Petrology and Geochemistry of Basalts from the Red Sea Axial Rift at 18° North

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ABSTRACT

Detailed studies by submersible were carried out in the axial zone of the Red Sea Rift near 18°N during the Soviet Red Sea expedition of the Oceanological Institute of the Academy of Sciences (December 1979-March 1980). The initial bathymetric, magnetic and seismic surveys established the general organization of the symmetric tectonic steps (1-3) descending towards the axial rift. The 4-5 km wide inner floor of the rift was explored during 21 dives. It is occupied by 100-300 m high, young pillowed volcanoes, isolated or grouped to form elongated hills, frequently cut by open fissures except in the zone of most recent extrusion.

The 42 samples collected are typical plagioclase \pm olivine \pm clinopyroxene \pm spinel, more or less porphyritic mid-ocean ridge basalts whose compositions were mainly controlled by polybaric fractionation of plagioclase, olivine and minor clinopyroxene. They have been separated into porphyritic and sub-aphyric groups using modes and mineralogical criteria. Mineral-liquid equilibria, crystal zonation, and modal proportions indicate some magma mixing but probably only of closely related magma batches within each described group, as can occur inside a single magma chamber. Crystal accumulation is believed to have played a significant role in only a few porphyritic samples.

Three sub-groups (from less to more evolved; (a) $FeO^*/MgO < 1.22$; (b) $1.16 < FeO^*/MgO < 1.48$; and (c) $FeO^*/MgO > 1.49$) were distinguished on the basis of glass and whole-rock major element chemistry. Glass compositions follow the multisaturated cotectic-like curve for MORB-type basalts and show a general evolution very comparable to what is observed on the Mid-Atlantic Ridge near 36°N, but are less diverse than in the FAMOUS area. ⁸⁷Sr/⁸⁶Sr, ¹⁴³Nd/¹⁴⁴Nd, and ²⁰⁶Pb/²⁰⁴Pb data for 4 samples show strong similarities to those from the Mid-Atlantic and East Pacific Ridges, and indicate no continental contamination despite the fact that they have been produced during recent continental break-up and ocean opening. ²⁰⁶Pb/²⁰⁴Pb values, Th/Ta vs. Th/Tb correlations, and rare earth element patterns allow recognition of three different groups of samples, indicating that the Red Sea Rift near 18°N is fed by a heterogeneous mantle source. The chondrite-normalized LREE

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patterns strongly suggest that variations in the degree of melting also played a significant role. A Tb/Th vs. Th diagram shows that the crystal fractionation inside each group of samples cannot exceed 20%. Transition element versus hygromagmatophile element correlations show that: (a) these magmas have already fractionated olivine; (b) no mixing between a very primitive magma and a more evolved one occurred; and (c) clinopyroxene did not play a major role in the fractionation.

The geochemical and petrological data clearly identified two magma groups related to two distinct volcanic ridges and a third magma group related to the rift floor. The new data set, compared with the FAMOUS-Narrowgate set for the Mid-Atlantic Ridge provides some constraints on magma chamber evolution beneath low spreading rate ridges. It suggests the possibility of ephemeral (10^2-10^3 y) magma chambers producing magma batches which build the volcanic ridges, alternating with more steady-state magma chambers (on the scale of less than a few 10^5 y), less than 2 km wide, which could produce larger volumes of magma erupting over much larger surfaces, smoothing the topography of the axial rift. Each of these magmas would have its own geochemical and petrological signature while remaining within a restricted range of MORB-type compositions.

INTRODUCTION

The Red Sea axial trough marks the separation of the Arabian and African lithospheric plates, which began some 5 Ma (Roeser, 1975) to 6 Ma B.P. (Izzeldin, 1982, 1987). It is generally accepted that ocean floor spreading is occurring at the axis of the trough: the magnetic anomaly pattern, at least in the central and southern part of the Red Sea, supports this hypothesis. The spreading rate inferred from these anomalies is one of the slowest known from present day active oceanic spreading centres, with an average of 1.5 cm/y of total spreading over the last 5 Ma (Chase, 1969; Schilling, 1969; Schneider *et al.*, 1976; Joron *et al.*, 1980*a*; Bonatti *et al.*, 1984).

In spite of the many studies of the Red Sea over the past twenty years, very few samples of the basaltic basement had been recovered from the sea-floor. Except for the very axial part of the Red Sea, basalt outcrops are rare, because a thick layer of Miocene evaporites covers most of the sea-floor. Moreover, the evaporites in the central and northern part of the Red Sea have undergone gravitational flow towards the axial lows, masking the basement outcrops. Even in the central deeps and in the axial rift where bare oceanic crust is exposed, the roughness of the topography makes dredging difficult and several recent cruises have failed to get good collections of basement rocks by dredging. The DSDP programme was no more successful when the drilling ship *Glomar Challenger* visited the Red Sea area on Leg 23 in 1972: of the six holes drilled (Whitmarsh *et al.*, 1974), only one penetrated 9 m of the basaltic basement, recovering several fragments less than 4 cm in diameter (Hole 226, cores 1–2, Atlantis II Deep).

In the southern part of the Red Sea $(16^{\circ}N-19^{\circ}N)$, the axial trough exhibits a regular and well-delineated rift valley (Fig. 1). This segment is probably the main zone of sea-floor spreading in the Red Sea, since the maximum age of accretion appears to be younger going north or south of this segment (Pouit & Guennoc, 1983; Izzeldin, 1987). It was selected for extensive bathymetric, geophysical and geological studies by Prof. A. S. Monin and his research team from the Shirshov Institute of Oceanology, Academy of Sciences of the USSR, Moscow. The Soviet Red Sea Cruise, with the participation of the second author of this paper (T.J.), took place from December 1979 to March 1980 and involved three research vessels. One of them, the Akademik Kurchatov, supported the manned submersible Pisces XI used for detailed bottom investigations.

Basaltic samples were recovered from a restricted area of the inner floor of the axial rift at 18°N, most of them collected by submersible and precisely located. Several preliminary papers have given the general results of this cruise (Zohnenshain *et al.*, 1981; Eissen, 1982; Matveenkov *et al.*, 1982; Monin *et al.*, 1982; Al'mukhamedov *et al.*, 1983; Juteau *et al.*, 1983).



FIG. 1. General situation of the studied area (A) and bathymetry of the Red Sea Rift near 18°North (B). The delineated square shows the location of Fig. 3. Original scale 1/100 000. Isobath interval: 100 m.

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The samples were analysed for major and trace elements and for Sr, Nd and Pb isotopes. This paper presents the results of these combined petrological and geochemical investigations. In the following discussion we will: (1) describe the geological setting of the studied samples; (2) present a complete description of these samples, including: modal analyses, microprobe analyses of the minerals and fresh glasses, whole-rock, trace and REE analyses, and Sr, Nd and Pb isotopic ratios; (3) attempt to evaluate how partial melting, crystal fractionation and/or magma mixing processes have controlled their evolution; and (4) compare them with other oceanic ridge basalts and discuss their similarities and differences.

GENERAL RESULTS

Structural setting

The bathymetric, magnetic and seismic surveys were carried out by the R/V Professor Chtokman in a quadrangle lying between $17^{\circ}48'N$ and $18^{\circ}18'N$, and $39^{\circ}44'E$ and $40^{\circ}18'E$, over about 2500 km² (Fig. 1). Here the Red Sea rift has a pronounced linear structure with an average orientation N150°E. Its transverse section is typical for the segment between 16°N and 19°N (Fig. 2). It differs notably from the more northern part of the Red Sea rift (19°N-24°N) (Guennoc & Thysse, 1982; Bonatti *et al.*, 1984; Pautot *et al.*, 1986) which is characterized by the existence of deep, circular, isolated pools, some of them containing hot brines. The southern transverse sections are symmetrical with respect to the axis, and consist mainly of interlocked steps going down by successive stages toward the axis. These are, starting from the external zones (Fig. 2) (Zohnenshain *et al.*, 1981; Monin *et al.*, 1982; Juteau *et al.*, 1983):

(1) An external step, 500-700 m below sea level, established over a several kilometer thick mass of Miocene evaporites, which appears as a reflector on seismic profiles (Whitmarsh *et al.*, 1974) under 200-300 m of Plio-Quaternary sediments. The width of the rift between the internal margins of these upper steps is 35-40 km.

(2) 1 to 3 internal (or lower) steps in variable number, ranging between 1100 and 1500 m below sea-level. Their width is 10-15 km for the two external steps, and 4-5 km for the innermost step. They are bounded on the inner side by spectacular fault scarps (up to 600 m high), dipping 60° towards the axis, whereas their surfaces are gently inclined outwards and



FIG. 2. Geological section of the Red Sea rift at 18°N (after Monin et al., 1982). (1) Basalts of the axial zone (less than 0.5 Ma). (2) Basalts of the lower steps (0.5 to 3 Ma). (3) Plio-Quaternary sediment cover. (4) Miocene evaporites. (5) Normal faults. a: Axial Volcanoes. b: Lower Step. c: Middle step. d: Salt slope. e: External step.

covered by Plio-Quaternary sediments. Thus these internal steps constitute huge asymmetrical blocks whose internal margin is uplifted and forms steep scarps (the 'rift mountains'). The dredges made on these fault scarps during this cruise brought up only basalt pieces, showing that the basement of these blocks is probably of oceanic nature.

(3) An axial rift, 4-5 km wide, whose floor is at 1300-1600 m below sea-level. In a transverse section, this floor exhibits a central volcanic ridge made of up to 3 individual volcanoes (100-300 m high), separated by irregularly shaped topographic lows (Fig. 3). At the base of the fault scarps bounding the axial rift, narrow linear marginal deeps, down to 1700-1900 m, have been observed. This arrangement is rarely symmetrical, since the location of the central volcanic ridge is variable and the deeps often appear only at one side of the ridge.

(4) True transform faults have been observed in the studied area (Fig. 4), the most important of which is the fracture zone No. 3, named the '18°N Fracture Zone'. There the axial zone attains a width of 10 km, more than twice the usual width of the inner floor, and exhibits a series of deeps perpendicular to the rift axis, reaching 1700–1800 m below sea-level



FIG. 3. Bathymetric map of the axial area with location of the dive tracks (large numbers) and of the samples (closed circles with small numbers). An accuracy of the navigation of less than 50 m was obtained by using acoustic transponders (crossed circle).



FIG. 4. Distribution of the isochrons of the oceanic crust in the central part of the Red Sea near 18°N (after Zohnenshain *et al.*, 1981). Small numbers indicate the age (Ma). Large numbers refer to the fracture zones. The double dashed line represents a fossil spreading axis. The striped lines delineate the recent crust formed after the spreading jump. The salt slopes are hatched.

and separated by parallel ridges. Only the axial rift and the first faulted step are interrupted by transverse faults (Fig. 4). This indicates that the processes which initiated the fracture zone at 18°N occurred after the formation of the second tectonic step and are at present restricted to the rift axis.

Morphological and tectonic zonation

The relief and the bathymetry of the structures allows recognition of three tectonic zones in the studied area (Fig. 2): (1) the axial zone, 10–15 km wide, characterized by an active extensional regime with open fissures, normal fault scarps and recent volcanism; (2) a zone, including the lowermost steps and the associated bounding fault scarps, 10–25 km wide, characterized by an outward tilting of the blocks with tilts of 4–7°; the vertical displacements have an average amplitude of 400–500 m; and (3) the marginal parts of the rift (> 30 km wide), including the upper 'salt' steps, constitute a third inactive zone, devoid of significant tectonic activity, but undergoing slow subsidence.

Magnetic anomalies and evolution of the axial rift at $18^{\circ}N$

Isochron map and estimation of spreading rate

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A well-marked magnetic anomaly coincides with the narrow axial extrusive zone containing the most recent volcanoes (Zohnenshain *et al.*, 1981). The isochrons (Fig. 4) are interrupted by offsets at several places. Six fracture zones were outlined, with offsets ≥ 1 km (6 km for the 18°N fracture zone). The distribution of the isochrons suggests an approximate 6 km eastward jump of the accretion axis between 0.5 and 1 Ma B.P.

The average spreading rate for the last 2 Ma is 1.6 cm/y. However, the spreading rates calculated for shorter time spans have changed considerably: 2 cm/y from 1.71 to 0.89 Ma,

3 cm/y from 0.89 to 0.69 Ma, then 1.04 cm/y from 0.69 Ma to the present. Thus over the past 2 Ma, the accretion process appears to have been highly irregular.

Geodynamic reconstitution

A palynspastic reconstruction using the fracture zone directions to bring together isochrons of the same age from both sides of the rift, shows that at 3 Ma, the salt edges exhibit a considerable overlap (15–18 km) (Fig. 5): this means that the salt flowed out considerably since the formation of the crust 3 Ma B.P. Some 2 Ma ago, the salt formation probably closely resembled the brine-pool stage which exists at present between 19°N and 22°N. During its evolution, the axial zone of the rift at 18°N has probably already gone through the 'brine-pool stage'. If so, deep drill holes should find, at the bottom of the Plio-Quaternary sediments covering the second and third tectonic steps, local concentrations of metalliferous sediments formed in fossil brine-pools some 2 Ma ago.

Geological description of the inner floor of the axial rift from submersible observations

An extrusive axial zone, 0.5–1 km wide, consists of young volcanoes, entirely made of pillow lavas with reflective glassy surfaces more or less devoid of sediments. Taking into account the high sedimentation rate in the Red Sea (about 5–10 cm/1000 y after Monin *et al.*, 1982), and the very thin palagonite rims exhibited by the glassy margins of the pillowed basalts (Hékinian & Hoffert, 1975), it is clear that these axial volcanoes are extremely young, certainly less than 1000 y for most of them, and less than a few hundred years for some of them (Matveenkov *et al.*, 1982). The whole axial rift floor is dissected by countless open fissures (gjás) and normal fault scarps, most of them parallel to the rift orientation. In some places, however, swarms of transverse fractures and scarps have been observed, some with evidence of strike–slip movements corresponding to new transform fracture zones.

Along strike, the ridge is composed of fissure-volcanoes 200–300 m high, 500–1000 m wide, either isolated or clustered in short ranges 3–4 km long. More than 40 individual volcanoes of this type occur in the studied area. They have generally steep slopes (30–60°), and appear to be entirely constructed from pillow-lava tubes. No sheet flows or collapsed lava ponds were observed. Many of the tubes are hollow, especially towards the top of the edifices, and their fragile external glassy crust has frequently collapsed. The infrequent eruptions have not buried the tectonic scarps in the extrusive zone, as occurs in faster-spreading ridges (CYAMEX Team, 1981; Gente *et al.*, 1986). The talus at the feet of these scarps lacks sediment, proving that these fault scarps are at present active. Open craters, 8-10 m in diameter, are visible at the summit of some of the youngest volcanoes.

With the exception of these young volcanoes, the basaltic crust is obscured as a result of the high sedimentation rate. The general supersaturation of sea-water in carbonate at the bottom temperature of 22–24 °C leads to abundant encrustations restricting observation and sampling. For instance, the spectacular 500 m high eastern fault scarp bounding the axial rift at 18°N is entirely encrusted with carbonates, and the lava flows are visible only when the wall is overhanging. On both sides of the extrusion zone, the young oceanic crust is affected by numerous open fissures and active fault scarps with huge talus accumulations, most of which are devoid of any sediment cover.

No evidence of active hydrothermal vents was observed except for scarce hydrothermal rocks among the dredge samples and one black iron sulphide sample picked up during Dive 69 (Monin *et al.*, 1982).

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FIG. 5. Diagram of evolution of the Red Sea rift at 18° N from 3 Ma to the present, obtained from palynspastic reconstruction (after Zohnenshain *et al.*, 1981). At 3 Ma, there is a considerable overlap of the salt slopes. However, this overlap is just apparent and demonstrates that the salt slopes have flowed considerably after the formation of the oceanic crust at 3 Ma. Before its dislocation, the Miocene salt mass, covering the entire axial zone, has been subjected to strong distensional forces as basaltic magmas were probably intruding under it, forming sills or laccoliths. At 2 Ma, the salt slopes are just overlapping in the northernmost part of the studied area, whereas there is already an axial area 4 km wide in its southern part, where basalts start to erupt directly on the rift floor. At 1 Ma, the salt mass is already completely dislocated and the salt slopes are close to their present position. At this time the third faulted step appears. At 0.6 Ma, formation of the second faulted step. At 0.5 Ma, formation of the first faulted step with spreading jump and end of the activity of the transform fault number 5; start of activity of the transform fault number 3. At 0 Ma, present situation. (1) Active spreading axis; (2) non-active part of the transform faults; (3) fossil spreading axis; (4) at 0.6 Ma, situation of the future spreading axis; (5) boundary between the old and the new crust formed after the spreading jump; (6) isochrons of the age of the oceanic floor and their value (in Ma); (7) fault scarps; (8) seamount; (9) salt slopes; (10) theoretical position of the western salt slope in the reconstruction between 3 and 2 Ma.

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Finally, the geological features and the volcanic and tectonic zonation of the Red Sea at 18°N are closely comparable to those described from the FAMOUS area of the Mid-Atlantic Ridge at 36°30'N: both areas are characterized, with respect to faster spreading centers or to other segments of the Mid-Atlantic Ridge, by the absence of sheet flows, lava lakes, surface hydrothermal activity, and by the presence of relatively high volcanic edifices (200–300 m). They also display a similar morphologic and tectonic zonation (Fig. 2) (ARCYANA, 1977).

DEFINITIONS AND ANALYTICAL TECHNIQUES

Modal analyses were obtained with at least 1000 points counted on 42 thin sections of the basaltic samples plus one on the dredged picrite (Table 1), using internally consistent definitions of phenocrysts and microphenocrysts, based not only on the size of the minerals (microlites $< 200 \ \mu m <$ microphenocrysts $< 500 \ \mu m <$ phenocrysts $< 2 \ mm <$ megacrysts), but also on the texture, shapes and intergrowths of the coexisting phases as described by Bryan (1979) and already used in several petrologic studies (ARCYANA, 1977; Bryan & Moore, 1977; Juteau *et al.*, 1980; Staudigel & Bryan, 1981; Bryan, 1983; Juteau *et al.*, 1983). However, we did not make the distinction between phenocrysts may represent up to 10% of the total amount of plagioclase. Basalts are defined as aphyric, sub-aphyric, moderately porphyritic, and porphyritic when they contain a volume % of total microphenocrysts, phenocrysts of respectively, < 0.5%, 0.5-3.0%, 3.0-10.0%, and > 10.0%.

Microprobe analyses of the glass in all the glassy samples and of the main mineral phases from selected samples were made on the fully automated CAMEBAX microprobes of the University of Nancy or of the Centre IFREMER of Brest. For silicate analysis we generally used the usual procedure that analyses in three successive runs: (1) Fe, Na, Si, and K; (2) Mn, Mg, Al, and Ca; and (3) Cr and Ti. The peak and background counting times were both of six seconds with a filament current of 15 kV. This fast analytical procedure (each analysis lasts less than 2 min) is accurate enough when some analytical controls are made during a single probe session, by analysing reference homogeneous minerals, or from one session to another by using the same references at the beginning of each new session. Thus the accuracy of analysis, for concentrations >0.3%, is better than 0.5-1.0% depending on the element. Below 0.3%, the accuracy is less good.

Whole-rock analyses were done at the Institute of Geochemistry of Irkutsk by X-ray and atomic absorption spectrometry.

Very fresh glasses were hand-picked under a lens from four selected samples (53-2, 59-3, 68-2 and 71-2) for isotopic analyses. The chemical separation methods for Pb, Sr, and Nd have been described respectively by Manhes *et al.* (1978), White & Patchett (1984) and Richard *et al.* (1976). The analytical procedure on the mass spectrometer as well as the standards used are given in White & Dupré (1986).

Twenty-five samples were analysed for trace elements by instrumental neutron activation at the Laboratoire Pierre Süe (Groupe des Sciences de la Terre), Saclay. The samples were placed in cadmium containers and irradiated inside the Osiris reactor (C.E.N., Saclay). The γ -spectrometry was carried out with a Ge(Li) detector, having 1.75 keV of resolution for the 1333 keV ray of ⁶⁰Co. The analytical technique used, described by Chayla *et al.* (1973), is non-destructive. About 20 trace elements can be analysed simultaneously. The precision is about 5% for U, Th, Hf, Ta, Ba, Ce, Sb, Co, Ni, Sc, La, Tb, and between 5 and 10% for all the other elements.

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Summary of the description of the 45 samples collected by the submersible Pisces XI in the axial rift of the Red Sea at 18°N with their location (latitude, longitude and depth; see also Fig. 3) and their morphology

| Dive-Sple | Latitude | Sample loci Longitude | ations Depth | Morphology | Plagio. | Olivine | СРХ | SP | Mode | al analyses Totals | Grou | ndmass | | No | Geoch | emic | al anal | yses |
|-----------|-------------------|--------------------------|-----------------|------------|----------|---------|---------|----|-------|-----------------------|------------------|--------|----------|--------|-------|------|---------|------|
| no. | North | Ĕast | | • • | ph-mph | ph-mph | ph-mph | 51 | ph | ph+mph | % | text. | Vesicles | points | WR | G | T+R | Is |
| 53-1 | 17°56′51 | 40°05′04 | 1550 | crust | 3.2-2.6 | 0.1-0.8 | 0-1.1 | 0 | 3.3 | 7.8 | 92.2 | VFG | 9.4 | 1875 | Y | Y | N | N |
| 53-2 | 17°56′51 | 40°05′04 | 1600 | pillow | 3.0-1.5 | 0-0.4 | 0-1.0 | 0 | 3.0 | 5.6 | 94.4 | VG | 5.4 | 1433 | Ŷ | Ŷ | Ŷ | Ŷ |
| 53-3 | 17°56′51 | 40°05′04 | 1600 | bud | 1.6-1.4 | 0-0.4 | 0-1.0 | 0 | 1.6 | 4-4 | 95 ∙6 | GV | 5.7 | 1393 | Ŷ | Ŷ | Ŷ | Ň |
| 53-4 | 17°56′64 | 40°05′05 | 1530 | pillow | 1.5-5-1 | 0.5-0.5 | 0.2-0.2 | 0 | . 1.9 | 4.4 | 95.6 | VG | 5-2 | 1868 | Ň | Ŷ | Ň | N |
| 57-1 | 17°58′10 | 40°04′05 | 1310 | crust | 6.1-3.1 | 1.5-0.1 | 0-t | t | 7.3 | 10.5 | 89.5 | VG | 3.8 | 1693 | Y | Ŷ | Y | N |
| 59-1 | 17°58'38 | 40°03′05 | 1460 | massive | 0.4-0.2 | 0-0.1 | 0-0.4 | 0 | 0.4 | 1.6 | 98.4 | IM | 3.1 | 1853 | Y | ··N | Ŷ | N |
| 59-2 | 17°58'38 | 40°03′85 | 1470 | crust | 4.2-0.6 | 0-0.3 | 0-0.6 | 0 | 4.5 | 6.0 | 94.0 | VG | 3.2 | 1474 | Ŷ | Y | Ň | N |
| 59-3 | 17°58'38 | 40°03′85 | 1350 | crust | 1.3-1.3 | 0.10.6 | 0-0-8 | 0 | 1.4 | 4.1 | 95.9 | VG | 7.4 | 1758 | Ŷ | Ŷ | Y | Y |
| 59-4 | 17°58'38 | 40°03′95 | 1380 | crust | 0.2-0.9 | 0-t | 0-0.4 | 0 | 0.5 | 1.8 | 98·2 | VG | 4.9 | 1578 | Ŷ | Ŷ | Ň | Ň |
| 61-1 | 17°58′78 | 40°05′51 | 1510 | pillow | 0-t | 0-0 | 0-0 | 0 | 0 | t | 100.0 | VG | 2.8 | 1338 | Ŷ | Ŷ | Ŷ | N |
| 62-1 | 17°55′81 | 40°05′35 | 1600 | pillow | 2.4-1.1 | 0·5-t | 0·6−t | 0 | 3.5 | 4.6 | 95.4 | VG | 2.6 | 1835 | Ŷ | Ŷ | Ŷ | N |
| 62-2 | 17°55′81 | 40°05'35 | 1550 | pillow · | 0.2-1.1 | 0-t | 0-0.4 | 0 | 0.2 | 2.0 | 98·0 | VG | 5.6 | 1940 | Y | Y | N | N |
| 62-3 | 17°55'81 | 40°05′35 | 1500 | crust | 1.1-1.3 | 0.5-0.1 | 0.1-0.5 | 0 | 1.4 | 3.0 | 97.0 | VG | 2.3 | 1781 | Y | Y | Y | Ν |
| 62-4 | 17°56'34 | 40°06′27 | 1400 | pillow | 5.6-1.0 | 00-4 | 0-0.5 | 0 | 5.6 | 7.5 | 92.5 | VG | 6.3 | 1922 | Y | Y | Y | Ν |
| 65-1 | 17°57'53 | 40°04'43 | 1500 | crust | 10.7-3.2 | 0.6-0.5 | 0-t | 0 | 11.3 | 14.7 | 85.3 | VG | 3.5 | 3711 | Y | Y | N | N |
| 65-2 | 17°57′56 | 40°04′40 | 1470 | crust | 9.3-2.1 | 0.1-0.5 | 0-0 | t | 9.4 | 11.7 | 88.3 | VG | 6.1 | 2574 | Y | Y | Y | Ν |
| 65-3 | 17°57′56 | 40°04'40 | 1480 | crust | 3.2-1.4 | 0-0.6 | 00 | 0 | 3.5 | 5.5 | 94.5 | VG | 3.6 | 2097 | Y | Y | N | N |
| 65-4 | 17° <i>57′5</i> 6 | 40°04′40 | 1470 | crust | 12.2-2.2 | 0.2-0.1 | 0-0.2 | 0 | 12.7 | 15-5 | 84.5 | VG | 3.4 | 1842 | Y | Y | Y | N |
| 65-5 | 17°57'60 | 40°04′29 | 1500 | crust | 14.5-0.5 | 1.4-0.8 | 0.1-0 | t | 16.0 | 17.0 | 83·0 | G | 3.4 | 2141 | N | Y | N | N |
| 66-1 | 17°58'75 | 40°04'00 | 1500 | crust | 0.3-0.4 | 0-t | 0-0.8 | 0 | 0.3 | 1.6 | 98 •4 | VG | 5.4 | 1586 | Y | Ň | Y | N |
| 66-2 | 17°58′75 | 40°04′00 | 1500 | ? | | | | | | | | | | | Y | Ν | N | Ν |
| 66-3 | 17°58′95 | 40°04'00 | 1500 | ? | | | | | | | | | | | N | N | N | N |
| | | | | | | | | | | | | | | | | | | |
| 66-4 | 17°58'90 | 40°03′60 | 1410 | pillow | 2.2-1.0 | 1.5-0.3 | 0t | 0 | 3.7 | 5.0 | 95·0 | VG | 4.2 | 1671 | Y | N | Y | N |
| 67-1 | 17°58'75 | 40°03′30 | 1540 | bud | 0.2-1.5 | 0.1-0.2 | 0t | 0 | 0.3 | 2.0 | 98.0 | GV | 5-9 | 1740 | Y | Y | Y | N |
| 68-1 | 17°57'49 | 40°04'84 | 1520 | crust | 1.7-1.0 | 0.1-0.5 | 0-0-2 | 0 | 1.8 | 3-2 | 96-8 | VG | 5-3 | 1760 | Y | Ŷ | N | N |
| 68-2 | 17°57'49 | 40°04'84 | 1520 | pillow | 2.1-0.9 | 2.1-0.3 | 0-t | 0 | 4·2 | 5-4 | 94.6 | VG | 6.3 | 1643 | Y | Y | Y | Ŷ |
| 68-3 | 17°57'35 | 40°05'02 | 1460 | pillow | 1.8-1.1 | 0-0.6 | 0-t | 0 | 1-8 | 3.5 | 96.5 | VG | 4.6 | 1427 | Y | Ŷ | Ŷ | N |
| 70-1 | 17°55'86 | 40°05′77 | 1600 | pillow | 0.2-0.2 | 0-0-1 | 0-0-1 | 0 | 0.5 | 1.2 | 98 .8 | VG | 3.4 | 1559 | Ŷ | Y | N | N |
| 70-2 | 17°55'96 | 40°05'75 | 1670 | pillow | 0.6-1.6 | 0.1-0.5 | 0-0.8 | 0 | 0.7 | 3.3 | 96.7 | VG | 5.8 | 2046 | Ŷ | Ŷ | N | N |
| 70-3 | 17°55'96 | 40°05'75 | 1670 | pillow | 0.1-0.2 | 0-t | 0-0-2 | 0 | 0-1 | 0.8 | 99.2 | VG | 2.9 | 1545 | Ŷ | Y | Y | N |
| 70-4 | 17°55′96 | 40'05'75 | 1670 | pillow | 1.5-1.5 | 0-0-2 | 00-7 | 0 | 1.5 | 3.3 | 96.7 | VG | 5.6 | 1375 | Ŷ | N | Y | N |
| 70-5 | 17°55'96 | 40°05'75 | 1670 | pillow | 2.0-1.3 | 0.3-0.5 | 0-0-2 | 0 | 2.3 | 4·0 | 96.0 | VG | 5.5 | 1876 | Ŷ | Ŷ | Y | N |
| 71-1 | 17°56'75 | 40°05'20 | 1500 | pillow | 2.0-1.9 | 0-0-3 | 0-1.0 | 0 | 2.0 | 5-2 | 94.8 | VG | 6.3 | 1247 | Y | Y | Ŷ | N |
| 71-2 | 17°56'75 | 40°05'20 | 1500 | crust | 1.2-2.0 | 0-0-2 | 0-0.3 | 0 | 1.5 | 3.7 | 96.3 | VFG | 2.2 | 1241 | Y | Y | Ŷ | Ŷ |
| | | | | | | | | | 2.6 | | 05.4 | VEG- | 0.1 | 1644 | V | v | Y | |

| 66-4 | 17°58'90 | 40°03'60 | 1410 | pillow | 2.2-1.0 | 1.5-0.3 | 0t | 0 | 3.7 | 5.0 | 95·0 | VG | 4.2 | 1671 | Y | N | Y | N |
|---------|-----------|----------|--------|------------|----------|----------|---------|-----|-------|-------------|-------|-----|------|------|---|---|---|---|
| 67-1 | 17°58′75 | 40°03′30 | 1540 | bud | 0.2-1.5 | 0.1-0.5 | 0t | 0 | 0.3 | 2.0 | 98·0 | GV | 5.9 | 1740 | Y | Y | Y | N |
| 68-1 | 17°57'49 | 40°04′84 | 1520 | crust | 1.7-1.0 | 0.1-0.5 | 0-0-2 | 0 | 1.8 | 3.2 | 96.8 | VG | 5-3 | 1760 | Y | Y | Ν | N |
| 68-2 | 17°57′49 | 40°04′84 | 1520 | pillow | 2.1-0.9 | 2.1-0.3 | 0-t | 0 | 4·2 | 5-4 | 94.6 | VG | 6.3 | 1643 | Y | Y | Y | Y |
| 68-3 | 17°57'35 | 40°05′02 | 1460 | pillow | 1.8-1.1 | 0-0.6 | 0-t | 0 | 1.8 | 3-5 | 96.5 | VG | 4.6 | 1427 | Y | Υ | Y | N |
| 70-1 | 17°55′86 | 40°05'77 | 1600 | pillow | 0.5-0.5 | 0-0-1 | 0-0-1 | 0 | 0.5 | 1.2 | 98.8 | VG | 3.4 | 1559 | Y | Y | Ν | N |
| 70-2 | 17°55′96 | 40°05′75 | 1670 | pillow | 0.6-1.6 | 0.1-0.5 | 0-0-8 | 0 | 0.7 | 3.3 | 96.7 | VG | 5.8 | 2046 | Y | Y | Ν | N |
| 70-3 | 17°55′96 | 40°05′75 | ·1670 | pillow | 0.1-0.5 | 0t | 0-0.2 | 0 | 0.1 | 0.8 | 99-2 | VG | 2.9 | 1545 | Y | Y | Y | N |
| 70-4 | 17°55′96 | 40'05'75 | 1670 | pillow | 1.2-1.2 | 0-0-2 | 0-0.7 | 0 | 1.2 | 3.3 | 96.7 | VG | 5.6 | 1375 | Y | Ν | Y | N |
| 70-5 | 17°55′96 | 40°05'75 | 1670 | pillow | 2.0-1.3 | 0.3-0.5 | 0-0-2 | 0 | 2.3 | 4 ·0 | 96.0 | VG | 5.5 | 1876 | Y | Y | Y | N |
| 71-1 | 17°56′75 | 40°05'20 | 1500 | pillow | 2.0-1.9 | 0-0-3 | 0-1.0 | 0 | 2.0 | 5.2 | 94.8 | VG | 6.3 | 1247 | Y | Y | Y | N |
| 71-2 | 17°56′75 | 40°05′20 | 1500 | crust | 1.5-2.0 | 0-0-2 | 0-0-3 | 0 | 1.2 | 3.7 | 96.3 | VFG | 2.2 | 1241 | Y | Y | Y | Y |
| 71-3 | 17°56'75 | 40°05'20 | 1500 | crust | 2.2-1-8 | 0-0-1 | 0-0-2 | 0 | 2.5 | 4.6 | 95.4 | VFG | 9-1 | 1644 | Y | Y | Y | N |
| 71-4 | 17°56'75 | 40°05'20 | 1500 | ? | | | | | | | | | | | Y | N | N | N |
| 71-5 | 17°56′75 | 40'05'20 | 1500 | pillow | .1.3-0.8 | 0.4-0.6 | 0-0-2 | 0 | 1.7 | 3.3 | 96.7 | VFG | 2.9 | 1853 | Y | Y | Ν | N |
| 72-1 | 17°56'00 | 40°05′60 | 1630 | pillow | 0.8-1.0 | 0-0-1 | 0-0.7 | 0 | 0.8 | 2.6 | 97.4 | VFG | 4-3 | 1902 | Y | Y | Ν | N |
| 72-2 | 17°56′00 | 40°05′60 | 1630 | massive | 0.4-1.3 | 0-0.5 | 0-0.2 | 0 | 0.4 | 2.1 | 97.9 | IM | 4.3 | 1466 | Y | N | Y | N |
| 72-3 | 17°56′00 | 40°05′60 | 1630 | massive | 1.1-0.7 | 0-t | 0.1-0.7 | 0 | 1.2 | 2.6 | 97.4 | IM | 4.8 | 1891 | Y | Ν | Ν | N |
| 72-4 | 17°56′00 | 40°05′60 | 1630 | massive | 2.9-1.1 | 0·1−t | 0-0.4 | 0 | - 3.0 | 4.5 | 95.5 | I | 16.2 | 2387 | Y | N | Y | N |
| 72-5 | 17°56′00 | 40°05′60 | 1630 | pillow | 1.1-0.6 | 0·2-t | 0t | 0 | 1.3 | 1-9 | 98-1 | VG | 3.8 | 1243 | Y | Y | Ν | N |
| 72-6 | 17°56'00 | 40°05′60 | 1630 | pillow | 0.7-1.7 | 0-0.3 | 0-0-8 | 0 | 0.7 | 3-5 | 96-5 | VG | 2.5 | 1424 | Y | N | Ν | N |
| 72-7 | 17°56'00 | 40°05′60 | 1630 | pillow | 0.6-0.7 | 0-0.1 | 0-0-3 | 0 | 0.6 | 1.7 | 98.3 | VG | 3.7 | 3567 | Y | Y | N | N |
| 72-8 | 17°56′00 | 40°05′60 | 1630 | pillow | 1.6-1.6 | 0.5-0.5 | 0-0-4 | 0 | 1.8 | 3.0 | 97·0 | VG | 3.9 | 1736 | Y | Y | Y | N |
| Dredged | 18°01′00 | 40°02′05 | 1533 · | | | | | | | | | | | | | | | |
| picrite | to | to | to | hypabyssal | 0-t | 12.7-5.8 | 0-0-1 | 0.1 | 12.7 | 21.9* | 78.1* | D | 0.0 | 1492 | Y | N | N | N |
| 23-2b | 1,8°01′03 | 40°02′03 | 1490 | - | | | | | | | | | | | | | | |

Modal analyses: The modes have been obtained by at least 1000 points counted on internally consistent definitions of the crystal distinctions (see the text); ph = phenocrysts; mph = microphenocrysts; CPX = clinopyroxene; SP = spinels; t = traces; text. = groundmass texture: D = hypabyssal, F = fluidal, G = glassy, I = intersertal, M = microlitic, V = variolitic. The vesicle content has been counted apart from the crystal content; *=the groundmass of the alkali picrite also contains biotite, albite, K-feldspar. Geochemical analyses: the four last columns summarize the geochemical analyses: WR = whole-rock; G = glass; TR = trace and rare earth elements; Is = isotopes.

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J.-P. EISSEN ET AL.

PETROLOGY OF THE LAVAS

Almost one ton of basaltic samples was collected from the axial area and its immediate surroundings by the submersible *Pisces XI*. Additionally, several dredges extended the sampling to older parts of the oceanic crust. The present study, however, deals with 42 of the 45 basaltic samples taken by the submersible in the axial area (Fig. 3). It also includes a single dredged sample of picritic dolerite showing a peculiar petrology.

General description of the samples

The samples recovered by submersible are all recent basaltic lavas, presenting the usual morphologies for submarine tubes or pillow lavas (Ballard & Moore, 1977), rimmed by a glassy crust about 1.5 cm thick. A description of the samples is given in the Appendix and is outlined in Table 1. In summary, the recovered samples belong to the following lavas types: 21 pillow lava fragments; 15 lava crusts with one or two glassy rims; 2 glass horns; and 4 massive fragments without glassy rim (pillow cores or massive flows). The glass buds and the lava crusts have usually been sampled *in situ*, whereas the pillow and massive fragments most frequently represent fault-scarp rubble.

Their average vesicularity is 4.2% in volume (Matveenkov *et al.*, 1982) to 4.9% (Table 1): i.e., about twice that of deeper basalts from the Mid-Atlantic Ridge axis (Hékinian *et al.*, 1976; ARCYANA, 1977; Bryan, 1983), or from the East Pacific Rise axis in the RITA area at 21°N (Juteau *et al.*, 1980), extruded at about 2500 m below sea-level, as compared with 1500 m in the present study. This difference accords quite closely with the theoretical curve proposed by Moore (1979), which relates vesicularity of MORB to depth of extrusion, indicating a similar gas content for all these basalts.

Petrography

The basalts present the typical MORB plagioclase \pm olivine \pm clinopyroxene \pm spinel mineralogy with similar crystal shapes and relationships (Table 1). Plagioclase always represents more than 50% of the total phenocryst content, with an average around 80% (Fig. 6). Plagioclase phenocrysts never exceed 2 vol.% (average = 0.9%) in the sub-aphyric basalts whereas they reach 14.5 vol.% in the moderately porphyritic and porphyritic basalts (average = 5.4%). Olivine phenocrysts are present in 20 of the 42 samples, most commonly in the plagioclase + olivine porphyritic to moderately porphyritic basalts. Clinopyroxene appears rarely as phenocrysts, or intimately associated with plagioclase in glomerocrysts. Spinels are scarce, generally associated with plagioclase or olivine phenocrysts or xenocrysts.

The modal analyses (Table 1) lead to the distinction of two main groups of samples:

(1) Moderately porphyritic to porphyritic basalts (MPB) consisting of 16 samples (53-1, 53-2, 57-1, 59-2, 62-1, 62-4, 65-1, 65-2, 65-3, 65-4, 65-5, 66-4, 68-2, 71-1, 71-3 and 72-4), containing $4\cdot5-17$ vol.% of total phenocrysts and microphenocrysts (phenocrysts alone 2–16 vol.%). The crystals are resorbed to sub-euhedral phenocrysts and megacrysts of plagioclase (3–14.7 vol.%), olivine (0.1–2.4 vol.%), scarce clinopyroxene (<1.1 vol.%), and traces of spinel, often associated in large glomerocrysts. The largest glomerocrysts have a texture similar to coarse-grained gabbro and are likely to represent gabbros crystallized at shallow depth as suggested by Bryan (1983) and described by Hékinian *et al.* (1985) and Dixon-Eaby *et al.* (1986). The euhedral to skeletal microphenocrysts are of plagioclase and olivine associated in clusters, sometimes with clinopyroxene. The presence and relative proportions of the different crystals in the MPBs allow us to distinguish 3 sub-groups: (i) 5 porphyritic samples (57-1, 65-1, 65-2, 65-4, and 65-5) with phenocryst abundances plagioclase \gg



FIG. 6. Modal proportions of plagioclase phenocrysts + microphenocrysts vs. total crystal content (in volume %) of the Red Sea basalts. Open circles are plagioclase + olivine porphyritic lavas. Striped circles are plagioclase + olivine porphyritic lavas + microphenocrysts of clinopyroxene. Closed circles are plagioclase + olivine + clinopyroxene porphyritic lavas (data from Eissen (1982) and this study). For comparison, Bryan's diagram (1983) shows the field of the LIL element depleted basalts (the delineated field corresponds to the Red Sea 18°N samples).

olivine > \pm clinopyroxene > \pm spinel. Plagioclase was probably the liquidus phase with clinopyroxene appearing only at a late stage in a few of these samples. The abundance of plagioclase in these basalts and the resorbed shape of some of them (xenocrysts) may be attributable to plagioclase enrichment and/or mixing of magmas of slightly different compositions; (ii) 3 samples (65-3, 66-4, and 68-2) which are moderately plagioclase + olivine porphyritic, with the two last samples having almost equivalent amounts of these two minerals. For all these samples, olivine was probably the liquidus mineral followed closely by plagioclase, but these magmas never reach the 4-phase cotectic; and (iii) the remaining 8 samples are moderately porphyritic samples with phenocryst abundances plagioclase \gg clinopyroxene \ge olivine. Plagioclase was the liquidus phase in these and they all attained the cotectic prior to eruption.

(2) A second group of aphyric to sub-aphyric basalts (SAB) of 26 samples (53-3, 53-4, 59-1, 59-3, 59-4, 61-1, 62-2, 62-3, 66-1, 67-1, 68-1, 68-3, 70-1, 70-2, 70-3, 70-4, 70-5, 71-2, 71-5, 72-1, 72-2, 72-3, 72-5, 72-6, 72-7 and 72-8), contains $0-2\cdot3$ vol.% phenocrysts and always less than $4\cdot5$ vol.% of total microphenocrysts and phenocrysts of euhedral to sub-euhedral plagioclase ($0\cdot6-3\cdot3$ vol.%), clinopyroxene ($0-1\cdot1$ vol.%), and olivine (0-1 vol.%). Their mesostasis contains numerous euhedral microphenocrysts to skeletal microlites of plagioclase and clinopyroxene generally associated in clusters, with few olivines. If we exclude the totally aphyric sample (61-1), the SABs can also be subdivided into 2 sub-groups based on their relative mineralogical proportions: (i) 6 samples (67-1, 68-1, 68-3, 70-5, 71-5, and 72-5) with

phenocryst abundances plagioclase > olivine > clinopyroxene in which plagioclase was probably the liquidus phase. These magmas reached the 4-phase cotectic only shortly before extrusion as clinopyroxene phenocrysts are small and rare; (ii) the remaining 19 samples with phenocryst abundances plagioclase > clinopyroxene \geq olivine. These are very similar to the last sub-group of the MPBs, except for their lower crystal content.

The most calcic plagioclases (An_{91-90}) (Fig. 7, Table 2) and the most magnesian olivines (Fo_{89-87}) correspond to large sub-euhedral phenocrysts or to rounded crystals included in larger ones, and appear in the porphyritic MPBs. They might represent xenocrysts, at least for samples 65-1 and 65-2 where the chemical variations observed correspond to a reverse zoning in the plagioclase (Table 2a). The similarity of xenocryst core compositions to that of the surrounding phenocrysts suggests that they are cognate xenocrysts produced in a



FIG. 7. Histograms of the plagioclase (An %), olivine (Fo %) and clinopyroxene (En %) compositions (Table 2a, b, and c). Large numbers are sample numbers (see Fig. 3 or Table 1 for sample location).

slightly cooler and more crystalline portion of the magma chamber or alternatively that they equilibrated at slightly higher pressures (Bender *et al.*, 1978).

The other plagioclase and olivine phenocrysts and some of the microphenocrysts can be separated into two chemically coherent groups even when they cannot be distinguished morphologically. The first group consists of calcic plagioclase (An_{89-86}) and magnesian olivine (c. Fo₈₅) found dominantly in the MPB (e.g., 57-1, 59-4, 65-1, 65-2 and 65-4) but also in some SAB (e.g. 70-1, 70-3, 70-5 and 71-2). The second group is made of less calcic plagioclase (An_{85-80}) and less magnesian olivine (Fo_{85-81}) found dominantly in SAB (e.g., 53-4, 68-3, 70-1, 70-3, 71-2, 72-5 and 72-7) and some MPB (e.g., 65-1, 65-2 and 65-4). These crystals are mainly euhedral phenocrysts or microphenocrysts; some of those phenocrysts can be subhedral as in samples 62-1, 68-3 and 71-2. The plagioclase crystals may be optically zoned, but their cores are usually homogeneous (Table 2a). However, some normal zoning (e.g., 71-2) and occasional reverse zoning (e.g., 72-7) is observed. Their rims reach An₇₀ (e.g., 59-4) and correspond to the crystals of the mesostasis represented mainly by microlites and skeletal microphenocrysts with compositions more sodic than An_{80} . Olivine phenocrysts and microphenocrysts are usually very homogeneous. The clinopyroxene phenocrysts and microphenocrysts are endiopside but they do not evolve toward augitic compositions as frequently observed in MORB. In three samples (68-3, 71-2 and 72-7), the clinopyroxene phenocrysts evolve from Mg-rich pigeonitic (e.g., Wo₁₁ En₇₄ for sample 72-7) compositions, to more calcic endiopsidic compositions, even within one crystal of sample 71-2 (Fig. 8 and Table 2c). This could be the result of a re-equilibration of a Mg-rich pigeonite or a clinoenstatite from rather high pressure (c. 5-10 kb) to shallower conditions close to the roof of the magma chamber (<1 kb).

The morphology, mutual relationships and chemistry of the different minerals of the basalts of the Red Sea at 18°N generally exhibit a 'normal' evolution from mafic or calcic phenocrysts to more evolved microphenocrysts and microlites in all the lavas analysed. As frequently observed in MORB, the high-temperature minerals seem to be restricted to the most porphyritic lavas. Several of these lavas may be the result of some magma mixing between at least two magmas as indicated by the reverse zoning of the plagioclase, the strong zoning of the clinopyroxene and the zoning of one olivine (sample 68-3; Table 2b). The different components of the mixing were probably closely related magmas, more or less differentiated, inside a small zoned magma reservoir.

Mineral-liquid equilibria

During crystal fractionation, the partition of the elements between the melt and the growing crystals follows the law of mass action, expressed by specific partition coefficients $(K_D \text{ mineral/liquid})$. The calculation of these K_Ds for some major element pairs using the glass composition and its coexisting minerals can provide a guide for gross disequilibrium between phenocryst and their present host glass. We used the following low pressure experimental partition coefficients for (assuming that $X_{\text{Liq}}^{\text{FeO}} = 0.9 \times X_{\text{Liq}}^{\text{FeO}}$):

| Olivine: | $\frac{X_{\text{Pol}}^{\text{FoO}}}{X_{\text{Ol}}^{\text{MgO}}} \times \frac{X_{\text{Liq}}^{\text{MgO}}}{X_{\text{Liq}}^{\text{FoO}}} = 0.30 \text{(Roedder \& Emslie, 1970)}$ |
|----------------|--|
| Clinopyroxene: | $\frac{X_{Cpx}^{FeO}}{X_{Cpx}^{MgO}} \times \frac{X_{Liq}^{MgO}}{X_{Liq}^{FeO}} = 0.23 - 0.27 (Grove \& Bryan, 1983; Irving \& Frey, 1984)$ |
| Plagioclase: | $\frac{X_{Pl}^{Ca}}{X_{Pl}^{Na}} \times \frac{X_{Liq}^{Na}}{X_{Liq}^{Ca}} = 1.2 - 1.4 \text{(Drake, 1976; Grove & Bryan, 1983).}$ |

TABLE 2

Selected electron microprobe analyses of some microphenocrysts and phenocrysts of the Red Sea Rift basalts near 18°N. Analyses performed on the CAMEBAX microprobe of the Centre IFREMER de Brest (analyst J. P. Eissen) and of the University of Nancy (analyst T. Juteau). Arrows link centre and edge of a single crystal. (a) Plagioclase; An = anorthite content in relative % of the plagioclase; * = K-feldspar (orthoclase % ~63·5); $K_D \sim$ partition coefficient of Ca and Na between the plagioclase and its host glass. (b) Olivine; Fo = forsterite content in relative % of the olivine; $K_D =$ partition coefficient of Mg and Fe between the olivine and its host glass. (c) Clinopyroxene; En, Wo, and Fs = respectively enstatite, wollastonite, and ferrosilite content in relative % of the clinopyroxene. $K_D =$ partition coefficient of Mg and Fe between the clinopyroxene and its host glass. (d) Spinels; c = calculated; iron distribution calculated assuming a stoichiometric partitioning of Fe²⁺ and Fe³⁺, Ti occupying octahedral sites (Carmichael, 1967). FeO* = total iron as FeO, Cr/Cr+Al+Fe³⁺, Fe³⁺/Cr+Al+Fe³⁺ and Mg/+Fe²⁺ = cationic ratios

TABLE 2a

| Sample no. | 62-1 centre | 62-1 edge | 62-1 | 62-1 | 65-1 centre | 65-1 → | 65-1 edge | 65-1 | 65-1 | 65-1 | 65-1 | 65-1 centre | 65-1 edge | 65-1 | 65-4 centre | 65-4 → | 65-4 edge |
|--------------------------------|----------------|--------------|--------|-------|----------------|----------------|--------------|--------|--------|-------|-------|----------------|--------------|-------------|----------------|---------------------|--|
| SiO ₂ | 45.67 | 46.04 | 52-18 | 50.09 | 47.07 | 46.79 | 45.67 | 49.79 | 48.51 | 46.51 | 46.75 | 45.18 | 45:08 | /0.11 | 46.04 | 47.09 | AC ((|
| Al ₂ O ₃ | 33.94 | 33.67 | 28.63 | 30.79 | 32.90 | 32.93 | 34.10 | 31.21 | 31.58 | 33.46 | 32.71 | 34.00 | 13.88 | 31.40 | 22.16 | 47.08 | 40.00 |
| FeO | 0.35 | 0.45 | 0.90 | 0.73 | 0.32 | 0.33 | 0.17 | 0.44 | 0.66 | 0.34 | 0.41 | 0.30 | 0.47 | 0.60 | 0.43 | 0.26 | 0.22 |
| MgO | 0.21 | 0.24 | 0.33 | 0.35 | 0.28 | 0.30 | 0.29 | 0.41 | 0.49 | 0.28 | 0.27 | 0.23 | 0.19 | 0.20 | 0.35 | 0.25 | 0.10 |
| CaO | 18-15 | 18.10 | 13.33 | 14.60 | 17.22 | 17-56 | 18.04 | 15.77 | 15.85 | 17.32 | 17.33 | 18.53 | 18.52 | 16.16 | 16.06 | 17.22 | 17.04 |
| Na ₂ O | 1.19 | 1.39 | 3.61 | 2.85 | 1.81 | 1.67 | 1.03 | 2.57 | 2.21 | 1.65 | 1.55 | 0.98 | 1.04 | 2.20 | 1.02 | 1.65 | 1.04 |
| K ₂ O | 0.00 | 0.00 | 0.03 | 0.00 | 0.00 | 0.00 | 0.01 | 0.00 | 0.03 | 0.00 | 0.00 | 0.01 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Total | 99.51 | 99-91 | 99.02 | 99.40 | 99.60 | 99.58 | 99·31 | 100-21 | 99.33 | 99.56 | 99.02 | 99.33 | 99·17 | 99.85 | 98.79 | 99.49 | 100.43 |
| An | 89-4 | 87.8 | 67·0 | 73.9 | 84.0 | 85.3 | 90.6 | 77.3 | 79.7 | 85.3 | 86.1 | 01.7 | 00.9 | 70.6 | 020 | 05.2 | 100.45 |
| KD | 2.35 | 2.00 | 0.57 | 0.79 | 1.26 | 1.39 | 2.32 | 0.81 | 0.95 | 1.39 | 1.48 | 2.50 | 2.36 | 0.93 | 1.21 | 85·3 1·43 | 89·1 2·01 |
| Sample no. | 65-4 | 65-4 | 65-4 | 68-3 | 68-3 | 68-3 centre | 68-3 edge | 68-3 | 68-3 | 68-3 | 68-3 | 68-3 | 68-3 | <i>68-3</i> | 68-3 | <i>68-3</i> | 70-5 |
| SiO ₂ | 46.95 | 45.58 | 49.24 | 47.83 | 47.43 | 50.96 | 51.60 | 47.30 | 51.54 | 48.35 | 48.23 | 47.61 | 46.07 | \$2.20 | 51.24 | 51.50 | 47.69 |
| Al_2O_3 | 33.29 | 33.73 | 31.45 | 33.35 | 33.63 | 31.19 | 30.67 | 33.78 | 30.81 | 33.36 | 32.92 | 33.38 | 33.67 | 30.55 | 21.12 | 20.52 | 4/.08 |
| FeO | 0.62 | 0.58 | 0.62 | 0.49 | 0.46 | 0.69 | 0.68 | 0.39 | 0.74 | 0.36 | 0.64 | 0.61 | 0.65 | 0.75 | 0.76 | 0.72 | 32.82 |
| MgO | 0.37 | 0.00 | 0.07 | 0.18 | 0.16 | 0.13 | 0.16 | 0.18 | 0.18 | 0.25 | 0.20 | 0.16 | 0.15 | 0.26 | 0.16 | 0.27 | 0.35 |
| CaO | 17.02 | 18.22 | 15.75 | 15.83 | 15.58 | 13.71 | 13-34 | 17.06 | 13.46 | 15.39 | 15.28 | 15.38 | 15.88 | 12.42 | 13.00 | 12.06 | 16.60 |
| Na ₂ O | 1.68 | 1.31 | 2.54 | 1.91 | 1.86 | 3.08 | 3.14 | 1.40 | 3.16 | 1.92 | 1.99 | 1.96 | 1.73 | 3.55 | 3.36 | 3.50 | 1.00 |
| K ₂ O | 0.03 | 0.00 | 0.00 | 0.00 | 0.01 | 0.00 | 0.01 | 0.02 | 0.05 | 0.01 | 0.00 | 0.00 | 0.00 | 0.03 | 0.00 | 0.05 | 0.00 |
| Total | 99.95 | 99-36 | 100.04 | 99-59 | 99.13 | 99.75 | 99.60 | 100.15 | 99.96 | 99.63 | 99.26 | 99·12 | 99.05 | 99.95 | 99.64 | 99.71 | 99.87 |
| An | 84.7 | 88.5 | 77.4 | 82.1 | 82.2 | 71.1 | 70.1 | 87.0 | 70-0 | 81.6 | 81.0 | 01.7 | 026 | (50 | 22 OF | <i>(</i> 7 0 | <i>,,,,,</i> ,,,,,,,,,,,,,,,,,,,,,,,,,,,,,,, |
| K _D | 1.38 | 1.90 | 0.85 | 1.89 | 1.91 | 1.02 | 0.97 | 2.78 | · 0·97 | 1.83 | 1.75 | 1.79 | 2.09 | 0.80 | 08.2 | 6/·2 0·85 | 82.2 |

Plagioclase analyses

| Sample no. | 70-5 | 70-5 | 70-5 centre | 70-5 → | 70-5 edge | 70-5 | 70-5 | 70-5 | 70-5 | 70-5 | 70-5 | 70-5 | 70-5 | 70-5 | 70-5 | 70-5 | 71-2 centre |
|--|----------------|----------------|----------------|----------------|----------------|------------------------|------------------------|----------------|------------------------|------------------------|----------------|------------------------|----------------|------------------------|------------------------|------------------------|------------------------|
| $\overline{SiO_2}$ Al ₂ O ₃ | 47·04 32·96 | 47·62 32·70 | 46·91 32·81 | 47·12 32·63 | 47·56 32·46 | 50·25 30·36 0·71 | 50.08 29.96 0.91 | 50·38 30·54 | 50.54 30.12 0.83 | 46.55 32.85 0.40 | 47·31 32·29 | 50.78 29.99 0.72 | 47·22 32·65 | 46·80 33·15 0·52 | 46·85 32·84 0·44 | 45·89 33·84 0·40 | 50.55 30.04 0.50 |
| MeO | 0.47 | 0.23 | 0.42 | 0.48 | 0.26 | 0.30 | 0.39 | 0.29 | 0:33 | 0.30 | 0.24 | 0.27 | 0.25 | 0.20 | 0.18 | 0.22 | 0.31 |
| CaO | 17.21 | 16.79 | 16.79 | 17.05 | 16.63 | 14.47 | 14.88 | 14.41 | 14.67 | 16.92 | 17.53 | 14.35 | 16.85 | 16.78 | 17.34 | 17.66 | 15.20 |
| Na ₂ O | 1.79 | 2.03 | 1.52 | 1.97 | 2.04 | 2.77 | 3.04 | 3.13 | 3.11 | 1.94 | 1.99 | 3.23 | 1.95 | 1.74 | 1.65 | 1.32 | 2.56 |
| K ₂ O | 0.00 | 0.00 | 0.00 | 0.03 | 0.01 | 0.00 | 0.01 | 0.00 | 0.01 | 0.00 | 0.01 | 0.05 | 0.00 | 0.00 | 0.00 | 0.03 | 0.00 |
| Total | 99.71 | 99.57 | <u>98</u> .65 | 99.48 | 99.43 | 98.89 | 99.26 | 99.45 | 99.62 | 98.95 | 99.70 | 99-39 | 99.28 | 99.19 | 99.30 | 99.35 | 99.15 |
| An K | 84·2 | 82·1 | 85·9 | 82·6 | 81.8 | 74·3 | 73·0 | 71·8 | 72·3 | 82·8 | 82·9 1·57 | 70·9 0·79 | 82·7 1·54 | 84·2 1·72 | 85·3 1·87 | 87·9 2·38 | 76·7 1·04 |
| к _D | 1.11 | 1.41 | 1.90 | 1.24 | 1.42 | 0.93 | 0.07 | 0.95 | 0.04 | 1 55 | 1.57 | | 1.54 | 172 | 107 | 2 50 | 104 |
| Sample | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | | |
| no. | edge | | | centre | edge | | | • | | | | centre | edge | | | | |
| SiO, | 50·33 | 51.24 | 50.38 | 51.14 | 50.67 | 50.42 | 47.77 | 47.40 | 51.16 | 47.37 | 47.47 | 46.16 | 46.63 | 50.82 | 45·28 | | |
| Al ₂ Õ ₃ | 29.89 | 29.90 | 29.44 | 30.64 | 29.81 | 30.43 | 32.74 | 32.72 | 30.39 | 32.98 | 33.30 | 34.51 | 33.10 | 30.47 | 33-58 | | |
| FeO | 0.60 | 0.89 | 0.74 | 0.69 | 0.50 | 0.69 | 0.51 | 0.45 | 0.62 | 0.36 | 0.48 | 0.44 | 0.59 | 0.57 | 0.37 | | |
| MgO | 0.26 | 0.31 | 0.31 | 0.29 | 0.26 | 0.30 | 0.24 | 0.19 | 0.25 | 0.19 | 0.24 | 0.24 | 0.25 | 0.31 | 0.20 | | |
| CaO | 15.14 | 14.68 | 15.08 | 14.12 | 14.41 | 14.62 | 16.78 | 16.62 | 13.94 | 10.73 | 10.43 | 1/.5/ | 1.60 | 13.92 | 18.20 | | |
| Na ₂ O | 2.80 | 3.00 | 3.18 | 3.08 | 0.00 | 0.05 | 0.01 | 0.02 | 0.01 | 0.01 | 0.02 | 0.01 | 0.02 | 0.03 | 0.01 | | |
| Total | 003 | 100-11 | 99-19 | 00.00 | 99.07 | 99.58 | 99.86 | 99.21 | 99.53 | 99.59 | 99.66 | 100.06 | 99.14 | 99.36 | 98.93 | | |
| Am | 74.4 | 72.6 | 77.1 | 71.6 | 70.2 | 77.3 | 83.6 | 83.5 | 70.8 | 82.5 | 84.0 | 80.5 | 84.6 | 70.2 | 88.6 | | |
| Kn | 0.93 | 0.84 | 0.83 | 0.80 | 0.75 | 0.83 | 1.62 | 1.62 | 0.77 | 1.49 | 1.67 | 2·70 | 1.75 | 0.75 | 2.47 | | |
| | | | | | | | | | | | | | | | | | |
| Sample | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 72-5 | 72-5 | 72-5 | 23-2b | 23-2b | 23-2b | 23-2b | 23-2b | | | |
| <i>no.</i> | centre | eage | | | | | centre | euge | | | | | | | | | |
| SiO ₂ | 45.53 | 46.32 | 46.44 | 47.02 | 47.33 | 47-21 | 46.84 | 47.23 | 50.58 | 46.26 | 46.02 | 46.23 | 52.45 | 64.47 | | | |
| Al ₂ O ₃ | 33.96 | 32.94 | 33.13 | 33.03 | 32.87 | 32.67 | 32.73 | 32.91 | 30.77 | 34.02 | 34.34 | 33.74 | 29.31 | 20.14 | | | |
| FeO | 0.37 | 0.41 | 0.43 | 0.24 | 0.51 | 0.35 | 0.45 | 0.59 | 0.49 | 0.62 | 0.48 | 0.20 | 0.00 | 0.47 | | | |
| MgO | 0.26 | 0.26 | 0.22 | 0.23 | 0.20 | 0.24 | 0.23 | 0.20 | 0.48 | 0.10 | 16.19 | 17.20 | 11.20 | 0.65 | | | |
| CaO Na O | 1/91 | 1/-24 | 1.49 | 1/*/4 | 10.71 | 1,90 | 1/09 | 10.92 | 14.11 | 10'41 | 1.04 | 1.92 | 11.38 | 3.65 | | | |
| Na ₂ O | 1.23 | 1.11 | 1.48 | 0.00 | 0.00 | 0.01 | 1.14 | 0.00 | 0.00 | 0.07 | 0.14 | 0.00 | 4.70 | 10.62 | | | |
| K ₂ U Tatal | 0.05 | 00.04 | 00.07 | 0.01 | 00.29 | 00.29 | 00.00 | 00.52 | 100.55 | 00.27 | 00.32 | 00.01 | 0.77 | 100.00 | | | |
| Total | 99-26 | 98.84 | 98.97 | 99.81 | 99.38 | 99.28 | 99.08 | 99.52 | 70.0 | 92.27 | 99.33 | 99.21 | 90.11 55.0 | 2.2* | | | |
| An | 88.7 | 84.8 | 86.6 | 86.4 | 84.4 | 83.9 | 84.5 | 84.2 | 72.8 | 82.9 | 81.2 | 84.0 | 22.2 | 3.3* | | | |
| Λ _D | 2.35 | 1.10 | 2.04 | 2.00 | 1.12 | 1.02 | 1.22 | 1.21 | 0.10 | | | _ | | | | | |

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|--------------------------------|---------|---------------|--------|----------------|-----------|----------------|--------|-----------|---|--------|--------|-----------------|----------------|--------|--------|----------|-------|
| Sample no. | 65-4 | 65-4 | 65-4 | 68-3 centre | 68-3 → | 68-3 edge | 68-3 | 68-3 | 68-3 | 68-3 | 70-5 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | |
| SiO ₂ | 39.89 | 39.93 | 39.79 | 40.31 | 39.48 | 39.60 | 39.10 | 39.32 | 39.23 | 39.83 | 39.74 | 39·46 | 38.88 | 38.97 | 38.59 | 39.13 | |
| FeO | 12.95 | 12.23 | 13.14 | i 13∙77 | 16.74 | 17.22 | 18.85 | 17.82 | 18.59 | 17.81 | 15.60 | 17.72 | 17.64 | 16:75 | 16.52 | 18.53 | |
| MnO | 0.13 | 0.50 | 0.33 | 0.36 | 0.22 | 0.20 | 0.28 | 0.23 | 0.13 | 0.29 | 0.22 | 0.24 | 0.35 | 0.10 | 0.33 | 0.35 | |
| MgO . | · 46·70 | 47.50 | 47.27 | 46.34 | 44·14 | 43.65 | 42·19 | 41.71 | 42.75 | 42.96 | 44·25 | 43.20 | 43.49 | 43.30 | 43.73 | 42.63 | |
| CaO | 0.33 | 0.36 | 0.32 | 0.20 | 0.27 | 0.19 | 0.27 | 0.32 | 0.25 | 0.32 | 0.29 | 0.31 | 0.31 | 0.29 | 0.32 | 0.23 | |
| Cr_2O_3 | 0.04 | 0.02 | 0.09 | 0.00 | 0.00 | 0.00 | 0.01 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | |
| NiÒ | 0.23 | 0.31 | 0.06 | 0.00 | 0.10 | 0.13 | 0.01 | 0.16 | 0.04 | 0.00 | 0.00 | 0.06 | 0.04 | 0.10 | 0.11 | 0.06 | |
| Total | 100-28 | 100.55 | 101.00 | 100-98 | 100.95 | 100.99 | 100.74 | 99-56 | 100.99 | 101-21 | 100-10 | 100.99 | 100.71 | 99.51 | 99.60 | 100.93 | |
| Fo | 86.5 | 87.4 | 86.5 | 85.7 | 82.5 | 81.9 | 80.0 | 80.7 | 80.4 | 81-1 | 83.5 | 81-3 | 81.5 | 82.2 | 82.5 | 80.4 | |
| K _D | 0.26 | 0.24 | 0.26 | 0.18 | 0.23 | 0.24 | 0.28 | 0.26 | 0.27 | 0.26 | 0.25 | 0.27 | 0.27 | 0.26 | 0.25 | 0.29 | |
| | | | | | | | | | | | | | | | | | • |
| Sample | 23-2b | 23-2b | 23-2b | 23-2b | 23-2b | 23 - 2b | 23-2b | 23-2b | 23-2b | 23-2b | 23-2b | 23 - 2b | 23 - 2b | 23-2b | 23-2b | 23-2b | 23-2b |
| no. | centre | | → | edge | centre | → | edge | centre | · • • • • • • • • • • • • • • • • • • • | edge | centre | _ _→ | → | edge | centre | <i>→</i> | edge |
| SiO ₂ | 38.41 | 39.66 | 38-26 | 37.42 | 39.74 | 39.00 | 38.78 | 40.41 | 39.80 | 37.65 | 39.44 | 39.67 | 39.42 | 37.32 | 39.00 | 39.88 | 38.62 |
| FeO | 14.40 | 16.25 | 23.35 | 30.85 | 15.17 | 20.06 | 21.03 | 13-33 | 16.13 | 28.43 | 13.43 | 14.63 | 18.83 | 26.03 | 13.28 | 15.77 | 23.22 |
| MnO | 0.00 | 0.19 | 0.19 | 0.12 | 0.69 | 0.06 | 0.43 | 0.23 | 0.00 | 0.48 | 0.22 | 0.00 | 0.39 | 0.52 | 0.49 | 0.15 | 0.30 |
| MgO | 46.30 | 43·23 | 38.41 | 32.17 | 45.22 | 41·27 | 38.01 | 46.52 | 44·11 | 33.85 | 45.39 | 45.72 | 41.80 | 36.62 | 46.89 | 44·34 | 37.03 |
| CaO | 0.24 | 0.36 | 0.42 | 0.46 | 0.23 | 0.49 | 0.55 | 0.50 | 0.46 | 0.35 | 0.27 | 0.31 | 0.40 | 0.22 | 0.21 | 0.44 | 0.32 |
| Cr ₂ O ₃ | 0.00 | 0.00 | 0.00 | 0.03 | 0.00 | 0.00 | 0.00 | 0.00 | 0.14 | 0.00 | 0.00 | 0.00 | 0.00 | · 0.00 | 0.01 | 0.00 | 0.00 |
| NiO | 0.00 | 0.08 | 0.03 | 0.02 | 0.01 | 0.01 | 0.08 | 0.02 | 0.00 | 0.03 | 0.05 | 0.17 | 0.09 | 0.00 | 0.08 | 0.00 | 0.06 |
| Total | 99-35 | 99 •77 | 100.66 | 101.07 | 101-06 | 100-89 | 98-88 | 100.74 | 100.64 | 100-79 | 98-80 | 100-50 | 100-93 | 101.01 | 99·96 | 101-21 | 99.55 |
| Fo | 85-1 | 82.6 | 74·6 | 65·0 | 84·2 | 78 ∙6 | 76.3 | 86.1 | 83·0 | 68·0 | 85.8 | 84·8 | 79 ·8 | 71.5 | 86.3 | 82·0 | 74·0 |
| KD | — | | | _ | — | | | _ | | | — | — | | — | — | | |

| TABLE 2b |
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| TABLE | 2c |
|-------|----|
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STATISTICS TO A STATISTICS

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Clinopyroxene analyses

| Sample no. | 68- <u>3</u> | <u>68-3</u> | 68-3 | 68-3 | 68-3 | 71-2 centre | 71-2 → | 71-2 edge | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 | 71-2 |
|-------------------|-------------------|----------------------|----------------|---------------|--------------|----------------|-----------|--------------|--------|--------|-------|--------|--------|-------|
| SiO ₂ | 52.70 | 53.05 | 51.46 | 51.70 | 51.65 | 53·03 | 52.98 | 51.92 | 52.28 | 51.24 | 51.82 | 51.62 | 52.58 | 51.53 |
| TiO ₂ | 0.48 | 0.44 | 0.66 | 0.56 | 0.69 | 0.29 | 0.31 | 0.35 | 0.47 | 0.47 | 0.42 | 0.48 | 0.38 | 0.45 |
| $Al_2\bar{O}_3$ | 3.36 | 2.78 | 3.99 | 3.27 | 3.37 | 2.92 | 2.85 | 3.15 | 3.43 | 3.43 | 3.08 | 3.11 | 2.85 | 3.68 |
| FeO | 9.30 | 10.56 | 7.84 | 9.64 | 8.45 | 7.75 | 8.10 | 7-62 | 8.33 | 7-47 | 6.48 | 7.82 | 8.07 | 6.52 |
| MnO | 0.29 | 0.36 | 0.07 | 0.20 | 0.07 | 0.38 | 0.21 | 0.29 | 0.22 | 0.21 | 0.26 | 0.28 | 0.18 | 0.23 |
| MgO | 20.68 | 23.98 | 17•34 | 20.83 | 17.17 | 20.11 | 21.00 | 18.99 | 19.78 | 17.71 | 18.06 | 20.25 | 19.93 | 16.92 |
| CaO | 11.86 | 7.43 | 17.10 | 12.36 | 17.51 | 16-13 | 15-33 | 17.46 | 15.02 | 18.34 | 18.61 | 15-42 | 15.75 | 19-11 |
| Na ₂ O | 0.22 | 0.11 | 0.20 | · 0·09 | 0.24 | 0.10 | 0.07 | 0.18 | 0.15 | 0.18 | 0.21 | 0.16 | 0.20 | 0.19 |
| Cr_2O_3 | 0.23 | 0.34 | 0.19 | 0.14 | 0.03 | 0.08 | 0.17 | 0.23 | 0.34 | 0.13 | 0.06 | 0.23 | 0.30 | 0.21 |
| NiO | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.04 | 0.00 | 0.00 | 0.00 | 0.04 | 0.00 |
| Total | 99 ·11 | 99·03 [/] . | 98.83 | 98·79 | 99-19 | 100.82 | ` 101·01 | 100-19 | 100.06 | 99·18 | 98.92 | 99·37 | 100-31 | 98-84 |
| En | 60.0 | 68·0 | 50.9 | 59.3 | 49.7 | 55.7 | 57.4 | 53·0 | 56-1 | 50.4 | 51-5 | 56.7. | 55.7 | 49.3 |
| Wo | 24.8 | 15.2 | 36-2 | 25.3 | 36.5 | 32.2 | 30.2 | 35.1 | 30.7 | 37.6 | 38.2 | 31.1 | 31.7 | 40.1 |
| Fs | 15-2 | 16.8 | 12.9 | 15.4 | 13.7 | 12.1 | 12.4 | 11.9 | 13-3 | 11.9 | 10.4 | 12.3 | 12.7 | 10.7 |
| K _D | 0.28 | 0.27 | 0.58 | 0.28 | 0.30 | 0.26 | 0.26 | 0.27 | 0.28 | 0.28 | 0.24 | 0.26 | 0.27 | 0.26 |
| Sample | 71-2 | 71-2 | 71-2 centre | 71-2 edae | 23-2b | 23-2b | 23-2b | 23-2b | 23-2b | 23-2b | 23-2b | 23-2b | 23-2b | |
| | | | | | | | | | | | | | | - |
| SiO ₂ | 52.28 | 52.37 | 54.09 | 51.34 | 47.99 | 49.19 | 44.15 | 46.13 | 44.50 | 48.45 | 48.29 | 44·25 | 46.59 | |
| TiO ₂ | 0.50 | 0.48 | 0.06 | 0.36 | 1.49 | 1.70 | 3.91 | 3.10 | 2.88 | 1.84 | 1.65 | 4.10 | 2.14 | |
| Al_2O_3 | 3.27 | 3.39 | 0.92 | 3.82 | 5.49 | 5.73 | 9.80 | 7.77 | 9.40 | 6.00 | 5.90 | 9.79 | 8.10 | |
| FeO | 8.90 | 7.90 | 9.56 | 7.09 | 6.09 | 5.09 | 6.74 | 7.26 | 7.45 | 5.79 | 5.85 | 7.59 | 6.08 | |
| MnO | 0.24 | 0.23 | 0.37 | 0.10 | 0.00 | 10.01 | 0.53 | 0.08 | 0.00 | 0.00 | 0-39 | 0.24 | 0.32 | |
| MgO | 20.04 | 20.43 | 20.08 | 17.50 | 14.43 | 14.02 | 11.28 | 12.05 | 11.89 | 14.03 | 13.82 | 11.19 | 12.86 | |
| | 14.20 | 14.09 | /.45 | 19.23 | 23.82 | 22.30 | 22.73 | 22.44 | 22.44 | 22.91 | 22.54 | 22.07 | 23.23 | |
| Na_2O | 0.06 | 0.11 | 0.10 | 0.10 | 0.20 | 0.33 | 0.73 | 0.01 | 0.50 | . 0.25 | 0.37 | 0.00 | 0.33 | |
| NIO | 0.00 | 0.21 | 0.02 | 0.17 | 0.10 | 0.15 | 0.17 | 0.00 | 0.03 | 0.03 | 0.07 | 0.00 | 0.07 | |
| NIO | 0.00 | 0.02 | 0.03 | 0.00 | 0.10 | 0.02 | 0.00 | 0.00 | 0.00 | 0.00 | 0.08 | 0.00 | 0.00 | |
| Total | 100.22 | 99-23 | 98.73 | 99 ∙77 | 99.68 | 98.55 | 100.04 | 99-34 | 99.09 | 99-29 | 98.96 | 100.46 | 100.22 | |
| En | 57.6 | 58.4 | 70.9 | 49.5 | 41.2 | 42.6 | 35.9 | 37.3 | 36.9 | 41.5 | 41-5 | 35-2 | 39.0 | |
| Wo | 28.5 | 29.0 | 14.6 | 39.2 | 49 ·0 | 48.8 | 52-1 | 50.1 | 50-1 | 48-8 | 48.7 | 51.4 | 50.7 | |
| Fs | 13-9 | 12.7 | 14.6 | 11.3 | 9.8 | 8·7 | 12.0 | 12.6 | 13·0 | 9.6 | 9.8 | 13.4 | 10.3 | |
| K _D | 0.29 | 0.26 | 0.24 | 0.27 - | | | | | | _ | _ | | _ | |
| | | | | | | | | | | | | - | | |

| Sample no. | 23-2b | 23-2b | 23-2b | 23-2b | 23-2b | 23-2b | 23-2b |
|--|--------|--------|--------|--------|--------|--------|-------|
| SiO ₂ | 0.04 | 0.00 | 0.08 | 0.00 | 0.02 | 0.07 | 0.14 |
| TiO ₂ | 3.06 | 1.72 | 1.68 | 1.54 | 1.54 | 2.10 | 2.09 |
| $Al_2\tilde{O}_3$ | 23.36 | 29.65 | 27.19 | 28.69 | 28.70 | 28.61 | 27.88 |
| FeO | 36.72 | 31.41 | 31.88 | 30.44 | 29.43 | 31.69 | 31.77 |
| MnÖ | 0.11 | 0.23 | 0.32 | 0.12 | 0.30 | 0.00 | 0.30 |
| MgO | 9.35 | 10.35 | 10.57 | 11.74 | 12.47 | 10.59 | 11.38 |
| Cr ₂ O ₃ | 26.71 | 26.15 | 28.04 | 27.31 | 27.56 | 26.41 | 25.23 |
| Total , | 100.77 | 100.42 | 100-91 | 101.00 | 100.74 | 100.50 | 99.84 |
| Fe ₂ O ₃ | 14.13 | 10.13 | 11.46 | 11.57 | 11.95 | 10.35 | 12.36 |
| FeO | 24.01 | 22.29 | 21.57 | 20.03 | 18.67 | 22.38 | 20.65 |
| $Cr/Cr + Al + Fe_{a}^{3+}$ | 0.43 | 0.37 | 0.41 | 0.39 | 0.39 | 0.38 | 0.38 |
| $\operatorname{Fe_{c}^{3+}/Cr+Al+Fe_{c}^{3+}}$ | 0.18 | 0.12 | 0.14 | 0.14 | 0.14 | 0.12 | 0.15 |
| $Mg/Mg + Fe_c^{2+}$ | 0.41 | 0.45 | 0.47 | 0.51 | 0.54 | 0.46 | 0.50 |

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| Spinel | and | alyse | s |

Olivine crystals are systematically slightly less evolved than their host glass (Table 2b and Fig. 9a) with $K_{\rm D}s$ between 0.23 and 0.29, as expected for liquids which evolved slightly after the start of their crystallization. The olivines at present in equilibrium with the glass should have calculated compositions of Fo_{85.0}, Fo_{80.8}, Fo_{79.6}, or Fo_{78.5} for samples 65-4, 70-5, 71-2, or 68-3 respectively.

Sample 68-3 (and probably also sample 70-3) contain olivine xenocrysts as attested by one of the olivine phenocrysts which is in complete disequilibrium with its host glass ($K_D = 0.18$ for sample 68-3).

The clinopyroxene crystals analysed in ten samples exhibit $K_{\rm D}s$ covering the entire range of experimentally determined values (Table 2c, Fig. 9) as described for some Mid-Atlantic Ridge (MAR) samples near 36°N (Stakes *et al.*, 1985). If it is assumed that $K_{\rm D}s$ around 0.29 are more realistic, it follows that several crystals are close to equilibrium with their host glass whereas the cores of the zoned crystals and some others are less evolved than their host glass. Samples 70-3, 71-2, and 72-7 also contain clinopyroxene microphenocrysts which are in complete disequilibrium with their host glass ($K_{\rm D}s < 0.20$) and can be considered as xenocrysts.

The scattering of the plagioclase plots is greater (Table 2a, Fig. 9). However, the K_D values allow several groups of crystals to be identified:

All the samples contain a group of plagioclase crystals with K_D s of 0.8–1.1, i.e., more evolved than their host glass.

Some samples contain some plagioclase crystals with K_D s between 1·2 and 1·5, e.g., 65-4, 65-1, 70-5 and 71-2) believed to have been in, or near equilibrium with their host glasses.

Several samples exhibit plagioclase phenocrysts slightly less evolved than their host glass with K_D s of 1.6–2.0 and which may have grown along with the olivine phenocrysts described above.

Finally, in the majority of the samples, there are some plagioclase crystals which appear to be in complete disequilibrium with their host glass, with $K_{\rm D}s > 2.0$ (up to 2.7 in sample 71-2). These are probably xenocrysts corresponding to the olivine xenocrysts in samples 68-3 and 70-3, and clinopyroxene xenocrysts of samples 70-3, 71-2, and 72-7.

A similar distribution has been observed in the FAMOUS-Narrowgate area by Stakes *et al.* (1985). These data support the hypothesis for mixing of two magmas belonging to the



FIG. 8. Clinopyroxene compositions plotted in the En%, Wo%, Fs% pyroxene triangle (Table 2c). Leg LV: clinopyroxenes from alkali lavas of DSDP Leg 55 (data from Clague et al., 1980).

same trend or geochemical group. The less evolved of these two could have been the host magma for the xenocrysts, whereas the more evolved one could have been the host magma of the more evolved crystals. Mixing would have resulted in the complex zoning observed in some plagiocrase and clinopyroxene crystals.

The extrusion temperatures of the lavas, calculated from the equilibrium equation between olivine microphenocrysts and their host glass in sample 71-2 (olivine-liquid equilibrium, after Roedder & Emslie, 1970; Leeman & Scheidegger, 1977), are around 1200°C, which is compatible with temperatures calculated for other oceanic tholeiitic basalts (Hékinian *et al.*, 1976; Juteau *et al.*, 1980).

Description of a dredged alkali picrite

One dredged sample (23-2b), dredged just north of the studied area near a small fracture zone (Table 1), is holocrystalline and probably represents an intrusive rock. It contains numerous olivine megacrysts (Table 1), strongly zoned (Fo₈₅₋₆₅; Table 2b), containing small spinel inclusions in their margins. These Ti-rich (TiO₂% = 1.5-3.0; Table 2d) and Fe-rich (Fe³⁺/Al+Cr+Fe³⁺ = 0.12-0.18) spinels are typical of alkali basalts (Clague *et al.*, 1980). The groundmass comprises microphenocrysts of olivine, clinopyroxene, plagioclase (An₈₃₋₈₁; Table 2a), and ilmenite-bearing titanomagnetite, cemented by microlites of biotite, alkali feldspar (Or₆₄) and plagioclase (An_{<55}). The clinopyroxenes are Ca- and Ti-rich



FIG. 9. Composition of the minerals as a function of the degree of evolution of their host glass given by its FeO*/MgO ratio. The range of the calculated partition coefficients (K_D) is given for each sample. For olivine, the dashed line indicates the experimentally determined low pressure K_D ^{Olivine}/_{Liquid} equilibrium (=0.30; Roedder & Emslie, 1970). For clinopyroxene, the dashed lines give the range of experimentally determined low pressure K_D ^{Olivine}/_{Liquid} equilibrium (=0.30; Roedder & Emslie, 1970). For clinopyroxene, the dashed lines give the range of experimentally determined low pressure K_D ^{Clinopyroxene}/_{Liquid} equilibrium (=0.23, Grove & Bryan 1983; or =0.27, Irving & Frey, 1984). For plagioclase, the dashed lines give the variations of the K_D ^{Plagioclase}/_{Liquid} as a function of the evolution of the glass composition. The experimentally determined low pressure K_D ^{Plagioclase}/_{Liquid} equilibrium should be around 1.3 (Drake, 1976; Grove & Bryan, 1983).

(Wo₅₀, TiO₂% = 1.5-4.2; Table 2c), and show a compositional trend similar to those described for other oceanic alkali lavas (Clague *et al.*, 1980). This petrography corresponds to an alkali picrite whose presence in the Red Sea Rift could testify either to an early stage of rifting, or to special magma compositions linked to specific transform fault structures.

GEOCHEMISTRY

- Analyses of major elements in whole-rocks and glasses

The variation diagrams of the bulk-rock oxides (Table 3) and glasses (Table 4) vs. the FeO*/MgO ratio (Fig. 10) show broad linear correlations for the glasses. For the bulk-rock compositions, the correlations are also good except for SiO₂ and Al₂O₃. The observed dispersion mainly involves the less evolved lavas (FeO*/MgO < 1.1), which are also the most porphyritic (Table 1), and is caused by the presence of plagioclase, and to a lesser extent, olivine phenocrysts. This dispersion results probably from crystal accumulation in these lavas, either by gravity settling, flotation or dynamical sorting in the lava tubes (Staudigel & Bryan, 1981), after or just before eruption.

On the basis of major element chemistry of their glasses, three groups of lavas can be distinguished (Fig. 10, Tables 3 and 4):

(1) Less evolved basalts, with FeO*/MgO < 1.22, 0.59 < mg-number < 0.63, MgO wt.% 7.9-8.7 and TiO₂ wt.% 0.75-0.90, including the most porphyritic MPBs (samples 57-1, 65-1, 65-2, 65-3, 65-4, and 65-5) and some less porphyritic samples (59-2, 59-3, and 59-4).

(2) Moderately evolved basalts, with 1.16 < FeO*/MgO < 1.48, 0.53 < mg-number < 0.57, 6.9 < MgO wt.% < 8.4, and $0.84 < \text{TiO}_2$ wt.% < 1.04, including some less-phyric MPBs as well as some of the SABs: samples 62-1, 62-3, 70-1, 70-2, 70-5, 72-1, 72-5, and 72-8.

(3) The most evolved basalts with FeO*/MgO > 1.49, mg-number < 0.54, MgO wt.% < 7.4 and TiO₂ wt.% > 1.02, including the rest of the SABs and some MPBs: samples 53-1, 53-2, 53-3, 53-4, 61-1, 62-2, 62-4, 67-1, 68-1, 68-2, 68-3, 70-3, 71-1, 71-2, 71-3, 71-5, and 72-7.

On the TiO₂ vs. FeO*/MgO diagram (Fig. 10), the glasses generally have a higher TiO₂ content (+0.05% to +0.20%) (Tables 3 and 4) than their host rocks at a constant FeO*/MgO ratio. Although analytical bias cannot be excluded, these variations can be explained by the plagioclase content (Fig. 6) of some of these basalts, which causes a relative 'dilution' of the TiO₂ content at a constant FeO*/MgO ratio. When compared with the composition of the basaltic glasses of the FAMOUS area (Bryan, 1979), the average TiO₂ content of the Red Sea basalts is systematically near half a percent lower at the same FeO*/MgO ratio as has been previously noted for samples near 16°N in the Atlantis Deep area (Chase, 1969). This might be an indication of a higher degree of partial melting of the mantle source.

The general evolution from less to more evolved basalts is very similar to that of the MAR near 36°N in the AMAR and Narrowgate areas (Stakes *et al.*, 1984) which has a comparable spreading rate (1·1 cm/y). The evolved basalts are compositionally close to their counterparts from the FAMOUS area (MAR 36°50'N), whereas the least evolved ones resemble the moderately evolved basalts of the FAMOUS area. In fact, the latter basalts range from distinctly more primitive (MgO-rich) compositions, and several petrographic types from the FAMOUS area (Hékinian *et al.*, 1976; ARCYANA, 1977; Bryan & Moore, 1977) have not been found in the Red Sea at 18°N, e.g., the Highly Plagioclase-Phyric Basalts and the 'Mount Venus type' picritic basalts. Thus the tholeiites extruded at the axis of the Red Sea are generally more evolved than those of the FAMOUS area (MAR 36°50'N) but closely resemble those of the AMAR and Narrowgate areas (MAR 36°N).

The glass compositions are shown within the CMAS tetrahedron using the projections of Walker *et al.* (1979), modified by Grove *et al.* (1982) (Fig. 11). On the two projections used, it appears that crystallization of the Red Sea basalts was mainly controlled by very low pressure crystal fractionation as all the glass compositions coincide with the 1 atmosphere multisaturated cotectic (Walker *et al.*, 1979; Grove *et al.*, 1982).

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FIG. 10. Evolution of the oxide compositions (on a total recalculated at 100%) of the glasses vs. the FeO*/MgO ratio. Glass analyses were made with the CAMEBAX microprobe of the University of Nancy (France) (see Table 4). Symbols used for samples of each dive: closed triangle; Dive 53. Open triangle; Dive 57. Inverted closed triangle; Dive 59. Inverted open triangle; Dive 61. Closed square; Dive 62. Cross; Dive 65. Open diamond; Dive 66. Closed circle; Dive 67. Asterisk; Dive 68. Open square; Dive 70. Open square with NE–SW diagonal; Dive 71. Open square with NW–SE diagonal; Dive 72. Contoured fields are: (1) Analyses of the FAMOUS area glasses (M.A.R. 36°50'N (Bryan, 1979); (2) analyses of the AMAR and Narrowgate areas glasses (M.A.R. 34–35°N) (Stakes *et al.*, 1984); (3) analyses of the Red Sea whole rock near 18°N (this study, see Table 3).

TABLE 3

Major element analyses of 45 whole-rocks of the Red Sea Rift basalts at 18°N. Analyses done at the Institute of Geochemistry of Irkutsk analyst A. Al'mukhamedov). $FeO^* = total$ iron expressed as FeO; mg-number = Mg/(Mg + Fe) with Fe = total iron. Fo% and An% = respectively calculated forsteritic content (for a $K_D = 0.30$) of the olivine and anorthitic content (for a $K_D = 1.3$) of the plagioclase at equilibrium with this magmatic composition

| Sample | | | ······································ | | | ····· | • | | ······ | | | | | ····· | |
|--------------------------------|-------|-------|--|-------|-------|-------|-------|-------|--------------|-------|-------|-------|-------|--------|-------|
| no. | 53-1 | 53-2 | 53-3 | 57-1 | 59-1 | 59-2 | 59-3 | 59-4 | 61-1 | 62-1 | 62-2 | 62-3 | 62-4 | 65-1 | 65-2 |
| SiO, | 51.02 | 50.62 | 51.36 | 50.10 | 50.69 | 49.98 | 50.52 | 50.04 | 50.84 | 51.02 | 51.18 | 51.27 | 51.02 | 49·08 | 49.76 |
| TiO, | 1.14 | 1.15 | 1.09 | 0.84 | 0.84 | 0.88 | 0.88 | 0.94 | 1.30 | 1.09 | 0.89 | 1.04 | 1.14 | 0.75 | 0.78 |
| Al ₂ Õ ₃ | 14.30 | 14.10 | 14.20 | 16.20 | 14.30 | 14.30 | 14.20 | 14.20 | 13.70 | 13.70 | 13.85 | 13.80 | 15.40 | 16.75 | 16.20 |
| Fe ₂ O ₃ | 1.45 | 2.34 | 1.45 | 1.93 | 1.64 | 3.04 | 2.74 | 2.54 | 1.65 | 1.35 | 2.40 | 0.55 | 1.25 | 0.70 | 0.82 |
| FeO | 9.87 | 9.34 | 9.87 | 7.00 | 8.80 | 7.72 | 7.72 | 8∙44 | 9.69 | 9.87 | 9.16 | 10.59 | 9.15 | 8.35 | 7.90 |
| MnO | 0.19 | 0.19 | 0.19 | 0.16 | 0.18 | 0.19 | 0.18 | 0.18 | 0.19 | 0.50 | 0.18 | 0.19 | 0.18 | 0.16 | 0.15 |
| MgO | 7.50 | 7.42 | 7.40 | 7.90 | 8.30 | 8.40 | 8.30 | 8.30 | 7.50 | 7.70 | 7.60 | 7.80 | 7.30 | 8.40 | 8.30 |
| CaO | 11.50 | 12.15 | 11.70 | 13.40 | 12.70 | 12.90 | 12.80 | 12.70 | 11.60 | 12.00 | 12.15 | 11.90 | 11.90 | 13-80 | 13.70 |
| Na ₂ O | 1.88 | 1.91 | 1.91 | 1.71 | 1.82 | 1.82 | 1.92 | 1.85 | 2.11 | 1.88 | 1.88 | 1.86 | 1.80 | 1.85 | 1-59 |
| K,Ō | 0.15 | 0.16 | 0.16 | 0.08 | 0.08 | 0.06 | 0.07 | 0.02 | 0.20 | 0.10 | 0.10 | 0.10 | 0.14 | 0.04 | 0.02 |
| P ₂ O ₅ | 0.12 | 0.10 | 0.12 | 0.04 | 0.04 | 0.04 | 0.04 | 0.04 | 0.16 | 0.06 | 0.06 | 0.06 | 0.09 | 0.00 | 0.06 |
| P1000 | 0.30 | 0.26 | 0.23 | 0.25 | 0.22 | 0.44 | 0.15 | 0.19 | 0.45 | 0.45 | 0.29 | 0.28 | 0.38 | 0.17 | 0.15 |
| Total | 99.42 | 99·74 | 99.68 | 99.61 | 99.61 | 99.77 | 99.52 | 99·47 | 99.39 | 99.42 | 99.74 | 99.44 | 99.75 | 100.05 | 99.49 |
| FeO*/MgO | 1.49 | 1.54 | 1.51 | 1.11 | 1.24 | 1.24 | 1.23 | 1.29 | 1.49 | 1.44 | 1.49 | 1.42 | 1.41 | 1.07 | 1.04 |
| mq-number | 54.5 | 53.6 | 54.1 | 61.7 | 59.0 | 58.9 | 59.2 | 58-0 | 54-5 | 55.3 | 54-5 | 55.6 | 55.9 | 62-5 | 63·1 |
| Fo% | 79.9 | 79-4 | 79.7 | 84·3 | 82.8 | 82.7 | 82.9 | 82.1 | 79.9 | 80.5 | 80.0 | 80.7 | 80-8 | 84.7 | 85.1 |
| An% | 81.5 | 82·0 | 81.5 | 84.9 | 83·4 | 83·6 | 82·7 | 83.1 | 79 ·8 | 82·1 | 82.3 | 82.1 | 82.6 | 84·3 | 86.1 |

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|--------------------------------|-------|-------|-------|-------|-------|-------|----------|-----------|-------|-------|-------|--------|-------|--------|--------------|
| Sample | 65-3 | 65-1 | 65.5 | | 66.2 | | | <u> </u> | | (8.2 | | | | | |
| | | | | | 00-2 | 00-4 | 07-1 | 08-1 | 08-2 | 08-2 | 08-3 | . 70-1 | 70-2 | 70-3 | 70-4 |
| SiO ₂ | 49.92 | 49.96 | 49.96 | 51.70 | 51.26 | 51.45 | 50.99 | 51.26 | 50.68 | 50.98 | 50.93 | 51.03 | 51.28 | \$1.73 | 51.05 |
| TiO | 0.90 | 0.75 | 0.82 | 1.30 | 1.20 | 1.30 | 1.30 | 1.31 | 1.50 | 1.52 | 1.52 | 0.99 | 1.12 | 1.12 | 0.80 |
| Al ₂ O ₃ | 15.40 | 16.10 | 15.00 | 14.30 | 13.40 | 14.25 | 14.30 | 14.10 | 14.40 | 14.30 | 14.20 | 14.00 | 14.25 | 14.10 | 13.70 |
| Fe ₂ O ₃ | 0.56 | 0.56 | 0.65 | 0.63 | 2.86 | 0.21 | 0.56 | 2.36 . | 0.95 | 1.75 | 1.95 | 2.32 | 1.83 | 1.15 | 2.22 |
| FeO | 8.51 | 7.97 | 8.61 | 10.34 | 8.62 | 10.99 | 10.77 | 9.70 | 10.77 | 9.78 | 9.78 | 8.98 | 8.00 | 0.78 | 8.80 |
| MnO | 0.17 | 0.16 | 0.16 | 0.17 | 0.17 | 0.19 | 0.19 | 0.17 | 0.19 | 0.18 | 0.18 | 0.17 | 0.18 | 0.10 | 0.07 |
| MgO | 8.60 | 8.10 | 8.70 | 7.40 | 7.60 | 7.20 | 7.30 | 6.90 | 7.10 | 7.20 | 7.20 | 7.80 | 7.90 | 7.80 | 7.50 |
| CaO | 12.90 | 12.80 | 13.10 | 11.00 | 11.90 | 11.20 | 11.10 | 11.30 | 10.80 | 10.90 | 10.90 | 12.10 | 11.80 | 11.30 | 12.10 |
| Na ₂ O | 1.77 | 1.76 | 1.56 | 1.96 | 2.22 | 2.21 | 2.19 | 2.12 | 2.28 | 2.22 | 2.31 | 1.93 | 1.99 | 1.00 | 1.05 |
| K ₂ O | 0.05 | 0.05 | 0.05 | 0.13 | 0.19 | 0.12 | 0.13 | 0.19 | 0.18 | 0.18 | 0.19 | 0.11 | 0.12 | 0.10 | 0.11 |
| P_2O_5 | 0.00 | 0.00 | 0.00 | 0.04 | 0.08 | 0.03 | 0.04 | 0.14 | 0.08 | 0.08 | 0.08 | 0.07 | 0.03 | 0.02 | 0.05 |
| P1000 | 0.13 | 0.16 | 0.21 | 0.43 | 0.33 | 0.26 | 0.22 | 0.34 | 0.39 | 0.36 | 0.33 | 0.24 | 0.26 | 0.20 | 0.03 |
| Total | 98-91 | 98·37 | 98.82 | 99.40 | 99.83 | 99·41 | 99:09 | 99-89 | 99·32 | 99.45 | 99-57 | 99.74 | 99.75 | 99.48 | 99.87 |
| FeO*/MgO | 1.05 | 1.05 | 1.06 | 1.47 | 1.47 | 1.55 | 1.54 | 1.71 | 1.64 | 1.58 | 1.60 | 1.42 | 1.35 | 1.30 | 1.46 |
| mg-number | 63·0 | 63·0 | 62.8 | 54.7 | 54.8 | 53.4 | 53.6 | 51.0 | 52.1 | 53.1 | 52.7 | 55.7 | 57.0 | 56.2 | 540 |
| Fo% | 85.0 | 85.0 | 84.9 | 80.1 | 80.1 | 79.3 | 79.4 | 77.6 | 78.4 | 79.0 | 78.8 | 80.7 | 81.5 | 81.1 | 24-9 80.7 |
| An% | 84·0 | 83.9 | 85.8 | 80.1 | 79.4 | 78.5 | 78.5 | 79.3 | 77.3 | 77.9 | 77.2 | 81.8 | 81.0 | 80.3 | 81.7 |

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| Sample no. | 70-5 | 71-1 | 71-2 | 71-3 | 71-4 | 71-5 | 72-1 | 72-2 | 72-3 | 72-4 | 72-5 | 72-6 | 72-7 | 72-8 | 23-2b |
|--------------------------------|-------|-------|--------------|-------|--------|---------------|-------|--------|-------|-------|-------|-------|-------|-------|--------------|
| SiO ₂ | 51.41 | 51.36 | 50.88 | 51.08 | 51.77 | 51-29 | 51.70 | 51.12 | 51-99 | 51.45 | 51.67 | 51.74 | 51.49 | 50.95 | 45.42 |
| TiO ₂ | 0.99 | 1.20 | 1.10 | 1.20 | 1.10 | 1.25 | 1.04 | 0.99 | 1.04 | 1.05 | 1.14 | 1.05 | 1.25 | 0.99 | 2.12 |
| $Al_2\bar{O_3}$ | 14.00 | 14.10 | 14.20 | 14.10 | 14.20 | 14.20 | 14.10 | 14.55 | 13.70 | 13.90 | 13.80 | 14.00 | 13.70 | 13.90 | 11.10 |
| Fe ₂ O ₃ | 2.09 | 1.05 | 2.39 | 1.35 | 3.52 | 2.34 | 1.84 | 2.09 | 1.35 | 2.36 | 1.15 | 1.54 | 2.64 | 2.40 | 3.34 |
| FeO | 9.52 | 9.87 | 9.34 | 9.51 | 7.90 | 6.82 | 8.80 | 9.61 | 9.33 | 8.98 | 9.69 | 9.34 | 8.44 | 9.16 | 7.54 |
| MnO | 0.18 | 0.18 | 0.20 | 0.18 | 0.18 | 0.18 | 0.19 | 0.20 | 0.18 | 0.19 | 0.19 | 0.19 | 0.19 | 0.18 | 0.17 |
| MgO | 7.62 | 7.40 | 7.29 | 7-50 | 7.22 | 7.30 | 7.70 | 7.50 | 7.60 | 7.50 | 7.60 | 7.50 | 7.60 | 7.62 | ` 13.10 |
| CaO | 11.70 | 11.60 | 11.72 | 11.90 | 12.22 | <u>`11∙70</u> | 11.80 | 11.80 | 11.90 | 11.60 | 11.70 | 11.95 | 11-80 | 12.10 | 11.40 |
| Na ₂ O | 1.97 | 1.98 | 1.97 | 2.03 | 2.02 | 1.99 | 1.97 | 2.06 | 2.02 | 2.09 | 2.02 | 2.09 | 1.98 | 2.02 | 2.73 |
| K₂Ō | 0.11 | 0.16 | 0.20 | 0.17 | 0.17 | 0.15 | 0.11 | 0.13 | 0.16 | 0.14 | 0.12 | 0.13 | 0.12 | 0.12 | 1.05 |
| P_2O_5 | 0.09 | 0.14 | 0.09 | 0.14 | 0.08 | 0.14 | 0.12 | 0.06 | 0.12 | 0.08 | 0.06 | 0.02 | 0.06 | 0.08 | 0.40 |
| P1000 | 0.18 | 0.33 | 0.28 | 0.33 | 0.34 | 0.22 | 0.23 | 0.21 | 0.22 | 0.17 | 0.22 | 0.18 | 0.19 | 0.26 | 0.00 |
| Total | 99.86 | 99.37 | 99.66 | 99.49 | 100.72 | 97.58 | 99.60 | 100-32 | 99.61 | 99-51 | 99.36 | 99.78 | 99.46 | 99.78 | 98·37 |
| FeO*/MgO | 1.50 | 1.46 | 1.58 | 1.43 | 1.53 | 1.22 | 1.36 | 1.53 | 1.39 | 1.48 | 1.41 | 1.43 | 1.42 | 1.49 | 0.81 |
| mg-number | 54.4 | 54.9 | 53-1 | 55.5 | 53.8 | 59.3 | 56.8 | 53.8 | 56-2 | 54.6 | 55.8 | 55.5 | 55.6 | 54.5 | 68.9 |
| Fo% | 79.9 | 80.3 | 79 ·0 | 80.6 | 79.5 | 82.9 | 81.4 | 79•5 | 81.1 | 80.1 | 80.8 | 80.6 | 80.7 | 80.0 | 88-1 |
| An% | 81·0 | 80.8 | 81·0 | 80.8 | 81.3 | 80.9 | 81-1 | 80.5 | 80.9 | 80.0 | 80.6 | 80.4 | 81.1 | 81-1 | 75 ∙0 |

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Major element analyses of 36 glass rims of the Red Sea Rift basalts near 18°N. Analyses performed on the CAMEBAX microprobe of the University of Nancy (analysts E. Humler and J. P. Eissen). $FeO^* = total$ iron expressed as FeO; mg-number = Mg/(Mg + Fe) with Fe = total iron. Fo% and An% = respectively calculated forsteritic content (for a $K_D = 0.30$) of the olivine and anorthitic content (for a $K_D = 1.3$) of the plagioclase at equilibrium with this magmatic composition

| Sample no. | 53-1 | 53-2 | 53-3 | 53-4 | 57-1 | 59-2 | 59-3 | 59-4 | 61-1 | 62-1 | 62-2 | 62-3 | 62-4 | 65-1 | 65-2 | 65-3 | 65-4 | 65-5 |
|-------------------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|------------------|------------------|-------|--------|-------|-------|-------|-------|
| SiO | 51.87 | 51.84 | 51.99 | 52.13 | 50.79 | 51.36 | 50.96 | 51.31 | 52.37 | 51.91 | 52·25 | 52.40 | 52.43 | 50.95 | 51.02 | 51.19 | 51.06 | 51.15 |
| TiO ₂ | 1.09 | 1.05 | 1.16 | 1.01 | 0.75 | 0.77 | 0.77 | 0.77 | 1.12 | 1.02 | 0.96 | 0.84 | 1.03 | 0.65 | 0.70 | 0.69 | 0.67 | 0.67 |
| AL-O- | 14.18 | 14.26 | 14.19 | 14.10 | 14.41 | 14.46 | 14.37 | 14.43 | 14.12 | 14.29 | 14.42 | 14.17 | 14.16 | 15.09 | 14.47 | 14.72 | 14.79 | 14.82 |
| FeO* | 11.10 | 10.97 | 11.02 | 10.87 | 9.06 | 9.41 | 9.97 | 9.80 | 11.02 | 10.24 | 11.25 | 10.84 | 10.58 | 9.35 | 9.05 | 9.27 | 8.91 | 9.02 |
| MnO | 0.09 | 0.17 | 0.28 | 0.15 | 0.17 | 0.36 | 0.20 | 0.17 | 0.30 | 0.28 | 0.06 | 0.03 | 0.30 | 0.22 | 0.28 | 0.21 | 0.10 | 0.16 |
| MgO | 6.93 | 7.23 | 6.76 | 7.17 | 8.48 | 8.05 | 8.15 | 8.01 | 7.06 | 7.60 | 7.20 | 7.35 | 7.08 | 8.78 | 8.61 | 8.43 | 8.47 | 8.67 |
| CaO | 12.11 | 11.82 | 11.60 | 11.43 | 13.50 | 12.46 | 12.71 | 12.41 | 11.46 | 11.90 | 11.47 | 11.77 | 12.10 | 13.30 | 13.56 | 13-20 | 13.30 | 13.32 |
| Na ₂ O | 2.16 | 2.21 | 2.22 | 2.13 | 1.90 | 2.22 | 1.94 | 2.01 | 2.18 | 1.83 | 2.09 | 2.20 | 2.04 | 1.76 | 1.78 | 1.87 | 1.82 | 1.81 |
| K.O | 0.13 | 0.13 | 0.18 | 0.50 | 0.02 | 0.04 | 0.02 | 0.00 | 0.19 | 0.18 | 0.07 | 0.12 | 0.08 | 0.02 | 0.03 | 0.09 | 0.02 | 0.02 |
| Cr.O. | 0.04 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.03 | 0.00 | 0.00 | 0.00 | 0.00 | 0.03 | 0.02 | 0.07 | 0.00 | 0.05 | 0.07 |
| NiO | 0.02 | 0.01 | 0.14 | 0.03 | 0.05 | 0.00 | 0.07 | 0.04 | 0.00 | 0.04 | 0.00 | 0.07 | 0.00 | 0.00 | 0.03 | 0.00 | 0.06 | 0.01 |
| Total | 99.72 | 99.69 | 99.54 | 99.52 | 99.13 | 99.13 | 99.16 | 98-98 | 99.82 | 99-29 | 99·77 | 99.79 | 99-83 | 100.14 | 99.60 | 99.67 | 99.28 | 99·72 |
| FeO*/MgO | 1.60 | 1.52 | 1.63 | 1.52 | 1.07 | 1.17 | 1.22 | 1.22 | 1.56 | 1.35 | 1.56 | 1.47 | 1.49 | 1.06 | 1.05 | 1.10 | 1.05 | 1.04 |
| ma-number | 52.7 | 54·0 | 52.2 | 54.0 | 62.5 | 60.4 | 59.3 | 59.3 | 53.3 | 56.9 | 53-3 | 54.7 | 54.4 | 62.6 | 62.9 | 61.8 | 62.9 | 63.1 |
| Fo% | 78.8 | 79.7 | 78.5 | 79·7 | 84.8 | 83.6 | 82.9 | 82.9 | 79·2 | 81.5 | 79·2 | 80.1 | 79.9 | 84.8 | 85.0 | 84.4 | 85.0 | 85.1 |
| An% | 80.1 | 79.3 | 79·0 | 79·4 | 83.6 | 80.1 | 82.5 | 81.6 | 79-1 | 82.4 | 79 ·8 | 79 •4 | 81.0 | 84.4 | 84.6 | 83.5 | 84·0 | 84.1 |

| Sample no. | 66-I | 66-4 | 67-1 | 68-1 | <i>68-2</i> | <i>68-3</i> | 70-1 | 70-2 | 70-3 | 70-5 | 71-1 | 71-2 | 71-3 | 71-5 | 72-1 | 72-5 | 72-7 | 72-8 |
|--------------------------------|-------------------|-------------------|--------------|-------|-------------|-------------|-------|-------|-------|-------|-------|-------|-------------------|--------------|--------|--------|--------------|-------|
| SiO ₂ | 51.33 | 51.04 | 51.88 | 52.42 | 51.97 | 51.85 | 52.46 | 51.66 | 52·52 | 51.85 | 52·30 | 52.01 | 51·90 | 52.14 | 52.05 | 52.09 | 52·29 | 51.91 |
| TiO, | 1.11 | 1.17 | 1.16 | 1.25 | 1.29 | 1.32 | 0.91 | 0.93 | 0.96 | 0.94 | 1.15 | 1.08 | 1.04 | 1.01 | 0.81 | 0.90 | 1.08 | 1.03 |
| Al ₂ Õ ₃ | 14.41 | 14.30 | 14.02 | 13.93 | 14.11 | 14.52 | 14.60 | 14.11 | 14.62 | 14.04 | 13.88 | 14.08 | 13.91 | 14.04 | 14-25 | 14.30 | 13.96 | 14.15 |
| FeO* | 9.90 | 11.38 | 11.06 | 11.42 | 11.58 | 11.52 | 9.69 | 10.47 | 10.77 | 10.58 | 11-11 | 10.83 | 10.92 | 10.87 | 10.27 | 10.97 | 10.93 | 10.72 |
| MnO | 0.24 | 0.24 | 0.09 | 0.13 | 0.20 | 0.14 | 0.10 | 0.19 | 0.25 | 0.20 | 0.17 | 0.23 | 0.16 | 0.24 | 0.24 | 0.13 | 0.17 | 0.14 |
| MgO | 7.51 | 6.84 | 7.17 | 6.75 | 6.98 | 7.08 | 6.83 | 7.17 | 6.16 | 7.48 | 6.81 | 7.12 | 7.15 | 7.22 | 7.54 | 8.02 | 7.23 | 7.51 |
| CaO | 11.78 | 11.19 | 11.57 | 11.27 | 11.04 | 10.52 | 12.39 | 11.95 | 11.00 | 11.81 | 11.33 | 11.85 | 12.17 | 11.56 | 12.19 | 12-26 | 11.45 | 11-41 |
| Na ₂ O | 2.09 | 2.53 | 2.19 | 2.27 | 2.32 | 2.40 | 2.39 | 1.96 | 2.52 | 2.10 | 2.19 | 2.07 | 2.09 | 2.17 | 1.79 | 1.93 | 2.08 | 2.10 |
| K ₂ Ô | 0.16 | 0.08 | 0.13 | 0.15 | 0.17 | 0.20 | 0.09 | 0.09 | 0.15 | 0.06 | 0.17 | 0.12 | 0.16 | 0.16 | 0.06 | 0.03 | 0.13 | 0.11 |
| Cr,O3 | 0.03 | 0.01 | 0.00 | 0.00 | 0.02 | 0.00 | 0.02 | 0.00 | 0.07 | 0.00 | 0.00 | 0.00 | 0.00 | 0.02 | 0.00 | 0.07 | 0.00 | 0.00 |
| NiÔ | 0.09 | 0.00 | 0.03 | 0.05 | 0.01 | 0.00 | 0.00 | 0.05 | 0.00 | 0.00 | 0.06 | 0.00 | 0.05 | 0.00 | 0.01 | 0.00 | 0.06 | 0.37 |
| Total [,] | 9 8·65 | 98 •78 | 99.30 | 99.64 | 99.69 | 99.55 | 99•48 | 98.58 | 99.02 | 99.06 | 99-17 | 99.39 | 99 ∙52 | 99.43 | 99·21 | 100.70 | 99.38 | 99.45 |
| FeO*/MgO | 1.32 | 1.66 | 1.54 | 1.69 | 1.66 | 1.63 | 1.42 | 1.46 | 1.75 | 1.41 | 1.63 | 1.52 | 1.53 | 1.51 | 1.36 | 1.37 | 1.51 | 1.43 |
| mg-number | 57.5 | 51.7 | 53.6 | 51.3 | 51.8 | 52.3 | 55.7 | 55.0 | 50.5 | 55.8 | 52.2 | 54.0 | 53.9 | 54.2 | 56.7 | 56.6 | 54.1 | 55.5 |
| Fo% | 81.8 | 78.1 | 79.4 | 77.8 | 78·2 | 78.5 | 80.7 | 80-3 | 77.3 | 80.8 | 78.5 | 79.6 | 79.5 | 79 ·8 | 81.3 | 81.3 | 79.7 | 80.6 |
| An% | 80.2 | 76-1 | 79 ·1 | 78·1 | 77.4 | 75.9 | 78.8 | 81.4 | 75.8 | 80.2 | 78-8 | 80.4 | 80.7 | 79.3 | . 83.0 | 82·0 | 7 9·8 | 79·6 |

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The holocrystalline alkali picrite cannot be compared with these tholeiitic lavas, since it was dredged in a fracture zone. However, it could represent an early product of the rifting of the Red Sea, showing the effect of some contamination by the continental crust.

Isotope geochemistry

The isotopic ratios for Nd and Sr (Table 5) show rather small variations (⁸⁷Sr/⁸⁶Sr: 0.702861-0.702914 and ¹⁴³Nd/¹⁴⁴Nd: 0.512990-0.513056). Variations are slightly more



FIG. 11. Compositions of the Red Sea basaltic glasses at 18°N reported in two projections from the CMAS tetrahedron. The projection scheme calculation is described by Grove et al. (1982). The 1 atmosphere cotectics and reaction curves for North Atlantic MORB were determined experimentally by Walker et al. (1979) and Grove et al. (1982). Glass compositions are projected from: (a) plagioclase onto the olivine (OI)-clinopyroxene (Cpx)-silica (Qz) plane; (b) clinopyroxene onto the olivine (OI)-plagioclase (PI)-silica (Qz) plane. On the enlarged triangles, the symbols used are the same as for Fig. 10.

TABLE 5

Isotope data obtained on 4 hand-picked glasses of the Red Sea basalts at $18^{\circ}N$, at the Max Planck Institut für Chemie (analyst B. Dupré). The analytical procedure, on the mass spectrometer as well as the standards used are given by White & Dupré (1986)

| Sample no. | 53-2 | 59-3 | 68-2 | 71-2 |
|--|--|--|--|---|
| ²⁰⁶ Pb/ ²⁰⁴ Pb ²⁰⁷ Pb/ ²⁰⁴ Pb ²⁰⁸ Pb/ ²⁰⁴ Pb ¹⁴³ Nd/ ¹⁴⁴ Nd ⁸⁷ Sr/ ⁸⁶ Sr | 19.116 15.590 38.857 0.512990±10 0.702906±23 | $18.78715.55538.5750.513056 \pm 360.702861 \pm 20$ | 19.141 15.577 38.876 0.513020±11 0.702914±19 | $ \begin{array}{r} 19.163 \\ 15.588 \\ 38.895 \\ 0.513014 \pm 20 \\ 0.702906 \pm 22 \end{array} $ |

important for ²⁰⁶Pb/²⁰⁴Pb, with three samples showing ratios between 19·116 and 19·163, and sample 59-3 giving 18·787. These data suggest a relatively homogeneous mantle source for the sample suite, although some of the groups described above have not been analysed. The results show similarity to isotopic data from Iceland and the FAMOUS area, i.e., they are slightly more radiogenic than normal 'depleted' MORB, and plot in the isotopic domains of the MAR and East Pacific Rise (EPR) for their ⁸⁷Sr/⁸⁶Sr,¹⁴³Nd/¹⁴⁴Nd, and ²⁰⁶Pb/²⁰⁴Pb ratios (Fig. 12) and also for ²⁰⁶Pb/²⁰⁴Pb, ²⁰⁷Pb/²⁰⁴Pb, and ²⁰⁸Pb/²⁰⁴Pb (diagrams not shown).

These data permit the following statements:

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(1) Although the Red Sea basalts at 18°N erupted from a spreading centre at an early opening stage, their isotopic values are not distinct from those of other MORB. At this stage of opening there appears to have been no interaction with continental crust. Only sampling perpendicular to the spreading-axis over a wider area would provide an adequate test as to whether contamination with continental crust had occurred during the earlier stages of opening.

(2) Morgan (1971) proposed that the opening of a spreading centre would result in a hotspot track. Without giving any proof, those results are not inconsistent with this model. In fact, isotopic values close to 19 for ²⁰⁶Pb/²⁰⁴Pb and 0.7030 for ⁸⁷Sr/⁸⁶Sr are often associated with hot-spots (Iceland, Azores, Afar: Schilling, 1973; Sun *et al.*, 1975; Dupré & Allègre, 1980; Hamelin *et al.*, 1984).

(3) In the Indian Ocean, the three spreading ridges converging toward the Rodriguez triple junction have high 208 Pb/ 204 Pb and Sr ratios, compared with those observed in the Atlantic or in the Pacific (Dupré & Allègre, 1980; Hamelin & Allègre, 1985; Hamelin *et al.*, 1986), and delineate a regional mantle province. However, this anomaly could only be distinguished when 206 Pb/ 204 Pb ratios are < 18. Since this ratio for our sample suite is > 18, we cannot know if the Indian Ocean anomaly extends towards the Red Sea area.

Trace element and REE geochemistry

To interpret the data (Table 6) and identify the effects of the main factors which control the genesis and differentiation of these magmas, we shall compare the geochemical behaviour of the different families of trace elements, and especially the properties of the so-called 'hygromagmatophile' elements (HYG) (Joron *et al.*, 1978; Allègre & Minster, 1978), since these highly incompatible elements are weakly or not fractionated during crystal fractionation (Treuil & Varet, 1973; Treuil & Joron, 1975; Treuil *et al.*, 1979). The evaluation of the respective effects of fractional crystallization and melting processes in the analysed samples



FIG. 12. Plots of: (a) ⁸⁷Sr/⁸⁶Sr vs. ²⁰⁶Pb/²⁰⁴Pb; and (b) ¹⁴³Nd/¹⁴⁴Nd vs. ⁸⁷Sr/⁸⁶Sr. These diagrams show the relative homogeneity of the isotopic ratios of the Red Sea basalts near 18°N, compared with those of other oceanic floor basalts (data from Dupré & Allègre, 1980; Dupré et al., 1981) with, however, a slight difference in ²⁰⁶Pb/²⁰⁴Pb for sample 59-3 relative to the others. IR = Indian Ridge; CIR = Central Indian Ridge; MAR = Mid-Atlantic Ridge; EPR = East Pacific Rise; triangle = 53-2; inverted triangle = 59-3; asterisk = 68-2; source = 71-2.

can be made by examining the different behaviour of transition elements and hygromagmatophile elements. The basic principles of the method have been set out in several previous publications (Joron & Treuil, 1977; Joron *et al.*, 1978, 1982).

Primary magma compositions—heterogeneity of the mantle sources

An estimation of the composition of the primary magma can be obtained using the HYG elements, as the slope of the correlation lines gives the ratio of the concentration of the elements in the mantle source (Joron *et al.*, 1978). This method is, however, restricted to the less differentiated samples. In the La vs. Ta diagram (Fig. 13), all the samples follow the same correlation, with a ratio La/Ta close to 10. This ratio is characteristic of one of the mantle domains defined by Joron *et al.* (1978), Bougault *et al.* (1979), and Bougault & Treuil (1980) in the North Atlantic and Pacific oceans. This domain has a chondritic distribution of the

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|---------------|--------------|------|--------|------|------|-----------|-----------|----------|--------|-------|------|------------|
| Sample no. | 53-2 | 53-3 | 57-1 | 59-1 | 59-3 | 61-1 | 62-1 | 62-3 | 62-4 | 65-2 | 65-4 | 66-1 |
| Sc | 41.7 | 41.5 | 37.2 | 43.1 | 41.8 | 41.0 | 40.4 | 41.5 | 40.0 | 36.5 | 38.4 | 40.8 |
| Cr | 105- | 104 | 404 | 83. | 91· | 85. | 164 | 30. | 162. | 394. | 398. | 40. |
| Co | 49.6 | 49.2 | 43·0 | 52.2 | 51.1 | 49.7 | 48·2 | 50.8 | 47.3 | 43.9 | 46.5 | 48.4 |
| Ni | 71. | 70· | 119. | 89. | 87. | 67. | 77. | 69. | 73. | 127. | 124 | 72. |
| Zn | 100 | 102. | | | | 100. | 126. | | 85. | | | · - |
| Rb | 3.2 | 2.5 | 1.4 | | 1.5 | 3.5- | 2.5 | 2.0 | 3.7 | | 0.8 | 4.0 |
| Sr | 148 ∙ | 136. | | 123. | 124 | 152. | 117. | 137. | 151 | | 122. | 156 |
| Zr | 50.0 | 54·0 | 42·0 | 44·0 | 31.0 | 51·0 | 49·0 | 48·0 | 68.0 | 37.0 | 28.0 | 56.0 |
| Sb | 0.03 | 0.02 | 0.04 | 0.08 | 0.16 | 0.08 | 0.05 | 0.14 | 0.04 | 0.03 | 0.10 | 0.15 |
| Cs | 0.03 | 0.03 | 0.01 | | 0.02 | 0.03 | 0.02 | 0.02 | 0.03 | | 0.01 | 0.03 |
| Ba | 29.2 | 26-2 | 13.1 | 7.7 | 7.4 | 40.2 | 15.5 | 24.6 | 30-6 | 10.3 | 10.4 | 42.7 |
| La | 4.3 | 4.6 | 1.64 | 1.2 | 1.2 | 4.7 | 2.4 | 2.6 | 4.3 | 1.0 | 1.45 | 5-3 |
| Ce | 11.4 | 11.3 | 4.8 | | | 13.5 | 7.7 | 7.7 | 10.9 | 3.5 | 4.2 | 13.0 |
| Eu | 0.94 | 0.9 | 0.6 | 0.45 | 0.72 | 0.92 | 0.71 | 0.84 | 0.83 | 0.51 | 0.59 | 0.98 |
| Tb | 0.55 | 0.55 | 0.38 | 0.45 | 0.45 | 0.59 | 0.47 | 0.52 | 0.54 | 0.35 | 0.37 | 0.28 |
| Hf | 1.5 | 1.49 | 0.84 | 0.94 | 0.9 | 1.72 | 1.13 | 2.48 | 1.47 | 0.72 | 0.81 | 1.6 |
| Та | 0.48 | 0.49 | 0.16 | 0.12 | 0.12 | 0.53 | 0.26 | 0.29 | 0.45 | 0.106 | 0.11 | 0.54 |
| Th | 0.57 | 0.56 | 0.18 | 0.12 | 0.07 | 0.65 | 0.32 | 0.31 | 0.54 | 0.11 | 0.13 | 0.57 |
| U | 0.5 | 0.14 | 0.02 | | 0.21 | 1.21 | 0.19 | 0.18 | 0.27 | 0.06 | 1.16 | 0.17 |

Trace and rare earth elements analyses of 25 of the Red Sea basalts at 18°N (expressed in ppm) done by neutron activation at the C.E.N. of Saclay (analyst J. L. Joron)

| | | | | | Г | TABLE 6 | (continue | ed) | | . मुझ् | •* | • | • |
|---------------|------|-------------|-------------------|--------------|------|---------|-------------|------|-------------|------------|------|------|------|
| Sample no. | 66-4 | 67-1 | 68-2 | 68- <i>3</i> | 70-3 | 70-4 | 70-5 | 71-1 | 71-2 | 71-3 | 72-2 | 72-4 | 72-8 |
| Sc | 40.1 | 40.1 | 41.2 | 41.9 | 42.0 | 41·4 · | 41.9 | 40.5 | 41.6 | 39.8 | 42·1 | 41.9 | 41.8 |
| Cr | 37. | 85. | 129. | 160. | 20. | 159. | 114 | 71· | | 102· | 33. | 68. | 25∙ |
| Co | 48.1 | 48 · | 49·5 [°] | 49.9 | 50.8 | 49.8 | 50·3 | 48.4 | 47·2 | 53-4 | 50.7 | 51.9 | 50.1 |
| Ni | 70∙ | 72· | 83. | 77. | 64. | 79· | 78 ∙ | 71. | 78 ∙ | 77. | 68· | 67. | 66. |
| Zn | | — . | | | _ | | | | _ | _ | | — | |
| Rb | 3.2 | 2.1 | 3-7 | 2.9 | 2.4 | 3.3 | 2.2 | 3.2 | 3.4 | 3.1 | 1.3 | 2.5 | 2.1 |
| Sr | 173 | · 201· | 155 | 153. | 132. | 152- | 140. | 140 | — | 142. | 118 | 116 | 112. |
| Zr | 55. | 74· | 69. | 86. | 70· | 74· | 58- | 57. | _ | 91· | 48· | 55- | 50· |
| Sb | 0.24 | 0.23 | 0.09 | 0.07 | 0.08 | 0.04 | 0.29 | 0.06 | 0.21 | 0.28 | 0.04 | 0.06 | 0.02 |
| Cs | 0.03 | 0.02 | 0.03 | 0.05 | 0.01 | 0.03 | 0.05 | 0.04 | 0.04 | 0.03 | 0.03 | 0.03 | 0.05 |
| Ba | 42. | 22· | 37.8 | 29.9 | 23.8 | 37.1 | 22.2 | 30.9 | 35.6 | 35.4 | 27.6 | 32.8 | 24.0 |
| La | 4.7 | 4.3 | 5.4 | 5.2 | 2.7 | 5.3 | 2.9 | 4.4 | 4.6 | 5∙0 | 2.5 | 2.8 | 2.7 |
| Ce | 13.4 | 11.5 | 12.8 | 14.1 | 8∙5 | 13.8 | 6.4 | 10.4 | | 15.5 | 7.6 | 8.9 | 7.8 |
| Eu | 0.98 | 1.06 | 1.13 | 1.1 | 0.76 | 1.12 | 0.79 | 1.0 | | 0.83 | 0.79 | 0.77 | 0.85 |
| Tb | 0.57 | 0.62 | 0.65 | 0.66 | 0.53 | 0.66 | 0.20 | 0.23 | 0.55 | 0.53 | 0.51 | 0.20 | 0.51 |
| Hſ | 1.66 | 2.42 | 1.85 | 1.92 | 1.27 | 1.89 | 1.32 | 1.45 | 1.60 | 1.49 | 1.52 | 1.2 | 2.4 |
| Та | 0.54 | 0.43 | 0.55 | 0.26 | 0.3 | 0.55 | 0.31 | 0.48 | 0.49 | 0.46 | 0.29 | 0.58 | 0.28 |
| Th | 0.57 | 0.47 | 0.57 | 0.67 | 0.27 | 0:66 | 0.26 | 0.48 | 0.54 | 0.53 | 0.27 | 0.35 | 0.28 |
| U | 0.33 | 0.21 | 0.24 | 0.32 | 0.14 | 0.4 | 0.4 | 0.13 | 0.33 | 13.8 | 0.12 | 3.4 | 0.18 |

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FIG. 13. Plot of La (ppm) and Th (ppm) vs. Ta (ppm) (same symbols as Fig. 10).

HYG elements. Other domains recognised in the North Atlantic and Pacific oceans, with La/Ta ratio as high as 20, are characterized by a depletion in HYG elements compared with chondrites and corresponds to typical N-type MORB. In the Th vs. Ta diagram (Fig. 13), the ratio is close to 1, which is a characteristic of the spreading-centre domains, whereas this ratio is much higher in the convergence domains; Th/Ta ≥ 10 (Table 7) (Joron *et al.*, 1977). However, in the Red Sea Rift samples near 18°N we note slight variations in the Th/Ta ratio, with slightly different correlations. These differences could have resulted either from variations in the degree of partial fusion or from variable source compositions.

The respective contribution of mantle source and of degree of partial fusion can be distinguished by studying variations of ratio versus ratio diagrams of HYG elements. In such diagrams, populations of samples coming from one mantle source are represented along a single correlation line whose y-axis intercept and slope characterize the mineralogy and the chemistry of the mantle source. On the other hand, populations of samples generated from compositionally distinct mantle sources are represented along different correlation lines. This method has been used in several areas of the oceanic lithosphere and demonstrates, on the basis of HYG elements distribution, the complex heterogeneity of the mantle (Joron et al., 1980b, Treuil et al., 1982). The distinction between partial melting and mantle source heterogeneities effects in regard to HYG elements fractionation has been emphasized in previous works (Joron et al., 1980b; Treuil et al., 1982) and these distinctions have been confirmed by the use of both HYG element ratios and isotopic data (Joron et al., in press). From such diagrams (e.g., Th/Ta vs. Th/Tb; Fig. 14), we can observe three main trends showing that these samples originated from at least three different sources. The first trend involves samples 59-1, 59-3, 65-2, and 65-4. The second is composed of samples 62-1, 62-3, 70-3, 70-5, 72-2, 72-4, and 72-8 and the last, samples 53-2, 53-3, 61-1, 62-4, 66-1, 66-2, 68-2, 68-3, 70-4, 71-1, 71-32, 71-3. Sample 57-1 is intermediate between groups 1 and 2 but closer to group 1 in other chemical and petrological characteristics. Sample 67-1 lies intermediate between groups 2 and 3 but its other characteristics indicate affiliation to group 3. By

Table 7

Comparison of some remarkable trace element ratios of the Red Sea basalts at 18°N with some oceanic lavas as a function of their geodynamic setting

| Oceanic ridges | | Th/Ta | Th/La | La/Ta | |
|--|---------------------------|--------------------------------|-----------------------------|----------------------|--|
| Red Sea 18°N | | 0.59–1.2 | 0.06-0.126 | 10 | |
| FAMOUS MAR 36°4 36°45'N IPOD Leg 45 East Pacific Rise | 15'N | 1·03 0·73–0·76 0·72–0·77 | 0·1 0·040·05 0·020·04 | 10 18–20 16–40 | (Joron <i>et al.</i> , 1978) and (Bougault <i>et al.</i> , 1979) |
| Arc magmatism linked to subduction zones | Chile Vanuatu Japan | 10 10–22 10 | 0·20·3 0·030·16 | 120-420 | (Briqueu et al., 1984) |



FIG. 14. Plot of Th/Ta vs. Th/Tb (same symbols as Fig. 10). Aden and Tadjoura Gulfs (Joron; unpublished data).

comparison with isotopic data from the 4 samples studied, we observe that the three samples having similar isotopic signatures belong to the same trend (Th/Ta vs. Th/Tb = 0.8) and that the only sample which is isotopically distinct (59-3) corresponds to a different correlation line (Th/Ta vs. Th/Tb = 0.25). Agreement between the HYG ratios and the isotopic data for the characterization of the mantle sources has already been observed elsewhere (Schilling, 1973; Hart *et al.*, 1973; Sun *et al.*, 1975; Wood *et al.*, 1979; Dupré & Allègre, 1980, 1983). Thus the Red Sea Rift in this area is fed by a heterogeneous mantle. Similar source heterogeneities have been observed on Zabargad island in the southern Red Sea (Petrini *et al.*, in press).

From the REE patterns (Fig. 15), these three groups of samples can also be distinguished. The first group shows a slightly LREE-depleted pattern, the second one has a rather flat pattern, and the last one has a slightly LREE-enriched pattern.



FIG. 15. Rare earth element abundances normalized to chondrites for selected samples of the Red Sea basalts at 18°N; note the presence of slightly depleted, flat, and slightly enriched patterns (same symbols as Fig. 10), for basalts collected in a small area of the axial trough of the Red Sea.

Thus the distinction between the different groups on petrology and bulk-rock chemistry is even more clearly brought out by the isotopic data, HYG ratios and REE patterns.

Crystal fractionation processes

In the Th/Tb vs. Th diagram (Fig. 16), the effect of crystal fractionation gives rise to horizontal shifts towards higher Th contents, the Th/Tb ratios remaining constant (Treuil *et al.*, 1982). As this shift is very limited in the set of samples analysed (Fig. 16), it shows that the crystal fractionation could not have exceeded 20%.

The variation of transition elements vs. HYG elements gives complementary information. The variation of Ni vs. Th, for instance, shows remarkably low contents for Ni (64–127 ppm) in the whole set of samples (Fig. 17). The three samples with the highest Ni contents (119–127 ppm) are the least evolved samples of the collection (Dives 57 and 65). These diagrams show that: (1) these basalts had already fractionated olivine before being extruded and are therefore relatively evolved; and (2) there is no evidence of any magma mixing between very primitive and more evolved compositions as observed, e.g., in the FAMOUS area (Fig. 17). This does not preclude the possibility of mixing of relatively evolved magmas inside a single zoned magma chamber, as is suggested by the petrography.

The Cr vs. Th diagram (Fig. 17) shows more scatter than that for Ni. The less evolved basalts have Cr contents between 394 and 404 ppm, whereas values in the others fall to 20 ppm. The general aspect of the trend is quite similar to the Ni trend and implies that these basalts have fractionated Cr-bearing phases such as chrome-spinels and/or clinopyroxenes.



FIG. 16. Plot of Th/Tb vs. Th (same symbols as Figure 10).

However, the flat trend in the Co vs. Th and Sc vs. Th diagrams (Fig. 17), with typical MORB-like contents (44–53 ppm for Co, and 37–43 ppm for Sc), indicates that fractionation of clinopyroxene was distinctly subordinate.

DISCUSSION AND CONCLUSION

The volcanic rocks collected by this cruise constitute a unique set of samples from the Red Sea at 18° N, collected *in situ* by submersible inside a *c*. 50 km² area (Fig. 2). We consider that they are statistically meaningful and reflect the variations of the basalts extruded at that latitude on the inner floor of the axial rift.

Excluding the holocrystalline alkali picrite from a fracture zone just north of the studied area, the whole collection comprises typical N-MORB, as evidenced by their mineralogy, major and trace element contents, and isotopic ratios, and some LREE-enriched MORB. Despite the fact that they have been produced during recent continental break-up and ocean opening, they are similar to the basalts extruded at long established ocean spreading ridges.

An important point is that no significant interaction with continental crustal materials has been observed, as attested to for instance by the Sr and Nd isotopic ratios. This means that at the latitude and at the stage of opening represented near 18°N, the oceanic spreading process has already reached a quasi-steady state involving only sub-oceanic mantle sources and their partial melt derivatives. This statement is probably valid for the whole Red Sea spreading segment between 17°N and 19°N, where a linear axial rift is well delineated and 1

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FIG. 17. Plot of Sc, Cr, Co, and Ni (ppm) vs. Th (ppm). For the Red Sea samples (same symbols as Fig. 10). Other data (contoured fields) given for comparison are from the FAMOUS area axial (Bougault et al., 1979; Bryan et al., 1979), DSDP sites 332 and 333 (Aumento et al., 1977), and DSDP sites 411 and 412 (Wood et al., 1978).

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expressed in the bathymetry. The Red Sea opening probably propagated northwards and southwards from this segment (Pouit & Guennoc, 1983) as, e.g., the oldest magnetic anomalies identified between 18°30'N and 19°45'N, 19°45'N and 20°N, near 21°N and near 20°N are respectively 3A (6 Ma), 2A, 2, and only 1 (Izzeldin, 1987). Therefore, 'along strike' variations in the petrology of the axial basalts all along the Red Sea Rift could be expected: e.g., transitional basalts richer in alkalis have already been described both in the northern Red Sea (Maury *et al.*, 1985) and in the southern Red Sea (Gass *et al.*, 1973).

The mineralogical study of the samples of the 18°N collection shows that their bulk compositions have been controlled mainly by polybaric fractionation of plagioclase, olivine and minor clinopyroxene, as usual in MORB. The plots of the fresh glass analyses follow the multisaturated cotectic-like curve for MORB in the normative olivine-silica-clinopyroxene projection (Fig. 11). Most of the samples are nearly aphyric and some of them are moderately porphyritic with a high plagioclase/olivine ratio (the range of total phenocrysts content is 0-15% by volume). It appears that crystal accumulation only played a significant role for the few porphyritic samples as indicated by the similarities between whole-rock and glass analyses (Fig. 10). The range of magmatic evolution, or degree of crystallization, is rather limited in the whole collection: we found neither primitive. Mg-rich basalts such as were described in the FAMOUS area for instance (Hékinian et al., 1976), nor strongly evolved Fe-Ti-rich basalts such as the ferrobasalts described, e.g., in the Galapagos rift (Clague et al., 1981). Despite this restricted compositional range, separate groups of samples have been distinguished by petrology (the SABs and MPBs) and glass chemistry (3 groups), two of which can be related to individual volcanic ridges in the axial rift (Fig. 18, Table 8). The third group, more widespread at the scale of the studied area, could represent the background signature of this portion of the floor of the Red Sea Rift.

The mantle source appears to be heterogeneous as demonstrated by both isotopic ratios $(^{206}\text{Pb}/^{204}\text{Pb}, \text{Fig. 12})$, and HYG elements ratio vs. ratio diagram (Fig. 14). The HYG elements in particular, show variations far beyond those that can be attributed to crystal fractionation or partial melting processes alone (Figs. 13 and 16). Mantle source heterogeneity is also generally invoked to explain such variations in MOR-type basalts (Bougault *et al.*, 1979; Bougault & Treuil, 1980). Again, three groups of samples can be distinguished, which in general correspond well to the ones identified by studying the whole-rock and glass chemistry and by petrological data. These data suggest that some of the volcanic ridges of the axial rift were produced by individual magma batches, each with specific geochemical and petrological characteristics.

The chondrite-normalized REE patterns (Fig. 15) strongly suggest that variations in the degree of melting also played a significant role. This is particularly so with respect to the LREE variation, with La/Sm ratios varying from 0.4 to 1.5.

Diagrams using hygromagmatophile elements either alone (Fig. 16) or combined with transition elements (Fig. 17) give complementary information and permit evaluation of crystal fractionation vs. partial melting processes. Despite several attempts at modelling, the different groups of magmas identified here cannot be related by crystal fractionation, confirming that their parental magmas were different. The general depletion in Ni and Cr indicates that significant fractionation of olivine and spinel has occurred prior to eruption, whereas the Sc and Co contents indicate that relatively little clinopyroxene fractionation had occurred. The possibility of polybaric crystal fractionation may explain the zonation of the clinopyroxene phenocrysts. Cores of these zoned pyroxenes may represent the only relicts of higher pressure stages, as no primary liquids are observed in this collection.

Furthermore, the Ni and Cr vs. Th diagrams (Fig. 17) indicate the absence of magma mixing between primary and more evolved magma batches, as described from the



FIG. 18. (a) Geographic distribution of the three different magma types distinguished in the axial valley of the Red Sea Rift at 18°N, presented on a schematic bathymetric map (large numbers are dive numbers). A summary of the characteristics of the magma types is given in Table 8. (b) Extrapolation of distribution of the three magma types, drawn from the sample distribution and the bathymetrical volcanic ridge identification (see the text for explanations).

| Т | AI | BL | E | 8 |
|---|----|----|---|---|
|---|----|----|---|---|

Petrographic and geochemical characteristics of the three different magma types defined in the Red Sea Rift at 18°N. See Fig. 18 for the geographic distribution.

| Magma type | MgO (%) | TiO ₂ (%) | Th (ppm) | Th/Tb | La (ppm) | REE pattern | ²⁰⁶ Pb/ ²⁰⁴ Pb |
|---------------|------------|-------------------------|--|---------|-------------|----------------|--------------------------------------|
| 1 | > 8·0 | <0.9 | < 0·2 > 0·4 0·20·4 | <0.5 | < 2 | depleted | ~ 18·79 |
| 2 | 6·8–7·5 | 1.0–1.3 | | >0.7 | 4–5·5 | flat | ? |
| 3 | 6·0–8·0 | 0.9–1.1 | | 0.5–0.7 | 2–3 | enriched | ~ 19·14 |

FAMOUS area (Fig. 17) and other Atlantic collections (Rhodes *et al.*, 1979; Walker *et al.*, 1979; Bryan *et al.*, 1981). However, this does not preclude magma mixing events between moderately evolved and slightly more evolved batches of MORB-type magmas having occurred inside each of the geochemically distinguished groups, as is suggested by the reverse zoning in some plagioclase phenocrysts.

Size of the magma chamber

The data presented here provide some constraints on the dimensions of the magma chambers beneath the Red Sea axial rift in this area. It has been possible to identify several magma batches with distinct geochemical signatures which are thought to have originated from varying degrees of melting of a heterogeneous mantle source. As described from the FAMOUS area (Hékinian et al., 1976), these observations do not support the concept of a wide, zoned, steady-state magma chamber, but rather they support a narrow, zoned, steadystate magma chamber, as wide as the inner floor of the rift in the FAMOUS area (Hékinian et al., 1976; Bryan & Moore, 1977; Bryan et al., 1979). Other investigators have argued against this model and concluded that magmas erupted within the FAMOUS rift valley as small discrete batches that cannot be related to one another by crystal fractionation (Langmuir et al., 1977; Byerly & Wright, 1978; Nisbet & Fowler, 1978; Flower & Robinson, 1979). In an attempt to overcome these apparent contradictions, Stakes et al. (1985) proposed that a small steady-state magma chamber (about 2 km wide) is maintained beneath the mid-section of each rift valley (FAMOUS, AMAR,...) but that it thins and loses its steady-state character toward the ends of the valleys: in this model as applied to the FAMOUS area, ephemeral magma bodies near the fracture zones at the ends of the valley connect with a larger steady-state body beneath the valley mid-section, and a continuous 'growth and contraction pattern' develops between them.

An interesting consequence of this model is that these ephemeral magma bodies will produce both primitive Mg-rich basalts during contraction periods, when relatively unfractionated mantle melts ascend directly to the surface, and Fe–Ti-rich evolved basalts produced by virtually closed-system fractionation in isolated magma pockets. By contrast, the steady-state magma chambers at the mid-sections of the spreading segments will produce a more limited compositional range of basaltic melts, showing ubiquitous evidence for magma mixing, suggestive of eruption from an open-system magma chamber. An expression of this duality in the MAR basalts at 36°N is visible in the plot of TiO_2vs . MgO of Stakes *et al.* (1985) which shows the bimodality of the FAMOUS basaltic glasses, and the limited, regular evolution of the glasses of the Narrowgate mid-section valley segment. Comparison of our Red Sea data with the FAMOUS–Narrowgate data (Fig. 10) shows a striking similarity between the Red Sea and Narrowgate glasses, whereas neither the Mgrich nor the Ti-rich glasses typical of the FAMOUS area are represented in our collection.

In conclusion, we infer that in such a slow-spreading environment as that represented by the Red Sea rift at 18°N, magma chambers wider than 2 km are unlikely. The basalts are relatively evolved, show a limited range of evolution and were derived from a heterogeneous mantle source. As proposed for the Narrowgate valley segment of the MAR at 36° N by Stakes *et al.* (1985), the magma group 1 (Fig. 18) could have formed in a small magma chamber (steady-state over a time-scale of less than 10^5 y) in which the major element content of the magma was largely buffered and which yielded the most evolved samples of the entire collection. The small chamber size, however, would prevent the buffering of trace elements during the major influxes of primitive melts from the mantle, so that the different magma batches erupted as a consequence of these influxes would retain their distinctive trace element characteristics, largely inherited from partial melting processes and mantle source compositions. The volcanic ridges 2 and 3 (Fig. 18) were formed by individual magma batches, each having their own geochemical and petrological signature. They probably came from ephemeral (10^2 to 10^3 years) magma chambers, which generally had insufficient time to evolve as much as would have occurred in a larger magma chamber.

Thus it seems that the recent activity (≤ 1 Ma) of the Red Sea Rift in the studied area was characterized by an initial period of high eruptive rate smoothing the topography with production of relatively geochemically homogeneous basalts. This was followed by two periods of intermittent activity building the two individual volcanic ridges, each of which has its own geochemical signature. These changes reflect the complex nature of the mantle source chemistry and the changing regional stress regime and/or upper mantle fusion rate, which directly influence the morphological-tectonic style of this 10-km-long spreading ridge segment.

The alternation of low and high eruptive rates proposed for this slow-spreading centre may well be applicable to other steady-state ridge sections with faster spreading rates, but on a longer time-scale and with their own geochemical and petrological signatures.

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APPENDIX

Short description of the samples

For each sample, the first paragraph gives a macroscopic description, the second paragraph a microscopic description.

Sample 53-1

Fragment of lava crust with double parallel glassy margins, hanging septa of glass and some trapped sediments. Plagioclase (PL) and olivine (OL) moderately porphyritic, vesicular basalt. Fresh, some scarce traces of palagonite.

Euhedral phenocrysts of OL and PL, often showing a rounded core (optical zonation). Euhedral to skeletal microphenocrysts of PL, OL, and clinopyroxene (CPX). Glassy to variolitic groundmass with fluidal texture.

Sample 53-2

Pillow lava of a moderately PL porphyritic, vesicular basalt. Microscopically similar to 53-1.

Sample 53-3

Pillow lava glassy bud of a PL sub-aphyric, vesicular basalt. Scarce euhedral phenocrysts of PL. Euhedral to skeletal microphenocrysts of PL, CPX, and OL often clustered in glomerocrysts. Glassy to variolitic groundmass, fluidal texture.

Sample 53-4

Pillow lava fragment of a PL, OL, and clinopyroxene (CPX) sub-aphyric, unaltered, vesicular basalt. Sub-euhedral phenocrysts of PL, often associated in mega-glomerocrysts. Euhedral phenocrysts of OL and one sub-euhedral CPX phenocryst containing numerous inclusions of PL microphenocrysts. Euhedral 'to skeletal microphenocrysts of PL, CPX, and OL. Glassy to variolitic groundmass.

Sample 57-1

Lava crust with glassy margin. PL and OL moderately porphyritic basalt. Euhedral OL phenocrysts. Sub-euhedral to euhedral megacrysts and phenocrysts of PL with sometimes an anhedral core (optical zonation). Euhedral to slightly skeletal microphenocrysts of PL and OL. Glassy to variolitic groundmass.

Sample 59-1

Angular massive fragment of PL sub-aphyric basalt. Skeletal microphenocrysts of PL, CPX and OL clustered in scarce glomerocrysts. Glassy to variolitic groundmass.

Sample 59-2

Slightly ropy lava with glassy margin of a moderately PL porphyritic basalt. Presence of one mega-glomerocryst made of sub-euhedral PL phenocrysts. Numerous glomerocrysts made of skeletal microphenocrysts of PL, OL, and/or CPX. Glassy to variolitic groundmass.

Sample 59-3

Glassy lava crust of a PL and OL sub-aphyric, vesicular basalt. Skeletal microphenocrysts of PL, CPX and OL clustered in scarce glomerocrysts. Glassy to variolitic groundmass.

Sample 59-4

Lava crust with a double glassy margin. PL sub-aphyric, slightly vesicular basalt. Microscopically similar to 59-3.

Sample 61-1

Pillow fragment of an aphyric basalt.

Very scarce euhedral PL microphenocrysts. Opaque glassy to variolitic groundmass.

Sample 62-1

Pillow fragment of a PL, OL, and CPX moderately porphyritic basalt. Presence of a mega-glomerocryst made of sub-euhedral PL and CPX phenocrysts. Euhedral to skeletal microphenocrysts of PL, OL, and CPX. Glassy to variolitic groundmass.

Sample 62-2

Pillow fragment of a PL sub-aphyric, vesicular basalt. Microscopically similar to 62-1.

Sample 62-3

Folded glassy lava crust of a PL, OL, and CPX sub-aphyric basalt. Euhedral PL phenocrysts with anhedral cores, and OL. Euhedral to skeletal PL, CPX and OL microphenocrysts. CPX and OL often associated in glomerocrysts. Glassy to variolitic groundmass.

Sample 62-4

Pillow lava crust of a PL and OL moderately porphyritic, vesicular basalt. Sub-euhedral PL megacrysts and phenocrysts, sometimes zoned. Scarce sub-euhedral OL phenocrysts. Rare skeletal PL and OL microphenocrysts. Glassy to variolitic groundmass.

Sample 65-1, 65-2 and 65-4

Pillow lava fragments of PL and OL porphyritic basalt (more vesicular for 65-2). Mostly euhedral PL megacrysts and phenocrysts, with sometimes anhedral cores. They cluster in mega-glomerocrysts containing small PL crystals and magmatic inclusions. Euhedral OL phenocrysts sometimes containing inclusions of PL crystals. Euhedral microphenocrysts of PL and OL. Glassy to variolitic groundmass.

Sample 65-3

Pillow lava crust of a PL moderately porphyritic basalt. Microscopically similar to 65-1 but less porphyritic.

Sample 65-5

Glassy lava crust of a PL, OL, and CPX porphyritic basalt. Sample similar to sample 65-1, with some CPX microphenocrysts and scarce spinels associated with OL. Glassy to variolitic groundmass.

Sample 66-1

Ensedimented glassy lava crust of a PL sub-aphyric, vesicular basalt. Rare sub-euhedral PL phenocrysts. Skeletal PL, OL, and CPX microphenocrysts clustered in glomerocrysts. Glassy to variolitic groundmass.

Sample 66-4

Pillow lava of a PL and OL moderately porphyritic, slightly vesicular basalt. Microscopically similar to 66-1.

Sample 67-1

Glassy lava bud of a PL and OL sub-aphyric, vesicular basalt. Euhedral to skeletal PL and OL phenocrysts associated in glomerocrysts. Glassy to variolitic groundmass.

Sample 68-1

Glassy lava crust of a rare OL sub-aphyric PL, vesicular basalt. Scarce sub-euhedral PL phenocrysts. Glomerocrysts made of skeletal PL and OL microphenocrysts. Glassy to variolitic groundmass.

Sample 68-2

Pillow fragment of a PL and OL moderately porphyritic, vesicular basalt. Sub-euhedral PL and euhedral OL phenocrysts. Skeletal PL, OL, and rare CPX microphenocrysts. Glassy to variolitic groundmass.

Sample 68-3

Pillow fragment of a PL sub-aphyric, slightly vesicular basalt. Sub-euhedral to euhedral PL phenocrysts. Euhedral to skeletal PL and OL microphenocrysts, associated in glomerocrysts. Glassy to variolitic groundmass.

Sample 70-1 and 70-3

Pillow fragments of PL sub-aphyric basalt.

Rare euhedral PL phenocrysts isolated or clustered in glomerocrysts. Skeletal PL, OL, and CPX microphenocrysts. Glassy, variolitic and microlitic groundmass.

Sample 70-2, 70-4 and 70-5

Pillow fragments of PL and OL sub-aphyric, vesicular basalt. Euhedral PL and OL phenocrysts, associated in mega-glomerocrysts. Skeletal PL, OL, and CPX microphenocrysts. Glassy to variolitic groundmass.

Sample 71-1 and 71-3

Pillow lava fragments of PL sub-aphyric, vesicular basalt.

Euhedral PL phenocrysts. Glomerocrysts made of skeletal PL, OL, and CPX microphenocrysts. Glassy groundmass and fluidal texture parallel to the margin (for 71-3).

Sample 71-2

Glassy lava crust of a PL sub-aphyric basalt.

Euhedral or anhedral PL phenocrysts often associated in glomerocrysts. Skeletal PL, OL, and CPX microphenocrysts. Glassy to variolitic groundmass.

Sample 71-5

Ensedimented glassy pillow rim of a PL and OL sub-aphyric basalt.

Sub-euhedral PL and OL phenocrysts associated in glomerocrysts. Skeletal PL, CPX, and OL phenocrysts associated in glomerocrysts. Opaque glassy to variolitic groundmass, fluidal texture.

Sample 72-1

Pillow fragment of a PL sub-aphyric basalt.

Rare sub-euhedral to skeletal PL phenocrysts and PL, OL, and CPX microphenocrysts. Glassy to variolitic groundmass, fluidal texture.

Samples 72-2 and 72-3

Massive basaltic fragments without glassy rim. PL and CPX sub-aphyric, vesicular basalts. Microscopically similar to 72-1 with an intersertal to microlitic texture.

Sample 72-4

Similar to 72-2 but highly vesicular.

Samples 72-5, 72-6, 72-7, and 72-8

Pillow fragment of a $PL \pm OL$ sub-aphyric basalt.

Euhedral PL and OL phenocrysts in mega-glomerocrysts. Euhedral to skeletal PL, OL, and CPX microphenocrysts. Glassy to variolitic groundmass.

Sample 23-2b (picrite)

Small fragment of massive, micro-crystalline and doleritic picrite, dredged at $18^{\circ}01'N$, $40^{\circ}02'E$. Numerous sub-euhedral OL phenocrysts and microphenocrysts showing a slight optical zoning, but a strong chemical one (Fo₈₆-Fo₆₅). They include small spinel crystals in their external margins. Groundmass formed of CPX and PL microlites, with intersertal Ti-magnetite, biotite, K-feldspar, and albite grains.

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The following samples are not in our possession: 66-2, 66-3 and 71-4.