

12. OXYGEN AND STRONTIUM ISOTOPE COMPOSITIONS OF HESS DEEP GABBROS (HOLES 894F AND 894G): HIGH-TEMPERATURE INTERACTION OF SEAWATER WITH THE OCEANIC CRUST LAYER 3¹

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ABSTRACT

The Hess Deep Rift Valley exposes a young (≈ 1 Ma) section of lower oceanic crust generated at the East Pacific Rise (EPR). The drillers recovered 154 m of plutonic rocks from Hole 894G. The mineralogy and textures of these rocks suggest that they represent the roof of a magma chamber. These high-level gabbros, olivine-gabbros, and gabbro-norites are slightly hydrated (average LOI = 0.75%, $N = 27$) and crosscut by a few olivine basaltic dikes. The dikes were altered under greenschist facies conditions, and the gabbros were amphibolitized. Most of the gabbros have Sr-isotopic ratios that range from 0.70247 to 0.70309, indicative of low water-rock ratios (< 1), with a circulating fluid having a $^{87}\text{Sr}/^{86}\text{Sr}$ ratio calculated at 0.7032. The $\delta^{18}\text{O}$ values of the gabbros range from 2.2 to 6.5 and are generally much lower (average $\delta^{18}\text{O} = 4.8$, $N = 31$) than the mantle reference value (5.7 ± 0.2), whereas olivine basalts do not show any ^{18}O depletion. The $\delta^{18}\text{O}$ values of clinopyroxenes are often anomalously low (5.1–5.3). Clinopyroxene-plagioclase fractionation values reach 1.5 and reveal strong isotopic disequilibrium attributed to high-temperature isotopic exchange with a discrete aqueous fluid. Clinopyroxenes with $\delta^{18}\text{O} \approx 5$ contain minute Mg-rich amphibole lamellae (1–5 mm) that represent a very early stage of high-temperature alteration before altering totally into amphibole. Except for a few samples with isotopic compositions that were reequilibrated at low temperatures ($< 200^\circ\text{C}$), the gabbros contain low- $\delta^{18}\text{O}$ plagioclase ($\delta^{18}\text{O} = 3\text{--}5$) that recrystallized at high temperatures ($400\text{--}600^\circ\text{C}$) without exchanging major cations (An_{50-65}). The Hess Deep gabbros record mechanisms of isotopic exchanges governed by percolation of fluids along grain boundaries and self-diffusion of oxygen through plagioclase without involving macroscopic brittle deformation. These water-rock interactions produced a sequence of low- ^{18}O oceanic rocks early in the spreading history of the EPR that can be compared with the gabbro sequence of the Oman ophiolite.

INTRODUCTION

The opening of the Hess Deep Rift has tectonically exhumed young (≈ 1 Ma) lower crustal and upper mantle rocks that were generated at the East Pacific Rise (EPR). During Leg 147 of the Ocean Drilling Program (ODP), we cored gabbroic sections (Holes 894F and 894G) that provide an unique opportunity for studying the hydrothermal interactions between seawater and the lower levels of the oceanic crust generated at a fast-spreading ridge. Within the high-level gabbros of Holes 894F and 894G, a well-developed vein and crack network records low-temperature channeled fluid flow ($< 350^\circ\text{C}$) that was contemporaneous with the thermal contraction of the crust (Gillis, Mével, Allan, et al., 1993).

The role of solid state diffusion during fluid circulation is generally neglected in the total budget of the hydrothermal alteration of the oceanic crust even though it should provide information about the early on-axis high-temperature hydrothermal activity ($> 350^\circ\text{C}$). The pervasive circulation of fluids at grain boundaries through plutonic rocks must be examined at a location where hot hydrothermal fluids are expected to react with high-level gabbros. If temperatures in the hydrothermal system are high enough, hot fluids percolate along grain boundaries and react at the grain scale through dissolution-pre-

cipitation or diffusion mechanisms (e.g., self-diffusion of oxygen in plagioclase). Consequently, mineralogic alteration and isotopic exchange may develop far from major discontinuities (veins, cracks) and participate in the thermal cooling of the crust as well as in chemical and isotope exchange with seawater.

If subsolidus hydrothermal alteration of gabbros is pervasive and occurs at high temperatures, their initial oxygen isotopic compositions must be shifted toward lower $\delta^{18}\text{O}$ values at the contrary of basalts and most dikes of oceanic layer 2. An accurate knowledge of the oxygen isotope budget established during oceanic crust-seawater interactions is an important key to determine how the oxygen isotope composition of the world ocean evolves through time. On the basis of oxygen isotope measurements of dredged oceanic gabbros and mass balance calculations, Muehlenbachs and Clayton (1976) proposed that the $\delta^{18}\text{O}$ value of the oceans is buffered at a value near zero by isotopic exchange between seawater and oceanic crust. Gregory and Taylor (1981) and Stakes and Taylor (1992), who have studied the Oman ophiolite, found that the ^{18}O enrichment of basalts and dikes balances the ^{18}O depletion of the gabbros in different profiles of the fossil oceanic crust. However, oxygen isotope studies of ophiolites like Xigaze (Agrinier et al., 1988) and Trinity (Lécuyer and Fourcade, 1991) revealed a global ^{18}O enrichment of the deep-level rocks. Holes 894F and 894G in the Hess Deep Rift Valley provide the opportunity to validate the existence of low $\delta^{18}\text{O}$ gabbros drilled in an intact part of oceanic layer 3. We sampled undeformed gabbros mainly metamorphosed under amphibolite facies conditions (called "static metamorphism") to study the thermal and chemical effects of their high-temperature alteration by seawater. Strontium and oxygen isotope measurements were performed on gabbros selected far from any kind of major discontinuity (shear zone, crack, vein). In situ oxygen isotope ratios were obtained on single plagioclase and pyroxene

¹Mével, C., Gillis, K.M., Allan, J.F., and Meyer, P.S. (Eds.), 1996. *Proc. ODP, Sci. Results*, 147: College Station, TX (Ocean Drilling Program).

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grains to examine how water-rock interactions extend down to the subgrain scale. Thus, the "background alteration" of Hess Deep gabbros will be characterized in terms of (1) metamorphic assemblages, (2) rates of hydration, (3) closure temperatures for oxygen isotopes, and (4) water-rock ratios.

ANALYTICAL TECHNIQUES

We determined mineral modes (volume percentages) visually or by point counting (see also Gillis, Mével, Allan, et al., 1993). Loss on ignition (LOI) was either determined on the *JOIDES Resolution* or at the University of Rennes (Table 1).

Oxygen was extracted from rock powders using the BrF_3 method (Clayton and Mayeda, 1963) and analyzed as CO_2 gas on a VG SIRA 10 mass spectrometer at the University of Rennes. Isotopic compositions are quoted in the standard δ notation relative to SMOW. Results from the NBS28 standard gave $\delta^{18}\text{O} = +9.5\text{‰} \pm 0.2\text{‰}$ ($2\sigma_m$).

We obtained $\delta^{18}\text{O}$ values for mineral separates using microcores extracted from thick sections of rock chips; consequently, these analyses gave the compositions of individual mineral grains, not averaged values as typically provided by conventional techniques of mineral separation. The amount of material extracted by this method is typically 1–3 mg of powder. Thick sections are used to check if contamination by neighboring minerals occurs during the drilling procedure. The identification of minerals was performed by Raman spectroscopy, either in the vicinity of microcores or on residual powders.

We determined the Sr-isotope compositions (Table 1) following the method of Jahn et al. (1980). Sr blanks were lower than 1.10^{-9} g. $^{87}\text{Sr}/^{86}\text{Sr}$ ratios were measured using a five-collector Finnigan MAT-262 mass spectrometer. All ratios were normalized against $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ for isotopic fractionation. Results from the NBS987 Sr standard on 15 separate runs yielded $^{87}\text{Sr}/^{86}\text{Sr} = 0.71023 \pm 2$ ($2\sigma_{\text{pop}}$), and all $^{87}\text{Sr}/^{86}\text{Sr}$ ratios presented in this paper were corrected by $+0.00002$ to be consistent with the NBS reference value of 0.71025.

PETROLOGY

Hole 894F and the upper part of Hole 894G are dominated by gabbros and olivine gabbros that are crosscut by basaltic dikes with chilled contacts against the wallrocks. Deeper in Hole 894G, gabbro represents the most abundant rock type and contains a few olivine gabbro layers.

The Hess Deep gabbros are characterized by a "static metamorphism" (rocks metamorphosed in the absence of deformation at the rock or grain scale) whose extent is highly variable with depth; the percentage by volume of secondary minerals ranges from 5% to 80%. Mineral assemblages suggest that this metamorphism began under amphibolite facies conditions characterized by the following mineral association: Mg-rich amphibole (tremolite) + Ca-rich plagioclase. Retrograde metamorphic reactions under greenschist facies conditions can be also found in some samples showing an association of Na-rich plagioclase (albite-oligoclase) + actinolite \pm chlorite \pm talc \pm epidote.

Brittle deformation, which accompanied the thermal contraction of the plutonic rocks, has developed a dense vein network observable all along the core. Both macroscopic (≥ 0.01 mm wide) and microscopic veins occur in the gabbros of Site 894 (Manning and MacLeod; Manning et al., this volume). Optical microscopy reveals that the early generation of veins are microscopic and filled with amphibole: they are considered to be related to the pervasive and static amphibolite facies mineral assemblages ("background alteration"). All macroscopic veins postdate the microscopic veins and are characterized by greenschist or zeolite facies assemblages. A first generation of these veins are filled with amphibole \pm chlorite \pm titanite. They are followed by anastomosing fractures filled with actinolite \pm chlorite \pm

epidote \pm prehnite \pm smectite and typically associated with extensive wallrock alteration that extends at least 10 cm from the veins. The youngest macroscopic veins, commonly subvertical, are filled with a typical zeolite facies mineral association: chlorite \pm smectite \pm mixed-layer clays \pm zeolites \pm calcite, which can be found as coatings on core pieces or as late residual pore filling in earlier vein types.

Olivine is generally strongly altered in the range from 50 to 100 vol%. When olivine relicts are still present, they are rimmed by a coronitic association of chlorite, iddingsite, talc, and tremolite. Replacement of pyroxenes by amphibole and minor chlorite is the main manifestation of alteration in Hess Deep gabbros but remains heterogeneous from one sample to another one (10%–100% alteration by volume). Tiny secondary clinopyroxene patches, commonly associated with brown-green amphibole lamellae, occur as replacement of magmatic clinopyroxenes, whereas discrete patches of talc sometimes invade the orthopyroxene cores.

Plagioclase looks fresh in most of the gabbro samples, but turbid grains containing numerous and very tiny iron exsolutions are also present. Fluid inclusions are often abundant along the crystal rims. In highly altered zones, the formation of albitic plagioclase dominates. Inclusions or patches of clays and zeolites are common in plagioclase grains close to anastomosing macroscopic vein networks filled with low-temperature minerals.

The extent of alteration in basaltic dikes is also highly variable, with 10% to 85% replacement of primary mineralogy by volume. The metamorphic mineral assemblage lacks any evidence of amphibolite facies. The mineral associations that characterize the greenschist and zeolite facies observed in gabbros occur both in the background alteration of the dikes and in their associated macroscopic and microscopic veins.

For studying the mechanisms of background alteration affecting the Hess Deep gabbro succession of Holes 894F and 894G, samples were selected randomly along the core independent of the texture and primary mineralogy variations. However, except for Sample 147-894G-2R-3, 20–26 cm, which was collected close to a low-temperature macroscopic vein, sites close to any kind of macroscopic open fracture, vein network, and associated alteration halo were avoided.

RESULTS

Oxygen Isotopes

We analyzed 35 whole rocks (Table 1) and 16 microcored minerals (Table 2) for oxygen isotope compositions. In Figure 1, whole-rock $\delta^{18}\text{O}$ values are plotted vs. depth (meters below seafloor [mbsf]) for Holes 894F and 894G. The $\delta^{18}\text{O}$ values for gabbros vary widely and range from 2.2 to 6.5, but 60% of the data range from 4.5 to 5.0, revealing a significant ^{18}O depletion for this sequence of high-level gabbros relative to the mantle reference of 5.7 ± 0.2 (Pineau et al., 1976). Sample 147-894G-2R-3, 20–26 cm, the only gabbro sampled close to a macroscopic vein network, has the highest $\delta^{18}\text{O}$ value of 6.5. $\delta^{18}\text{O}$ variations with depth commonly reach 1‰ by meter. The distribution of $\delta^{18}\text{O}$ values does not show any regular trend of isotopic enrichment or depletion with depth (Fig. 1). The lowest $\delta^{18}\text{O}$ values are found within 100 m below the surface whereas $\delta^{18}\text{O}$ values close to 5.3‰ characterize Hole 894G from 100 to 150 m. Hess Deep gabbros show a combination of low $\delta^{18}\text{O}$ values (down to 2.2) and low LOI ($< 0.5\%$). However, the inverse is also seen: $\delta^{18}\text{O}$ values lower than 5 associated with high LOI values up to 2.5% (Table 1).

Four olivine basalts (Table 1), which are interpreted as dikes crosscutting the gabbro sequence (Gillis, Mével, Allan, et al., 1993), have $\delta^{18}\text{O}$ values slightly lower (5.3 and 5.5) or slightly higher (6.1) than the mantle reference value ($5.7\text{‰} \pm 0.2\text{‰}$).

$\delta^{18}\text{O}$ values of microcored minerals are only available for pyroxene, plagioclase, chlorite, and ilmenite (Table 2). Amphibole was not analyzed because of the common mixing of this mineral with minute

Table 1. Oxygen and strontium isotope compositions of gabbros and basalts from Holes 894F and 894G.

Core, section, interval (cm)	Piece no.	Depth (mbsf)	Rock type	Alteration (vol%)	LOI	$\delta^{18}\text{O}$ ‰(SMOW)	Sr (ppm)	$^{87}\text{Sr}/^{86}\text{Sr}$
147-894F-3R-1, 98-103	17	17.78	gb	16	0.86	4.4	98	0.702618
147-894G-2R-1, 60-65	9	29.20	pl-ol basalt	12	1.80	5.3	88	0.702642
2R-1, 85-92	11	29.45	pl-ol basalt		1.38	5.3		
2R-2, 95-100	15	31.05	pl-ol basalt	20	1.84	5.5	88	
2R-3, 20-26	4	31.75	ol-gb	85	4.05	6.5	65	0.704297
2R-3, 109-115	14	32.64	ol-gb	33		5.2	79	
4R-1, 113-119	16	46.13	gb-norite	26	0.40	5.1	92	0.702577
5R-1, 30-34	5	50.30	gb-norite		0.57	4.7	92	0.702588
6R-1, 94-100	8A	55.74	gb-norite		0.95	4.9		
6R-2, 79-87	6	57.02	gb-norite	22	0.52	4.6	83	0.702629
7R-1, 56-66	11	65.36	ol-gb-norite	23	0.70	4.7	89	0.702569
7R-1, 76-84	13	65.56	gb-norite	31	0.48	4.1	100	
8R-2, 87-90	8	70.87	gb-norite		0.34	4.8		
9R-2, 88-93	5B	76.53	gb-norite		0.35	4.8		
9R-3, 133-139	11	78.24	ol-gb-norite	42	0.72	5.2	99	0.702688
9R-4, 100-144	8-12	79.33	ol-gb-norite	33	0.70	4.9	105	
10R-1, 105-111	15	80.15	gb		0.77	5.0		
11R-2, 100-107	10D	86.56	gb-norite	15	0.49	5.0	92	0.702513
11R-3, 1-3	1	87.01	gb-norite		0.13	4.7		
11R-3, 17-28	3A	87.17	gb-norite			4.9	79	0.702489
11R-3, 31-38	4	87.31	gb-norite	30	0.45	5.2	90	
12R-1, 45-51	6	94.25	gb-norite		1.27	5.5		
12R-2, 37-45	4E	95.63	gb-norite	22	1.27	4.7	111	0.702584
12R-3, 137-142	8A	98.1	gb-norite		0.75	5.3		
12R-5, 122-125	11B	100.95	gb-norite		0.27	4.4		
13R-2, 6-13	1B	104.83	gb-norite	30	0.65	2.2	111	0.702574
13R-2, 90-98	8	105.67	gb-norite	33	0.41	4.3	93	
13R-3, 17-24	1B	106.40	gb-norite	32		4.1	88	0.702511
15R-1, 32-35	6	119.12	gb-norite		1.59	5.7		
17R-1, 50-58	10A	126.30	gb-norite	6	0.42	5.4	85	0.702474
18R-1, 82-85	10A	131.72	gb-norite		1.42	5.2		
19R-1, 49-55	9	140.99	pl-ol basalt	80	2.83	6.1	91	0.703090
19R-1, 84-96	13	141.34	gb-norite	56	0.40	4.4	106	0.702966
20R-1, 87-91	15	146.47	gb-norite		2.58	4.8		
20R-3, 66-71	10	148.08	gb-norite		0.75	5.6		

Notes: Estimated volume of alteration, loss on ignition, and Sr concentrations are also reported for discussion in the text. The error presented on the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is the statistical error obtained during the mass spectrometry run. The true error, including other error sources such as instrumental reproducibility, is estimated to be close to ± 30 . Total uncertainty on the $\delta^{18}\text{O}$ values with respect to the SMOW is estimated to be lower than 0.1‰ to 0.15‰. LOI = loss on ignition. gb = gabbro, pl-ol basalt = plagioclase-olivine basalt, gb-norite = gabbro-norite, ol-gb = olivine gabbro, and ol-gb-norite = olivine gabbro-norite.

grains of iron oxides, talc, and pyroxene. Plagioclase has a wide range of $\delta^{18}\text{O}$ values (3.0‰–6.4‰), contrasting with the more homogenous compositions of pyroxenes (5.1‰–5.7‰). Pyroxenes from Hess Deep have slightly lower $\delta^{18}\text{O}$ values than those typical of fresh magmatic pyroxene from a MORB mantle source (Muehlenbachs and Clayton, 1972; Pineau et al., 1976; Kyser, 1986). The exception is Sample 147-894G-15R-1, 32–35 cm, with a $\delta^{18}\text{O}$ value of 5.7 (Table 2). One analysis of clinopyroxene in Sample 147-894G-20R-1, 87–91 cm, gave a surprisingly low $\delta^{18}\text{O}$ value of 3.5. The powder produced during the microcoring was examined by Raman spectroscopy, which revealed a mixture of clinopyroxene and amphibole. P. Agrinier (pers. comm., 1994) and G. Früh-Green (pers. comm., 1994) reported $\delta^{18}\text{O}$ values of amphibole in gabbros as low as 2‰, and this could explain the low $\delta^{18}\text{O}$ value for our pyroxene. Analyses of two chlorites gave $\delta^{18}\text{O}$ values of 8.2 and 5.3 in gabbro samples characterized by whole-rock $\delta^{18}\text{O}$ values of 5.7 and 5.5, respectively. One analysis of fresh ilmenite ($\text{FeTiO}_3 = 96.5\%$, $\text{MnTiO}_3 = 3\%$, and $\text{MgTiO}_3 = 0.5\%$) yielded a $\delta^{18}\text{O}$ value of 2.6, typical of Ti-Fe oxides that preserve a magmatic composition with limited subsolidus reequilibration in the presence of pyroxene and plagioclase (Taylor and Epstein, 1962a, 1962b; Anderson et al., 1971).

Strontium Isotopes

Sr-isotope compositions of Hess Deep rocks range from 0.70247 to 0.70430 but most of them are lower than 0.7027. Sample 147-894G-2R-3, 20–26 cm, the only sample close to a macroscopic vein network, has the highest initial Sr ratio and $\delta^{18}\text{O}$ value (Table 1). This sample, which will be discussed further in the next section, clearly

records late and low-temperature hydrothermal alteration and is not considered as belonging to the background alteration. Sr-isotope compositions for fresh mid-ocean ridge basalts (MORB) are commonly between 0.7023 and 0.7029, with an average value of 0.70265 (Hart et al., 1974; Barrett and Friedrichsen, 1982). Any increase in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio can be attributed to the addition of seawater Sr (Dasch et al., 1973). No Hess Deep samples are totally fresh with respect to their mineralogy. Thus, to estimate the initial magmatic Sr ratio, which is required to quantify isotopic exchange with seawater, we need to use another approach. There is a rough correlation of the initial Sr-isotope compositions with the estimated volume percentage of alteration, suggesting that Sr-isotope variations within Holes 894F and 894G are mainly controlled by the extent of hydrothermal alteration (Fig. 2). By extrapolation to a percentage of alteration equal to zero, an initial "magmatic" Sr-isotope ratio of 0.7024 is inferred. This value fits well with the previous range given for fresh MORB. We do not observe any positive correlation between the Sr-isotope ratios and Sr concentrations, which suggests that the main mechanism of ^{87}Sr increase in Hess Deep gabbros was isotope exchange rather than a global strontium addition caused by mineral neoformation.

DISCUSSION

The Hess Deep gabbros show large $\delta^{18}\text{O}$ variations (4‰) at the meter scale within 150 m of drilled rocks (Fig. 1). These whole-rock $\delta^{18}\text{O}$ values may depend on four parameters, including (1) the modal abundance of a mineral very sensitive to isotope exchange, (2) integrated fluid fluxes, (3) temperature, and (4) the isotope composition

Table 2. Oxygen isotope compositions of microcored minerals for gabbros from Hole 894G.

Core, section, interval (cm)	Piece no.	Depth (mbsf)	$\delta^{18}\text{O}\text{‰}$ (SMOW)			
			Plagioclase	Pyroxene	Chlorite	Ilmenite
147-894G-						
6R-1, 94-100	8A	55.74	4.0	5.7		
8R-2, 87-90	8	70.87	4.5	5.3		
9R-2, 88-93	5B	76.53	3.9	5.2		
11R-3, 1-3	1	87.01	4.8	5.2		2.6
12R-1, 45-51	6	94.25	6.4	5.1	5.3	
15R-1, 32-35	6	119.12	5.7		8.2	
20R-1, 87-91	15	146.47	3.0	3.5		

Notes: Total uncertainty on the $\delta^{18}\text{O}$ values with respect to the SMOW is estimated to be lower than 0.1‰ to 0.15‰.

of the hydrothermal fluid (Brikowski and Norton, 1989). When we plot the whole-rock $\delta^{18}\text{O}$ values against the estimated volume percentages of plagioclase, we do not see any significant correlation that could reveal a "modal effect" on the oxygen isotope compositions of the gabbros (Fig. 3). Such a correlation would not be expected if fluid fluxes are larger than reaction rates (Gregory et al., 1989).

Strontium Isotopes

The Sr-isotope compositions of the gabbros can be used to estimate water-rock ratios and the composition of the hydrothermal fluid. The water-rock ratios can be calculated using a simple mass balance equation (Albarède et al., 1981). This calculation requires that the Sr-isotope composition of a totally altered rock should be equal to the fluid in equilibrium. Holes 894F and 894G do not provide a 100% altered rock; however, by extrapolating the linear relation of Figure 2, we can deduce a Sr-isotope composition of 0.7032 as a rough estimate of the hydrothermal fluid that reacted with the Hess Deep gabbros. For the hypothetical 100%-altered rock, the mass balance equation provides a water-rock ratio of 1.4. Consequently, most of the Hess Deep gabbros reacted with a "mature" hydrothermal fluid developed in a rock-dominated system under water-rock ratios less than 1. It means that this fluid, whose composition is far from the initial Sr composition of seawater (≈ 0.7091 ; Burke et al., 1982), had already exchanged Sr isotopes before altering the high-level gabbros. McCulloch et al. (1980, 1981) found a similar relationship in the Samail ophiolite. The same phenomenon was also inferred from the calculated Sr-isotope composition of the end-member fluid ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7038$) that circulated through the sheeted dike complex of the East Pacific Rise sampled at Hole 504B (Kawahata et al., 1987). Sr-isotope compositions of Hole 504B veins crosscutting sheeted dikes indicate also that these hydrothermal solutions lie between 0.7031 and 0.7039 (Kawahata et al., 1987), well bracketing the calculated Hess Deep solution. These estimates also are in very good agreement with those measured in the hot fluids discharged at the East Pacific Rise at 21°N ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7030$; Albarède et al., 1981) and 13°N ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7041$; Michard et al., 1984).

Oxygen Isotopes

The study of Holes 894F and 894G gabbros provides the first evidence of the net ^{18}O depletion (Fig. 1) of an intact part of layer 3 formed at a modern fast-spreading ridge. The EPR basalts and dikes sampled at Hole 504B are ^{18}O enriched (Alt et al., 1986). Because no complete section of modern oceanic crust has been sampled, it is not yet possible to perform the types of mass balance calculations that have been done on the Samail ophiolite (Gregory and Taylor, 1981; Stakes and Taylor, 1992).

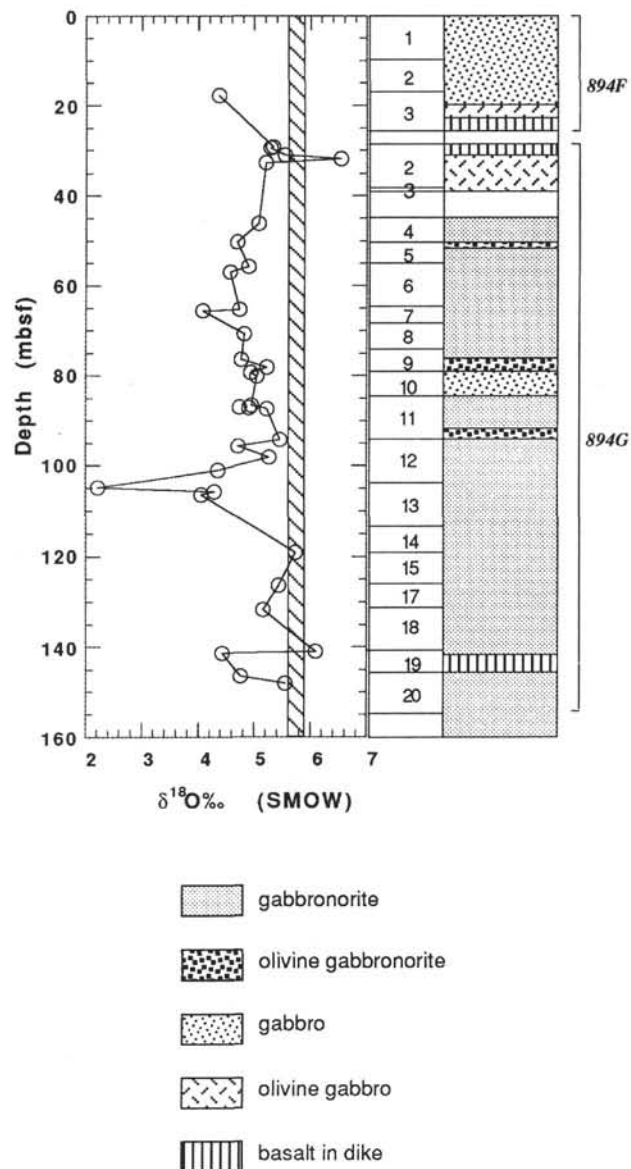


Figure 1. Distribution of whole-rock $\delta^{18}\text{O}$ values with depth (meters below seafloor). The shaded area covers the oxygen isotope compositions of fresh magmatic rocks derived from an oceanic mantle source. Numbers in column refer to core sections.

Gabbros in the Hess Deep area were recovered for the first time in 1988 during a *Nautilus* submersible dive program (Francheteau et al., 1990). P. Agrinier (pers. comm., 1994) investigated the stable isotope compositions of a large variety of rocks and found that most gabbros have $\delta^{18}\text{O} \approx 5.6$, which is very close to their initial magmatic values. Calcic plagioclase has magmatic oxygen isotope compositions and does not show evidence of isotope exchange with a fluid phase. When isotope exchange occurs, this is mainly under greenschist facies conditions (P. Agrinier, pers. comm., 1994) and ^{18}O modified gabbros are those in which plagioclase is unstable (partly replaced by albite and/or prehnite).

The conditions of water-rock interactions recorded in the gabbros from Holes 894F and 894G are drastically different from those dredged by the *Nautilus* and reveal the complexity and heterogeneity

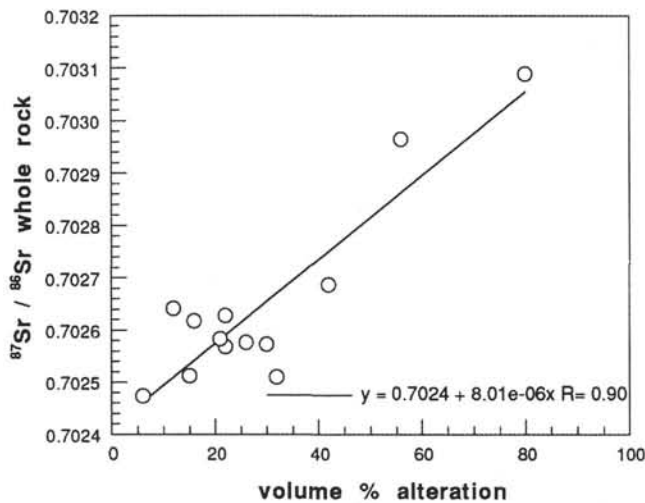


Figure 2. Variations of $^{87}\text{Sr}/^{86}\text{Sr}$ vs. the volume percentage of alteration. Sample 147-894G-2R-3, 20–26 cm, a gabbro very close to a macroscopic vein network, is excluded from this treatment. Only the background alteration is examined here.

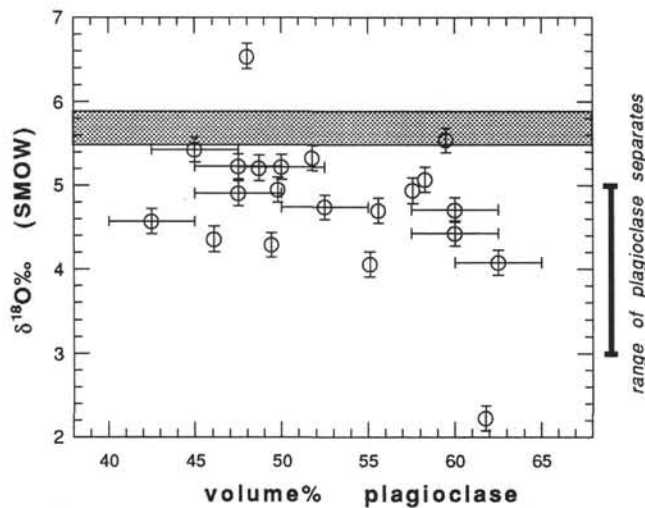


Figure 3. Variations of whole-rock $\delta^{18}\text{O}$ values vs. the modal abundance of plagioclase.

of the alteration processes affecting the oceanic crust. The ^{18}O depletion that affects our collection of Hess Deep gabbros occurs in slightly hydrated rocks where the plagioclase remains calcic under amphibolite facies conditions (An_{50-65} for plagioclases in Samples 147-894G-6R-1, 94–100 cm; 11R-3, 1–3 cm, and 15R-1, 32–35 cm). Figure 4 gives the variation of $\delta^{18}\text{O}$ values for some rock-forming minerals and shows that plagioclase ($3 < \delta^{18}\text{O} < 5$) is principally responsible for the global decrease of whole-rock $\delta^{18}\text{O}$ values. In pyroxene, $\delta^{18}\text{O}$ values range from 5.1 to 5.7. The apparent oxygen isotope fractionation between pyroxene and plagioclase can reach up to 1.5, a disequilibrium pattern already mentioned by Gregory and Taylor (1981) in the Oman ophiolite gabbros and by Ito and Clayton (1983) in samples from the Mid-Cayman slow-spreading center.

Assuming $\delta^{18}\text{O}$ values of hydrothermal fluids that range from 0 to 2, as measured in hydrothermal vents (Craig et al., 1980; Craig, 1981) or as predicted by thermodynamic models (Bowers and Taylor, 1985), and water-rock ratios lower than 1, closure temperatures of hydrothermal alteration for plagioclases are calculated in the range from 400°C to 600°C using the mass-balance equation for open sys-

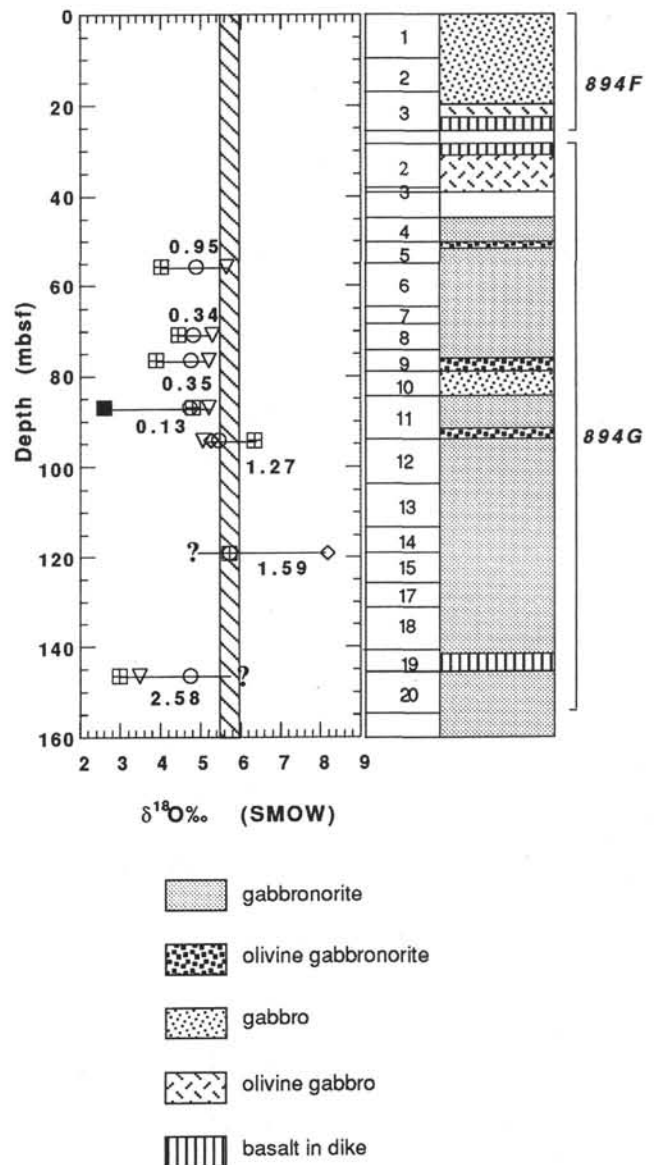


Figure 4. Distribution of mineral $\delta^{18}\text{O}$ values with depth (meters below seafloor). The hatched area shows the range of $\delta^{18}\text{O}$ values of fresh magmatic rocks derived from an oceanic mantle source. Numbers in bold represent the loss on ignition. Other numbers refer to core sections.

tems proposed by Taylor (1977). These “isotopic temperatures” are in agreement with the pervasive presence of amphibolite assemblages throughout Hole 894G.

Most magmatic pyroxenes are characterized by $\delta^{18}\text{O}$ values in the range from 5.1 to 5.3, which are lower than the expected value of 5.7 ± 0.2 for mantle-derived pyroxenes (Taylor and Epstein, 1962a, 1962b; Anderson et al., 1971). Ito and Clayton (1983) analyzed clinopyroxenes from Mid-Cayman Rise gabbros that also have low $\delta^{18}\text{O}$ values (from 4.8‰ to 5.0‰). Ito and Clayton (1983) attributed these isotope compositions to subsolidus exchange between clinopyroxenes, plagioclase, and Fe-Ti oxides during the slow cooling of the plutonic rocks. As a consequence of the subsolidus isotope reequilibration, plagioclase becomes ^{18}O enriched (slightly higher than 6), whereas pyroxene (4.8–5.0) and Fe-Ti oxide (down to 0.5) become ^{18}O depleted. A few mineral separates from Hess Deep gabbros do not support this hypothesis (Table 2). The existence of clinopyroxene

(Sample 147-894G-6R-1, 94–100 cm) and ilmenite (Sample 147-894G-11R-3, 1–3 cm) with $\delta^{18}\text{O}$ values of 5.7 and 2.6, respectively, suggests that these rocks that did not undergo significant subsolidus exchange.

The oxygen isotope composition of pyroxenes can be modified by exchange with a fluid phase at high temperature and with large amounts of fluids, as proposed by Gregory and Criss (1986) who studied pyroxene-plagioclase disequilibrium in gabbros from the Oman ophiolite. In the case of the Hess Deep gabbros, the Raman spectroscopic analysis reveals the existence of OH^- groups trapped in the magmatic pyroxenes: these correspond to the presence of tiny (1–5 mm) Mg-rich amphibole lamellae only detectable by a scanning electron microscopy (SEM) study (Lécuyer et al., 1993). These results reveal that plagioclase (the mineral most sensitive to oxygen isotope exchanges) governs the lowering of whole-rock $\delta^{18}\text{O}$ values and that most pyroxenes were involved in oxygen isotope exchange with an aqueous phase. Despite the low water-rock ratios determined by the Sr isotopes, the fluid circulation is pervasive without involving macroscopic deformation such as faulting, cracking, or shearing. This is one of the most striking results inferred from the Hess Deep gabbros. Indeed, Ito and Clayton (1983) emphasized the strong relationship between the deformation (fracturing and shearing) and alteration of Mid-Cayman gabbros. In this area, similar oxygen isotope compositions were measured on amphibolitized gabbros.

In the case of Hess Deep, we propose that at temperatures of 400°–600°C, compatible with an early hydrothermal activity close to the ridge system, the fluid circulation is likely dominated by fluid percolation along grain boundaries and the isotopic exchange by oxygen self-diffusion through plagioclase. Indeed, the preservation of magmatic compositions and zonations of plagioclases (rims An_{50} and cores An_{65}), and the absence of crystal deformation demonstrate that no major cation exchange with a fluid occurred, the low $\delta^{18}\text{O}$ values can be simply accounted for by solid state diffusion of oxygen under hydrothermal conditions (Giletti et al., 1978). On the other hand, self-diffusion of oxygen is too slow in pyroxenes (Farver, 1989), their oxygen isotope ratios were only slightly modified by the growth of tiny amphibole lamellae. In these conditions of high-temperature fluid-rock interactions, significant oxygen isotopic exchanges (Fig. 5) can thus occur without involving a dramatic hydration of the rocks. These results reveal that channeled fluid flow in the oceanic layer 3 (Manning and MacLeod, this volume) is accompanied by pervasive fluid flow down to the grain scale during the high temperature hydrothermal stage that is responsible for both mineralogical and isotopic alteration of gabbros.

The four olivine basalts analyzed for their oxygen isotope compositions (Table 1; Fig. 1) do not show any evidence of ^{18}O depletion, a result that must be linked to the lack of amphibolite facies mineral assemblages. However, their oxygen isotope compositions remain close to those of a few strongly hydrated gabbros at the bottom of the Hole (Table 1; Fig. 1). Higher $\delta^{18}\text{O}$ values relative to gabbros could be explained by lower closure isotopic temperatures as the consequence of the small grain size (0.5 mm) of these rocks. Higher hydration rates could be also caused by a fluid flow channeled along the dike margins, a process suggested by Nehlig and Juteau (1988) to promote fluid circulation in the sheeted dikes of the Oman ophiolite.

In terms of the oxygen isotope record, the alteration of Hess Deep gabbros is dominated by a high-temperature stage. Nevertheless, a low-temperature circulation overprints the early stage of hydrothermal alteration in the areas very close to the vein networks (for a more detailed study, see Früh-Green et al., this volume). This is the case for Sample 147-894G-2R-3, 20–26 cm, which has a whole-rock $\delta^{18}\text{O}$ value of 6.5. Traces of the low-temperature event can also be found, surprisingly, in gabbros that show mantle-like oxygen ratios such as Sample 147-894G-15R-1, 32–35 cm ($\delta^{18}\text{O} = 5.7$), and Sample 147-894G-12R-1, 45–51 cm ($\delta^{18}\text{O} = 5.5$). Chlorite is generally present in gabbros that experienced some retrograde reactions into the green-

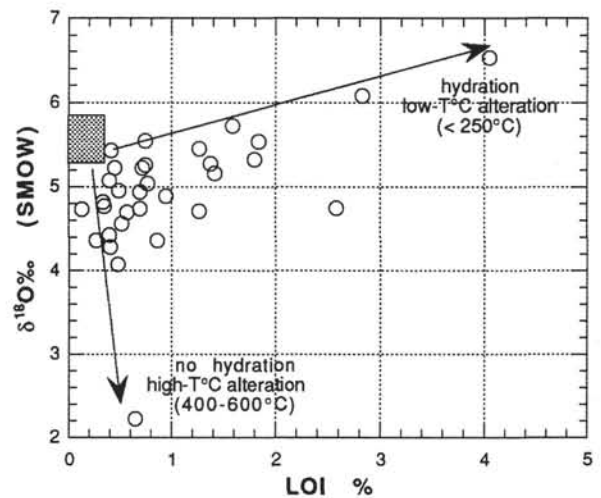


Figure 5. Variations of whole-rock $\delta^{18}\text{O}$ values vs. loss on ignition. The hatched area represents the field of fresh mantle-derived rocks.

schist or zeolite facies. Oxygen isotope composition of chlorite can be used as a thermometer if we take the fractionation equation serpentine/chlorite-water of Wenner and Taylor (1973) as a good approximation. In Sample 147-894G-15R-1, 32–35 cm, chlorite occurs as isolated patches and has a high $\delta^{18}\text{O}$ value (8.2), which suggests precipitation from seawater at a temperature that does not exceed 100°C. This chlorite probably was contemporaneous with the vein network filled by the zeolite facies assemblage: chlorite, prehnite, and zeolite. Chlorites ($\delta^{18}\text{O} = 5.3$) crystallized also at higher temperatures ($\approx 300^\circ\text{C}$) in the stability field of greenschist facies assemblages that occur in Sample 147-894G-12R-1, 45–51 cm (Table 2; Fig. 4). Früh-Green et al. (this volume) calculated approximate temperatures of 400°–450°C for chlorites in serpentinized peridotites from Holes 895C and 894E, which are characterized by lower $\delta^{18}\text{O}$ values (from 1.5 to 3) than those analyzed in gabbros from Holes 894F and 894G. The large range of oxygen isotope compositions measured in chlorites from Hess Deep rocks reveals that this mineral is stable over a large span of temperatures during the hydrothermal alteration of the oceanic crust.

These results suggest that in some restricted areas the low-temperature fluids can overprint the high-temperature event that characterizes the background alteration (Fig. 5). This isotopic overprint is linked to the high densities of anastomosing macroscopic vein networks (0.3 cm^{-1}) that occur in Cores 147-894G-2R, 4R, 6R, 12R, 15R, and 20R (Gillis, Mével, Allan, et al., 1993; Manning et al., this volume). Apparently, the low-temperature fluids percolated through the wallrock and precipitated the greenschist facies or zeolite facies minerals responsible for the strong oxygen isotope disequilibrium between minerals observed in these gabbros.

CONCLUSION

Pervasive background alteration of the gabbros occurred without deformation under amphibolite facies conditions (400°–600°C) and was responsible for ^{18}O depletion of these rocks ($\delta^{18}\text{O} = 4.8$, $N = 31$). Plagioclase controls most of the oxygen isotope exchanges with the aqueous fluid phase. As a consequence, many low $\delta^{18}\text{O}$ gabbros are only weakly hydrated ($\text{LOI} \leq 0.7\%$).

Water-rock ratios lower than 1 were calculated using Sr concentrations and isotope ratios. However these probably underestimate the real water-rock ratios because, as suggested by the oxygen isotope study, the $\delta^{18}\text{O}$ value of plagioclase grains was lowered by self-

diffusion of oxygen under hydrothermal conditions without modification of the major element chemistry (An₅₀₋₆₅).

A low-temperature fluid circulated in gabbroic levels where the vein density reaches 0.3 cm⁻¹. This circulation led to whole-rock δ¹⁸O values close to normal mantle values (5.5–5.7) or slightly higher values (6.5), and stronger hydration (LOI = 1%–2%).

Holes 894F and 894G define, for the first time, a remarkable section of low δ¹⁸O gabbros hydrothermally altered beneath a fast-spreading ridge. We do not yet know whether the ¹⁸O depletion of Hole 894G gabbros is representative of the entire lower oceanic crust. In the absence of ¹⁸O-depleted gabbro zones, the oceans should evolve to lower δ¹⁸O values.

The existence of these low δ¹⁸O gabbros clearly validates the models of oceanic hydrothermal circulation elaborated on the basis of the Oman ophiolite. Gregory and Taylor (1981) presented an oxygen isotope profile in which gabbros clearly show a ¹⁸O depletion similar to our Hess Deep samples.

These observations confirm that seawater-derived fluids react with the deep layers of the oceanic crust generated at fast-spreading ridges.

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