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ABSTRACT

We present oxygen and strontium isotopic data for olivine tholeiites, evolved (that is, differentiated and (or) contaminated) lavas, rhyolites, and crustal-derived xenoliths from the Snake River Plain. These data show that the olivine tholeiites are fairly uniform in $\delta^{18}O$ (5.1 to 6.2) and $^{87}Sr/^{86}Sr$ (0.7056 to 0.7076) and reveal no correlation between these ratios. The tholeiites are considered representative of mantle-derived magmas that have not interacted significantly with crustal material or meteoric water. The evolved lavas display a wider range in $\delta^{18}O$ (5.6 to 7.6) and $^{87}Sr/^{86}Sr$ (0.708 to 0.717) with positive correlations between these ratios in some suites but not in others. Crustal xenoliths have high and variable $^{87}Sr/^{86}Sr$ (0.715 to 0.830) and $\delta^{18}O$ values that vary widely (6.7 to 9.2) and are a few permil greater than $\delta^{18}O$ values of the Snake River basalts. Thus, isotopic data for the evolved lavas are permissive of small degrees of contamination by crustal rocks similar to the most ^{18}O-depleted xenoliths. The $\delta^{18}O$ enrichments in some evolved lavas also are consistent with crystal fractionation processes and do not necessarily require bulk interaction with crustal rocks. Enrichment in ^{18}O but not in $^{87}Sr/^{86}Sr$ in one suite of evolved lavas suggests that crustal contamination may not be essential to the petrogenesis of those lavas. Other suites of evolved lavas display large variations in $^{87}Sr/^{86}Sr$ that reflect at least some selective contamination with ^{87}Sr.

Bulk solid/liquid oxygen-isotope fractionation factors (α's) calculated for the evolved lavas from Craters of the Moon National Monument are comparatively large. These α's are dependent upon the nature and proportions of phases removed by crystal fractionation; basaltic lava α's differ from latitic lava α's in accordance with different phenocryst assemblages in these rocks.

Snake River Plain rhyolites are isotopically distinct from both the analyzed crustal xenoliths and olivine tholeiites. Their origin remains poorly understood, but crustal or sub-crustal sources may be viable. In the first case, they must be derived by anatexis of material distinct from the analyzed crustal xenoliths. In the second case, they must be derived from material unlike the source for tholeiites. No cogenetic relation with the tholeiites seems likely on the basis of available data.

INTRODUCTION

The Snake River Plain (SRP) is a bimodal basalt-rhyolite volcanic province that extends for 500 km across southern Idaho from the Columbia River Plateau eastward to Yellowstone National Park. Structurally, the western SRP is considered to be a graben bounded by high-angle faults (Malde, 1959), whereas the eastern SRP is essentially a downwarp controlled in part by margin faults (Kirkham, 1931; Trimble and Carr, 1976). Stratigraphic relationships within the SRP (for example, Malde and Powers, 1962; Carr and Trimble, 1963; Stearns and others, 1938) and absolute ages of the volcanic rocks (Armstrong and others, 1975) indicate that volcanic activity became progressively younger toward the east. In any locality within the SRP, volcanism began with voluminous eruptions of high-silica rhyolitic flows and ash-flow tuffs, and was followed by dominantly basaltic eruptions. Initial rhyolitic volcanism ranges in age from about 15 m.y. old to about 2 m.y. old from west to east whereas the oldest basalts range in age from about 13 m.y. old in the west to less than 1. m.y. old in the east.

The basaltic rocks are dominantly olivine tholeiites characterized by comparatively high iron enrichment, high TiO_2 (average 2.7 weight percent) and P_2O_5 (0.6 percent), low SiO_2 (46.9 percent) and total alkalies (3.1 percent) (Powers, 1960a; Stone, 1967; Leeman, 1974). The basalts are relatively uniform in composition regardless of age or location. In addition, minor volumes of "evolved" lavas occur situated mainly on the margins of the SRP. These rocks are typified by lavas at Craters of the Moon (COM) National Monument, which range widely in composition (from ferrobasalt to tholeiitic latite) as the result of low pressure crystal fractionation (Leeman and others, 1976; Stone, 1970). The COM-type lavas may be related to olivine tholeiite parental magmas via high-pressure differentiation (Thompson, 1975; Leeman and others, 1976). However, the presence of crustal-derived xenoliths in some of the COM-type lavas, and anomalies in their strontium and lead isotopic compositions (Leeman and Manton, 1971; Leeman, 1974 and in preparation) suggest that they may also be contaminated by crustal material.

The eastward younging of SRP volcanism, as well as geologic and geophysical information for this province, have led numerous authors to suggest that volcanism there is related to movement of the North American plate over a mantle hot-spot presently located beneath the Yellowstone volcanic complex (for example, Smith and Sbar, 1974; Suppe and others, 1975; Eaton and others, 1975; Thompson, 1977; Brott and others, 1978). To evaluate this model it is desirable to determine the source(s) of the SRP magmas.

Here we report oxygen isotopic analyses of representative basalts and crustal xenoliths from the SRP for which strontium isotopic analyses are available. Analyzed samples include: (a) 19 olivine tholeiites covering a broad temporal and spatial distribution, (b) 26 evolved COM-type lavas from five localities, (c) 8 crustal xenoliths from three localities, and (d) 6 rhyolite flows and ash-flow tuffs. These data augment analyses by Muehlenbachs and Stone (1973) for nine other SRP samples and, in combination with strontium isotopic and major-element data, provide important constraints on possible source materials for SRP magmas and on processes by which they evolved.

ANALYTICAL METHODS

Identical splits of sample powders (about 200 mesh) were used in this study for both strontium and oxygen isotopic analyses. Many of the strontium isotopic analyses are taken from Leeman and Manton (1971), and the remaining samples were analyzed on a 6-inch-radius NBS design mass spectrometer with triple filament ionization located at the U.S. Geological Survey, Denver, Colorado. All isotopic ratios are normalized to a $^{87}Sr/^{86}Sr$ ratio of 0.1194, and $^{87}Sr/^{86}Sr$ ratios are adjusted to a value of 0.7080 for the Eimer and Amend $SrCO_3$ standard. Within- and between-run precision for this ratio is about ±0.0002 (two-sigma) or better.

After drying the sample powders for one hour at 200°C, oxygen was extracted by reaction with BrF_5 at 550°C in nickel vessels (Clayton and Mayeda, 1963). Oxygen was converted to CO_2 by combustion with a resistance-heated carbon rod (Taylor and Epstein, 1962). Isotopic analyses of CO_2 were performed on a Nuclide 60 RMS mass spectrometer with modified source, collector, and emission regulator. The isotopic data are expressed as δ, the per mil deviation relative to SMOW (Craig, 1961). Twenty-two analyses of NBS-28 reference standard performed during this study yielded a mean $\delta^{18}O$ value of 9.53 ± 0.36 (two-sigma), and three analyses of St. Peter sandstone yielded a mean value of 11.10 ± 0.16 (two-sigma). The unusually large 2-sigma of NBS-28 reflects some severe O_2-extraction problems during the final stages of the study. The average $\delta^{18}O$ difference between replicate analyses of 22 individual samples (tables 1, 2, and 3) is 0.1 permil.

RESULTS

Olivine tholeiites

SRP olivine tholeiites were selected for analysis to provide representative spatial and temporal coverage. Our samples are divided into two age groups in table 1; lavas of the Snake River Group (that is, less than about 1 m.y. old), and those equivalent to the Glenns Ferry and Banbury formations of Malde and Powers (1962). The latter rocks range in age from about 3 to 10 m.y. old and, despite an attempt to sample fresh material, some of these rocks display incipient iddingsitic alteration of olivines. Both groups include samples from the western and eastern parts of the SRP.

The Snake River Group lavas display a wide range in K_2O content (0.14 to 0.83 weight percent), but narrow ranges in $\delta^{18}O$ (5.12 to 5.99) and $^{87}Sr/^{86}Sr$ (0.7056 to 0.7068), and show little or no correlation between these compositional features. The older lavas also display a wide range in K_2O (0.10 to 0.64) and similar ranges in $\delta^{18}O$ (5.06 to 6.21) and $^{87}Sr/^{86}Sr$ (0.7065 to 0.7076). Average $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ ratios (table 1) are slightly higher for the older group, but the differences are statistically insignificant. Other compositional features of these respective groups are similar except that the older lavas have higher average potassium/rubidium (580) than the younger group (470) (Leeman, 1974). The average $\delta^{18}O$ (5.59 ± 0.30) and $^{87}Sr/^{86}Sr$ (0.7064 ± 0.0005) for all analyzed SRP tholeiites are considered representative of these lavas (Leeman and Manton, 1971).

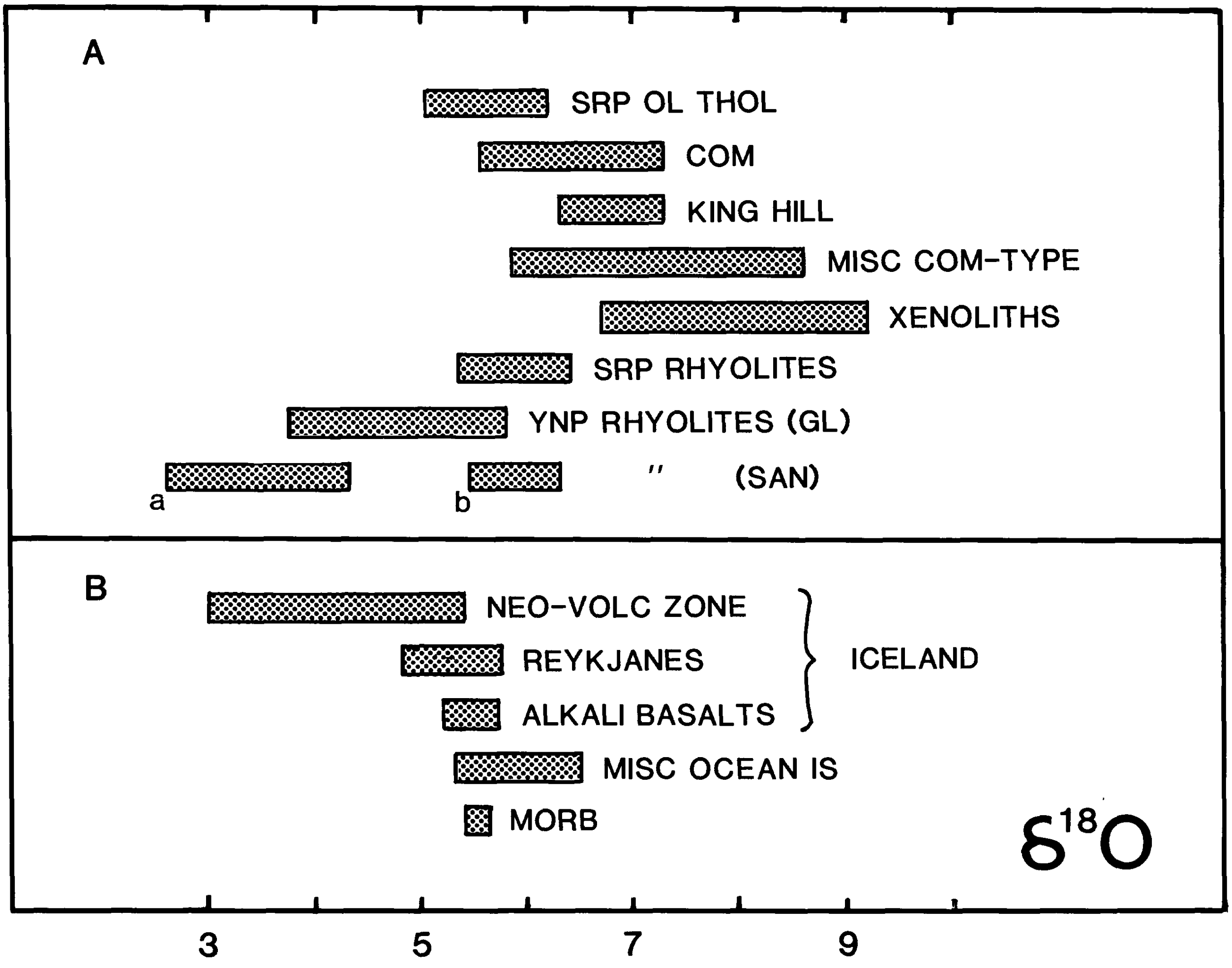

Figure 1A. Range in $\delta^{18}O$ values for Snake River Plain (SRP) and Yellowstone National Park (YNP) volcanic rocks and crustal xenoltihs. Data for basalts from this report. Data for rhyolites: SRP (this report), YNP non-hydrated rhyolite glasses (Friedman, personal commun., 1975), YNP sanidines (Friedman and others, 1974) (a - Mesa Falls Tuff and Plateau Rhyolite; b - Huckleberry Ridge Tuff). (1B) Range in $\delta^{18}O$ values for selected oceanic island and mid-ocean-ridge basalts. References are given in text.

Table 1. Oxygen and initial strontium isotopic compositions and K_2O content of Snake River olivine tholeiites

Analysis No.	Sample No. and description	K_2O	$\delta^{18}O$	$^{87}Sr/^{86}Sr$
1	6YC-142, Gerrit Basalt Island Park	0.19	5.5 ± 0.05	0.7061
2	73-64, unnamed flow, W of Island Park	0.31	5.5	0.7063
3	73-77, Monument Butte SW flow	0.32	5.7 ± 0.01	0.7056
4	SBR-2, unnamed flow, Idaho Falls	0.54	5.6 ± 0.00	0.7056
5	SBR-3, Hells Half Acre lava field	0.60	5.3	0.7065
6	SBR-6, unnamed flow, E of Shoshone	0.65	5.5	0.7066
7	69-G, Crystal Ice Caves flow	0.54	5.5 ± 0.13	0.7056
8	69-14, Wapi lava field	0.52	5.5	0.7063
9	69-49, Wendell Grade Basalt	0.83	5.3	0.7068
10	69-56, McKinney Basalt	0.67	5.1	0.7064
11	70-15Gl, do, pillow glass	0.97	6.0 ± 0.01	0.7066
12	70-15Pl, do, plagioclase	0.17	5.8 ± 0.07	0.7066
13	74-26, unnamed flow, NE of COM lava field	0.14	5.8 ± 0.15	0.7063
Average of 1-13 (Snake River Group)		0.69[a]	5.55 ± 0.24	0.7063±0.0004
14	69-47, upper Banbury Basalt	0.40	5.1	0.7066
15	70-1, Banbury Basalt	0.44	6.0	0.7065

Table 1. Oxygen and initial strontium isotopic compositions and K_2O content of Snake River olivine tholeiites (cont'd)

Analysis No.	Sample No. and Description	K_2O	$\delta^{18}O$	$^{87}Sr/^{86}Sr$
16	70-16, dike in Deer Gulch Basalt	0.64	5.7	0.7076
17	70-18, Massacre Rocks plug-vent	0.49	5.7	0.7067
18	74-C8, capping basalt Malta Range	0.10	6.2 ± 0.05	0.7070
Average of 14-18 (Glenns Ferry and Banbury Fms.)		0.49[a]	5.74 ± 0.42	0.7069±0.0004
Average of 1-18		0.58[a]	5.60 ± 0.30	0.7064±0.0005
19	72-36B, highly altered Deer Gulch Basalt	--	8.5	0.7086

[a] average K_2O contents are given for 27 Snake River Group samples, 32 pre-Snake River Group (that is, Glenns Ferry and Banbury) samples, and for 59 combined Snake River basalts.

All oxygen isotopic values are given relative to SMOW, with standard deviations of averaged data where replicate analyses were made. Strontium isotopic ratios are given relative to a value of 0.7080 for the E&A standard; in-run and between-run precision of these ratios are both about ± 0.0002 (two sigma).

A single sample (72-36B) of highly altered basalt has significantly higher $\delta^{18}O$ (8.5) and $^{87}Sr/^{86}Sr$ (0.7086) than the fresh tholeiites. This sample represents the deuterically altered lower portion of a flow that now contains a large proportion of montmorillonite-rich phyllosilicates and secondary calcite (which was removed from the analyzed portion prior to crushing). These data are indicative of the low-temperature alteration processes which result in significant changes in oxygen and strontium isotopic compositions (Lawrence and Taylor, 1972). It is possible that the higher average $\delta^{18}O$ values and $^{87}Sr/^{86}Sr$ ratios in the older SRP tholeiites (samples 14-18) reflect mild alteration, although none of these samples contain obvious alteration minerals, nor are their water contents significantly higher than younger basalts. There is no correlation between $\delta^{18}O$ and either water content or Fe_2O_3/FeO ratio for the analyzed tholeiites.

Muehlenbachs and Stone (1973) analyzed four Snake River Group basalts. Their $\delta^{18}O$ values (5.0 to 5.6) are in good agreement with our data. The only discrepancy between the data sets is in analyses of pillow glass from McKinney Basalt; they reported a value of 5.0 per mil compared to our value of 6.00; however, different samples were analyzed.

Oxygen isotopic data for SRP tholeiites are similar to results for most fresh oceanic island and mid-ocean-ridge basalts (Garlick, 1966; Taylor, 1968; Anderson and others, 1971; Muehlenbachs, and Clayton, 1972; Pineau and others, 1976) which display a narrow range in $\delta^{18}O$ (about 5.3 to 6.2). In contrast, tholeiitic basalts of Iceland are distinct from most other oceanic island basalts in having variable and systematically lower $\delta^{18}O$ (Muehlenbachs, and others, 1974). As also suggested for igneous complexes in northwestern Scotland (Taylor and Forrester, 1971) and at Yellowstone National Park (Friedman and others, 1974), these lower values may reflect interaction with low-$\delta^{18}O$ meteoric water or hydrothermally altered crustal rocks (Muehlenbachs and others, 1974). The SRP tholeiites show no evidence of such interaction with altered crustal rocks or meteoric water. Figure 1 shows the contrast between oxygen isotopic data for SRP tholeiites and those for evolved COM-type lavas, SRP and Yellowstone rhyolites, and oceanic volcanic rocks.

The range in $\delta^{18}O$ for SRP tholeiites is somewhat greater than that for fresh mid-ocean-ridge tholeiites, suggesting the possibility that the SRP basalts may be contaminated by high $\delta^{18}O$ crustal rocks such as the xenoliths listed in table 3. Figure 2 shows that no correlation exists between $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ for the tholeiites as would be expected if they were contaminated. Furthermore, figure 3 indicates a slight negative correlation between K_2O and $\delta^{18}O$ in these rocks, which is inconsistent with contamination by ^{18}O- and K_2O-rich sialic crustal rocks. We conclude that the tholeiites are representative of uncontaminated mantle-derived magmas.

Evolved COM-type lavas

Lavas from Craters of the Moon National Monument range in composition from ferrobasalt to tholeiitic latite, (Leeman and others, 1976). Lavas from four other COM-type localities are more restricted in composition but include varieties typical of the COM suite. Each sample in table 2 represents a different flow unit except for numbers 21 and 22 (Lava Creek flow) and numbers 29 to 31 (Devils Orchard flow). The latter flow is somewhat heterogenous

Table 2. Oxygen and initial strontium isotopic compositions and rubidium and K_2O contents of Craters of the Moon-type lavas

Analysis No.	Sample No. and Description	K_2O (%)	Rb (ppm)	$\delta^{18}O$	$^{87}Sr/^{86}Sr$
Craters of the Moon lava field (Holocene)					
20	69-28, rift flow	1.21	24	5.6 ± 0.00	0.7075
21	69-22, Lava Creek flow	1.59	30	5.9 ± 0.10	0.7077
22	70-42, do	1.55	32	5.9 ± 0.06	0.7081
23	V-28, Grassy Cone flow	1.85	38	6.1	0.7090
24	WPL-87, unnamed flow	2.05	41	6.0 ± 0.03	0.7087
25	69-33, rift flow	2.15	44	6.2	0.7100
26	SBR-5, Big Craters NW flow	2.34	48	6.5	0.7097
27	69-17, unnamed flow	2.70	55	6.5	0.7100
28	60-35, ejecta block	3.35	71	6.5 ± 0.03	0.7120
29	69-19, Devils Orchard	3.63	79	6.7	0.7103
30	69-20, do, glassy block	4.50	94	6.8	0.7104
31	V-31, do	4.67	96	6.6	0.7108
32	V-27, Highway flow	4.81	104	6.6	0.7113
King Hill lavas (Pliocene-Pleistocene)					
33[a]	OB-32, basalt	2.00	31	6.3	0.7100
34	69-61, basalt	2.49	30	6.4 ± 0.06	0.7106
35	69-62, basalt	2.37	34	6.4 ± 0.01	0.7118
36	70-9, basalt	1.96	33	6.5 ± 0.02	0.7110
37	69-60, basalt	2.90	47	7.0 ± 0.08	0.7128
38	69-63, hybrid lava	4.57	79	7.3 ± 0.01	0.7163

Table 2. Oxygen and initial strontium isotopic compositions and rubidium and K_2O contents of Craters of the Moon-type lavas (cont'd)

Analysis No.	Sample No. and Description	K_2O (%)	Rb (ppm)	$\delta^{18}O$	$^{87}Sr/^{86}Sr$
Cold Springs Creek lava (Pliocene-Pleistocene)					
39	70-5, latite	--	138	5.9 ± 0.10	--
40[a]	59P-276A Glass	--	--	5.9	
Square Mountain Basalt (Pliocene)					
41	SM-1B, hybrid lava	2.80	60	7.6	0.7144
Island Park west rift (Pleistocene-Holocene)					
42	73-68GM, Crystal Butte SW flow, groundmass	1.18	28	6.4	0.7093
43[a]	72S-15GM, do, vent lava, groundmass	--	--	8.9	
44[a]	72S-15Pl, do, plagioclase	--	--	7.1	--
45[a]	72S-15Px, do, augite	--	--	6.3	--
46[a]	72S-16A, do, vent lava	--	--	8.6	--
47[a]	72S-16A Pl, do, plagioclase	--	--	7.2	--
48	73-106GM, Swan Butte scoria bomb, groundmass	1.64	52	6.0	0.7090
49	73-106Pl, do, plagioclase	--	--	6.9	0.7081
50	73-71, High Point latite	2.19	68	6.0	0.7060
51	73-59, do	3.66	120	6.2	0.7060
52	73-42, do	3.97	132	6.5 ± 0.07	0.7064

[a] analyses from Muehlenbachs and Stone (1973)

(table 2; Leeman and others, 1976) and contains xenoliths of crustal rocks. The King Hill lavas (Powers, 1960b; Leeman, unpublished data) comprise at least three flow units; all are basaltic except one (sample 69-63) which is a latite containing xenocrysts of quartz and alkali feldspar. Some of the basaltic samples contain megacrysts of intergrown orthopyroxene and labradorite, which are interpreted as high-pressure phenocrysts (Leeman, unpublished data). The first four samples (analyses 33 to 36, table 2) are thought to represent the lowermost flow of the King Hill lavas, but correlations between outcrops are indefinite and it is possible that more than one flow is represented. The Cold Springs Creek lavas (Stone, 1970) include several flows ranging in composition from basalt to mafic latite. Muehlenbachs and Stone (1973) sampled the most mafic flow (analysis 40) whereas our sample analysis is from a comparatively silicic latite flow that contains sparse small crustal xenoliths. The Square Mtn. Basalt (Schmidt, 1961) is a single flow unit that contains abundant crustal xenoliths and actually has a latitic composition (Leeman, unpublished data). The Island Park west rift comprises a fairly large volcanic field with two distinct groups of lavas -- olivine tholeiites similar to those elsewhere in the SRP (analyses 2 and 3, table 1), and transitional to slightly alkalic basalts (analyses 42 to 49, table 2). Several of the latter flows contain crustal xenoliths. In addition, latitic lavas similar to those at Craters of the Moon National Monument issued from the High Point vent on the west rim of Island Park caldera (Hamilton, 1965); three of these lavas were analyzed.

Both $\delta^{18}O$ (5.6 to 7.6) and $^{87}Sr/^{86}Sr$ (0.708 to 0.717) ratios are generally higher in the COM-type lavas than in the SRP tholeiites (fig. 1). At Craters of the Moon National Monument and at King Hill these ratios are strongly correlated as might be expected if these lavas are contaminated by crustal materials such as the analyzed xenoliths in table 3 (fig. 2). However, the High Point (Island Park) latites display no such correlation, and in fact resemble the SRP tholeiites with regard to strontium isotopic ratios. All three groups of COM-type lavas display a trend betwen $\delta^{18}O$ and K_2O contents (fig. 3), which may be consistent with either crystal fractionation or with crustal contamination processes.

Crustal xenoliths

Crustal xenoliths from three localities have been analyzed (table 3) in order to evaluate possible contamination processes. All of these rocks are granulite facies gneisses, some containing both ortho- and clinopyroxene. All of these gneisses are silicic (65 to 70 weight percent SiO_2), except 73-68X (57 weight percent SiO_2). High $^{87}Sr/^{86}Sr$ ratios (0.715 to 0.830), and the lead isotopic and other compositional features (Leeman, 1979 and in preparation) of these rocks indicate that they may be derived from ancient (ca. 3 b.y. old) crust beneath the SRP. Their $\delta^{18}O$ values (6.7 to 9.2) are in the range of many silicic crustal rocks (for example, Taylor, 1968; Flagstaffe, 1979); those with higher $\delta^{18}O$ values contain higher proportions of quartz than the others. The $\delta^{18}O$ values in these xenoliths are 1 to 2 permil higher than $\delta^{18}O$ values in most SRP basalts and rhyolites (fig. 1).

Table 3. Oxygen and initial strontium isotopic composition and K_2O content of crustal xenoliths in COM-type lavas

Analysis No.	Sample No.	K_2O	$\delta^{18}O$	$^{87}Sr/^{86}Sr$
Craters of the Moon Lava Field				
52	SI-1, two-pyroxene granulite	0.86	6.7 ± 0.03	0.7173
53	70-40, orthopyroxene granulite	2.01	8.0 ± 0.09	0.7152
54	CKI, two-pyroxene granulite	2.18	9.2	0.7271
Square Mountain Basalt				
55	SM-2A, granitic gneiss	4.34	7.5	0.7313
56	SM-2E, hybrid rock (latite and xenocrysts)	3.24	7.4	0.7423
57	SM-2F, granitic gneiss	4.59	7.0	0.8295
58	SM-2G, granitic gneiss	2.66	7.6	0.7212
Island Park West Rift				
59[a]	72S16A-X, granitic	--	9.1	--
60	73-68X, two pyroxene granulite	0.77	7.3	0.7255

[a] analysis taken from Muehlenbachs and Stone (1973)

Table 4. Oxygen and initial strontium isotopic composition and K_2O content of selected Snake River rhyolites

Analysis No.	Sample No.	K_2O	$\delta^{18}O$	$^{87}Sr/^{86}Sr$
61	69-12, Walcott tuff obsidian (6 m.y.)	5.36	6.1	0.7108
62	74-144, Wolverine Ck. tuff, obsidian (~4 m.y.)	5.17	5.7	0.7098
63	74-83, Cougar Ck. tuff, obsidian (13 m.y.)	5.51	-0.8	0.7096
64	74-23, Big Southern Butte, obsidian (0.3 m.y.)	4.67	5.8	n.d.
65	70-36, East Butte, rhyolite (0.6 m.y.)	4.88	5.4	0.7080
66	70-26, China Hat, rhyolite (0.05 m.y.)	4.11	6.4	n.d.

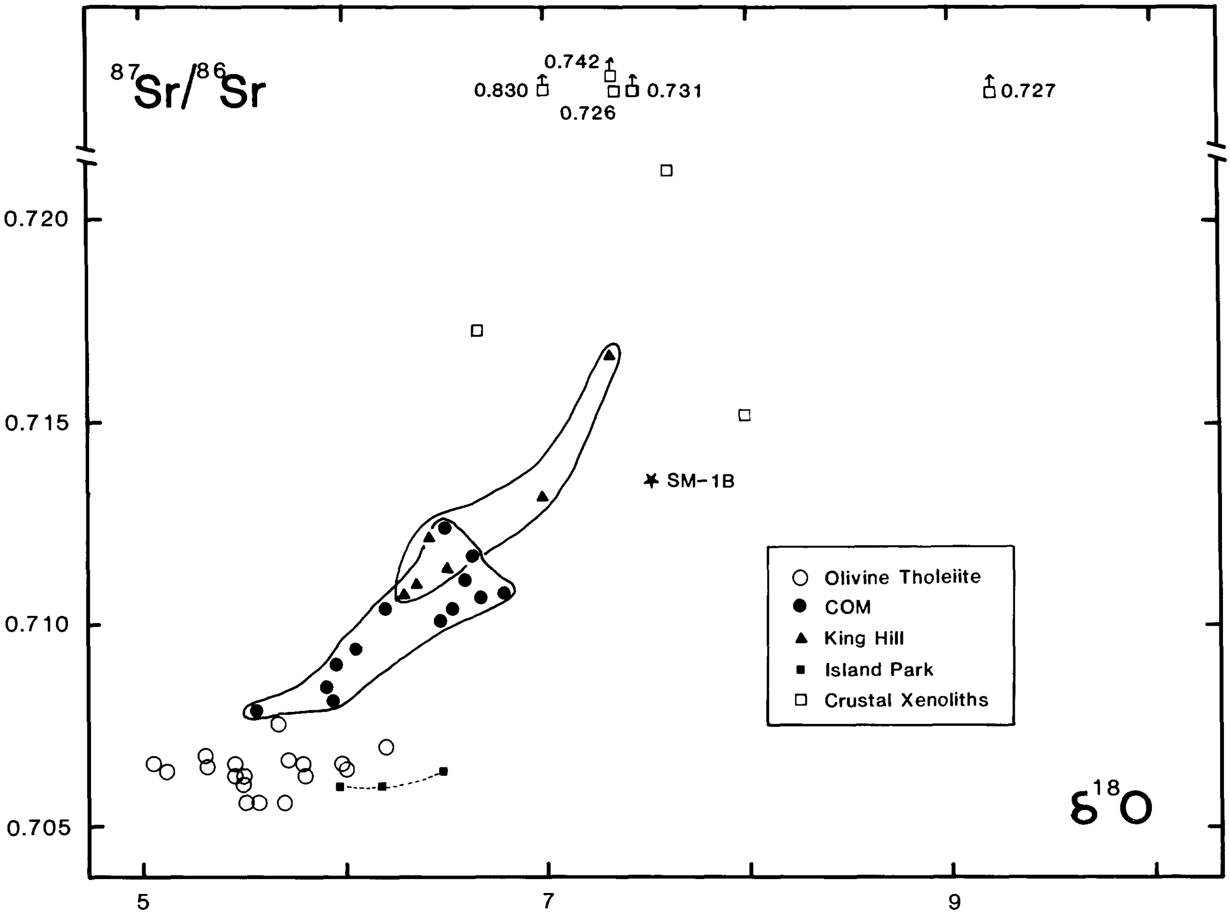

Figure 2. $^{87}Sr/^{86}Sr$ versus $\delta^{18}O$ plot for SRP olivine tholeiites, evolved COM-type lavas, and crustal xenoliths.

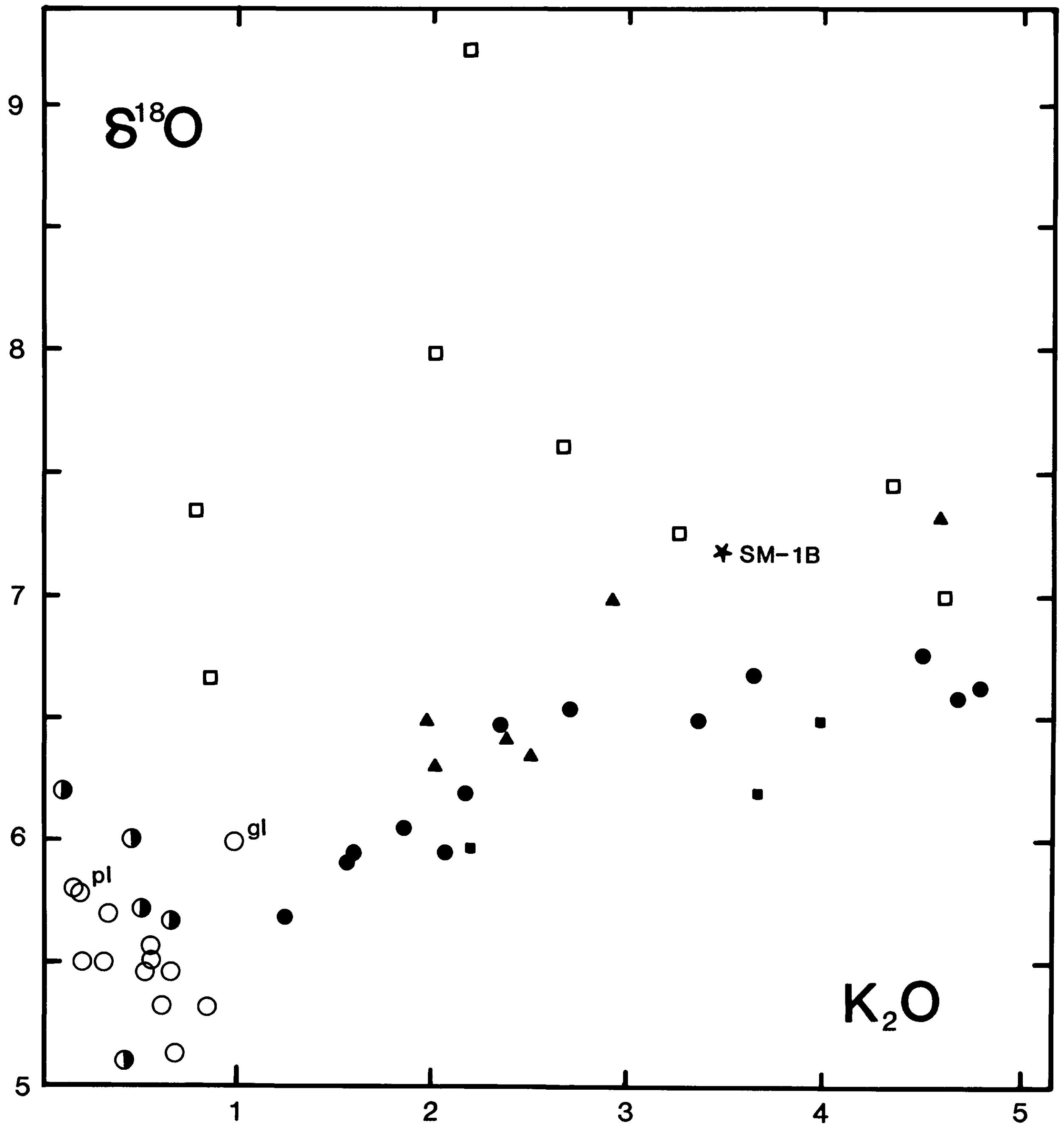

Figure 3. $\delta^{18}O$ versus K_2O plot for SRP olivine tholeiites, evolved COM-type lavas, and crustal xenoliths. ○ = young olivine theoliites, ◑ = old olivine tholeiites, ● = COM, ▲ = King Hill, ■ = Island Park, and □ = crustal xenoliths. gl and pl denote pillow glass and plagioclase, respectively, from sample 70-15.

Rhyolites

Six typical SRP rhyolites were selected for analysis (Table 4) to examine their role as possible contaminants for the younger basaltic magmas and to place some constraints on their origin. Excluding one sample (74-83), they define a narrow range in $\delta^{18}O$ (5.4 to 6.4) similar to that of many other rhyolites (Taylor, 1968) as well as that of the SRP olivine tholeiites. Obsidian from the Cougar Creek Tuff (74-83) is quite depleted in ^{18}O. This is not likely the result of low-temperature hydration processes, as the total H_2O content of this sample is only 0.30 weight percent. It is possible that depletion of ^{18}O took place via isotopic exchange between low-^{18}O hydrothermal fluids and the Cougar Point magmas as has been suggested for low-^{18}O rhyolites from the Yellowstone volcanic complex (Friedman and others, 1974).

DISCUSSION

Variations of strontium and oxygen isotopic compositions with major element contents (figs. 2-5) place certain constraints on the generation and evolution of SRP magmas. There are no significant correlations between these parameters for the olivine tholeiites. As concluded by Leeman and Manton (1971) and Leeman (1977) the range in $^{87}Sr/^{86}Sr$ in these rocks likely reflects isotopic heterogeneities in their mantle source regions. If, for example, the source area was isotopically homogenized during Archean time (ca. 2.5 b.y. ago, Leeman, 1975, 1977), the present-day range in $^{87}Sr/^{86}Sr$ could be explained by a range of only about 0.03 in its Rb/Sr ratio. The range in $\delta^{18}O$ may be enhanced by post-eruptive processes (deuteric alteration, weathering), but in any case the mean $\delta^{18}O$ value of 5.6 is equivalent to that of fresh oceanic basalts and precludes significant isotopic exchange with sialic crustal rocks. In summary, the available data are consistent with more or less direct derivation of olivine tholeiitic magmas from a mantle source that is slightly enriched in $^{87}Sr/^{86}Sr$ compared to sub-oceanic mantle (Leeman, 1977).

Isotopic data for COM-type lavas indicate a more complex petrogenesis. There are significant correlations between the isotopic data and most other compositional parameters that may be indicative of crustal contamination. On the other hand, evaluation of major and trace element data for the COM lavas (Leeman and others, 1976) indicates that much of their compositonal variation resulted from extreme fractional crystallization. Evaluation of these alternative processes is given below.

Evidence for Crustal Contamination

The large variations in $^{87}Sr/^{86}Sr$ in COM-type lavas are seemingly incompatible with a simple crystal fractionation model because the small relative mass difference between the various isotopes of strontium precludes significant chemical fractionation. In any event, routine correction for the mass fractionation that occurs during strontium isotopic analysis would obscure any natural fractionation effect. Therefore, it is concluded that the COM, Square Mountain, and King Hill magmas may have been contaminated by ^{87}Sr, perhaps as they cooled and differentiated within crustal-level magma reservoirs. On the other hand, lack of a correlation betwen $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ in the High Point

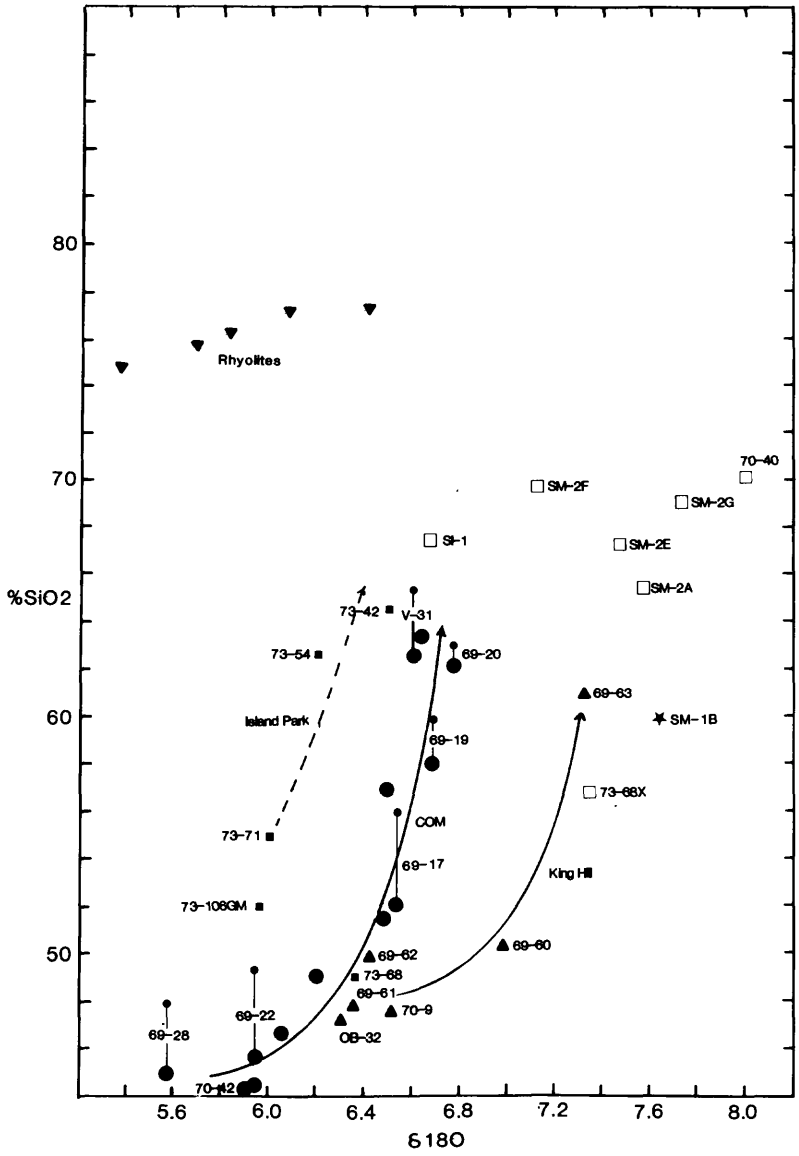

Figure 4. SiO_2 versus $\delta^{18}O$ plot for SRP samples. ○ = olivine tholeiites, ● = COM, ▲ = King Hill, ■ = Island Park, ▼ = rhyolites, and □ = crustal xenoliths. Compositional trend of Island Park lavas is dashed due to small sample population.

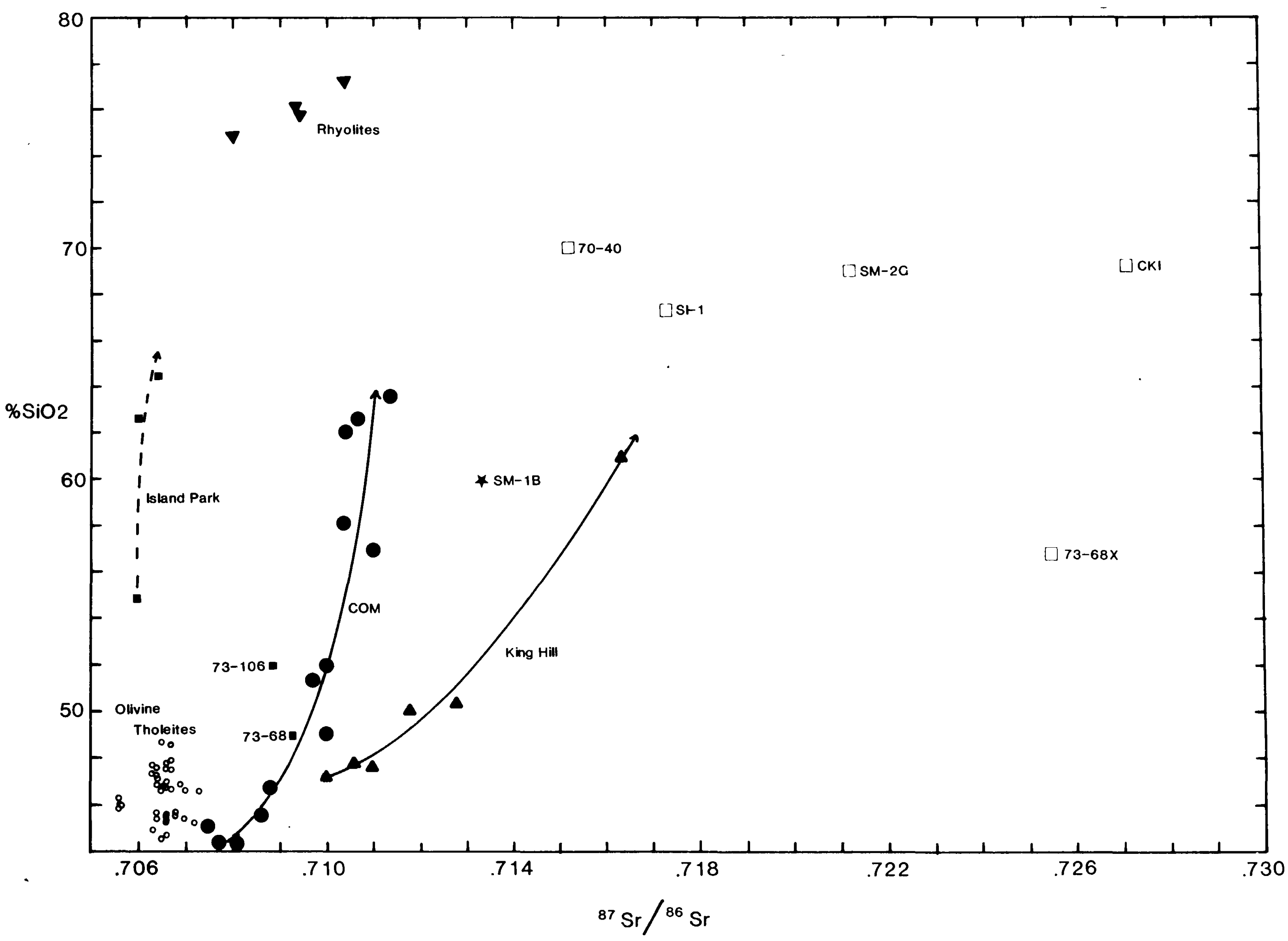

Figure 5. SiO_2 versus $^{87}Sr/^{86}Sr$ plot for SRP samples. ○ = olivine tholeiites, ● = COM, ▲ = King Hill, ■ = Island Park, ▼ = rhyolites, and ☐ = crustal xenoliths. Compositional trend of Island Park lavas is dashed due to small sample population

lavas suggests that such contamination processes are not an essential factor in the evolution of COM-type magmas. The low and uniform $^{87}Sr/^{86}Sr$ ratios in the High Point lavas are believed to represent the initial ratios in their parent magmas. They do not reflect contamination by ^{87}Sr which probably indicates that these magmas evolved in an environment that was isolated from ^{87}Sr-rich crustal rocks, either within the mantle or within the comparatively non-radiogenic sub-structure of the Island Park volcanic complex. In contrast, some xenolith-bearing basaltic lavas from the rift zone west of Island Park are enriched in ^{87}Sr and ^{18}O and display isotopic disequilibrium between groundmass and phenocryst minerals. In basalts, plagioclase $\delta^{18}O$ values are typically ~0.3 greater than those of coexisting glassy matrix (Taylor, 1968). Analyses 42 to 49 (table 2) show that this is not the case for the rift zone west of Island Park; this probably reflects interaction between crustal rocks and at least some of the basaltic magmas erupted there. As noted earlier, olivine tholeiites (analyses 2 and 3, table 1) from this rift zone have oxygen and strontium isotopic compositions typical of olivine tholeiites elsewhere in the SRP. Thus, details on the magma plumbing systems and the accessibility of sialic crustal rock control whether or not crustal contamination occurs. In this regard, most COM-type lavas issued from vents located near margins of the SRP, where access to deep crustal sialic rocks is most probable.

In general, xenocryst- or xenolith-bearing suites from COM, Square Mountain, and King Hill display strong enrichments in both $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ coupled with parallel enrichments in SiO_2, K_2O and other large-ion lithophile elements. In figures 4 and 5, SiO_2-enrichment trends for these suites and the High Point latites are distinct, but more or less internally coherent, and are compatible with incorporation of crustal material similar to some of the analyzed xenoliths. Most of the xenoliths are too enriched in $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ to lie on reasonable mixing lines with the COM-type lavas. It is possible that the xenoliths are not entirely representative of the sialic crust beneath (or near margins of) the SRP, and that suitable materials do exist but have not been sampled. Alternatively, it is possible that selective contamination occurred by assimilation of partial melt components from suitable crustal rocks. Isotopic disequilibrium between phenocrysts and groundmass for some xenolith-bearing lavas clearly suggests that contamination is a viable process (Muehlenbachs and Kushiro, 1974; Magaritz and others, 1978). In general, the contamination processes must be complex and differ from one place to another, due in part to differences in compositions, volumes, etc. of the magmatic and crustal components.

The analyzed rhyolites are not likely contaminants for evolved COM-type lavas as they have relatively lower $\delta^{18}O$, and in some cases lower $^{87}Sr/^{86}Sr$. Thus they do not lie on suitable mixing curves for the evolved lava suites.

The systematic isotopic variations observed for COM-type lavas are not observed for the olivine tholeiites, and this relationship strengthens the conclusion that the tholeiites have not experienced comparable interactions with crustal materials. Further considerations of contamination processes, based on lead isotopic data, will be given elsewhere (Leeman, in preparation). Despite the strong evidence for at least local crustal contamination, the oxygen isotopic data display correlations with major and trace element data in COM lavas that indicate that bulk assimilation of crustal material is not a major factor in the petrogenesis of these lavas. Specifically, the data are also compatible with the operation of extreme crystal fractionation.

Evaluation of crystal fractionation processes

Because oxygen isotopic compositions of minerals and silicate melts differ slightly (Taylor and Epstein, 1962; Garlick, 1966; Anderson and others, 1971), fractional crystallization can produce isotopic fractionation in the residual liquids. Such isotopic fractionation can be expressed by the Rayleigh equation

$$R/R_0 = F^{(\bar{\alpha} - 1)}, \qquad (1)$$

where R_0 is the initial $^{18}O/^{16}O$ ratio in the parental magma, R is that ratio in the residual liquid at a given stage of solidification, F is the fraction of original liquid remaining at that stage, and α is the isotopic fractionation factor between bulk solids and liquid. Considering that in δ-notation, $\delta^{18}_{A}O - \delta^{18}_{B}O = 10^3 \ln (R_A/R_B)$, the above equation can be rearranged as follows:

$$\underset{\text{liquid}}{\delta^{18}O} = 10^3 (\bar{a} - 1) \ln F + \underset{\text{parent}}{\delta^{18}_{0}O} \qquad (2)$$

Using Kuno's Solidification Index to approximate F, Matsuhisa and others (1973) applied this expression to estimate α for several Japanese volcanoes.

Leeman and others (1976) computed values of F for COM lavas using a least-squares mixing program (Bryan and others, 1969) and major element analyses of rocks and their constituent phenocrysts, and assuming that these lavas were derived by crystal fractionation from a common parental magma. For the purposes of the calculations, the Lava Creek flow (sample 69-22) was taken as the parent magma because it is the oldest as well as the most mafic lava in the series, except for a comparatively young rift lava (sample 69-28). The Lava Creek flow can be related to sample 69-28 by removal from the latter of 23 percent of olivine, plagioclase, and minor apatite. Because the exact parent magma is not known, the F values calculated by Leeman and others (1976) are only relative, and all are normalized to F = 1.0 for sample 69-22 (table 5). F values can be estimated independently from data for incompatible elements (that is, those with small solid/liquid bulk distribution coefficients) using the relationship

$$C_L/C_0 = F^{(\bar{D}-1)}, \qquad (3)$$

Table 5. Crystallization indices for COM-type lavas

Sample	F	$\frac{(K_2O)\ \text{ref}}{(K_2O)\ \text{rock}}$	$\frac{(\text{Rb})\ \text{ref}}{(\text{Rb})\ \text{rock}}$	Average "F"
Craters of the Moon lava field (reference lava = 69-22)				
69-28	1.3	1.31	1.25	1.29
69-22	1.0	1.00	1.00	1.00
70-42	1.0	1.03	0.94	0.99
V-28	0.84	0.86	0.79	0.83
WPL-87	0.77	0.78	0.73	0.76
69-33	0.68	0.74	0.68	0.70
SBR-5	0.65	0.68	0.63	0.65
69-17	0.56	0.59	0.55	0.57
69-35	0.47	0.47	0.42	0.45
69-19	0.42	0.44	0.38	0.41
69-20	0.34	0.35	0.32	0.34
V-31	0.33	0.34	0.31	0.33
V-27	0.32	0.33	0.29	0.31
King Hill lavas (reference lava = OB-32)				
OB-32	--	1.00	1.00	1.00
69-61	--	0.80	1.03	0.92
69-62	--	0.84	0.91	0.88
70-9	--	1.02	0.94	0.98
69-60	--	0.69	0.66	0.68
69-63	--	0.44	0.39	0.42

Table 5. Crystallization indices for COM-type lavas (cont'd)

Sample	F	$\frac{(K_2O)\ \mathrm{ref}}{(K_2O)\ \mathrm{rock}}$	$\frac{(Rb)\ \mathrm{ref}}{(Rb)\ \mathrm{rock}}$	Average "F"
High Point lavas, Island Park west rift (reference lava = 73-71)				
73-71	--	1.00	1.00	1.00
73-59	--	0.60	0.57	0.59
73-42	--	0.55	0.52	0.54

F (=fraction of original liquid remaining) values in column one are taken from Leeman (1976) and are derived from least-squares fractionation models based on major element compositons of lavas and their phenocrysts. Columns two and three give estimated values of F based on contents of K_2O and rubidium as explained in the text. Column three gives an average of the first three columns; these values are used in all computations.

where C_L is the concentration of an element in the residual liquid, C_0 is the concentration in the parent magma, and $\bar{D}$ is the bulk solid/liquid distribution coefficient. Because $\bar{D}$ is small (<<1) for K_2O and Rb, these elements are used to estimate F using the rearranged expression

$$F = C_0/C_L. \tag{4}$$

Agreement between F values reported by Leeman and others (1976) and those computed using rubidium and K_2O contents in COM lavas is good (table 5). Consequently, the latter approach was used to estimate F values for the King Hill and High Point lavas, and averages of the various estimates (table 5) are used in all subsequent calculations. It is again emphasized that F values for any given series of COM-type lavas are estimated relative to the most mafic sample in that series and cannot be directly compared with F values for a different series.

Equation (2) predicts a linear relationship between $\delta^{18}O$ and lnF for cogenetic lavas related by crystal fractionation. The data for COM lavas produce two distinct linear trends (figure 6) -- one for basalts and another for latites. Leeman and others (1976) showed that compositional variations in these lavas are consistent with removal in constant proportions (in parentheses) of plagioclase (0.462), olivine (0.193), magnetite (0.255), and apatite (0.090) from the basaltic lavas and removal of plagioclase (0.40), olivine (0.057), clinopyroxene (0.40), magnetite (0.114), and apatite (0.029) from the latitic lavas. Thus, the two trends in figure 6 correspond to removal of different mineral assemblages in the early and late stages of crystallization of the COM lavas. Least-squares regression of lnF vs. $\delta^{18}O$ for the basalts and for the latites have the slopes ($m = 10^3(\alpha - 1)$ and intercepts ($b = \delta^{18}O$) given in table 6, the regression lines are shown in figure 6. The King Hill and High Point lavas also display linear trends between $\delta^{18}O$ and lnF (figure 7), but the slopes (table 6) are different for each of these two series. The slope for the King Hill lavas resembles that for the COM basalts, whereas the slope for the High Point latites resembles that for COM latites. Thus, differences between these two series could reflect removal of different mineral assemblages, as in the respective COM magmas.

Bulk solid/liquid fractionation factors calculated from the slopes of the regression lines ($\alpha = 1 + 10^{-3}m$) are given in table 6 for the COM-type lavas. The α values for basaltic lavas from COM and King Hill are in close agreement, as are those for COM and High Point laties. Values of $\delta^{18}O$ estimated for these lava series are in the range 5.9 to 6.4. These values are similar to or slightly higher than $\delta^{18}O$ values in SRP tholeiites, but because the exact parent magmas are unknown the $\delta^{18}O$ estimtes are considered as upper limits.

The magnitude of α can be estimated independently for COM lavas using the weight proportions (x_i) of minerals i in the bulk solid removed from the basaltic and latitic magmas, respectively, and individual mineral/liquid α_i values, according to the equation

$$\bar{\alpha} = \Sigma x_i \cdot \alpha_i \tag{5}$$

Table 6. Linear regression parameters for oxygen isotopic ratios versus ln(F) and inferred bulk fractionation factors for oxygen isotopes

	m	b	r	α
Craters of the Moon lava field				
basalts (anal. 20-27)	-1.115	5.86	0.95	0.9989
latites and evolved basalt (anal. 26-32)	-0.264	6.36	0.71	0.9997
King Hill lavas				
all samples (anal. 33-38)	-1.165	6.36	0.95	0.9988
High Point lavas				
all samples (anal. 50-52)	-0.678	5.96	0.88	0.9993

α values are derived from the equation $\delta^{18}O = m\ln(F))+b$, where $m = 10^3 (\alpha-1)$ and $b = \delta^{18}O_0$ (the initial isotopic ratio of the magma); r = correlation coefficient.

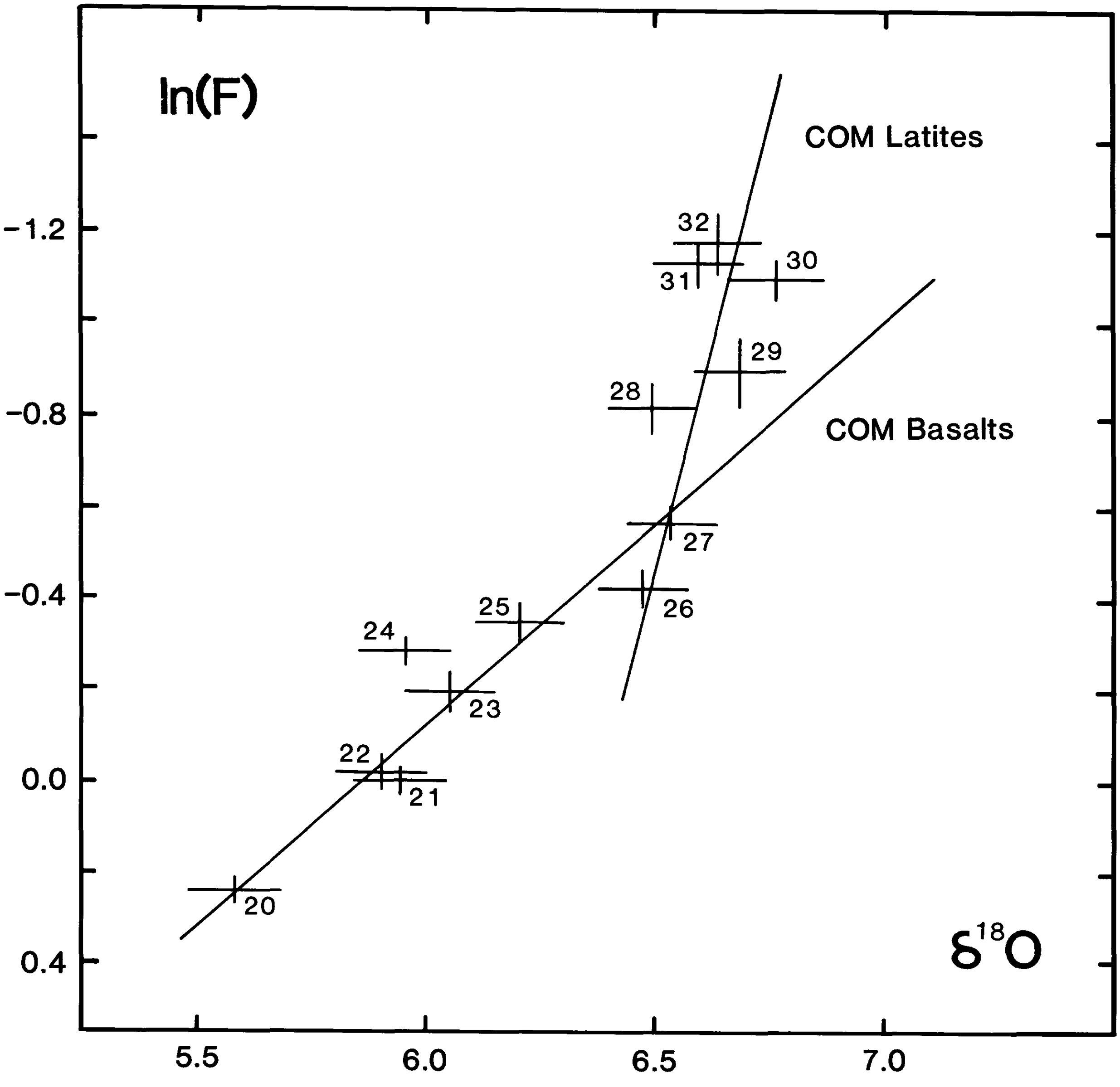

Figure 6. ln(F) versus $\delta^{18}O$ plot for COM lavas. Numbers by data points refer to analysis numbers in table 2. Lines through the basalt and latite data are fitted by least-squares linear regression.

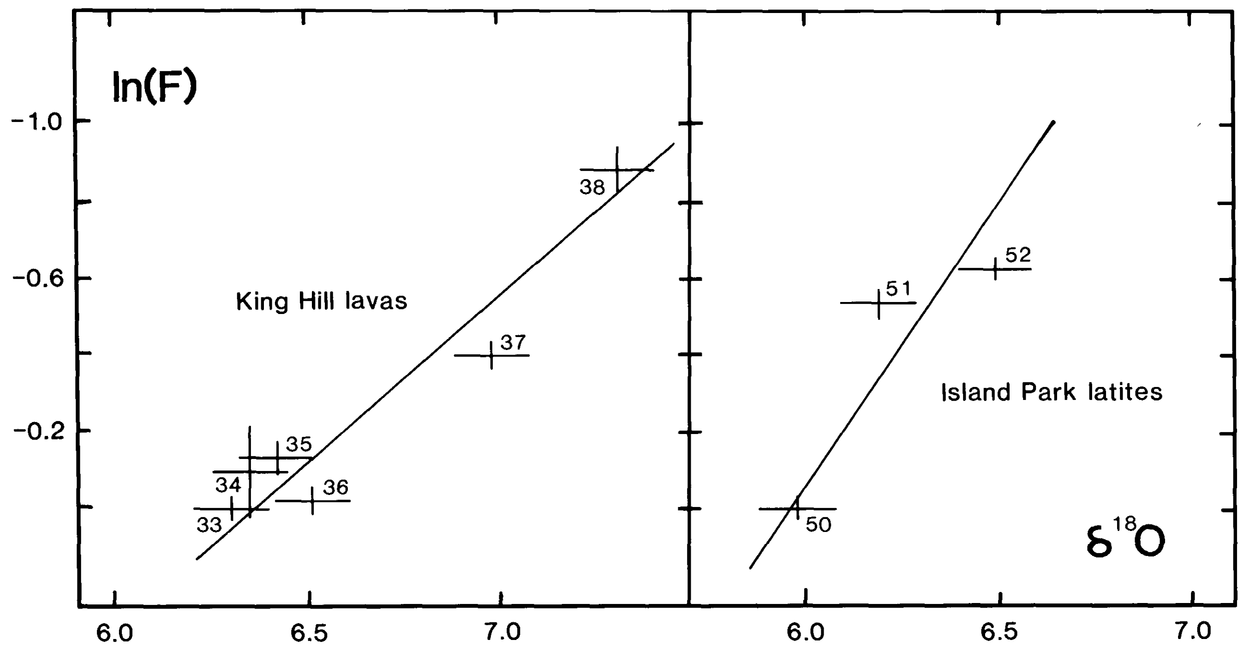

Figure 7. ln(F) versus $\delta^{18}O$ plot for King Hill and Island Park (High Point) lavas. Analysis numbers from table 2 and lines are fit by least-squares linear regression.

However, experimental determinations of α_i values are unavailable for the minerals involved and it is necessary to estimate α_i's from analyses of coexisting groundmass and phenocryst phases in volcanic rocks using the relation

$$\Delta^{\text{mineral}}_{\text{gm}} = \delta^{18}_{i}O - \delta^{18}_{\text{gm}} \simeq 10^{3} \cdot \ln\alpha_{i} \qquad (6)$$

where the subscipts i and gm refer to mineral i and groundmass (that is, liquid).

Excluding a few extreme values, Δ(mineral,gm) values calculated from the data of Anderson and others (1971) for predominantly basaltic lavas show the following ranges: plagioclase (+0.4 to -0.2), olivine (-1.4 to -1.3), pyroxene (-0.2 to -0.8), and magnetite (-2.1 to -2.5). These ranges may reflect variations in temperature of crystallization, kinetic or disequilibrium effects, and(or) analytical uncertainties. The basalts analyzed by Anderson and others were quenched at temperatures around 1,200° to 1,050°C, thus the Δ values should approximate the fractionation factors relevant to crystallization of COM basalts (about 1,100°C) and latites (ca. 1,000°C) (Leeman, 1974, in preparation). The average value for Δ (plagioclase, gm) is 0.0 ± 0.3, and it is assumed that the corresponding value of $\alpha^{\text{pl}}_{\text{gm}}$ ($\approx\alpha^{\text{pl}}_{\text{Liq}}$) of 1.000 ± 0.0003 is representative throughout the COM crystallization sequence. Using this value and plagioclase/mineral fractionation factors calculated from the equations of Bottinga and Javoy (1975), we estimated "mineral/liquid" α's for all the liquidus phases in COM lavas except apatite (table 7; see footnote). It is assumed that Δ(ap, Liq) is similar to Δ(ol, Liq) based on experimental studies of isotopic fractionation between these minerals and water (J. R. O'Neil, personal communication, 1975). In any case, the uncertainty in this value is not significant because only small amounts of apatite are involved in the fractionation models. Using equation (5), α_i values in table 7, and the weight fractions of solids removed, values of α were computed for COM basalts (0.9991 ± 0.0003 at 1,100°C) and latites (0.9993 ± 0.0003 at 1,000°C).

Plagioclase-, olivine-, and pyroxene-liquid α_i's from Kyser and others (1981) are somewhat different from those in Bottinga and Javoy (1975) (at 1000°C $\alpha^{\text{pl}}_{\text{gm}}$ is 1.0003, $\alpha^{\text{ol}}_{\text{gm}}$ is 0.9982, and $\alpha^{\text{px}}_{\text{gm}}$ is 0.9990); however, bulk α's computed with these α_i's are only slightly different because the increase in $\alpha^{\text{pl}}_{\text{gm}}$ is offset by decreases in $\alpha^{\text{ol}}_{\text{gm}}$ and $\alpha^{\text{px}}_{\text{gm}}$. Computed α's for COM basalts at 1,100°C and COM latites at 1,000°C are both 0.9993 ± 0.0003.

These values are in close agreement with those determined from graphs of $\delta^{18}O$ vs. lnF, suggesting that within the uncertainties involved, variations in the $\delta^{18}O$ values of COM lavas are compatible with the crystal fractionation model proposed for these rocks.

Table 7. Estimated mineral/melt fractionation factors for oxygen-isotope ratios

$T(^{o}C)$	α^{pl}_{liq}	α^{mt}_{liq}	α^{ol}_{liq}	α^{px}_{liq}
1100	1.0000	0.99789	0.99877	0.99943
1000	1.0000	0.99755	0.99857	0.99934
900	1.0000	0.99711	0.99832	0.99922

Method of derivation

The average Δ(pl,gm) ~Δ(pl,liq) value for 10 lavas in Anderson and others (1971) and this paper is 0.0±0.3, corresponding to a temperature range of about 1,200-900°C. As $\Delta(A,B)$ ~1,000 ln α^{A}_{B}, it is assumed that α^{pl}_{gm} = 1.0000±0.0003 for conditions appropriate to crystallization of SRP evolved lavas. Fractionation factors for the other minerals relative to liquid were obtained from Δ(pl,gm) and values of Δ(pl,mt), Δ(pl,ol), and Δ(pl,px) calculated at appropriate temperatures from the equations of Bottinga and Javoy (1975)., Errors in the calculated α values are at least ±0.0003 based on the one-sigma range in Δ(pl,gm) noted above. Corrections for plagioclase compositional variations (An_{60} to An_{30} in COM lavas) are an order of magnitude smaller and are ignored accordingly. Finally α^{ap}_{gm} is assumed to be similar to a^{ol}_{gm} (J.R. O'Neil, personal communication, 1975), but in any case uncertainties in this value will have little effect on the results of our calculations as the proportion of apatite is always small relative to that of plagioclase, olivine, magnetite and pyroxene.

Oxygen isotope fractionation factors (α's) between bulk solids and silicate liquids have been previously estimated by Anderson and others (1971) and Matsuhisa and others (1973). Their values are compared with ours for COM-type lavas in table 8. As noted earlier, these α values are dependent upon the nature and proportions of phases removed. Thus, the comparatively large α's calculated for COM-type lavas and for Japanese calkalkaline lavas may reflect removal of relatively large proportions of ^{18}O-depleted minerals (magnetite, olivine, and so forth) from these magma types. In any case, the data for these suites indicate that significant ^{18}O-enrichment (as much as 1 to 2 per mil) can be achieved via fractional crystallization under favorable conditions. We also note that α values need not remain constant for a given magma series, but will vary according to changes in the proportions and nature of removed phases; this point is well illustrated for the COM-type basaltic and latitic lavas (Matsuhisa, 1979).

Origins of rhyolites

The isotopic data for SRP rhyolites, though limited, seemingly preclude their direct derivation from a crustal source with $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ similar to observed values in the analyzed crustal xenoliths (figs. 4 and 5). It may be argued that the xenoliths are not wholly representative of all possible crustal sources, nor can a sub-crustal source be discounted. However, the distinction in $^{87}Sr/^{86}Sr$ between SRP rhyolites and tholeiites apparently does rule out a common source material for these magma types. For the same reason, it is improbable that the rhyolites represent differentiates or immiscible liquids derived from precursor olivine tholeiite magmas. Specific models for the origin of the rhyolites must await further work and are best deferred until these rocks have been more fully characterized.

CONCLUSIONS

Combined oxygen and strontium isotopic data for SRP olivine tholeiites indicate that these basalts are not contaminated by crustal materials. The $\delta^{18}O$ values for these rocks are typical of many ocean island and mid-ocean ridge basalts. There is no evidence that the mantle source for SRP tholeiites is unusual with regard to oxygen isotopic composition, nor have parental magmas interacted with low-^{18}O meteroic waters. $^{87}Sr/^{86}Sr$ ratios in the SRP tholeiites range slightly (0.7056 to 0.7076) and are higher than such ratios in most oceanic basalts. We consider that these ratios are representative of mantle source regions that are slightly heterogeneous and enriched in $^{87}Sr/^{86}Sr$ compared with typical sub-oceanic mantle material.

Evolved basaltic to latitic COM-type lavas from several localities display variations in $\delta^{18}O$ that are largely consistent with crystal fractionation effects. Nevertheless, many COM-type lava suites display a strong correlation between $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ that we interpret as reflecting selective contamination by ^{87}Sr derived from the crust while the respective magmas were differentiating in shallow chambers. Small amounts of contamination may not be readily apparent in the oxygen isotope data if the difference in $\delta^{18}O$ between the COM-type lavas and crustal rocks is small, as is the case for some crustal xenoliths and their host COM-type lavas. However, some COM-type

Table 8. Estimated Oxygen-Isotope Fractionation Factors

	$\bar{\alpha}$	Δ^b	$\delta^{18}O$-enrichment in residual liquid[a]	Ref
COM-type basalt	0.9989	-1.10	0.76	1
COM-type latite	0.9997	-0.30	0.21	1
Ocean Island basalt	0.99985(min)	-0.15	0.10	2
	0.99975(max)	-0.25	0.17	2
Japanese calcalkalic lavas	0.9997(min)	-0.30	0.21	3
	0.9990(max)	-1.00	0.70	3

[a] enrichment is calculated for 50 percent crystallization (F=0.5) according to the equation: $\delta^{18}O - \delta^{18}O = 10^3 (\alpha - 1) \cdot \ln F$.

[b] $\Delta = 10^3 \cdot \ln\alpha$

References: (1) this report, (2) Anderson and others (1971), (3) Matsuhisa and others (1973), and Matsuhisa (1979).

suites display no correlation between $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ indicating that crustal contamination is not necessarily essential to the petrogenesis of these lavas.

Oxygen isotope fractionation factors (α's) between bulk solid and liquid are sufficiently large to result in significant enrichments in $\delta^{18}O$ as basaltic magmas crystallize (up to 1 permil for 60 percent crystallization of COM-type basalt). For any given magma series α-values need not remain constant, and will vary if the nature and(or) proportions of removed phases changes. Such α values are dependent upon magma composition, pressure, temperature, and other factors which control either the individual mineral/liquid fractionation factors or the phase relationships.

Rhyolites from the SRP have similar $\delta^{18}O$ but much higher $^{87}Sr/^{86}Sr$ values than the associated olivine tholeiites. Crustal anatexis may be a viable mechanism for producing the rhyolite magmas, but the analyzed crustal xenoliths are not likely source materials. A sub-crustal source cannot be discounted, but in this case different source materials must be postulated for the rhyolites and olivine tholeiites. The isotopic data seemingly preclude any cogenetic relation between the tholeiites and rhyolites.

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