sediments (0,2-0,4 g/cm). This points to the deep source of the anomalies.

Based on the spectral analysis of the observed value of the gravity field on the rise agreeing with the approximation of the geological objects by vertical steps, the upper level of the main source of the anomalies lies at depths of 22-25 km, 11km, 7km and 5km.

The main contribution in the regional element of the gravity field made by density boundaries lying at the following depths:

Sierra Leone Rise -4 km; 11 km

Sierra Leone Basin - 5km; 8,4-9,4 km.

All the magnetic anomalies are timed to the fault zones. No good correlation between the magnetic and gravity anomalies is observed that attest to smaller redundancy density of the magnetic bodes.

The results of the density modeling are compared to meridian profile on meridian 20°00' W made by "Sevmorgeologia" and to models by E. Jones and K. Mgbatogu (17-1982).

According the model of these latter authors the value of the density of the sediments onshore and on the Sierra Leone Rise are assumed to be 2,30 g/cm . This is too high for the oceanic region. The values of the density of the layers 11 and 111 are assumed to be 2.70 g/cm and 2,90 g/cm, and for the upper mantle the value of the density is assumed to be 3,27 g/cm. In comparison with the "standard column" of the oceanic crust made by J. Worzel (1974), the values of density in this model of layer 11 is more than f), 15 g/cm and of layer 111 - more than 0,04 g/cm. The density of the upper mantle seems to us also excessively high.

In accordance with the above, E. Jones and K. Mgbatogu modeled the profile in which he thickness of the crust under the shelf of the Guinea marginal plateau is 7 km, and close to 10 km under the northern part of the Sierra Leone Rise (with some breeches of the isostatics). In comparison to the "standard column" the thickness of layer 11 is reduced and the thickness of layer 111 is increased.

In the meridian profile of the "Sevmorgeologiya" the sediment density is assumed at 1,90 g/cm (it is equal to "standard column" and to average density of the sediments in the borehole N 366). The density of layer 11 is assumed to be 2,70 g/cm (except for the section with the seamounts). The density of layer 111 is assumed at 2,90 g/cm,

When modeling the density profiles 8701 and 8702, the density of the sediments was assumed to be 1,70-2,40 g/cm., and the density of layer 11 -2,70 g/cm and of layer 111-2,80-2,90 g/cm (Plate 17, fig. 2,3). During the process of work it was necessary to assume that under the Sierra Leone Rise the density of the upper mantle must he 3.20 g/cm.

The resulting profile shows that the thickness of crust under the neighboring part of Cape Verde Basin is 5.0-5.3 km, under the North Block of the Sierra Leone Rise - 7.5 km, and under the Southern Block of the Rast -9 km. Under the neighboring part of the Sierra Leone Basin the thickness of crust is assumed at 3.0-4.5 km. The comparison of the density models on the lines 8701 and 8702 with the model of E, Jones and K. Mgbatogu and "Sevmorgeologiya" shows content values of the depths of the crust's bottom. At the point of the crossing of the line 8701 and the profile of E. Jones and K. Mgbatogu, the depths of the crusts bottom are equal or differ in depth by not more than 1.5 km.

The difference in depth in the point of crossing line 8702 and the E. Jones and Mgbatogu profile is about 3.0 km. These differences in depth may be explained by differences in assumed densities.

The correlation of the assumed density profiles with the "standard column" of the J. Worzel for the Atlantic show the departure from the latter in profiles 8701,8702. The depth of the crust bottom under the Cape Verde Basin is more 1.5 -2,0 km in our profiles. The correlation of the profile of the Cape Verde and Sierra Leone Basins show that in the latter the thickness of layer 11 is less, and layer 111 is greater. The thickness of the entire crust for both basins, however, is very similar. The thickness of the crust under the Sierra Leone Rise is 2.5-3,0 km greater in comparison with the "standard column". At the expense of layers 11 and 111 the thickness of the crust here is twice as great.

The seismostratigraphy of the Sierra Leone Rise.

The problem of the seismostratigraphy and chronostratigraphy of the sedimentary sequence of the Sierra Leone rise encounters difficulties.

Only borehole N366 IPOD on this rise reached at a depth of 550,5 m (the bottom of the hole) the marlstone of the Maastrichtian (Plate 17, fig.9).

The geological age of the lower part of the sedimental sequence has a thickness of more than 2 km, and the nature of these laminated substrata is unknown, The borehole 13 IPOD drilled in the western part of the neighboring Sierra Leone basin meets at the bottom of the hole (463 m) the Cenomanian strata. The thickness of sediment cover here is not as great in comparison with the Sierra Leone rise. In this borehole the hiatus of Oligocene time is observed.

In both boreholes -13 and 366- the main seismic reflector A* is not reached. only the borehole 367 in Cape Verde basin north of the Sierra Leone rise reached the top of layer 11, but only if this is not simply the basaltic sill of the Jurassic sediments.

our seismic profile 8001-8002-8003 transects the Cape Verde basin from borehole 367, west ankle of the Quinean marginal plateau and Sierra Leone basin and print of the borehole 366 on the eastern past of the Sierra Leone rise (Plate 17, fig. 1, 2, 3, 4). The chronostratigraphic correlation of the main seismic reflectors of the Cape Verde and Sierra Leone basins, is based on our seismic profiles CDP 8001,8002,8003 (Yunov, 1987). All this gives us the possibility for seismostratigraphic correlation of the main reflectors and seismic sequences of the Sierra Leone rise.

The seisrnostratigraphic complex of the Cape Verde basin presents the profile 8705 (Plate 17,fig.6). Here, and on the profiles 8001, 8002,8003 the main seismic reflectors named A ,A ,A ,A , A ,C,C 1 and C2 are distinguished. (Plate 17, fig. 10, 12).

The seismostratigaphic correlation of these reflectors is based on eustratic cycles modelled by B. Hag and others (1987). The reflector A is the best seismic reference for the Atlantic ocean from Florida to Senegal and from Newfoundland to Angola. It is diachronous in the boundaries of the ocean. In the zone of the lower part of the continental slope this horizon disappears and is replaced by reflector A.

This latter is widely present in the continental margins of the Atlantic and other aquatoria and is correlated with the great regressive phase with an absolute age of 29,5-30 mln years.

Main seismic reflector A in borehole 367 IPOD corresponded to the top of lithologic complex 4a (black lignite shale upper apt - lower Turonian age). The top of this shale is bedded 620 m (0,7 sec) under the bottom and is considered to have had dolomitic strata 700 m under the bottom. On the seismic profiles 8001 the reflector A* is located in the upper part of the polyphased wavelet 0,3 sec and under reflector AC. Reflector A* is well correlated along the 400 km profile and is well distinguished on the other profiles in the main regions of the Atlantic. Correlation of the reflector A* in the Fast Atlantic with borehole 367 shows that it is in keeping with the eustatic cycle on the diagram by Hag and Vail (B.Hag et al 1987) in accordance with the inter-cenomanian hiatus aged 97 mln years. P.Vail et al 1981 marks this event for the West Atlantic, not for the east. But this seismic horizon is well distinguished in the Cape Verde basin (and in other parts of the east and South Atlantic). The reflectors A* with the reflector A presents a significant seismostraphic wavelet. Both reflectors A and A are divided by a specific semitransparent interval 0,3-0,4 sec on the upper part of the continental rise. This interval in borehole 367 corresponded to sand shale. The number of sand layers in the lower part of interval decreased, and here the first beds of black shale appeared. The age of this complex is upper Cretaceus - upper Paleocene.

The reflector C is well distinguished, two or one phases fixed on 7,3 sec. In borehole 367 it corresponds to the top of the carbonate complex (900 m under the bottom). It is considered as calestones, and olive marlstones laminated with black shale. It is a stratigraphic analogue of upper Titonian-Hauterive nanno-chalk and organics.

The intensive reflector 0,1 sec under the horizon C corresponds to the top of brown-red and gray argillitic calstones laid on the basalts of the layer II 1100m under the bottom. This complex had analogues in the West Atlantic, where the age is 137-138 mln years - main reflector Y1 of K.Klitgord and 1980). This is reflector C1. Horizon corresponds to the regressive phase aged 131-135 mln years (B.Hag et al. 1987). In the direction towards the continent both reflectors are well distinguished and join under the continental rise. Here and there the lower horizon joins with the reflector of the top of the layer II basalts, but under the zone of the continental rise and the continental slope the top of layer II sinks, and the thickness of the Jurassic increases to 4 km and more. The seismic feature allows us to assume the carbonate nature of Jurassic complex,

The seismic complex between the reflectors C and A* (their thickness in borehole 367 260 m 0,3-0,35 see) is characterized in the upper part by high frequency wavelets. The lower part of this complex is acoustically transparent. In the direction towards the continental slope in this part relatively weak reflectors appear. In the borehole 367 this complex is represented by black shale rich in organic material (8-28%), with fish and plant remnants,

Along the profile 8701 from borehole 367 toward the Guinea Plateau all horizons are well distinguished, Under the continental rise new reflectors appear - A",A"',A**,A*** corresponding to the eustatic curve of B.Hag et al, with events of an absolute age of 60 mln, 70 or 75 mln, 112 mln and less than 131 mln years.

The reflector from top of layer-II is not distinguished in all places. Here and there it is very smooth, complicated by reverberations, secondary waves and is probably connected with thin acoustically hard sediments bedding on the top of the layer II basalts. The other parts of the profile point to a typically “oceanic” top surface of layer II, with hyperbolic reflections.

In the Cape Verde basin toward the Guinea marginal plateau all seismic sequences are well distinguished. The thickness of the Jurassic under reflector C is variable. On the top of the surface of layer 11 the Jurassic is absent, and in the depression it increases.

The new reflectors in Jurassic are probably analogues of the mid - and lower Jurassic West Atlantic (K. Klitgord et. al). Under the upper part of the continental slope the Jurassic is characterized by the presence of carbonates with reef facies. South of the Guinea Plateau in Sierra Leone basin seismic stratigraphy is comparable and rather like seismic stratigraphy of the Cape Verde basin. The stratigraphy of the Sierra Leone Basin and Rise is supported by boreholes IPOD N 13 and 366, drilled in the basin and on the eastern border of the rise. The seismic investigations prior to drilling immediately displayed the key horizons here except horizon A, and therefore the stratigraphication of the seismic complexes was approximate (Initial Reports of the DSDP XI,). The base of borehole 366 reaches Maastrichtian and shows the considerable carbonate character of the sediments of the rise and the considerable terrigenous character of the Basin.

The horizon A is well distinguished on the profiles CDP 8002-8004(Plate 17, fig 3, 4, 5) transecting the Sierra Leone Basin between boreholes 13 and 366. This is a well-expressed high amplitude reflector of two or three phases 0,3-0,5 sec under the bottom. Near borehole 366 on the eastern part of the Sierra Leone Rise this reflector corresponds to the top of the mid - upper Eocene complex (interbedding of the chalk cherts and porcellanites). Bedded on this complex are the upper Eocene-low Miocene and chalk. Then the reflector A is recognized as A It is an analog of the same reflector in Cape Verde basin, and their approximate age is 40-49 mln years, Here and there 0,1 sec above this reflector is the reflector - satellite, probably A“ (29-30 mln of years),

Approximately 0,25-0,3 sec below horizon A and subparallel to this, the reflector recognized as horizon A* is clearly distinguished. In borehole 366 at a depth of 850,5 m (the marlestone of the Maastrichtian) this reflector is not reached. As in the Cape Verde Basin the age of this horizon is approximately 97 mln years. In the Sierra Leone basin below this reflector lies the transerved bedding on the basalts of layer II. The low amplitude and frequency reflectors are present in the upper part of this complex. In the depths of the top of layer 11 the thickness of these sediments increases to 2,0 sec due to transparent complex. Here and there the lower complex decreases and horizon A*, and rarely A, are bedded on the basalts. Lower amplitude reflections in the lower part of the sediments are distinguished in the deep depression of the top of layer II.

Analogues of the main Jurassic horizons C and C1 typical for Cape Verde Basin, however, are not distinguished here. This means that the basal sediments of the deepest parts of the Sierra Leone basin are not older than the boundary between Jurassic and Cretaceus (131-137 mln years).

The reflections from the top of layer II are very different - hyperbolic (typical for most oceanic basins), sub-horizontal, or dipping correlatable two phase, intensive. The nature of the latter is unknown.

Many intrabed reflections, sometimes very intensive, make it impossible to recognize the true top of basalts in such areas. When this intensive reflector on the basalts is sedimental in nature it corresponds to conventional horizon C or reflector A***. In such a *case* its age is 131-137 mln of years-lower neocomiat.

In the Sierra Leone basin the thickness of the seismic sequence between the reflectors A* and Ac is equal to the latter in Cape Verde basin.

On the Sierra Leone Rise the thickness of the upper sequence - above the horizon A* - is one and half times more that in the Sierra Leone Basin, whereas the thickness of the complexes of the horizons A* and Ac are equal. There is predominance of the carbonate facies on the Rise, and predominance of the terrigenous - carbonate facies in the Basin (Plate 17, fig.1, 13).

On the Sierra Leone Rise the stratigraphic correlation of the reflectors H, A’, Ac, A* and A** is relatively reliable.. The reflector H corresponds to the top of turbidites of the upper Miocene - low Pliocene (pre-Miocene regressive phase of P. Vail). The reflector A’ is distinguished on the eastern part and probably corresponds to the upper Oligocene regressive phase (Plate 17, fig.9,10,11,12).

The reflector Ac - is diachronous and corresponds to the top of upper - Eocene cherts (borehole 366), and in the Sierra Leone Basin corresponds to the pre-upper-Oligocene regressive phase(30 mln years); in such case it is equal to horizon Au.

The reflector A* corresponds the top of the black shale of the Albian - Cenomanian in the Cape Verde basin and probably to synchronous sediments on the Rise, not reached in borehole 366. The reflectors on the Rise were not reached in borehole 366. The reflector A** corresponds to intra-aptian hiatus.

The age of the acoustic basement on the Sierra Leone Rise and basin *is* different in different places. It is present either as volcanogenic strata covering the older sediments, or bedded on the top of the basalts of layer II.

The intra-bedrock reflections are evidence the stratified character of the upper part of layer II. The stratified character of the upper part of layer 11 or the bedding below the “upper”, clearly sedimental complex may be volcanogenic-sedimental or sedimental in nature. The age of that is unknown, but it is probably pre-Cretaceous or even pre-mid-Jurassic. Here and there the angle discordance between the latter and upper-aptian-quarternary sediments is considerable , in other places the unconformity is practically imperceptible. But a common “picture” of the seismic reflections and the high formational velocities in this complex allow us to assume significant time interruption of the “upper” and “lower” complexes between sedimentation and the considerable changes in composition. This is evidence of strong tectonic events in the time interval corresponding to the interruption in sedimentary process.

It was interesting to correlate this “lower” sedimentary (or half sedimentary) sequence with the pre-upper-Cretaceous sequence of the Chines continental Plateau, where the seismic investigation of recent years distinguished thick lower Cretaceus - Triassic sedimentary strata and on the Guinea Shelf - the sedimentary Paleozoic sediments. The latter are the prolongation of the Paleozoic complex of the Bouval depression.

If the latter is correct, the Sierra Leone rise may be recognized as a fragment of the former African continent. In general the correlation of the “upper” and “lower” sedimental sequences on the seismic data are analogous to the correlation of the Voring Plateau and of the east edge of the Norwegian basin (Hinz, 1981). In these places the top of the acoustic basement along the profile shifts to unconformity or discordance between two strata sequences. The lower stratum corresponds to volcanogenic-sedimental or sedimental bedrocks, including those pre-Cretaceous and even Permian in age.

In general the character of the seismic material and sediments known by drilling in borehole 13 and 366 attest to a stable tendency of sinking and of appropriate replacement of the coastal or pelagic facies by more deepwater facies.

This process begins not later than the upper Cretaceus.

Tectonics of the Sierra Leone Rise.

On tectonic and geomorphologic maps the Sierra Leone Rise is divided on the North and South Block Rises separated by Central Depression. The Guinea Saddle unites the Sierra Leone Rise and the Guinea marginal Plateau. On the east edge of the Rise the East Step is clearly distinguished. The east border of the Rise is bordered by a faulted steep projection of complex, configuration stretching north-north east.

The deep faults complicated the structure of the border between the Sierra Leone basin and the Eastern Step. The character of the faulting blocks is lystric.

In the Guinea Saddle the SW-NE and the NW-SE faults predominate. In the border zone of the Sierra Leone basin near the foot of the Rise the acoustical basement coincides with the surface of layer II, composed by a number of inclinal intensive seismic reflectors. It attests to hummocking of the crust here by horizonal movements.

The position of the Sierra Leone Rise structure corresponds to the border region between the differently aged Central - and South Atlantics. The different directions of the fracture zones to south and to north of the Sierra Leone Rise are evidence of the probability of horizonal movements in this region.

The Sierra Leone Basin is characterized by several types of structure in the upper part of layer II (Yunov A.,1981).

on the northern part of this basin series of lystric blocks are observed, positioned along the prolongation of the Guinea fault. The latter stretch under the Guinea Saddle and separate the Guinea marginal plateau from the floor of the Sierra Leone Basin.

The nature and origin of the latter are not fully known. Linear magnetic anomalies of the “spreading” type have not been distinguished here. Aside from the Guinean transform Fault, the fact has not been established here of the prolongation of the other transform faults sich as the Doldrums and Strakhov f.z., which “dive” under the Sierra Leone Rise. Much evidence of the non-riftogenic nature of the basement is distinguished in the Sierra Leone Basin. It is remarkable evidence of the primary continental crust which was submerged to the oceanic depth and converted into oceanic crust by the nonspreading processes. The arguments in favour of such a conclusion are as follows:

The north part of the Sierra Leone Basin is distinguished by numerous lystrical blocks with laminated structure. In the central part of the basin the basement consists of crustal blocks with ananomalously flat surface and the rough laminated structure of layer II top (Yunov A., 1981). The seismic velocities in the Third crustal layer beneath the basin are anomalously low - 6,2-6,7 km/see and comparable with velocities 6,1-6,5 km/see in the granite layer of the Liberian Shield (Sheridan, 1969).

In the different versions of the paleo-reconstructions it is dificult to observe the space for the Sierra Leone Rise between the North- and South American and African continents. Due to this the different paleogegraphic maps show this rise in different places.

Of course, if this Rise arose through the elevation of the oceanic lithosphere, there then no need to find the space for it in the paleoschemes of the Jurassic - Cretaceus. However, according to all models of reconstructions, the break-up of the Gondwana began not earlier than at the end of Jurassic - beginning of the Cretaceus. For the beginning of the upper Cretaceus age, the “continental bridge” must be present between South America and West Africa. It is possible that this “bridge”, if it existed, included part of the present Sierra Leone Basin, which developed as the result of the submergence of that ancient continental massif.

The borehole GU2B1, drilled near the edge of the Guinea marginal plateau, entered into the thick volcanogenic complex of the Barremian and Aptian sediments. This borehole shows the predominance of the continental- and shallow-water facies in the lower part of the upper Cretaceus. To the north of the Guinea marginal plateau, however, in the vicinity of the Cape Verde Basin on the shelf of Guinea and Bissau, in lower and upper Cretaceus the normal sea facies were predominant. According to this, between the regions of the shelfs of Guinea and Bissau. and of the southern part of the Guinea marginal plateau, there was some kind of barrier - the shallow water area or land “bridge” on the site of the present Sierra Leone Rise, and part of the Sierra Leone Basin.

It is possible to suppose that under the present Guinea marginal plateau and Sierra Leone Rise lies the continuation of the south - west branch of the Mauritanide fold belt . It is known from the Precambrian Reguibat massif that the linear zone of the Mauritanide fold - belt stretched to the south in the form of an arc on the territory of Guinea. There the metamorphosed upper Precambrian rocks divided into two branches. The first branch - the Kuluntu ridge - turn towards the south-west, the other - the Basaris ridge - stretched to the south. In the area between both branches the paleozoic sediments lie subhorizontally. The Mauritanide fold - belt system presents the broad folding belt of the low metamorphised rock. The prolongation of the Kuluntu fold belt stretched to the shelf of Guinea and probably to the Guinea marginal Plateau and the Sierra Leone Rise. Although the magnetic investigations data (Mc Master et. al, 1970, Jones, Mgbatogu, 1982) did not find extended magnetic anomalies here, using the data obtained the existence of short linear anomalies to the south-west can be demonstrated. In addition, some of the magmatic intrusions of the south-west stretches occur on the shorelines of the Guinea and Sierra Leone. The largest of intrusion occurs on Cap Verga and near the capital of Guinea - Konakri. These are the large laccolithes of gabbro and dunite framed by the old, but rejuvenated granites. Many dykes occur here stretched SW-NE with the absolute age 180+-10 -230 +-10 millions. This age corresponds to the time of the initial riftogenesis which led to the formation of the north subequatorial segment of the Atlantic ocean.

The map of the gravity anomaly of the World Ocean (W. Haxby, 1987, Sandwell, Smith, 1996) derived from SEASAT Radar Altimetry of the Ocean Surface shows that the distinct system of the subparallel breeches of the gravity fields stretched from SW to NE in the region includes the Sierra Leone Rise, Guinea marginal Plateau and adjacent parts of the ocean. These linear zones stretch to the continent in Bissau, Guinea and Sierra Leone. All of these may be evidence of the prolongation of the deep faulting zone from the continent to the south-west far in the ocean.

This zone of the deep faults is bound up with the fold belt Kuluntu and with many of the basic, alkali and acid intrusions on the shoreline and on land.

The configuration and character of some magnetic anomalies of the Sierra Leone Rise, the stretching of the main faults here and of the morphostructural features corroborates the southwest-north-east stretching. The entire Sierra Leone Rise also stretched in this direction. If the assumption of the prolongation of the fold - belt Kuluntu to Sierra Leone Rise is true, then all this probably also signifies that the structure of this Rise is primarily continental. This may signify that the Sierra Leone Rise, during the process of the forming of the oceanic basins to its north, south and west of, did not move far from its modem position in spite of spreading or other movements around it.

The crust under this Rise is reduced primary continental, thinned by deep processes. The thinning of the crust was followed by isostatic submergence, but not as great as in the neighboring oceanic basins.

The lystrical character of the step blocks on the edges of the Rise is not evidence of the rising primary oceanic crust which formed the structure of the Rise. Rather, it is evidence of the collapse of the edge blocks, which formed the deep oceanic basin. The arguments for constant deepening of the oceanic bottom are the facial contents of the drilled sediments, evidence of the large hiatus associated with the subaerial conditions between two of the laminated sequences and others. The forms which developed in subaerial conditions occur in seabottom topography as well.

All of the data may be evidence of the assumption that in preaptian time to the north of the Sierra Leone Rise a small oceanic basin existed, while to the south of the Rise begins the initial phases of the formation of the South Atlantic. On the place where Sierra Leone exists today, was the the large continental block which, in addition to the Sierra Leone Rise, included also the basement of northern parts of the recent Sierra Leone Basin and the recent Guinea marginal plateau.

This block was a part of the “continental bridge” between the South American and African continents.

The following processes of the formation of the South Atlantic oceans led to thinning, breaking of the crust, intensive volcanic activity on the edges of these blocks, on the Mid-Atlantic Rise and on the transform faults.

The sedimentations in upper Cretaceus on the Rise, where the sea bottom was higher than CCD (oxigenic carbonate level), was more intensive as compared to the Sierra Leone Basin because of the carbonate sedimentation.

The intensive volcanic activity formed the volcanic acoustic “basement”, which in places overlies the older sediments. The breaks on the edges of the Rise were synphased to deep processes which led to a reduction of the area of the Rise. The slow submergence of the Rise accompanied it and continued later.

From this position the structure of the Rise is close to the concept of the “microcontinent”.

The Structure of the Guinea marginal plateau based on reflection CDP data.

The Guinea marginal Plateau represents the triangular step-block on the continental slope of Bissau and Guinea. The surface of the slope forms two undersea terraces on the level of 900 and 1800 m inclined to the South-West. The southwest wedge - shaped edge of the plateau is complicated by a third faulted block - terrace with a depth of 2000-2500 m.

The southern slope of the plateau is very steep and complicated with irregularities, while the western slope is relatively gentle, From west to east along the southern slope the Plateau stretches for more then 350 km, from south to north -150-200 km.

The Guinea marginal plateau divides the Cape Verde and Sierra Leone oceanic basins, and is one of the great submarine plateaus of West Africa. Near the southern slope at a depth of 113 m the wildcat GU2B1 was drilled. It ended at a depth of 3351 m under the ocean floor in the Barremian (Promotion... 1987).

On the Guinea shelf from 1974 to 1986 the different companies made many of the reflection seismic profiles, but only in the eastern part. Only separate profiles transect the western deepwater part of the plateau (Promotion . . . . 1987).

The several seismic profiles made in the deepwater part of the plateau included the southern and western parts (de Ruiter, 1977; R. Mascle, M. Marinho, 1986; M. Marinho, 1985; K. Emery et.al., 1975). The western part of the Guinea Plateau and adjacent parts of the Sierra Leone and Cape Verde basins were investigated by Soviet geophysists. The main method was the reflection CDP accompanied by magnetic and gravity survey (Yunov, 1978, 1981,1987,1992). The northern part of the plateau and its transition to the Bissau shelf was investigated by seismic reflection profiles (M. Dumestre, F. Carvalho, 1985) (Plate 18). On the Bissau shelf a few wildcats holes were drilled. They discovered the indications of oil in Aptian and gas indications in the Cenomanian strata. Two wells ended in salt domes and the Triassic age of the salt was determined. These salt bodies are bedded in the submeridional rifting trough on the western part of the Bissau shelf. The trough stretched to the south in the marginal plateau. In the Guinea sector of the latter the salt bodies do not occur.

The complex geophysical investigations on the Guinea shelf allowed this to be divided into two regions. The eastern one is characterized by many faults and basic and hyperbaric intrusions along the three main. faults zones. The latter, prolonged on the shelf in Cap Verga, Katum peninsula and the mouth of the Fatala river, was investigated on land. Then these faults transected the large riftogenic trough stretched NW-SE. In the trough lie the paleozoic and other old strata, and in the axial part of trough are found the basic pre-upper Jurassic intrusions (V. Kozlenko et.al., 1990).

In the central part of the shelf are the thick sediments which overlap the large massif of the Precambrian granitization rocks.

To the west of the Guinea shelf is the Guinea marginal plateau, where the progradation prism of the sediments over the edge of the pericontinental mesosoic riftogen occurs (V. Kozlenko et. al., 1990). The “Petroconsultants” assumed under the marginal plateau the presence of the thick (5-6 km) Paleozoic sedimental sequence, overlapped by Triassic and Cretaceus strata (Plate 18, fig. 15-20 (Petroconsultants) (Promotion, 1987).

To the north of the marginal plateau lies the narrow Bissau shelf The structure of the latter is made up of the submeridional rift zone where the paleozoic sediments dips to 8-10 km. In the rifting zone occur the saltdomes from the Triassic salt layers. The slope is built of Jurassic, Cretaceous and Cenozoic strata with a thickness of 6 km (M. Dumestre, F. Carvalho, 1985) (Plate 18,fig. 16) On the south-western end of the marginal plateau the results of seismic refraction allow for us to suppose the presence in the basement of Paleozoic sediments - probable an analog of such sediments on the land of Guinea and Bissau (Sheridan et.al.,1969).

Our CDP investigations accompanied by the gravity and magnetic observations were conducted on the outer dipped part of the Guinea Plateau between 1974-1982.

The first profiles of 1974 and 1975 distinguished the complicated structure of the outer part of the Plateau and the different types of its connections with the Cape Verde basin to the west and with the Sierra Leone basin on the South. (Plate 18).

The later CDP seismic investigations of 1981 and 1982 yielded more data about the structure of the upper part of the thick sedimental sequence of the Plateau. Of greatest interest are the lines 8001 and 8002 which cross the outer southwestern part of the Plateau from the Cape Verde basin to the Sierra Leone basin.

The main reflectors distinguished in the Cape Verde basin identified in borehole 367 IPOD clearly traced more than 350 km up to the western slope of the Plateau. (A. Yunov 1987, 1989-92,1991). The best reper is the reflector Ac correlated to cherts of the Eocene. In the direction of the edge of the Plateau this reflector gradually lost its significance and was replaced by another. The latter, named Au, and correlated to the large pre - upper - Oligocene - pre- miocene regressive phase with absolute age 24 - 30 mln. years, connected with the relative drop in sea level.

The main reflector A* correlated in borehole 367 to the top of lithological sequence 4a. The latter was formed by upper - aptian lower Turonian black lignitic shale or by the dolomitic layer in the upper part of this shale. This reflector traced continuously to the western slope of the Plateau. This reflector probably connected with the intracenomanian hiatus and correlated to the intensive regressive phase with an absolut age of 97 mln of years, or to pre-santonian unconformity (90 min. of years) according the model of B.Hag et al (1988).

The main reflector C in borehole 367 is identified as the top of the carbonate sequence. The latter is represented by upper - tithonian- hauterivian grey limestone and marl interbedded with black shale.

The intensive reflector C1 0,1 sec below the C, in borehole 367 is identified with the top of the reddish-brown and grey limestone bedded on the surface of the layer II basalts. Analogous limestone in the West Atlantic is Tithonian-Kimmeridgian (137 million years old).

In the direction of the western slope of the Plateau both reflectors - C and C1 can be clearly traced. In some places the C1 reflector connnects to the reflection from the top of basalts. Under the Plateau’s slope both reflectors connect.

The thickness of the Jurassic sediments below reflector C in the approaches to the Plateau are decreased on the rises of layer II and increase in its depressions. Inside the Jurassic sequence the new reflectors appear. They may be correlated to their analogs in the West Atlantic - to the mid-lower Jurassic reflectors J2 and J3 (K. Khitgord, J. Grow, 1980). Beneath the upper part of the continental rise and the slope of the marginal Plateau the thickness of the Jurassic sediments increases to 3 km and more. They are probably represented by limestone and in some places by reef facies.

The thickness of the seismic sequence between reflectors C and A* (lower Cretaceons) in borehole 367 is 260 m. In the direction of the marginal Plateau the thickness of the latter increases to 2.6-3.2 km beneath the western slope. Inside the lower Cretaceons sequence are distinguished the reflectors A** and A***. On the global cyclic model of relative change of sea level (B.Hag et al., 1988) both of these reflectors may be correlated to the regressive phases between the Aptian and Albian or inside the Aptian (108 mln or 112 mln years) and in lower Valanginian (126 mln years) respectively. Both reflectors - A** and A** * are gradually dissipated beneath the slope.

In the direction of the marginal Plateau the thickness of the Berrias - lower Neocomian between reflectors C and A*** increases sharply - to 1.5-2.0 km. In the upper part of this sequence beneath the continental rise is observed the distal delta facies - the clinoforms, connected with the lower Valanginian regressive event (126 mln years). The sequence between Valanginian to Albian (between reflectors A*** and A*) is represented in its middle and upper parts by black shale, which increase in thickness towards the continental rise. These facies are replaced here by clay and sandy clay. In the middle and upper parts of the section here low thickness clinoforms are observed. Beneath the upper continental rise the thickness of this sequence decreases, and connects to burial paleostep of the upper Jurassic - lower Cretaceus continental slope. The clinoform bodies are probably connected here with the Aptian (near 112 mln years) and upper Albian or Cenomanian (98 or 94 mln years) regressive events.

The thickness of the upper Cretaceons-Paleogene sequence between reflectors A* and Ac represented by chalk. On the Guinea marginal Plateau, the surface of unconformity (reflector A) connected with the pre-upper Oligocene regressive phase truncated the Eocene and older sediments.

The western edge of the Plateau is complicated by many faults. In the northern part of the western edge of the Guinea plateau on the travers of many rivers of the land of Bissau, in the upper part of the section the rough erosional unconformity is observed, Thi unconformity sinks steeply to the north to the depths more than 2 km below the bottom and truncated the upper Cretaceons carbonate sequence here (Plate 18,fig.15,16,17,18). The broad - more than 50 km - buried erosional valley is filled by a thick clinoform sequence. This clinoform sequence with top discordance is overlaid by a low thickness of quarternary sediments and is connected with the paleodelta sequence of the system of the paleorivers Geba-Orubal, Monsoa, Bololo. This sequence was probably formed synchronously to the pre-upper-oligocene- pre miocene regressive phase of the global relative decline of the Ocean level (29-30 mln years ago). The modem delta of these rivers is located 200 km to the east.

on the CDP profile 8001, which transected the deep part of the Plateau, some reflectors are distinguished. The correlation of the latter to reflectors of the adjacent basin is diffilcult because of faults which truncated the western and especially southern slopes of the Plateau. On the deep part of the latter some seismic sequences divided by the unconformities are observed. The upper Quarternary seismic sequence 100-200 m thick lies horizontally. Below it lies the thin (200 -300 m) clinoform sequence, probably of Miocene age, with the pre-upper-oligocene unconformity (reflector Au). The clinoforms gently slopes to the South, to the Sierra Leone basin.

Below the unconformity lies thin (100-200) horizontally the carbonate sequence sloping towards reflector A*. Below the latter, corresponding to the Intracenomanian regressive phase and to rough unconformity, lies a carbonate sequence of more than 1 km thickness near the western slope of the Plateau and about 200-300 m from its southern slope. In the meridional sections this sequence is horizontally laminated, and in the northern part of the profiles it lies below reflector A*. On the seismic profiles in NE-SW directions the unconformity on top of this sequence is well distinguished (Plate 18,fig. 19). It is important to note that the beds are inclined towards the north-west, towards the Cape Verde basin. The angular discordance is more distinct in the area near the southern slope of the Plateau. Here the beds of this sequence are folded and faulted, and according to the configuration of the seismic reflection, are carbonate reef fascies.

J. Mascle and M. Marinho (1986) assert the intrusive nature of folded sequence along the southern edge of the Plateau. Many features, however, show evidence contary to this point of view.

The character of the gravity and magnetic fields of these structures is not typical of intrusive bodies. Laminated structure is visible here, and these lamina gradually change to the flat carbonate sequence of pre-cenomanian age in the central part of the Plateau. The value of the seismic velocity in these beds is near 5000 m/sec on the CDP data (A. Yunov, 1987). All of this speaks to the reef carbonate nature of this bodies.

The age of the base of the lower clinoform sequence is not clear. On the seismic section of profile 7508 the subhorizontal character of this boundary is visible. Below the latter the laminated sequence, folded under the southern edge of the Plateau, in the north-east direction loses the inchoations of folding, increases in thickness, and submerges.

The correlation to boreholeGU2B1 unfortunately is difficult. This borehole is located too far from this point, and in the geological features of the region near the this bore hole are peculiar. The borehole is located 320 km to the east and stopped drilling at the depth of 3355 m in Barremian. According to this borehole, the continental facies of the Barremian with the volcanic breccia are more 1100 m in thickness and are overlaid by a 670 m sequence of the basalts, volcanic breccia with the argillite shale, and sandstone of the Aptian-Albian. Above the latter lies the continental facies 1600 m in thickness of the Cenomanian-Maastrichtian - the sandstone and shale interbedding with the dolomitic and lignite.

In this latter section occurs the unconformity in the base of the Cenonian. After the rough unconformity and sign of erosion, the continental facies of the Cretaceus is overlaid by the marine mostly carbonate sediments of the Eocene-quarternary age. Beginning with Oligocene, the proportion of the clay facies increases. The most significant features of this section are the continental character of the ('setaceous sediments, the intensive basal volcanism of the Barremian, Aptian and Albian, the presence of the pre-cenomanian, pre-cenonian, pre-eocene unconformities, and the great thickness of the marine facies in Eocene and Neogene on the outer shelf of Chines 300 km to the east of the marginal Plateau (Plate 18, fig. 21.22).

In comparison with the Mesozoic section of the Bissau shelf, here, on the outher shelf of the South Guinea, the facies and the thickness are very different. The Barremian-Aptian-Albian section of the Bissau shelf is characterized by the marine and mainly carbonate facies, and by lack of volcanism. The thickness of that is twice that of the synchronous sediments of the South Guinea shelf. The Cenomanian-Turonian strata of both regions are characterized by their similar facies and thickness. The thickness of the marine Cenonian of the Bissau shelf is twice that of the Guinea southern shelf. The Cenozoic parts of the sedimental sections of both regions are similar in thickness, excluding Paleocene. The latter, on the Bissau shelf, is terrigenous and the thickness of Pliocene-Quarternary is here more that in the area of borehole GU2B1.

Further north, in the region near Dakar and in the South of Senegal, the lower Cretaceous sediments are represented by marine clay facies of great thickness. The deep-water basin here sagged deeply in lower and upper Cretaceous , having undergone inversion in the time between Paleogene and Neogene. The thickness of Neogene sediments here is only a third of that of the north Guinea shelf. On the other hand, on the south-eastern part of the Guinea shelf the predominance of continental conditions of the sedimentation in lower and upper Cretaceus was changed on the Paleocene-Eocene boundary by deepening and transgression. Thus the southern edge of the Guinea Plateau - as the orographic structure on the edge of the Sierra Leone basin - like the basin itself - must be younger that the western slope of the marginal Plateau and the Cape Verde basin. From this position the presence is clear of the continental facies on the southern edge of the Guinea shelf. This allows for an understanding of the reason for the north-western inclination of the prograding of the “lower” clinoform sequence of the deep parts of the Plateau in the direction of the Cape Verde basin in Neocomian, Aptian and Albian times. The beginning of the formation of the Sierra Leone basin and of the southern slope of the Guinea marginal Plateau corresponds in time to the general riftogenic process in Neocomian-Aptian. The presence of the basic magmatic bodies drilled by borehole GU2B1 is evidence in favour of this supposition.

The southern slope of the marginal Plateau is complicated by the lystric faults. Some of the latter, buried and half - buried by sediments, occur in the adjacent part of the Sierra Leone basin (A. Yunov, 1978, 1981). The maximum thickness of the sediments, near 2000 m., occurs in the deepest rear sections of these lystric blocks -

The age of the basal synrift sequence here is not known. The deepest reflectors here are distinguished as A** - the Aptian-Albian.

On this basis, the acoustically transparent synrift sequence in this part of the Sierra Leone basin is lower-cretaceous, probably upper Neocomian. The Jurassic strata represented in the Cape Verde basin is not observed in the Sierra Leone basin. This circumstance confirms the relatively young age of the Sierra Leone basin and of the southern slope of the marginal Plateau. Only after the formation of the abyssal Sierra Leone basin in the region around borehole GU2B1 did deepening take place, starting with the Eocene time. The transgressions and the appearance of marine facies, and erosion on the south-western part of the marginal Plateau were accompanied by the transportation of sediments into the deep basin, with the formation of the “upper” clinoform sequence. The transportation of the sediments took place to the south, to the young Sierra Leone basin. The relatively intensive drop in the level of the Atlantic ocean on the boundary between the Paleogene and Neogene was the cause of the intensive formation of the deltaic clinoform sequences in the direction of both basins - Cape Verde and Sierra Leone.

The presence of the thick Paleozoic sedimentary sequence under the inner and central parts of the marginal Plateau and under the Guinea shelf is assumed by R. Sheridan et al. (1969). This was confirmed by the geophysical investigations on the Guinea and Bissau shelfes. However, so far there is no data regarding the presence of Paleozoic sediments under the outer part of the marginal Plateau.

The presence in the past of land, or of shallow basins, in the region of the southern edge of recent marginal Plateau and in the adjacent part of the Sierra I.cone basin, and the folding of the lower Cretaceus sediments on the southern edge of the Plateau allow us to propose the presence here of the tectonic “barrier”.

Such a barrier may be the prolongation on the shelf of the Kuluntu fold belt - the south-western branch of the fold-belt Mauritanide. The latter in the form of the broad belt of the weakly metamorphised rocks, are faulting from the west to east in the form of scales. The fold-belt stretched in an arc from the Precambrian Reguibat massif of Mauritania in the north to the Guinea shore in the south.

All the data shows evidence of the past presence of the “continental bridge” connecting West Africa and South America in Aptian time and earlier. The results of dredging during the 7-nth cruise of the R/V “Akademik Nikolaj Strakhov” show the presence of the continental Mesozoic sediments on the Equatorial segment of the Mid-Atlantic Ridge, including coal argillite, probably similar to the lignites of the Guinea shelf in borehole GU2B1 (G.Udintsev et al., 1990). The age of the palynologic remnants of the dredged coal argillite - upper Cretaceous-Paleogene - is not far from the age of the rocks containing the lignite in borehole GU2B1 (Cenomanianmaastrichtian), although the distance from the point of dredging to the borehole is about 1800 km. The numerous finds of rocks which are continental in genesis at the point of dredging leads G. Udintsev et al. to write that on the flanking plateau of the Equatorial segment of the Mid-Atlantic Ridge “up to the end of the (’setaceous there existed platform conditions with many islands and shallow water” (G.Udintsev et al., 1990).

Chapter 8

Gravity Field of the Equatorial Segment.

J. Makris, U. Vogt, V. M. Golod, T. Funk, V. G.Udintsev

During the 11-th cruise of the research vessel "Akademik Nikolaj Strakhov", gravity data were collected on the Mid-Atlantic Ridge between 2°30'N and 5°N. Gravity data cover the crest of the Mid-Atlantic Ridge and the ridge flanks up to 200 kilometers to the east and west from the spreading center axis. As a result of the interpretation of this data set were defined a gravity field expression of the second order segments and discontinuities not only in the crestal part of the Mid-Atlantic Ridge but also in its flanks. Unfortunately, the survey did not extend to the active part of the Strakhov Fracture Zone (f.z.) (only multibeam bathymetry data are available for this area).

Most of the geophysical surveys in the Mid-Atlantic Ridge were carried out over the lithosphere, which is not older than 2 million years (Romevaux et al., 1994). These studies show that between any two large transform faults the Mid-Atlantic Ridge consists of a narrow crestal volcanic zone which is morphologically divided into second, third and fourth order segments. The segments are hounded by non-transform discontinuities with up to 20-30 kilometers offset of the spreading center segments. But the gravity expression of these second order discontinuities and second order segments off-axis of the Mid-Atlantic Ridge has not yet been reported. This study is thus important in that it reveals the extent and appearance of the second order segmentation in the off-axis parts of the Ridge in the gravity field.

Three-dimensional gravity anomaly studies carried out by Kuo and Forsyth (1990), Lin et al. (1990), Morris and Detrick et al. (1994) have shown that the Mantle Bouguer anomaly mover slow spreading centers have closed-contoured gravity lows, separated by relative gravity highs. These features are interpreted as a consequence of crustal thickness and/or density variation in the crust and/or upper part of the mantle. Such an interpretation led Kuo and Forsyth (1990) to the concept of concentrated or focused mantle upwelling beneath the centers of spreading segments. In this paper an analysis of gravity anomalies is presented for the central zone of the spreading center, as well as for the flanks of the Equatorial Segment (ES) of the Mid-Atlantic Ridge (MAR). Mantle Bouguer Anomaly maps were generated and analyzed over a broad area of the Mid-Atlantic Ridge to investigate the spatial distribution of gravity anomalies, their relations with the morphological segmentation of the Mid-Atlantic Ridge, and the evolution of the second order segments.

The gravity data acquired during this survey were analyzed and interpreted in terms of crustal thickness variation. The crustal thickness variation, according to Kuo and Forsyth (1990), can be determined from the gravity field, though it is not as detailed and accurate as seismic refraction experimental data. Gravity data provides a method to evaluate 2-D lateral variations of crustal thickness, and this is a distinct advantage in comparison with seismic refraction data. No seismic refraction experiments have been carried out in the study area. Thus, the interpretation of the gravity field cannot be calibrated by the seismic refraction as was done in major gravity field interpretations from other parts of the Mid-Atlantic Ridge.

Data Description and Initial Data Processing

The gravity survey was carried out on board the research vessel "Akademik Nikolaj Strakhov" during its 11-th cruise. The ship was equipped with a Global Positioning System navigation receiver with the spatial accuracy of approximately 25-50 meters. The gravity meter for the survey was generously provided by the Department of Geophysics, University of Hamburg, Germany. The gravity meter was supplied with the interface to the IBM PC with which the raw data were recorded. Gravity data have been corrected for the drift of the instrument. In each port the gravity meter was calibrated. The drift of the instrument was approximated by a linear fit. Calibrations were carried out in Hamburg (Germany), Madeira, Recife (Brazil), and once more in Hamburg. The output of the system was a binary file which contains Greenwich time, gravity meter readings in mGal and geographical coordinates.

The raw data were corrected for latitudinal, tidal and Eotvos corrections. The data were divided into profile files. The data for 20 minutes intervals after the vessel turn were excluded from each profile because of instrument stabilization problems. A despiking of the data was carried out manually before the Eotvos and latitudinal correction. Corrected gravity data were awn-aged by a 5-minute interval using the "moving window" algorithm in order to create the Free Air Anomaly Map.

The cross-over errors were the main concern during the construction of the Free Air gravity Anomaly. The ship-track map (Plate 2, 23) shows that this data-set was rather poorly designed because of the lack of the regional linking ship-tracks that would have crossed all the ship-tracks. Nor did a search in the NGDC data-set provide any crossings with the gravity data. The cross-over error correction of the gravity data had to be completed by polygons which have local linking crossings. The cross-over errors are generally a combined result of non-linear drift of the gravity meter, cross-coupling effect in the gravity meter (Talwani, 1966) and incorrect calibration of the instrument, Minimization of the cross-over errors was carried out using the Prince and Forsyth (1984) approach which implies the least mean square minimization of the cross-over errors by shifting Free-Air Gravity anomaly on the ship-tracks by a constant value. Then all polygons and "hanging" straight segments were shifted and normalized using the Sandwell's (1994) Satellite Altimetry Free Air Anomaly grid. A series of synthetic Free Air Gravity profiles were generated using the Sandwell's (1994) Satellite Altimeter Free Air Anomaly grid in order to shift the ship-borne gravity data. The locations of the synthetic gravity profiles were chosen using the criteria of the minimum gravity field gradient and at some distance (> 30 min. distance) from the ship-track turning points. After all applied correction, the root mean square of the remaining cross-over error is 4.7 mGal.

Gridding of the corrected data was carried out using the least mean square method with the search radius of 5 kilometers and grid spacing of 1 kilometer; sharp weighting of the data points depending upon the distance to the grid node was applied. Similar gridding parameters were used by Kuo and Forsyth (1990) when they were processing the data-set which consisted of data collected during different cruises on different ships with different gravity meters. The similarities with this data-set allowed the author to use their approach, The contour interval for the resulting maps is 10 mGal, which is roughly twice the root mean square of the remaining cross-over error.

Free Air Anomaly Map

Plate 11 presents the combined ship-borne Free Air Gravity Anomaly map of the study area and an extracted Sandwell's (1994) Satellite Altimeter Free Air Anomaly map. Color changes correspond to Sandwell's (1994) Satellite Altimetery Free Air anomaly with contour interval of 10 mGal. Black contour lines represent the ship-borne Free Air Gravitty Anomaly infield. This combined plot shows reasonably good fit of its two components. The Free Air anomaly of both grid system also show good correlation with the major bathymetric features in the area: they outline the spreading center and Strakhov f.z., its active part and passive branches. The two major differences between the two Free Air gravity anomaly gird systems are: a) the amplitude of the anomalies at the spreading center axis; b) presence of the higher frequency component in the ship-borne Free Air gravity anomaly map in comparison with the Sandwell's Free Air gravity anomaly map.

The spreading center is very well defined in the ship-home Free Air gravity anomaly map. It has a relative anomaly amplitude of-70 mGal, while on the Sandwell's Free Air gravity anomaly map the maximum relative anomaly does not exceed 40 mGal. Such differences are highly predictable, because of much greater distance from the source of the gravity signal in the case of satellite altimetry, and relative proximity to the source in the case of the ship-borne gravity. Appearance of the higher frequency signal like local closures and undulations of the contour lines on the ship-borne Free Air gravity anomaly map is explained by the same reason. The orientation of the main components, the spreading center, the Strakhov f.z. traces, are the same on both grid systems. Ship-borne Free Air gravity anomalies perfectly outline the second order segmentation of the spreading center. The rift mountains located at the senters of the second order spreading center segments are expressed as local gravity highs with relative amplitude ranging from 80 up to 140 mGal. Maximum relative Free Air gravity anomaly amplitude occurs at the intersection of the spreading center and the Strakhov f.z. The inner comer high is represented by local gravity high, and the nodal basin is represented by local gravity low.

The ship-borne Free Air gravity anomaly field has some features which were not clearly visible on the bathymetry map. These structures or lineaments have NNE-SSW orientation. They are located between the Strakhov f.z. and 4NS-3 second order discontinuity. These features are expressed as a sequence of NNE-SSW trending elongated direction local gravitty highs and low closures in the ship-borne Free Air gravity anomaly field. Such a gravity field pattern is visible and prominent starting at 60 kilometers from the spreading center axis and is observed up to 180 kilometers away from the spreading center axis. The amplitudes of these features progressively diminish in an off-axis direction. The maximum relative amplitude of these features is -80 mGal, their minimum relative amplitude is less than 30 mGal. These anomalies are also present on the Sandwell's Free Air gravity anomaly field, butt not as clearly visible as on the ship-home Free Air gravity anomaly field. The spreading center segment located to the north from Strakhov f.z. is characterized by a combination of forms with orientation parallel to the spreading center and parallel to the Strakhov f.z.

Surprisingly, the 4NS-3 second order discontinuity is well defined on the Sandwell's Free Air Gravity Anomaly field. It is represented by a long linear trough similar to those of a transform fault with a well defined nodal basin and aseismic fracture zone traces up to 600 kilometers away from the spreading center axis. Maximum relative amplitude of the Sandwell's Free Air Gravity Anomaly field associated with the intersection of the spreading center with the 4NS-3 second order discontinuity is more than 80 mGal, and it is gradually decreasing to the west from the spreading center axis.

At 250 kilometers to the west from the spreading center axis the relative amplitude of the Sandwell's Free Air Gravity Anomaly field at the 4NS-3 second order discontinuity off-axis trace is still up to 30 mGal. Relative amplitudeoftheship-home Free Air Gravity Anomaly field at the same site is even higher -~60 mGal. It is thus obvious that 4NS-3 second order discontinuity is very sharply defined in the Free-Air anomaly field as a transform-fault-looking feature. The orientation of the 4NS-3 second order discontinuity off-axis traces is also very unusual: on the Sandwell's Free Air Anomaly map they are not parallel to the orientation of the neighboring fracture zones (Strakhov f.z. and Saint Paul f.z.). In addition, at 270 kilometers segments of the eastern and western limbos of the 4NS-3 second order discontinuity off-axis htraces do not fit a straight line: the angle between the limbos is ~160III. At ~270 kilometers distance from the spreading center axis the orientation of the 4NS-3 second order discontinuity off-axis traces starts being roughly parallel to those of the Strakhov f.z. and Saint Paul f.z.

According to Schilling et al. (1994), the 4NS-3 second order discontinuity serves as a structural barrier of the Sierra Leone Mantle Plume. Schilling's et al., 1994 plate-tectonic reconstructions suggest that during 48-11 Ma Sierra Leone Plume was captured by the westwardly migrating Mid-Atlantic Ridge, which "acted as a sink". Than during 11-0 Ma Mid-Atlantic Ridge axis jumps back close to the 1 the present 1.7°N anomaly. Though there is no doubt about the existence of the 1.7°N geochemical anomaly and its connection with the Sierra Leone Mantle Plume, the timing of the ridge jump (11 Ma) is questionable. The Ridge jump probably should be associated with the bending of the 4NS-3 second order discontinuity off-axis traces (4NS-3 second order discontinuity possibly was a transform fault before the ridge jump). Westward jump of the Mid-Atlantic Ridge significantly diminished the offset of the "paleo" 4NS-3 Transform Fa, since the ridge jump the "paleo" 4NS-3 Transform Fault rejuvenated as a second order discontinuity. Bending of the 4NS-3 second order discontinuity off-axis traces occurs roughly symmetrically at 270 kilometers from the ridge axis on both sides of the Mid-Atlantic Ridge which corresponds to ~25 Ma (using 1,3 cm/year spreading rate). Thus, the author suggests that the ridge jump towards the Sierra Leone Mantle Plume occurred -25 Ma. Since that time the NS-33 second order discontinuity was prominently moving northward due to the influence of the Sierra Leone Mantle Plume captured by the Mid-Atlantic Ridge segment between 2°30'N and Saint Paul f.z. This suggests that first order discontinuities can be transformed into second order discontinuities due to the major ridge jumps. The crustal basement relief is the main source of the Free Air anomaly field. But the other important source of the gravity signal is the crustal thickness variations and crustal density variations. If the feature is expressed as a transform fault in the Free Air Anomaly field and as a second order discontinuity in the bathymetry, then this particular feature can be interpreted as a major crustal thickness boundary (which is supported by Schilling et al., 1994 crustal thickness calculations based on the geochemical data).

Thus the main source of the segmentation of the slow spreading centers is the shape of the 3-D focused mantle upwelling centers and mantle plumes that "feed" and outline the spreading center. Behavior of 3-D focused mantle upwelling centers can be different in time and space. In the Case when 3-D focused mantle upwelling centers are far enough from each other and if they evolve separately, the transform fault is formed. If these mantle diapirs have the common "root" and are simple multiple "arches" of larger mantle diapir not offset by a fracture zone, then the second order discontinuity is formed at the saddle between these "arches". Evolution of the crest-mantle interface and the mantle diapirs causes the evolution of the second and first order segments and their discontinuities. Substitution of the discontinuity order can also take place, as it seems to have occurred with 4NS-3 second order discontinuity. Major crustal and magmatic reorganizations of the spreading center in the vicinity of the 4NS-3 second order discontinuity occured approximately 25 million years ago: this is a time mark of the 4NS-3 second order discontinuity off-axis trace bending in the Free Air gravity anomaly field.

Bouguer and Mantle Bouguer Gravity Anomaly Maps

Bouguer and Mantle Bouguer gravity anomaly maps (Plate 12) were constructed using the approach of Kuo and Forsyth (1988). The water/crest interface influence on the gravity signal was modeled using density values 1.03 gram/cub. cm for the water layer and 2.73 gram/cub. cm for the crust (Kuo and Forsyth, 1988). The first correction that was applied to the Free Air Gravity anomaly gird in order to construct the Bouguer Gravity Anomaly map was the 3-D correction that "filled" or substituted the water layer with a single layer having crustal density. Kuo and Forsyth (1988) stress that the numerical difference between the values with a stratified model (densities increasing from 2.4 gram/cub.cm to 2.95 gram/cub.cm and a single-layer crustal model with the average density of 2.73

gram/cub. cm are on average less than 1 mGal. In order to produce a three-dimensional correction, the 3-D gravity processing software, generously provided by Amoco, was used.

The 3-D gravity correction software uses a 2-D Taylor series expansion of the 2-D Fourier transform of the imput bathymetry grid with the chosen density contrast and grid spacing values.

Bouguer correction accounts for the gravity signal from the water-crust interface, but in order to account for the gravitational attraction from the Moho a Mantle Bouguer anomaly is generally calculated for the marine gravity datasets. For this correction the density of mantle was assumed to be 3.33 gram/cub.cm (Kuo and Forsyth, 1988). six kilometers thickness of the crust was used in the Mantle Bouguer correction. Similar calculations were carried out using the gravity 3-D processing software, but 6 kilometers were added to the values of the bathymetry grid and the density contrast was 3.33-2.73 =0.6 gram/cub.cm.

Constructed Bouguer and Mantle Bouguer Gravity Anomaly maps were based on the numerical procedure, which included the “grid-to grid” arithmetic operations. The bathymetry grid, used for 3-D correction has the same grid spacing as the Free Air Anomaly grid. The back-interpolation of the corrected grid to the original ship-tracks was not carried out (this approach was described by Sempere et al., 1992) because the grid spacing of the bathymetry grid is much smaller than the Free Air Anomaly grid.

According to Morris and Detrick (1991) the Mantle Bouguer Gravity Anomaly represents the gravity signal from contributors from both crustal and mantle density anomalies. Thus, with the assumption of constant crustal thickness of 6 kilometers, any variations of the residual anomaly can be interpreted as spatial variations of crustal thickness and/or lateral crustal and/or mantle density variations. If the crustal thickness remains constant and there is no isostatic compensation, changes of the Mantle Bouguer Gravity Anomaly can be interpreted as deviation of the constant density assumption. Morris and Detrick (1991) also showed that even if some bathymetric features of the spreading center are compensated or partly compensated, the Mantle Bouguer Gravity Anomaly will have an anti-correlation with the bathymetry, which may be interpreted as an impact of low density crustal root effect (Sempere et al., 1992).

The Bouguer Gravity Anomaly map and the Mantle Bouguer Gravity Anomaly map are presented on Plate 12. The Mantle Bouguer Gravity Anomaly map shows a variation of the gravity signal from -30 to 110 mGal.

Unfortunately, the Strakhov f.z. area was not adequately covered by the gravity measurements. Using data collected on several ship-tracks crossing the active part of the Strakhov f.z. a schematic map for this portion of the study area has been constructed, though its reliability is rather poor. The trough of the Strakhov f.z. is expressed on the Mantle Bouguer Gravity Anomaly map as a relative gravity high. A significant gravity anomaly amplitude is associated with this transform fault maximum relative amplitude is ~110 mGal (this value corresponds to the eastern portion of the transform fault. The anomaly gradient at the Strakhov f.z. is much higher at the western intersection of the fault and the spreading center segment than at its eastern intersection, where the maximum anomaly amplitude does not exceed 40 mGal. A large positive Mantle Bouguer Gravity Anomaly at the Strakhov f.z. can arise from the extremely low crustal densities beneath the transform fault. Such an explanation was proposed by Morris and Detrick, 1991 for the Kane f.z. area, where they overestimated crustal thickness beneath the Kane f.z. as a result of anomalously low crustal densities. Detrick et al., 1993 also stressed that generally low crustal densities are rather typical beneath transform fault areas.

The East-West orientation of the anomaly associated with the Strakhov f.z. can be traced only for ~30 kilometers to the west from the western Ridge-Transform Intersection and only for ~15-20 kilometers to the east from the eastern Ridge-Transform Intersection. A broad area between 32III50’W and 33III10’W does not bear any traces of east-west oriental anomalies on the traverse of the Strakhov f.z.

To the west from 33III10’W the Strakhov f.z. trace is noted and it is associated with the sharp step-like gravity field structure. The strike of this form corresponds to the off-axis trace of the Strakhov f.z. To the north from it a Mantle Bouguer gravity low is located, and to the south - Mantle Bouguer high. Such feature can be interpreted as a juxtaposition of two blocks with significantly different crustal thickness the southern block has higher thickness than the northern. Such notion corresponds to the general fracture zone model: the northern block is younger than the southern one by ~6 million years. The amount of offset counted by the Mantle Bouguer gravity anomaly map is smaller than the actual 4IIINorth transform fault offset by ~20 kilometers. The measurement was performed using the ~30 mGal contour along the western off-axis trace of the Strakhov f.z.. Such a difference between the offsets is explained by smoothening of the gravity field signal as a result of heat conductivity between blocks; also, in theory, the gravity effect of the deeply buried step-like feature is smooth too. The juxtaposition of two different age blocks with the fracture zone aseismic trace as a boundary means that two blocks with different crustal thickness are juxtaposed. Such geometry can be approximated by a semi-infinite horizontal slab with high density contrast. According to Nettleton (1976), the gravity signal from such feature has a smooth profile with two asymptotical limbs, which is very close to the picture observed along the Strakhov f.z. off-axis traces area .

Such step-like geometry of the gravity field cannot be seen in the areas located close to the ridge axis, where it is overprinted by the influence of the spreading center. With the distance from the ridge axis the relative value of the crustal thickness difference between blocks is also dramatically diminishing, and thus the gravity signal, arising from this effect is also overprinted.

The relative gravity lows with the shape of “bull’s eye” are vVery important features that dominate the Mantle Bouguer Anomaly image at the crestal part of the Mid-Atlantic ridge in the area studied. This pattern of gravity lows and highs corresponds to morphological second order segmentation of the spreading center, which was discussed above. The centers of the second order segments, picked at the bathymetry map, approximately correspond to the centers or central areas of Mantle Bouguer Gravity anomalies. Thus second order morphological segmentation of the slow spreading centers can be interpreted as an effect of significant crustal thickness and/or crustal density variation, A crustal thickness variation plot along the spreading center axis in the study area is presented in Plate 22, figure 3.5. Crustal thickness for this plot was calculated using the assumption that the Mantle Bouguer Gravity anomaly is only an effect of crustal thickness variations. Figure 3.6 presents combined bathymetric and Moho depth profiles along the spreading center in the study area. Moho depth was calculated using the data presented on fig. 3.5 and axial depth profile. On this figure we can see that the Moho relief amplitude is at its maximum near the Strakhov f.z. In general, crustal thickness is significantly diminishing towards the Strakhov f.z. Also figures 3.5 and 3.6 suggest that the crustal thickness variations have a third order segmentation pattern superimposed on second order forms.

Second order discontinuities are expressed on the Mantle Bouguer Gravity Anomaly map as local gravity highs (in comparison with adjacent second order segments). The extent or length of these second order discontinuity off-axis traces differs from ones on the bathymetry map. For example D4N-1 second order discontinuity can be traced much farther east and west on the Mantle Bouguer Gravity Anomaly map than by the bathymetry map (the difference is ~30-40 kilometers. On the contrary, the D4N-2 second order discontinuity’s expression on the Mantle Bouguer Gravity Anomaly map is similar to that expressed on the bathymetry map.

The traceable width of the Mantle Bouguer Gravity Anomalies over the second order segments also differ from one segment to another in a non-systematic manner. The gravity anomaly over the: D4N-2 second order segment is the widest in the study area: its size in the east-west direction is at least 180 kilometers. The width of the gravity low over the 4NS-3 second order segment does not exceed 90 kilometers, while the anomaly over the 4NS-1 second order segment is at least 70 kilometers wide. The differences in width of the Mantle Bouguer Gravity Anomaly at different second order segments can be interpreted, following Rommevaux et al. (1994), as a consequence of the different geological history of these segments. Rommevaux et al. (1994) suggest that the width (in the direction perpendicular to spreading center axis) of the anomaly closure at the second order segment is proportional to the longevity of the particular segment. In such constraints, the D4N-2 second order segment seems to be the oldest in the study area.

It is very important that the spatial size of the gravity low over the second order segment does not have any correlation with the distance to the closest transform fault (in our case Strakhov f.z.). One can see that the 4NS-1 second order segment gravity low (closest to the Strakhov f.z.) has a much shorter wavelength than the D4N-2 second order segment anomaly (the center of the D4N-2 second order segment is located -60 kilometers to the north from the Strakhov f.z.). On the Mantle Bouguer anomaly map one can see this non-systematic (in respect to the distance from the Strakhov f.z.) variation of the anomaly size over different second order segments. This may indicate that the shape of the gravity anomaly, and thus the crustal thickness is highly dependent on the temporal history of the second order segments. Thus, each of these second order segments have a unique evolution, but their superposition provides the constraints for the first order segmentation: large first order segments necessarily consist of second order segments.

Another important observation of the Mantle Bouguer Gravity anomaly is that a significant gravity low closure is located directly to the north from the northern edge of the 4NS-1 second order segment just opposite the Strakhov f.z. Relative amplitude of this closure is even higher than the adjacent 4NS-1 anomaly (up to 50 mGal). Although the lateral size (~30 kilometers) of the anomaly closure is much smaller than those of the observed anomalies over the second order segments. Such a feature can be interpreted a significant density fluctuation in the 5 million year old crust. Such density variation may be caused by the proximity to very young and hot spreading center, or by tectonic activity in the Strakhov f.z. Similar effects were described by Kuo and Forsyth (1990) for the MARK area in their combined interpretation of gravity and seismic refraction data. Seismic refraction profile in the area near the transform fault presented evidence of major density contrasts in the lower crust and upper mantle. Kuo and Forsyth (1990) stressed that sharp high-gradient Mantle Bouguer Anomalies near the fracture zones are more likely to be associated with the density zones rather than with crustal thickness variations.

Conclusions

1. Ship-borne Free Air gravity anomalies clearly outline the second order segmentation of the spreading center. Rift mountains located at the centers of the second order spreading center segments are associated with local gravity highs with amplitude ranging from 80 up to 140 mGal.

2. Second order discontinuities are expressed in the Sandwell's Free Air Gravity Anomaly field as saddles between the highs associated with the rift mountains in the centers of second order segments. Some of the second order discontinuities (4NS-3 second reorder discontinuity) are defined on the Sandwell's Free Air Gravity Anomaly field by long linear troughs similar to those of transform faults with a very well defied nodal basins and straight non-disturbed off-axis aseismic fracture zone traces up to 270 kilometers away from the spreading center axis.

3. An analysis of gravity data together with multibeam bathymetry between 2III20'N and 5IIIN along the Mid-Atlantic Ridge axis and its flanks revealed pronounced segment-scale variations of mantle Bouguer gravity anomaly which may correspond to crustal thickness variations. The mantle Bouguer local low is located near the center of spreading center second order segments, which indicates that at the centers of second order segments the crustal thickness is at its local maximum, The Mantle Bouguer gravity anomalies at the second order discontinuities are local highs, which suggests that the thickness of the crust here is at its local minimum.

4. The magnitude of the local Mantle Bouguer gravity lows or "bull's eyes" does not have any correlation with the spatial dimensions of the corresponding second order segments. This notion contradicts the correspondence observed by Detrick et al. (1994) of along axis gravity magnitude variation with the length of the spreading center segment (Mid-Atlantic Ridge between 33IIIN and 40IIIN). Such difference once more shows that in the study area the shape of the Mantle Bouguer gravity anomalies is influenced by the Strakhov f.z., which obscures a generally observed correspondence of along axis gravity magnitude variation with the length of the spreading center segments.

5. An important result of the study is that it showed that some mantle Bouguer gravity anomalies associated with the second order segments are extended far away (up to 200 kilometers) from the spreading center axis. Such spatially large mantle Bouguer gravity lows coexist with relatively small features, which suggest that the crustal thickness variations over the second order segments have significant variance across as well as along the ridge axis.

Chapter 9

Manifestation of Tectonic and Petrochemical Segmentation during the Development of the Equatorial Zone of the Mid-Atlantic Ridge.

N. M. Sushchevskaya, L.P.Volokitina, M. M.Morozova, M.M.Fuguan

The problem of the Mid-Atlantic Ridge (MAR.) rift zone formation in the equatorial region is very interesting, first of all because here the initiation of the ridge has had many interesting prerequisites. The system of extended transform faults integrated in the so called mobile belt global system is supposed to have existed long before the Atlantic Ocean started to dilate (Neev et al.,1982, Nurnberg et al. 1991). Trans-atlantic faults Chain, St. Paul, and Romanche can be traced both in Africa (Benouet trough) and in South America (Amazon mouth). The equatorial zone separates two provinces of newly developed oceanic lithosphere which were formed at different times. These provinces are the Central one (15°-33° N.) whose formation started about 170 million years ago, and the South one (to the South from the Equator) whose spreading started independently about 120 million years ago (Nurnberg et al., 1991). The equatorial zone began to be formed later, about 90 roil, years ago at the junction between these two provinces. This work is dedicated to the comparison of the peculiarities of the geological structure of different regions of the M.A.R. equatorial zone with peculiar features of magmatism.

<u>Geological structure and tectonic segmentation of the M.A.R equatorial zone.</u>

Equatorial Atlantic segmentation, as revealed on the bathymetric profile drawn along the crest of the ridge (Plate 18, Fig.1), showed that crust disruptures separating segments of different orders are reflected in the local maximum depth, while small depths are typical for the surface of the crest of the ridge. (IGG Atlas Atlantic ocean, 1989-1990). The equatorial M.A.R. region as a whole is characterized by pronounced fracturing of the bottom by transversal faults which are pronounced to a variable degree and separated by 15-35 miles, (Plate 18, Fig.2). The whole spreading ridge was broken up into small segments, which is clearly demonstrated by the ridge morphology (Udintsev et al,, 1991). Right hand displacement of the ridge is displaced to the right, as it occurs to the South of the Strakhov fracture zone (f.z.) and sticks blindly to the rift crests of the trough bottom, forming nodular depressions where maximal depths of the bottom or of the small hollows are observed, Several kinds of displacements of the ridge rift zone have been established. The degree of manifestation of these displacements is determined by the segment structure. Segmentation of the lowest (3d) order prevails within 1st order segmentation (Plate 18, Fig.1). Complexity and variability of the rift valley morphology have been revealed, in particular vast fluctuations of the bottom depth (from 400-1500 m to 3500-4200 m) and differences in ridge morphology between central and peripheral parts of the segment. Asymmetry is manifest most clearly in the structure of the rift valleys, in the disposition of the domes (elevations), and in the variations of the crust structure. Altimetric data demonstrates this asymmetry most clearly for regions of St.Paul-Romanche faults (Calcagno, Cazenava, 1993),

The Cape Verde f.z. which intersects the MA. R. along 14°25’-14°45’ N,. line is the Northern border of the region under consideration. Marathon and Mercury faults intersect the M.A.R. axis along the 12°40' and 12°10’ line. Here M.A.R relief changes dramatically, The rift valley is extremely well pronounced to the North of this zone. Valley bottom depth varies from 3800 m in the North to 4800 m in the South. To the South of the Marathon and Mercury faults the ridge is crossed by several large faults separated by 30-40 miles each: Vima, Arkhangelsky, Doldramo, Vernadsky, Strakhov, St. Paul, Romanche (Plate 18, Fig.2). The rift zone is thus divided into separate sections; the width of the mid-ocean ridge diminishes gradually from the North to the South. The rift valley bottom is elevated most of all between the Cape Verde f.z. and 13°45’ N. where its depth is no more than 900 m (Mazarovich, Morozov et. al.,1992, Pushcharovsky, 1989). In the interval between 7°30’ and 9° N.. Arkhangelsky (8°50’N.), Doldrams (8°10’N..) and Vernadsky fault zones can be easily traced. Rift zones go orthogonally to the faults. The rift zone is separated from the trough bottom by a step (ledge) up to 1500 m high (Mazarovich et al., 1991), The most differentiated morphology is typical of the central region between the Marathon and Strakhov f.z.. Maximum depths are found in the Western interriftal part of the fault, The width of the definitely riftogenic axial part of the ridge is 80-100 miles. Depressions of the rift crest up to 2800-3200 m deep determine ridge segmentation to the North and to the South from. Strakhov f.z.. St.Paul and Romanche are the longest and most complex faults in their structure (Agapova, 1994, Belderson et al., 1984; Bonatti et al.,1994).

Fragments of the riftogenic ridge in the St. Paul f.z. complex are from 10 to 20 miles wide, and consist of separate blocks separated by small saddle-like structures (Agapova, 1993). The highest and most massive blocks fringe rift canyons. The transverse fault profile is asymmetric, and its Northern block is bordered by a high fault ridge along which large massifs and mountains are situated. The more ancient and leveled relief in the Northern part and more contrast relief in the Southern one can be seen in its structure. The morphology of the Romanche fault is rather complicated, especially in the Eastern part, where regions with a leveled relief, asymmetry of the South and North slopes and relic transverse ridge in the North slope can be pointed out. It is interesting that the fault ridges are very shallow: their relative height is 1550 m in the St. Paul fault, and less than 2000 m in the Romanche fault. Yet the depth rift regions associated (coupled) with the faults reach 4500 m in St. Paul and 5000 m in Romanche f.z.. These are the deepest bathymetric points of M.A.R. rift valley. It should be pointed out that the depth of the Romanche f.z. is more than 7000 m.

Petrochemical and geochemical segmentation and peculiarities of magmatism.

We have recognized 7 segments in the MAR, taking into consideration the results of probing in the region between 15°20' and Chain f.z. and having regard to tectonic segmentation (Plate 18, Fig. 2). Total sampling of the glass composition comprised some 300 analyses. Classification considering 10 elements allowed for the recognition of several stable petrochemical types, most of which are products of differentiation of the primary melts of TOR-2 ocean rift tholeites (Dmitriev, 1985). Melts of this kind were generated from the mantle of lerzolite composition in the course of polybaric melting . Total degree of melting reached 14-17% and the depth of the last equilibration with the mantle substrate was about 20 km. TOR 2 melts which are of widespread occurrence along the MAR, to the South of 33° N.. erupted within 3 segments in the central part of the equatorial zone: between the Marathon and St. Peter faults (2°30' N..). Average compositions of cluster groups of the glasses are shown in Fig. 3, Plate 18.. These glasses according to Susshchevskaja et al, 1994, belong to the TOR-2 type. Lines of differentiation of the primary melts of the TOR-2 type at a pressure of between 0.1 and 10 kbar calculated with the help oh the TOLEMAG program (Ariskin, Tsekhonova, Frenkee, 1992) are shown. The program based on the dependence of the distribution coefficients on P and T for the phases of the crystallizing basalt systems describes conditions of dry tholeite melts crystallization well enough. The main result of such modeling is that crystallization of the melts in the central part of the equatorial region occurred in conditions when pressure was decreasing from 8 to 4 kbar and temperature was between 1250-1170°C. The maximum degree of crystallization was about 55%. Spreading alongside these segments were hyperbasites with features normal for the Oceanic mantle: harbugites with medium degree of depletion, spinels in which were of composition between 20 and 40 100Cr/(Cr + Al). All these facts indicate that conditions of the modern rift formations were stable enough, his is typical for the greater part of the Atlantic. At the same time for magmatism in the regions of the MAR to the South of the 2nd and to the North of the 4th segments, drastic enrichment of the tholeitic magmas in lithophilic elements is very typical. Here low-potassium tholeites are rather scarce. The 1st segment between 15°20' and Marathon f.z. and the 5th one between St. Paul and St. Peter f.z. belong to such "geochemically anomalous" segments (Fig. 3, Plate 18), Petrochemical features of these two anomalies are very similar. The only difference is that more differentiated magmas are found near the St. Paul fault. Estimation of the crystallization and magma generation conditions showed that primary melts are characterized by broad dispersion of their compositions; primarily, this concerns such components as Fe, Ca, K (Plate 18, Fig.4). A broad spectrum of primary melts could be the result of an unstable melting regime of the separate diapirs in various intervals of pressure (4 kbar on the average). Beginning of melting and of separation of melts took place at a depth of 30-40 km. Lines of low depth primary melt crystallization are shown in Fig. 3. These melts were formed at various pressures between 4 and 20 kbar. It is seen that compositions of different cluster groups of glasses from the 1st and 5th segments lie between these lines. More thorough study of melts belonging to each group requires further probing. Moreover, further improvement of modeling of the melting processes of the mantle of various compositions is necessary. This will allow for a more precise estimate of the values of enriched tholeite melt crystallization parameters. It should be noted that outcrops of hyperbasites with a maximum degree of depletion (100Cr/(Cr+Al)>40) are associated with the 1st and 5th segments, which testifies to a prolonged melting process in which mantle rocks have been involved. It can be stressed that enrichment character has been changing for the St. Paul-St. Peter segment during at least 12 million years. Probing made during the 7th and 12th expeditions of RV "Akademik Nikolaj Strakhov" referred to peripheral regions of the ridge as well, while this anomaly was observed by Shilling (Shilling et al., 1990) who studied the rift valley itself.

The most specific magmas are found in the region of th M.A.R. to the South of St. Paul and near Romanche f.z., as well as in active parts of these faults. Magmas essentially enriched in Na predominate here. Melts of a similar type were found earlier only in the Cayman trough, in the zone of the Australian-Antarctic ridge (Klein , Langmuir, Staudigel, 1991) and in the Knipovich ridge (Neumann, Schilling, 1984). In the samples of the glass from the eastern intersection of the Romanche f.z. with the N.A.R., its composition is constituted by more than 70%. Na tholeites. They were also found in one of the stations near the eastern intersection of St. Paul f.z. (Sushchevskaja, Udintsev, et al, 1990). Primary melts of Na-type are enriched with Si whose content can reach 52% and are depleted of Fe.

Melts whose 50-60% differentiation gives rise to observed magmas were formed during mantle melting at rather low depths (corresponding to 4 kbar). Melt compositions given in ref. [Kinzler, Grove ,1992] are closest to estimates for the primary magmas. Magmas of this type can be melted out during polybaric critical melting of primitive mantle when about 1% of melt is permanently present. Total degree of melting reaches 10-14%. The mantle is being melted in the pressure interval from 20-15 up to 4 kbar. Differentiation occurs in the same manner as in the case of normal tholeites at a pressure of about 3-4 kbar. Melts that are most enriched in Na are associated with the Romanche fault and related Southern zone of MAR Occurrence of basalts with alkaline composition can be regarded as an important feature of magmatism in the St.Paul-Romanche region. They can be associated both with peripheral parts of MAR (3° N. to the North from St. Paul, Zolatarev in this book), with the faults (slope of St. Paul, Northern side of the Romanche f.z.) as well as with the rift valley (Fig. 2). Thus glasses and basalts of alkaline composition were found in the Eastern slope of MAR (Suschshevskja, Peive, Tsekhonova et al., 1995)) near its junction with Romanche fault. Joint occurrence of normal tholeites, Na-tholeites and of alkaline magmas indicates primarily that the system of intermediate magma chambers in which magma mixing could have taken place has been absent. It is interesting to underline that outcrops of hyperbasites are associated with Romanche f.z. Among these hyperbasites there are those with a minimum degree of depletion (100Cr/(Cr+ A1)<20) as well as those in which presence of trapped melts can be observed. (Bonatti et al, 1992).

The geochemistry of melts of various genesis shows that they constitute a unique field whose compositions range from slightly to highly enriched lithophilic elements. Basalts and glasses belonging to TOR-2 and to TOR-Na types from the St, Paul - Romanche fault regions depleted of Ce/Yb form a unique field with those slightly enriched in light REE (TOR-NA, Sushchevskaya et al., 1994), including maximally enriched ones (alkaline varieties, Fig.6). The appearance of these magmas enriched to various extents can be explained by melting of the mantle of mixed composition when the primitive mantle (of the Ronda, Zabargad type) and fractions of metasomatized mantle essentially enriched in light TR have been involved in melting (Hekinian, 1991). The vastest areas with which occurrence of enriched magmas is associated, as it has been already pointed out, are the 1st and 5th segments (La/Sm>2). Melts with intermediate compositions erupted in the active parts of St. Paul and Romanche faults. But locally enriched magmas occur to the North from this area as well: the eastern side of MAR about 5°30’ N.. A small segment of MAR near its intersection with the Eastern region of Romanche f.z.. proved to be very highly enriched. This segment is some 20 km long. Isotopic 144Nd/143Nd ration obtained by us for glasses (Sochevanov, Suschchevskaja, Kononkova, 1990) and for basalts are plotted in Fig. 4,5., Plate 18. These data indicate that the most important shifts in Nd isotope ratios (up to 0.5129) have been found for the anomaly of the 1st segment and St. Paul-Romanche f.z.. Metamorphosed St.Paul hyperbasites are distinguished by high isotope ratio values close to 0,5127-0.5129. This substantiates the possibility of involvement of the mantle whose characteristics were similar to those of the subcontinental mantle in the process of melting.

Evolution of the Equatorial Atlantic Province

Thus, when considering the Equatorial segment(ES) of MAR magmatism it is possible to recognize three major dominating features:

1. wide occurrence of specific thoelites of Na-type which are mostly variable in the Romanche f.z. region and to a smaller extent in the Eastern part of the St, Paul f.z.;

2. two vast geochemical anomalies which are similar in many parameters and are situated in the North and in the South parts of the region considered;

3. wide occurrence of magmas of alkaline composition which occur most frequently near the Romanche and St. Paul f.z. and along them.

Regions near 13-14°N., 0-2.°50’N. (geochemical anomalies of the 1st and 5th segments) and Romanche f.z. region (where diapirs are subject to multiple melting and melts formed at various depths and with various degree of melting have the possibility to reach the surface) are marked in a special way, Melting near the Romanche-St. Paul zone occurred in a broad interval of pressures with permanent retaining of the melt in the mantle substrate matrix. The mantle in this region is not homogeneous and is partially metamorhized. The mantle of subcontinental composition was present in the form of non-spread blocks from the early stages of spreading. Partial crystallization of melts takes place practically ubiquitously at a depth of 10-12 km. Melts of alkaline composition are an exception, they can melt deeper. Magma generation processes and subsequent differentiation for other segments (central region 13-2°N..) are typical for the total MAR region to the South from 33° N. where polybaric fractional accumulating melting occurs at a depth of 60-20 km (Sushchevskaya, Tsekhonya, 1994.). Such complicated structure of the ES, which is manifested in the diverse composition of the erupted magmas, could be possibly due to its formation on the border between two deep convectional streams in the mantle, similar to the situation in the region of the Australian-Antarctic Rise. ES MA.R. accommodated to the processes of deep dynamics and of lithosphere spreading, remained relatively cold during the whole period, and underwent changes, permanently alternating periods of spreading and of relative stability, during which separate blocks could be subjected to complicated motions relative to each other, these motions being both vertical and horizontal.

Preliminary absolute age determinations of gabbroid from the two regions: the zone of intersection of the Doldrams fault with the rift zone MAR. and from the west of the rift zone of the ES MAR. in the region 2°N.. (Fig.2) are given in Table 1. The analyzed apatite gabbro sample was taken from the top of the Peive mountain. This mountain is composed of gabbroids of different composition and degree of differentiation (17 Mazarovitch et al., 1991). Determined K/Ar age is 21 million years. It shows that the process of spreading had a discrete character and earlier formed regions of the ocean crust could remain in the form of unspread blocks. The age of plagioclase-pyroxenic gabbro, whose sample was taken 130 miles to the West from the rift zone of ES MAR. (2°N..), is close to 29 million years and reflects evidently normal conditions of rift formation in this region. The complexity of the evolution of this region is reflected in the morphology of the MAR structure. The transverse ridges, observed in the Romanche and St.Paul f.z., delimit equatorial can be traced till the continents and show early borders of the transforms. Changes in the direction of individual sections of these ridges point to changes in the direction of transform faults. At the same time MAR segmentation is due to initially formed fault zones and preserved during a long time, though changes have occurred in the segment direction and length. Spreading axis migration and reorientation of the transform borders was found in several regions of the Romanche, St.Paul and Doldrams faults (Agapova,1994, Mazarovich et al., 1991, Bonatti et al. 1994).

The role of transform faults in the process of magma formation is not clear up to now. They were inherited from the Precambrian system of African and American continental equatorial faults. They were marked by eruptions of the deep alkaline magmas. In this sense, manifestation of alkaline magmatism in the Eastern part of equatorial region of MAR could be of the same origin. Future comparison of all geochemical characteristics will help to answer the question. The analysis performed enabled us to come to the following main conclusions:

1. Depth of melt crystallization in the equatorial part off MAR (15°-0) lies within 12 km, which is typical for the whole MAR region to the South of 33°N..

2. Segments of anomalous tholeite magmatism exist where enriched tholeites (15-13° N..,0-2°30’N..) as well as melts of Na-type of shallow genesis (0-1° S.) occur.

3. Occurrence of Na-tholeites and enriched TOR is correlated with morphological peculiarities of relief relative elevation of the rift valley for the segments with enriched tholeites and depression near Romanche f.z. where Na-type tholeites are found.

4. Absolute age determination substantiates complexity and step-type behavior of the spreading processes in the evolution of the equatorial region of the MAR.

Chapter 10

Igneous Rocks from the Equatorial Atlantic.

P.I.Fedorov, B.P.Zolotarev, V.A.Gutsaky

The Equatorial zone of the Atlantic ocean is particularly interesting for a combined geophysics and petrology study for several reasons. The first reason lies in the fact that the zone is anomalous by geological parameters (cf. relief shape, mosaic crust construction, pressure and extension displacements etc.). Secondly, this zone contains a unique object - the St. Peter and Paul Islands which are built mostly of milonitic peridotites and serpentinites.

During the 12th Leg of RV “Akademik Nilcolaj Strakhov”, three sites were dredged, including the Strakhov fracture zone (f.z.) and the areas of Equatorial Segment (ES) of the Mid-Atlantic Ridge (MAR), including the vicinity of the St.Peter and Paul Islands (Plate 3, 16).

Four dredgings of the bedrock near the Strakhov Fault have produced various igneous rocks which may be classified into three groups: serpentinites, gabbroids and basalts.

Serpentinites (st. 1210; 3°56’ N - 30°36,68 W) are represented by rare block fragments (up to 15-20 cm), in which only ore minerals - magnetite and picotite - have been preserved as original rock relicts. The rock is characterized by reticulate structure, which is accentuated by distribution of newly generated magnetite.

Gabbroids comprise a group of rocks, including bipyroxene gabbros, gabbro-syenites and gabbro-dolerites are recognized. Bipyroxene gabbros (st. 1203, 4°06,98 N -29°11,64 W; st. 1210) are dense and medium-grained. The most typical gabbro varieties consist of clino- and orthopyroxenes, plagioclases and ore minerals. Pyroxenes are partially replaced by hornblende of two types: basaltic hornblende, which is the product of rhombic pyroxene replacement, and the common one, which results from basaltic hornblende transformation. Apatite is often present as an accessory mineral. Gabbro-syenites (st. 1203) are represent by dense, medium-grained rock, consisting of plagioclase, potassium feldspar, hornblende and ore minerals. Apatite, zircon, sphene are present among accessory minerals. The rocks have hypidiomorphic-granular structure. Gabbro-dolerites (st. 1203) are distinguished by dolerithic and poikiloohitic structures and consist of plagioclases and clinopyroxenes. Plagioclases are represented by short-columnar crystals occasionally containing poikilitic ingrowths of clinopyroxene. Light greenish clinopyroxenes are partially replaced by hornblende and are situated within plagioclase interstitions. Apatite is often present it forms acicular and fusiform crystals; rare picotite grains are present among accessory minerals.

Basalts were encountered in two field sites, Within the northern border of the Strakhov f.z. (st. 1203), basalts are represented by fine-grained aphyric rocks with trachytic matrix structure. The matrix is composed of plagioclase laths, clinopyroxene grains, small feldspar crystals and fine-grained magnetite aggregate. Rare plagioclase microphenocrysts are often distorted and broken,with fragments being slightly displaced. Clinopyroxene is often replaced by hornblende. One basalt sample (12D1-13) contains an xenolith of amphibole-pyroxenic gabbro. Xenolith is distorted at the contact with basalt. The second group of basalts is developed in the northeastern slope of the northwest striking structure within the southern border of the Strakhov f.z. (st. 1213). They are represented by pillow- lava fragments divided by cracks of sector jointing, with generally well preserved thin glassy crusts. The rock is subporphyric, rarely vesicular. Phenocrysts are represented by plagioclases, olivines and clinopyroxenes, which sometimes form glomeroporphiric aggregations. Matrix structure varies from intersertal in central pillow parts to paniculate in marginal zones. Occasional emulsion structure occurs, which has been formed at the earlier stages of rock generation. This is characterized by the presence of almost perfectly rounded entities which are often grouped into filamentous, cellular nets. These spheroids are formed by spherical or sheaf-shaped plagioclase aggregates surrounded with hyalomelane. The glass within interspherolite spaces is lighter and contains larger plagioclase microlites and skeletal clinopyroxene microlites. Nontypical ooids occur, lacking radiolith orientation of component crystals. They consist of randomly oriented crystals, which are translucent enough to see a sagenite-like aggregate of fibrous crystallite occupying the most part of an ooid.

Hardening glass, which form pillow crusts, is unusual in appearance. The distal part of the crest consists of pure sideromelane glass, comprising porphyric phenocrysts of pyroxene of the same habit and size as those in the central pillow parts. At some distance from the crust surface, non-uniformly distributed and irregularly oriented glass areas with slight birefringence occur. Skeletal pyroxene crystals are situated in the middle of such areas; filamentous crystallite accrete round them, forming parallel-fibrous or sagenite-like aggregates. Glass, saturated with such aggregates, is distinguished by its brownish hue. Extinguishing of such glass areas generally coincides with central skeletal crystal extinction. It is important to note that linearly stretched ooid entities, described previously, are found within the zone of tesselately extinct glass occurrence. A variolated basalt zone occurs within the intermediate zone between the glassy hardening crust and central pillow part, possessing intersertal structure. Varioles are of common structure. Clinopyroxene or plagioclase crystallite may occur within their central parts, and the radial-fibrous aggregate is composed of plagioclase laths with dendrite-shaped undiagnosed crystallite between them.

Generally, the basalts are rarely vesicular. Segregation and degassing vesicles are occasionally present their content does not vary in various structural rock types and constitutes less than 1% by volume.

Dredging from the axial part of ES MAR in vicinity of the St, Peter and Paul Islands (st. 1215; 3°14’15 N -31°04’42 W) has produced a vast amount of basic and ultrabasic rocks. Rock fragments range in size from a few cm to 0.5 m across, and often posses fresh detachment planes. Serpentinites and serpentinized peridotites and cataclased basaltoids are recognized among the rocks.

Serpentinites are massive dense rocks of dark gray and greenish-gray colors and schlieren-net (lenticular) and porphyroblastic structure. Structure design is conditioned by antigorite filling of the lenses, fringed with cross-fibrous chrysolites or lizardite. Picotite up to 0.8 mm across is frequently isolated, sometimes with magnetite pseudomorphs. Magnetite is also widely developed as fine-grained accumulations along the borders of serpentinite loops. Cracks are seldom filled with sepiolite aggregate. The least altered rock varieties contain clino-, and orthopyroxene relicts with widely developed bastite pseudomorphs. The rock compositions indicate them to be originally harzburgites. Bastite isolations within harzburgites represent coarsely crystalline antigorite aggregates up to 1mm across. Lizardite loops over olivine are also noted; lawsonite constitutes loop cores.

Cataclased basaltoids represent completely altered rocks with relict phenocrysts of dark-colored minerals, replaced with amphibole-hydromicaceous aggregates.

Within the meridional structure in the eastern flank of the ES MAR (1°20’5 N- 30°13’1 W), about ten miles from the rift valley (st.1216),a large amount of stone material was excavated, represented mainly by base volcanic and subintrusive rocks.

Among the basalts, aphyric and plagiophyric varieties are recognized. Aphyric basalts are dark gray, dense, vesicular. Rock matrix is composed of brow devitrified glass, small plagioclase laths, dendrite-shaped clinopyroxene seclusions, often forming variolites, and magnetite grains. Plagiophyric basalts are massive and vesicular. Vesicles (10-15%) are filled with smectite and chlorite. Phenocrysts are represented by plagioclases, measuring from 0.5 to 1-2 mm, often fused, with simple twinning. The matrix is composed of plagioclase laths, containing both rounded and dendrite clinopyroxene grains, magnetite and volcanic glass within the interstitions. The matrix is intersertally structured.

Microdolerites are distinguished by their dark green color and massive texture. Matrix structure is microdoleritic and poikilophitic, filled mainly with plagioclase laths measuring 0.7-0.8 mm, rarely with tabular crystals and grains of clinopyroxene and magnetite. Small olivine grains and rare seclusions of brown volcanic glass are also registered.

Gabbro-dolerites are cataclased and partially amphibolized. The rocks are of ophitic structure.

Aphyric, subaphyric and plagiophyric basalts are developed to the south, in the western flank of the ES MAR (st. 1226,0°55’5 N -30°14’7 W).

Aphyric basalt matrix is composed of plagioclase laths, often with cleaved terminations, small clinopyroxene grains, hyalomelane and magnetite. The matrix is of intersertal structure.

Subaphyric basalts are gray-coloured and vesicular (5-10%). Phenocrysts are represented by plagioclase, olivine and picotite. Plagioclases are formed by large (0.8-2 mm and more) tabular crystals. Simple and multiple twins are common: some cases of oscillatory extinction are observed. Plagioclases are often reactive with the matrix, which is manifested in dissolution and recrystallization. Olivines form isometric crystals measuring 0.8-2mm and containing numerous inclusions of spinel, plagioclase, glass, which are often polyphase. Picotites, less common among the phenocrysts, are represented by crystals measuring 0.4-0.6 mm. Besides phenocrysts, plagioclases, olivines and picotites in some cases form megalomeroporphiric aggregation sizing up to 7-8 mm. The composing crystals are generally badly fused and often rounded. The matrix is composed of plagioclase, clinopyroxene, magnetite and glass. The structure is intersertal, at some sites-glomeroporphiric.

Plagiophiric basalts are highly vesicular (15-20%). Plagioclase phenocrysts measuring up to 1.5-2 mm are of tabular shape, often of fusion structure; they contain glass and spinel inclusions. The matrix is formed by plagioclase laths up to 0.4- 0.5 mm long, clinopyroxene grains, magnetite hyalomelane. The structure is intersertal, occasionally variolitic.

Besides basalts, dredging at this station has produced fragments of yellowish-gray hyalomelanic volcanic glass. Separate sites of crystallite growth are registered within the glasses. Rare porphyric phenocrysts are represented by plagioclases of various habits, and brown spinel crystals.

Igneous rock chemistry is presented in Tables 1-3. Major oxides were determined by the wet chemical methods at the Geological Institute of the Russian Academy of Sciences. Cr, Ni, Co and V were determined by the quantity spectral method; Rb, Ba, Sr, Zr, Y and Nb by X-ray fluorescence and the abundance of REE were determined by instrumental neutron activation at the Geological Institute of the Russian Academy of Sciences.

It is evident from the variation diagrams (Plate 19, Fig.1) that basalts and dolerites from the areas studied are poorly differentiated with respect to silica; they are characterized by high iron content and total alkalinity at increased titanium concentrations. High Na_2O/K_2O ratios (3-14) are characteristic of lavas. At the same time, there are certain differences in silica/total alkalis ratio between the lavas from Strakhov f.z. and ES MAR. The basalts from the Strakhov f.z. and some lavas from the southern sector in J. McDonald's petrochemical diagrams and AFM (Plate 19,Fig.1) fall into the composition field of abyssal ocean tholeiites, and evolution of rock chemistry of close to the "fennerian" differentiation trend, which is manifested in iron accumulation with a slight increase in total alkaline elements content. At the same time, figurative points for the basalts and dolerites from the subequatorial part of the Mid-Atlantic Ridge fall into the field for rock of subalkalic soda series.

The differences among the recognized rock associations become more evident after the analysis of rare and rare-earth elements (REE) behavior. The contents and chondrite-normalized distributions of REE within the lavas of the tholeiitic series are on the whole similar to those of normal and transitional types of oceanic tholeiites (Lan/Smn=0.7-1.2; Lan/Ybn=0.9- 1.9; Plate 19, Fig. 2). The lavas of subalkalic series are more enriched in light REE (Lan/Smn=1.7-2.4; Lan/Ybn=l.6-5.4).

Igneous rock chemistry makes wide use of diagrams, demonstrating covariations of high field-strength element (HFS), large ion lithophile element (LIL) and coherent elements. J.A.Pearce and M. J. Norry [1979] and M. Meschede's [1986] diagrams are among them; the authors believe that they do reflect heterogeneities within the lithosphere, but do not depend upon composition variations, caused by various degrees of partial fusion and fractional crystallization. The figurative points relevant to the tholeiitic series, are situated either within the abyssal tholeiites field (Plate 19, Fig.3A), or are shifted into the area of transitional basalts (Plate 19, Fig.3B), while the subalkalic basalts form an independent field of enriched tholeiites lavas. A similar character of enrichment is clearly seen in Figure 4 (Plate 19), where rare element concentrations in rocks are normalized relatively to N- type MORB [Sun, McDonugh, 1988]. The spectrum of enriched basalt type (P-type MORB) is shown in spider diagram for comparison. It follows from the diagram analysis that the lavas from both series are diversely enriched relative to N-MORB. In tholeiite resies lavas, basalts are poor in high-charge elements, somewhat enriched in LILE, while the HFSE within subalkalic series basalts behave similarly to those in P-type MORB.

Aplite samples collected from the Station 15 have low Na/K ratios and plot within the subalkalic field on the Na_2O+K_2O versus SiO_2 diagram (Plate 19, Fig. 1). Aplites are light-REE-enriched, with Lan/Ybn ratios ranging from 19.2 to 19.7. In this respect they are similar to syncollizionation environment aplites (e.g. E.Sayan, see Plate 19, Fig.5).

Ultrabasic and basic rocks from the vicinity of the St.Peter and Paul Islands are described in a number of papers [cf. Frey, 1970, Melson, 1967, Prinz, Keil, Green et al., 1976], presenting detailed features of their composition and metamorphic transformations. The data on serpentinite and serpentinized peridotites, obtained by the authors, generally correspond to those described by F.Frey [1970]. The rocks are, as rule, badly altered the degree of serpentinization makes about 80-90%, reaching 100%, which is evident from the chemical composition of the rocks - the amount of constitution water makes up to 8-11%, sorbed water - up to 1.2%. Sodium, aluminum and titanium content are low. REE distribution (Plate 19,Fig.6) within serpentinites and harzburgite is also distinguished for depleted spectrum (Lan/Smn=0.3-0.8; Lan/Ybn=0.3-0.9) situated within F.Frey's field of peridotites from St. Peter and Paul Islands.

Our studies of igneous rocks from the Equatorial segment of the Mid-Atlantic Ridge indicate a wide variety both of volcanic and plutonic types. The volcanic rocks include tholeiitic and subalkalic Pl phyric basalts, Ol-Pl phyric basalts and aphyric basalts as well as hyalobasalts and volcanic glasses. The portion of basalts are vesicular (up to 20%) and this genesis is accounted in shallow-water situation. Among plutonic rocks are dredged a diverse suite of samples including peridotites, serpentinites, gabbros and dolerites.

Basalts composition are similar to other data from this zone of Mid-Atlantic Ridge at 0o-4o N and suggest that all basalts analyzed were from a heterogeneity region source that is typical of Equatorial segment of the Ridge.

Table 1. Representative compositions of the 12 LEG "Academician Nikolaj Strakhov" volcanic rocks

Station	1213		1220			1226		
Sample	D10-1	D10-7	D16-18	D16-4	D16-8	D20-5	D20-10	D20-11
SiO_2	48.41	44.12	46.89	45.96	45.86	48.25	48.95	48.75
TiO_2	1.53	1.69	1.86	1.53	1.66	1 52	1.62	1.44
Al_2O_3	15.41	16.35	15.22	13.64	14.23	14.60	13.97	14.88
Fe_2O_3	4.12	8.03	4.13	5.10	5.82	3.42	5.00	3.90
FeO	6.39	4.63	5.48	5.68	5.25	6.50	5.43	6.42
MnO	0.18	0.35	0.07	0.14	0.49	0.14	0.13	0.10
MgO	8.50	7.32	8.46	9.68	8.61	9.58	7.35	8.48
CaO	12.20	12.01	9.76	10.02	8.18	12.16	13.15	12.10
Na_2O	2.51	2.43	2.93	1.97	3.20	2.42	2.57	2.32
K_2O	0.19	0.17	1.44	0.41	1.42	0.67	0.67	0.69
P_2O_5	0.13	0.33	0.26	0.19	0.36	0.18	0.18	0.14
$H2O^+$	0.41	0.98	2.57	3.56	3.49	0.44	0.63	0.80
$H2O^-$	0.34	1.01	0.76	0.72	0.79	0.31	0.60	0.40
Total	100.32	99.82	99.83	100.20	99.58	100.30	100.25	100.42
Cr	415	338	43	182	195	332	235	275
Ni	205	200	75	160	155	185	115	140
Co	43	62	29	31	112	35	26	25
V	330	400	240	255	250	225	230	205
Sr	100	170	210	150	200	190	210	180
Ba	23	30	320	190	190	94	85	83
Zr	68	88	110	99	130	92	87	95
Y	24	27	15	17	23	16	22	16
Nb	8	8	20	17	26	14	16	16

Table 2. Representative compositions of the 12 LEG "Academician Nikojaj Strakhov" plutonic rocks

Station	1210		1215					1220
Sample	D7-7	D7-11	D12-12	D12-19	D12-22	D12-3a	D12-19a	D16-21
SiO_2	45.49	48.04	37.62	36.96	38.78	68.12	67.75	47.11
TiO_2	4.85	5.74	0.13	0.13	0.13	0.30	0.34	0.75
Al_2O_3	11.17	10.63	1.73	1.14	1.36	14.98	14.78	16.20
Fe_2O_3	9.10	3.14	7.37	9.56	6.96	-	-	4.26
FeO	9.07	12.30	0.86	0.63	0.99	2.07	2.28	4.91
MnO	0.35	0.49	0.31	0.29	0.14	0.07	0.07	0.04
MgO	5.90	4.63	36.87	37.23	37.47	0.72	0.69	10.37
CaO	8.95	7.88	1.08	0.84	0.96	1.82	1.86	8.20
Na_2O	2.82	4.21	0.21	0.17	0.17	3.41	3.33	2.57
K_2O	0.39	0.45	0.09	0.04	0.04	4.55	4.24	0.95
P_2O_5	0.03	-	0.01	0.01	0.01	0.17	0.17	-
H_2O^+	1.66	1.25	11.92	11.46	11.95	3.55	3.60	3.38
H_2O^-	0.58	0.35	1.15	1.06	1.05	-	-	0.84
Total	100.36	99.71	99.35	99.52	100.61	99.76	99.11	99.58
Cr	10	10	1120	710	825	-	-	332
Ni	43	25	1050	1000	1100	-	-	185
Co	38	27	100	82	120	2.9	4.8	35
V	370	400	35	22	20	-	-	225
Sr	81	45	-	-	-	-	-	190
Ba	-	40	40	-	-	-	-	94
Zr	80	250	-	-	-	-	-	92
Y	47	10	-	-	-	-	-	16
Nb	-	-	-	-	-	-	-	14

D7-7, D7-11 - gabbro; D12-12, D12-19 - serpentinites; D12-22 - harzburgite; D12-3a, D12-19a - aplites; D16-21 - gabbro-dolerite

Table 3. REE concentrations of the 12, LEG RV "Akademican Nikolaj Strakhov" igneous rocks

	D10-1	D10-7	D16-18	D16-4	D16-8	D20-5	D20-10	D20-11
La	4.2	10.0	20.0	12.0	24.0	9.1	12.0	11.0
Ce	11.0	21.0	40.0	24.0	70.0	24.0	24.0	22 0.
Nd	9.6	16.0	24.0	16.0	30.0	15.0	16.0	14.0
Sm	3.3	4.3	5.0	4.0	5.7	3.9	3.7	3.50
Eu	1.1	1.3	1.3	1.4	1.8	1.2	1.2	1.1
Tb	0.81	1.0	0.86	0.78	1.1	0.87	0.80	0.76
Yb	3.0	3.3	2.4	2.7	3.1	3.0	2.5	2.4
Lu	0.44	0.50	0.36	0.38	0.46	0.45	0.37	0.37

	D7-7	D7-11	D12-12	D12-19	D12-22	D12-3a	D12-19a	D16-21
La	3.4	2.8	0.6	1.1	0.47	48.0	46.0	2.2
Ce	11.0	6.7	3.0	6.8	1.10	94.0	95.0	4.7
Nd	12.0	6.0	1.0	1.5	0.60	-	-	3.3
Sm	4.9	2.4	0.21	0.31	0.16	5.1	4.5	0.9
Eu	1.55	0.94	0.03	0.09	0.02	0.71	0.79	0.44
Tb	1.53	0.72	0.05	0.05	0.04	0.66	0.57	0.22
Yb	5.6	3.6	0.23	0.18	0.26	1.60	1.50	0.86
Lu	0.83	0.57	0.03	0.03	0.05	0.23	0.22	0.14

\- not determined

For samples D12-3a and D12-19a also have been determined the Th (24 and 26 ppm, accordingly), U (9.9 and 8.6 ppm), Hf (3.9 and 4.4 ppm) and Ta (1.3 and 1.3 ppm).

Chapter 11.

Magmatic Basement of the Equatorial Segment.

N.A.Kurentsova, V.M.Tchubarov.

Samples of igneous rocks were dredged in the axial rift zone of the Equatorial Segment (ES) of the Mid-Atlantic Ridge (MAR) and on its flank plateaus, for which is possible to assume non-rifted origin, and in the transverse fracture zones (f.z.) (Timofeev et al., 1990) Plate 4, 16, 19, Table 1].

Studies of the large collection of dredged rocks have shown a significant variety, which does not correspond to the accepted elementary scheme of the magmatism of mid-oceanic ridges (Aumento et al., 1973). This scheme implies a limited set of the complimentary interconnected rocks: lercolite-residual harcburgite-hipabissal gabbro-tholeitic basalt. Petrochemical and mineral-geochemical peculiarities allow only one(1) part of the rocks collected to be attributed to such a set. The other part (2) must be compared with continental hyperbasits, gabbroids and basalts. Yet another group (3) can be considered as transitional between the first two. We can thereby assume the presence here of the Jurassic basement of continental origin.

Riftogenous magmatic oceanic association (1) is represented here by glassy magnesial porfiric olivin-plagioclase non-metamorphic tholeitic basalts and by rocks of a differentiated small-depth complex of ferritic lercolite-tractolites. The age of that association can be considered as upper Cretaceus to Recent. That allows us to consider this association as younger than the basement (2, 3) rocks which it covers.

In comparison with the older ferruginous oceanic trapic basalts (3), the basalts of the rift zone are enriched with magnesium, nickel, and chromium and depleted of titanium,, ferrum, alkalies and light rare earths elements (Plate 19, Table 2). On the correlative diagrams (Lutz, 1980) the points of content of tholeitic basalts are close to the lower part of the oceanic riftogenous trend. I. According to the results of microsondal analysis, the olivines from riftogenous basalts are more magnesial (15-16% Fe) than olivines from oceanic trappic ferriferous basalts (17-29%. Fe). In comparison with the composition of olivines from riftogenous basalts (10-12% Fe) of the North Atlantic (30°-40° N) (Kurentsova, 1979), the olivines from basalts of ES MAR are more ferruginous. The difference is also very clear between the composition of clinopyroxenes from magnesial and ferruginous basalts: on the diagram Ca-Mg-Fe the points of compositions of clinopyroxenes from the former are compactly positioned closer to Ca-Mg than are the clinopyroxenes from the latter. The plagioclases from the basalts of rift zone ES are more basic (An 79-92) than plagioclases from ferruginous trapic basalts (An 49-70).

Gabbroids of this association are characterised by wide dispersion of all components and diversity of composition (Plate 19, Table 3). This can probably be explained by the by the broad involvement of volatile components in the process of liquation during magma creation. (Marakushev, et al., 1977). On the diagram AFM the points of composition of these gabbroids lay along the ferruginous trend 1, close to the trend of composition of gabbroids from Skjergard and of South Aldan (Lennikov et al., 1987). Microsondal analysis demonstrates that olivines from leucocratic gabbroids of ES are more ferruginous (24% Fe), than olivines fron tractolites of the North Atlantic (14-17% Fe) (Kurentsova, 1979).

The typical feature of the riftogenous oceanic magmatic association is clear depth differentiation. It is expressed in the alternation of the outpouring of magmas of intermediate composition. Those magmas produce a highly diverse mix of effusives and intrusive - leicobasalts,-picrites-troctolites-olivine and pyroxenic gabbro- plagioclase hyperbasites. The rocks of this association barely have have the marks of metamorphism. On the continents the closest to these geochemical analogs are stratiform intrusion of gabbroids (Shtcheka, 1979).

Magmatic association of the continental type (II) significantly differs from that described above. It is represented by undepleted, weakly-depleted and highly-depleted amphybolitised hyperbasites, ferrousitive-titanic sphen-apatite gabbro-amphibolites, alcalic gabbroids, trachybasalts, alkalic olivine basalts of the type of continental trapic basalts. The rocks of this association are distributed in the axial part of ES, but predominate in the zone of marginal dislocations and in the flank plateaus (Plate 10, 16). (St.5, 6, 34, 31, 12, 26, 27, 39,43, 52-54,40, St.Paul Isl.). In age thiu is the most ancient association of the ES.

Special attention should be “given to ancient mylonitised amphybolitised plagioclase lherzolites of Saint Paul Island, associated with alkaline gabbroids and alkaline basalts (Melson, 1972). In comparison with the peridotites of the rift zone (St. 43, 54, 39) and from the Saint Paul f.z. (St. 12, 26, 27, 59) hyperbasites of Saint Paul Island are enriched with Si, Ti, Al, Fe, Mg, Ca, P, Ni, Cr, Ci, Va, Na, and deplete of llithophylic elements (Plate 19, fig. 1).. In ferruginosity, , alumiferousity , high content of

AL203 in ortopyroxenes, in ferruginosity of olivines, and in low chromisity (13-22) these undepleted hyperbasites of St.Paul Island are close to peridotites of the Zabargad Island. (Bonatti et al. 1992). Hyperbasites from the Saint paul f.z. are weakly depleted (chromousity of spinel -30-40 at.%), and hyperbasites from the rift zone - highly depleted (chromousity of spinel -46-47 at %). This demonstrates the heterogeneity of the upper mantle beneath ES (Bonatti et al., 1992). The points of composition of the spinels from hyperbasites of ES lie on their own subcontinental trend to the right of the field of spinels of oceanic hyperbasites (Plate 19, fig. 1).

Gabbro-amphybolites are ferruginous-titanous, sphen-apaptitive, pericrystalled, underwent dynamometamorphism (St.5, Plate 19, Table 1-3).

On the AFM diagram the points of composition of the amphibolites lie on ferruginous trend -1, close to trends IV (Skjergard) and V (Southern Aldan shield) [Lennikov et al., 1987]. In clinopyreoxene composition (Plate 19, Table 6), the amphibolites of the western flank plateau are close to the amphibolites of the Zabargad Island, Red Sea. On the diagram Ti-Al-Cr the points of composition of clinopyroxenes are close to those of the clinopyroxenes from gabbroids of the Cambrian continental intrusions (Shtcheka, 1977). On the diagram, in variations of silicity, alkalinity and magnesium content, and also in the Rb-Sr correlation, the amphibolites of the equatorial segment are also close to continental rocks (Silantyev et al. 199), and also to high intensity titanium gabbroids of the Pejve seamount (Kepezhinskas et al. 1990) and to the rocks of the Angola-Brazil geotraverse (Trukhalev et al., 1992),

Alkalic olivine basalts and trachybasalts (of the continental trap type) are characterized by enhanced ferruginosity, titanium content, alkalinity, a high concentration of litofile elements and light rare earths. Such composition brings the basalts of the equatorial segment of MAR close to platobasalts of the Siberian platform (Makarenko, 1983; Udintsev, Kurentsova, 1990). On the correlational diagrams (Plate 19, Fig. 3) the points of composition of olivine basalts of the sodium series of the equatorial segment occupied a transitional position between the geosynclinal and riftogenous trends, closer to the patform trend of the continents and trends of the continental rifts.

Oceanic trap formation (III) which is transitional between the two described above, is presented with crystallized pyroxen-plagioclase basalts and trachybasalts distributed over the entire equatorial segment. They are characterized by a mandelstone texture and porosity. As distinguished from young magnesial riftogenous basalts, oceanic traps (sill complex) are significantly enriched with iron, titanium, alkalides, litofile elements and light rare earths (Plate 19, Table 2). On diagrams K-Ti, K-P, Zr-Ti, Zr-Rb (Lutz, 1980) the points of composition of oceanic traps occupy a position between geogynclinal and riftogenous trends, but lower than the points of composition of alkaline olivine basalts (association II). Oceanic traps precede the riftogenous phase with the complex of rocks I.

We reported earlier on the discoveries in outcrops of the basement of the equatorial segment of rocks exotic for the Mid Atlantic Ridge: sericitic coaly phyllites, microquarttzes, which have undergone regional metamorphosis, and also coal shales (Udintsev et al., 1990), Shterenberg et al. 1990).

The results of the study of magmatic rocks of the equatorial segment of the MAR slow us now to formulate several assumptions. In the basement of the equatorial segment lies a relict fragment of the crust of the continental type, of upper Jurassic age, which experienced transformation in the form of oceanization. This means that the geophysical parameters of the crust were converted from the continental to the oceanic type. Here, in the early Cretaceus there was probably an outpouring of basalts which in their chemical composition are similar to continental ones which underwent greenstone metamorphism (Udintsev, Kurentsov, 1990). Later, in the Paleo-eocene, there were outpourings here of oceanic traps with the fullest reworking of the crust during the process of further riftogenesis. This process continued down to our time in the axial zone of the equatorial segment.

In assuming such a sequence of the development of the equatorial segment, we posit the following role of the magmatic associations described earlier. In the lower part of the basement there are ancient magmatic rocks: undepleted, weakly and highly depleted hyperbasalts, alkalic gabbroids, ferro-titanic gabbroooo-amphibolites, alkalic basalts (of the type of continental traps). Higher, during the initial stage of oceanization, ferruginous oceanic traps were formed. The riftogenous magnesial basalts, associated with shallow lherzolite-tractolite intrusive complex, form the uppermost part of the basement. Structurally it is laid on the oceanic trap formation in the axial part of the equatorial segment.

Chapter 12

Ultrabasic Mylonites of Saint Paul Island

Ye.A.Denisova

St. Paul's Rocks is a group of tiny islets and rocks located in the equatorial part of the Atlantic ocean. These are the summits of the transverse ridge in the northern fringing of the large transform fault gorge of the same name (Plate 19, fig. 1). We understand the latter as being the two closely lying faults (Emery and Uchupi, 1984), along which the axial rift valley of the Mid-Atlantic Ridge experiences sinistral displacement for 450 and 100 km (along the northern and southern faults, respectively).

The St. Paul's rocks have long interested scientists, They were first studied in 1832 by Charles Darwin from the Beagle. Some specimens were later collected by the Challenger Expedition in 1873, and by the Quest Expedition in 1921, Detailed study of the collections performed by Tilley (1947) has convincingly shown the dynamic origin of the rocks defined as mylonitized dunites and peridotites. Metrological and petrochemical examination of the rocks was subsequently done by Melson and Thompson with co-authors (Melson et al., 1967, 1972), who carried out the rock geologic survey in 1966 and described the mylonitic outcrops, and were also described by Roden et al. (1984), and Bonatti (1990).

In 1988, the sea floor in the vicinity of St. Paul's Rocks was dredged by the expedition of the 7th cruise of the r/v "Akademik Nikolaj Strakhov" (Geological Institute of the USSR Acad. Sci.). An abundance of ultrabasic mylonites has been recovered from the northern slope of the transverse ridge in the St. Paul's fracture zone (f.z.) (site 15; 0°59, 3'N, 29°23.7' W; depth 3130-2680 m). I was granted permission to study the material by N.A. Kurentsova, a cruise participant, whose courtesy I sincerely acknowledge.

For detailed petrofabric examination four ultrabasic mylonite samples have been chosen, showing an exquisite fine banding, The bands (from 0.1 to 1-2 mm thick) differ in mineralogical composition and, accordingly, in colour. The olivine-rich bands are grey, while a high content of spinel and/or secondary serpentine with magnetite contributes a darker hue to them. The microscopic examination reveals distinct yellowish and brownish bands with a brown hornblende. The bands are coloured greenish by the secondary amphibole and chlorite.

The primary mineral assemblage is determined by the composition of porphyroclasts set in the mylonitic matrix whose mineralogical composition is difficult to determine due to a small grain size (from 0.02 to 0.04 mm on the average, sometimes to 0.005 mm). In the mylonites studied, the porphyroclasts are most commonly represented by olivine, with subordinate pyroxenes, brown hornblende, and spinel. Their content varies in different samples, though the assemblage as a whole indicates that the source rocks, after which mylonites formed, were mainly hornblende peridotites.

The porphyroclasts are generally rounded or ellipsoidal in shape, with smoothed edges. Also common are more complicated irregular grains (Plate 18, fig, 2). Their size ranges from 0.1 to 2 mm, rarely reaching 5 mm. The grains are isometric to strongly elongated in shape, with a length-to-width shape ratio of 7:1. The grain foliation is usually parallel to the banding plane, though some porphyroclasts are oriented at different angles to the banding, sometimes at right angles. T.N. Kheraskova believes this reliably to confirm that the rock has a tectonic and not a sedimentary origin.

The olivine porphyroclasts, up to 2 mm long, are unevenly serpentinized. One third of them are virtually devoid of serpentine, another third contains it from 10 to 20%, while the rest of porphyroclasts, confined to darker bands, show stronger alternation with magnetite segregations. The serpentine and magnetite veinlets in porphyroclasts are usually subparallel to the banding. Provided large olivines are extended over two or three bands, the degree of their serpentinization drastically changes at the bands boundaries. Consequently, the main serpentinization was presumably synchronous with the concluding phases of mylonitization. At the same time, there typically occur thin serpentine veinlets containing magnetite (evidently of later origin), which cut the banding at high angles.

The olivine porphyroclasts show abundant traces of strong intragranular deformations: undulatory extinction, subgrains, and also kink bands, indicating intracrystalline slipping as the deformation mechanism. The kink band boundary examination has shown the slip to occur in Ng=[100] direction which takes effect at high-temperature, usually mantle, plastic deformations. This suggests the porphyroclasts and their reformational structures to be relic and characterize the tectonized peridotites which are the source rocks for mylonite formation.

P y r o x e n e s a n d h o r n b 1 e n d e form smaller porphyroclasts, usually under 1 mm size, though separate pyroxenes may amount to 5 mm. The pyroxenes show enstatite and diopside (2V{EMBED Equation |}= 50-55°}) in different relations with each other. Porphyroclasts of possibly titanium (kaersutite) hornblende (2V{EMBED Equation |}=83-84°) exhibiting a brown to pale-brown and yellowish pleochroism, are unevenly distributed,and usually confined to bands. The deformed porphyroclasts of these minerals are less common, possibly due to their smaller size.

S p i n e 1 s represent dark-brown (almost opaque), brown and olive-brown chrome-rich spinels, as well as olive and greyish-green grains, tentatively assigned to the alumniferous spinel group. The latter occur m specimens containing hornblende. The spinel porphyroclasts are usually under 0.5 mm, and some grains goup to 2-4 mm.

The mineral chemical composition conforms with that in the tectonized ultrabasic ophiolite units, while a high aluminous content of both pyroxenes and spinels indicates a relatively undepleted composition of the source peridotites (Melson et al., 1972; Roden et al., 1984; Bonatti, 1990). According to Roden et al. (1984), the mineral equilibriums in the pyroxene porphyroclasts were set at the temperature of 940-1010°C (the grain cores) to less than 800°C (the grain edges).

Besides porphyroclasts, the studied mylonites contain more or less rounded fragments of rocks (mostly dunites), some 3 to 5 mm across. According to kink bands, the direction of intracrystalline gliding in the olivines from rock fragments was similar to that in the olivine porphyroclasts.

To study the ultrabasic mylonite microstructure, the specimens were cut in three mutually-perpendicular planes: parallel and perpendicular to the banding (Plate 18, fig. 2). The banding plane usually shows a distinct lineation manifested by a subparallel distribution of elongated porphyroclasts of different minerals and of small lenses of different composition. Only one sample is poorly lineated. The section, perpendicular to the banding and parallel to the lineation, is the most informative for microstructural examination. In this plane, porphyroclasts are most elongated, and the long axis orientation is most perfect. The banding, showing the alternation of subparallel thin stripes and lenses enriched in some or other minerals, is clearly observed to flow smoothly around porphyroclasts and rock fragments. Some stripes are slightly undulated, sometimes split; their thickness varies smoothly, up to a total disappearance. As a whole, a highly plastic, flowing mylonitic microstructure takes place. At the same time, in this plane isometric or elongated porphyroclasts occur, lying at high angles to the banding. The third section, perpendicular both to the banding and lineation, shows either a well- or a poorly-oriented microstructure (Plate 18, fig. 2), the banding strongly complicated by branching of rare stripes of different composition.

All specimens thus exhibit distinct mylonitic features. However, their linear and plane structural components are developed to different extent one of the samples has a good foliation and almost no lineation; another one, in fact, exhibits a clear linear orientation and a very poor plane one . In the rest of the specimens, both the foliation and lineation are well-developed. Such variations, generally typical of tectonites, reflect the stress field variability characteristic of the plastic flow process. The environmental changes could be both temporal and spatial.

Petrofabric examination implies the identification and analysis of preferred optical orientations of rock-forming minerals. We examined one of the mylonites formed after a hornblende peridotite. It shows clinopyroxene sharply predominant over orthopyroxene, brown hornblende, and alumniferous spinel, olive and greyish-green in transparent light. The preferred orientation of the olivine indicatrix axes in porphyroclasts is utterly missing in the obtained petrofabrics (Plate 18, fig. 3). The diagrams for each of the axes feature from six to eight 3% maxima scattered over its area and unrelated to the banding and lineation. On the separately constructed orientation diagrams for elongated olivines lying in the banding plane (some 55% of total number), from nine to eleven 3% maxima are randomly arranged and occur at regular intervals. Consequently, the shape of most grains is unrelated to their optical orientation. At the same time, some peculiarities of the diagrams permit the conclusion that part of the grains are foliated along Ng axis and elongated along Ng or Np. Due to their small size, however, we unfortunately could not measure the olivine orientation in the matrix. At one time, Tilley (1947) indicated their significant orientation, with Np perpendicular to the banding. However, he did not petrofabric diagrams, or described them; most probably he did not measure the orientation on a Universal stage.

The above permits the following interpretation of the history of formation of the St. Paul's mylonites. Their source rocks were ultrabasic ones, plastically deformed in the upper mantle. One of the mechanisms of their early high-temperature deformations was intracrystalline slipping (in [100] direction in olivine). Having suffered strong fluid effect in the course of the St, Paul's f.z. origination, the ultrabasics transformed into the hornblende ultrabasics (Bonatti, 1990). It should be noted that the ultrabasic mylonites frequently contain amphibole or other water-bearing minerals both when they transect massifs composed of waterless ultrabasic parageneses (Miller and Mogk, 1987) and while occurring in the soles of large ophiolite sheets (Nicolas et al., 1980; Girardeau and Nicolas, 1981; Cannat and Boudier, 1985). Later, in the course of the St. Paul's f.z. evolution, the hornblende ultrabasics were intensely mylonitized. This presumably occurred in the area of plastic deformations due to syntectonic recrystallization of minerals. At first, the latter embraced small areas and zones - the traces of these initial mylonitization stages remain visible in the rock fragments included into mylonites as well as in several highly serpentinized harzburgite specimens dredged at the same site. Later, these areas and zones expanded, extended and merged with each other, so that the relic grains of the source ultrabasics (porphyroclasts) were included in the recrystallized matrix.

According to the average size of olivine neoblasts (0.03 mm), the maximum differential stress calculated from the Post and Mercier formulas (Boudier and Nicolas, 1980), at which the St. Pauls's mylonites formed, made up 195 and 254 MPa, respectively. The following stress values have been obtained for ultrabasic mylonites in other regions: 50-100 MPa and more - in the soles of ophiolite sheets of Newfoundland (Girardeau and Nicolas, 1981); 120 MPa - in the marginal mylonitic zones of the Lanzo massif (Boudier and Nicolas, 1980); 200 MPa - in different scale zones, transecting massifs of the Antalya ophiolite complex (Reuber, 1984); 275 MPa - in a thick mylonitized pendotite zone, crosscutting the Ingalls ophiolites (Miller and Mogk 1987). In the rare thin (a few millimetres to several centimetres thick) small mylonitic zones of the Kraka massif (South Urals), concordant with the peridotite banding and foliation, the intensive recrystallization is manifested by the maximum grain size reduction (to 0.01-0.05 mm and less), indicating the local stress rise to 200-300 MPa. Thus, the tentative range of differential stress necessary to ensure the ultrabasic mylonitization is 100-300 MPa. It should be recalled that the asthenosphere flow stage stress is commonly under 40-55 MPa.

In several regions, the petrofabric study of ultrabasic mylonites has yielded rather good preferred optical orientations of enstatite (Boudier, 1978; Nicolas et al., 1980), as well as much poorer olivine orientations (Boudier, 1978; Nicolas et al., 1980; Reuber, 1984; Cannat and Boudier, 1985). These were interpreted as resulting from the intracrystalline slipping. Random olivine orientation of the St. Paul's mylonites is possibly due to extremely intense plastic deformations combined with the cataclastic flow (intergranular sliding) elements. The effect of the latter mechanism is revealed by rotation of porphyroclasts in the course of differential displacements along the adjacent bands. This accounts for the fact that porphyroclasts are partly oriented at high angles to the banding and partly isometric in shape. The olivine elongation tends to increase and subsequently decrease in the one-metre-thick mylonitic zone of the Finero massif, studied in detail, toward its central, most deformed part (Brodie, 1980). This is accompanied by a gradually decreasing size of porphyroclasts and neoblasts, and the latter's growing abundance? The shape of some porphyroclasts in the St. Paul's mylonites is quite unusual for mantle peridotites - it is foliated along Ng or elongated along Np. The appearance of grains whose shape is unrelated to their optical orientation or isometric is probably due to Iong-term processes of rotation of porphyroclasts and recrystallization of their edges.

Thus, the ultrabasic mylonites examined bearing traces of early high-temperature upper-mantle deformations have been shown to suffer intensive plastic deformations in the lower crustal horizons during the St. Paul's transform f.z. generation and the source peridotite mylonitization. The syntectonic recrystallization intricately combined with the cataclastic flow elements has yielded a typical mylonitic structure and led to the destruction of the primary preferred optical orientations of minerals. No new petrofabric of porphyroclasts has been created. The modern occurrence of mylonites in the St. Paul's Rocks (Melson et al., 1967, 1972) discordant to the transform fault orientation suggests complicated surface-directed tectonic movements of the rock mass.

Chapter 13

Native Metals in the San Paulo Fault Basalts of the Atlantic Ocean.

L.E.Shterenberg, G.B.Udintsev, M.P.Antipov, B. I.Voronin,
D.I. Frikh-Khar, E.A.Korina, A.L.Sokolova, N.A.Kurentsova.

Recently, increased attention has been focused on native metals discovered in rocks of differing genesis in the continental Earth block. The published work of M.I.Novgorodova (1983) contains abundant factual data on these formations, and also discusses problems related to their chemical-mineralogical composition, structure and genesis.

There are significantly fewer contributions which examine the occurrence of native metals in a variety of oceanic sediments. Few publications dealing with this problem are known today, Among them are a monograph by G.N.Baturin (1986), papers by G.N.Baturin et al. (1984), G.Yu.Butuzova et al. (1987). L.E Shterenberg and G.L.Vasilieva (1979), L.E.Shterenberg et al. (1981,1986,1988), O.E.Yushko-Zakharova et al.(1984), P.Rena (1988). These works provide information on the composition and structure of the native metal particles in Fe-Mn nodules; metalliferous and pelagic sediments accumulated near the tectonically-active zones, in hydrothermal formations and the Red Sea sediments. No data, however, are available today on native metals in the basalts composing the second oceanic crustal layer. For our study we have chosen the basalts (both affected and unaffected by hydrothermal fluids) recovered during the 7th cruise of R/V "Akademik Nikolaj Strakhov".

Their first portion was dredged at Station 55 (0°49,0' - 0°51,2'N, 30°13,7'- 30°14,2'W) where the lower northern flank of the Saint Paul facture zone (f.z.) was dredged near the point of its intersection with the rift valley at the depth of 3900-3300 m. The second one was obtained in the same cruise at Station 21 (0°53,3' - 0°54,5'N, 27°06,5' - 27°37,1'W) at a depth of 3740-3630 m. The basalts unaltered by hydrothermal solutions were dredged on the southern flank of a small fault located to the south of the Saint Paul f.z..

The combined methods were used to study the basalts. These included microscopy (studies of thin and polished sections, X-ray analysis, DPOH-2,Cu,Y-ray, RAK,Co,Ka-ray), wet chemistry method, spectral analysis (JMA-1), electron-microsound "Kameka" (MS-46),etc.

After macro- and microscopic studies, the samples were pestled in an agate mortar, washed through the nylon net with a cell diameter of about 0.1 mm. After drying, the fraction more than 0.1 mm in size was examined under a binocular magnifier and the particles of native metals, sulfides and magnetite as well as other mineral formations were selected to be further examined in detail. The sorted out particles of native metals were rolled up in the rubber adhesive and subsequently analyzed in an X-ray chamber (RAK).
The time of exposure equaled 10-15 hrs. After the X-ray analysis, the rubber adhesive was removed from the metal particles acetone or gasoline. The particles were placed in special moulds and filled with an epoxy matrix resin. These moulds containing the sealed metal particles were used to produce the polished sections. The latter were examined first under the microanalyzer "Kameka". Subsequently, they were subjected to the semi-quantitative spectral analysis (JMA-1).

Another technique was used to analyze sulfides, iron-manganese oxides, magnetite, epidote, chlorite,etc. discovered in the samples. If the amount of the compound extracted from basalt was small, a part of it was analyzed in the RGK chamber, and special balls were made. If the amount of the sampled material was large enough, it was also X-rayed with a difractometer DPOH-2. Studies of thin sections have shown that the altered basalts were represented by clinopyroxene-plagioclase metabasalts containing phenocrysts of clinopyroxene, plagioclase, olivine and spinel.

Metabasalts have a porphyric structure. The groundmass contains abundant microlites of plagioclase with clinopyroxene and olivine crystals occurring between them.

Both the groundmass and phenocxysts have been affected by secondary alternations related to hydrothermal processes. Chlorite, saussurite aggregates, epidote, a rarer carbonate develop after plagioclase. Pyroxene is replaced by amphibole and chlorite.

Chlorite also fills in caverns and forms thin secondary veins along with calcium carbonate. Considerable importance of chlorite in the structure of metabasalt is evidenced by the diffractometer analysis.

Among the ore minerals resulting from hydrothermal processes, pyrite is the most widespread (Table 1). It does not usually form large uninterrupted emanations, being developed metasomatically in the groundmass after plagioclase, pyroxene and several other minerals.

The results of semi-qualitative analysis (JMA-1) illustrate that the pyrite recovered from metabasalts at Station 55 does not contain such elements as copper, zinc, nickel and cobalt. At the same time, a certain enrichment of pyrite with beryllium and silver should be noted.

The olivine-plagioclase tholeitic basalts found at Station 21 have a porphyric structure. Olivine is very slightly altered.

The plagioclase crystals are up to 1 mm long, The matrix has a chiefly microlitic structure. Together with basaltic blocks very slightly altered by halmyrolysis whose surface is covered by Fe-Mn crusts (1-3 mm), the dredging has yielded Holocene-Pleistocene carbonate oozes, containing well-preserved foraminifera and coccoliths. A diffractogram of a fine-grinded basalt sample shows that it contains no chlorite in contrast to metabasalts at Station 55. The pestling of metabasaltic samples dredged at Station 55 has yielded particles of native metals. Micrographs show these particles very clearly. It should be noted however that though the plate shows an abundance of the native metal particles, they are rather uncommon in the meta- and tholeitic basalts studied. A great bulk of initial material had to be handled (crushed and pestled in an agate mortar). The particles found were commonly very small (not more than 0.1 mm); on a few occasions they were slightly larger.

The particles of extended shape (threads, short sticks, needles, ovals, etc.) whose iron content approached 99.0-99.7 mass % were most common.

In the reflected light under the microscope and the binocular magnifier it is sometimes visible that the native iron particles are welded with minerals and the metabasalt matrix. The presence of “foreign” material is detected by the X-ray analysis of native iron particles, though it is virtually impossible to identify it due to very weak debayegram lines.

The data yielded by the MS-46 X-ray microanalyser add considerable evidence supporting the native iron particles and the enclosing metabasalt intergrowth,

The results of sounding of metabasalt adjoining the native iron particle have shown it to contain (in mass %): SiO_2 - 25.9; CaO - 25.1; TiO_2 - 23,06; FeO - 17.30; MgO -1.17; Al_2O_3 -2.53.

The second in distribution native metal discovered in the grinded metabasalt dredged at Station 55 is the original zinc. Its particles are somewhat irregular in shape. They are colored light grey.

In samples clear from mechanical admixtures, the zinc content is 95-96 mass %. A minor part of zinc is probably bounded with zincite (ZnO) which is sometimes observed in the studied native zinc particles in the form of strips and most thin rims.

Semi-quantitative spectral analysis of native zinc samples has revealed them to contain (in mass 0/0): copper - about 0.03; manganese - 0.02 and silver -0.002. Cobalt, nickel, plumbum, chromium and vanadium are either missing or minor.

The detailed study of native metal particles magnetized to a greater or lesser extent (which was determined with a magnetized needle) has shown them to contain native iron and zinc intergrowth.

Sometimes when the particles of both native metals are large in size their conjunction is distinct Examination of thin sections of such samples shows that zinc has emanated somewhat later than native iron. The X-ray analysis clearly indicates the existence of two separate independent phases - native iron and zinc (Table 1). Intergrowth of these minerals is confirmed by the X-ray microanalyzer (MS-46) which determines the occurrence of pure iron and zinc admixtures.

In other cases, the native zinc forms in separate microareas taken in absorbed electrons (c) and in the X-ray emanation (T-Fe; Y-Zn), at magnification 66<x> illustrate that the general iron field comprises only very small sites with a high zinc content (amounting to 93-94%), Neither has the X-ray analysis shown any mixed iron and zinc phases. The debayegram displays prominent native iron lines and very weak zinc lines. The relationship revealed by the microscopic study suggests the zinc was formed after the iron.

The third place in the distribution in the Station 55 metabasalt belongs to native aluminum It is a light soft metal.

One side of a native iron particle is frequently smooth while an- other has a coarser scratched surface. The aluminum content in the samples amounts to 95-96 mass %. The semi-quantitative spectral analysis of aluminum particles has shown that they contain (in mass %) on the average: manganese - 0.05; iron - 0,3; copper - 0.07; silver, zinc, cobalt, nickel, vanadium and chromium were undetected.

The X-ray analysis of aluminum particles has revealed that both the native aluminum and native iron occur in close association with the matrix and some other metabasaltic minerals (Table 1).

Thus, iron, aluminium and zinc particles have been discovered by pestling of the metabasalt samples recovered at Station55. Sometimes the metal particles were coalesced with metabasalt. Some micrographs show zinc to have been formed after the native iron (and possibly aluminum) with which it is aggregated. The pestling of tholeitic basalt dredged at Station 21 (in contrast to metabasalts) has yielded only the native iron particles. Their iron content according to the "Kameka" electron microsound analysis amounted to 98-99%.

According to the semi-quantitative spectral analysis (JMA-1), the above particles do not differ essentially from the native iron ones recovered from the Station 55 metabasalt in the content of minor elements.

As was noted above, several publications dealing with the study of formations of different genesis, metalliferous and pelagic sediments and the oceanic Fe-Mn nodules discuss the composition and structure of native metals and provide an inference on their genesis. The P.Ronas (1988) generalizing papers contain data on specific ore mineralization properties in the oceanic spreading centers where he reports the occurrence of such native metals as silver and copper among the thermal spring formations.

According to Baturin G.I.(Baturin, 1988; Baturin et al., 1984), the metals found among the oceanic Fe-Mn nodules have a biochemical origin. The authors consider that the volcanic-hydrothermal formation of native metals in Fe-Mn nodules is hardly probable since they are located far from the volcanic chambers and tectonically active rift zones.

In the opinion of G.Yu. Butuzova et al. (1987) and L.E.Shterenberg et al. (1979, 1981,1986,1988), the formation of native metals in oceanic sediments and in the Red Sea oozes may be caused by hydrothermal processes. The authors suggest that the smallest particles of Al, Fe, Cu, Zn, Au, Ag, Pb and other metals could be transported to the near-surface zone by hydrothermal springs. Also noteworthy is a viewpoint by M.I.Novgorodova (1983). She believes that the metals encountered in the various rocks of different composition on the continents usually are formed in reduction conditions in the course of the basaltoid magma origination and subsequent differentiation at the expense of hydrothermal activity with the active participation of dry reduction fluids which primarily come from deep sources.

The data obtained on the occurrence of native iron, aluminum and zinc in metabasalts (while tholeitic basalts contain only native iron) allow for the assumption of a variety of their formation modes. The native iron and aluminum have possibly formed as early as the pre-crystallization phase of magmatic melt in the intermediate chamber and are later discharged into the oceanic subsurface zone, where- as zinc, epidote, chlorite and some other mineral have formed under the effect of hydrothermal processes.

It is true that the data we suggest here are insufficient for resolving the problem of the genesis of the native metals discovered in the oceanic conditions. However we believe that these basite inclusions/formations should be given more attention.

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Table 1. X-ray pattern of Fe sulfides and native metals from station 55 metabasalts.

Pyrite				Native iron						Native zinc						Native aluminium					
St. 55		V.I.Mikheev (1957)		St. 55		V.I.Mikheev (1957)		In junction with meta-basalt		St. 55		V.I.Mikheev (1957)		In junction with meta-basalt		St. 55		V.I.Mikheev (1957)		In junction with meta-basalt	
I	d*	I	d	I	d	I	d	I	d	I	d	I	d	I	d	I	d	I	d	I	d
1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22
-	-	-	-	-	-	-	-	2	3.52	-	-	-	-	-	-	-	-	-	-	-	-
-	-	-	-	-	-	-	-	I	3.35	-	-	-	-	-	-	-	-	-	-	-	-
I	3.I3	2	3.I0	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
I0	2.68	8	2.69	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	2	2.96
-	-	-	-	-	-	-	-	2	2.54	-	-	-	-	-	-	-	-	-	-	3	2.54
-	-	-	-	-	-	-	-	-	-	3	2.44	4	2.47	3	2.48	-	-	-	-	-	-
I0	2.39	3	2.4I	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
-	-	-	-	-	-	-	-	2	2.36	-	-	-	-	-	-	-	-	-	-	-	-
-	-	-	-	-	-	-	-	-	-	3	2.30	5	2.3I	4	2.30	I0	2.33	I0	2.34	I0	2.33
3	2.20	7	2.20	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-
-	-	-	-	-	-	-	-	-	-	I0	2.08	I0	2.09	I0	2.09	-	-	-	-	-	-
-	-	-	-	I0	2.02	9	2.02	I0	2.02	-	-	-	-	-	2.0I	8	2.03	9	2.03	8	2.01
6	I.90	6	I.90	-	-	-	-	2	I.95	-	-	-	-	-	-	-	-	-	-	-	-
2	I.75	4	I.79	-	-	-	-	-	-	4	I.68	8'	I.68	3	I.68	-	-	-	-	-	-
I0	I.62	I0	I.62	-	-	-	-	-	-	4	I.68	8	I.68	3	I.68	-	-	-	-	-	-
2	I.55	3	I.56	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	I	I.54
3	I.49	4	I.49	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	-	I	I.50
4	I.40	6	I.44	4	I.43	7	I.43	4	I.43	-	-	-	-	2	I.43	6	I.43	8	I.43	6	I.43
I	I.23	3	I.23	-	-	-	-	3	I.29	2	I.32	9	I.33	6	I.33	-	-	-	-	-	-
2	I.I2	4	I.I2	-	-	-	-	-	-	2	I.I24	4	I.23	I	I.23	6	I.27	I0	I.22	8	I.21
3	I.II8	4	I.I7	6	I.I6	I0	I.I6	8	I.I7	6	I.I4	9	I.II	8	I.I7	3	I.I5	5	I.I6	4	I.I7
2	I.I5	2	I.I5	-	-	-	-	-	-	4	I.I3	6	I.I2	2	I.I2	-	-	-	-	-	-
3	I.I0	5	I.I0	-	-	-	-	-	-	2	I.I0	5	I.09	2	Io05	-	-	-	-	-	-
-	-	-	-	-	-	-	-	-	-	3	I.04	5	I.04	I	I.04	-	-	-	-	-	-
-	-	-	-	4	I.0I	7	I.0I	4	I.0I	-	-	-	-	4	I.0I	2	I.0I	4	I.0I	6	I.0I

I - line intensity; d - interplane distance

Table 1 X-ray pattern of Fe sulfides and native metals from station 55 metabasalts

Chapter 14

New Data of Pleistocene Stratigraphy and Paleoceanology of the Equatorial Atlantic

N.S.Oskina, O.B.Dmitrenko

The investigation of the Atlantic bottom sediment cores dates back more than one hundred years. At first short cores (up to one meter long) were taken. The stratigraphic volume of cores was small, Now the length of the studied cores is 4 to 5 meters and the stratigraphic volume is 100-150 thousand years. Therefore in the past few decades important paleoceanologic results were obtained for the Holocene and late Pleistocene. "At present paleoceanologic reconstructions of the world ocean were proposed (CLIMAP,1976,1981,Barash,0skina,1979,Barash et al,1989), but without reconstruction for the middle and early Pleistocene, because the investigators did not have sufficiently long and complete cores. The distribution of the cores studied is not equal in different regions; for example the North Atlantic and Equatorial Pacific are well studied, but the South Atlantic, Indian ocean, Arctic, and Antarctic regions are not sufficiently studied. Interesting material was obtained through deep-sea drilling, but the interval between probes turned out to be usually large, and hence some Pleistocene stages were not investigated. Summarizing the reasons. we may say that long and complete Pleistocene cores are rare and very interesting for stratigraphy and paleoceanology. In the equatorial zone the cores do not penetrate the middle Pleistocene horizon because the equatorial zone is always characterized by a high rate of sedimentation. During the 7-th cruise of r/v "Academik Nikolaj Strakhov" eleven cores were obtained, among them two cores containing the middle Pleistocene sediments.

The authors studied planktonic foraminifers and nannoplankton assemblages in the sediments from core T-11. The interval between the samples usually reached 10 cm . The tests of planktonic foraminifers had good preservation. The samples of the bottom sediments were washed, using the O,1 mm mesh, dried and the fractions obtained were examined. The fractions were thoroughly mixed and divided until the remaining split contained at least 300 tests. All the tests were identified and counted in this split under the microscope. Additionally, all the fractions larger than 0.1 mm were investigated to discover the rare species which might have been omitted from the split. Paleotemperature estimates were made, using the method of M.S.Barash and N.S.Bluym (1975) , which is based on quantitative counts of the ratio of different species in the assemblage (Barash, 1988). The fractions are also composed with the tests of benthic foraminifers, radiolaria and sometimes mineral grains. The ratio of planktonic and benthic foraminifers was accounted for in each sample because it indicates the dissolution of the deposit.

The core T-11 (6°27' N, 31°28' W, depth 3420 m) was taken from the north-eastern part of the Equatorial segment of the Mid-Atlantic Ridge, close to the Sierra Leone fracture zone. The length of the core is 4 meters. The core is composed of light homogeneous carbonate ooze. The studied samples contain rich assemblages of planktonic foraminifers. The average number of species in most samples is 23-25. In the lower part of the core the samples have a maximum number of 28-29 species.

The same planktonic foraminifera species were in all the samples, but in different proportions. In the lower part of the core Neogene species were present - Globoquadrina acostaensis, G. humerosa, Globorotalia miocenica, G. cultrata, G. Tosaensis, Pulleniatina primalis, Globigerinoides obliquus. Probably the Neogene species were reworked.

It is interesting to note the presence of Globorotalia crassaformis hessi in the interval of 290-400 cm, because it is one of the Pleistocene index-subspecies, We also noted the presence of another index-subspecies - Globigerina calida calida - in the interval of 60-280 cm. The level of the first appearance of Globorotalia crassaformis hessi in the tropical Atlantic (Bolli, Premoli Silva, 1973) is about 1.4 Ma years and the level of the first appearance of Globigerina calida calida is near 140 thousand years (Bolli, Premoli Silva, 1973). In the subtropical zone of the south Atlantic (Barash, 0skina,Bluym,1983) Globorotalia crassaformis hessi appeared in the end of the Pliocene -1,9 Ma years ago and disappeared near 0.3 Ma years ago in the middle of the Pleistocene. The second index-subspecies - Globigerina calida calida - appeared near 0.8 Ma years and disappeared in the late Pleistocene. It is evident that both subspecies coexisted during the middle Pleistocene, therefore authors state that the lower part of core T-11 is of the mid- Pleistocene.

Other index-species - Globoquadrina hexagona and index- subspecies Globorotalia menardii flexuosa -are widespread in the samples of the core T-11. The last occurrence of the former species is marked at 30 cm and the last occurrence of the latter species is marked at 20 cm. According to micropaleontological data supported by radiocarbon data (Barash, Kupsov, Oskina, 1987), both species disappeared at the end of the late Pleistocene between 20 and 14 thousand years ago in the equatorial zone of the Atlantic. The Holocene index-species Globorotalia fimbriata appeared at 20 cm in the studied core. According to all obtained data the location of the Holocene-Pleistocene boundary is near 20 cm.

Using only the biostratigraphic method based on the distribution of the index-species of planktonic foraminifers we cannot determine the location of the late-middle Pleistocene boundary in our core, but the paleotemperature method and the oxygen-isotope method help us to determine this location. This boundary corresponds to stage 5e of the oxygen-isotope curve, or the maximum of the last Interglacial. However, we will discuss the location of middle-late Pleistocene in the studied core further.

It is possible to detect the location of some age boundary by means of the nannoplankton analysis, too. A great number of well-preserved nannoplankton occurs in the studied samples.

The history of the investigation of the calcareous nannoplankton in the Atlantic ocean dates back to the 19th century (Wallich, 1877), and then other investigators (Ostenfeld, 1899, 1910, Lohmann, 1912, Huxley, 1968) later continued this work. In the 50-s of this century the electronic microscope was actively used in studies. It was very important for the development of nannoplankton investigation (Deflanre,Fert,1952, 1954). The second important factor in the development of the nannoplankton method was deep-sea drilling, which began in the year 1968. The first nannoplankton zonal scale was proposed by W. Hay (Hay et al., 1967) on the material of the Caribbean sea and the Gulf of Mexico. Three zones (Plate 20, Table 1) Emiliania huxleyi, Gephyrocapsa oceanica, G. caribbeanica were involved in this scale. Later this scale was used by K. Geitzenauer (1969) for investigations in the subantarctic regions of the Pacific. D. Bukry (1978) and E.Martiny and T.Worsley (1970) used this scale as the basis for the construction of their own scales of the World ocean. The latter scales were used to study the sediments of different regions, and they were detailed, At present the scale introduced by D.Bukry (1978) with age data by H.Okada and D.Bukry (1980) were used for investigation in the low and middle latitudes, while the scale by E.Martiny, and the first primitive scale by D.Bukry were used to work in the high latitudes.

The Pleistocene nannoplankton datum levels are summarized in Table 2, Plate 20..

Nannoplankton in the sediments of the core T- 11 was studied with the light microscope “Amplival” (magnification 1500) and with the scanning electron microscope JSM-U3 (magnification 20 000). Nannoplankton and planktonic foraminifers were obtained in the same samples. Nannoplankton was represented by more than 20 species (Plate 20.).

Little coccoliths belonging to genus Gephyrocapsa dominated. Their quantity increases down the core, Other dominating species are Emiliania huxleyi (Lohm.) Hay, Mohl., Umbilicosphaera sibogae (Weber van Bosse) Gaarder, Neosphaera coccolithomorpha Lecal- Schlauder.

The species Cyclococcolithus leptoporus (Murr.,Blackm.) Kpt., Helicosphaera carteri (Wall.) Kpt., Syracosphaera pulchra Lohm., Discolithina japonica Takayama, Umbellosphaera irregularis Paasche were found in all the samples but their quantity was smaller. The species Scapholithus fossilis Defl., Discolithina multipora (Kpt.,Defl.) Mart., Umbellosphaera tenuis (Kpt.) Paasche, Thoracosphaera heimi Kpt. occupied a small interval. The species Gephyrocapsa protohuxleyi McIntyre, G.ericsonii McIntyre occurred in the lower part of the core only. The species Ceratolithus cristatus Kpt., Hayaster perplexus (Brain.,Ried) Bukry were rare in the studied sediments. The species Helicosphaera sellii Bukry was marked on 380 cm, but it was possibly redeposited.

The upper part of the core (0-120 cm) belongs to the zone Emiliania huxleyi Acme due to an increase in the quantity of the index-species in this horizon. The next horizon (120-310 cm) belongs to the zone Emiliania huxleyi, as the index-species are present only (Gartner, 1977) in this horizon. The first appearance of Emiliania huxleyi was marked at 310 cm (Plate 20, fig. 1). The index- species is absent in the lower part of the core, which belongs to the Gephyrocapsa oceanica zone (Gartner,1977).

Gephyrocapsa protohuxleyi is also an important species of Pleistocene stratigrafy. The last occurrence of this species in the studied core was marked at 220 cm. The calculated age of the layer at 220 cm was 180-190 thousand years, based on the fact that the rate of the sedimentation was continuous, and on the age of the zonal boundaries . The age of zonal boundaries was determined with the help of the material from other regions; hence our calculation was not absolutely correct,

The age of the last occurrence of Gephyrocapsa protohuxleyi is 620 thousand years in the subtropical zone of the North-West Pacific (Matsuoka,Okada,1988). In the Atlantic and Indian oceans the age estimates of this datum level are not so great. The age of the last occurrence of Gephyrocapsa protohuxleyi is 100 thousand years in the Rio-Grande region and in the Sierra-Leone region

(Dmitrenko,1987, Samtleben,1978), 70-75 thousand years in the Red Sea (McIntyre,1969), and varies from 50 to 70 thousand years in the subtropical zone of the Indian Ocean (Dmitrenko,1990). It is evident that the age of this datum level decreases from high latitudes toward low latitudes.

The age of the first appearance of Umbellosphaera irregulars was determined as 190-240 thousand years in the North Atlantic (Pujos, 1985), In other articles the age of this datum level was estimated as 244-279 thousand years. The first appearance of Umbellosphaera irregulars in the studied core is fixed at 340 cm. The calculated age of layer 340 cm is 290 thousand years. The age of the first appearance of Emiliania huxleyi varies from 230 to 270 thousand years (Dmitrenko,1987,fig.).

The samples contain a great number of warm-water nonresistant species of nannoplankton. Their abundance is probably due to the warm conditions that existed during the sedimentation and the weak dissolution of the deposit. There are tropical Hayaster perplexus, subtropical-tropical Umbellosphaera irregularis, U. tenuis, Scapholithus fossilis, Discolithina multipora and Oolithothus antillarum.

The changes in the diversity of nannoplankton species are the result of changes of the surface water conditions on the one hand, and changes of the conditions that existed during the earlier period of the sedimentation in the deposit on the other hand. The maximum of species diversity (13-15) occurs at 20 cm, 80 cm, 220-240 cm, 280 cm and 360 cm. The minimum of species diversity (less than 12 species) occurs at 40-60 cm, 100-110 cm, 130, 180, 250-260 cm, and 340 cm, The absence of warm-water nannoplankton and low species diversity (Plate 20, fig. 1) in the upper layer (0-1 cm) of the deposit are explained by the absence of a large part of Holocene sediments, formed in conditions of higher temperatures.

The curve of the species diversity has five maximums and six minimums. These changes in nannoflora are prior to changes of the surface-water paleotemperature, reconstructed on planktonic foraminiferal data (Plate 20, fig.2)

Thirty six samples were analyzed to study the fauna of the planktonic foraminifers and to reconstruct the surface water paleotemperatures. The recent mean year temperature of the surface water determined by the hydrological method is 26.5 C in the studied region. The reconstructed paleotemperatures vary from 22.5 -23.0 C to 29.0-29.5 C. The amplitude of the Holocene- Pleistocene paleotemperature changes reached 6-7 C. The paleotemperature curve has four peaks of low paleotemperature (Plate 20, fig.2) and five peaks of high paleotemperatures, when surface water paleotemperature was the same as the present temperature or higher. The lowest paleotemperature existed during the formation of the lower part of the core. Up the core the cold peaks become smaller - 22,0 C, 23.5 C and 22.5 C correspondingly. The thickness of the horizons with warm-water fauna is less than the thickness of the horizons with the cold-water fauna of planktonic foraminifers.

The biostratigraphic investigation of the sediments with both micropaleontologic methods allowed for the definition of the stratigraphic volume of the studied sediments as Holocene-middle Pleistocene and to correct some of the age boundaries based on the datum levels. The paleotemperature method allowed us to subdivide the core in greater detail than the biostratigraphyc methods. For this purpose the reconstructed paleotemperature curve was compared with the standard oxygen- isotope curves to determine the age of the corresponding peaks. Now the oxygen-isotope scale is well-known (Van Donk, 1976), and is often used to study the young Pleistocene deposits. In the interval corresponding to the entire Pleistocene near 1.8- 1.9 Ma years 41 stages were defined (Van Donk, 1976), and the late Pleistocene plus middle Pleistocene intervals involved 11-12 oxygen-isotope stages (Emiliani,1966, Shacleton,Opdyke,1973). The ages of the late Pleistocene and middle Pleistocene stages are summarized in (Barash, 1988).

The obtained paleotemperature peaks correspond to 1-9 stages of the standard oxygen-isotope scale. The clearest peak of the increase in surface water paleotemperatures takes place at 180 cm, and is correlated with substage 5e, and has the age of 125 thousand years. Substage 5e is the boundary between the middle and late Pleistocene. Upon the analysis of the paleotemperature curve we may correct the location of the Holocene-Pleistocene boundary. It lies at 17-18 cm.

The horizon 20-100 cm was formed during the last Glacial. The Glacial period consisted of two cold epochs, which correspond to the second and fourth stages of the oxygen-isotope curve, and a warm stadial, which corresponds to the third stage. The age of the boundary between stage 4 and stage 5 of the oxygen-isotope curve is near 70-75 thousand years. The age of the Holocene-Pleistocene boundary is 11.5 Ma years. The authors calculated the rates of sedimentation in the studied region. During the late Pleistocene the rate of the sedimentation was 0.13 cm/1000 years, and during the Holocene the rate was 0.13-0.14 cm/1000 years.

The lower part of the core T-11 was formed during the middle Pleistocene Glacial, which consisted of two cold epochs divided by an interstadial epoch. This period corresponds to stages 6,7 and 8 of the oxygen-isotope curve. The age of this middle Pleistocene Glacial is from 290 to 130 thousand years. The lower boundary of this horizon occurs at 370-380 cm, below this lies the horizon with warm-water fauna of the planktonic foraminifers. The lowest horizon corresponds evidently to stage 9 of the oxygen-isotope curve, The rate of sedimentation was 0.12-0.13 cm/1000 years during the last Interglacial (Riss-Wurm) and was about 0.13 cm/1000 years during the middle Pleistocene Glacial (Riss). The calculated rates for the large cold periods are similar but the rates of the short cold epochs vary from 0.8-0.9 cm/1000 years during the beginning of the late Pleistocene to 0.17-0.18 cm/1000 years at the end of the late Pleistocene.

The authors corrected the age of some datum levels by nannoplankton using the obtained paleotemperature curve. The last occurrence of Gephyrocapsa protohuxleyi is marked at 220 cm. This layer was formed on stage 6, the age of stage 6 was from 190 to 130 thousand years and the age of the layer 220 cm was about 160 thousand years, if the rate of sedimentation was continuous. This age estimate is close to the age estimate based on biostratigraphic data. The first appearance of Umbellosphaera irregulars was fixed at 340 cm and coincided with the boundary between stage 7 and stage 8 of the oxygen-isotope curve. The age of this boundary is 250 thousand years, The age of the first appearance of Umbellosphaera irregulars in the North Atlantic was also defined as 250 thousand years. The first appearance of Emiliania huxleyi is at 310 cm and corresponds to stage 7 of the oxygen-isotope stage. The age of this datum level may be determined as 210 thousand years, This age is equal to the age of the boundary of Emiliania huxleyi zone.(Plate 20, fig.1) .

Conclusions

1. The large changes of the average paleotemperature of the surface water took place during the late and middle Pleistocene in the studied region. The amplitude of the changes reached 6-7 C and was greater than the amplitude of paleotemperature changes in the central part of the subtropical and tropical zones in the Atlantic Ocean.
2. The exchange of normal and contrasting types of planktonic foraminiferal assemblages (Oskina, Bluym, 1984) marks the changes of the hydrological regime in the region studied during cold and warm periods. The frontal hydrological conditions probably existed during cold periods. This is supported by the high productivity of the planktonic foraminifers during cold periods.
3. The mid-Pleistocene paleotemperature changes were smaller than the late Pleistocene changes.
4. The upper part of Holocene sediments is absent.
5. The maximums of nannoplankton species diversity do not coincide with the maximums of the paleotemperature of surface water, while this often correlates with the cold water conditions.

Chapter 15

Rocks with Anomalous Age or Origin in the Equatorial Atlantic. Review

I.A.Basov.

Until recently, the Equatorial Atlantic remained insufficiently studied as compared with the rest of this oceanic basin, At the same time, it is characterized by a most unusual and complicated structure and, accordingly, geological history. The most peculiar features of this area include the block structural patterns of its bottom and lack of linear magnetic anomalies, whereas a system of distinct anomalies parallel to the mid-oceanic ridge axis and shorelines is observed both north- and southward of the equatorial Atlantic.

Geological-geophysical investigations, and primarily deep-sea drilling, revealed that different segments of the Atlantic differ in geological history, i.e. in age and in the succession of events, It was established, for instance, that the basin of North Atlantic is the oldest one and that during the early stages of its formation it developed as a part of the Tethys. The formation of deep-water basins with oceanic crust started here as far back as in the Late Jurassic and real oceanic conditions existed since the Neocomian, According to benthic foraminifers the depths in the western North Atlantic at that time were about 2000-3000 m (Basov, 1980). It should be noted, however, that some scientists argue that spreading in the Central Atlantic initiated much later. For instance, Storetvedt (1985) dated this event as the late Cenomanian.

In the South Atlantic, oceanic crust formation started later and proceeded diachronously in its different parts. According to Larson and Ladd (1973), Sclater et al.(1977) the initial opening of the South Atlantic occurred in the Valanginian starting with the spreading in its southernmost part, though real open-ocean conditions set in here at the Early/Late Cretaceus boundary after the final separation of the Falkland Plateau from South Africa.

The northern half of the South Atlantic up to the end of the Early Cretaceus was an area of accumulation first of terrigenous non-marine facies, and then of thick salt-bearing deposits. By the beginning of the Late Cretaceus the continental and lagoon environments gave way to a marine one with accumulation of relatively shallow-water black shales. In the Angola Basin, for instance, accumulation of these deposits continued in the first half of the Late Cretaceus. Real open-ocean environments expressed in the initiation of pelagic carbonate accumulation set in here only in the Santonian when a morphological barrier formed by the Walvis Ridge - Rio Grande Rise system disappeared as a result of proceeding spreading. While the Central and South Atlantic had already been formed as separate oceanic basins, the continental and marine shallow-water environments most likely continued to exist in its equatorial part.

There is no uniform point of view on the age of the Equatorial Atlantic opening. Benthic foraminifers show that some connection and water exchange between the Central and South Atlantic could have existed since the late Albian (Gradstein, 1978; Sliter, 1980; Basov and Krasheninnikov, 1983). First finds of Tethyan planktonic foraminifers in the South Atlantic (the Angola Basin) are also confined to the beginning of the Late Cretaceus (the early Cenomanian) (Caron, 1978). These data indicate that restricted connection between the North and South Atlantic appeared somewhere close to the Early/Late Cretaceus boundary.

Stable water exchange between these parts of the Atlantic, however, set in most likely in the second half of the Late Cretaceus and probably even closer to the end of the Late Cretaceus. Planktonic foraminifers assemblage in the Coniacian sediments of the Angola Basin (Caron, 1978) and in the Santonian of the Rio Grande Rise (Premoli Silva and Boersma, 1977) with a large share of Tethyan species testifies to the strong and stable influence of warm Tethyan waters on these areas. This influence increased persistently during the second half of the Late Cretaceus, and that is reflected in the growth of calcareous plankton productivity and in the accumulation of thick sequences of Campanian-Maestrichtian nannofossil-foraminiferal chalk, which is widespread in the South Atlantic.

D.K.Patrunov (1987) who analyzed the continental, marine and oceanic facies distribution in this area came to the conclusion that a destruction of continuous continental crust between the Central and South Atlantic with appearance of depositional environments similar to those of the present-day occurred at the end of the Late Cretaceus.

However, wide deep-water exchange between these parts of the Atlantic seems to have appeared later. When analyzing the facies distribution in the South Atlantic, T.van Andel with co-authors (1977) stated that deep-water exchange initiated here in the Paleocene-Eocene, and not earlier.

Nevertheless, as E.Bonatti and A.Chermak (1981) noted, it still could sometimes be restricted or cease completely because of the uplifts along major fracture zones.

During the last fifteen years intensive studies were conducted in the Equatorial Atlantic. They included deep-sea drilling and geological-geophysical investigations with various research vessels. Among others, the American "Robert Conrad" and Russian "Akademik Nikolaj Strakhov" should be singled out. These studies revealed very complicated block structure of the bottom and showed that intensive vertical tectonic movements occurred here throughout its entire geological history.

These works also provided sufficiently a large collection of rock (and sediment) samples represented by both magmatic and sedimentary varieties (Plate 16, Fig. 1). Along with "normal" samples quite often rocks are present of "anomalous" age and origin, i.e. inconsistent with are those of the surrounding oceanic crust and, accordingly, with the normal spreading model of its formation. These data revived interest in the earlier idea that land bridges connecting South America and Africa existed in the Equatorial Atlantic up to the end of the Cretaceus. These land bridges complicated both bottom and surface circulation in this area and at the same time could serve as a way for land fauna migration between two continents.

In the light of this hypothesis such unusual rocks take on great importance. Their lithology as well as the composition and ecological peculiarities of organic remains contained in some of them provide an opportunity to reconstruct paleooceanographic and paleobathimetric depositional environments and to shed light on some aspects of geological history of the region under consideration.

The purpose of this paper is to summarize all the finds in the Equatorial Atlantic of rocks and sediments with anomalous age or origin, their temporal and spatial distribution here, with consideration of other data testifying to the unusual geological history of at least some tectonic blocks in the equatorial segment of the Mid-Atlantic Ridge.

The rocks with an age anomalous from the viewpoint of their present-day occurrence were obtained from both the western and eastern flanks of the MAR as well as from its axial part. They include both magmatic and sedimentary varieties (Table 1).

The most striking and well-known example in literature of such magmatic rocks is represented by mafics and ultramafics of the St. Peter and Paul Islets with an age of 150-180 Ma (Roden et al., 1984) uplifted above sea surface in the intersection of the fracture zone with the same name and the MAR rift valley. Unusual magmatic rocks were also dredged in the Doldrums Fracture Zone in tectonic block near MAR axis. Fe-Ti gabbro and granites of continental type revealed an absolute age of 150-170 Ma (the Late Jurassic) and even 300 Ma (the Carboniferous) (Kepezhinskas, 1991; Kepezhinskas et al., 1991; Kepezhinskas and Dmitriev, 1992).

The sedimentary rocks with age anomalous for the places of their sampling were obtained from several sites.

In the western part of the Romanche Fracture Zone 90 miles off its intersection with MAR rift valley at the R/V "Akademik Nikolaj Strakhov" station 14 the fragments of volcano-sedimentary breccia were dredged (Udintsev et al., 1990). The carbonate matrix of the breccia contains a diverse assemblage of planktonic foraminifers including Acarinina mckannai, A. nitida, A. pentacamerata, A. broedermanni, Globorotalia formosa gracilis, G. aequa, G. caucasica, G. aragonensis. G. trinidadensis, Turborotalia griffinae, Truncorotaloides rohri, G.triloculinoides. According to Brazilian micropaleontologists who studied this assemblage it attributes the host rock to the uppermost lower Eocene. They also did not find the presence of middle Eocene foraminifers Globorotalia lehneri and late Paleocene Globorotalites compressa in this sample, or the Eocene nannoplanktonic species Discoaster barbadiensis.

The analysis of this foraminiferal list shows, however, that, if species determinations are correct, the assemblage has a mixed composition, Beside the proper early Eocene forms it includes species typical of the middle Eocene (T. rohri, G. lehneri), the early-middle Eocene (G. caucasica, G. aragonensis, A. Pentacamerata, A. broedermani, T. griflinae), the early Eocene (G. Formosa gracilis), the late Paleocene-early Eocene (A. nitida, A. Mckannai, G. aequa), the early-late Paleocene (P. compressa, G. triloculinoides) and even the early Paleocene (G. Trinidadensis). Indeed, such mixed assemblage of species with different ages is quite natural in the breccia matrix that has been formed in the fault zone with sediments non- or semilithified. At any rate, however, the finds of the Paleocene-Eocene species relatively near the rift valley undoubtedly testify to the existence here of blocks with an older crust as compared with that in the neighboring areas.

Fragments of polymictic breccia with a carbon matrix containing relatively old microfossils were also encountered along with tholeithic calcalkaline and alkaline basalt, serpentinized harzburgite and gabbro at st. 54 from the block just near the rift valley northeast of St. Peter and Paul Islets, The breccia matrix incorporates the Cretaceus and Eocene-Oligocene nannofossils.

The nannofossil species of the early Eocene age were also found in the matrix of volcano-sedimentary breccia dredged at st. 70 located south of st. 54 in the same tectonic block.

Sedimentary rocks unusual from the age point of view were obtained from tectonic block on the southern side of the Vema Transform Fault on the 21st leg of the “Robert Conrad” (Bonatti et al., 1983). The rocks dredged here contain definable organic remains permitting age determination and are mostly represented by young shallow-water biomicrites (see below). However, one biomicrite sample (RC-21/04-26) contained rare, poorly preserved older planktonic foraminifers including Chiloguembelina sp. and Globorotalia aff pusilla. The latter species refers the host rock to the upper Paleocene (Globorotalia pusilla and Globorotalia pseudomenardii zones) and testifies to its formation in the open ocean environment of the tropical zone. According to Bonatti et al (1983), this age (approximately 55-58 Ma) is inconsistent with the geographical and structural position of the host block because if the spreading rate for the Paleocene is assumed to be equal to the present-day one, i.e. 1.1 cm/y, the crust age should not exceed here 35 Ma,

In addition to the above mentioned sedimentary rocks with older microfossil assemblages found in situ, several dredge and gravity core stations on several “Akademik Nikolaj Strakhov” cruises provided reworked older planktonic foraminifers in young sediments. Such reworked microfossils were observed at stations 65, 68, 73, 2 and 61, i.e. on both the western and eastern flanks of the Mid-Atlantic Ridge.

At st. 65 located in the rift valley near st. 54 where polymictic breccia with Cretaceus and Eocene-Oligocene nannofossils was obtained the Quaternary foraminiferal-nannofossil ooze contains rare specimens of Paleogene foraminiferal genera Acarinina, Hantkenina and Chiloguembelina (determination by Oskina N.S). The stratigraphic interval of the first genus representatives in other areas spans the upper Paleocene-middle Eocene. The range of genus Hantkenina is limited by the middle-upper Eocene, whereas the first occurrence of Chiloguembelina species is recorded in the lower Paleocene, and their last occurrence is confined to the upper Oligocene. Taken together, the assemblage indicates the presence of at least middle Eocene deposits in this area.

Similar reworked foraminiferal assemblage of Paleogene age also occurs in the Quaternary sediments penetrated by gravity corer at stations 68 and 73 on the MAR western flank farther from its rift valley. And if redeposition by turbidite fows from the North Brazilian Rise can be assumed for the latter station, such a possibility should most likely be precluded for the former one, because it is located on the uplifted tectonic block. Reworked Upper Cretaceus Globotruncana species in the Pliocene sediments dredged on the MAR western flank in the area between the St.Paul and Strakhov fracture zones (St. 2) are most likely of local origin, too.

On the eastern flank of the Mid-Atlantic Ridge, relatively old reworked foraminifers are recorded at St.61 approximately 180 miles off the rift valley. Several large pieces (20x30 cm) of white chalk, that along with the early Miocene forms contained specimens of Chiloguembelina? sp., ,Hantkenina sp. and Pseudohastigerina micra, most probably of Eocene age, were obtained here from the depth of 3640-3820 m.

Thus, these finds of both in situ rocks with age anomalous for the place of present-day occurrence and reworked old microfossils in younger sediments are indicative of the wide distribution here of deposits inconsistent in age with the surrounding oceanic crust. Such crystal blocks with anomalous age occur both near the axial part of the equatorial Mid-Atlantic Ridge and on its western and eastern flanks.

In addition to the above data on the unusual nature of some tectonic blocks in the MAR equatorial part, there is also other evidence that confirm these observations. They include first of all the finds of sedimentary rocks of doubtless shallow-water origin which is indicated by their lithology or by various fossils contained in them and which is inconsistent with geological environments prevailing in the neighboring areas. Rock of this type were obtained by both deep-sea drilling and dredging (Table 2).

DSDP Site 25 drilled at the top of the North Brazilian Rise that is considered to represent western structural and morphological continuation of the Romanche Fracture Zone recovered at the base of sedimentary sequence shallow-water organogenic limestone of Eocene age. Diverse and abundant macro- and microfossils are represented by corals, mollusks, echinoderms, algae and foraminifers.

Fossil composition evidences that during the Eocene the rise top was situated near the sea surface and was occupied by coral reefs. It was then subsided to the recent depth. At the same time Site 354, also drilled near the Brazilian margin on the Seara Rise, penetrated thick (about 800 m) Upper Cretaceus (Campanian-Maestrichtian)-Cenozoic sequence of pelagic carbonate sediments accumulated in the bathyal depth (Bolli, Ryan et al,, 1978).

Dredging provided geographically wider data on the distribution of sedimentary rocks with anomalous origin in the Equatorial Atlantic. Bonatti and Chermak (1981) were the first to discover such kinds of rocks near the intersection of the Romanche Fracture Zone with the Mid-Atlantic Ridge axis. Here, shallow-water limestones of Pliocene age were dredged from the uplifted block in the northern fault slope. The age of these rocks is consistent with the block location near the ridge axis, whereas their lithology and fossils indicate that they were formed near sea surface. Abundant corals testify to sedimentation depths of the first few meters. This means that the top of this tectonic block at that time was near sea level or even above the latter and then it was subsided to the recent depth (about 1 km) with the rate of an order of magnitude higher than that predicted by Sclater's back-tracking curve, The seismic profiling records suggest the presence of similar rocks also in other neighboring blocks where they form buried flat surfaces interpreted as formed as a result of wave erosion near sea surface.

Similar quite unusual sedimentary rocks were obtained from the tectonic block on the southern side of the Vema Transform Fault west of the MAR axis (Bonatti et al., 1983). The top of this block is now situated 3 km above the level predicted by the normal oceanic subsidence model. The bulk of sedimentary rocks dredged at this locality is represented, as was mentioned earlier, by shallow-water biomicrites with abundant fossils: corals, echinoderms, bivalves, red algae (Lithothamnium, Lithophyllum), planktonic and benthic foraminifers. Assemblage of planktonic species consisting of Globigerinoides trilobus, G. ruber, G. Sp., Globigerina bulloides, G. sp., Orbulina universa, Globoquadrina altispira and Sphaeroidinella dehiscens attributes these deposits to the middle part of the Pliocene (the Clloborotalia miocenica Zone).

Both the composition of various macrofossils and large benthic foraminifers (for instance, Amphistegina genus) and lithology unanimously relate deposition of these sediments to the photical zone at a depth of 100-150 m (or less).

Taking into consideration the fact that along with these shallow-water deposits the upper Paleocene rocks of bathyal pelagic origin (see above) were dredged here Bonatti et al.(1983) suggested a scenario of this block geological history. According to their model, at the beginning of the Cenozoic the block was situated in the tropical zone at the mid-upper bathyal depth. Pelagic depositional environments existed here up to the middle Miocene. In the late Miocene through the middle Pliocene the uplifting occurred and the block surface reached the sea level or even crossed it. The shallow-water parts of the block were populated by coral reefs with pertinent fauna and flora. During the late Pliocene-Quaternary it quickly subsided again to its recent depth with a rate of 0.3 mm/y, i.e. an order of magnitude quicker than is predicted by the spreading model.

Though no similar reef deposits were observed in any of the R/V "Akademik Nikolaj Strakhov" legs, dredging at some sites brought in rocks of various types confirming the existence of shallow-water environments in certain periods on some tectonic blocks in the MAR equatorial part. For example, a piece of coal shale with organic matter of land origin was obtained at St. 54 near the rift valley. The fragment of similar coal shale was also found along with magmatic rocks, breccia and chalk at St. 1134 on the MAR western flank north of the St. Paul Fracture Zone. The oldest assemblage of planktonic foraminifers observed in these samples includes Globigerina nepenthes, Globigerinoides bisphericus, G. quadrilobatus, G. diminutus, G. juvenilis and attributes the host rocks to the lower Miocene. Benthic foraminifers are also typical of the Neogene-Quaternary and indicative of bathyal environments. Both the age of sediments and their accumulation environments are consistent with the location of this block relatively far from the MAR axis, At the same time the finds of coal shale seem to suggest shallow-water conditions. It is worth noting that along with angular rock fragments, well rounded pebbles of size 1 to 10 cm were found at this site,

The dredging at St. 1203 in the northern side of the Strakhov Fault provided a large fragment of fresh coal formed, according to V.A.Kotov (Geological Institute of RAS), in swampy environments. Large coal debris was also observed in the Quaternary sediments penetrated by gravity corers at Stations 1101 (interval of 40-45 cm) and 1116 (interval of 0-7 cm) on the western flank of the MAR and at St. 1207 on its eastern flank.

Pebbles of various lithology and roundness class indicative of shallow-water environments with active near-surface hydrodynamics were observed at several sites. In addition to the above-mentioned St. 1134 on the MAR western flank they were also met on its eastern flank near the Romanche Fracture Zone.

Pebbles of limestone were also observed at St, 1324 in the eastern side of rift valley between the Romanche and Chain Transform faults. Among other finds indicative of shallow-water environments which existed on some tectonic blocks in the area under consideration, vesicular basalt should be mentioned. The latter was dredged at St. 48 near rift valley on the ridge eastern flank. Highly vesicular volcanic glass occurred at St. 1322 on the block near the intersection of the rift valley with the Chain Transform Fault.

And, finally, the dredging near the MAR axial zone south to the Strakhov and north of the St. Paul Transform faults at Stations 48 and 75, respectively, yielded along with rocks the debris and entire bivalve shells understood by scientists of the "Akademik Nikolaj Strakhov's" 7th cruise as being of a shallow-water nature.

Conclusions

Geological-geophysical investigations conducted in the Equatorial Atlantic provided strong evidence that this segment of the Atlantic Ocean differed in the mode of lithosphere evolution in comparison with that in other parts of this oceanic basin. The analysis of the materials both on age, origin and distribution of magmatic and sedimentary rocks and on structural patterns in this region allow to draw the following conclusions:

1. There are several types of evidence confirming the existence of tectonic blocks in the equatorial segment of the Mid-Atlantic Ridge, the origin and age of which are inconsistent with the spreading mode of ridge evolution. This is supported by lithological, paleontological, geochronological and geophysical data.

2. The lithological data includes finds, close to the ridge axis, of both sedimentary and magmatic rocks formed at shallow-water or near-surface environments and which subsequently subsided to present-day depths. Such rocks are quite widespread in the Mid-Atlantic Ridge area between the Vema and Romanche Fracture zones. They are represented by reef limestones with diverse benthic rnicro- and macrofossils (algae, corals, mollusks, echinoderms, foraminifers), organic-rich rocks (coal shale, fragments and debris of coal) 'with land plant fossils, large pebbles of various roundness class and composition as well as highly vesicular basalt and volcanic glass. The age of these rocks ranges from the Eocene on the North Brazilian Rise to the Pliocene in the west of the Vema Fracture Zone and in the east of the Romanche Fracture Zone.

3. Strong evidence exists that some tectonic blocks in the MAR equatorial segment are composed of rocks with anomalous age inconsistent with that of the surrounding oceanic crust. This is confirmed by both absolute geochronology (150-170 Ma and 300 Ma in some blocks near Doldrums Fracture Zone) and paleontological datings. Microfossil assemblages (planktonic foraminifers, nannofossils) with anomalous age are present both in situ in pelagic deposits or in carbonate matrix of various breccia and reworked in younger sediments. The geographical and morphological position of sites with reworked Cretaceus and Paleogene microfossils indicates that most likely they are of local origin rather than having been transported by turbidite flows from the ocean marginal parts.

4. Geophysical data shows that some tectonic blocks situated near the intersection of the MAR axial zone with transform faults have flattened top surfaces that were most likely formed near sea level owing to wave erosion. This is supported by the finds of young shallow-water rocks within these blocks. The calculations estimate the subsidence rate for these blocks is an order of magnitude higher than that typical of normal oceanic subsidence.

5. Wide distribution of sedimentary and magmatic rocks with anomalous age and origin in the equatorial segment of the mid-Atlantic Ridge along with geophysical data is indicative of intensive differently directed vertical movements in this region throughout the Cenozoic.

Table 1. Characteristics of rocks with anomalous age

Station	Geographical and structural position	Lithology	Age	Comments
	R/V "Akademik Nikolaj Strakhov"			
23, 48, 64	Doldrums F.Z.	Gabbro, granites of continental type	150- 170 Ma, 300 Ma	
14	MAR western flank in the Romanche F.Z. area	Volcano-sedimentary breccia with carbonate matrix	E. Paleocene- L. Oligocene	Planktonic foramini-fers of mixed compo-sition
54	Intersection of rift valley with the St. Paul F.Z.	Polymictic breccia with carbonate matrix	L. Cretaceous, Eocene-Oligo-cene	
61	MAR eastern flank	Nannofossil-foraminiferal ooze	Eocene	Reworking
65	Rift valley	-"-	-"-	-"-
68	MAR western flank	-"-	-"-	-"-
70.	Intersection of rift valley with the St. Paul F.Z.	Volcano-sedimentary breccia with carbonate matrix{ESC}	E. Eocene	
73	MAR western flank	Nannofossil-foraminiferal ooze	M. Eocene	Reworking
	R/V "Robert Conrad'			
21/04	Southern side of the western Vema F.Z.	Carbonate rocks	L. Paleocene	Middle-upper bathyal environments

Table 2. Characteristics of rocks with anomalous origin

DSDP Site, Station	Geographical station and structural position	Lithology	Age	Comments
Site 25	North Brazilian Rise	Limestone	M. Eocene	Reefal environments
	R/I/ "Akademik Nikolaj Strakhov"			
48	MAR eastern flank	Vesicular basalt; bivalve shells	Pliocene	Shallow-water environments
54	Intersection of MAR rift valley with MAR rift valley	Coaly shale	-"-	-"-
75	Near intersection of MAR rift valley wit h h e St. Paul F.Z.	Bivalve shells		Probably, shallow-water environments
1101	MAR western flank	Coal debris in Quaternary sediments		
1116	MAR western flank	." .	.	
1132	Intersection of MAR rift valley with the Chain F.Z.	Highly vesicular volcanic glass		Shallow-water environments
1134	MAR western flank	Pebbles of variable roundness class; coaly shale		Wave erosion zone
1203	MAR eastern flank	Coal fragments		Swampy environments
1207	MAR eastern flank	Coal debris in Quaternary sediments		
1324	Western side of MAR rift valley	Rock pebbles of variable composition and roundness class		Wave erosion zone
1348	Intersection of MAR rift valley with the Romanche F.Z.	-"-		-"-
	R/V "Robert Conrad"			
Romanche F.Z.	Intersection with MAR rift valley	Limestone	Pliocene	Reefal environments
21/04	Western Vema F.Z.	-"-	." .	." .

Chapter 16

Finds of Rocks of Continental type and Sediments of Anomalous Age on the Equatorial Segment of the Mid-Atlantic Ridge (MAR).

G.B.Udintsev, N.A.Kurentsova, N.V.Pronina, S.B.Smirnova, M.G.Ushakova

During the 7th expedition of the “Akademik Nikolaj Strakhov” (1988,) the investigations focused on the equatorial segment (ES) of the Mid-Atlantic Ridge (MAR.), During the dredging of native rocks from the bottom in places of the exposed basement, apart from the usual magmatic rocks (toleitic basalts et al.) samples of rocks of continental type and sediments of an older age were obtained.

On station 43 (2°15.5’N; 30°25.2’W) from a steep west slope of trench with a depth of 2100-2900 m, about 150 kg of stone material (Plate 10,16,) was lifted by dredging. This trench lies 12 miles (20 km) to the east of the axis seismic and volcanic active rift valley and corresponds to the marginal zone of intensive tectonic dislocations on the contact point of the axial rift zone of the ridge and its gently sloping eastern flank plateau.

Among the samples obtained there are fragments of full crystallizated ferriferous basalts (60%) of polymict carbonaceous volcanic-sedimentary breccia (30%), serpentiuous peridotites of dunite-harzburgite series and gabbro (10%) and also a single fragment of coal schist.

The age composition of the microremains of the sedimentary breccia matrix is unusual. According to paleo-reconstructions based on the concept of plate-tectonics, the age of sediments in the place of this dredging must not be more ancient than Miocene-Pliocene. Meanwhile, in the composition of lime nannoplanctone there are not only modem forms such as (Cyclococcolithus leptoporus (Murray, Blackman), Helicosphaera kamptheri (May, Mohler), H. inversa (Garther), Rhabdosphaera claviger (Murray, Blackman)) and late Pliocene (Discoaster brouweri (Tan)), but in the sedimentary matrix there also are Eocene-Oligocene forms such as (Discoaster adamanteus (Bramlette, n), D.tanti (Bramlette, Ridel, Wilcoxon), D.cf robustus (Hag), Chiphragmalithus cf.calathus (Bramlette,Sullivan), Ch.of.barbathus (Perch-Nielsen)) and even late Cretaceus (Maastrixt) - Archangelskiella cymbiformis (Vekshina), Prediscosphaera cretacea (Archangelsky garther), Lucianorhab dageuxi (Deflandre).

Side by side with modem forms of foraminiferas there are also Pliocene forms such as (Glodorotalia miocenica (Palmer), Globogerinoides fistulosus (Schubert) and others, according to the analyses of Tchekhovskaja..

Judging by the spectrum of lime nannoplancton, it can be supposed that sedimentation in this part of the Equatorial segment of MAR began in late Cretaceous-Eocene, and was interrupted in the interval between Eocene and Pliocene. The breccia probably originated in Pliocene. It is now covered with a thin cover of recent sediments. There has been earlier information [e.g. Barash, Lavrov, 1977; Cifelli et al., 1968; Bonatti, Honnorez, 1971; Bonatti, Chermak, 1981; Bonatti, Crane, 1982] about finds of relatively ancient (Paleocene-Eocene) sedimentary deposits in the equatorial part of MAR. On the basis of such findings several hypotheses were proposed as alternatives to the hypothesis about the spreading nature of the Ridge. These were suppositions about the non-spreading origin of the Ridge as a whole [Barash, Lavrov, 1977] or of its segment only [Bonatti, Honnorez, 1971], or about repeated rejumps of axis of spreading and transform faults [Bonatti, Crane, 1982].

Sediments of Cretaceous-Eocene age which we discovered by us on station 43 led us to assume extremely low scales of spreading at ES MAR(Udintsev, Kurentsova, Pronina et al,, 1990). These sediments lie at 20 km to the east of the active axial rift valley in the zone of intensive tectonic dislocations, on the contact point of the rift system with the flank plateau, whose nature, according to geomorphological and geophysical data, is different from riftogenous, and rather close to platform type. Sedimentation occured there on the basement of Cretaceous-Eocene age, which since that time had not undergone any large horizontal movements to the east of the active rift valley, but was broken by tectonic forces and experienced differential vertical movements (perhaps in conditions of compressional collision), which causedan interruption in sedimentation after Eocene up to late Pliocene, and the origin of the polymict breccia, covered by a thin layer of modern sediments.

The composition of the breccia includes unrounded sharply angular fragments of siliceous schist and porhyraceous andesito-basalts with the character of amygdaloidal structure, which corresponds to spilito-ceratofirous formation and have experienced regional metamorphism. This reeks are analogous to the volcano-sedimentary basement for the elementary stage of development of euglosynclinic belts

[Lutz, 1980]. Judging by the fragments of siliceous radiolaries (Pseudodictyomitra sp., Thanaria lacrimula gr., Stichocapsa sp.),the age of siliceous schists is Early-Middle Cretaceus.

During the dredging on station 43 one unrounded fragment of coalschist (argillite), perhaps from sedimentary breccia, was found, The coal schists in the Equatorial Atlantic were earlier found at the Sierra Leone rise [Gevorkan et all, 1988), but in the central part of the equatorial segment coal schist was found for the first time. Ash content in this sample is 40-50%, Ash substance is thinly laminated with small grained quartz sand with clay cement. The grains of quartz are badly rounded. The coal substance is humus, represented by components of vitrinite (collinite) and inertite (fusinite with remains of the primary structure of wood) groups. Transformation of humus substance into coal is not higher than that of the stage of brown coal corresponding to the early stage of regional metamorphism. L.E.Shterenberg (1990) attribute this coal schist to meagre coals (according to data of chemical and mineralogical investigations), i.e. to a higher stage of metamorphism.

Coal schist was formed in conditions usual for humic coals: in peat bog, drained periodically and washed by sandy-clay flows. In damp anaerobical conditions the organic substance was reformed in vitrinite and in drier oxidized conditions - in fusinite, consisting of half of the coal mass.

A piece of coal schist was macerated to pick out spores and pollen. Palynological analyses of the maceration results established the presence in the in sample of very small quantity of spores and pollen. In 30 preparations 25 specimens of miospores were discovered, Among them were: spores sphagnum mosses (Sphagnum sp.) - 1 specimen; plauns (Lycopodium sp.) - 3 sp,; ferns (Concavisportites sp.) - 3 sp.; indefinite spores -3 sp. The pollen of Gareseed is represented by pine (Pinus sp.) - 3 sp. and heirolepidiev (Classopollis sp.) -1 sp. Very dark ekzina of spores seed Classopollis points to the idea of its redeposition here. The rest of the pollen belongs to Coveredseed plants: birch (Betula plicoides Zakl) -2 sp.; alder (Alnus sp.) - 1 sp.; elm (Umoideipites planeraeformis Anderson) -2 sp.; nut-grove (Corylus sp.) -1 sp.; ash-tree (Fraxinus sp.) -1 sp.; (Rhus sp.) -1 sp.; indefinite pollen of coverseed - 3 sp. There are primarily elements of moderate-climate flora. Only Rhus is characteristic of subtropical vegetation, Such a small quantity of spores and pollen does not allow for the determination with sufficient accuracy of the age of the deposits contained. The most probable age is the end of Paleogene, because of the absence of the pollen of Normapolles, which is characteristic of upper cretaceous-paleocene.

The very dark colour of pieces of vegetable tissue in the maceration, and the normal light-yellow color in pollen and spores as well as the small quantity of the latter allow for the assumption that a process of erosion had taken place in the greater part of the rocks.

It is imposible to explain the find of coal schist on station 43 station as a result of transportation by floating ice (ice does not reach this area), or by coal thrown off ships (brown coal is not usable in steamship boilers).

The fragments of volcano-sedimentary breccia had also been dredged on 54 station (2°05.3’N; 30°18.4'W), on the east margin of thee rift zone. Here in the composition of the matrix lime nannoplancton of the early Eocene age (Stephanolithus anakrhopus Bukry Percilab, St. Primus Perch-Nielsen, St. radians Deflandek) was again discovered.

About 20 kg of brecciated material were dredged on station 10(1°22.8’N, 32°28.1’W) from the steep southern slope of the Saint Paul fracture zone, at a depth of 3290-3500 m. Numerous fragments of metamorphic sedimentary rocks: sericitic siliceous phyllites, sometimes with streaks of graphite, and microquartzites enter into the composition of breccia. There are products of regional metamorphism of quartz-feldspar sandstone and clays with admixtures of coaly organic matter. On the continents phyllites are found in the foundation of folded geosyncline formations [Lutz, 1980]

The Eocene age of breccia matrix base been defined by lime nannoplancton and by the plancton foraminifers. The Paleocene species Planorotalites compressa is the most ancient. Forms of the end of early Eocene are predominant (Acarinina Mckaanat; A. Pentacamerata; A. brocdermanni; A. nitida; Morozovella trinidadiensis; M. Acqua; M. caucasica; M. formosa gracilis; M. aragonensis; Turborotalia griffinae; Trucorotaloides Rohri; Globigerina triloculinoides), the forms of middle Eocene are rare (Morozovella lehneri). From ancient coccoliths eocene form was met (Discoaster Garbadiensis Tan Sihnon). The shells of genus Globigerinoides are absent in complex, but many shells from genuses: Globigerina, Acarinia, Globorotalia [N. S. Oskina]. From ancient coccoliths eocene specie Discoaster Garbadiensis Tan was met.

Given the distance between the dredging site on the 10th station from the active rift valley, the age of this sediments does not exceed that required by the hypothesis of the spreading origin of the whole ridge, including the flank plateau, The presence in the breccia of coal phyllites, however, allows for the assumption of the nonspreading nature of the basement of the west flank plateau , and for considering it as a relict of continental structure.

The finds of rocks of continental type, and sediments of a more ancient age than was posited using the canons of the plate-tectonic theory, serve as an important addition to geophysical data, allowing for the supposition of the nonspreading origin of flank plateaus. We suppose that to the end of Cretaceus the Equatorial Segment represented a wide platform, on which were positioned some islands, shoals, and shallow water basins, in which, probably, coal schists were formed, At the end of the Cretaceous-beginning of Paleocene the platform regime had been replaced in the axial zone of the segment by riftogenesis, as a result of the prograding of the rift, which connected riftogenals of the North and South Atlantic. The unity of the structure barrier had been disturbed, and ES experienced progressive submergence to modern depths. Some features of the platform structure at the basement of the flank plateaus are still exist, however, since from Cretaceus until now spreading has been localized solely to the axial zone of the segment.

Chapter 17

Phyllites of the Equatorial Segment of the Mid-Atlantic Ridge

A.N.Fenogenov

About 20 kg of breccia material was dredged on the station 10 (1°22,8'S; 32°28.1'W) at a depth of 3290-3500 m on the southern slope of the transversive ridge of the St. Paul fracture zone (Plate 2, 3, 16) on the west flank of the Mid-Atlantic Ridge, on the eastern edge of the Guyana basin. Unrounded fragments of metamorphic sedimentary rock of sericitic phyllites make up this breccia. The Eocene Matrix of breccia Udintsev et al., 1990) is represented by carbonate coccolith-foraminifcral silt.

These phyllites are characterized by 15 thin sections and they are represented everywhere by sericitic species, in which admixtures of dusty coal material with transition to coal phyllites often play an essential part. There are large admixtures of quartz material - from separate clastogene grains to a few lenses and streaks to microquartzites.

Structures of monomineral scricitic phyllites are microlepidoblastic. with growths of an admixture of quartz and plagioclase distinctly exchanged by microgranolepidoblastic. As a rule, they vary from relict aleuro-pilitic to psammitic structures (Plate 20, fig, 1-8).

Textures of phyllites are stably slaty, but are sometimes complicated by striation (by alteration of stripes of almost monomineral sericitic composition with a few admixtures of 0.02-0.04 mm rounded grains of quartz with stripes, enriched fragmental material, represented by quartz and plagioclase rounded and comer-rounded forms of the same sizes, these and in sericitic stripes, and larger - to 0.15 mm, cemented by chlorite-sericitic material); by presence near short lenses (5- 1 () mm in length) of microquartzites or simple quartz compositions (Plate 20, fig, 4, 5)). Thin sections are noted from microflexure folds to microstriated (Plate 20, fig. 1, 2).

The phyllites in many slides are crossed by quartz and quartz- albite vein accompaniments, often containing different quantity admixtures of thin-scale chlorite-pennine; sometimes quartz is very strong. Sometimes chlorite forms independent rnonomineral very thin (0.03-0.06 mm thickness) veins, crossing the foliation at an acute angle (Plate 20, fig. 8).

The phyllites are subjected to brecciation with the formation of microbreccias from little (0.1-1.0 mm) angular fragments of phyllites, and are differently oriented, cemented by microgranular mass chlorite-albite-quartz composition, Sometimes brecciation is marked on near (to 2.5 mm wide) zones, at an acute angle crossing the foliation. The orientation of these zones is the same as with crossing the foliation of microgranular quartz.

Side by side with brecciation there is also mlcrogranulation, visible in quartzite streaks. Grains of quartz are recrystallized, converting into an extremely thin microgranular aggregate (Plate 20, fig. 4). Hypergenetic changes of phyllites are shown in development of oxides' iron hydroxides, coloring rocks in brown tones.

In their mineral composition the phyllites of the region are very monotonous. Sericite is the leading mineral, but in individual lenses and streaks - quartz in transition to microquartzites, containing thin streaks of sericite in small quantities,

There are in different quantities plagioclases, chlorite, leicocsen, coal material, carbonate, and ore minerals; from accessory minerals there are sphene, apatite, zircon; from hypergenetic - iron hydroxides.

The sericite is in thin scales 0.01-0.1 mm in size, possessing as a rule bare-axis orientation, and emphasized slaty and striate textures. Besides sericite, composed of phyllites formed from clay material of initial sedimentary rocks, there is also sericite, developing on plagioclase.

Quartz is represented by clastogene grains and hydrothermal formations in the form of veins, marked on rocks which are already quartz. Clastogene quartz has rounded and comer-rounded grains 0.01-02 mm in size, either uniformly diffused in the rock, or collected in monomineral lens and striped aggregates, m which quartz often possesses sharply pronounced wavy extinction, pointing to the development of cataclasis, up to full granulation of the primary fragmental product, In the hydrothermal veins crossing the phyllites quartz sometimes is the only mineral, but more often it is associated in them with albite, or with albite and chlorite, or only with chlorite.

Plagioclase, just like quartz, is present in the form of fragmental grains and in the veins crossing the phyllites. In the clastogene plagioclase there is usually no twinning, but there is seritization. Plagioclase is xenomorphic, and the sizes are commensurable with quartz and reach 0.5 mm. The albite of veins is clean, without signs of seritization.

Chlorite is found only in quartz and quartz-albite veins, in the regions of quartz reef and in the form of monomineral veins crossing phyllites. The chlorite is pennine, thin-scale, light-green in color with pleochroism going to colorless, with the characteristic anomalous purple interference of hues. The scales of chlorite are either unevenly diffused, or collected in spherolite or worm-like inclusions in quartz. The chlorite veins cross the stripes of coal-sericitic composition, sometimes separating them into a series of short blocks. In all cases, the imposition of chloritization on cericitic phyllites is visible, already with a slaty texture.

The leucoxene is not fixed in all thin sections, but when it is present in quantity, it is collected in stream-like, thin accumulations, translucent in transitional light and partially changed in xenomorphic sphene. The latter is also found without leucoxene, but also in the form of xenomorphic grains, size to 0.015-0.025 mm.

Carbonate was found only in one thin section,in single xenomorphic grains of a size of up to 0.1 mm in cericitic mass.

Ore minerals in the phyllites are rare and are found in one thin section; it is xenomorphic untransparent grains, size in first hundredth parts of millimeter and to 0.1 mm.

Apatite was observed in one thin section in the form of prismatic crystals, size to 0,05 mm, related to quartz-plagioclase particles applied to the main material. The prismatic form of apatite indicates its origin together with quartz and albite veins, and non-detrital genesis.

Zircon is also very rare; it appeared in one thin section in the form of single grains, well rounded, size to 0.035 mm.

The phyllites described are among typical pararocks, which is obvious from:

1) relict detriral structures;
2) slaty and lenticularly slaty (originally-thin layered and lenticularly-layered) textures;
3) presence of well rounded zircon, corresponding in size to clayish fraction (aleurolitic).

These rocks experienced two stages of metamorphism:

1) regional metamorphism of phyllitic facies and
2) hydrothermal- imposition of quartzitization, albitization and chloritization.

Chapter 18

Plutonic and Metamorphic Rocks in the Crestal Zone of the Mid-Atlantic Ridge: on the Problem of Composition and age of the Lower Horizons of the Oceanic Crust

A.I.Trukhalev, Yu.E.Pogrebitsky, B.V.Belyatsky, N.A.Kurentsova, I.M.Vasilieva.

The present work concerns the results of the study of basic and ultrabasic plutonic and metamorphic rocks, dredged in the crestal zone of the Mid-Atlantic Ridge (MAR) in the course of the regional geologic-geophysical survey on the Angola-Brazil Geotraverse (ABGT) and on the Canary-Bahamas Geotraverse /Trukhalev, Silantjev, Kurentsova et al., 1990/, and the findings of exploration for abyssal complex sulfide, carried out by "SEVMORGEOLOGIA" Association. The study of the rocks is of special interest since it provides a means for obtaining firm data on the composition and age of lower horizons of the oceanic crust which underlie oceanic basalts. It is worth noting that in spite of many publications, dealing with abyssal rocks of the MAR /Kovalenko, Ionov, Jarmoljuk et al., 1990, Lazko, 1987, Pljuskina, 1983, For, 1989, among others/, they usually do not deal with the problem of their age. It goes without saying that they are young reeks, coeval with the basalts of the MAR rift zone. However, this opinion seems to be far from being unquestioned. Sharp differences in metamorphic grade and tectonic disturbance between the basalts and the plutonic rocks, and the usual occurrence of the latter as blocks-protrusions or tectonic breccias would rather suggest that they are genetically unrelated rocks, differing in age.

First datings of plutonic rocks supported the assumption to a degree, leaving some doubts in the reliability of ancient age values obtained(about 1.9 Ga for metagabbro from a site at 145 0S, and 0.55 and 1.63 Ga for metagabbro and migmatite-granite from the area of 265 ON /14, 25/. Further studies of the rocks in question provided new data, supporting, in the author's view, the primary age estimates and substantially widening existing notions of the time of formation of abyssal rocks of the MAR.

On the ABGT, the plutonic and metamorphic rocks were dredged in the course of a cruise of R/V "Professor Kurentsov" during 1986-1987 at localities adjacent to the transform faults, at 11 5 OS (Stations 8726-4 and 8728-2) and at 145 0S (Stations 8717 through 8720, Plate 21, Fig. 1,) in the area of block uplifts in the MAR rift zone. Petrographic and geochemical examination of the rocks offered a means for dividing them into the following groups, differing in genesis and, probably, in age:

1. Coarse- to gigantic-grained amphibolized and tectonized metagabbro, titanomagnetite, metagabbro, metagabbro-norite, serpentinized spinel- and phlogopite-bearing peridotites. In petrographic-geochemical features, rock and mineral composition (Plate 21, Figs 2,3,4), they can be correlated with the reeks of the layered intrusions of the Bushveld, Stillwater and other types.
2. Products of multistage metamorphism of gabbroids, generally represented by plutonic rocks, partly recrystallized and replaced by secondary minerals, as well as gabbro-amphibolites (sample 87 19/23) and gabbro-granulates (sample 8719/19). In texture, association and composition of minerals, they represent rocks of granulite (T=740-940 5 0C, P=5 kbar,), epidote-amphibolite (T=530 5 0C, P less than 2 kbar, Plate 21, Fig. 5), and greenschist facies.
3. Metadolerites and metabasalts, representing products of greenschist etamorphism of volcanic and hypabyssal tholeitic rocks and probably belonging to the complex of rocks, enclosing metagabbroids and hyperbaric rocks.
4. Serpentinites and talc-serpentine schists, probable products of metasomatic reworking of gabbroids. They are characterized by very high values of the indicator (Nb/Zr)4 N0 ratio (20.4-27.7), sharply differing from that in metagabbroids and peridotites (0.49-0.84 or 1.40-2.53 in the reeks, dredged at Station 8728-2), and suggesting their relation to some other, more deep-seated magma source.

Of the above four rock complexes, isotopic analyses were made of the rocks of the first three complexes. The oldest age values (about 2.5-2,6 Ga) were obtained with the aid of Pb-Pb thermal ion technique for zircons from metadolerites. The zircons drastically differ in size, color, morphology and isotopic composition (they contain practically no common lead) from those, extracted from metagabbro. The Sm-Nd model age of one of metadolerite samples (3.74 Ga) far exceeds those for metagabbroids (2.5-2.7 Ga). These peculiarities suggest that the metadolerites are independent igneous rocks, probably belonging to the rock complex, enclosing metagabbroids and peridotites. The metabasalts, dredged at Station 8726-4, probably rank among the same complex. A very young K-Ar date of the reeks (less than 1 Ma /14/) probably shows the time of the last reworking of ancient rocks in the zone of recent rifting.

A Pb-Pb date of zircons of magmatogenic habit, extracted from metagabbro (sample 8719/2) is the most realistic age of metagabbroids (about 1.9-2.0 Ga). It is somewhat supported by results of

Sm-Nd isotope analysis. Points plotted on an isotope diagram (Plate 21, Fig, 6), showing compositions of bulk samples of metagabbro, recovered at Stations 8719 and 8728-2, as well as metagabbro from the Canary-Bahama Geotraverse (sample 105-4, see below), lie on a straight line which is probably errorchron; its slope corresponds to an age of 2322+-560 Ma. An assumption of approximately coeval gabbroids, dredged at stations far removed from each other, is based on general petrographic-geochemical features of the rocks and their association with a single lithospheric block, characterized by a common evolutionary trend of tectonomagmatic processes, The assumption is confirmed by close dates of zircons, extracted from samples 8719/2 and 105-4, approximately coinciding with their errorchron age.

Sm-Nd model ages of metagabbroids (see Table 5) prove to be somewhat older as compared to zircon dates. It is as yet impossible to determine with some certainty the cause of the older ages. Judging from 7 0Nd(T) valuesfrom the rocks (+ 1.5- +2.9), which are lower than 0Nd of the depleted mantle for the assumed time of their crystallization, the older ages may be caused by contamination of the gabbroid magma with material of the ancient continental crust.

In the authors' view, Pb-Pb dates of other morphotypes of zircons, extracted from sample 8719/2 (see Table 2), the internal Sm-Nd isochron for the rock (on amphiboles and bulk sample, Fig.6), as well as K-Ar age determinations of metagabbro and metadolerites (220-336 Ma for sample 8719/2; 55-57 Ma for sample 8728-2/1; 125 Ma for sample 8719/5, see Fonarev, Grafchikov, 1987) reflect the polymetamorphic nature of the rocks and the repeated action on them of various superposed processes. Most of these dates and determinations roughly coincide either with epochs of activation (isotope rejuvenation), widely displayed in Precambrian rocks of adjacent continental areas (Pan-African Epoch, Brazilian cycle of metamorphism, 500-650 Ma), or with outbursts of basic magmatism (Cretaceus /Ozima Sasito, Matsuda et al,, 1976/, Triassic-Jurassic) over the coasts of Africa and South America.

As for the internal Sm-Nd isochrons for samples 8719/2 and 8728-2/1, constructed for plagioclase, whole rock and pyroxene, they are probably devoid of geologic sense. This is especially evident for sample 8728-2/1, whose internal isochron shows a negative age.

On the Canary-Bahama Geotraverse, the authors studied the rocks, dredged in the course of Legs I and V of R/V Geolog Fersman (1986, 1988) in the area of 26 5 0N (Stations 67 and 74), 24 5 0N (Stations 130, 162, 192), and 23 5 0N (Station 105) - (Plate 21, Figures 4,5.).

Analysis of our own material and publications concerning abyssal rocks of the region shows that fragments of layered hyperbaric-basic complexes, usually tectonized to a greater or lesser degree and bearing evidence of multistage metamorphism, are exposed there in tectonic blocks,

According to /Pljuskina, 1983/, plutonic rocks, exposed in the area of 23 5 0N and southward, are composed of coarse-grained tectonized gabbroids, diabases,and amphibole gabbros, containing a wealth of Fe-Ti oxides and peridotitic tectonites, In petrography and geochemistry, these rocks, building up the western slope of the median valley over a length of 30 km, are sufficiently similar to plutonic rocks, dredged in the area of 14 5 0S. This is supported by isotope studies as well. A Pb-Pb date on zircons, extracted from pegmatoid metagabbro, dredged at Station 105 in the northern part of abyssal rock exposures, studied by J. Karson (about 2 Ga for sample 105-4), and an ancient Sm-Nd model age of the rock (about 2.4 Ga) probably point to a Proterozoic age for the rock. As noted above, the errorchronous relations of the rock with the metagabbroids, dredged on the ABGT (samples 8719/2 and 8728-2/1), and a sufficiently ancient (about 2.3 Ga) Sm-Nd errorchron age of the rocks can be inferred,

An origin from a source, corresponding in composition to a depleted mantle, containing a tangible proportion of continental crustal material, is more confidently inferred for the rock in question than for other samples studied, The origin from a source of this kind is demonstrated, along with the ancient model age of the rock which markedly exceeds its inferred crystallization age (about 2 Ga), and the value of 74 Nd0(T)=(+- 1.5), which is lower than for the depleted mantle of that time, by the unusual radiogenic isotope composition of lead in plagioclase from sample 105-4, close to that in the cratonized crust (5206 0Pb 5207 0Pb= 19. 120; 52070Pb/ 5204 0Pb= 15.816; 5208 0Pb/5204 0Pb=40.820).

The plutonic rocks, dredged in the area of 24 5 ON, are represented by serpentinized peridotites, metatroctolites and metagabbroids (Plate 21,Fig.7,). The least altered samples of metagabbroids were taken to be studied isotonically. Fine-grained gabbro-granulates (sample 162-5), probably representing metamorphosed basic dikes in gabbroids and peridotites, differ markedly from other rocks in structure and geochemistry. A geochemical “alienation” of the plutonic rocks from riftogenic basalts, dredged in the same area, should be singled out. This is most evident in different values of the (Nb/Zr) 4 N 0ratio, peculiar of asalts (0.46-0.52) and plutonic rocks (5.46-19.1), suggesting their relation with different magma sources.

Metamorphic granoblastic textures, displayed by the most recrystallized metagabbroid varieties, the composition of incorporated metamorphic pyroxenes (Plate 2l,Fig. 8) and amphiboles suggest that the rocks underwent multistage metamorphism at granulite (T=840 5 0C, P=5 kbar for samples 162-5, 192-21), epidote-amphibolite (T=420-540 5 0C, P=2-4 kbar for samples 162-5, 67-3), and greenschist facies.

Results of Sm-Nd isotope analysis suggest a Proterozoic age for the plutonic rocks. The model ages for all the samples under study are sufficiently ancient, Precambrian; they form two groups, falling into the age ranges of 1600-1700 Ma (samples 162-5 and 192-21) and 600-1100 Ma (samples 130-14, 130-17 and sample 67-3, taken in the area of 265 ON). Plotted on an isotope diagram (see Plate 21,Fig. 6), points, showing the composition of the rocks of these groups, lie separately, suggesting the errorchronous relations between them. The position of the points relative to 2.0 and 0.6 Ga isochrons, plotted for comparison, indicates that an errorchron age for Group I samples should be somewhat less than 2 Ga and should roughly correspond to their model age. Group II samples (especially Nos 130-17 and 67-3) obviously tend to the 0.6 Ga isochron; their probable errorchron age is about 600 Ma. The assumption is confirmed to some extent by K-Ar dates of one of Group II samples (5 17-562 Ma for sample 67-3 /Siedner, Miller, 1068).

Internal Sm-Nd isochrons for the rocks in question, representing, as a rule, sufficiently young ages (0-64 Ma, Figures 6B and 6C show the most characteristic isochrons) probably reflect recurrent actions on them of different superposed processes (mantle diapirism, metasomatism, etc.). The exception is gabbro-granulite sample No. 162-5; no internal isochron, common to the rocks under study, have been obtained for it; the minerals and whole rock allow two isochrons (?), corresponding to 1574 t-31 Ma and 250+-64 Ma ages, to be delineated (see Plate 21,Fig.6D), The first value proves to be sufficiently close to the Sm-Nd model age of the rocks and probably shows an approximate time of its formation. The second date probably fixes one of the episodes of thermotectonic transformations of ancient rocks.

Isotope data obtained indicate that hyperbaric-basic complexes, differing in age, are exposed in the area of 24 5 0N. More ancient complexes (about 1600-1700 Ma), represented by the rocks, dredged at Stations 162 and 192, make up tectonic blocks, occurring on the lower slope of the rift valley; younger complexes (1100-600 Ma), forming a separate block (see Fig. 7), were taken at Stations 130 and 171 on the upper slope.

Characteristics of the metagabbro and migmatite-granite, dredged in the area of 26я5oя0N, and the preliminary results of age determinations were published The data were confirmed by further studies. Komarov A.N. made fission-track dating of zircons from metagabbro (sample 67-3), indicating an ancient age for the rock, 280+-70 Ma or probably higher because natural annealing of spontaneous tracks was not allowed for in calculations, Data on Sm and Nd isotopy also point to a sufficiently ancient (about 600 Ma) age for the rock.

From the outset, age determination of the migmatite-granite appears to be sufficiently reliable since K-Ar datings on biotite rarely provide overestimation. A local origin of this typical continental rock uncommon with the mid-oceanic ridge, would only be dubious. However, isotopic evidence (sample 105-14) for the presence of continental crust relics in the area (235 ON) has been obtained by now. Hence, the find of migmatite-granite here appears to be quite natural.

It should be noted in this connection that metamorphogenic granitoids, granulite-charnockites and other continental rocks were also dredged in other areas of the Atlantic (Atlantis Fault Zone (about 30я5oя0N); Azores-Gibraltar Bar; Reykjanes Ridge and others /Roden, Hart, Frey, Melson, 1984/). The studied migmatite-granite is very similar in structure, composition and age to metamorphogenic granites of the Reykjanes Ridge; like the latter, it may have been formed in granitization of Precambrian metabasic rocks (orthoamphiboles and others /Reynolds, Clay, 1977/). This suggests the presence of ancient (Lower Proterozoic and, possibly, Archean) metabasic rocks within the MAR. The suggestion is directly supported by the find of 2.5-2.6 Ga old metadolerites on the ABGT. Ancient metabasic rock exposures next to the site of rnigmatite-granite dredging are believed to be located south of the Kane Fracture. The age of the metagabbro dredged here (about 2 Ga for sample 105-4) suggests an Early Proterozoic age for the associated gabbro-granulites, diabases and greenschists.

The above-cited age dates or, rather, estimates of the abyssal rocks of the MAR are tentative and approximate, and hence they need refinement. Even now, however, there are grounds for believing that they have a geological meaning and reflect actual geological processes. First of all, this is suggested by close dates, obtained by different isotope techniques. For instance, Pb-Pb dates of zircons from the metagabbro (samples 8719/2 and 105-4) roughly coincide with the Sm-Nd errorchron age of the rocks. It is also known from the practice of geochronological studies that the probability of obtaining overestimated Pb-Pb dates (especially for zircons having low common lead) is insignificant . Therefore the ancient age values, obtained by this technique, can be regarded with confidence as the minimal time for formation of the rocks under study. The fact that the ancient age values for the abyssal rocks of the MAR are realistic is indirectly supported by the coincidence of most of them with peaks of planetary endogenic activity (2.6; 1.6; 0.6 Ga and others), ascertained on the basis of many precise isochron dates for igneous, metamorphic, and volcano sedimentary rocks of the Earth /Tscheka, 1977/.

As for young age values, obtained from Sm-Nd mineral isochrons and K-Ar determinations, they are believed to show the time of the most intense reworking of the abyssal rocks by heat and fluid flows and melts, ascending in the rift zone. A similar pattern was observed for peridotites, incorporated as xenoliths in Cenozoic volcanites of the Central Asiatic region /Peive, 1975, Pushkarev, 1990/. Sm-Nd mineral isochrons for the rocks of xenoliths represent very young ages (2-16 to 30-48 Ma), close to the age of enclosing volcanites, while the Sm-Nd whole rock isochron (2Ga) and ancient Rb-Sr and Sm-Nd model ages (1.7-2.3 Ga) of most xenoliths suggest a Proterozoic age for them. It is the opinion of the above-mentioned authors that the Cenozoic mineral isochrons fix the time of reworking of the lithospheric mantle in the process of recent activation of the Central Asiatic region.

The data presented above and earlier well known publications point to the relatively wide distribution of metamorphosed hyperbaric and basic rocks at the MAR. The ancient isotope dates of the rocks (about 4.5 Ga for ultrabasic rocks from St.Paul Island /Hebert, 1982, Roden, Hart, Frey, Melson, 1984/; 635 Ma for the metagabbro from Borehole 334 /Bonatti, Honnorez, 1970/; 169 Ma for the metabasalt from the area of 30 5 0N ; our data), as well currently available and permanently incoming information, suggesting their geochemical and isotopic similarity with continental rocks of the same type /Angola-Brazil geotravers... 1989, Pogrebitskij et al., 1990, Roden, Hart, Frey, Melson, 1984, Stedner, Miller, 1968, Prinz et al., 1976/, suggest that they are probably relicts of pre-oceanic substratum, reworked in the process of oceanic rifting and volcanicity. It is also evident that the substratum is composed of geologically heterogeneous rocks which differ in age.

The data currently available allow for the recognition (or inferral) of the following rock complexes in the substratum

- Archean ultrabasic rocks, representing ancient mantle blocks or, according to Peive, relicts of the primary serpentinite-ukrabasic solid earth;
- Lower Proterozoic (2.5-2.6 Ga) and, possibly, Archean volcanic and Hypabyssal basic rocks, metamorphosed in the epidote-amphibolite and greenschist facies;
- layered hyperbaric-basic plutonic rocks, differing in age (1.9-2.2; 1.6-1.7; 1.1-0.6 Ga);
- Lower Proterozoic (1.6-1.7 Ga) metamorphogenic granites, formed in the process of granitization of ancient basic rocks;
- Phanerozoic metasomatic and igneous rocks, formed in the epoch of planetary endogenic activity.

In addition, intrusive basic rocks, comagmatic with the oceanic basalt complex, should also occur. The oceanic basalt sequence may have formed on the ancient Precambrian substratum. By now the substratum has been transformed, “removed part by part”, and incorporated into the mantle, expanded under the ocean, from where it has been squeezed out as blocks-protrusions or tectonic breccias on the sea floor surface.

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Table 1

Age of Zircon from metagabbro (method of thermo-emission)

N samples	Isotopic relations, Pb			Age, mln years
	208/206	207/206	204/206	
8719/2, metagabbro				
Colourless and pale-orange idiomorphic crustalls of Zr (60x70 mkm) magmatic habitus (combination of prisms with dipiramid, K = 2,0)	0,0000±0,0214	0,1137±0,0020	0,0027±0,0017	1859±292
Ohric more large Zircon (200-80 mkm, K = 0,3-0,4)	0,0029±0,0184	0,0872±0,0233	0,0017±0,0009	1367±441
Colourless isometric Zircon (200-80 mkm)	0,0000±0,0388	0,0616±0,0451	0,0033±0,0005	664±108
Pink isometric Zircon (200-80 mkm)	0,0000±0,0093	0,0572±0,0043	0,0016±0,0003	500±156
8719/8, metadolerit				
Small (0.02 mm and smaller) light-brown rounded, rarely short-prismatic crustals; habitus unclear. Homogenous with normal be-refraction, sometimes with fragment of zonation.	0,000000±0,003208	0,163161±0,00354	0,000605	2488±37
Small (0,07 mm and smaller) pink short-prismatic crustalls with lengthed dipiramidas side: habitus unclear. Homogenous non-zonal with normal bi refraction, numerous inclusion of gase,gase-liquid and dark oore.	0,102157±0,0034	0,172393±0,00912	0,000007	2580±85
105-4, metagabbro				
Average sample of Zircon of different colour and size (60-150 mkm). Prevailing are Zr similar to magmatic Zr from sample 8719/2	0,0000±0,0045	0,1234±0,0094	0,0022±0,0009	2005±128

Note: Analysed in laboratories of IGGD RAN. Zircons are selected by I.K.Shuleshko, Isotopic analisys - by B.V.Beljatskiy. For Zircons from samples 8719/2 and 105/4 corrections inserted for normal lead.

Table 2

Age of Zircon from metagabbro (tracks method)

N samples	Number of grains	Spontan tracks		Induced tracks		Age, mln years
		Number	Density $10^6/cm^2$	Number	Density $10^6/cm^2$	
67-3	1*	100	14.7	70	2.41	280±70**
Stand. Zr GS-3	4	403	3.31	653	5.36	27.9

Made by A.N.Komarov in IGGD RAN

* - Only in one grain of 15 contain of Uran stipulate for permit of tracks

** - Possible reduction of the age due to nature of spontan track have not been considered.

Chapter 19

The Structure and Composition of a Coal Shale Fragment from the Equatorial Atlantic

L.E.Shterenberg, R.N.Smirnov, A.P.Sokolova

During dredging in the Atlantic equatorial zone in the course of the 7th cruise of R/V “Akademik Nikolaj Strakhov” (March 1 - August 4, 1989), in addition to the magmatic rocks common to mid-oceanic ridges (tholeiitic basalts, dolerites, metabasites), several rock specimens were recovered which were attributed by some of the participants to continental deposits, Among them there was a coal shale fragment (Plate 16). Such single finds (unique for the oceans) have been encountered earlier, though not examined in detail. We received a fragment of the dredged coal shale sample to study its composition and genesis.

Various methods were used to examine the coal shale sample: microscopy, thermography, chemical analysis, etc. The sample is a compact, black coal rock with a thin-layered foliated structure peculiar to metamorphosed sediment. The rock easily foliates into thin sheets laminae. The shear surfaces are of bright blue, red and orange tarnish colors. The coal shale vertical section features laminae and lenses of lustrous coal. A coal trace on porcelaneous surface shows up black,

We have encountered serious difficulties while producing transparent and polished thin sections permitting of the coal in translucent light under an optical microscope. Numerous attempts to prepare the quality sections were in vain. Only very small sites of sections were transparent. The microscopic of thin sections, or more precisely their transparent areas in translucent light, and of polished sections in the refracted light have permitted conclusions on specific microstructural properties of the coal (present in the coal shale). Studies of thin and polished sections prepared from coal shale have shown main gelified mass to be generally homogeneous. It looks red-brown in the translucent light. The groundmass contains both fusainized (fusain, xylovitrainfusain, etc.) and gelified (vitrain, xylovitrain, etc.) components. The latter are relatively poorly distinguished. No cutinized (cuticules, macro- and microspore) or resinous bodies have been revealed in the thin sections translucent areas. The pyrite inclusions are observed in cellular areas of separate fusain lenses. The groundmass and gelified components are virtually isotropic in the translucent light with crossed nikoles. In the reflected light, the boundaries between the gelified components and the groundmass are difficult to place. It is only fusain and the similar components showing higher relief and reflection coefficient which differ drastically that from the coal’s gelified components.

Summing up the results of macro- and microscopic studies we may define it as a humic hard coal, which has evidently formed within the continental block, experienced metamorphic alternation and finally reached the near-OC stage.

In the polished section reflection values were measured with a “Leits” microscope-photometer (oil immersion, magnification 600 g). The reference used was the artificial glass whose reflection index in Kotg oil equaled 0.60%, a wave length of monochromatic light being 546 km. The results of measurements (done by I.E.Stukalova, GIN, Russ,Acad.Sci.) are listed in Table 1. The values obtained were compared with reelection values of gelified components of the Donets basin coal (Table 2). Comparison of data (Tables 1 and 2) permits attribution of the studied coal to meagre coals. Table 3 presents the results of chemical analysis performed in the Chemical Laboratory at the Institute of Fuel Minerals which also assign this coal matte to hard coals at the OC-T stage. What catches the eye is somewhat elevated the yet unexplained hydrogen content (H=4.72%).

Figure 1 (a-d) (Plate 21) demonstrates the heating curves of brown and hard coals of D-G and G grades (Ivanova VP. et al.). The same figure shows the heating curve of the examined coal shale sample (fig. le). The similarity of thermograms of our sample comprising the shining coal lenses and layers and showing high exomaximum value in the area around 760-770°C and the Donets basin meager coal is evident.

At the same time it is unclear whether the results of analyses are related to the regional metamorphism or presumably the coal was thermally affected later, possibly by hot thermal springs. The possible effect of relatively low-temperature (T- 500) hydrothermal fluids is inferred from the following: 1) The coal shale bedding planes are tarnish colored; 2) The anisotropy typical of regionally-metamorphosed hard coals (grades OC and T) distinct in the translucent light microscopy with crossed nikoles is missing; 3) Besides quartz, kaolinite, hydromica and probably small amounts of the mixed-layer mica-montmorillonite phase, dickite is also present. This dickite contained in the slate and coal(PIate 21, fig.2) is possibly a product of kaolinite transformation; 4) Very small single particles of native iron and aluminum (possibly hydrothermally introduced into a highly-porous rock) were revealed while resettling the rock in an agate mortar (Table 4, Plate 21, fig.1e). These were presumably the highly-porous rock by hydrothermal fluids. We note that besides the native metals the coal shale pestling has revealed magnetite, pirite and other minerals.

Thus, the integrated study of the available coal shale sample with the coal laminae and lenses has permitted to refer it (by several characteristics to the OC-T grade of hard coal. In addition to the fact that the regional metamorphism has alternated both the coal proper and the enclosing coal /clay, the coal matter transformation was presumably influenced by hydrothermal processes at the T<500. At the same time, the original locality of the studied coal shale sample has not been determined. It is difficult to judge whether the sample was dredged from the bedrock, or it was a chance fragment found on the ocean bottom.

Chapter 20

Relies of Continental Lithosphere in the Atlantic by Data of $(Th/U)_{Th}$, $(Th/U)_{Pb}$ and K/Ti Systematic

N.A.Titayeva, Yu.V.Mironov.

The Atlantic Ocean, which has been rather well-studied and which up to very recent times was considered as simple in terms of its structure and development, which were considered easy to explain on the basis of an "elementary" spreading theory, appears to be a useful subject to which one can repeatedly revert for discussion of a wide variety of global problems, including the problem of mantle reservoirs.

The considerable body of data now accumulated shows that variations of many geochemical and isotopic parameters of oceanic rocks exhibit a regional character, and cannot be explained by the proposition that all the diversity of magma varieties is derived from a homogeneous mantle [Pushtsharovsky, Peyve, 1994; Kenneth, 1982]. Although some scientists are of the opinion that mantle inhomogeneity (both horizontal and vertical) dates back to as early as the stage of Earth's accretion [Wood et al., 1979; Kurz et al., 1982], a concept of upper mantle vertical stratification going back to the stage of protocrust formation [Dupre, 1981; Wederpole, 1981; White, Hofman, 1982] is more commonly accepted. Lateral variability of basalt composition is thereby associated, first and foremost, with regional (and, not uncommonly, local) distinctions in dynamics of development of magma-forming systems involving one or more mantle reservoirs localized at different depths. At least two mantle reservoirs are recognized [Shilling et al., 1983]. One of them presents the upper mantle layer depleted (relative to the primary mantle) with incoherent elements with large ionic radii and radiogenic isotopes during the process of continental crust formation. This reservoir is considered to be a source involved in melting-out of tholeitic magmas, and the differentiates of these latter (N-MORB) are known to prevail in mid-oceanic ridges. Another reservoir, apparently deeper-seated, represents the primary mantle material probably enriched with a number of incoherent elements due to mantle metasomatism. It is deemed that this "enriched" mantle material goes up to the surface in the form of individual plumages, therewith determining the geochemical and isotopic specificity of the majority of alkaline lavas innate to oceanic islands. The interpretation of data on Pb isotopes shows that these two essential mantle reservoirs retain their specificity over a prolonged period of 1-2 billion years [Sun et al., 1975].

"Mixed''-type reservoirs, such as one apparently located under the Reykjanes Ridge [Sun et al., 1975] are also presumed to exist. Reservoirs of this type, which are intermediate by composition between "depleted" and "enriched" ones, are known to produce T-type of MORB. Magrnatic series of different oceanic island groups also show pronounced distinctions in geochemical and isotopic characteristics. For example, basalts of St. Helen are distinguished by strongly "radiogenic" Pb and "non-radiogenic" Sr, while in basalts of the Gough and Tristan da Cunha Islands precisely the reverse of these relationships is observed [White, 1985]. Based on this fact, a conclusion is made that basalts of the Gough and Tristan da Cunha island group appear to be derived from old continental lithosphere (except for the upper portion of crust), whereas basalts of St.Helen appear to originate from a mixture of depleted mantle material and submerged continental crust.

A conclusion that some "continental" component is likely to be involved in the composition of magma sources of some alkaline island series is based on the following facts:

1) geochemical and isotopic compositional similarity of these series to alkaline volcanites of intracontinental rifts known to be integrated into a common group of "intraplate basalts" in well-known discrimination diagrams [Pearce, Cann, 1983; Pearce, Norry, 1979; Wood et al., 1979];

2)higher values of "incoherent element with large ionic radius/incoherent element with small ionic radius" and "radiogenic isotope /non-radiogenic isotope" ratios relative to those typical for the primitive mantle.

The first fact convincingly indicates that the reservoirs associated with these different geodynamical situations are identical; on the contrary, the unambiguity of interpretation of the second above-mentioned fact turns out to be a subject of permanent discussion. This is due to the vagueness of the mechanism of crust material contamination by mantle matter as well as to a now prevailing view that the oceanic floor was formed as a result of spreading.

Methods of Investigation.

In our attemps to make clear the character of mantle heterogeneity we have used Th/U ratios which have been calculated using the data on Th and Pb isotopic compositions [Titayeva, 1990] and K2O/TiO2 values. These ratios belong to a group of parameters based on the relationship between incoherent elements and commonly used to discriminate basalts from different geodynamical situation [Lutz, 1980; Pearce, Cann, 1983; Pearce, Norry, 1979; Wood et al., 1979]. The values of these parameters are dependent on melting conditions (initial matter composition, location depth of melting region, degree of melting), but are practically invariable over the whole course of further evolution of melts [Cox et al,, 1979]. When a direct correlation is found to exist between these parameters and values of 87Sr/86Sr (or an inverse one with 143Nd/144Nd values), as it has been firmly established for a vast expance of the North Atlantic region [Shilling et al., 1983], variations of their values are primarily due to the composition of the source. When such a correlation is not observed, values of these parameters turn out to increasewith a decreasing degree of melting and/or increasing location depth of melting region. Despite these common features exhibited by the parameters in question, they also show some essential distinctions.

The Th/U ratio is a sensitive geochemical tracer exhibiting significantly different values in sources of volcanic rocks confined to major global reservoirs, i.e., primitive mantle, enriched continental lithosphere and depleted oceanic lithosphere. This fact is satisfactorily explained from the geochemical standpoint. U and Th are lithophile incoherent elements belonging to the actinoid group. When presented in their +4 oxidation number, they feature close ionic radii and similar geochemical properties. Low (and close) coefficients of their distribution between essential rock-forming minerals and melt [Volpe, Hammond, 1991] cause the similarity of their behaviour in the processes of partial melting and crystallization differentiation (within the limits of range of variations which we have choosen here). In the presense of oxidated aqueous fluids or solutions U tends to take on its higher oxidation number, i.e, + 6, and, as distinct from Th, turns out to be a rather labile constituent. The succeeding behaviour of U is governed by redox reactions. This results in a steady growth of Th/U ratio in continental crust rocks due to partial removal of dissolved uranium into the ocean [Taylor, McLenan, 1985].A certain proportion of the evacuated portion of uranium is thereupon fixed in the oceanic lithosphere, thereby causing a decrease in Th/U ratio in oceanic lithosphere and in magma derived from it. Changes in U/Th ratios caused by redox reactions are far in excess of those due to other processes. Higher mobility of U as compared to that of Th in the exogenesis zone (which is the only one open to sampling) hinders the use of Th/U elemental ratio as an indicative parameter.

(Th/U)th=Kth and (Th/U)pb=Kpb parameters, are free from the aforementioned drawback, as they are calculated on the basis of "single-element" isotope ratios, i.e., 232Th/230Th and 208Pb/206Pb, correspondingly; hence 230Th and 206Pb are decay products of 238U, and 208Pb is a product of 232Th decay [Titayeva, 1990].

The systematic of (Th/U)th=Kth is valid , by and large, for Quaternary volcanic rocks because of the relatively short 230Th half-life (80.000 years). The parameter Kth characterizes the Th/U ratio in a recent volcanic chamber and (with somewhat lower accuracy) in a magmatic source [Titayeva, 1986].

In distinction to Kth, the Kpb parameter is an integral quantity characterizing the "pre-oceanic" history of a magmatic source covering a period of time beginning from the onset of Earth formation and ending about 150 million years ago. Thus, the variations of Th/U ratio in recent oceanic lithosphere exert no effect on Kpb values. By virtue of sharp distinctions in magnitude of Ti and K ionic radii, the K/Ti systematic is one of the most highcontrast variants in a series of systematic equivalent in their content and based on relationships between incoherent elements [Mironov et al., 1993]. For a comparative analysis of rocks we have used a discrimination diagram constructed to discriminate basalts from dissimilar geodynamical situations [Mironov, 1990]. Its feasibility field is restricted to essentially basaltic compositions (by the classification given in [Marakushev, 1989]) with II < Na2O+K2O+0.41(SiO2-30) <14. This diagram would not do for more acid rocks, because in many differentiated serii evolving according to Fenner's type, going from basalts to the aforementioned rocks is accompanied by a mass-scale release of fern-titanic oxides. Thereby Ti becomes a coherent element, whereas K goes on accumulating in residual melt, which results in a steep rise in values of the ratio in question within an individual series.

Results of investigations.

The Kth systematic is represented in Fig. I by two histograms characterizing continental and oceanic reservoirs, which are thereby complementary to each other by the "mantle" value (plate 21, M area in Fig. I). According to the data of this systematic, Th/U ratio in the mantle is equal to 3.4 я_+я. 0.2. Depleted and non-depleted mantle regions are not distinguished in these histograms within the adopted step. Continental volcanites exhibit higher values of Kth parameter (>3.6), while the oceanic ones feature lower Kth values (< 3.2).

Values of Kth typical for volcanic rocks of the Atlantic Ocean and its western framing are represented in Fig.2 (Plate21) as a Kth-U diagram. Clearly seen in the diagram is a "mantle sequence" which is a horizontal band involving rocks characterized by different compositions and dissimilar U contents. In particular, the members of the Iceland serii (from basalts to rhyolites) differing correspondingly in U content, fit rather well in this sequence [Condomines et al., 1981; Titayeva et al., 1982]. At the same time, volcanites of some of the shoreland volcanic centers of Iceland fall into the "oceanic" part of the diagram located below. Also localized in this area are points representing tholeites of Mid-Atlantic ridge [Reinitz,Turekian, 1989; Condomines et al., 1982], and those of the Azores (E. Faial) [Oversby, Gast, 1988].

Above the "mantle sequence" (partially covering it) are localized the compositional points of volcanic rocks. Most remote from the "mantle sequence" are rocks of potassium alkaline serii of the East-African rift zone [Polyakov et al., 1986], as their compositions are thought to be associated with metasomatic processes in mantle. The compositional points of rocks of some Atlantic islands Tristan da Cunna Islands [Oversby, Gast, 1968], Cape Verde Islands [Titayeva et al., 1988] also fall into these area. Thus, basalts of recent eruptions such as the Fogo volcano (Cape Verde Islands), the Cameroun volcano in the African shoreland [Kochemasov et al., 1988] are, within the framework of Kth systematic, identical to oceanic basalts slightly contaminated with uranium (Kth = 3.0). At the same time, alkaline rocks of these islands (picrites, phonolites, carbonatites) and nephelinites (etindites) of the Cameroun volcano subordinate vent correspond in this diagram to the field of metasomatized mantle. The correlation is noteworthy which has been found existing between Th and Sr isotopic compositions both in continental rift [Polyakov et al., 1986] and oceanic rocks [Condomines et al., 1988].

The Kpb systematic is also represented in Fig.3 (Plate 21) in the form of two histograms characterizing oceanic and continental Earth regions. Cenozoic continental volkanites feature values of Kpb varying from 4.0 to 4.2 and showing steep rises in the areas of metasomatized mantle.Oceanic volcanic rocks, except for the areas of isotopic anomalies and islands corresponding to them, exhibit Kpb values varying within a restricted range from 3.7 to 3.9. At the boundary between those former and these latter a band of intermediate values is always present. The cause of quantitative distinctions between Kpb and Kth values is beyond the scope of our report.

Distribution of Kpb within the borders of the Atlantic Ocean (Plate 21, Fig.4) follows the same regularity which has been noted for Kth. However, comparatively larger amount of data on Pb isotopic composition has allowed for the marking out a number of anomalies. The largest band-type anomaly is located in the Southern Atlantic. Beginning at the shore of Africa, it is traced by Walvis Ridge [Richardson, Erlank, 1982], Discovery submarine mountain, Gough and Tristan da Cunna Islands [Sun, 1980; Hanan et al., 1986]. Within the borders of the Mid-Atlantic Ridge this anomaly covers a segment between 3455' S.Lat. and 46 21' S. Lat. [Hanan et al., 1986]. The Bouvet island falls out beyond the boundaries of this anomaly [Sun, 1980]. Another anomalous area is noted near the equator (O 56' N. Lat.) at the St, Paul ultrabasite massif [Roden et al., 1984]. It is intriguing that young olivine basalts occurring within the boundaries of this massif feature a typical oceanic value of Kpb = 3.87. One more Kpb anomaly is located to the north of the Azores, from 45 51' N. Lat. to 46 32' N.Lat. [Dupre, Allegre, 1980]. Individual islands of Azores differ from each other in isotopic composition. They exhibit values varying from typical "oceanic" ones (3,82 at St. George) to "subcontinental" ones (4.02 at Flores). Other anomalies may conceivable also exist, for which the isotopic data are yet not available.

It is noteworthy that alkaline serii are not necessarily characterized by anomalous Kpb and 87Sr/86Sr values. In particular, volcanites of St. Helena exhibit a oceanic composition [Sun, 1980]. In Ascension Island [Weis,1983] trachites, pantelleries, comendites and recent tholeites feature rather close, typical oceanic values of Kpb = 3.8. Much the same values are characteristic of granite of this island. At the same time, Kpb values of gabbro and monzodiorite xenoliths are close to the continental ones.

Another group of anomalies, apparently of different tectonic settings, is represented by islands located along the shores of continents:the Cape Verde Islands, Canaries, Madeira, Fernande de Noronha [Sun, 1980]. It is worth nothing that in certain segments of Kpb anomalies (for instance, in Cape Verde Islands) young basalts, as distinct from other alkaline rocks, exhibit typical oceanic Kpb (and Kth) values.

The K/Ti systematic of basalts indigenous to the Atlantic Ocean and some regions of intracontinental volcanism is shown in Fig. 5 (Plate 21). All the basaltoids occurring within the borders of the Ocean are characterized by $K_2O/TiO_2 < 0.65$. Midland basaltoids typically feature the values of this parameter larger than 0.32. In accordance with this fact, a sector of (O. 32< K_2O/TiO_2 < 0.65) is marked out in the diagram so that compositions of both oceanic and continental rocks fall into this sector. The boundaries between oceanic, intermediate and continental sectors coincide with the boundaries of corresponding fields in a discrimination diagram proposed earlier which has been worked out based on more extensive factual data [Mironov, 1990].

Purely "oceanic" K_2O/TiO_2 values (<0,32) are typical for a majority of basalts of mid-oceanic ridges and oceanic islands. As it has been shown earlier, each of the provinces of the Mid-Atlantic ridge separated by large transform faults is characterized by a certain relationship between Ti and K represented by a parameter of $A=K_2O/(TiO_2-0{,}75)$ [Mironov, 1991]. For instance, basalts with the lowest A values (<0,2) corresponding to N-type of MORB occur to the north of the Charlie Gibbs transform fault in Reykjanes and Kolbensey Ridges and to the south of the Oceanographer fault approximately up to 25 S. Lat. Hygher A values (0,2-0,6) typical for T-type of MORB are observed between the Oceanografer and Charlie Gibbs faults, to the North of the Jan-Mayen fault and in the southern region of Atlantic. The boundary between N- and T-types of MORB coincides (within the accuracy of construction) with a vector of K_2O/TiO_2=0,13, which is in agreement with the value of this parameter in primary mantle [Teylor, McLennan, 1985]. This fact is consistent with concepts stating that that the N-type of MORB is derived from depleted mantle, while N-type is thought to originate from a somewhat higher-enriched source.

The boundary between the N- and T-types of MORB in the coordinates of TiO_2 and K_2O contents also coincides with a demarcation line between TOR-1 and TOR-2 types whith are melted out at different depths [Dmitriyev et al., 1979]. This fact indicates that magma generation from a T-type enriched reservoir occurs at larger depth (25-30 km) as compared to that originating from depleted mantle, It has also been found that in the course of spreading development within the boundaries of a province an evolution of volcanism from T-type of MORB to a shallow N-type is observed in cases [Dmitriyev at al., 1984]. This fact implies that there is vertical inhomogeneity in the ocean mantle.

Tholeites and, to even larger extent, alkaline rocks of volcanic islands, the composition of which fall into the purely "oceanic" sector of the diagram (Iceland, Faeroe islands and many islands of the Pacific Ocean), are enriched with incoherent elements as compared with MOREL This fact might be thought of as reflecting the common tendency of increase in depth of melting-outs in the course of their development. It is seen in the example of Iceland (Fig. 5) that the compositional line of basalts innate to these islands is an extension (accompanied by a moderate but progressing deviation towards an increase in K_2O/TiO_2) of a trend of compositional changes exhibited by basalts inherent in the Reykjanes and Kolbeinsey submarine ridges localized in the same province. Although tholeites typically feature the values of A=0,2-0,3, and alkaline basalts exhibit A values running up to 0,4-0,5, in the Reykjanes and Kolbeinsey submarine ridges they are always less than 0.2 [Mironov, 1991]. Similar relationships are also noted in other regions, in particular, they may be exemplified by submarine T-type tholeites and somewhat higher-alkaline island rocks of the Red Sea Rift. However, an overall systematic shift of these formations towards higher values of K_2O/TiO_2 relative to those typical for volcanic serii of the Iceland dome results in the fact that the most alkaline rocks of the Red Sea islands turn out to be localized in the diagram in the vicinity of a boundary with the "intermediate" sector. One could also consider volcanites of the Cape Verde Islands and Bouvet island as alkaline analogues of T-type of MORB, since their compositions being also localized nearby this boundary.

Anomalies characterized by enhanced K_2o/TiO_2 values relative to conjugated segments of mid-oceanic ridges are, in addition, also manifested out of any association with island alkaline volcanism. For instance, the anomaly at 45 N.Lat. is bounded in the north by an area adjacent to the Charlie Gibbs fracture zone, and this area is characterized by extremely low A values (<0,2); in the south it is bounded by a series of segments with intermediate values equal to 0.2-0.4. In the anomaly by itself the values of 0.4-0.6 are prevalent. It is worth noting that Iceland volcanic complexes exhibiting rather close A values are characterized by significantly higher differentiation degree and alkalinity, which, along with some other petrochemical parameters (Fe8 [Klein, Langmuir, 1987], B [Mironov, 1991]), point to a larger (as compared to the region of 45 N.Lat.) depth of magma separation from mantle diapir. Thus, compositionally common-type mantle reservoirs in different provinces may generate magmas at different depths. Thus, enhanced depth of lithosphere and overall relief elevation in the region of the islands arc, first and foremost, associated with the increase in depth of melting-out. A correlation between "enrichment degree" of a mantle reservoir and relief elevation, which has been established with a high degree of confidence by the example of the Iceland Dome [Hart et al., 1973], holds good only within the boundaries of individual provinces, but will not work if used to account for relief distinction between the provinces IMironov., 1991]. These distinctions are more closely associated with depths of magma formation [Klein, Langmuir, 1987].

Volcanites of the Atlantic Ocean, which are characterized by "intermediate" K2O/TiO2 values, appear to be credible candidates for the role of indicators of continental-lithospheric blocks. They typically occur in two distinct position. Tholeitic basalts (E-type MORB) form local but intensive anomalies (A= 0.6-1.2) confided to areas of intersections of mid-oceanic ridges with certain transform fault zones {the oceanographer, Jan Mayen and some other faults) and to the Azores triple junction. They are characterized by a high degree of differentiation and extremely low values of some parameters (Fe8 [Klein, Langmuir, 1987], B [Mironov, 199 1]) which might point to the separation of magmas from injections of strongly differentiated abyssal metter. This latter appears to be the most enriched and, presumble, the most deep-seated of all essentially oceanic mantle reservoirs. Relatively differentiated variants of the tholeites under discussion, as considered in the framework of K-Ti systematic, are similar to tholeites of midland traps. However, relying solely upon the methods used in this study, we would not be entitled to make a conclusion about the complete identity of reservoirs feeding volcanic systems that are so dissimilar in geodynamical position.

Volcanites of some islands located within the boundaries of the aforementioned regions of isotopic anomalies (Gough, Tristan da Cunha, St. Paul islands) could be considered as indicators of subcontinental lithosphere with a somewhat higher degree of confidence. However, in this case different interpretations are also possible, In particular, these volcanites may be considered as alkaline "island analogues of the E-type of MORB. At the same time, it is of interest that, in distinction to other evolutional island serii, a noticeable decrease in K2O/TiO2 values with increasing alkalinity (and, correspondingly, depth of melting-outs) is observed. In Fig. 5 (Plate 21) this effect is reflected in a steep rise in TiO2 at K2O held constant. The same regularity is also noted for a volcanic series of continental-marginal Cameroun volcano. Tholeitic shallow basalts exhibit typical continental values of K2o/TiO2 closely approaching the crustal values [Taylor, McLennan, 1985]. As alkalinity increases, the K2O/TiO2 values take on an "intermediate" character, reaching (in most alkaline basalts varieties) values close to purely "oceanic" ones. As has been pointed out above and in the literature [Kochemasov et al,, 1988], such an oceanization of subcontinental lithosphere in both cases is also established based on the results of a diversity of isotopic studies.

Among the volcanites of East-African rifts, rocks of moderately-alkaline potassium-sodium formation and potassic alkaline formation are distinguished [Belousov et al., 1974]. The field of moderately-alkaline basalts in the diagram turns out to coincide almost absolutely with the field of alkaline rocks innate to "anomalous" islands and to the Cameroun volcano, thus exibiting geochemical similarity to a reservoir finding trappean fields. Only highly-alkaline rift rocks fall into the purely "continental" sector of the diagram. Their melting-out depth is estimated to be about 100 km [Belousov et al., 1974], and high K2O/TiO2 values are thought to be associated with the presence of metasomatized subcontinental mantle. However, by the virtue of the fact that in Fig. 5 the field representing the rocks under consideration, as well as that portraying highly-alkaline etindites from the Cameroun volcano subordinate vent, is essentially limited by a vector corresponding to the K2O/TiO2 value involved in the modelled lithosphere composition [Taylor, McLennan, 1985], it is not inconceivable that one of the causes of the aforementioned enrichment of rocks with K against Ti may consist in the extraction of incoherent elements from a crustal source.

Conclusion.

Using three independent methods based on relationships between incoherent elements, by and large consistent variants of systematic for volcanites of the Atlantic Ocean and regions of midland magmatism in Africa have been developed. Within the framework of (Th/U)th and (Th/U)pb systematic, clear distinctions between "depleted oceanic" and "enriched continental" reservoirs have been determined. The establishment of geochemical specificity of these reservoirs dates back to no less than 150 million years ago. The area of overlap of their compositional fields corresponds approximately to the modelled composition of primitive mantle. In the Atlantic Ocean individual areas (such as the region of 45° N. lat. in the Mid-Atlantic Ridge or the St.Paul island) and even whole provinces (the South-Atlantic Province, including Walvis Ridge, Gough and Tristan da Cunha Islands and a segment between 35° and 46° S.Lat.) are noted, within which volcanites exhibit typical “continental” composition. This allows for the suggestion of the presence of subcontinental lithosphere in these areas.

Through proportional relationships between Ti and K (elements radically differing from each other in the magnitude of ionic radius), four basaltoid groups are distinguished, Two of them (N- and T-types of MORB with corresponding relatively higher alkaline rocks of the majority of islands) are known to be found in oceans only, Inasmuch as the demarcation line between these groups goes through the K2O/TiO2 value equal to that typical for modelled primitive mantle, these groups may be thought to be associated with depleted and intermediate (or mixed-type) mantle sources. The depleted source virtually absolutely corresponds to the “oceanic” reservoir marked out by (Th/U)th and (Th/U)pb. Oceanic (but not continental) basalts predominantly falling into the overlap area are associated with the “mixed-type” source by (Th/U)pb systematic. Exception is made for tholeites of the 45 N. Lat. region of the Mid-Atlantic Ridge, which turn out to be “subcontinental” by (Th/U)pb systematic. Although they feature enhanced (relative to basalts of adjacent areas) K2O/TiO2 values, these values are restricted to the limits of purely “oceanic” interval .

Maximum K2O/TiOa2 values due to melting-out from intensively metasomatized mantle and, presumably, subsequent interaction of melts with continental crust matter are exhibited solely by the most potassium-rich and, as a rule, alkaline variants of basaltoids innate to intracontinental rifts and the Cameroun volcano. In oceanic volcanites with such extremely “continental” characteristics have not been found. All the rocks featuring “continental” values of (Th/U)pb and (Th/U)th parameters and, thus, considered as credible indicators of continental-lithospheric blocks in oceans feature lower values of K2O/TiO2, the latter corresponding to values typical for rocks of continental-marginal volcanois (such as Cameroun), traps and moderately-alkaline potassium-sodium variants of basaltoids innate to intracontinental rifts. Tectonic settings of oceanic formations included in this group are rather diversified.

Tholeitic “enriched” basalts (E type of MORB) are developed predominantly in the areas of intersections of mid-oceanic ridges with certain transform fault zones. They are associated with the most “enriched” and, conceivably, the most deep-seated of all the oceanic reservoirs. The assumption of complete identity of this reservoir and reservoir feeding intracontinental tholeitic and moderately-alkaline magmas lead one to infer that some geochemically homogeneous source is present in the upper mantle, and that this source is almost universally widespread in both oceanic and continental regions, In oceans this source is likely to be located beneath the depleted and intermediate-type mantle, and in the continents - immediately underneath the crust but above the regions of mantle subjected to a higher-intensive metasomatism. However, relying solely on the methods used in this study, we would not be entitled to state such an identity.

The most likely areas of continental lithosphere in oceans are fixed in the regions of islands which are “anomalous” by data of (Th/U)pb, (Th/U)th and K2O/TiO2 systematic (Gough, Tristan da Cunha, St. Paul islands). Here, as distinct from typical oceanic islands (such as Iceland and others), where alkaline rocks, as a rule, appear to originate from a more “enriched source than tholeites, “continental” values of all the parameters tend to give way to the “oceanic” ones as the alkalinity of rocks increases. A similar regularity is observed in continental-marginal regions of Africa (the Cameroun volcano). This suggests in both cases the presence of one of the mantle oceanic reservoirs beneath the crustal source. The number of areas with subcontinental lithosphere relics can be substantively enlarged if one includes here also the islands lying along the coasts of continents (Cape Verde Islands, Canarie Fernande de Noronha and some others), where oceanization of subcontinental lithosphere is reliably established by the results gained in (Th/U)th and (Th/U)pb systematic.

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Chapter 21

Residual Peridotites from the 15-20 Mid-atlantic Fracture Zone - a possible Analogy of the Ancient Metasomotized Mantle below St.Paul Rocks.

S.A.Silantyev, L.V.Dmitriev, B.A.Bazilev, L.K.Levsky, J.F.Casey, H.Bougault, M.Cannat

Introduction.

After the appearance of the well-known work by M. K. Roden and colleagues (Roden et al., 1984) many investigators studying mantle-derived peridotites from the Mid-Atlantic Ridge (MAR) suggested that amphibole-bearing peridotites of St.Paul Rocks (Equatorial Atlantic) are members of the metasomatized mantle enriched by light REE and characterized by the ancient age of their metasomatic and metamorphic recrystallization: 155 m, y,. This ancient age is absolutely unique for igneous rocks from MAR. In fact, such an old age of mantle derived rocks obtained at the crest zone of MAR allows for the assumption that not only crustal non-spreading blocks (Bonatti, Honnorez, 1971), but also lithosphere ones that avoided lateral movement related to contemporary spreading and preserved their position for at least 150 m. y., exist at the crest zone of the MAR. Data on the isotope composition of Sr, Nd and Pb in peridotites and hornblendites the St. Paul Fracture Zone (f.z.) gave M, K. Roden with colleagues reason to propose that the mantle below this segment of MAR is characterized by isotope heterogeneity. Data on variations of the geochemical parameters, observed along the strike of the axial zone of the MAR supplied since that time, led many investigators to ideas on the existence of longitudinal geochemical segmentation of the MAR crest zone, which correlates with character of distribution of the geophysical fields and bathymetry at the MAR Crest (Schilling et al., 1983; Bougault et al., 1988; Dosso et al., 1993; Dmitriev et al., 1994). In the past years features of the compositional segmentation of the MAR have also been established among layer 3 rocks: in gabbro (Silantyev et al., 1989) and residual peridotites (Michael, Bonatti,1985;Dick,1989;Bonatti et al., 1992). It is accepted that the geochemical segmentation of MAR reflects variations of melting degrees of mantle sources below the MAR Crest,. The application of this conception allows us to distinguish two anomalous MAR megasegments below which the mantle was affected by the greatest degree of partial melting: areas of the Azores Mantle Plume and 15-20 Fracture Zone intersection with MAR - 15-20 RTI. At the same time, taking into consideration data adduced in (Roden et al.,1984), it is impossible to rule, out the manifestation of at least short wavelength compositional segmentation at some MAR segments defined by geochemical heterogeneity of the mantle substratum below the MAR Crest.. Mantle derived peridotites are a most gratifying object for study in connection with this problem because these contain the most complete information about the geochemical and petrologic peculiarities of the mantle substratum. Here the 15-20 Fracture Zone is a unique polygon for investigations, since residual peridotites located here are characterized by the greatest variety of mineral types among other MAR regions, and also because the largest outcrop of peridotites among those known at Mid-Oceanic Ridges was discovered here (Silantyev et al., 1991). Residual peridotites studied in this work were dredged at the western slope of the Rift Valley of the MAR near western RTI - Transform-Ridge Intersection (Site 16ABP-56,15°37’N,46°42’ W) during the 16-th cruise of r/v “Akademik Boris Petrov”, and were obtained by the submersible “NAUTILE” on the western slope of the Rift Valley of thr southern MAR segment in eastern RTI (Site F03, 15°04’ N, 44°57’ W, South Inner Corner High) and on the eastern slope of the Rift Valley of the northern MAR segment to the north from western RTI (Site F12, 15°37’ N ,46°32’ W) (Plate 22, Fig. 1).

Analytical methods.

The mineral composition has been examined by N, N. Kononkova with the electron probe “CAMEBAX-MICROBEAM” in the Vernadsky Institute. Contents of major elements in hyper-basites were detected by T. V. Romashova with the help of the XRF method and spectrometer PW - 1600 “Phillips”, and contents of REE for the same samples - by G. M. Kolesov with the help of the method NAA according to (Kolesov, 1976). Study of the isotope composition of Sr, Nd, U и Pb in peridotites was conducted by G.V.Ovchinikova and B.V.Belyatzky with multi-channel mass-spectrometer “Finnigan-MAT 261” in the regime of simultaneous registration of ion currents of different isotopes in the Institute of Geology and Geochronology of the Precambrian.

Petrography, mineralogy, and petrochemistry of studied rocks.

As emphasized above, 15-20 Fracture Zone is characterized by a wide compositional spectrum of residual peridotites presented here. The compositions of the local mantle rocks varied from pyroxenites to peridotites close to dunites (the latter were discovered at eastern RTI only) on the classification diagram, based on the dependence existing between the variations of parameters MgO/SiO_2- $A1_2O_3/SiO_2$, and the melting degree of the mantle rocks proposed by E.Jagoutz with colleagues (Jagoutz et al.,1979) (Tab. 1., Plate 22, Fig.2), In the application of this diagram for oceanic mantle derived peridotites, it is necessary to remember that these rocks are almost entirely serpentinized, and that MgO is acquired in low temperature serpentinization and SiO2 - at high temperature (>250-300°C) serpentinization (Silantyev et al., 1992). However, deviations in the compositions of the studied rocks from the main trend of mantle rock composition variation constructed by E.Jagoutz and co-workers for relatively non-altered mantle xenoliths is so insignificant that it is possible to consider that distribution of compositions in 15-20 FZ peridotites into the selected coordinates reflects the realistic variety of the petrochemical and mineral types of the mantle derived rock at the investigated MAR segment, Fig, 2 shows that residual peridotites from eastern RTI of 15-20 FZ are distinctly close in their composition to most magnesium (depleted) compositions, corresponding to dunites and strongly depleted harzburgites. It is true also for the character of the distribution of the Cr# [Cr/(Cr+Al)] in relic spinel from residual peridotites across and along the strike of the 15-20 f.z. (Plate 22,Fig.3). Fig.3 gives a clear picture of the compositional segmentation of the MAR in the area of its intersection with 15-20 FZ, documented in the mantle peridotites, It is remarkable that the geochemical anomaly discovered at the South Inner Comer High (Site F03 is located in this area, Plate 22, Fig. 1) and related with extremely depleted peridotites, corresponding to the strongly manifested geochemical anomaly revealed early in layer 2 basalts (Casey et al., 1992). Basaltic rocks from this geochemical anomaly are characterized by enrichment of incompatible elements and belong to E-MORB. With regard to data from Fig. 1-3 it is possible to note that the ultramafic rock collection selected for the present study includes rocks obtained from the most principal points on the profile of geochemical segmentation for this area of the MAR,

Peridotites from Site 16ABP-56 were dredged on the western rift valley slope, 30 km north of western RTI (Plate 22, Fig. 1), presented by strongly tectonized harzburgites, and associate with recrystallized gneiss gabbro. Ultramafic rocks from Site 16ABP-56 in representative thin sections have protogranular textures; porphyroclastic and sub-porphyroclastic textured varieties, representing later subsolidus deformation overprinted on the protogranular textures, were also found. No evidence of diopside other than rare granular exsolution and exsolution lamellae in enstatite was seen, and they have accordingly intergrown with pseudomorphed enstatite. The relic phases are represented by reddish-brown spinel and olivine (composition in Tab. 2). Composition of the primary spinel from residual peridotites obtained on Site 16ABP-56 (sample 16ABP-56-99) corresponds to spinel compositions characteristic of moderate depleted mantle harzburgites.

Ultramafic rocks associated with different gabbro, including biotite- and potassium feldspar-bearing ones and relatively fresh basalts were obtained on Site F12, located on the opposite western slope of the rift (almost opposite Site 16ABP-56). These hyperbasites are represented by massive serpentinites with a characteristic pseudomorph-plate texture, determined by predominantly plate shapes of serpentine which composes the rock matrix. In its composition serpentine corresponds to the colorless kind and reflects low a temperature stage (<250°C) of serpentinization. Higher temperature metamorphic phases have not been detected in these rocks. Thin-prismatic segregations of brown spinel are equally disseminated in this rock. This spinel both in morphological features, and in its composition, has a high content of FeO, TiO_2 and MnO (Tab.2), corresponding to the stage of subsolidus metamorphic recrystallization of the primary peridotites, and therefore characteristic parameters of its composition can not be used as indicators of depletion degree of the mantle substratum in this case.

Ultramatic rocks from Site F03 were obtained on the eastern slope of the rise almost totally composed by residual peridotites, on the western slope of the rift valley at eastern RTI. The maximum depletion degree of mantle peridotites conforms to this point on the profile represented on Fig. 3 (Plate 22). Peridotites from here associate with normal tholeiitic gabbro and fresh basalts and dolerite-basalts. Sample F03-04b is typical serpentinite characterized by protogranular texture. A lot of thin chlorite-actinolite veins cutt this rock; chlorite is also present in the matrix. This mineral in its composition (Tab. 2) corresponds to clynochlore. Relics of primary reddish-brown spinel have also been detected in this sample.

Petrochemical and mineralogical peculiarities of the above-mentioned ultramafic rocks studied make it possible to state that these are close to residual peridotites obtained at many other MAR areas. Content variation of aluminum, alkaline and talc-alkaline elements in the described rocks reflects the modal proportions of primary silicate phases (at first clynopiroxene) At the same time, data on distribution of REE obtained for samples F12-04 and F03-04b (Tab.3) show essential enrichment of LREE in sample F12-04. Even taking into consideration some possible LREE mobility during low temperature recrystallization, it cannot be observed that the enrichment level of sample F12-04 is so high that it does not seem possible that enrichment of LREE is a geochemical peculiarity inherited from primary mantle derived rocks. It is important for clarification of this problem to obtain data on the isotopic composition of some elements, which are indicators of the primary nature of prematomorphic substratum for different igneous rocks including residual peridotites.

Data on isotope composition of Sr, Nd и Pb and their application for genetic interpretation.

Isotope compositions of Sr and Nd in sample F12-04 is represented in Table 4 and can be compared with analogous data obtained for samples 16ABP-56-76 and F03-04. Variations of isotope composition of these elements in studied samples allows us to make a few preliminary inferences, important for the identification of possible protolith types. The ultramafic rocks represented by sample F03-04b, as visible on Fig.4, close to mantle peridotites are complementary with tholeiitic basalts MORB by isotopic characteristics. Therefore this rock is possibly a residual component originated by melting of MORB mantle source. Insignificant shift toward ^{87}Sr enrichment, observed in this rock may reflect preservation of secondary phase, which retained its isotope composition of strontium in spite of the leaching procedure which has been used for studied samples, This assumption seems more true for sample 56-76 in that the value of ratio $^{87}Sr/^{86}Sr$ reaches 0.7097. Such a high value of this ratio alongside the accompanying "usual" level of $^{143}Nd/^{144}Nd$ value (0.5132), undoubtedly indicate that low temperature sea water originated solution participated in metamorphic recrystallization of primary rocks. On the other hand, isotope composition of Nd in sample 56-76 allows us to suggest that this rock belongs to the same group of residual peridotites as sample F03-04b. Extremely unusual isotope compositions of Sr and Nd are characteristic for sample F12-04. This sample demonstrates a very low value of $^{143}Nd/^{144}Nd$ (Tab.4,Plate 22, Fig.4) by strong enrichment by ^{87}Sr. Influence of marine sediment matter on Nd isotope composition must be excluded because features of brecciation and admixture of alien substance have not been observed in the studied rock. High value of strontium isotope ratio obtained for this sample can be explained by the same reason as analogous shifts established for samples F03-04b and 16ABP-56-76. However, isotope composition of Nd is a reliable indicator of a genetic type for mantle derived rocks and usually does not depend on secondary low temperature recrystallization of primary rock. (Faure, 1986). It is necessary to emphasize that peridotite from Site F 12 by its Nd isotope composition is essentially more unusual than amphibole-bearing peridotites and hornblendites from St. Paul Rocks, known until now as the most anomalous mantle peridotites below MAR (Plate 22, Fig.4). Another remarkable geochemical peculiarity of Sample F12-04 is a high concentration of Sm and particularly Nd (Tab.4). By this parameter ultramafic rock represented by this sample differs from mantle harzburgites, pyroxenites, websterites and almost undepleted lherzolites. It is known (Faure, 1986), that concentration in igneous rocks of both these elements rises as the differentiation degree grows. Alternative explanation of such a high concentration can be related to specific mineralogical composition of protolyth, The value of Sm/Nd in sample F12-04 (0.217) allows us to suppose the presence of primary biotite in this rock. It is notable that rare small scales of biotite were revealed recently in some samples of milonitic peridotites from the same MAR segment (unpublished data). Therefore it seems important to compare concentrations of Sm and Nd observed in residual peridotites from 15-20 FZ with those in different kinds of mantle xenoliths well-enough classified by their chemical and mineralogical compositions. Mantle xenoliths of the lherzolites (including garnet-lherzolites), metasomatized lherzolites with biotite and amphibole, different pyroxenite and websterite were chosen as objects for this comparison; data on geochemistry and mineralogy of these rocks are published by the following authors (Kramers et al., 1977; Obata, 1980; Frey, 1980; Roden et al., 1984; Bonatti et al., 1986). Note that data on St. Paul Peridotites and Zabargad Peridotites were used also in this work. The results of the comparison confirm that two types of mantle derived peridotites are present in the examined MAR segment: 1- with Sm concentration 0.1 -0.2 ppm and Nd -0.35-0.65 ppm (Sample 16ABP-56-76 and F03-04b), and 2- about 0.9 ppm and more than 4.0 ppm, accordingly (Sample F12-04). The first type corresponds to normal residual MAR-peridotites related with MORB suite. The second represents anomalous in the geochemical, and perhaps mineralogical sense peridotites that are close to compositional fields of metasomatized mantle peridotite hydrous phases bearing - biotite and amphibole (Plate 22, Fig. 5 and 6). Note that by their composition amphibole-bearing peridotites from St.Paul Rocks are located in the same compositional field,

Data on lead isotope composition in ultramafic rocks from site F12-04 are presented in Table 5. These data confirm the geochemical anomaly of sample F12-04 established above. A remarkable geochemical peculiarity of this rock is the enrichment of the radiogenic lead isotopes and similarity with amphibole-bearing peridotites of St.Paul Reeks in its isotope composition to lead (Plate 22, Fig. 7). The values of ratios $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}$, determined in sample F12-04 essentially surpass those characteristic of tholeiitic basalts of mid-ocean ridges (Stacey, Kramers, 1975), The most reasonable explanation of this phenomenon means ancient age for mantle derived peridotites from site F 12. This admission seems true because studied rocks are close to ancient mantle peridotites from St. Paul Rocks by other geochemical characteristics as mentioned above. Another notable peculiarity of peridotite represented by sample F 12-04 is a very low ratio of U/Pb. The value of this ratio may be used as an indicator of depletion degree of mantle substratum because U as well Th have been concentrated in liquids by melting. It is necessary to emphasize that sample. F12-04 is characterized by the lowest ratio of MgO/SiO_2. among the studied samples. Thus, contradiction appears between data on U and Pb concentrations testifying to a strong high degree of depletion of sample F12-04, and petrochemical data indicate that same rock is less depleted than the other peridotites studied in this work (Plate 22, Fig. 8). This contradiction may by eliminated by assuming that low concentration of U in sample F12-04 has been explained by its ancient age. It is necessary to add to what has been mentioned above that consideration that sample F12-04 has an unusually high concentration of lead (about 8ppm) (Plate 22, Fig.9) and a very low U/Pb ratio in this rock This phenomenon again conforms to the idea of the ancient age of studied rocks. Available data show that U is very mobile during low temperature metamorphism of MAR-Peridotites: positive correlation between its concentration and values of LOI (losses by Ignition) established for peridotites from site 16ABP-56 (Fig. 9) confirms this assumption.

Conclusion.

The study conducted of geochemical peculiarities of mantle derived peridotites collected at principle tectonic settings of MAR - 15-20 FZ RTI shows that within the examined MAR segment different scale compositional segmentation exists. Two types of compositional segmentation have been detected here. The first type is related to variations of mantle sources melting degree below MAR between 14° and 16%; the second is characterized by short wavelength segmentation and perhaps reflects geochemical and isotopic heterogeneity of the mantle substratum. Two kinds of mantle peridotites are present at MAR Crest between 15° and 15°40’N at least. The first one is residual peridotites depleted to a different degree associated with possible singenetic basalts and gabbro. The second is represented by unusual peridotites maybe corresponding to metasomatized ancient mantle substratum not related to manifestations of zero age volcanism at axial part of MAR. These unusual peridotites are close to peridotites from the St. Paul Rocks by geochemical characteristics. It is possible that these rocks, just like St.Paul peridotites (Bonatti, 1994) have been subjected to insignificant melting or none at all, E.Bonatti (Bonatti, 1994) assumed that peridotites from St.Paul Rocks are representatives of a mantle which has been located below the pre-oceanic (subcontinental) rift zone and that metasomatic recrystallization of these rocks took place during the separation of Africa and South America in the equatorial Atlantic (breakup of Pangaea)(150 m.y.). Discovery of possible geochemical doubles of St.Paul peridotites at 15-20 FZ allows us to suggest that representatives of anomalous mantle below MAR can be exposed at sea floor not only in the equatorial zone of MAR Crest but also significantly to the north of this region. Anomalous peridotites from site F12 are exposed as mentioned above on the eastern slope of Rift Valley; on the other hand peridotites from site 16ABP-56, normal in geochemical sense, were dredged at the opposite (western) slope of Rift Valley at almost the same latitude. Thus, serious reasons emerge for the assumption of the existence of short wavelength geochemical and isotope heterogeneity at the examined MAR segment similar to the one established by Roden with colleagues (Roden et al,, 1984) below the crest zone of MAR at St. Paul Rocks.

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Chapter 22

The Origin and the History of the Development of the Equatorial Segment of the Mid-Atlantic Ridge.

G.B.Udintsev

This monograph includes a collection of materials on the geology and geophysics of the ES MAR, obtained during three expeditions under the “Equaridge” program. The authors also took into account the materials published by other researchers who have studied this area of the ocean. The majority of the authors concluded that due to its structure the ES is substantively different from the existing image of the standard structure of the mid-oceanic ridge. Analysis of the data allows us to formulate our ideas regarding the specific characteristics of the structure of the ES and the main stages of its development.

At the same time, it would be wrong not to indicate reservations about the lack of basic data, the indefinite and varied nature of our conclusions, and to fail to understand the possibility and likelihood of an alternative interpretation of these data. This is possible because of the lack of information about the deep structure and the substance of the basement, Only a very limited amount of such data was available to our expedition because of the lack of needed equipment and of time, even during the three cruises, although that the opportunity for systematic study of the same object during three expeditions is rather rare and in academic work, and the experience was extremely positive. A more definite solution to the questions posed by the research therefore depend on the possibility of carrying out at the ES deeper seismic sounding, fuller gravity and magnetic survey, deep sea drilling and active collection of samples of the rocks of the basement from its outcrops using a manned submersible.

In this chapter we shall attempt to summarize the conclusions regarding certain anomalies of the ES, its origin and history, bearing in mind the information given in the previous chapters.

A comparison of the data regarding the structure of the axial rift zone of the ES with the data regarding the structure of the flank plateaus seems to provide convincing evidence of their difference, and that the structure of the flank plateaus cannot be considered as having been created during the process of spreading localized in the rift zone.

The axial rift zone of the ES processes all the typical and characteristic features of oceanic rifts, such as the linear regularity of the morphostructure, relatively large length of the cells of the rift valleys and rift ridges, linearity of the anomalies of the magnetic and gravity fields, an absence of noteworthy sedimentary cover, predomination in the composition of the basement of glassed pillow basaltic lavas of the typical MORB type, a segmentation of the rift with the appearance of discontinuity of all known categories of hierarchy, reflecting the irregularity and inhomogeneity of the mantle upwelling.

At the flank plateaus the morphostructure is characterized by a predominance of the system of wide plain-surfaced grabens between which are scattered narrow and relatively short horst crests (blocks). Side by side with horst, lineated in parallel to the general lineation of the rift zone, horsts are observed with lineation oblique to it, oriented NW-SE, closer to parallel to lineations of the shore lines of south-west Guinea to the east and of north-west Brazil to the west. The zones of intensive dislocations of the basement and sedimentary cover at the contact of the flank plateaus with the axial rift zone are notable, Such a zone of marginal dislocations overlaps the structures of the transform faults. The thickness of the sedimentary cover, negligible in the rift zone, increases very sharply with the transition to flank plateaus and are rather significant there.

Many outstanding features of the structure of the basement are visible in the structure of the anomalous gravity field both in the Free Air Anomalies and in the Bouguer and Mantle Bouguer Anomalies (Plate 11, 12). Along with the reduction of density of the mantle material under the axial rift and under all transverse fractures, which is usual for the Mid-Atlantic ridge, what is striking is the significantly smaller fragmentation of our segment by such transverse fractures, and also the substantive variations in the structural expression of these fractures and in their relations with the structures of the axial rift.

Judging by the nature of the gravity anomalies, in all transverse fractures of the Equatorial Segment a thinning of the crust and a rise of the densely reduced reeks of the upper mantle are observed. Since these transverse fractures represent yawning cracks as grabens, we follow the assumption regarding the origin of transverse fracture zones as a result of stretching, directed along the axis of the mid-oceanic ridge which was expressed by Larin and Solovjeva (1979). The link between the transverse fractures with such possible stress was demonstrated by data on the structure of the gravity field for the Vema fracture zone(Prince, Forsyth, 1988) and according to the characteristics of the focal mechanism of the earthquakes for the Kane fracture zone (Wilcock, Purdy, Solomon, 1990). In the same way, in our case characteristics of the structure of the gravity field apparently speak to the origin of these transverse fractures as a result of stress directed along the axis of the ridge.

The thinning of the crust to 4-5 km and corresponding rise of the roof of the Upper Mantle in the St. Peter fracture zone of the Equatorial Segment is clearly demonstrated by the results of deep seismic profiling (Plate 13, 14).

As distinguished from the Strakhov fracture zone, other transverse fractures of the Equatorial Segment as a rule do not cut through the axial rift and to varying degrees are masked by sedimentary and lava covers. Thus, for example, to the north of the Strakhov fracture zone, on the western flank plateau in the Gravity Anomaly maps through the stripe of the local lows a fracture is noticeable which is virtually not expressed either in the relief of the seabed, or in the relief of the basement. At the same time the St.Peter fracture zone, which is well expressed in the topography of the ocean floor, is noticeable in the anomalous gravity field for relative heights, attesting to the rise of the roof of the mantle. In addition, fractures are evident which are distinguished in their direction from perpendicular to the axis of the rift. What is noticeable here is the fracture in the southern part of the western flank plateau, oriented in the NE-SW direction. It is clearly indicated on the map of Free Air Anomaly based on satellite altimetry data (Sandvell, Smith, 1996) (Plate 11).

Another important feature of the gravity anomalous field observed in the Equatorial Segment is the clearly expressed local gravity highs and lows along the rift valley demonstrating the 3-d focused mantle swelling centers and mantle diapirs that feed and outline the spreading center (See crust and Moho profiles at Plate). Moreover, here one can speak of a definite hierarchy of mantle diapirs: diapirs of the first order are divided by trans-ridge and trans-oceanic fracture zones, while diapirs of the second and third order determine the discontinuities of the structure of the rift valley and the segmentation of the rift zone without splitting it through transverse fracture.

According to observations of seismic activity with ocean-bottom seismographs, positioned on the Strakhov fracture and in the rift valley to the north of it, a higher concentration was noted of the epicenters of earthquakes inside the nodal basin, on the site of the rift-fracture zone intersection (Plate 14). These observations can be considered as important evidence in favor of the assumption earlier expressed regarding the origin of the nodal basins as a result of the overlapping of stress vectors in the rift and in the transverse fracture (Udintsev, Agapova, 1991).

In the structure of the anomalous magnetic field there is a clear marking of the axial rift zone by characteristic regular positioning of linear anomalies (Plate 15). However, beyond the boundaries of the axial zone, on the flank plateau, this regular positioning is lost or changes direction. Thus, for example, regular lineations are noticed along the Strakhov, Saint Paul, Saint Peter and other transverse fracture zones. The between-fracture space of the flank plateau is characterized by an irregular, mosaic position of the anomalies, which are also considerably shorter in length (Plate 15). Proceeding from this, the structure of the flank plateaus should be distinguished from the riftogenous structure of the axial zone. The orientation of the anomalies of the magnetic field along the transverse fractures can be explained by the role of serpentization of ultrabasic rocks of the mantle diapirs under the impact of water seeping into the depths along the fracture. The serpentization can also explain the elevation of the structures of the flank plateaus along the edges of the Strakhov fracture zone, noted in the topography of the ocean floor. Such a reason for the rise of the crests in the rift zone was proposed in the past by T. Francis (1981) and by Zonnenshein and co-authors (1989).

In the structure of the sedimentary cover of the flank plateaus (see Plates 23-33) there are characteristic deformations of extension inside grabens and compressions over the horsts. Not infrequent diapiric structures are noticeable. Their origin can be associated with piercing of sedimentary cover with diapirs from evaporite layer - if we assume the possibility of the existence here of such a layer, analogous to known evaporates of the eastern part of the Angola, Sierra Leone and Gambia basins. In the case of the ES, such a layer very likely is armored with a basaltic cover, but in conditions of regional extension, at the fragmentation of that cover the uprise of the salt diapirs can be expected. It is remarkable that the morphology of the diapirs observed at the ES is very similar to the morphology of the diapirs from messinian layer at the floor of the Mediterranean Sea. It is not, however, excluded that the diapirs at the ES can be the diapirs of serpentites.

The rocks of the basement, outcropped at the normal faults of the horst blocks of flank plateaus, are as a rule represented with the usual set of samples of rocks of the second and the third seismic layers of the oceanic crust, but together with them were found rather numerous samples of rocks which are rather exotic for the ocean crust: granodiorites, granosienites, phyllites, brown coal, coal shale. Grains of the brown coal are present in some levels in the upper part of the core samples of the sedimentary cover from flank plateaus.

Together with the tholeitic basalts and magnesial gabbro of usual content for the volcanicity of the ocean areas, the basement of the flank plateaus contains basalts of amygdaloidal texture, laminated ferriferous gabbro passed through metasomatos, and amphibolitisied peridotites not so typical for the rift zone. Amphibolites from flank plateaus are metasomatic, formed on the base not only of gabbro and peridotites but also on phyllite. The plagio-granite, quartz-syenite ferriferous-quartzite can considered as rocks exotic for the oceanic area. The presence in the contents of the basement of the flank plateaus of the rocks of a deep origin, exposed. to regional metamorphism and metasomatos in conditions of high temperature and great pressure, in our opinion, attest to the primordial greater thickness of the continental or sub-continental crust (in contrast to thinner oceanic crust), experienced later with submergence and thinning.

All these data on the structure of the flank plateaus do not correspond to the usual image of the rifted structures. We cannot definitely identify the structure of the flank plateaus as platform, but we recognize in that structure a very close analogue to platform structures of the continents. That is most probably quasi-platform structure, corresponding to the status of a certain stage of reworking of the former platform in conditions of its stretching and thinning synchronously with tectono-magmatic activation, which led to the formation of vast lava fields (large igneous provinces), Here morphostructure, structure of the sedimentary cover, gravity and magnetic fields, and the specter of dredged rocks corresponding to this are observed.

It is important to decide which of the possible active factors can be defined as applicable to development of the Equatorial segment of the Mid-Atlantic Ridge. The structure of the margins of the lithospheric plates in the rift zone of the ES does not display the evidences of their rigidity. There are, on the contrary, some marks of the capacity for destruction in the en echelon type of structure of the rift valley, in its non-transform discontinuities. Evidence is available of varieties of development in space and time in comparison with neighboring segments of the Mid-Atlantic Ridge. The clearly expressed features of the spreading are presented only in the axial zone limited to 60-80 miles. The intensive marginal dislocations of the flank plateaus testify to the role of transverse compressional stresses directed from the axial rifted zone. It allows for the assumption of the local development of the mantle upwelling,and local and focused mantle diapirism in the narrow axial zone, marked with negative gravity anomalies, and also corresponding to this local spreading in the axial zone. For the flanks plateaus, however, we must suppose only moderated scattered spreading or general extension.

The model we require must provide an explanation for the existence within the limits of the Equatorial segment of the Mid-Atlantic Ridge of the narrow riftogenal axial zone, and of the wide flank plateaus with quasi-platform structures. There are previously proposed models by E.Bonatti designed to explain the remains of the relics of the non-spreading blocks inside the limits of the riftogenal structure of the Equatorial part of the Mid-Atlantic Ridge (e.g. Bonatti, Honnorez, 1971) by the position exactly above the zone of the divergence of the upwelled mantle jets (or streams), or by the migration of the rift and transform faults (Bonatti, Crane, 1984). J. Johns proposed an explanation of such events by a delay in displacement of the lithospheric plates in the South Atlantic relative to displacement of the plates in the North Atlantic (Jones, 1987).

We would prefer to use a combination of the elements of the models proposed by K.Hinz (1981) for the development of continental margins with the creation of lava covers leading to submergence (fig. 1) with the version of the model proposed by P. White and D.Mackenzie (1989) for the creation of the rift fracture as the result of development of the wide mantle plume and large igneous province above it (fig,2). This is followed by localization of the mantle upwelling, leading to the creation of the rift fracture as the result of the splitting of the crustal dome above the mantle jet, but not as the result of the horizontal displacement of the plates). These elements would be combined with the ideas of importance of the mechanism of the general extension of the lithosphere (Mutter, Carson, 1990) for creation of the large igneous provinces (Coffin, Eldholm, 1991), the creation of the oceanic basins above the area of the lack of the mantle material on the sides of local mantle upwelling, and with the ideas regarding the history of the development of the Atlantic Ocean by K. Storetvedt (1985, 1987), and J. Pogrebitsky et al. (1993). On the base of such a composite model we would use the idea of the mantle upwelling and diapirism as the result of modest Earth expansion (Barsukov, Urusov, 1984, Udintsev, 1984, 1987).

According to the model we initially accepted (White and Mackenzie 1989) the character of the magmatic activity in the process of the development of the ocean depends on the conditions of the thinning of the lithosphere and on the uprise of the hot mantle stream or mantle diapir. That can occur according to various models (fig. 3). Model “a” - in unthinned lithosphere temperature in the mantle increases with the depth , but at the same time increasing pressure keeps the ductile rock under the lithosphere from melting, If the lithosphere is stretched, (model "b") it leads to the thinning of the continental crust and its submergence beneath sea level, However, if it occurs without connection to the uprise to the base of the lithosphere of relatively hot mantle jet (or diapir) - little or no melt is ordinarily produced. If the hot mantle rises and temperature at the foot of the lithosphere increases with approaches to the melting point (model "c"), the same amount of stretching as in model "b" , caused with uprise of the swell above mantle stream, leads to a volcanic outburst and the reduction of the pressure in the upwelling mantle generates magma that spills out and has influence on wide areas on the land and floor of the sea basin created due to stretching, and on the continental margins around it as flood basalts of the large igneous provinces.

The lithosphere could be rifted by the divergence of the lithospheric plates , driven by the mantle convection, or even if mantle plumes do not drive plates, because the uplift caused helps create the rift fracture through gravity, which stimulates crustal scales to slide apart down opposite slopes of the swell. After the lithosphere has been rifted (model "d") the hot mantle rock can well up to the surface. Because the melting temperature falls as pressure is reduced, part of the initially solid material melts as it moves upward. The resulting magma solidifies and can formmagmatic structures in the axial zone of the newly created mid-oceanic ridge.

For their original model, White and MacKenzie (1989), like Hinz (1981), use the most popular concepts of plate tectonics. According to this concept the formation of oceanic rifts and basins of the ocean as a whole is due to the movement of lithospheric plates as an active factor. The active movement of plates is accompanied by a passive buildup on their edges of magmatic rocks forming a crust of an oceanic type. In such an outline of development of the ocean bed it is difficult to explain the existence within the limits of the mid-oceanic ridges of exotic crustal blocks similar to the flank plateaus of the equatorial segment. However, the system of views of White and MacKenzie provides for the possibility of initiating a rift crack through the rise itself of the mantle diapir,and by its dynamic impact, and not only by the movement of lithospheric plates. Although the extension of the lithosphere leads to the immersion of its surface over a broad area, nevertheless above the axial zone of the mantle plume, where the mantle is at least 50 degrees hotter, the mantle material with low density can raise the crust by 1000-2000 meters above the periphery parts of the newly formed basin. Under the impact of the force of gravity the litosphere scales slip down from the slope of such a rise, moving towards the sides and promoting the opening up of rift cracks (Artjushkov, 1995) and the formation of transverse fractures (Iljin, 1983). Thus, the mantle plume can initiate riftogenesis.

The possibility for such active impact of the mantle diapirs, the hot fields or plumes, and a blob of material 50-100 degrees hotter than the lithosphere, which expanded over an area of about 2000 km, leading to regional outpouring of basalt fields, assumed by White and MacKenzie (1989), was recently also considered by Storey (1995) as a probable active factor, Earlier, in a number of works (Udintsev 1984, 1987) we already laid down the geomorpholidical basis for the concept that the crust of ocean areas is heterogeneous, and that the active factors for its formation are the rise of the mantle diapirs resulting from the expansion of the mantle substance, accompanied by a very moderate extension of the Earth's crust. We therefore propose an outline of development of the Equatorial Segment according to a model virtually similar to the model of White and MacKenzie, but with a different sequence of events.

We assume the original stage of evolution of the Equatorial Segment to be the development of the intercontinental bridge. The creation of such a bridge could be stimulated by a rise of the large mantle diapir or plume, which led to the heating of a large area of the lithosphere. It also led to a moderate regional extension of the crust and to the outpouring of a large field of basalts onto the surface of the future bridge and onto the continental margins of Africa and South America. This episode was finalized with the thermal subsidence of a newly created intercontinental bridge . We propose to call such a process pontogenesis from the Latin word *pontis* - bridge). It seems very likely that the northern structural limitations of that bridge was a cluster of enormous fractures (transoceanic in future) from the Vema fracture zone in the North to the Doldrums fracture zone in the South. It seems possible that this cluster of fractures represents a kind of system of steplike faults along the former continental slope pointing towards the north of the intercontinental bridge to the previously formed basin of the North Atlantic.

The central area of this bridge was buoyed by the dynamic impact of the mantle diapir and it led to the rifting of the crust at the axis of the arch and the sliding of plates apart down its slopes. The tectono-magmatic activity was localized then in the axial rift zone (riftogenesis).

The scales of the mantle diapir are determined by spatial boundaries of the development of the axial rift ridge. Outside of it the continuing moderate extension leads to an even greater immersion of the former continental bridge, at a depth greater than that of the shelf and coming closer to that of the oceanic (oceanization, thalassogenesis). In conditions of this extension there is a formation of a system of fractures determining step-like faults of the continents of Africa and South America, and on the former continental bridge there is the formation of a system of fractures, and the grabens and horsts determined by them as “remnants of subsidence”, aimed approximately parallel to the contours of the shorelines of West Africa and North-East South America.

Simultaneously, an extension in the meridional direction leads to the formation of a system of transverse fractures of a lower category than transoceanic and transridge. Some of these fractures, formed before the formation of the ridge crack are later covered by riftogenous structures of the axial zone. On the other hand, the fractures formed after the formation of the rift system or before it but which are actively continuing to expand, cut across it. Relics of the continental bridge are preserved as flank plateaus of the Equatorial Segment and they act as the structural threshold between the North and South Atlantic.

Along the continental margins of Africa and South America a complex of structures of accommodation is developing in the form of steplike faults, marginal horsts and grabens and also basaltic-sedimentary accretive prisms of oceanward dipping layers.

Thermal contraction along the periphery of the mantle plume while it was cooling determined the maximum submersion in the basins of Sierra Leone and Guinea to the east and Guyana and Brazil to the west.

In the annex to such a model, and using paleogeographical formations by Storedvedt (1985, 1987), Patrunov (1987), Pogrebitskij et al. (1990, 1993), Andreev (1996), we find that the history of the formation and development of the Equatorial Segment has undergone the following stages.

1) The beginning of regional extension in the space of the future Equatorial Continental Bridge linking Africa with South America (ponto- genes) in the late Triassic-early Jurassic time. The rise of the mantle plume leads to a flow of a bubbles of melted material moving significantly to the side and to the formation of an enormous lava bubble with a large hot field above it. The temperature of that bubble is 50-100 degrees higher than it can be in a stationary flow, when the stream is later localized in the rift crack. In conditions of developing extension there is a progressive immersion, diffused cracking of the basement and outpouring of basaltic fields - large igneous province, which is widespread over the area of the intercontinental bridge and marginal areas of continents..

These basalts were covered by red-colored deposits of the Triassic and by evaporates of the early Jurassic periods. The diapiric structures inside the sedimentary cover observed on the flank plateaus of the Equatorial Segment can possibly be represented by diapirs of Jurassic evaporates. In conditions of a continuous extension of the crust repeated explosions of magmatic activity occur and also outpouring of other portions of lavas.

In the late Jurassic-early Cretaceus the regional extension of the Equatorial Intercontinental Bridge above the uprising mantle stream continues, but due to its gradual cooling and narrowing there was localization of tectono-magmatic activity in the axial part of the arch created under the dynamic impact of the mantle diapir. The development of the arch proceeds on the background of regional submersion of the whole area of the former intercontinental bridge and is accompanied by the formation of the deepest basins on both sides of the newly created rise.

2) In the late Cretaceous the continued uprise of focused mantle diapirs stimulated cracking of the axial zone of the crustal arch (riftogenesis). The reduction of pressure stimulated partial melting of upraised material. The outpoured lavas now created fields which were not large, butrather local and narrow stripes of basalts, younger than basalts of flank plateaus. Vertical tectonic movements in the rift zone can be considered as a combination of the normal downfaulting at both sides of the rift cracked valley with an uprise of fragmented blocks as a result of the serpentinisation of mantle ultrabasic rocks at their foot. The uprise of the axial rift zone led to separation of the flank plateaus.

In conditions of continuous extension in the limits of flank plateaus and cracking of basaltic cover, an uprise of diapirs from the layer of hypothetical Jurassic evaporates appears possible. On the flank plateaus wide grabens and narrow horst are formed, as remnants of subsiding. one of the largest horsts is represented by the Sierra Leone Rise which is viewed as a huge block of a former intercontinental bridge A wedge-shaped form of such horsts stipulates local compressions and corresponding deformations of the sedimentary cover on their top, while inside of the grabens exists a situation of extension of sedimentary layers.

3) In the early Cenozoic (Eocene-Miocene) the formation of the basins on both sides of the Equatorial Segment of the Mid-Atlantic Ridge led to the submersion of their floors to the oceanic depths (thalassogenes). That was accompanied by the activisation of oceanic water circulation, with the formation of the abyssal plains due to outpouring inside of the basins of numerous turbidity currents from continental margins, and due to the activity in transportation of sediments with currents along the slope and near the oscean floor. The submersion of the floor of these basins creates a possibility for penetration

through them of Antarctic waters from the South to the North Atlantic in the Eastern part of the Equatorial Atlantic. In its Western part such deepening of the oceanic basins permits turbidity currents to move along the Equatorial Sea channel from the Ceara abyssal plain to the south, to the Pernambuco abyssal plain (Gorini, 1981).

The former intercontinental bridge, now the structural barrier in form of the Equatorial segment, preserved the elevated position for some time. We see evidence of this in intensive transportation of sediments from its side, from the north to south to the trenches of complex trough of the St. Paul f.z. We also see the relics of the higher position of the ES in the flat tops of the Belousoff, Muratov and Vinogradov seamounts, now submerged to their modem depth.

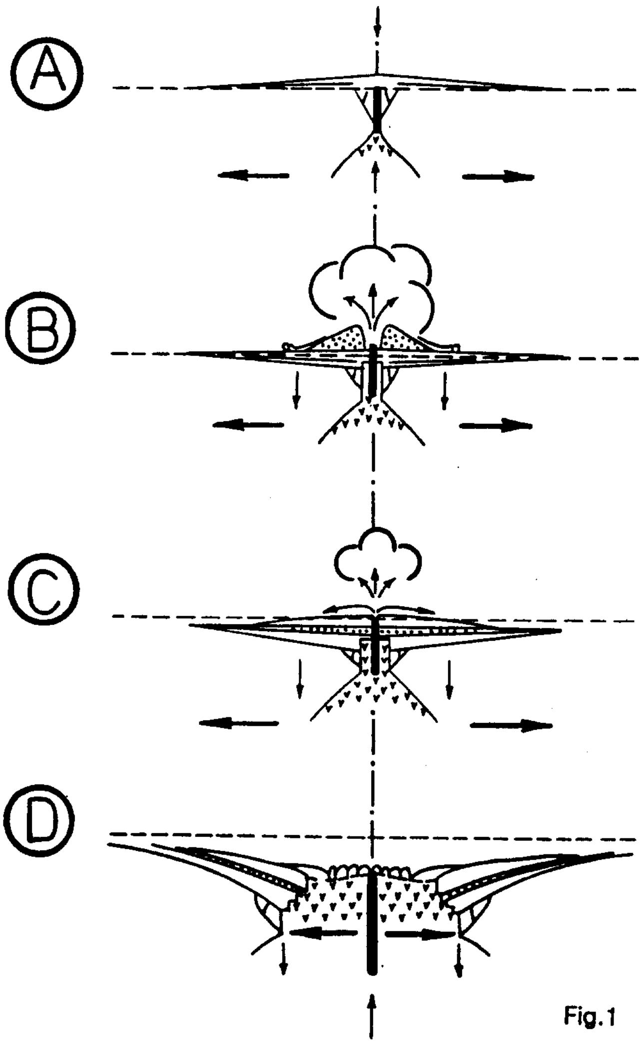
A
B
C
D

Fig.1

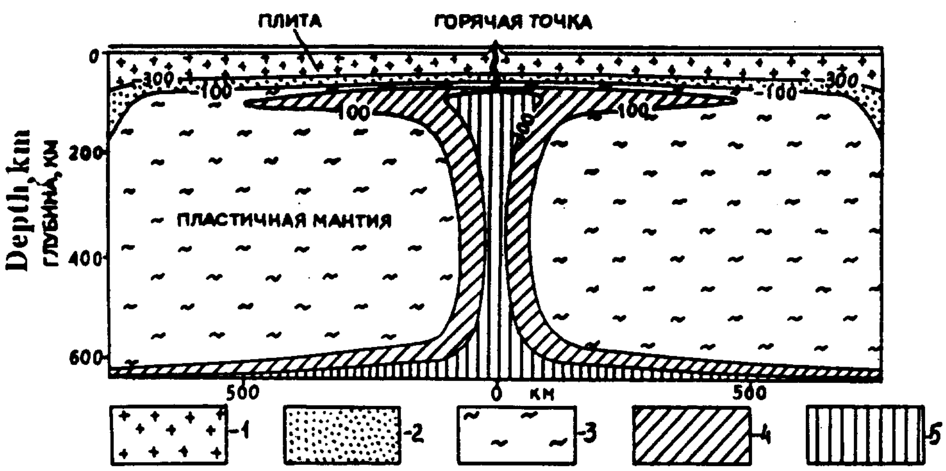
ПЛИТА
ГОРЯЧАЯ ТОЧКА
ПЛАСТИЧНАЯ МАНТИЯ
Depth, km
ГЛУБИНА, КМ
0
200
400
600
500
0 КМ
500
300
100
1
2
3
4
5

Fig.2

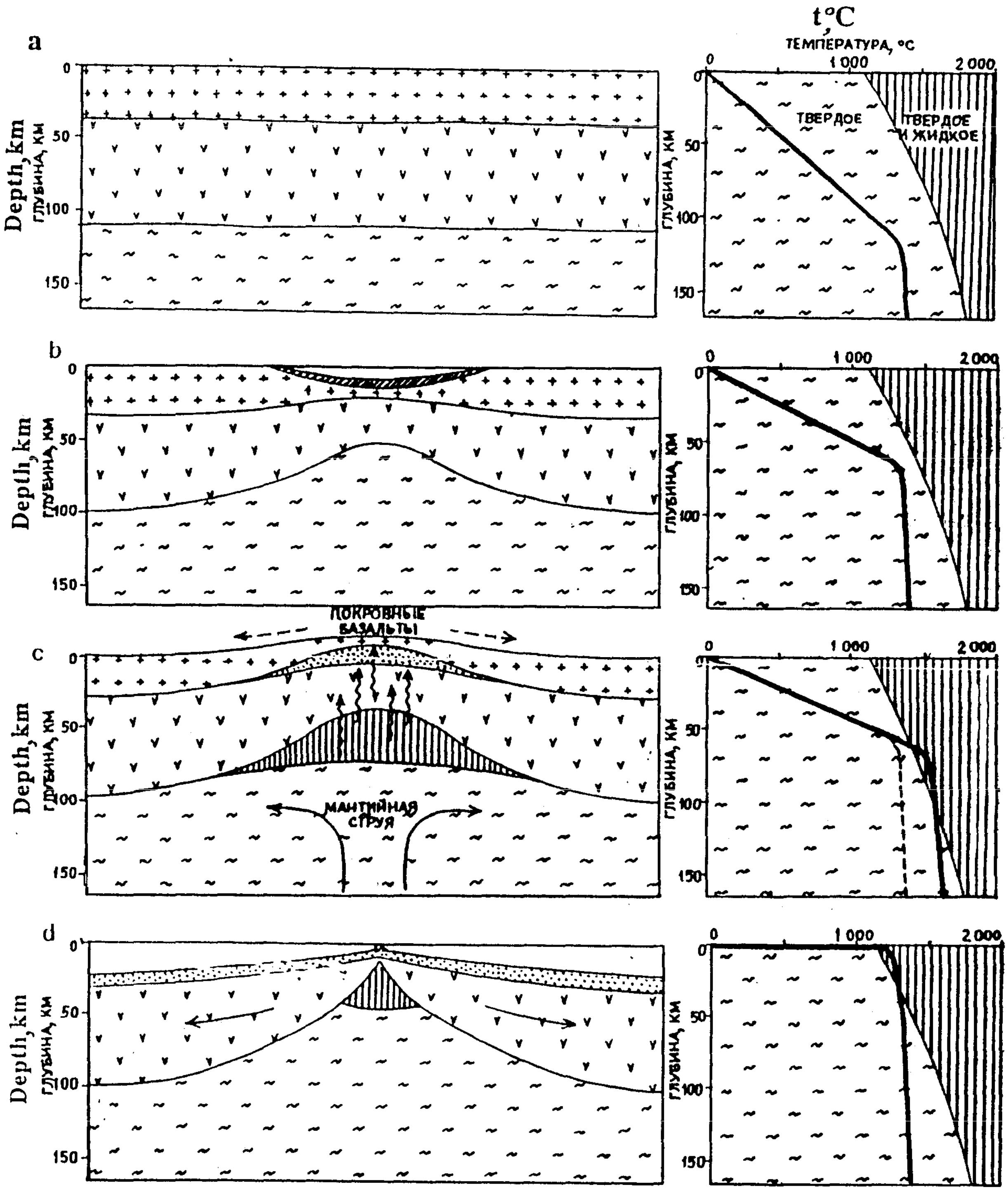
a
b
c
d
t°C
ТЕМПЕРАТУРА, °С
Depth, km
ГЛУБИНА, КМ
ТВЕРДОЕ
ТВЕРДОЕ И ЖИДКОЕ
ПОКРОВНЫЕ БАЗАЛЬТЫ
МАНТИЙНАЯ СТРУЯ
0
50
100
150
1 000
2 000

Fig.3

Conclusion

G.B.Udintsev

In considering the results of the investigations of the Equatorial Segment of the Mid-Atlantic Ridge, we concluded that in comparison with adjacent segments of the Ridge to the North and South, this segment is truly anomalous. It is also anomalous in comparison to the ideal standard scheme of the structure of the Mid-oceanic Ridges usually used when geologists operate with the concept of plate-tectonics.

We also come to the idea that an active factor in the origin and development of the mid-oceanic ridges must be the active mantle upwelling, which stimulates progressively variable tectono-magmatic activity, but does not" cause considerable horizontal displacements of the lithospheric plates in conveyer style, accompanied by accretion of the converged plate margins with newly created oceanic crust. In our view, the mantle upwelling causes the transformation of the original crust (probably continental in some cases, and oceanic in others) along four principal ways, which led to the creation of the four principal genotypes of the morphostructures of the oceanic floor:

a) general heating, extension and fracturing of the original crust above spacious mantle plume (astenolith) , overlapping of the original crust with large igneous fields (basalt lava fields), submergence, accompanied by migration up the Moho discontinuity, and visual (according to seismic sounding) thinning of the crust (extensional or pontogenal genotype),

b) swelling of the crustal arch (dome) by the localized mantle jet, breaking of the dome of the future mid-oceanic ridge with the creation of a rift fracture and localization of the tectono-magmatic activity in the axial rift zone, where neo-volcanic activity increases along with vertical movements, stimulated by the serpentisation of the mantle rocks, together with horizontal movements and fracturing of the crustal scales which are gravitationally sliding down the slopes of the dome (riftogenal genotype),

c) progressive crustal submergence above peripheral parts of the coding mantle plume followed by the formation of oceanic basins on both sides of the mid-oceanic ridge (subversive or thalassogenal genotype) and

d) downwarping of the continental margins in the form of continental flexures, complicated by a series of step-like faults and asymmetrical grabens, usually accompanied by the formation of volcanic or sedimentary accretion prisms (continental flexure genotype).

The spatial scales of the manifestation of the results of the combination of the application of these four types of processes in the creation of the final combination of the above-mentioned genotype lead to a variety of morphostructure of different segments of the mid-oceanic ridge. Some of them look anomalous compared to some segments, but are very similar as compared to others. Examples of such variations of morphostructure we can see in referring to the General Bathymetric Chart of the Oceans (GEBCO, 1984), recent maps of the gravity field of World Ocean based on satellite altimetry (Haxby, 1985, Sandwell, Smith, 1993, 1994, 1996) , to a complex of geological-geophysical information,and results of the detailed survey of different segments of the mid-oceanic ridges. We thereby observe not only differences between the Equatorial Segment and adjacent segments of Mid-Atlantic Ridge, but also similarity with some other segments of the same Ridge and other branches of the planetary system of the mid-oceanic ridges. Features of morphostructural similarity are observed in examples with the Icelandic, South Azores , Ascension , St. Helena and Tristan da Kugna segments of the Mid-Atlantic Ridge,and with the South Madagascar-Del Cane, Amsterdam-St. Paul, Red Sea segments of the Mid-Indian Ridge. We also simultaneously observe rather remarkable features of a certain similarity between these and other segments of the above-mentioned ridges. This leads us to suppose that the cause of such differences and similarities lies in the reflection in the morphostructure of various segments of the different combinations of genotypes varied through space and time due to variation of some deep process. It creates the impression that in a variety of morphostructures of different segments of mid-oceanic ridges we observe the evolutionary sequences of the development of the various parts of this planetary structure. It probably reflects the evolutionary development of a deep process.

In our view the supposition is logical regarding major variations in the development of the deepseated process of mantle upwelling through space and time as the active factor of development of the structures of the ocean floor through the evolutionary complex, including the above- mentioned genotypes of the morphostructures. Such evolution has to undergo interaction between the mantle upwelling of variable rates, thermoactivity and contents, as the most active factor, and with the variable behavior and resistance to transformation of the original crust in the area, subjected to the influence of the mantle upwelling, as the more passive factor. In this supposition the role of the active factor of the origin and further development of the mid-oceanic ridges, accompanied with their segmentation, belongs to mantle diapirism without horizontal movements of great lithospheric plates on a large scale, but followed by a reworking of the original crust accompanied by the development of the above-mentioned genotypes of the morphostructures. The correlation between the areal scale of development of such genotypes - and the prevailing type of structure in certain segments of the mid-oceanic ridge and in the entire part of the heterogeneous ocean - depend upon the local combination of the active and passive factors.

Naturally, such suppositions lead us to ideas on the heterogeneity of the mid-oceanic ridges and of the whole ocean floor, rejecting the idea of their uniformity as a result of the active movements of the great lithospheric plates. We have to look back at how that idea developed.

Along with studies of the mid-oceanic ridges, demonstrating their substantive heterogeneity , studies in other morphostructural provinces of the oceans also provided evidence of the definite heterogeneity of the structures of ocean areas and the nature of the tectonic-magmatic activity, defining their structural appearance. Thus, deep seismic profiling of the flanks of the mid-oceanic ridges have demonstrated the substantive distinction between the structure and foundation and the characteristic riftogenous structure of the mid-Atlantic ridges. (White et al, 1984, 1990, Mutter et al., 1985). Simultaneously studies of a number of continental margins of a passive type showed the substantive role of the formation of volcanic-sedimentary prisms which accompany the formation of the flexure of the continental slope (Hinz, 1992), and progressive oceanization of the continental crust away from the continental margin (e.g. Rosendahl et al., 1992).

Evidence of substantive heterogeneity of the structure of mid-oceanic ridges, reflected in their segmentation, and signs of structural distinctions between mid-oceanic ridges and the floor of the ocean basins, as well as the specific characteristics of the structure of the continental margins evoked doubts regarding the uniformity of the structure of oceanic areas, which are assumed to exist by the concept of plate tectonics (e.g. Artiushkov, Schlesinger, Yanshin, 1984, Udintsev 1988, 1989, Rezanov, 1990).

The question arises as to what deep-seated process can lead to the formation of mid-oceanic ridges with their division into a multitude of segments reflecting the evolutionary development of the ridge and the oceanic basins associated with them. First and foremost, this can be a rise of the mantle material as a result of radioactive heating and an approximately 5 per cent decrease in its density (Barsukov, Urusov 1984). The same kind of radioactive heating of the of the mantle material is insufficient for the organization of closed convection, and can lead only to its rise in the process of quasi-convection, accompanied by zonal melting (Vinogradov 1963, Vinogradov ,Yaroshevsky, 1965).

According to Vinogradov, with the coming of zonal melting there is only a partial melting of the original material starting with extremely great depths through radioactive decay of potassium, uranium, and thorium. This decay gives clearly insufficient heat for genuine melting, but sufficient for melting of a small fraction and degasation of the mantle. The rise in the melting in the earth’s body upwards, moreover, consists not only and not so much in the emergence of lighter liquid into the gravity field, as in the consistent welding of rocks with compensation of their crystallization at the bottom of the melting zone. Melting which arises at great depths - at several hundred or even thousand kilometers - as a consequence of even insignificant heterogeneity of the mantle’s properties, both in composition and in energy will function as a reason for major heterogeneous factors. Vinogradov believes this is an attractive explanation of the fundamentally localized nature in space of the character of the geological processes in the Earth’s crust on the background of a generally homogeneous picture of the heating of the originally cold planet. According to Vinogradov, zonal melting is the reason for localization in space and time, and of the shift of matter and heat inside of the Earth’s body. Moreover, the process, when it reaches the Earth’s crust and determines the “cosmetic” appearance of the planet’s surface, develops rhythmically, creating a striped structures observed, inter alia, in the mid-oceanic ridges.

Vinogradov emphasized that it should not be thought that a single column of a differentiated substance is active within the Earth’s mantle, and that it should be assumed that the area of total or partial melting at a depth can have a significant extension in length (the mantle plume of today’s western authors), but along with its rise there occurs a dissipating of this enormous zone into individual columns already penetrating into the upper mantle nearly simultaneously, but separately in space. The reasons for the disintegration of the primary source of melting can be any type of non-homogeneity of the mantle. Its reflection on the Earth’s surface will be both a heterogeneous segmentation of riftogenous mid-oceanic ridges, and the inhomogeneity of oceanic basins as a whole.

As Vinogradov saw it, the nature of the process is non-linear. This differs sharply from the linear nature of the geodynamic process proposed by the concept of plate tectonics. According to Vinogradov’s idea, the structure of riftogenous ridges will be accompanied by non-riftogenous or scattered riftogenous fields which have spurted up to the surface from great depths, basalt fields and the field of immersion, compensating for the shortage of material along the periphery of the zone of mantle upswelling. It is precisely for this reason that the idea was born that riftogenous mid-ocean ridges are a particular tectono-magmatic system - riftogenal- opposed to the platform system plates of the oceanic basins. The combination of these two most important systems determines the most important features of the heterogeneity of the oceanic areas of the Earth.

On the backdrop of the resounding successes of the concept of plate tectonics in the 1960-70s, Vinogradov's idea was not widely accepted, although it continued to be developed in the works of his colleagues (Barsukov, Urusov,1984) and others, e.g. (Belousov (1986). Today, however, a great deal of new data is appearing, which has revived interest in Vinogradov's assumptions. These are data about the very deep roots of the continents, and about the great depths of the melting of basalts, which spurt up to the surface at the initial stage of the process. The latter is confirmed by the discovery in basalts of trapp fields the inclusion of the isotope Helium-3 (Basu et al, 1995). This isotope is linked to the earlier stages of the Earth's history, and has been preserved only at the deepest horizons of the mantle. A clear picture of the depth emergence of sources of the mantle's melting and their disintegration along with the upwelling to the surface with the formation of a system of focused mantle, was demonstrated by seismic tomography (Dziewonsky, Woodhouse, 1988). The active role at the early stages of the development of the oceans of mantle upwelling in the form of hot fields or mantle plumes, has become clear through the understanding of the important role at this stage of broad fields of basalt trapps - large igneous provinces (Storey, 1995).

The process of zonal melting must be pulsating in nature and must be accompanied by spatial disintegration of the primary broad mantle diapirs. This leads to the reduction of the space of the basalt fields initially melted from it from large depths, and later from smaller ones. This takes place during consistent episodes of the spurting up to the Earth's surface of giant lava fields, then replaced by episodes of localized tectono-magmatic activity in the rift zones, with the formation of linear bodies of different scales in the riftogenous segments of the mid-oceanic ridges. The decrease in the density and increase in the volume of the heated mantle material leads to regional arching and extension of the earth's crust above the broad mantle diapirs during the first stage of its development. At subsequent stages of development when the mantle diapir is localized, its rise leads to a dynamic elevation of the riftogenous ridge, linked with the simultaneous immersion along the edges of the riftogenous arch - because of the shortage of the material along the sides of the localized diapir.

Spatial-temporal inequalities in the development of the mantle upwelling, leading to the formation of a system of chains of focused diapirs, in turn leads to segmentation of the arch of the mid-oceanic ridges. The differences in the morphostructure of the segments and in the petrology of magmatic rocks of their basement reflect the stages of the evolution of the cores of the upwelling. The structure of the floor of the ocean basins beyond the borders of the ridges is most likely consonant with the structure created at the stage of general extension. It must be distinguished not only from the structure of the riftogenal ridge. Variation in their structure depends not only on the complexity of the flexure but also on the role of volcanic-sedimentary accreative prism developing on it. The role of such a prism is very great at the early stage of development, but is reduced later because the sediments begin to be transported to the floor of the deepened basin. Naturally, everything expressed here represents only a first attempt at resolving the question of the homogeneity or heterogeneity of the ocean floor. We believe that heterogeneity is more in keeping with the data observed. The weakest point in our knowledge, however, continues to be the lack of information on the structure of the deep basement of the ocean basins.

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