Sm-Nd, Rb-Sr, and $^{18}$O/$^{16}$O Isotopic Systematics in an Oceanic Crustal Section: Evidence From the Samail Ophiolite

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The Sm-Nd, Rb-Sr, and $^{18}$O/$^{16}$O isotopic systems have been used to distinguish between the effects of seafloor hydrothermal alteration and primary magmatic isotopic variations. The Sm-Nd isotopic system is essentially unaffected by seawater alteration, while the Rb-Sr and $^{18}$O/$^{16}$O systems are sensitive to hydrothermal interactions with seawater. Sm-Nd mineral isochrons from the cumulative gabbros of the Samail ophiolite have an initial $^{143}$Nd/$^{144}$Nd ratio of $e_{Nd} = 7.8 \pm 0.3$, which clearly substantiates the oceanic affinity of this complex. The initial $^{143}$Nd/$^{144}$Nd ratios for the harzburgite, plagiogranite, sheeted diabase dikes, and basalt units have a limited range in $e_{Nd}$ of from 7.5 to 8.6, indicating that all the lithologies have distinctive oceanic affinities, although there is also some evidence for small isotopic heterogeneities in the magma reservoirs. The Sm-Nd mineral isochrons give crystallization ages of 128 ± 20 m.y. and 150 ± 40 m.y. from Ibra and 100 ± 20 m.y. from Wadi Fizh, which is approximately 300 km NW of Ibra. These crystallization ages are interpreted as the time of formation of the oceanic crust. The $^{87}$Sr/$^{86}$Sr initial ratios on the same rocks have an extremely large range of from 0.7030 to 0.7065 and the $\delta^{18}$O values vary from 2.6 to 12.7. These large variations clearly demonstrate hydrothermal interaction of oceanic crust with seawater.

INTRODUCTION

The purpose of this study is to compare the neodymium, strontium, and oxygen isotopic characteristics of the Samail ophiolite, Oman. These rocks are of significance, as it is now generally recognized that ophiolites represent transported slabs of oceanic crust and upper mantle emplaced upon continental margins. The Samail ophiolite provides a unique opportunity to study in detail a well-exposed and preserved section of oceanic crust. The Samail, unlike most other 'dismembered' ophiolites, is remarkably intact with a thick arcuate Oman Mountains which are on the eastern edge of the Arabian Peninsula (Figure 1). The samples analyzed in this study mainly come from the Ibra ophiolite section south of the Arabian Peninsula (Figure 1). A minimum age for the emplacement of the ophiolite is provided by Maastrichtian-Tertiary shallow water limestones which unconformably overlie parts of the ophiolite. A maximum age is given by K-Ar ages of ~90 m.y. (Turonian) [Lanphere, 1981] from metamorphic biotites in amphibolites from the detachment aureole at base of the Samail thrust sheet. Until now (see also Tilton et al. [1981]) evidence for the age of production of the oceanic crust is sparse, although Glennie et al. [1974] estimated a minimum age limit based on middle to


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late Cretaceous (Cenomanian) sediments which overlie the pillow lavas in northern Oman.

Sample Descriptions

The stratigraphic locality of the Ibra samples is shown in Figure 1.

**Basalt G54 (Ibra).** Microphyric basalt with acicular laths of plagioclase (An2). Groundmass is altered to palagonite, zeolites (?), saponite (?), and calcite. The calcite is present as 1–2 cm amygdules and also as smaller blebs dispersed throughout the rock. A whole rock sample which was free of large amygdules and calcite from an amygdule were analyzed. The mineralogy and oxygen isotope data (whole rock 8[^18]O = 12.7) are consistent with low temperature alteration (100°–200°C).

**Diabase dike G10 (Ibra).** Ophitic texture with turbid albitized plagioclase (An20). Alteration products include epidote, chlorite, antinolite-tremolite pseudomorphs after pyroxene, quartz, carbonate, and leucoxene. The mineralogy indicates alteration in the greenschist facies.

**Diabase dike K1 (Ibra).** Ophitic texture with turbid plagioclase (An30), clinopyroxene, and opaques. Alteration products include chlorite, uralite, and quartz and suggest alteration at greenschist facies.

**Plagiogranite G224-3 (Ibra).** Granophyric texture with zoned plagioclase (core An30, rim An20), hornblende, quartz, magnetite, and trace amounts of apatite. Hornblende is partially altered to uralite. The plagiogranite was collected from a nonchilled dike which cross-cuts uralite gabbro, G224-2, approximately 500 m below the gabbro-diabase contact.

**Uralite gabbro G224-2 (Ibra).** Poikilitic hornblende enclosing zoned plagioclase (core An40, rim An20) and opaques. Alteration products include fibrous amphibole pseudomorphs after hornblende, chlorite, and rare prehnite. The mineralogy and oxygen isotope data (whole rock 8[^18]O = 3.7, plagioclase 8[^18]O = 4.5, and uralite 8[^18]O = 2.6) suggest relatively high temperatures of alteration (>400°C).

**Gabbro K9 (Ibra).** Adcumulate with igneous lamination; 72% plagioclase (An75), 20% clinopyroxene, 7% olivine, and 1% hornblende. The brown hornblende is a minor alteration product of the clinopyroxene. The mineralogy and oxygen isotope data (whole rock 8[^18]O = 5.8, plagioclase 8[^18]O = 6.0, and clinopyroxene 8[^18]O = 5.3) indicate that this is one of the least altered samples.

**Gabbro OM251 (Wadi Fizh, northern Oman).** Late cumulate pyroxene poikilitically enclosing plagioclase; 77% plagioclase (An20), 20% clinopyroxene, and 3% olivine. The olivine has been partially serpentinized. Although the olivine is serpentinized, the oxygen data (plagioclase 8[^18]O = 6.4 and clinopyroxene 8[^18]O = 5.8) indicate that the plagioclase and clinopyroxene have exchanged only slightly with an 18O shifted seawater-derived hydrothermal fluid.

**Gabbro OM28 (Rustaq).** Adcumulate with igneous lamination; 72% plagioclase (An75), 15% clinopyroxene, 10% am-
phibole, and 2% serpentine, and/or tale alteration pseudomorphing olivine. In contrast to the other cumulate gabbros, the oxygen isotope data from this sample (plagioclase $\delta^{18}O = 4.7$ and whole rock $\delta^{18}O = 5.2$) indicate significant exchange of plagioclase and pyroxene with a seawater-derived fluid.

**Harzburgite K22 (Ibra).** Porphyroclastic texture with tektomite fabric; 25% orthopyroxene, 45% olivine, 30% serpentine, and trace amounts of spinel. Serpentine is pseudomorphing olivine.

**Experimental Procedures**

Samples were prepared by removing exterior surfaces with a rock splitter and crushing interior chunks. Only mechanical means were used for mineral separation and involved a combination of handpicking and magnetic separation. These procedures produced high purity ($\approx 100\%$) plagioclase separates and pyroxene separates of greater than 95% purity. The small amount of impurity in the pyroxene separate generally consisted of olivine which is unimportant as it contains extremely low trace element concentrations.

Oxygen was extracted and analyzed from approximately 20 mg of sample using the procedures outlined by Taylor and Epstein [1962]. For Nd and Sr approximately 200 mg of sample was dissolved using HF and HClO$_4$. After complete dissolution an aliquot consisting of about 10% of the solution was spiked and concentrations of K, Rb, Sr, Sm, and Nd were determined. From approximately 100 mg of the remaining unspiked solution, Nd and Sr were separated using ion exchange chemistry. For several samples (see Table 1) both Sm and Nd concentrations and $^{144}$Nd/$^{144}$Nd ratios were obtained from the same spiked sample. For the plagioclase separate from the gabbro K9, the $^{144}$Nd/$^{144}$Nd ratio was obtained from both spiked and unspiked samples and was the same within experimental error. The Sr isotopic composition was determined using a single oxidized tantalum filament, and typically $\approx 150$ ratios were collected. The Nd isotopic composition was determined by measuring NdO$^+$ on a single rhenium filament and normalized to $^{150}$Nd/$^{144}$Nd = 0.2096. The $2\sigma$ (mean) errors quoted for the $^{144}$Nd/$^{144}$Nd ratio were generally obtained from about 250 ratios. For the above procedures the blanks were Rb = 0.01 ng, Sr = 0.1 ng, Sm = 0.006 ng, and Nd = 0.02 ng and are negligible for the data presented here.

For the harzburgite the lower Sr and Nd concentrations required the use of a slightly modified procedure to enable the precise determination of the isotopic compositions. Approximately 400 mg of the harzburgite was dissolved and split into three equal portions after removing a 10% aliquot for concentrations. The three portions were passed separately through a cation column, and Sr and rare earths were separated. The rare earth fractions were combined and loaded on the lactic acid column, and Nd was separated. Using this procedure, a total of 10 ng of Nd was obtained from the harzburgite K22, with a Nd blank of $\approx 0.06$ ng. This blank is less than 0.6% of the total Nd and is therefore still insignificant. For 10 ng of Nd the precision of the $^{144}$Nd/$^{144}$Nd ratio is still better than 5 parts in $10^4$.

**Data Representation**

The oxygen data are given in the conventional $\delta$ notation of:

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

Samples are corrected to the SMOW scale using California Institute of Technology rose quartz $^{18}O = 8.45$. Following DePaolo and Wasserburg [1976a], an analogous notation is used for Nd and Sr, with the reference reservoir evolving with time where

$$
\epsilon_{\text{Nd}} = \left[\frac{^{144}\text{Nd}}{^{144}\text{Nd}}_{\text{HUR}} - 1\right] 10^6
$$

and

$$
\epsilon_{\text{Sr}} = \left[\frac{^{87}\text{Sr}}{^{86}\text{Sr}}_{\text{HUR}} - 1\right] 10^6
$$

The ($^{144}$Nd/$^{144}$Nd)$_{\text{HUR}}$ and ($^{87}$Sr/$^{86}$Sr)$_{\text{HUR}}$ are the measured ratios corrected for decay since the time of crystallization $T$.

<table>
<thead>
<tr>
<th>Samples</th>
<th>$^{87}$Rb/$^{86}$Sr</th>
<th>$^{144}$Sm/$^{144}$Nd</th>
<th>$^{87}$Sr/$^{86}$Sr</th>
<th>$^{144}$Nd/$^{144}$Nd</th>
<th>$\epsilon_{\text{Sr}}$*</th>
<th>$\epsilon_{\text{Nd}}$*</th>
<th>$\delta^{18}O$</th>
</tr>
</thead>
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<tr>
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<td></td>
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<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>G54 WR</td>
<td>0.104</td>
<td>0.192</td>
<td>0.70491 ± 4</td>
<td>0.512219 ± 19</td>
<td>5.3 ± 0.6</td>
<td>7.5 ± 0.4</td>
<td>12.7</td>
</tr>
<tr>
<td>G54 Calcite</td>
<td>0.004</td>
<td>0.209</td>
<td>0.70650 ± 15</td>
<td>0.512265 ± 29</td>
<td>30.5 ± 2.1</td>
<td>8.1 ± 0.6</td>
<td></td>
</tr>
<tr>
<td>Sheeted Dikes</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>G10</td>
<td>0.029</td>
<td>0.209</td>
<td>0.70535 ± 4</td>
<td>0.512250 ± 25</td>
<td>13.5 ± 0.6</td>
<td>7.8 ± 0.5</td>
<td>8.5</td>
</tr>
<tr>
<td>K1</td>
<td>0.008</td>
<td>0.194</td>
<td>0.70519 ± 4</td>
<td>0.512275 ± 20</td>
<td>11.7 ± 0.6</td>
<td>8.6 ± 0.4</td>
<td>6.8</td>
</tr>
<tr>
<td>Plagiogranite</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>G224-3</td>
<td>0.035</td>
<td>0.187</td>
<td>0.70362 ± 4</td>
<td>0.512249 ± 18</td>
<td>-11.3 ± 0.6</td>
<td>8.2 ± 0.3</td>
<td>5.2</td>
</tr>
<tr>
<td>Gabbros</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>G224-2 WR</td>
<td>0.010</td>
<td>0.201</td>
<td>0.70370 ± 3</td>
<td>0.512251 ± 19</td>
<td>-9.5 ± 0.4</td>
<td>8.0 ± 0.4</td>
<td>3.7</td>
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<tr>
<td>G224-2 PLAG</td>
<td>0.008</td>
<td>0.151</td>
<td>0.70352 ± 4</td>
<td>0.512196 ± 18</td>
<td>-12.0 ± 0.6</td>
<td>7.8 ± 0.3</td>
<td>4.5</td>
</tr>
<tr>
<td>G224-2 URAL</td>
<td>0.014</td>
<td>0.241</td>
<td>0.70425 ± 3</td>
<td>0.512285 ± 19</td>
<td>-1.6 ± 0.4</td>
<td>8.0 ± 0.4</td>
<td>2.6</td>
</tr>
<tr>
<td>K9 WR</td>
<td>0.0008</td>
<td>0.271</td>
<td>0.70304 ± 5</td>
<td>0.512310 ± 18</td>
<td>-18.6 ± 0.7</td>
<td>8.0 ± 0.3</td>
<td>5.7</td>
</tr>
<tr>
<td>K9 PLAG</td>
<td>0.0005</td>
<td>0.129</td>
<td>0.70296 ± 2</td>
<td>0.512180 ± 25</td>
<td>-19.7 ± 0.3</td>
<td>7.8 ± 0.5</td>
<td>6.0</td>
</tr>
<tr>
<td>K9 CPX</td>
<td>0.0008</td>
<td>0.322</td>
<td>0.70313 ± 3</td>
<td>0.512341 ± 15</td>
<td>-17.3 ± 0.4</td>
<td>7.7 ± 0.3</td>
<td>5.3</td>
</tr>
<tr>
<td>K9 WR</td>
<td>0.0008</td>
<td>0.316</td>
<td>0.70311 ± 5</td>
<td>0.512196 ± 18</td>
<td>-12.0 ± 0.6</td>
<td>7.8 ± 0.3</td>
<td>4.5</td>
</tr>
<tr>
<td>OM251 PLAG</td>
<td>0.0010</td>
<td>0.159</td>
<td>0.70315 ± 5</td>
<td>0.512203 ± 22</td>
<td>-18.0 ± 0.7</td>
<td>7.6 ± 0.4</td>
<td>6.4</td>
</tr>
<tr>
<td>OM251 CPX</td>
<td>0.0018</td>
<td>0.377</td>
<td>0.70315 ± 5</td>
<td>0.512346 ± 25</td>
<td>-17.5 ± 0.7</td>
<td>7.6 ± 0.5</td>
<td>5.8</td>
</tr>
<tr>
<td>OM28 WR</td>
<td>0.0014</td>
<td>0.264</td>
<td>0.70383 ± 5</td>
<td>0.512274 ± 22</td>
<td>-7.8 ± 0.7</td>
<td>7.7 ± 0.4</td>
<td>5.2</td>
</tr>
<tr>
<td>Harzburgite K22</td>
<td>0.150</td>
<td>0.216</td>
<td>0.70313 ± 9</td>
<td>0.512278 ± 23</td>
<td>-21.2 ± 1.1</td>
<td>8.3 ± 0.4</td>
<td></td>
</tr>
</tbody>
</table>

*The $\epsilon_{\text{Sr}}$ 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.

†The $^{144}$Nd/$^{144}$Nd determined from spiked sample.
The ($^{143}$Nd/$^{144}$Nd)CHUR $\delta$ and ($^{87}$Sr/$^{86}$Sr)UR $\delta$ ratios are those in the standard reservoirs [DePaolo and Wasserburg, 1976a, b] at the time $T$.

RESULTS

Oxygen Isotopic Characteristics

The oxygen isotopic systematics of the Samail ophiolite are discussed in detail by Gregory and Taylor [1981]. The three general categories of samples are (1) lower cumulate gabbros which exhibit close to 'normal' $^{18}$O characteristics (K9, OM251), (2) rocks below the diabase-gabbro contact which exhibit $^{18}$O depletions (G224-2, G224-3, OM28), and (3) rocks above the diabase-gabbro contact which exhibit $^{18}$O enrichments, relative to a MORB reservoir of $\delta^{18}$O = 5.7 [Muehlenbachs and Clayton, 1976]. The dichotomy between $\delta^{18}$O values of the rocks above or below the diabase-gabbro contact is a result of differing alteration environments which depend upon temperature, water/rock ratio, and the history of the exchange fluid.

In order to understand the $\delta^{18}$O variation due to temperature effects, let us consider the $\delta^{18}$O fractionation between rock and fluid defined as

$$\Delta^{18}O_{\text{rock}} = \Delta = \delta^{18}O_{\text{rock}} - \delta^{18}O_{\text{water}}$$

(1)

which is a function of temperature given by Taylor [1974]:

$$\Delta(T) = \frac{A}{T^2} + B$$

(2)

where $A$, $B$ are constants.

In the basalt-seawater system, temperatures of alteration in the vicinity of 250°C to 300°C produce no $\delta^{18}$O shifts in the rocks as a result of hydrothermal exchange (i.e., $\Delta \approx 5.7$). The gabbros which are altered at temperatures greater than 400°C are depleted in $\delta^{18}$O (i.e., $\Delta < 5.7$). The subsolidus alteration event is characterized by isotopic disequilibrium between coexisting plagioclases and clinopyroxenes and water/rock ratios $<1.0$ [Gregory and Taylor, 1981]. The Samail gabbro-water interactions are analogous to those observed in continental layered complexes such as the Skærgaard intrusion which have suffered meteoric hydrothermal exchange [Taylor and Foster, 1979]. The major difference between the continental and marine environments is the contrast between the initial compositions of the hydrothermal fluids: meteoric water with an initial $\delta^{18}$O $\approx -15\%$ (e.g., the Skærgaard [Taylor and Foster, 1979]), and seawater with $\delta^{18}$O $\approx 0\%$. The contrast between fluid and rock in the subaerial environments is large, $\approx 20\%$, so that the hydrothermal exchange with meteoric water produces almost exclusively $\delta^{18}$O depletions in the intrusion and its surrounding country rock over a wide range of temperatures [Taylor, 1974; Taylor and Foster, 1979]. In the marine case, because seawater is isotopically closer to the initial rock composition, the effects of high temperature alteration in the gabbros are less pronounced with a maximum $\delta^{18}$O depletion of approximately $3\%$, while in subaerial environments hydrothermally altered gabbros often display $\delta^{18}$O depletions of as much as $5-10\%$ [Taylor and Foster, 1979]. Because of the spreading environment of the seafloor the diabases and pillow lavas overlying gabbroic magma chambers 'see' a time-temperature superposition of alteration regimes that has no other analog in the subaerial examples described to date [Gregory and Taylor, 1981].

In order to illustrate these points, considering only the effects of temperature, the diabase dikes altered under green-schist to lower amphibolite facies are expected to be depleted in $^{18}$O relative to their primary values if pristine seawater with a $\delta^{18}$O $\approx -0.5$ is the hydrothermal fluid. However, diabases G10 and K1 with $\delta^{18}$O = 8.5, 6.8, respectively, are to the contrary enriched in $^{18}$O. This contradiction requires that these diabases have exchanged with a seawater-derived hydrothermal fluid that has become $^{18}$O enriched relative to its initial value. The fluids which have exchanged with the underlying solidified cumulate gabbro are likely candidates as the gabbro section is a region of high-temperature ($\approx 400^\circ$C) alteration and low water/rock ratio (<1); this is ideal for producing a fluid phase with $^{18}$O enrichments. Fluids discharging from the gabbros into the diabases have a calculated $^{18}$O enrichment relative to an initial $\delta^{18}$O = -0.5 from $\approx+2$ to $\approx+9\%$ [Gregory and Taylor, 1981]. Since seawater is only $\approx 6\%$ lighter than the rock, water exhibiting enrichments of this magnitude reacts with the diabases at lower temperature resulting in green-schist facies rocks ($2 \lt \Delta(T) \leq 6$) with a $\delta^{18}$O $> 6$. This suggests that the oxygen isotopes of the diabases are recording a final exchange event as the following discussion will illustrate.

The inferred funnel-shaped geometry of the oceanic crustal magma chamber [Hopson et al., 1981; Pallister and Hopson, 1980] requires that fluids which exchange with large amounts of solidified gabbraro discharge at the distal edge of the chamber away from the ridge axis [Gregory and Taylor, 1981]. Since thin sheets of magma (inferring from the crystallization of the most fractionated magma at the diabase-gabbro contact [Pallister and Hopson, 1980]; see Figure 11) act as impermeable barriers to fluid circulation [Taylor and Forester, 1979; Norton and Taylor, 1979], two hydrothermal systems exist during the early history of newly created oceanic crust: one affecting the roof rocks of the gabbroic magma chamber near the ridge axis and the second affecting the gabbros which have crystallized under the wings of the chamber [Gregory and Taylor, 1979]. Waters discharging from the second system react with the diabases which have been already altered by the upper system as they spread away from the ridge axis, and thus the diabases are affected by both systems. These geometrical constraints become important when evaluating the combined Sr and O effects discussed below.

Trace Element Abundances

The K, Rb, Sr, Sm, and Nd abundances are listed in Table 2. The most obvious feature is the contrast in K, Rb, Sm, and Nd concentrations between the layered gabbros (K9, OM28, and OM251) which are a factor of 10 to 100 times lower than those in the upper sequence (plagiogranite, uralite gabbro, sheeted dikes, and basalts). This feature reflects the mineralogy (clinopyroxene, plagioclase, and olivine) and the cumulative origin of the layered gabbros. Since the K, Rb, Sm, and Nd mineral/liquid partition coefficients of the cumulus phases are all less than unity, the gabbro cumulates contain much lower concentrations than the melts from which they crystallized. The resulting melt which is enriched in these elements is presumably represented by the overlying more fractionated rocks such as the diabases and pillow basalts. It is therefore unlikely that any particular ophiolite member represents the primitive melt composition. In contrast, the Sr concentration is relatively uniform throughout the layered gabbros and the upper sequence. This is probably due to the high concentration of Sr in plagioclase of the layered gabbro which compensates for
the lower concentrations in the other cumulate minerals, olivine and clinopyroxene, leaving Sr in the melt approximately constant.

Significant variations in concentrations also exist within the same lithologic units. The gabbro K9 has approximately a factor of 4 higher concentration of Sm and Nd and a factor of approximately 2 higher K and Sr concentrations than the gabbro OM251 (see Table 2). Similar variations have also been observed in the gabbros from the Troodos ophiolite [Kay and Senechal, 1976]. The variations in trace element concentration probably reflect the complex crystal fractionation and replenishment history of the magma chamber. In the Samail ophiolite, evidence for multiple cycles of primitive magma replenishment has been extensively documented within the gabbro complex by Pallister and Hopson [1981]. This is a common feature of ophiolite complexes [Moores, 1969; Jackson et al., 1973; Pallister and Knight, 1981] to test the hypothesis that the harzburgite tectonite is a residual fraction of upper mantle left after partial melting. REE data can also be used [Montigny et al., 1973; Allegre et al., 1973; Pallister and Knight, 1981] to test the hypothesis that the harzburgite is a residue co-genetic with the overlying cumulate gabbros and basaltic suites. This type of calculation requires assumptions concerning amount of partial melting, distribution coefficients, and composition of source material, all of which are uncertain. A more definitive constraint is provided by the Nd isotopic data, presented in the following section.

**Crystallization Age and Initial \(^{143}\)Nd/\(^{144}\)Nd**

Although current models for the formation of ophiolites at spreading centers indicate that within a single section all the igneous rocks, with perhaps the exception of the basal peridotites, should be of approximately the same age, there may still be significant variations in initial \(^{143}\)Nd/\(^{144}\)Nd ratios due to mantle source heterogeneities. For this reason, analyses of only whole rock samples from different parts of an ophiolite may have variable initial \(^{143}\)Nd/\(^{144}\)Nd ratios and may not define an isochron. In addition, the range in Sm/Nd ratios of total rocks may be inadequate to define an age for such young samples. Therefore crystallization ages and initial ratios were determined using coexisting minerals separated from gabbros.

We obtained Sm-Nd mineral isochrons from three gabbros

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**TABLE 2. Trace Element Abundances From the Samail Ophiolite**

<table>
<thead>
<tr>
<th>Samples</th>
<th>K</th>
<th>Rb</th>
<th>Sr</th>
<th>Sm</th>
<th>Nd</th>
<th>Sr/Rb</th>
<th>K/Rb</th>
<th>Sm/Nd</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basalt</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>G54 WR</td>
<td>4470</td>
<td>7.16</td>
<td>198</td>
<td>5.03</td>
<td>15.9</td>
<td>28</td>
<td>624</td>
<td>0.316</td>
</tr>
<tr>
<td>G54 Calcite</td>
<td>76</td>
<td>0.009</td>
<td>6.12</td>
<td>0.043</td>
<td>0.126</td>
<td>680</td>
<td>8440</td>
<td>0.341</td>
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<td>Sheeted Dikes</td>
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<td>G10</td>
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<td>162</td>
<td>3.85</td>
<td>11.2</td>
<td>99</td>
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<td>K1</td>
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<td>388</td>
<td>4.16</td>
<td>13.0</td>
<td>370</td>
<td>2201</td>
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<tr>
<td>Plagiogranite</td>
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<tr>
<td>G224-3</td>
<td>2200</td>
<td>2.31</td>
<td>192</td>
<td>5.36</td>
<td>17.3</td>
<td>83</td>
<td>952</td>
<td>0.310</td>
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<td>Gabbros</td>
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<tr>
<td>G224-2 WR</td>
<td>1320</td>
<td>0.674</td>
<td>196</td>
<td>2.78</td>
<td>8.36</td>
<td>290</td>
<td>1960</td>
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<tr>
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<td>1800</td>
<td>0.950</td>
<td>330</td>
<td>0.563</td>
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<td>3.79</td>
<td>9.51</td>
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<td>2640</td>
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<td>0.052</td>
<td>188</td>
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<td>24.8</td>
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<td>156</td>
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<td>0.081</td>
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<td>Average MOR* tholeiite</td>
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<td>2.7</td>
<td>8</td>
<td>130</td>
<td>130</td>
<td>0.34</td>
</tr>
</tbody>
</table>

Parts per million.

*Data from Engel et al. [1965], Tatsumoto et al. [1965], Kay et al. [1970], and Schilling [1971].

combined Ibra data give an age of 130 ± 12 m.y. and Esd = 7.8 ± 0.2. This will be used as a reference isochron for comparison of other data from the Ibra section.

In order to determine whether there exist any variations in either the age or initial $^{143}$Nd/$^{144}$Nd ratio we analyzed the gabbro OM251 which was collected in Wadi Fizh approximately 300 km NW from the Ibra section (see Figure 1). The plagioclase and clinopyroxene data points from this gabbro give an age of 100 ± 20 m.y. and an initial of $\epsilon_{Nd} = 7.5 ± 0.3$ (Figure 4). This initial ratio is slightly lower than that obtained from the Ibra section, and the crystallization age appears to be about 30 m.y. younger (note, however, that the error bars on the two isochron ages overlap). This age difference, if substantiated, would imply a tectonic break, for example, a transform fault, with a large lateral offset. Such a structure conceivably might be located in the Samail gap (Figure 1).

Tilton et al. [1981] have dated plagiogranites along the entire length of the Oman Mountains by using the uranium lead technique on zircons and found U-Pb ages which are virtually constant (94–98 m.y.) for the entire ophiolite. On the basis of these consistent results, Tilton et al. [1981] interpret the zircon ages as the time of igneous crystallization. Sm-Nd ages obtained from north of the Samail gap (Figure 1) agree with the zircon results. However east of the Samail gap, the Sm-Nd data from Ibra gives consistently older (~30 m.y.) ages than the plagiogranites. Since all Sm/Nd and $^{143}$Nd/$^{144}$Nd were determined from the same spike solution for samples northwest of the Samail gap, K9 clinopyroxene was repeated using this same technique. The results of the second run are within experimental error the same as the aliquot data (Table 1). Note that the second run was performed on a new mineral separate, yet the Sm-Nd systematics did not change, thereby indicating the reliability of the data. The contrast between the two Ibra ages suggests that the U-Pb and Sm-Nd clocks are recording different events southeast of the Samail gap.

The geologic significance of the age discrepancy is not resolved. However, the internal consistency of both the U-Pb [Tilton et al., 1981] and the Sm-Nd data indicates that extreme caution must be exercised when attaching a geologic inter-

(K9, G224-2, OM251) using plagioclase, clinopyroxene, and uralite mineral separates. The gabbro K9 from the Ibra section contains plagioclase, clinopyroxene, and olivine and is extremely fresh, with only a minor amount of alteration. It can be seen in Table 1 and Figure 2 that the $^{143}$Sm/$^{144}$Nd ratio is enriched in the clinopyroxene from K9 by 2 parts in 10$^4$ relative to the plagioclase. The clinopyroxene also has a higher $^{143}$Sm/$^{144}$Nd ratio of 0.320 compared to 0.129 for the plagioclase. Assuming that these minerals and the whole rocks have remained closed systems since crystallizing from an isotopically homogeneous melt, these data indicate a crystallization age of 128 ± 20 m.y. with an initial $^{143}$Nd/$^{144}$Nd ratio of $\epsilon_{Nd} = 7.7 ± 0.3$. Since the age and initial ratio for this gabbro are only defined by two independent points, the interpretation of this two-point isochron as the time of crystallization is not unique. To eliminate the possibility of two component mixing producing this array, three or more distinct mineral phases would have to be analyzed. This is not possible in this sample, as no other REE bearing mineral phase is present (olivine contains negligible REE).

We analyzed a second gabbro from the upper part of the same gabbro unit as K9. This gabbro (G224-2) contains uralite as a late-stage replacement of pyroxene and primary hornblende. The low $\delta^{18}O$ values for both plagioclase and uralite indicate that this sample has exchanged with seawater at high temperatures (>400°C). The Sm-Nd data from G224-2 are shown in Table 1 and Figure 3. The uralite has the highest $^{143}$Sm/$^{144}$Nd and $^{143}$Nd/$^{144}$Nd ratios and together with the plagioclase gives an age of 150 ± 40 m.y. and an initial $^{143}$Nd/$^{144}$Nd ratio of $\epsilon_{Nd} = 7.8 ± 0.4$. The larger uncertainty in the age compared to K9 is due to the smaller difference between the $^{143}$Sm/$^{144}$Nd ratios in the plagioclase and uralite mineral separates. However, within analytical uncertainty, both samples have the same age and initial $^{143}$Nd/$^{144}$Nd ratio. In addition, the results from G224-2 show that Sm-Nd systematics are not affected by the presence hydrous alteration minerals such as uralite. This is evidence that the correlated variations of $^{143}$Nd/$^{144}$Nd and $^{143}$Sm/$^{144}$Nd ratios in these minerals represent an isochron rather than an accidental mixing line. The combined Ibra data give an age of 130 ± 12 m.y. and $\epsilon_{Nd} =
Fig. 4. Sm-Nd evolution diagram for the cumulate gabbro OM251, from Wadi Fizh, northern Oman (see Figure 1).

Fig. 5. Sm-Nd evolution diagram showing whole rock data for basalt, sheeted dikes, and plagiogranite from the Ibra section, Oman. With the exception of the sheeted dike K1 the data from the other samples are consistent with the same crystallization age and initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio as the gabbros from Ibra. The deviation of K1 indicates the presence of small but significant isotopic heterogeneities in the magma reservoirs. The tectonized harzburgite K22 is within analytical uncertainty of the Ibra reference isochron. This is consistent with the harzburgite being a cogenetic of the partial melting event which produced the rest of the ophiolite.

Implications of Sm-Nd

The presence in the Samail ophiolite of only small variations in the initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios from different localities and stratigraphic units appears to indicate a relatively uniform isotopic source regions. It is therefore useful to compare...
The value of $\epsilon_{Nd} = 7.8$ in the Samail ophiolite is somewhat lower than 'typical' MORB with $\epsilon_{Nd} \approx 10$. This may be due to the older age of the Samail ophiolite compared to the zero age basalts plotted in Figure 6, or it may simply reflect the overall variability present today in ocean ridge basalts. The dispersion in $\epsilon_{Nd}$ of modern day oceanic basalts is to some extent correlated with chemistry. Alkaline rocks generally have lower $\epsilon_{Nd}$ values than MOR tholeiites [Carlson et al., 1978]. However, several oceanic tholeiites with relatively low $\epsilon_{Nd}$ values in the range of $+7.0$ to $8.5$ have also been found, for example, in the Atlantic [113152; DePaolo and Wasserburg, 1976b], and ARPs [Richard et al., 1976] and Pacific (BD17-1 [DePaolo and Wasserburg, 1976b]) oceans. These samples do not appear to be chemically or tectonically distinctive, although the isotopic data indicate an origin from a less depleted source reservoir. Tectonically distinctive oceanic settings which may be modern day analogues of the environment in which the Samail ophiolite was formed are the Lau marginal basin and Scotia Sea back arc. The $\epsilon_{Nd}$ values from these two areas [Hawkesworth et al., 1977; Carlson et al., 1978] are shown in Figure 6 and define a relatively narrow range in $\epsilon_{Nd}$ of from 8.1 to 9.0. These ratios are also somewhat low compared to normal MORB samples and are very similar to the Samail ophiolite.

To some extent the slightly lower $\epsilon_{Nd}$ value of the Samail ophiolite compared to typical oceanic crust is due to the evolution of $143\text{Nd}/144\text{Nd}$ in the oceanic mantle reservoirs during the past 130 m.y. To correct for this evolution, an estimate of the $147\text{Sm}/144\text{Nd}$ ratio in the Samail ophiolite source reservoir is required. This can be obtained by assuming that the composite of the cumulative gabbros, diabase dikes, and pillow basalts represents the primary unprocessed magma. From the Samail ophiolite data we estimate that this magma had $\text{Nd} \sim 4$ ppm and $\text{Sm} \sim 1.6$ ppm. This corresponds to $147\text{Sm}/144\text{Nd} \sim 0.24 \pm 0.02$. As Nd and Sm are strongly partitioned into the liquid during partial melting, this is also probably a reasonable estimate for the source reservoir. For this value of $147\text{Sm}/144\text{Nd}$ the Samail source reservoir evolves by $\sim 0.6$ $\epsilon_{Nd}$ units per 100 m.y. and would today have $\epsilon_{Nd} \sim 8.7$. Thus the $\epsilon_{Nd}$ value of the Samail ophiolite would now only be about $\sim 1$ $\epsilon_{Nd}$ unit below the center of the MORB distribution and is there-
fore fully consistent with a MORB affinity. It is also noted that for the MORB source reservoir to evolve in a single stage from $e_{Nd} = 0$ to the modern day value of $e_{Nd} = +10$ requires, for $^{142}$Sm/$^{144}$Nd = 0.24, at time interval of 1.7 eons. This result is similar to the single stage differentiation ages obtained by Pb-isotope studies [Church and Tatsumoto, 1975; Sun et al., 1975].

The relationship of the basal peridotites to the rest of the ophiolite sequence has been obscure due to the effects of deformation and post-emplacement alteration which have strongly modified the primary geologic structures and relationships. Various hypotheses summarized by Coleman [1977] have considered the harzburgite as a (1) cogenetic lowermost cumulate, (2) older accidental mantle substrate with is not petrogenetically linked to the rest of the ophiolite, or (3) a cogenetic refractory residue remaining from the partial melting which produced the rest of the ophiolite. The initial Nd value of $e_{Nd} = 8.3 \pm 0.4$ for the tectonized harzburgite K22 is within error the same as the rest of the ophiolite. This result, together with the low trace element concentrations, supports the third hypothesis.

There is, however, conflicting evidence from the U-shaped (convex downward chondrite normalized) rare earth element (REE) patterns which have been observed in the Samail harzburgite [Pallister and Knight, 1981] and a number of other peridotites [Allegre et al., 1973; Kay and Senechal, 1976; Frey, 1969; Suen et al., 1979]. On the basis of U-shaped REE patterns in harzburgite and dunite from Newfoundland, Suen et al. [1979] have concluded that the tectonite peridotite could not be related to the overlying oceanic crust by a simple partial melting event. Their argument against a cogenetic origin for the harzburgite is that residues of light REE (LREE) depleted MORB should also exhibit LREE depletion. The argument depends upon the choice of crystal-liquid partition coefficients for olivine and orthopyroxene. Natural olivines from meteorites and from peridotite [Schatzler and Bottino, 1971; Masuda, 1968; Frey, 1969] may exhibit U-shaped REE patterns. Combining natural olivine-orthopyroxene partition coefficient data from Frey [1969] with experimentally determined orthopyroxene-liquid data reviewed by Irving [1978] suggests that residual harzburgites can have U-shaped patterns because olivine dominates the light REE and orthopyroxene dominates the heavy REE. A cumulative dunite measured by Suen et al. [1979] (which on geologic grounds crystallized from the same melts that produced LREE depleted diabases [Malpas, 1978]) has a U-shaped pattern. If we assume that REE partitioning between cumulate and liquid is the same as partitioning between residue and liquid, then a cogenetic residue of harzburgite should also exhibit such a pattern but at lower concentrations.

The harzburgite REE concentrations are lower than expected from simple one-stage partial melting and are a consequence of the plumbing system that delivers melts to the surface. Two processes enrich the overall REE contents in ophiolitic diabases and by analogy in seafloor basalts: (1) open system fractionation in crustal magma chambers driven by periodic intrusion and mixing into the chamber of primitive magma [Stern, 1979; Pallister and Knight, 1981] and (2) fractionation of olivine during ascent of the primary melt bodies through the upper mantle [Hopson et al., 1981]. The second process is a consequence of the fact that enstatite is not an early liquidus phase at low pressure for any MORB compositions—confirmed experimentally by Green et al. [1979] and Stolper [1980] and in the Samail ophiolite [Hopson et al., 1981; Pallister and Hopson, 1980]. The implication of this relationship is that above depths of 30 km [Stolper, 1980] all harzburgite-melt contacts become sites of reaction between ascending melts and their host rocks which are not in magmatic equilibrium. The upward migration and reaction of the melts with the harzburgite is manifested in the Samail peridotite as cross-cutting dunite bodies [Boudier and Coleman, 1981]. Using field estimates of the amount of dunite present in the Samail peridotite, fractionation of olivine during ascent may enrich the REE of the melt by as much as a factor of 2.

The former process documented by Pallister and Knight [1981] has enriched the overall REE concentrations of Samail basalts by factors up to five. In samples K9 and OM251 a 4-fold enrichment is observed. These combined crustal and upper mantle processes may lead to order of magnitude enrichments of overall bulk REE between initial melt segregation and crystallization. Therefore any trace element concentration arguments concerning genetic relationships between rocks such as the harzburgites and diabases are subject to postseparation processes which cannot be constrained without other petrologic data (usually not available with seafloor samples). In contrast, the Nd isotope data which are independent of the vagaries of the plumbing system see through this history and thus provide a more straightforward test of the primary relationship between harzburgite and the overlying oceanic crust.

The partial melting event which produced the harzburgite residue is unlikely to be exactly contemporaneous with the crystallization of the immediately overlying pile of cumulate gabbros, sheeted dikes, and pillow lavas. The following sequence is suggested. Partial melting of an already LREE depleted source material (herzolite) at depths of 60 to 70 km [Green et al., 1979] is followed by melt segregation. The residual harzburgite ascends with a rate constrained by the spreading rate to form the basement of the cumulate gabbros and basaltic suites. For a half spreading rate of 1 cm/yr the harzburgite would take ~6 m.y. to rise from the partial melting zone. In contrast, the magma produced during this partial melting would rise much more rapidly at the rate of ~1 km/yr, otherwise the melt would crystallize before reaching the surface [Marsh, 1979]. Thus for the conservative spreading rate used in this example the formation of the harzburgite residue probably occurred, at most, 6 m.y. prior to the crystallization of the immediately overlying magmatic suite. This time difference is not resolvable with the present Sm-Nd data. Thus within the analytical uncertainties the harzburgite and overlying magmatic sequences are considered to be cogenetic.

Comparison of Rb-Sr and Sm-Nd

The most obvious contrast between the Sr and Nd isotopic systems is the larger variability of the $^{87}$Sr/$^{86}$Sr ratios ranging from 0.70296 ($e_{Sr} = -19.7$) to 0.70650 ($e_{Sr} = +30.5$) compared to a range in $e_{Nd}$ of from only 7.5 to 8.6. The lowest $^{87}$Sr/$^{86}$Sr ratios are found in the plagioclase separates from the layered cumulate gabbros K9 and OM251. In K9 the plagioclase has an $^{87}$Sr/$^{86}$Sr ratio of 0.70296 ± 2. However, the coexisting clinopyroxene has a distinctly higher value of 0.70313 ± 3. In Figure 7 reference isochrons of 130 and 4550 m.y. are shown. A line connecting the clinopyroxene and plagioclase from K9 gives an impossibly old age of >4.5 eons. In the high-level gabbro G224-2 there is a similar disparity between the plagioclase and uralite. The uralite has an $^{87}$Sr/$^{86}$Sr ratio of 0.70426 ± 3 compared to 0.70352 ± 4 for the coexisting plagioclase. These isotopic variations cannot be accounted for
by radiogenic decay since crystallization. Instead, this isotopic relationship between the coexisting mineral phases indicates partial exchange with a source of high \(^{87}\text{Sr}/^{86}\text{Sr}\). The enrichment in \(^{87}\text{Sr}/^{86}\text{Sr}\) is more pronounced in the clinopyroxene and uralite and is anticipated as they have a factor of 5 lower Sr concentration than the plagioclase. This enrichment is also consistent with the petrography as the uralite is clearly an alteration product and the clinopyroxene contains ~5% secondary hornblende. These results are in distinct contrast to Samail, where all mineral phases plot on a 130 m.y. isochron.

The highest \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios found in the Samail ophiolite are in the upper sequences consisting of basalts, sheeted dikes, and plagiogranites. The higher \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios in these rocks are consistent with a greater degree of exchange with seawater having a high \(^{87}\text{Sr}/^{86}\text{Sr}\) ratio. In an attempt to determine the \(^{87}\text{Sr}/^{86}\text{Sr}\) ratio in seawater we analyzed a calcite amygdule from the basalt OMG54. This has \(^{87}\text{Sr}/^{86}\text{Sr} = 0.7065\) which is significantly higher than that found for any of the igneous rocks and may be compared to 0.7076 estimated for Cretaceous seawater [Peterman et al., 1970; Veizer and Compston, 1974]. These observations are clearly consistent with hydrothermal interaction with seawater. The slightly lower ratio in the carbonate amygdule compared to Cretaceous seawater is probably due to the lowering of \(^{87}\text{Sr}/^{86}\text{Sr}\) in the fluid by interaction with the rock. There is also an approximate correlation between high \(^{87}\text{Sr}/^{86}\text{Sr}\) and stratigraphic height (see Figure 11) which suggest that water/rock increased upward in the section.

Because of 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 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. 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 and should thus also represent a primary value. The difference between the plagioclase separates from K9 and OM251 of 0.70296 ± 2 and 0.70311 ± 5 could therefore indicate a small but real difference in the \(^{87}\text{Sr}/^{86}\text{Sr}\) ratios of these samples and apparently of the original magmas from which they crystallized. This suggests that small \(^{87}\text{Sr}/^{86}\text{Sr}\) mantle isotopic heterogeneities may have been preserved during the formation of the Samail ophiolite because the same effect is also apparent in the Nd isotopic data (Table 1). In fact these two samples OM251 and K9 plot right on the MOR correlation line in the \(e_{\text{Nd}}, e_{\text{Sr}}\) diagram of Figure 8.

The importance of distinguishing between seawater alteration effects and mantle isotopic variations is illustrated in the Rb-Sr isochron diagram of Figure 7. In this figure, seawater alteration has produced an approximate correlation between Rb/Sr and \(^{87}\text{Sr}/^{86}\text{Sr}\). If interpreted literally, this correlation would imply that Rb/Sr heterogeneities have existed for at least several billion years in the mantle reservoirs of the Samail ophiolite. Similar correlations and interpretations have been suggested in other ophiolites [Peterman et al., 1971], oceanic basalts [Brooks et al., 1976; Hedge, 1978], and oceanic islands [Sun and Hanson, 1975; Duncan and Compston, 1976; Whitford et al., 1977]. Although some of these correlations are undoubtedly due to mantle features, the \(^{18}\text{O}\) and Sr data from the Samail ophiolite indicate that in this case the Rb/Sr-\(^{87}\text{Sr}/^{86}\text{Sr}\) correlation is clearly an artifact of seawater alteration.

A useful criterion to distinguish between isotopic variations due to either primary magmatic or seawater alteration effects is the \(e_{\text{Nd}}, e_{\text{Sr}}\) diagram. In Figure 8 the \(e_{\text{Nd}}, e_{\text{Sr}}\) values from the Samail ophiolite are shown. They have an origin at the mantle correlation line [DePaolo and Wasserburg, 1976b; O’Nions et al., 1977] and form a horizontal trajectory to the higher \(e_{\text{Sr}}\) values. Although this general effect has been observed by previous workers [DePaolo and Wasserburg, 1977; O’Nions et al., 1977], the present results provide a spectacular example of how the Nd isotopic composition is unaffected by seawater contamination, while the Sr composition is shifted drastically. It is illustrative to compare the observed \(e_{\text{Nd}}, e_{\text{Sr}}\) values with those calculated for contamination of oceanic crust by Cretaceous seawater. Assuming a simple closed system model for mixing of Sr and Nd in seawater with oceanic crust, mass balance between the initial and final products gives:

\[
W_{\text{water}} + F_{\text{rock}} = W_{\text{water}} + F_{\text{rock}}' = W_{\text{water}} + F_{\text{rock}} (3)
\]

...
where \( w \) and \( r \) are the atomic proportions of Nd or Sr in the water and rock systems respectively and \( \epsilon_{\text{rock}} \) is the initial \( \epsilon_{\text{Nd}} \) or \( \epsilon_{\text{Sr}} \) value in the rock before exchange, \( \epsilon_{\text{rock}}' \) is the final modified rock \( \epsilon_{\text{Nd}} \) or \( \epsilon_{\text{Sr}} \) value, \( \epsilon_{\text{water}} \) is the initial \( \epsilon_{\text{Sr}} \) value in water before exchange, and \( \epsilon_{\text{water}}' \) is the modified water \( \epsilon_{\text{Nd}} \) or \( \epsilon_{\text{Sr}} \) value. For equilibrium exchange of water and rock, \( \epsilon_{\text{water}}' = \epsilon_{\text{rock}}' \), and from (3) the water/rock ratio by weight is given by

\[
\frac{W}{R} = \left( \frac{\epsilon_{\text{rock}}' - \epsilon_{\text{water}}'}{\epsilon_{\text{water}}' - \epsilon_{\text{rock}}'} \right) \frac{C_{\text{rock}}}{C_{\text{water}}}
\]

(4)

where \( C_{\text{rock}} \) is the concentration of Sr or Nd in initial rock and \( C_{\text{water}} \) is the concentration of Sr or Nd in seawater.

For Sr the parameters used in this equation are \( \epsilon_{\text{rock}} = -20 \), \( \epsilon_{\text{water}} = +47 \) [Peterman et al., 1970], and \( C_{\text{water}} = 8 \) ppm [Goldberg, 1965]. The \( \epsilon_{\text{rock}}' \) values are given in Table 1. It is assumed that the initial Sr concentration in the rock \( (C_{\text{rock}}) \) is the same as the final measured concentration (Table 2). For Nd the parameters used in (4) are \( \epsilon_{\text{rock}} = +8.0 \), \( \epsilon_{\text{water}} = -7 \), and \( C_{\text{water}} = 2.8 \times 10^{-6} \) ppm [Piepgras et al., 1979]. The \( \epsilon_{\text{rock}} \) and \( \epsilon_{\text{water}} \) values are given in Tables 1 and 2.

Using (4) and the Sr and Nd parameters, the \( \epsilon_{\text{Sr}} \) and \( \epsilon_{\text{Nd}} \) values in the rock resulting from mixing of different proportions of seawater with oceanic crust have been calculated. The \( \epsilon_{\text{Sr}} \) and \( \epsilon_{\text{Nd}} \) mixing lines is shown in Figure 8 together with the water/rock ratios calculated for rock Sr and Nd concentrations of 180 and 10 ppm, respectively. From Figure 9 it can be seen that the \( \epsilon_{\text{Nd}} \) values are far less sensitive to seawater contamination than the \( \epsilon_{\text{Sr}} \) values. This is due to the significantly lower concentration of Nd in seawater of \( 2.8 \times 10^{-6} \) ppm compared to 8 ppm for Sr. For example, the Sr data indicate that the sample with the highest water/rock ratio is the diabase dike K1 with \( W/R = 43 \) (Table 3). Using the Sr water/rock ratio for K1 and solving for \( \epsilon_{\text{water}} \) show that a \( W/R = 43 \) causes an insignificant change of \( \epsilon_{\text{Sr}} = -0.0001 \). To produce a measurable shift in \( \epsilon_{\text{Nd}} \) of 0.5, a water/rock ratio of \( W/R = 1.6 \times 10^5 \) would be required. Thus relative to the most altered rock, an increase in the water/rock ratio of \( -4 \times 10^5 \) would be required before seawater contamination would measurably change the \( \epsilon_{\text{Nd}} \) values. A detailed analysis of the effects of seawater interaction using \( \epsilon_{\text{Sr}} \) and \( \epsilon_{\text{Nd}} \) variations will be given in the following section.

Comparison of \( \delta^47 \text{Sr}/\delta^{86} \text{Sr} \) and \( \delta^{18} \text{O} \)

It has already been shown that \( ^{87}\text{Sr}/^{86}\text{Sr} \) and \( \delta^{18} \text{O} \) values are sensitive to hydrothermal alteration. The effects of alteration on both these systems is shown in the initial \( ^{87}\text{Sr}/^{86}\text{Sr} \), \( \delta^{18} \text{O} \) diagram of Figure 9. In this figure there appears to be at least two distinct trends which are centered about the primary val-

**TABLE 3. Sr Water/Rock Ratios for Closed and Open Systems, Ibra Section**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Open System</th>
<th>Closed System</th>
</tr>
</thead>
<tbody>
<tr>
<td>G54</td>
<td>11.6</td>
<td>14.8</td>
</tr>
<tr>
<td>G10</td>
<td>14.0</td>
<td>20.2</td>
</tr>
<tr>
<td>K1</td>
<td>31.1</td>
<td>43.5</td>
</tr>
<tr>
<td>G224-3</td>
<td>3.3</td>
<td>3.6</td>
</tr>
<tr>
<td>G224-2</td>
<td>4.2</td>
<td>4.5</td>
</tr>
<tr>
<td>K9</td>
<td>0.5</td>
<td>0.5</td>
</tr>
</tbody>
</table>

![Graph](image-url)

**Fig. 10. (a)** The \( \epsilon_{\text{Sr}} \) and \( \delta^{18} \text{O} \) mixing lines for \( \Delta = 0 \) and \( C = 0.22 \). The closed system curve is calculated using (6) and maps out the locus of \( \delta^{18} \text{O} \) values for systems with fixed \( W/R \) and water recirculation. The open system curve allows each packet of water to pass through the rock only once; thus no recirculation is permitted. In areas where hot spring activity occurs, such as the Galapagos rift [Corliss et al., 1979], the open system equation applies. The 'one pass' curve is calculated by substituting the expression for open system water/rock (equation (9)) into (7). For \( \Delta \) and \( C \) fixed, the natural systems lie between the two curves. (b) The \( \epsilon_{\text{Sr}} \) and \( \delta^{18} \text{O} \) values in the rock for \( C = 0.022 \) and \( \Delta = 2 \) (line A), \( \Delta = 6 \) (line C), and \( \Delta = 8 \) (line B). For \( \Delta = 6 \), which corresponds to a temperature of \( -300^\circ \text{C} \), there can be shifts in \( \epsilon_{\text{Sr}} \) without any change in \( \delta^{18} \text{O} \). (c) The \( \epsilon_{\text{Sr}} \) and \( \delta^{18} \text{O} \) values in water for \( \Delta = 2 \), \( \Delta = 6 \), and \( \Delta = 8 \) (solid lines) are given by \( \delta^{18} \text{O}_{\text{water}} = \delta^{18} \text{O}_{\text{rock}} - \Delta \) and \( \epsilon_{\text{rock}} \). For \( \Delta < 6 \), which corresponds to temperature \( >300^\circ \text{C} \), seawater will be enriched in \( 18 \text{O} \) as long as the water/rock ratio is small. For \( \Delta > 6 \) (<300ºC) \( 18 \text{O} \)-depleted waters will be produced. The dashed lines show the \( \epsilon_{\text{Sr}} \) and \( \delta^{18} \text{O} \) values in the rock (from Figure 10b). The horizontal dashed line illustrates how the combined strontium and oxygen data can be used as a geothermometer for hydrothermal interactions if the \( W/R \) obtained from the \( \delta^{18} \text{O} \) is used to solve the oxygen \( W/R \) equation for \( \Delta \). (d) Example of mixing lines consistent with \( \epsilon_{\text{Sr}} \) and \( \delta^{18} \text{O} \) values in the uralite gabbro.

The values of \( \delta^{18} \text{O} \approx +6 \) and \( ^{87}\text{Sr}/^{86}\text{Sr} \approx 0.7030 \). The most obvious trend is the approximate inverse correlation corresponding to depleted \( \delta^{18} \text{O} \) and high \( ^{87}\text{Sr}/^{86}\text{Sr} \) ratios. The second, less distinctive, feature is the association of heavy \( \delta^{18} \text{O} \) values with high \( ^{87}\text{Sr}/^{86}\text{Sr} \) ratios in the sheeted dikes and basalt. In an attempt to understand these features we will now consider simple models for the exchange of Sr and O with seawater.

Let us consider a simple closed system model for the exchange of Sr and O with seawater as has already been discussed for Sr and Nd. Using this model, the water/rock ratio (by weight) calculated using \( \delta^{18} \text{O} \) variations is given by Taylor [1974]:

\[
\frac{W}{R} = \left( \frac{\delta_{\text{rock}} - \delta_{\text{water}}}{\delta_{\text{water}} - (\delta_{\text{rock}} - \Delta)} \right) \frac{C_{\text{rock}}}{C_{\text{water}}}
\]

(5)

This equation is identical to that used for Sr, apart from the temperature dependent fractionation factor \( \Delta \). Let us now assume that only a particular fraction of the rock exchanges oxygen with seawater. This may, for example, be the case for rocks with different modal proportions of easily exchanged minerals such as feldspar. The effective water/rock ratio \( (W/R)_{\text{effective}} \) is then given by

\[
\frac{W}{R}_{\text{(effective)}} = \frac{W}{R} \times \frac{1}{q}
\]

(6)

where \( q \) is the fraction of the rock that exchanges oxygen with seawater.
Fig. 11. The δ¹⁸O, εSr, and εNd values from a composite profile through the Samail ophiolite, Ibra section. The ophiolite section was compiled from the geologic map of Bailey et al. [1980]. The symbol V = pillowed volcanic, SD = sheeted dikes, PG = plagiogranite, HG = high-level gabbro, G = cumulate gabbro, WG = interlayered wehrlite and gabbro, D = dunite, and H = harzburgite. The plagiogranites and differentiated high-level gabbros are the last intrusive rocks to crystallize. It can be seen that the end values are, in general, within error of the primary magmatic value of εNd = 7.8. In contrast, the δ¹⁸O [from Gregory and Taylor, 1981] and εSr values show large deviations from their primary magmatic values of δ¹⁸O = 5.7 and εSr = -19.5. These large variations in δ¹⁸O and εSr are a consequence of hydrothermal interaction of seawater. The isotopic characteristics of the Samail ophiolite may be similar to those in oceanic crust undergoing subduction in present-day ocean basins.

where q is the fraction (by weight) of rock that exchanged with seawater. A similar equation can also be written for the Sr water/rock ratio, where, in this case, let us assume that different fractions of the rock, q' exchanged with seawater. Therefore the relationship between the Sr and O water/rock ratios is given by

\[ \frac{W}{R} (\text{Sr}) = \frac{1}{q} \frac{W}{R} (\text{oxygen}) = \frac{1}{q} \]

Using this relationship, (4) and (5) can now be combined to give

\[ \frac{\epsilon_{\text{rock}}}{\epsilon_{\text{water}}} = \left( \frac{\delta_{\text{water}} - \delta_{\text{rock}}}{\delta_{\text{water}} - (\delta_{\text{rock}} - \Delta)} \right) \times C \]

where

\[ C = \frac{C_{\text{Srwater}}/C_{\text{Owater}} 	imes q'/q}{C_{\text{Srrock}}/C_{\text{Orock}}} \]

An example of a mixing line using (7) is shown in Figure 10a for Δ = 0. It can be seen that the mixing line is a hyperbola with the curvature determined by the constant C. For q'/q = 1 and using average concentrations of Sr and O in the rock and seawater, C = 0.022 in both the closed and open system curves shown in Figure 10a. Using this arbitrary value of C, the effect of changing the oxygen water/rock fractionation factor from Δ = +2 to Δ = +6 or +8 is shown in Figure 10b. The value of Δ depends on the temperature of exchange and is given by (2). Thus in an environment with changing exchange temperatures no simple correlation of δ¹⁸O and ⁸⁷Sr/⁶⁶Sr would be expected. In fact, for Δ = 6 corresponding to a temperature of −250°C there can be changes in ⁸⁷Sr/⁶⁶Sr without any change in the δ¹⁸O (Figure 10b).

For Δ ≠ 0 the δ¹⁸O value of the water is different from the δ¹⁸O value of the rock, being related by \( \Delta = \delta_{\text{rock}} - \delta_{\text{water}} \).

The curves which map the water compositions are shown in Figure 10c for the different Δ values. In Figure 10c, for small W/R ratios (<3) when Δ < 6, the final water which equilibrates with rock will be enriched in δ¹⁸O (i.e., δ¹⁸Ofinal > 0) and for Δ > 6, the water will be depleted. Subsequent interaction of this exchanged water will result in another set of mixing curves. Also, in Figure 10c the ⁸⁷Sr/⁶⁶Sr value can be used independently to assess the water/rock ratio. Then either the temperature of equilibration or the final fluid composition can be in principle calculated from (8). Using this reasoning on sample G10 (Figure 9) exhibiting a greenschist assemblage suggests that diabase G10 equilibrated with a seawater-derived fluid which had been exchanged and enriched by at least 2% over its initial value. Evidence has been found in the diabase dikes for exchange with ¹⁸O-enriched water [Gregory and
Taylor, 1981], indicating that this water has had a history of high temperature (>300°C) exchange.

In Figure 10d, mixing lines are shown which could account for the Sr isotopic disequilibrium observed for example between the plagioclase and uralite in the gabbro G224-2. The mixing lines are different for each mineral as they have different Sr concentrations and presumably different exchange rates (i.e., different values of q'/q). As the factor q'/q is unknown, the mixing lines that are shown are arbitrary, but they qualitatively account for the observed 87Sr/86Sr and δ18O values in the minerals of G224-2.

The above calculations were based on a closed system model which assumes continuous recirculation and cyclic reequilibration of the water with the rock. However, some of the heated water will be lost from the system, for example, by escape to the ocean. In the extreme open system case in which the system reequilibrates with the underlying gabbros at higher temperature as suggested by Gregory and Taylor [1981]. This δ18O-shifted fluid would also presumably be shifted in 87Sr/86Sr (~0.704 indicated by uralite from G224-2) and may not have affected the diabases already enriched in 87Sr during alteration over the ridge axis. As stated above, this implies that there are at least two temperature and spatial regimes where exchange of oxygen and Sr occurs. The first is at the ridge axis above the magma chamber where fluid path lines do not cross the diabase-magma contact. It is in this environment that 18O-depleted rocks and 87Sr/86Sr enriched rocks would be produced during reaction and exchange with large volumes of seawater. Subsequently, with continued production of oceanic crust these rocks are transported away from the ridge into a cooler regime at the distal edges of the magma chamber where interaction between isotopically shifted fluids and the diabases occurs. A high 18O fluid discharges from a second hydrothermal system which operates under the wings of the ridge magma chamber. This fluid imposes a 18O enrichment upon the diabases which masks the earlier ridge axis oxygen exchange without appreciably affecting the Sr isotope ratios. This 'hydrothermal reworking' of oceanic crust may also explain the generally more altered nature of ophiolite pillow basalts and diabases compared to those dredged from ocean ridges.

**SUMMARY AND CONCLUSIONS**

1. In this study we have established the crystallization age of the Samail ophiolite by obtaining Sm-Nd mineral isochrons from gabbros. The gabbros from the southern part of the ophiolite in Ibra give an age of 128 ± 20 m.y. and 150 ± 40 m.y., while a gabbro from Wadi Fizh in the northern part of Oman gives a somewhat younger age of 100 ± 20 m.y. These results show that the Sm-Nd technique can be used to determine crystallization ages of relatively young mafic rocks.

2. The initial 143Nd/144Nd ratio of gabbros, plagiogranite, diabase dikes, and a basalt from the Samail ophiolite have a narrow range in εNd of from 7.6 to 8.6 (Figure 11). This indicates derivation from a relatively uniform reservoir that has been depleted in the LREE for at least a billion years or more.

3. The tectonized harzburgite has εNd = 8.3, which is the same as the overlying cumulate gabbros, sheeted dikes, and pillow basalts. This result, together with its low trace element concentrations, is consistent with the harzburgite being a cogenetic residue from partial melting event which produced the overlying magmatic sequences. However, considering the complexity and heterogeneous character of the peridotite in the Samail ophiolite [Boudier and Coleman, 1981], generalization of this result must await a more complete characterization of the peridotite.

4. In contrast to the initial 143Nd/144Nd ratios the initial 87Sr/86Sr ratios have an extremely large range of from 0.7030 to 0.7065. This large variation is consistent with hydrothermal interaction of seawater with oceanic crust. In addition, the 87Sr/86Sr ratios generally decrease downward (Figure 11), suggesting downward decreasing water/rock ratios. These results indicate that extreme caution must be employed before variations in initial 87Sr/86Sr ratios of ophiolitic rocks can be attributed to primary magmatic variations. In this study we have, in fact, found small variations in initial 87Sr/86Sr of gabbro plagioclase separates of from 0.70296 ± 2 to 0.70311 ± 5, which have been attributed to primary magmatic variations. However, this interpretation is supported by corresponding variations in the initial 143Nd/144Nd ratios and by analysis of coexisting pyroxenes and plagioclases.

5. The complete δ18O profile from Gregory and Taylor [1981] is also shown in Figure 11. With respect to the "normal" value of δ18O = 5.7 there are both 18O-depleted and enriched rocks. The depletions are a result of high-temperature hydrothermal alteration, whereas the 18O enrichments are due to hydrothermal exchange with either strongly 18O-shifted waters at high temperature or less 18O-shifted waters at much lower temperatures.

In Figure 11 there is approximate linear correlation between high 18O and high 87Sr/86Sr ratios above the diabase-gabbro contact. Below this contact, the correlation is approximately inverse, corresponding to depleted 18O and high 87Sr/86Sr ratios. These relationships are primarily due to the temperature dependence of the oxygen water-rock interactions. Because of this temperature effect it has been shown that, in general, no simple relationship between 18O and 87Sr/86Sr variations would be expected for the entire ophiolite section. In fact, at the appropriate temperature of hydrothermal interaction with pristine seawater (200°–250°C) it is possible to
have 87Sr/86Sr enrichments while retaining apparently normal 18O values.

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