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Contributions to Mineralogy and Petrology

Samarium-neodymium data on two late Proterozoic ophiolites of Saudi Arabia and implications for crustal and mantle evolution

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Abstract. Whole-rock and mineral samples from the Jabal al Wask and Jabal Ess ophiolites, northwestern Saudi Arabia, yield Sm–Nd isochron ages of 743±24 Ma and 782±38 Ma, respectively. These formation ages, which provide maximum limits for possible obduction ages, are in broad but not precise agreement with the previously known geologic history of the Arabian Shield. They indicate that the ophiolitic rocks are roughly coeval with nearby volcanic and plutonic rocks, supporting a back-arc origin for the two ophiolites. We suggest that the Jabal al Wask and Jabal Ess ophiolites were parts of the same northeast-southwest trending ophiolite belt, now offset along the Najd fault system. Initial ε_{Nd} values range from +6.6 to +7.6, indicating derivation from a mantle source that has been LIL-depleted for at least 2 Ga. Reported ε_{Nd} values from the Arabian Shield that are lower than this suggest the presence of older, reworked continental crust.

Introduction

Many workers have identified the Arabian-Nubian Shield as a type area for demonstrating that modern-type plate tectonic processes operated during the Late Proterozoic. Greenwood et al. (1976) first suggested that the Shield was produced by the accretion of island arcs. Schmidt et al. (1979) defined sutures, with associated ophiolitic rocks, that represent the remnants of subduction zones within the Arabian Shield. Fleck et al. (1980) outlined a plate-tectonic model for the evolution of the southern part of the Shield that culminated with arc accretion during collision with the African craton. Gass (1981) proposed a model that summarizes the results and ideas presented in many previous papers by various workers. In this model, the shield is composed of a number of intra-oceanic island arcs that have been pushed together, along with their sedimentary aprons and occasional slices of oceanic lithosphere, to form new continental crust.

Alternative models for the evolution of the Shield call for formation in an intracratonic tectonic setting either by repeated rifting and compression (Garson and Shalaby 1976; Kemp et al. 1980, 1982; Delfour 1981; Stern 1979, 1981), or as a Proterozoic greenstone belt (Engel et al. 1980). These models are not consistent with the tholeiitic to calc-alkaline major-element chemistry and with the trace-element chemistry of most of the volcanic and plutonic rocks of the Arabian Shield (Schmidt and Brown 1982; Roobol et al. 1983; Reischmann et al. 1983).

Isotopic data support an oceanic-mantle origin for the central parts of the Arabian-Nubian Shield (Bokhari and Kramers 1981; Duyverman et al. 1982; Stacey and Stoeser 1983). However, Stacey and Stoeser (1983) also show that the isotopic composition of lead from both the eastern and the westernmost parts of the Shield indicates the existence in these areas of an older continental crustal component. Calvez et al. (1983) and Stacey and Hedge (1983) have published geochronologic and isotopic evidence that suggests the presence of Early or Middle Proterozoic continental crust in the eastern Arabian Shield. Dixon (1979) found Early or Middle Proterozoic granite cobbles in a conglomerate from the Egyptian Shield. These data suggest that new continental crust of the Arabian-Nubian Shield formed in an ocean basin which was bordered both to the east and west by older cratons. Influx of material from this older continental crust was limited, however, and only affected the marginal parts of the accreting Shield.

Crossing the Arabian-Nubian Shield are a number of narrow, discontinuous and disrupted zones containing mafic and ultramafic rocks (Fig. 1) (Bakor et al. 1976; Baubron et al. 1976; Frisch and Al-Shanti 1977; Sustrac 1980; Gass 1981). In the plate tectonic model outlined above, these rocks are considered to be fragments of oceanic crust, originally parts of the ocean floor which separated the island arcs from each other. They were obducted to their present positions during the collision of the arc systems. Some of the mafic-ultramafic complexes have been identified as ophiolite suites in the commonly used sense of the 1972 GSA Penrose Conference on ophiolites (Bakor et al. 1976; Frisch and Al-Shanti 1977; Rehaile and Warden 1978; Shanti and Roobol 1979). Most of the complexes are incomplete, lacking various components of the ophiolite stratigraphy; and most are disrupted by faulting; locally forming melange belts. These ophiolitic complexes, together with other Late Proterozoic ophiolites of northern Africa (Leblanc 1976, 1981) and Wales (Thorpe 1978), are among the oldest fragments of modern-type oceanic lithosphere known. As such, they can provide more direct information about the Proterozoic mantle than any continental-type rocks.

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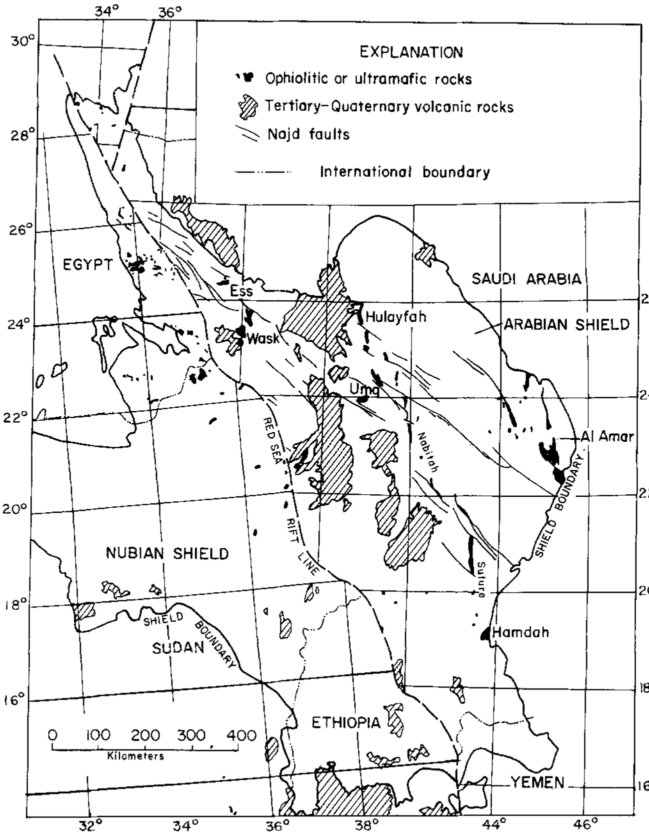

Fig. 1. Map of the Arabian-Nubian Shield prior to Red Sea rifting showing the distribution of ophiolitic rocks and other prominent geologic features. Based on maps by the U.S. Geological Survey and Arabian American Oil Company (1963), Merla et al. (1973), Vail (1974), Baubron et al. (1976), Frisch and Al-Shanti (1977), Schmidt et al. (1979), Dixon (1979), Egyptian Geological Survey (1981), Stacey and Stoeser (1983), and Fitches et al. (1983). Selected ophiolite localities are indicated by full or abbreviated names: Ess, Jabal Ess; Wask, Jabal al Wask; Umq, Bir Umq; Al Amar, Al Amar-Idsas suture

In northwestern Saudi Arabia, the Jabal al Wask (Bakor et al. 1976) and Jabal Ess (Shanti and Roobol 1979) mafic-ultramafic complexes are described as ophiolite sequences. All of the classical ophiolite components are present in the Jabal Ess suite. The Jabal al Wask ophiolite is structurally more complicated than originally thought (Kemp 1981), but most of the ophiolite components are present (Fig. 2). In this paper we present Sm–Nd isotopic data for whole-rock and mineral samples from these two ophiolites and discuss the implications of the data for the age of the ophiolites, the evolution of the Arabian Shield, and the evolution of the mantle.

Sample descriptions

Sample localities are given in Figs. 2 and 3, and petrographic descriptions are given in Table 1. The Jabal al Wask sample (175408) is from a noncumulus, deformed hornblende gabbro pluton that intrudes serpentinized peridotite near the center of the complex. The gabbro is intruded by a younger monzogranite pluton to the northeast.

The two Jabal Ess ophiolite samples are from near the southwest and northeast ends of the outcrop belt (Fig. 3). Sample 175403 is from a northwest-trending ridge of cumulus layered gabbro that intrudes serpentinized peridotite. The gabbro and peridotite are faulted against younger sedimentary rocks to the southwest. Sample 175578 is from a gabbro pluton that intrudes serpentinized peridotite and is intruded by a later alkali feldspar granite.

Each of the samples has undergone some degree of alteration and greenschist facies recrystallization characterized by saussuritization of plagioclase and replacement of pyroxene and hornblende by fibrous amphibole (uralite) and chlorite. The samples were selected because they are the least altered gabbros found by the second author in a broad reconnaissance study of the Precambrian Arabian ophiolites. Each of the samples retains some primary magmatic minerals and has a relict igneous texture.

Analytical techniques and precision

Plagioclase, clinopyroxene, and hornblende were separated from the gabbro samples using magnetic and heavy liquid methods. After handpicking, the mineral separates were ultrasonically cleaned several times in distilled acetone before dissolution. Analytical procedures for Sm–Nd isotopic determinations closely follow those published by Nakamura et al. (1976). About 100 mg of whole-rock powder, pyroxene, and hornblende, and up to 500 mg of plagioclase from each sample were dissolved in a mixture of hydrofluoric and perchloric acids in teflon bombs at 75° C for several days. Each dissolved sample was evaporated to dryness repeatedly with perchloric acid, dissolved in hydrochloric acid, centrifuged, and aliquoted in two portions. The smaller portion, about 20% of each sample, was spiked with a combined ^{146}Nd–^{150}Sm solution. Both aliquots were then eluted through cation exchange columns to separate the rare-earth elements (REE). Sm and Nd were separated from each other and from the other REE in a second series of over-pressure cation-exchange columns using hydroxy-α-isobutyric acid with pH adjusted to 4.495. The smaller samples were purified of hydroxy-α-isobutyric acid in small cation exchange columns eluted with hydrochloric acid. The blank for the whole procedure was about 50 pg Sm and 200 pg Nd. This blank level is significant for the spiked aliquots of the low-abundance level plagioclase fractions from samples 175403 and 175408. To account for a possibly erratic blank-correction, the estimated uncertainty in the ^{147}Sm/^{144}Nd ratio for these two samples was doubled to $\pm 1\%$, as described below.

The samples were loaded onto Ta side filaments in a reducing H_2+N_2 atmosphere and were run in triple filament configuration, with a Re center filament, as metal ions. The unspiked Nd samples were run on a Micromass 54*, 12″ mass spectrometer, and the spiked samples on a 12″ NBS spectrometer, both with computer-controlled data aquisition and reduction. For the composition runs, 200–300 scans over mass numbers 143, 144, and 146 were taken for each sample, with a beam intensity of 400–1600 mV. Sm interference was monitored at mass 147.

Typically, the 147/144 mass-ratio was smaller than 1×10^{-4}. A corresponding correction was applied to mass 144. Mass fractionation was corrected against $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. During the course of this study, 3 samples of the La Jolla Nd standard (Lugmair and Carlson 1978) and 7 samples of USGS rock standard BCR-1 were analyzed. The measured average ratios for these standard runs are listed in Table 2 together with published ratios for these standards from other laboratories. The good agreement in the ratios in Table 2, indicate the accuracy of the data, but the scatter in the reported values, corresponding to 0.9 ε_{Nd}-units for BCR-1, illustrates the problem of interlaboratory comparisons of high-precision isotopic data. Two of the samples listed in Table 3 were dissolved, separated, and analyzed in duplicate; the results from these runs indicate an excellent reproducibility. The seven

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* Any use of trade names is for descriptive purposes only and does not imply endorsement by the U.S. Geological Survey

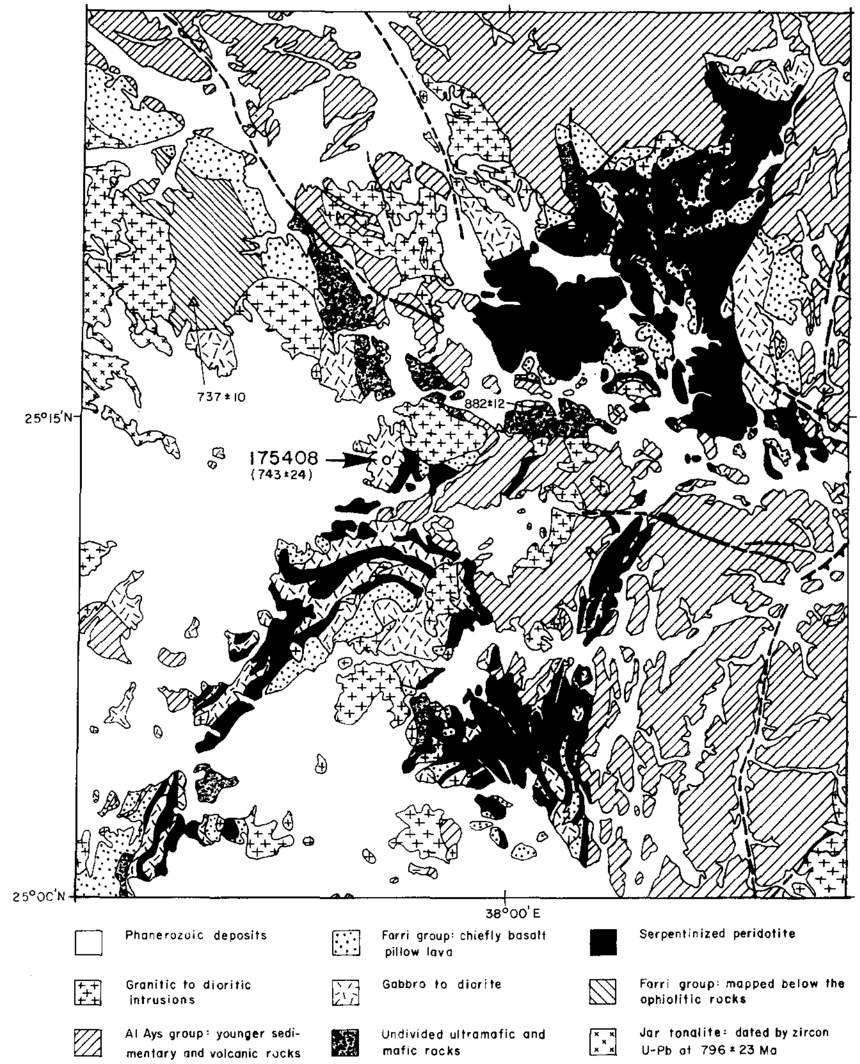

Fig. 2. Geologic map showing sample localities at the Jabal al Wask ophiolite. Circle indicates Sm–Nd sample; triangles indicate U–Pb zircon-dated samples from Kemp et al. (1980). Map simplified from Kemp (1981). Solid and dashed lines indicate mapped faults. Scale indicated by latitude grid: 1′ latitude = 1 nautical mile = 1.85 km

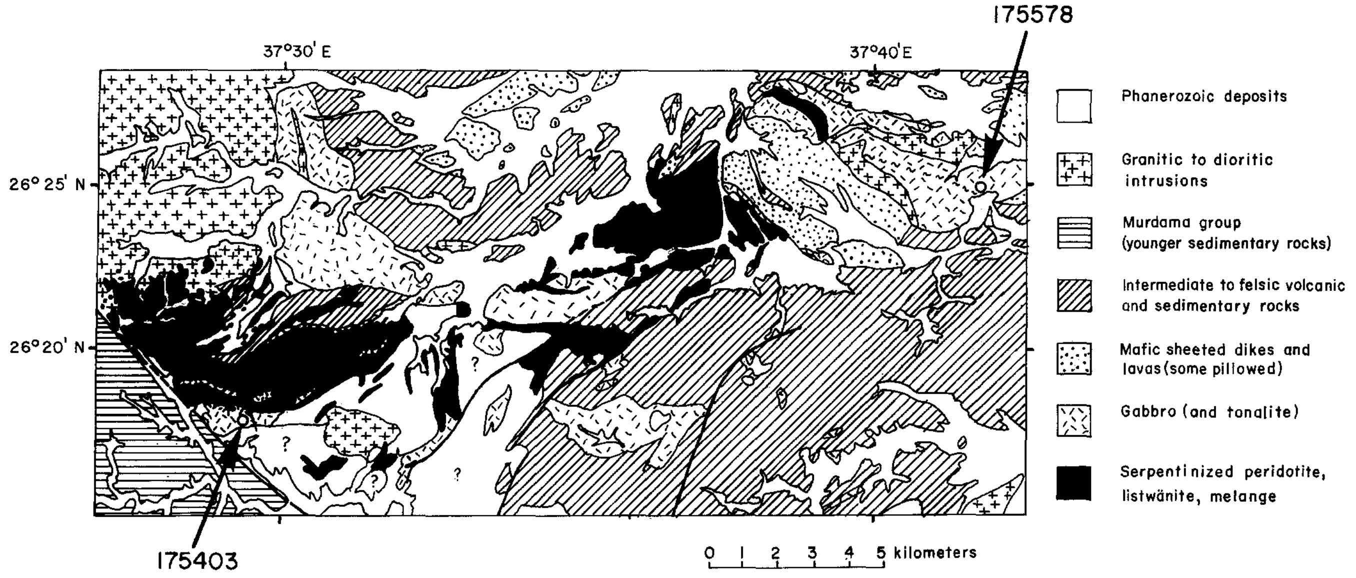

Fig. 3. Geologic map showing sample localities at the Jabal Ess ophiolite. Map simplified from Shanti and Roobol (1982) and Chevremont and Johan (1982). Areas labeled with question marks are unmapped. Circles indicate sample localities. Solid and dashed lines indicate mapped faults

Table 1. Petrography of Sm – Nd samples

Sample	Rock type and description
175408	Clinopyroxene-bearing hornblende gabbro; hypidiomorphic granular, medium grained. ~75% zoned andesine moderately saussuritic; ~20% pale green to tan magmatic hornblende, partly uralitic; ~5% clinopyroxene, cores of grains that were replaced by hornblende, then uralite; ~1% opaque minerals (titanomagnetite?). Jabal al Wask ophiolite
175403	Clinopyroxene gabbro; clinopyroxene-plagioclase adcumulate, medium grained. ~65% tabular unzoned labradorite, slightly to moderately saussuritic; ~35% clinopyroxene, largely replaced by uralite. Jabal Ess ophiolite
175578	Hornblende-clinopyroxene gabbro; hypidiomorphic granular, coarse grained. ~65% oligoclase-andesine, moderately to highly saussuritic; ~20% clinopyroxene, partly replaced by hornblende or uralite; ~15% hornblende, uralitic and chloritic; ~1% opaque minerals (titanomagnetite), hematitic; trace biotite, apatite and zircon. Jabal Ess ophiolite

Table 2. ^{143}Nd/^{144}Nd ratios for the La Jolla Nd-standard and the USGS rock-standard BCR-1 reported from different laboratories

La Jolla	BCR-1	Reference
0.511856	0.512653	This study
0.511859	–	Lugmair and Carlson (1978)
0.511861	0.512640	Wasserburg et al. (1981)
–	0.512664	Hamilton et al. (1983)
0.511880	–	Bokhari and Kramers (1981)
–	0.51262	Duyverman et al. (1982)

All values normalized to ^{146}Nd/^{144}Nd = 0.7219

Table 3. Sm – Nd analytical data

Sample		Sm (ppm)	Nd (ppm)	^{147}Sm/^{144}Nd[a]	^{143}Nd/^{144}Nd[b]
Jabal al Wask					
175408	plag	0.0324	0.2463	0.0794 ± 8	0.512457 ± 16
	wr	1.791	5.137	0.2107 ± 11	0.513087 ± 16
	most mag	1.359	3.719	0.2225 ± 11	0.513165 ± 36
	hbl	6.313	15.03	0.2538 ± 13	0.513304 ± 11
					0.513312 ± 12
Jabal Ess					
175403	plag	0.0509	0.1764	0.1744 ± 18	0.512899 ± 24
	wr	0.4383	0.8100	0.3271 ± 16	0.513635 ± 25
	pyr	0.7929	1.306	0.3671 ± 18	0.513867 ± 14
175578	plag	0.2666	1.392	0.1158 ± 6	0.512557 ± 19
					0.512557 ± 12
	wr	1.322	4.244	0.1882 ± 9	0.512935 ± 14
	pyr	2.538	6.463	0.2374 ± 12	0.513179 ± 7

plag = plagioclase, wr = whole rock, most mag = most magnetic mineral separate, hbl = hornblende, pyr = pyroxene

[a] Precision based on replicate analyses of BCR-1 standard

[b] Within-run precision from mass spectrometrical analysis

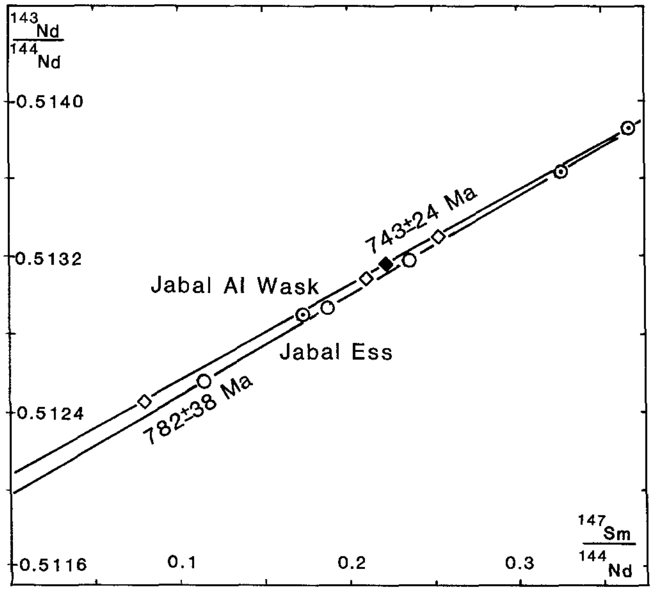

Fig. 4. Sm – Nd isochron diagram (^{143}Nd/^{144}Nd on the ordinate, ^{147}Sm/^{144}Nd on the abscissa) showing data points for rock and mineral analyses in Table 2. Diamond symbols = Jabal al Wask sample 175408, filled diamond = most magnetic fraction; open circles = Jabal Ess sample 175578; dotted circles = Jabal Ess sample 175403. Analytical uncertainty (Table 1) for data points is approximated by the size of the symbols

BCR-1 analyses give a between-run precision of $\pm 20 \times 10^{-6}$, slightly worse than the within-run precision for most samples (Table 3). For the isochron calculations, the larger value of the within-run precision or 20×10^{-6} was used. The precision of the ^{147}Sm/^{144}Nd ratios was estimated as ±0.5%, based on three BCR-1 runs.

Results

Analytical data for the ophiolite samples are listed in Table 3 and are plotted on a Sm-Nd isochron diagram in Fig. 4, where the data form a linear array with a good spread in Sm/Nd ratios. A York (1969) regresson of all 10 data points yields an age of 760 ± 39 Ma, an initial ^{143}Nd/^{144}Nd ratio of 0.512020 ± 0.000019, and a mean squared weighted deviation (MSWD) of 6.94.

Investigations of other ophiolite complexes (Jacobsen and Wasserburg 1979; McCulloch et al. 1981) have indicated that samples from the same ophiolite body may show a small but significant scatter in $(^{143}\mathrm{Nd}/^{144}\mathrm{Nd})_i$ (initial ratios). In the present case, there is no reason to believe that Jabal al Wask and Jabal Ess are of identical age. When the three samples are regressed individually, the Jabal al Wask sample (175408) yields 743 ± 24 Ma, $(^{143}\mathrm{Nd}/^{144}\mathrm{Nd})_i = 0.512069 \pm 0.000031$, and MSWD = 0.58, and the two Jabal Ess samples (175403 and 175578) yield 761 ± 27 Ma, $(^{143}\mathrm{Nd}/^{144}\mathrm{Nd})_i = 0.512025 \pm 0.000054$, and MSWD = 3.59; and 782 ± 36 Ma, $(^{143}\mathrm{Nd}/^{144}\mathrm{Nd})_i = 0.511966 \pm 0.000044$, and MSWD = 0.35, respectively. The two Jabal Ess samples together give 782 ± 38 Ma, $(^{143}\mathrm{Nd}/^{144}\mathrm{Nd})_i = 0.511968 \pm 0.000026$, and MSWD = 2.44. Within the analytical uncertainty, the ages of the three samples are not resolvable, but it appears likely that Jabal al Wask is slightly younger than Jabal Ess.

Initial ^{143}Nd/^{144}Nd ratios are commonly expressed in $\varepsilon_{\mathrm{Nd}}$ notation (DePaolo and Wasserburg 1977). The $\varepsilon_{\mathrm{Nd}}$ value indicates the deviation ($\times 10^4$) of the initial $^{143}Nd/^{144}Nd$ ratio of a sample from that of a CHondritic Uniform Reservoir (CHUR). The ε_{Nd} values for the three samples, 175408 (Wask), 175403 (Ess), and 175578 (Ess), calculated according to the method of Fletcher and Rosman (1982), are 7.6 ± 0.3, 7.2 ± 0.4, and 6.6 ± 0.3, respectively.

Whole-rock rare-earth element analyses of the three gabbro samples were made as part of a large suite of samples from the Arabian Shield ophiolites. Analyses were made using standard instrumental neutron activation analysis procedures. Chondrite normalized plots of the results from the gabbros, and diabase and pillow lava samples from Jabal Ess are given in Fig. 5 and are discussed below.

Discussion

Effect of alteration

The samples investigated in this study have been metamorphosed to greenschist facies and variably altered. Except for positive Ce-anomalies in two of the samples (Fig. 5), which may have been produced by metasomatism (Hellman et al. 1977), the REE patterns of the samples appear to reflect magmatic distributions. The good fit of the samples to the Sm–Nd isochrons (Fig. 4) indicates that alteration has not disturbed the Sm–Nd system in the samples. This is in accord with the observation of many other workers that the Sm–Nd system is very resistant to metamorphic disturbances. The colinear results from the "most magnetic" mineral fraction from sample 175408 also supports this conclusion (Fig. 4). The other mineral fractions represent the purest mineral separates that it was possible to achieve without inordinate effort; all are >90% pure. In contrast, the "most magnetic" fraction is highly enriched in Fe-rich intergranular material and alteration products of the primary minerals. The comparatively low Sm and Nd content of this fraction is probably due to dilution with plagioclase.

Age and tectonic setting of the ophiolites

The age of 740 to 780 Ma for the Jabal al Wask and Jabal Ess ophiolites is in broad agreement with the known temporal evolution of the area. According to the review of Gass (1981, p 397), 1000 to 600 Ma was "a period of numerous maturing ensimatic arc systems subsequently swept together by plate-tectonic processes to form, by the end of the division, larger 'protocontinental' masses." In detail, there are discrepancies between the ages presented here and the picture outlined by previous workers.

De La Boisse et al. (1980) determined U–Pb zircon dates (also reported in Kemp et al. (1980)) of 882±12 Ma for plagiogranite from the Jabal al Wask ophiolite, 796±23 Ma for the Jar tonalite that intrudes the upper Farri group (a chiefly pillow basalt sequence here considered part of the ophiolite), and 737±10 Ma for rhyolite in the lower Farri group (chiefly welded and unwelded tuff mapped below the ophiolite) (Fig. 2 and Kemp 1981). The plagiogranite and tonalite dates were based on a limited number of discordant points. Analysis of additional fractions and samples has led to revision of the plagiogranite date to 780±20 Ma (JY Calvez, oral communication, 1983). Kemp (1981) regarded the peridotite and gabbro of the ophiolite as intrusive into the upper Farri group pillow lavas and attributed the young age (737±10 Ma) for the lower Farri rhyolite to a post-Farri age silicic intrusion into the section. However, recent U–Pb zircon dating of a rhyolite flow within the lower Farri yielded a date of 742±6 Ma (JY Calvez, oral communication, 1983), in excellent agreement with the earlier determination of 737±10 Ma. The Sm–Nd date reported here for Jabal al Wask (743±24 Ma) overlaps the revised zircon date of the plagiogranite (780±20 Ma) and the zircon dates from the lower Farri rhyolite.

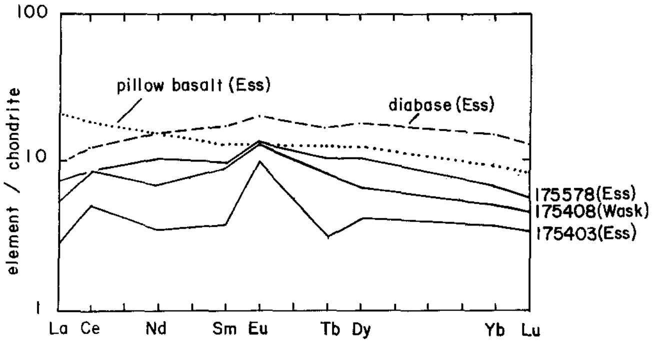

Fig. 5. Chondrite-normalized REE (rare earth element) patterns fro samples from the Jabal al Wask and Jabal Ess ophiolites

These dates show that the ophiolite is approximately coeval with, or slightly older than, the underlying volcanic rocks. Therefore, an overthrust structural setting is suggested for the ophiolite. The dates indicate that the ophiolite gabbro crystallized during a period characterized by silicic volcanism and tonalitic plutonism in the northwestern Arabian Shield. The oceanic crust represented by the gabbros and pillow basalts probably formed close to this region of apparent island-arc magmatism, not hundreds of kilometers away in a paleo-ocean basin.

Bakor et al. (1976) argued that pyroclastic and shallow water, volcanogenic sedimentary rocks overlie the Jabal al Wask ophiolite and indicate formation in a shallow basin adjacent to a volcanic landmass. They cited geochemical data that show the ophiolite basalts to be tholeiitic, with trace-element characteristics of ocean-floor or island arc basalts, and they suggested that the ophiolite formed in a back-arc tectonic setting.

The REE patterns for the gabbros from the two ophiolites are gently convex upward and have positive Eu-anomalies, indicating relative clinopyroxene and plagioclase accumulation (Fig. 5). The patterns are similar to those of high-level gabbros of Phanerozoic ophiolites (Pallister and Knight 1981). These data are compatible with crystal fractionation of the gabbros from basaltic magma with a nearly flat or slightly light rare-earth element (LREE)-depleted pattern.

The REE pattern for the Jabal Ess diabase is at an abundance and shows LREE depletion characteristic of basalts from mid-ocean ridges, immature island arcs, and small ocean basins. A magma with this REE distribution could have been parental to the gabbros. In contrast, the Jabal Ess pillow basalt sample shows LREE enrichment, indicating a different source. This LREE enrichment has some significance for the formative environment of the ophiolites, because normal-type mid-ocean ridge basalt is LREE-depleted, enriched or plume-type mid-ocean ridge basalt and transitional types are LREE-enriched, and island arc and back arc basalts are both LREE-depleted and LREE-enriched (Sun et al. 1979; Hawkesworth et al. 1977; DePaolo and Johnson 1979; McCulloch and Perfit 1981).

The age relations of the ophiolites and associated rocks and the REE chemistry are compatible with origin in a marginal back-arc tectonic setting. The association in time and space of the ophiolitic rocks with calc-alkaline volcanic and plutonic rocks is additional evidence for such a setting. The formation of sheeted dikes is likely to require a tensional tectonic regime. The apparent lack of sheeted dikes in the Jabal al Wask ophiolite may indicate that this complex formed during less tension than Jabal Ess, and this is consistent with both having formed in a marginal basin-island arc environment, where stress patterns should change over a short distance – from tension at the marginal spreading center to compression near the subduction zone. The observations of Kemp (1981) that some of the ophiolite plutonic rocks are intrusive into the Farri group volcanosedimentary sequence may also be explained by back-arc magmatism and spreading.

Gass (1981) suggested that the ophiolites of the Arabian shield were emplaced at about 1000, 800, and 600 Ma, based on the apparent ages of the surrounding volcanosedimentary sequences. These ages were based on radiometric determinations that, in some cases, are now known to be suspect for analytical or geologic reasons. The older episode of oceanic and immature island-arc magmatism (the "Lower Pan-African" of Gass 1981) is now believed to correspond to a period from about 800 to 950 Ma in the southern Arabian Shield during which the Baish, Bahah, and Jiddah groups were deposited (Reischmann et al. 1983; Pallister 1983; Stoeser et al. 1983). These rocks were sutured to a younger, more evolved arc complex to the east, the Halaban-Hulayfah group (680 to about 800 Ma). This composite package was accreted to a crustal block farther to the east that has Middle or Early Proterozoic continental roots during a major collisional episode from about 640–680 Ma (Stoeser et al. 1983). The contact between these crustal blocks is the belt of ophiolites shown in Fig. 1 that corresponds to the Nabitah suture or mobile belt (Schmidt et al. 1979; Stoeser et al. 1983).

Because the ages of the ophiolites date the formation of oceanic crust, they only give a maximum limit for possible obduction ages. Emplacement in the Wask-Ess area took place prior to deposition of the Al Ays group, which, according to Kemp (1981), unconformably overlies the Jabal al Wask ophiolite and is intruded by the Jabal Salajah tonalite. Zircon dating of the tonalite yielded 725 ± 12 Ma (De La Boisse et al. 1980, also reported in Kemp et al. 1980). Therefore, ophiolite emplacement occurred shortly after formation, between about 725 and about 750 Ma.

The large Najd fault system (Schmidt et al. 1979) crosses the Arabian Shield with a northwest trend (Fig. 1). Along these faults, cumulative left-lateral strike-slip of about 300 kilometers appears to have occurred, disrupting originally continuous linear features such as the Hulayfah-Hamdah ophiolite belt (the Nabitah suture) (Frisch and Al-Shanti 1977; Schmidt et al. 1979). The similar ages of the Jabal al Wask and Jabal Ess ophiolites suggest that they were once part of a single belt.

Bakor et al. (1976) and Frisch and Al-Shanti (1977) suggested that the Jabal al Wask and Jabal Ess ophiolites, together with several smaller ultramafic-mafic suites to the northwest and the Bir Umq complex to the southeast, form an ophiolite zone that branches off the Hulayfah-Hamdah ophiolite belt and continues in a northwesterly direction into Egypt (Fig. 1). However, about 50 kilometers left-lateral offset along the Najd faults between Jabal al Wask and Jabal Ess would bring the Jabal Ess ophiolite into an apparent northeast-southwest line of ophiolites that extends into the northeast Sudan (Fig. 1). This is in accord with the suggestion of Camp (1983) that the northwestern Arabian Shield was formed by northeast-southwest trending island arcs. Additional radiometric dating of the ophiolites is underway to test various plate tectonic models for the Shield.

Age of the arabian shield

Bokhari and Kramers (1981) obtained ε_{Nd} values between +7 and +9 for two sequences of volcanic rocks from a small area in the southern Arabian Shield. Duyverman et al. (1982) determined ε_{Nd} values between +1 and +7 for different rock types from 8 scattered localities in the Shield. As was pointed out by these authors, these data, especially the high ε_{Nd} values of Bokhari and Kramers, suggest that the rocks were derived from a LREE-depleted source, which suggests a depleted mantle origin. In addition, Bokhari and Kramers (1981) pointed out that the high ε_{Nd} values require a source that has been depleted for a long time. They calculated a T^{Nd}_{CHUR} (time of differentiation from the CHondritic Uniform Reservoir) model age of 2.2 Ga for this depletion event.

The initial ε_{Nd} values of +6.6 to +7.6 presented in this study for the Jabal al Wask and Jabal Ess ophiolites should be identical to the values in their contemporaneous mantle sources and are assumed to be representative of the mantle sources of the Arabian Shield. Comparison of ε_{Nd} values obtained from different laboratories should be made with caution because small systematic differences may exist (Table 2). However, some of the ε_{Nd} values reported by Duyverman et al. (1982) are significantly lower than the range +6.6 to +7.6 suggested here for the mantle source for the ophiolites. The presence of a component with low ε_{Nd} in these rocks was taken by Stacey and Hedge (1983) as supporting evidence for the presence of older crustal material in the eastern Arabian Shield. Indeed, a comparison between the Nd data of Duyverman et al. (1982) and the Pb data of Stacey and Hedge (1983), shows that all the samples with ε_{Nd} values lower than +5 come from areas with crustal- or transitional-type lead isotopic ratios.

Mantle evolution

Both Rb–Sr isotopic data (Brooks et al. 1976) and Pb–Pb isotopic data (Church and Tatsumoto 1975; Tatsumoto 1978) for oceanic rocks define linear arrays that may be interpreted as isochrons. Brooks et al. (1976) argued that the fact that both systems provide "mantle isochrons" with indistinguishable ages of about 1.6 Ga for Rb–Sr and 1.7 Ga for Pb–Pb hardly can be fortuitous, and they suggested that these dates reflect the age of a major differentiation event in the mantle. However, Tatsumoto (1978) showed that the linear trends should be interpreted as reflecting an average age for a progressive mantle differentiation that started directly after formation of the Earth. Determining whether the Sm–Nd isotope systematics of the mantle also can be forced into the two-stage evolution model will further clarify the meaning of mantle isochrons.

Carter et al. (1978), in a study of Tertiary volcanics from Scotland, obtained Rb–Sr and Sm–Nd "mantle isochrons" of ages 1.6 and 6.2 Ga, respectively. For obvious

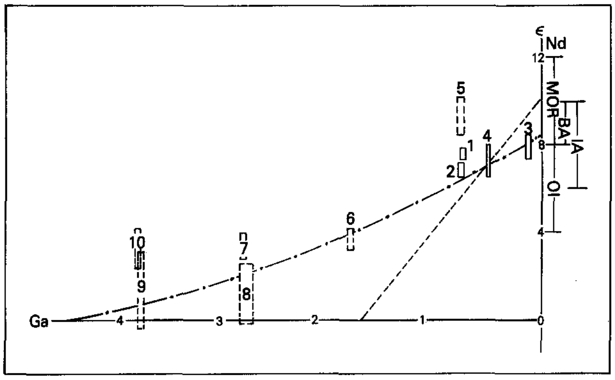

Fig. 6. ε_{Nd}-time diagram showing age on the abscissa and ε_{Nd} values, with ranges for different types of modern ocean crust (from McCulloch et al. 1981) on the ordinate. Data from the ophiolites reported in this paper are plotted along with data from other ophiolites (solid-line boxes) and from other rocks (dashed-line boxes). Symbols are as follows: *1*, Jabal al Wask; *2*, Jabal Ess; *3*, Samail ophiolite, Oman (McCulloch et al. 1981); *4*, Bay of Islands ophiolite, Newfoundland (Jacobsen and Wasserburg 1979); *5*, Matchless amphibolite, Namibia (Hawkesworth et al. 1981); *6*, Colorado Front Range crustal rocks (DePaolo 1981); *7*, Kambalda greenstone belt, Australia (McCulloch and Compston 1981); *8*, komatiites and tholeiites from Munrow township, Ontario (Zindler et al. 1978); *9*, metasediments and metavolcanics from West Greenland (Hamilton et al. 1983); *10*, gneisses from eastern India (Basu et al. 1981); *MOR*, mid-ocean ridges; *OI*, oceanic islands; *BA*, back-arc basins; *IA*, island arcs. Evolution lines are discussed in the text

reasons, they rejected the latter and, by inference, also the former date as chronologically meaningless. However, a more direct and reliable way to investigate the evolution of isotopic compositions in the mantle is to investigate the composition of old oceanic lithosphere: ophiolites. Such studies are likely to be more conclusive for older ophiolites. This study of the Jabal al Wask and Jabal Ess ophiolites takes the approach back to about 750 Ma. Younger ophiolites have been investigated, and for both the 500 Ma Bay of Islands ophiolite on Newfoundland (Jacobsen and Wasserburg 1979) and for the 100 Ma Samail ophiolite in Oman (McCulloch et al. 1981), the Nd data could be explained using the two-stage differentiation model discussed above.

The widely used T_{CHUR}^{Nd} model age (DePaolo and Wasserburg 1977) gives the time when a rock differentiated from CHUR, assuming a two-stage evolution. Calculation of T_{CHUR}^{Nd} ages for ophiolites requires an assumption about the Sm/Nd ratio in the depleted mantle from which the ophiolites were derived. Jacobsen and Wasserburg (1979), McCulloch et al. (1981), and Bokhari and Kramers (1981) have, on different grounds, argued for a $^{147}Sm/^{144}Nd$ ratio close to 0.24. Using this value, the three ophiolite samples have T_{CHUR}^{Nd} ages of 2090 Ma (175408), 2010 Ma (175403), and 1940 Ma (175578), which all are slightly but significantly higher than the 1.6 to 1.7 Ga suggested by the Rb — Sr and Pb — Pb mantle isochrons.

Another way of looking at these relations is in an ε_{Nd}-time diagram. In Fig. 6, the ε_{Nd} values for the Jabal al Wask and Jabal Ess ophiolites are plotted together with some other ε_{Nd} values from the literature. Shown on the ordinate are the ranges for different types of modern ocean crust, taken from the literature compilation of McCulloch et al. (1981). In this diagram, a reservoir with chondritic Sm/Nd and $^{143}Nd/^{144}Nd$ ratios, which is generally believed to approximate the bulk Earth ratios, would evolve with time along the horizontal line $\varepsilon_{Nd} = 0$. During a partial melting event in this chondritic reservoir, Nd would be more strongly partitioned into the melt than Sm. The unmelted mantle residue at the site of melting would have a higher Sm/Nd ratio than CHUR and would thereafter evolve along a practically straight line with poitive slope (its ε_{Nd} value would increase with time). If mantle differentiation were a more continuous process, its average composition would have evolved along a curve like the dot-dashed line in Fig. 6 (DePaolo 1981), rather than along straight lines.

In the two-stage mantle evolution model, ophiolites should plot along straight lines between $\varepsilon_{Nd} = 0$ at 1.6 to 1.7 Ga and modern ε_{Nd} values for the various mantle sources for ophiolites. The modern ε_{Nd} values for mantle sources are not well constrained, but assuming the Arabian ophiolites are of island-arc or back-arc origin, as discussed above, the best guess would be a value between 8 and 10, which is in the range of ε_{Nd} values for the few analyzed modern back-arc basalts (Hawkesworth et al. 1977; Carlson et al. 1978) and also falls within the upper range of island-arc values. A straight line between $\varepsilon_{Nd} = 0$ at 1.7 Ga and $\varepsilon_{Nd} = 10$ and 0 Ga (dashed line, Fig. 6) falls well below the Arabian ophiolites. Even if the modern value for the ophiolite sources is set to +12, the upper limit for mid-ocean ridge basalt, the Jabal al Wask point still lies above the line. Therefore, the ophiolite values are more consistent with a progressive differentiation of the type outlined by Tatsumoto (1978).

Also plotted in Fig. 6 are data for rocks from the Samail and Bay of Islands ophiolites; the 750 Ma Matchless amphibolite in Namibia (Hawkesworth et al. 1981); various 1.8-Ga rocks in the Colorado Front Range (DePaolo 1981), the 2.8-Ga Kambalda greenstone belt in Australia (McCulloch and Compston 1981); 2.8-Ga komatiitic and tholeiitic rocks of the Munrow Township, Ontario (Zindler et al. 1978); 3.8-Ga metasediments and metavolcanics from western Greenland (Hamilton et al. 1983); and 3.8-Ga gneisses from eastern India (Basu et al. 1981). If it is assumed that all these ε_{Nd} values reflect the values in contemporaneous mantle sources, it seems clear that depleted mantle has existed over most of the history of the Earth.

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