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## A NEODYMIUM, STRONTIUM, AND OXYGEN ISOTOPIC STUDY OF THE CRETACEOUS SAMAIL OPHIOLITE AND IMPLICATIONS FOR THE PETROGENESIS AND SEAWATER-HYDROTHERMAL ALTERATION OF OCEANIC CRUST

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Received August 17, 1979

Revised version received October 18, 1979

In the Samail ophiolite,  $^{147}\text{Sm}$ - $^{143}\text{Nd}$ ,  $^{87}\text{Rb}$ - $^{87}\text{Sr}$ , and  $^{18}\text{O}/^{16}\text{O}$  isotopic systems have been used to distinguish between sea-floor hydrothermal alteration and primary magmatic isotopic variations. The Rb-Sr and  $^{18}\text{O}/^{16}\text{O}$  isotopic systems clearly exhibit sensitivity to hydrothermal interactions with seawater while the Sm-Nd system appears essentially undisturbed. Internal isochrons have been determined by the  $^{147}\text{Sm}$ - $^{143}\text{Nd}$  method using coexisting plagioclase and pyroxene and give crystallization ages of  $130 \pm 12$  m.y. from Ibra and  $100 \pm 20$  m.y. from Wadi Fizh. These ages are interpreted as the time of formation of the Samail oceanic crust and are older than the inferred emplacement age of 65–85 m.y. The initial  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios for a tectonized harzburgite, cumulate gabbros, plagiogranite, sheeted dikes and a basalt have a limited range in  $\epsilon_{\text{Nd}}$  of from 7.5 to 8.6 for all lithologies, demonstrating a clear oceanic affinity and supporting earlier interpretations based on geologic observations and geochemistry. The  $^{87}\text{Sr}/^{86}\text{Sr}$  initial ratios on the same rocks have an extremely large range of from 0.70296 to 0.70650 ( $\epsilon_{\text{Sr}} = -19.7$  to  $+30.5$ ) and the  $\delta^{18}\text{O}$  values vary from 2.6 to 12.7. These large variations are clearly consistent with hydrothermal interaction of seawater with the oceanic crust. A model is presented for the closed system exchange of Sr and O, that in principle illustrates how the Sr isotopic data may be utilized to estimate the water/rock ratio and subsequently used to evaluate the temperature of equilibration between the water and silicates from the  $^{18}\text{O}/^{16}\text{O}$  water-rock fractionation.

### 1. Introduction

This report presents the results of a neodymium, strontium and oxygen isotopic study of a profile through Cretaceous oceanic crust, as represented by the Samail ophiolite, Oman. The Samail ophiolite consists of (from bottom to top) tectonized peridotite (dominantly harzburgite), layered gabbro, sheeted dikes, pillow basalts, and pelagic sediments. These rocks represent allochthonous fragments of oceanic crust and upper mantle that were emplaced upon the Arabian continental margin in the late Cretaceous [1–4]. The Samail ophiolite crops out in a desert region and, therefore, provides a unique opportunity

to study in detail an exceptionally well exposed and preserved cross section of oceanic crust. The characterization of oceanic crust is important as it presently represents over two thirds of Earth's surface and because nearly all of the oceanic crust created over Earth's history has been recycled into the mantle at subduction zones. This recycling of oceanic crust probably has had profound effects both on the geochemical evolution of the mantle and on the composition of island arcs and active continental margins. In addition, during the course of geologic time, the isotopic composition of oxygen and strontium in seawater has been affected strongly by hydrothermal interaction with the oceanic crust [5–8].

The aims of this present study are:

(1) To determine the crystallization age of this complex by using Sm-Nd mineral isochrons.

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Division Contribution No. 3309 (325).

(2) To determine the initial Nd and Sr ratios of the magma source which produced the Samail ophiolite and to compare these with possible modern-day analogues such as mid-ocean ridge basalts (MORB), island arcs and back arc basins.

(3) To compare the effects of seawater-hydrothermal alteration on the Nd, Sr and O isotopic systems. A more comprehensive report of the results of these studies will be published elsewhere [9,10].

Isotopic studies of a few other ophiolite complexes have been reported [11–19] with the Troodos massif of Cyprus at present being one of the better documented examples.  $^{18}\text{O}/^{16}\text{O}$  [13,14],  $^{87}\text{Sr}/^{86}\text{Sr}$  [11, 15,16] and trace and major elements studies [12,20–22] confirm its similarity to oceanic crust and identify the effects of seawater circulation in the pillow basalts, sheeted dikes, and uppermost gabbro complexes. As seawater also has largely altered the primary  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios, it has therefore not been possible to infer genetic relations between the different units within the complex using Sr isotopic data. A preliminary Nd isotopic study by Richard et al. [23] of single whole-rock gabbro samples from the Troodos, Samail, and several other ophiolite complexes has shown that the measured  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios are similar to those found in young ocean floor basalts.

## 2. Geology and sampling

The Samail ophiolite forms a major part of the 700 km long arcuate Oman Mountains on the eastern edge of the Arabian Peninsula (Fig. 1). Individual members of the ophiolite pseudostratigraphy are essentially internally undeformed [1] and have thicknesses comparable to geophysical estimates of modern-day oceanic crust [24].

The Samail ophiolite formed during Cretaceous spreading in the Hawasina ocean basin, which was a portion of the Tethys seaway [2,3]. A minimum age for the emplacement of the ophiolite is provided by Maestrichtian-Tertiary shallow-water limestones, which unconformably overlie parts of the ophiolite. A maximum age is given by Campanian sediments which underlie the ophiolite [3]. These constraints are consistent with K-Ar ages of  $83 \pm 5$  m.y. [25] from metamorphic biotites in amphibolites from the

contact aureole at the base of the Samail thrust sheet. Until now, no radiometric age determinations on any of the ophiolite members have been reported, although Glennie et al. [3] estimated a minimum crystallization age based on middle to late Cretaceous sediments which overlie the pillow lavas in northern Oman.

Samples representing the different lithologic units were chosen on the basis of their oxygen isotopic properties as described by Gregory and Taylor [9,26]. Most of these samples were collected from one of the standard ophiolite sections near Ibra as part of a cooperative study led by C.A. Hopson and R.G. Coleman. A detailed geologic map of the Ibra section in southeastern Oman Mountains has been prepared by Hopson et al. [27].

G54 is a microphyric basalt from the pillow lava section which exhibits igneous textures despite extensive recrystallization during hydrothermal alteration in the zeolite facies. The mineralogy and oxygen isotopic composition of the whole rock ( $\delta^{18}\text{O} = 12.7$ ), are consistent with low-temperature alteration by a hydrothermal fluid derived from seawater. G10 and K1 are from the sheeted dike complex; they have an ophitic texture and have been hydrothermally altered under greenschist facies conditions (K1 whole-rock  $\delta^{18}\text{O} = 6.8$ , and G10 whole-rock  $\delta^{18}\text{O} = 8.5$ ). G224-3 is a plagiogranite with a granophyric texture and consists of plagioclase, hornblende, quartz, magnetite and trace amounts of apatite. Hornblende is altered partially to chlorite. This plagiogranite was collected from an unchilled dike which cross-cuts the uraltite gabbro G224-2, approximately 500 m below the gabbro-diabase contact. The gabbro G224-2 contains poikilitic hornblende enclosing zoned plagioclase laths; alteration products include both chlorite and fibrous amphibole pseudomorphs after hornblende and rare retrograde prehnite. The mineralogy and oxygen isotope data of G224-2 (whole-rock  $\delta^{18}\text{O} = 3.7$ , plagioclase  $\delta^{18}\text{O} = 4.5$  and uraltite  $\delta^{18}\text{O} = 2.6$ ) suggest relatively high temperatures of alteration ( $>400^\circ\text{C}$ ). K9 is a sample from the cumulate gabbros and consists predominantly of cumulate plagioclase, clinopyroxene, and olivine. Trace amounts of brown hornblende are present as a minor alteration product of the clinopyroxene. This sample was chosen for Rb-Sr and Sm-Nd analysis as the mineralogy and oxygen data (whole-rock  $\delta^{18}\text{O} =$

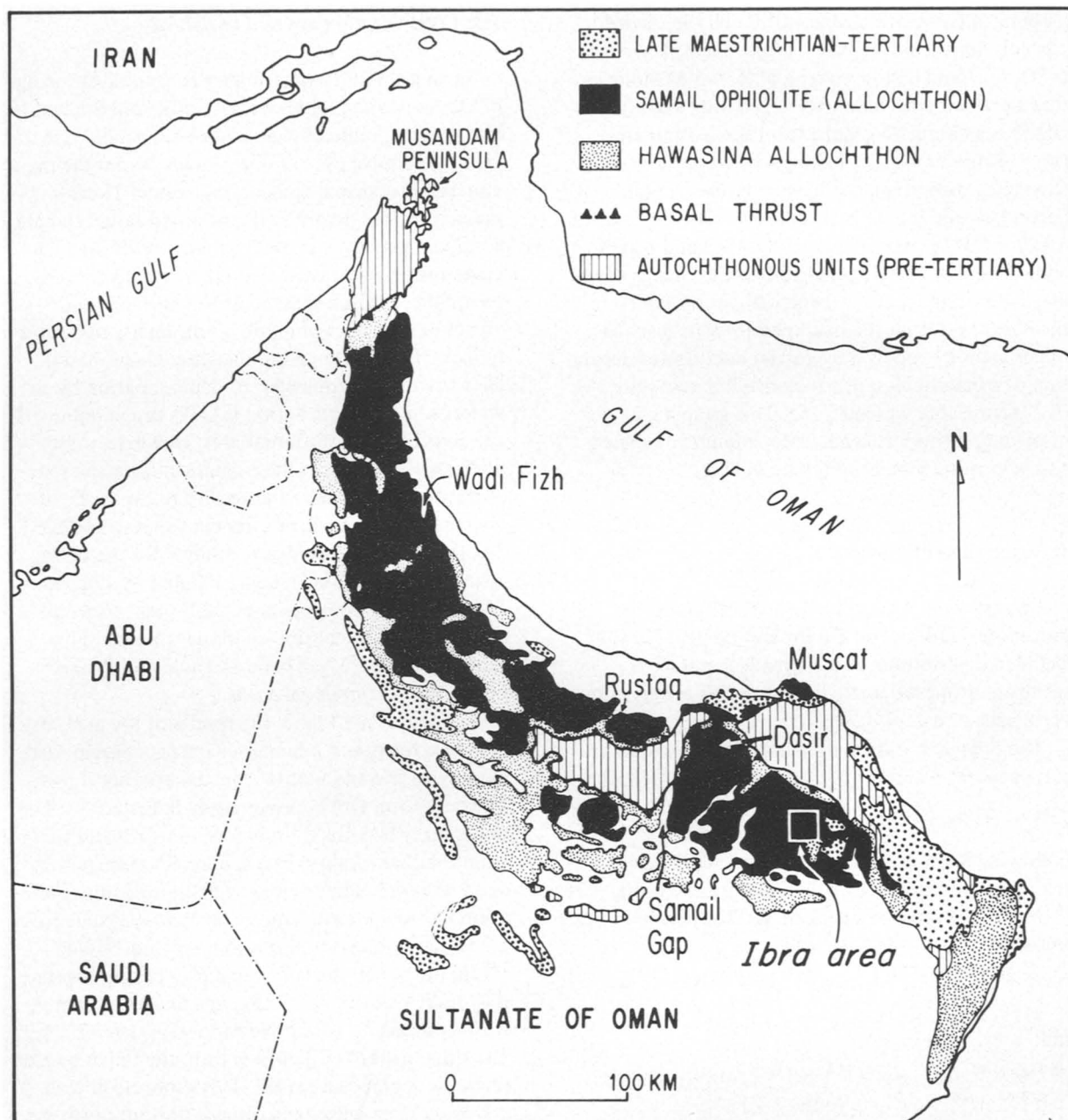


Fig. 1. Generalized geologic map of the Samail ophiolite, Oman (after Glennie et al. [3]), showing sample localities. The Samail ophiolite (solid) is an allochthonous body which is in thrust contact with the underlying rocks.

5.8, plagioclase  $\delta^{18}\text{O} = 6.0$  and clinopyroxene  $\delta^{18}\text{O} = 5.3$ ) indicate that this is one of the least altered samples.

To check for possible isotopic and age variations,

two additional gabbros (OM251 and OM28) were analyzed from widely spaced areas north of the Samail gap. OM251 is from Wadi Fizh (Fig. 1) and consists of cumulate plagioclase, clinopyroxene, and a minor

amount of cumulate olivine. Although the olivine is partially serpentinized, the oxygen data (plagioclase  $\delta^{18}\text{O} = 6.4$  and clinopyroxene  $\delta^{18}\text{O} = 5.8$ ) indicate that these minerals have exchanged only slightly. OM28 is a cumulate gabbro from the Rustaq area (Fig. 1) containing plagioclase and clinopyroxene. Olivine has been replaced by serpentine and talc. Secondary amphibole is also present. The oxygen data for OM28 (whole-rock  $\delta^{18}\text{O} = 5.2$  and plagioclase  $\delta^{18}\text{O} = 4.7$ ) indicate significant exchange of plagioclase and clinopyroxene with seawater. To test the hypothesis that the basal peridotite is residual upper mantle produced by partial melting and separation of a basaltic liquid, a tectonized harzburgite (K22) from Ibra was analyzed. This sample consists of olivine, orthopyroxene, trace amounts of spinel, and serpentine pseudomorphing olivine.

### 3. Results and discussion

Oxygen was extracted and analyzed using the procedures outlined by Taylor and Epstein [28]. Detailed descriptions of the chemical and mass spectrometric procedures for Nd and Sr are given by Papanastassiou et al. [29].

The oxygen results are reported in  $\delta$ -notation in parts per  $10^3$  where:

$$\delta^{18}\text{O} = [({}^{18}\text{O}/{}^{16}\text{O})_{\text{sample}}/({}^{18}\text{O}/{}^{16}\text{O})_{\text{standard}} - 1] 10^3$$

Following DePaolo and Wasserburg [30,31] an analogous notation is used for Nd with the initial  ${}^{143}\text{Nd}/{}^{144}\text{Nd}$  ratio of a rock of age  $T$  reported in the  $\epsilon$ -notation in parts per  $10^4$  where:

$$\epsilon_{\text{Nd}} = [({}^{143}\text{Nd}/{}^{144}\text{Nd})_{\text{INIT}}^T/({}^{143}\text{Nd}/{}^{144}\text{Nd})_{\text{CHUR}}^T - 1] 10^4$$

and:

$$({}^{143}\text{Nd}/{}^{144}\text{Nd})_{\text{CHUR}}^T = ({}^{143}\text{Nd}/{}^{144}\text{Nd})_{\text{CHUR}}^0 - ({}^{147}\text{Sm}/{}^{144}\text{Nd})_{\text{CHUR}}^0 (e^{\lambda_{\text{Sm}} T} - 1)$$

Present-day reference values are  $({}^{143}\text{Nd}/{}^{144}\text{Nd})_{\text{CHUR}}^0 = 0.511836$  and  $({}^{147}\text{Sm}/{}^{144}\text{Nd})_{\text{CHUR}}^0 = 0.1936$  [30];  $\lambda_{\text{Sm}} = 6.54 \times 10^{-12} \text{ yr}^{-1}$ .  $({}^{143}\text{Nd}/{}^{144}\text{Nd})_{\text{INIT}}^T$  is the measured ratio in the rock, corrected for decay since the time of crystallization  $T$ . An equivalent notation is used for Sr with  $({}^{87}\text{Sr}/{}^{86}\text{Sr})_{\text{UR}}^0 = 0.7045$ ,  $({}^{87}\text{Rb}/{}^{86}\text{Sr})_{\text{UR}}^0 = 0.0827$  [31,40], and  $\lambda_{\text{Rb}} = 1.42 \times 10^{-11} \text{ yr}^{-1}$ .

#### 3.1. Crystallization age and initial Nd

In an attempt to determine the crystallization age of the Samail ophiolite, we have analyzed Sm and Nd in coexisting mineral phases from the gabbros. In principle Sm-Nd ages can be obtained from the major minerals plagioclase and clinopyroxene. These minerals differ greatly in their Sm/Nd ratios and are common constituents of all gabbros, unaltered diabases and phyric basalts. Using other conventional techniques it often is difficult to determine the crystallization ages of mafic or ultramafic rocks due to lack of adequate parent-daughter fractionation, later thermal disturbances, or contamination by seawater. A possible exception is U-Pb zircon dating of siliceous late-stage differentiates. However, these bodies are volumetrically insignificant, not present everywhere, and their relationship to the mafic or ultramafic rocks is sometimes controversial [32,33]. The success of the Sm-Nd technique for the dating of mafic rocks has been demonstrated by Lugmair [34,35], Papanastassiou et al. [29] and Nakamura et al. [36] for meteoritic and lunar rocks and by Hamilton et al. [37,38] and DePaolo and Wasserburg [39] for terrestrial rocks.

In Table 1 and Fig. 2, the results of the analysis of whole rocks and mineral separates of plagioclase, clinopyroxene and uraltite from gabbros are shown. The data form two approximately linear arrays. The upper array is defined by two gabbros K9 and G224-2 from the Ibra locality. The gabbro K9 is extremely fresh with only clinopyroxene exhibiting minor alteration to brown hornblende. The clinopyroxene from K9 has a significantly higher Sm/Nd and  ${}^{143}\text{Nd}/{}^{144}\text{Nd}$  ratio than the coexisting plagioclase; together they indicate a crystallization age of  $128 \pm 20 \text{ m.y.}$  with an initial  ${}^{143}\text{Nd}/{}^{144}\text{Nd}$  ratio of  $\epsilon_{\text{Nd}} = 7.7 \pm 0.3$ . The other gabbro (G224-2) is from the upper part of the same gabbro unit as K9. This gabbro contains uraltite as a late-stage replacement of both pyroxene and primary hornblende. The uraltite has the highest  ${}^{147}\text{Sm}/{}^{144}\text{Nd}$  and  ${}^{143}\text{Nd}/{}^{144}\text{Nd}$  ratios for this rock, and together with the plagioclase gives an age of  $150 \pm 40 \text{ m.y.}$  and an initial  ${}^{143}\text{Nd}/{}^{144}\text{Nd}$  ratio of  $\epsilon_{\text{Nd}} = 7.8 \pm 0.4$ . The larger uncertainty in the age for this gabbro compared to K9 is due to the smaller difference in the  ${}^{147}\text{Sm}/{}^{144}\text{Nd}$  ratios for the plagioclase and uraltite. Both gabbros have the same age and

TABLE 1

Nd, Sr, and O isotopic data from the Samail ophiolite

Samples		Nd (ppm)	Sr (ppm)	$\frac{87\text{Rb}}{86\text{Sr}}$	$\frac{147\text{Sm}}{144\text{Nd}}$	$\frac{87\text{Sr}}{86\text{Sr}}$	$\frac{143\text{Nd}}{144\text{Nd}}$	$\epsilon_{\text{Sr}}$	$\epsilon_{\text{Nd}}$	$\delta^{18}\text{O}$
<i>Basalt</i>										
G54	WR	15.9	198	0.104	0.192	$0.70491 \pm 4$ *	$0.512219 \pm 19$ *	5.3 **	7.5 **	12.7
	calcite	0.126	6.12	0.004	0.209	$0.70650 \pm 15$	$0.512265 \pm 29$	30.5	8.1	—
<i>Sheeted dikes</i>										
G10		11.2	162	0.029	0.209	$0.70535 \pm 4$	$0.512250 \pm 25$	13.5	7.8	8.5
K1		13.0	388	0.008	0.194	$0.70519 \pm 4$	$0.512275 \pm 20$	11.7	8.6	6.8
<i>Plagiogranite</i>										
G224-3		17.3	192	0.035	0.187	$0.70362 \pm 4$	$0.512249 \pm 18$	-11.3	8.2	5.2
<i>Gabbros</i>										
G224-2	WR	8.36	196	0.010	0.201	$0.70370 \pm 3$	$0.512251 \pm 19$	-9.5	8.0	3.7
	plag	2.25	330	0.008	0.151	$0.70352 \pm 4$	$0.512196 \pm 18$	-12.0	7.8	4.5
	ural	9.51	58.7	0.014	0.241	$0.70426 \pm 3$	$0.512285 \pm 19$	-1.6	8.0	2.6
K9	WR	0.986	188	0.001	0.271	$0.70304 \pm 5$	$0.512310 \pm 18$	-18.6	8.0	5.7
	plag	0.399	311	0.001	0.129	$0.70296 \pm 2$	$0.512180 \pm 25$	-19.7	7.8	6.0
							$0.512181 \pm 19$ ***		7.8	
	cpx	2.630	24.8	0.001	0.322	$0.70313 \pm 3$	$0.512341 \pm 15$	-17.3	7.7	5.3
OM251	plag	0.081	156	0.001	0.159	$0.70311 \pm 5$	$0.512203 \pm 22$	-18.0	7.6	6.4
	cpx	0.411	54.2	0.002	0.377	$0.70315 \pm 5$	$0.512346 \pm 25$ ***	-17.5	7.6	5.8
OM28	WR	0.385	132	0.001	0.264	$0.70383 \pm 5$	$0.512274 \pm 22$ ***	-7.8	7.7	5.2
<i>Harzburgite</i>										
K22		0.035	0.43	0.150	0.216	$0.70313 \pm 9$	$0.512278 \pm 23$	-21.2	8.3	—

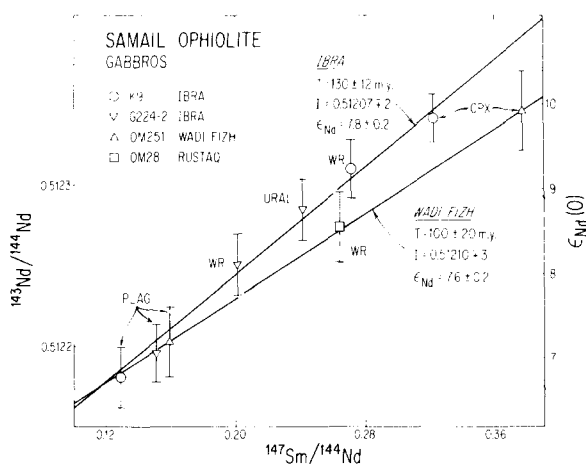
\* Errors are in the last figures given and represent  $\pm 2\sigma$  mean. These errors correspond to typical uncertainties in  $\epsilon_{\text{Sr}}$ , and  $\epsilon_{\text{Nd}}$  of  $\pm 0.6$  and  $\pm 0.4$   $\epsilon$  units, respectively. The uncertainty in  $\delta^{18}\text{O}$  is  $\pm 0.1\%$ .

\*\*  $\epsilon_{\text{Sr}}$  and  $\epsilon_{\text{Nd}}$  values calculated for a crystallization age of 130 m.y. except for OM251 and OM28 for which an age of 100 m.y. was used.

\*\*\*  $^{143}\text{Nd}/^{144}\text{Nd}$  determined from spiked sample.

initial ratio within analytical uncertainty, and the Sm-Nd systematics do not appear to be disturbed by the presence of hydrous alteration products such as uraltite. The combined data are shown in Fig. 2 and give a crystallization age of  $130 \pm 12$  m.y. and  $\epsilon_{\text{Nd}} = 7.8 \pm 0.2$ .

Fig. 2. Sm-Nd evolution diagram showing the data for whole rocks and coexisting pyroxene-plagioclase and uraltite-plagioclase pairs for gabbros from Ibra, Wadi Fizh and Rustaq. The gabbros from Ibra give an age of  $130 \pm 12$  m.y. and  $\epsilon_{\text{Nd}} = 7.8 \pm 0.2$ . The gabbro from north of the Samail gap gives an age of  $100 \pm 20$  m.y. and  $\epsilon_{\text{Nd}} = 7.6 \pm 0.2$ . These ages are considered to be the times of formation of the oceanic crust that was formed in the Hawasina ocean basin and is now preserved as the Samail ophiolite.



The lower array shown in Fig. 2 is defined by the gabbros OM251 and OM28. These two gabbros were collected from Wadi Fizh and Rustaq, which are located north of the Samail gap (Fig. 1). The clinopyroxene and plagioclase from OM251 give an age of  $100 \pm 20$  m.y. and an  $\epsilon_{\text{Nd}} = 7.6 \pm 0.3$ . Although the isochron ages from Ibra and Wadi Fizh overlap within analytical uncertainty, the Sm-Nd results suggest that there may be an age difference of approximately 30 m.y. between these two areas. More precise measurements will be required for this to be established definitively, but the present results clearly indicate that these rocks crystallized in the time interval from 100 to 130 m.y. These crystallization ages are greater than the time of emplacement of the Samail ophiolite onto the Oman continental margin which occurred between 65 and 85 m.y. [25]. This suggests that relatively mature ( $>15$  m.y. old) oceanic crust was obducted.

The initial  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios from the gabbros, together with those determined from a plagiogranite, sheeted dikes, basalt and a tectonized harzburgite from the peridotite, are shown in the histogram in Fig. 3. The latter samples are all from the Ibra area and their initial ratios were calculated assuming a crystallization age of 130 m.y. Most of the samples have  $\epsilon_{\text{Nd}}$  values within an extremely narrow range of from 7.5 to 8.2 (Table 1), indicating derivation from an isotopically uniform light-REE-depleted (depleted in Nd relative to Sm) source region. The sheeted dike K1 has  $\epsilon_{\text{Nd}} = 8.6 \pm 0.4$  compared to the mean  $\epsilon_{\text{Nd}}$  of  $7.8 \pm 0.2$  for the Ibra gabbros. This small deviation in  $\epsilon_{\text{Nd}}$  of K1 cannot be explained by seawater alteration as more altered samples have consistent  $\epsilon_{\text{Nd}}$  values, but it suggests the presence of small but significant isotopic heterogeneities in the oceanic mantle magma reservoirs.

The initial value of  $\epsilon_{\text{Nd}} = 8.3 \pm 0.4$  obtained from the tectonized harzburgite is also, within error, the same as the rest of the ophiolite. This result, together with the low trace element concentrations (Table 1), supports the hypothesis that the harzburgite is probably a cogenetic upper mantle residue produced by partial melting and removal of a basaltic liquid from which the rest of the ophiolite section formed.

In Fig. 3 the  $\epsilon_{\text{Nd}}$  values from the Samail ophiolite are compared with those found in MORB (+7 to +13), marginal basins (+8 to +9), island arcs (+7 to

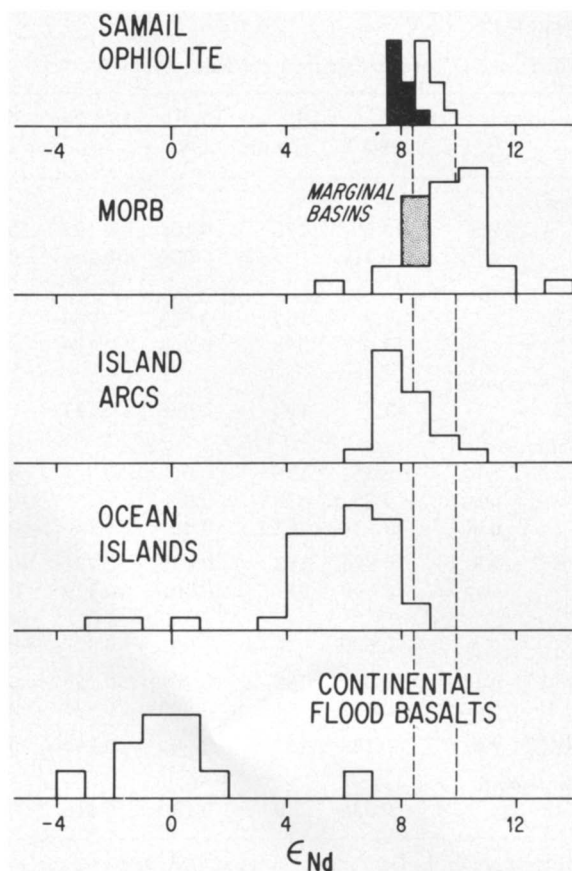


Fig. 3. Histogram comparing initial  $\epsilon_{\text{Nd}}$  values from the Samail ophiolite measured in this study with modern-day mid-ocean ridge basalts (MORB) [30,31,40–42], marginal basins (Lau, Scotia Sea) [42,44], island arcs [43,44], ocean islands [31,40,41,45], and continental flood basalts [30,31,42,46]. The small range in  $\epsilon_{\text{Nd}}$  for marginal basins (shaded) may be an artifact of the limited data (the smallest vertical division represents one sample). The Samail ophiolite clearly has an oceanic affinity. The initial  $\epsilon_{\text{Nd}}$  values of the Samail ophiolite (solid) give an average of  $\epsilon_{\text{Nd}} = 7.9$ , which is somewhat lower than typical modern-day MORB with  $\epsilon_{\text{Nd}} \approx 10$ . Correcting for differential evolution in the Samail ophiolite source region relative to the bulk earth during the past 130 m.y. increases the  $\epsilon_{\text{Nd}}$  values of the Samail ophiolite (delimited by dashed lines) to give a modern-day average of  $\epsilon_{\text{Nd}} = 8.8$ , which is well within the MORB and marginal basin range.

+10), ocean islands (+4 to +8) and continental basalts ( $\sim 0$ ). The  $\epsilon_{\text{Nd}}$  values from the Samail ophiolite clearly have an oceanic affinity. The Nd isotopic data alone cannot be used to distinguish definitively

among the various possible tectonic settings for this ophiolite because of the overlap in  $\epsilon_{\text{Nd}}$  between MORB, marginal basins, island arcs and to a lesser degree ocean islands (Fig. 3). However, the lack of light-REE-enriched samples [10,47] and of widespread alkalic volcanism usually associated with ocean island and island arc environments plus geologic arguments [27] suggests that the most reasonable scenario for the formation of the Samail ophiolite is at a ridge axis spreading center in an ocean basin.

The initial Nd value of the Samail ophiolite with an average of  $\epsilon_{\text{Nd}} = 7.9 \pm 0.2$ , is somewhat lower than the "typical" MORB with  $\epsilon_{\text{Nd}} \approx 10$ . To some extent this is due to its older age compared to the zero age MORB samples shown in Fig. 3. A more exact comparison requires correction for the evolution of Nd in the Samail ophiolite source region during the past 130 m.y. Assuming that the composite of the cumulate gabbros, diabase dikes, and pillow basalts represents the primary unfractionated magma of the Samail ophiolite, we estimate [10] that this magma had  $^{147}\text{Sm}/^{144}\text{Nd} = 0.24 \pm 0.02$ . This also probably represents the  $^{147}\text{Sm}/^{144}\text{Nd}$  ratio in the source region, as both Nd and Sm are strongly partitioned into the melt during partial melting. For this value of  $^{147}\text{Sm}/^{144}\text{Nd}$ , the Samail ophiolite source region today would have  $\epsilon_{\text{Nd}} \approx 8.8$  which is only 1  $\epsilon_{\text{Nd}}$  unit below the center of the present MORB distribution and is therefore fully consistent with a MORB affinity. The lower  $\epsilon_{\text{Nd}}$  value of the Samail ophiolite compared to typical MORB indicates derivation from a slightly less light-REE-depleted oceanic, mantle source.

### 3.2. Comparison of Nd and Sr

A striking contrast exists between the Sr and Nd isotopic systems, with an extremely large range in  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios from 0.70296 ( $\epsilon_{\text{Sr}} = -19.7$ ) to 0.70650 ( $\epsilon_{\text{Sr}} = +30.5$ ) compared to a narrow range in  $\epsilon_{\text{Nd}}$  values of from only 7.5 to 8.6. The highest  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios are found in the petrographically more altered rocks (basalts, sheeted dikes, high-level gabbros, and plagiogranites). This approximate correlation between high  $^{87}\text{Sr}/^{86}\text{Sr}$  and stratigraphic height (see Fig. 5) indicates that these rocks exchanged with Cretaceous seawater, the only geologically reasonable source of high  $^{87}\text{Sr}/^{86}\text{Sr}$ .

In Fig. 4 the  $\epsilon_{\text{Nd}}$  and  $\epsilon_{\text{Sr}}$  values from the Samail

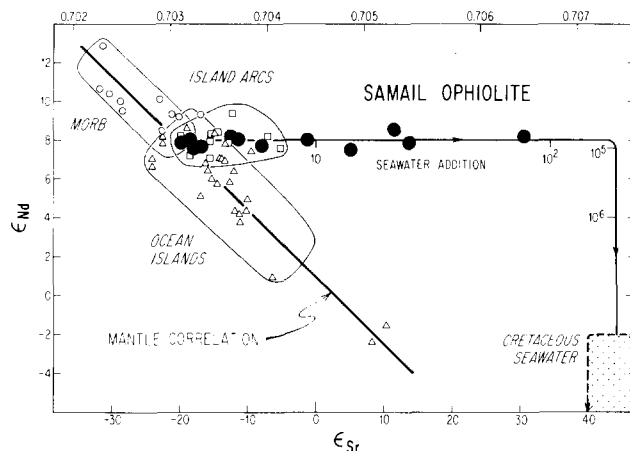


Fig. 4.  $\epsilon_{\text{Nd}}$  and  $\epsilon_{\text{Sr}}$  values for the Samail ophiolite. The arrows indicate the effect of contamination with Cretaceous seawater for different water/rock ratios and shown that  $W/R > 10^5$  is required before  $\epsilon_{\text{Nd}}$  values are affected. The  $\epsilon_{\text{Nd}}$  and  $\epsilon_{\text{Sr}}$  values of uncontaminated samples plot on the "mantle correlation" line [31,40] and are consistent with derivation from a less depleted MORB source region. The  $\epsilon_{\text{Nd}}$  and  $\epsilon_{\text{Sr}}$  values plot in a position where the fields for MORB and island arc basalts overlap, and hence do not discriminate between the oceanic environments in which the Samail ophiolite formed.

ophiolite are plotted. They have an origin at the mantle correlation line [31,40] and produce a horizontal trajectory to higher  $\epsilon_{\text{Sr}}$  values. Although this general effect has been shown by previous workers [40,43], the present results provide a spectacular example of how the Nd isotopic composition is unaffected by seawater contamination, despite the large shifts in the Sr isotopic composition. To illustrate further the effects of seawater addition, in Fig. 4, a mixing line for Sr and Nd is shown for the closed system exchange of seawater with oceanic crust for various water/rock ratios. For closed system exchange of seawater and rock the Sr water/rock ratio (by weight) is given by:

$$\frac{W}{R} = \left( \frac{\epsilon_{\text{rock}}^f - \epsilon_{\text{rock}}^i}{\epsilon_{\text{H}_2\text{O}}^i - \epsilon_{\text{rock}}^f} \right) \frac{C_{\text{rock}}^i}{C_{\text{H}_2\text{O}}^i} \quad (1)$$

where  $i$  = initial value,  $f$  = final value,  $C_{\text{rock}}^i$  = concentration of Sr in rock and  $C_{\text{H}_2\text{O}}^i$  = concentration of Sr in seawater. The parameters used in this equation are  $\epsilon_{\text{rock}}^i = -20$ ,  $\epsilon_{\text{H}_2\text{O}}^i = +47$  [7,8], and  $C_{\text{H}_2\text{O}}^i = 8$  ppm [48].  $\epsilon_{\text{rock}}^f$  values are given in Table 1. It is

assumed that the initial Sr concentration in the rock ( $C_{\text{rock}}^i$ ) is the same as the final measured concentration (Table 1). With these parameters,  $W/R$  ratios of 15–40 are required to account for the observed  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios in the sheeted dikes and basalt. For the gabbros lower  $W/R$  ratios of 0–4 are indicated, suggesting that  $W/R$  ratios increase upward in the section. The significantly lower concentration of Nd in seawater of  $3 \times 10^{-6}$  ppm [48,49] requires a  $W/R$  ratio exceeding  $2 \times 10^5$  to produce a measurable shift in  $\epsilon_{\text{Nd}}$ .

Due to the overprinting of rock Sr with seawater Sr, it is not possible to determine unambiguously the primary magmatic  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio in these rocks. An upper limit is given by the lowest  $^{87}\text{Sr}/^{86}\text{Sr}$  value of  $0.70296 \pm 2$  found in the plagioclase from the gabbro K9. This may be close to its magmatic value as the coexisting clinopyroxene with a lower Sr concentration has only a slightly greater  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio of  $0.70313 \pm 3$ . The  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios in the clinopyroxene and plagioclase from the gabbro OM251 also have very different Sr concentrations but have the same  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio within error ( $0.70315 \pm 5$  and  $0.70311 \pm 5$ ), and may also represent a primary value. The difference between the plagioclase separates from K9 and OM251, may suggest a small but real difference in the  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios of these samples and of the original magmas from which they crystallized. This small  $^{87}\text{Sr}/^{86}\text{Sr}$  isotopic heterogeneity is also consistent with the  $\epsilon_{\text{Nd}}$  values for these two gabbros as both plot right on the correlation line in the  $\epsilon_{\text{Sr}}-\epsilon_{\text{Nd}}$  diagram of Fig. 4. The correlation line shown in Fig. 4 is defined by the  $\epsilon_{\text{Nd}}$  and  $\epsilon_{\text{Sr}}$  values of MORB, island arcs and ocean islands [30,31,40–45]. This correlation of  $\epsilon_{\text{Nd}}$  and  $\epsilon_{\text{Sr}}$  has been interpreted as an important mantle feature [31,40,45]. For example, the correlation may represent mixing between two reservoirs [31]; an undepleted “continental”-type mantle with  $\epsilon_{\text{Nd}} \sim 0$  and  $\epsilon_{\text{Sr}} \sim 0$  and a LIL-depleted “oceanic”-type mantle with  $\epsilon_{\text{Nd}} = +10$  and  $\epsilon_{\text{Sr}} = -27$ .

### 3.3. Comparison of Sr and $^{18}\text{O}$

The data presented in Fig. 5 and in Table 1 show that: (1) measured  $^{87}\text{Sr}/^{86}\text{Sr}$  in altered samples shifts from the primary magmatic value of  $\sim 0.7030$  towards the higher values of Cretaceous seawater of 0.7076 [7,8] and (2)  $\delta^{18}\text{O}$  values of samples above

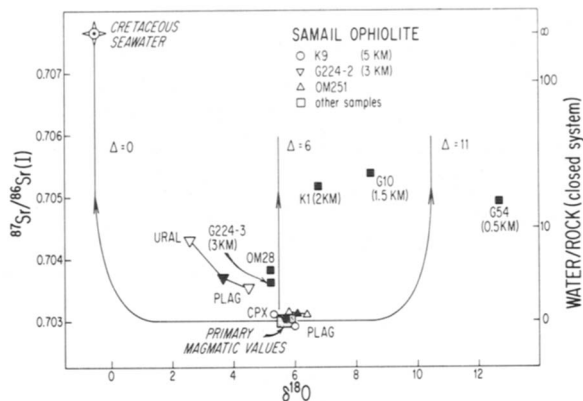


Fig. 5. Initial  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $\delta^{18}\text{O}$  values of whole rocks (solid) and minerals (open) from the Samail ophiolite. The approximate stratigraphic depth is given in parentheses. The curves describe the  $\delta^{18}\text{O}$  and  $^{87}\text{Sr}/^{86}\text{Sr}$  values in mixtures of seawater and oceanic crust for different temperatures of  $^{18}\text{O}$  exchange. The oxygen water-rock fractionation factor  $\Delta$  is temperature dependent ( $\Delta \propto 1/T^2$ ), being approximately equivalent to zeolite facies for  $\Delta = 11$ , to greenschist facies for  $\Delta = 6$ , and to upper amphibolite facies for  $\Delta = 0$ . The water/rock ratios on the right ordinate are calculated assuming closed system exchange of Sr (equation 1) in Cretaceous seawater ( $^{87}\text{Sr}/^{86}\text{Sr} = 0.7076$ ,  $\text{Sr} = 8$  ppm) with Sr in oceanic crust ( $^{87}\text{Sr}/^{86}\text{Sr} = 0.7030$ ,  $\text{Sr} = 160$  ppm). The arrows on the mixing lines show the direction of increasing  $W/R$  ratios. These lines illustrate that  $W/R$  ratios are higher while alteration temperatures are lower for the diabases and basalts (K1, G10, G54). The curves also show that the gabbros (G224-2, OM28) were altered at higher temperatures and lower  $W/R$  ratios. In principle, the Sr and O isotopic systems can be used together as a geothermometer for hydrothermal interactions.

the diabase-gabbro contact are enriched (relative to the primary magmatic value), whereas samples below the diabase-gabbro contact have depleted  $\delta^{18}\text{O}$  values. These data, therefore, indicate that both the Sr and  $^{18}\text{O}$  isotopic systems have been affected by exchange with a seawater-derived hydrothermal fluid. The major difference between the two systems is that during hydrothermal exchange, only oxygen isotopes are significantly affected by temperature-dependent fractionation. Oxygen water-rock fractionation factor  $\Delta$  is defined as  $\Delta = \delta^{18}\text{O}_{\text{rock}} - \delta^{18}\text{O}_{\text{water}}$  and is temperature dependent ( $\Delta \propto 1/T^2$ ). For zeolite facies alteration,  $\Delta > 6$ , and for greenschist or higher facies alteration,  $\Delta < 6$  [50]. As extremely steep thermal gradients are present near the ridge axis spreading



center, a wide range of  $\Delta$ 's ( $-2 < \Delta < +20$ ) is possible. In addition, since seawater is only 6‰ higher in  $\delta^{18}\text{O}$  than the initial magmatic reservoir, the changes in the alteration temperature regimes at active spreading environments can produce both  $^{18}\text{O}$  enrichments or depletions [9].

For closed system exchange of seawater and rock, the oxygen water/rock ratio (by weight) is given by [50]:

$$\frac{W}{R} = \left( \frac{\delta_{\text{rock}}^f - \delta_{\text{rock}}^i}{\delta_{\text{H}_2\text{O}}^i - (\delta_{\text{rock}}^f - \Delta)} \right) \frac{C_{\text{rock}}^i}{C_{\text{H}_2\text{O}}^i} \quad (2)$$

This is identical to the Sr exchange equation (equation 1) apart from the fractionation factor  $\Delta$ . In principle the combined Sr and O approach may be used as a geothermometer with the  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios being utilized to estimate the  $W/R$  ratio. Using this  $W/R$  ratio and the  $\delta^{18}\text{O}$  values, the fractionation factor  $\Delta$  and hence the final temperature of water-rock equilibration can be evaluated. This approach may be particularly useful in systems with relatively low water/rock ratios where equilibrium closed system exchange prevails and the Sr and O water-rock exchange rates are probably similar. Mixing lines can be calculated by combining equations (1) and (2) [10]. The effect of changing  $\Delta$  on the calculated mixing lines is shown in Fig. 5. Since there are many different alteration temperature regimes, no simple correlation between  $\delta^{18}\text{O}$  and  $^{87}\text{Sr}/^{86}\text{Sr}$  would be expected for the entire ophiolite.

#### 4. Conclusions

Using the Sm-Nd method, internal isochrons were obtained from three gabbros establishing a Cretaceous crystallization age for the Samail ophiolite. These results appear to be highly consistent in spite of extensive hydrothermal exchange which has altered the primary  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $^{18}\text{O}/^{16}\text{O}$  ratios. Gabbros from Ibra give an age of  $130 \pm 12$  m.y., and a gabbro in the northern part of Oman gives an age of  $100 \pm 20$  m.y. These results show that the Sm-Nd technique can be used to determine crystallization ages of relatively young, Mesozoic and Cenozoic mafic complexes.

The initial  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios for all of the major lithologic units of the Samail ophiolite have a narrow

range in  $\epsilon_{\text{Nd}}$  values indicating that they are derived from a common mantle source which has been depleted in Nd relative to Sm (light REE depleted) for at least a billion years or more. These data clearly demonstrate an oceanic affinity for this complex. The Nd and Sr isotopic data cannot by themselves be used to distinguish definitively among the various possible settings such as mid-ocean ridge, back-arc basin, or island arc.

In contrast to the small range in initial Nd with  $\epsilon_{\text{Nd}} = 7.5$  to 8.6, the same samples have an extremely large range in initial  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $\delta^{18}\text{O}$  with  $^{87}\text{Sr}/^{86}\text{Sr}(\text{I}) = 0.7030$ – $0.7065$  and  $\delta^{18}\text{O} = +2.6$  to 12.7. These large variations in both Sr and O are consistent with hydrothermal exchange of oceanic crust with seawater initially having a high  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio and  $\delta^{18}\text{O} \approx 0$ . The initial  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios generally decrease downward suggesting that the ratio of seawater to rock decreases with depth. As the  $^{18}\text{O}$  shifts depend upon  $W/R$  ratio and temperature, both  $^{18}\text{O}$ -depleted and  $^{18}\text{O}$ -enriched rocks have been produced. Utilizing a relatively simple model for relating the exchange of Sr and O, it may be possible to use the Sr data to estimate  $W/R$  ratios and subsequently to evaluate the temperature of equilibration between the water and silicates by using the  $^{18}\text{O}/^{16}\text{O}$  water-rock fractionation.

#### Acknowledgements

Field work was conducted in collaboration with the National Science Foundation–U.S. Geological Survey Oman project organized by R.G. Coleman and C.A. Hopson. Their help and support have been invaluable. This work has benefited from discussions with R.G. Coleman, C.A. Hopson, E.H. Bailey, J.S. Pallister, J. Chen, and M.A. Lanphere. R.G. Coleman provided additional samples from northwest of Ibra which greatly broadened the scope of this study. S.B. Jacobsen and the reviewers provided helpful comments on the manuscript. We appreciate Marvin Lanphere's indulgence in matters regarding an easily exchanged element. We express our gratitude to the Directorate General of Petroleum and Minerals, Sultanate of Oman, particularly Mohammed Kassim and Ismail M. El Boushi, for their support of this project. This work has been supported by N.S.F. grants PHY

76-83685, EAR-76-21310, EAR-7816874, and also indirectly by an N.S.F. grant to C.A. Hopson.

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