

MAJOR ELEMENT, VOLATILE, AND STABLE ISOTOPE GEOCHEMISTRY
OF HAWAIIAN SUBMARINE THOLEIITIC GLASSES

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Abstract. Tholeiitic glasses were dredged from the submarine rift zones of the five volcanoes comprising the island of Hawaii and Loihi Seamount. The major element composition of the glasses follows a systematic trend that is related to the stage of evolution of the volcano. Glasses from Loihi Seamount (the youngest Hawaiian volcano) are enriched in Fe, Ca, Ti, Na, and K and depleted in Si and Al relative to the glasses from the other, older volcanoes. Kilauea is intermediate in age and its glasses are intermediate in composition between those from Loihi and Mauna Loa, the largest and oldest of the active Hawaiian tholeiitic volcanoes. The volatile contents (H₂O, CO₂, S, F, Cl) of the glasses from these volcanoes follow the same trend (highest in Loihi; lowest in Mauna Loa). Glasses from Hualalai Volcano are similar in composition to those from Mauna Loa; those from Kohala Volcano are similar to Kilauea; Mauna Kea glasses range from Mauna Loa-like to Kilauea-like. The observed systematic variation in composition of Hawaiian tholeiites may be related to the progressive melting and depletion of the source of these volcanoes during their growth. Oxygen and hydrogen isotope analyses were made on many of the glasses from each volcano. The $\delta^{18}\text{O}$ values of Hawaiian tholeiites are distinctly lower than those of mid-ocean ridge basalt (MORB) (averages: 5.1 versus 5.7). These low values are probably a distinct feature of hot spot lavas. The δD values for these glasses (-88 to -61) are typical of mantle and MORB values. Thus the H₂O in the Hawaiian glasses is probably of magmatic origin. Previous isotopic and trace element data indicate that the source of Hawaiian tholeiites contains two distinct source components. Based on the results of this study, the plume component in the source for Hawaiian tholeiites is characterized by moderate $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7035-0.7037) and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios (18.6-18.7), a low $\delta^{18}\text{O}$ value (~5.0), and greater contents of volatiles, Fe, Ca, Ti, Na and K relative to the MORB source.

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Introduction

Trace element, rare gas, and isotopic data indicate that Hawaiian lavas (including shield and posterosional lavas) and other oceanic island basalts (OIB) evolved by mixing melts from several geochemically distinct sources. However, the number and compositions of the mixing end-members and the relationship of these components to specific mantle sources remain points of controversy (see Zindler and Hart [1986] for a review). The source for Hawaiian tholeiites apparently consists of two principal components based on Pb, Sr and Nd isotope data [Tatsumoto, 1978; Tatsumoto et al., 1987; West et al., 1987]. One of these components is thought to be from a plume but there is little agreement on the nature of the nonplume component (for example, 100 Ma oceanic lithosphere [Stille et al., 1986]; recycled oceanic lithosphere [Hofmann et al., 1986]). Isotopic end-members can be identified among Hawaiian tholeiites. Kilauea is considered one end-member and Koolau has been considered the other with tholeiites from other volcanoes in between. However, new data for Kahoolawe and Lanai tholeiites show that Koolau is no longer unique (as suggested by Tatsumoto et al. [1987]) and that the range of Hawaiian tholeiites must be extended [West et al., 1987].

In order to complement previous studies and more fully evaluate the nature of compositional variation among Hawaiian tholeiites, we undertook a comprehensive geochemical study of submarine glasses from each of the five volcanoes comprising the island of Hawaii (Mauna Kea, Mauna Loa, Kilauea, Kohala and Hualalai) and Loihi Seamount (Figure 1). This is the first such study where the geochemistry of a single suite of Hawaiian lavas is thoroughly documented. Fresh submarine glasses were used to avoid the effects of alteration and crystal accumulation. Such glasses may represent virtually unmodified magmatic liquids. Furthermore, submarine erupted glasses may contain preeruption concentrations of volcanic gases [Moore, 1970; Moore and Schilling, 1973] which provide another important parameter in the evaluation of the source of Hawaiian lavas. Finally, since all the glasses analyzed from each of the volcanoes are tholeiitic, we were able to study lavas that are broadly representative of the shield-building stage of development for each of the volcanoes. This would not be possible if only subaerial lavas

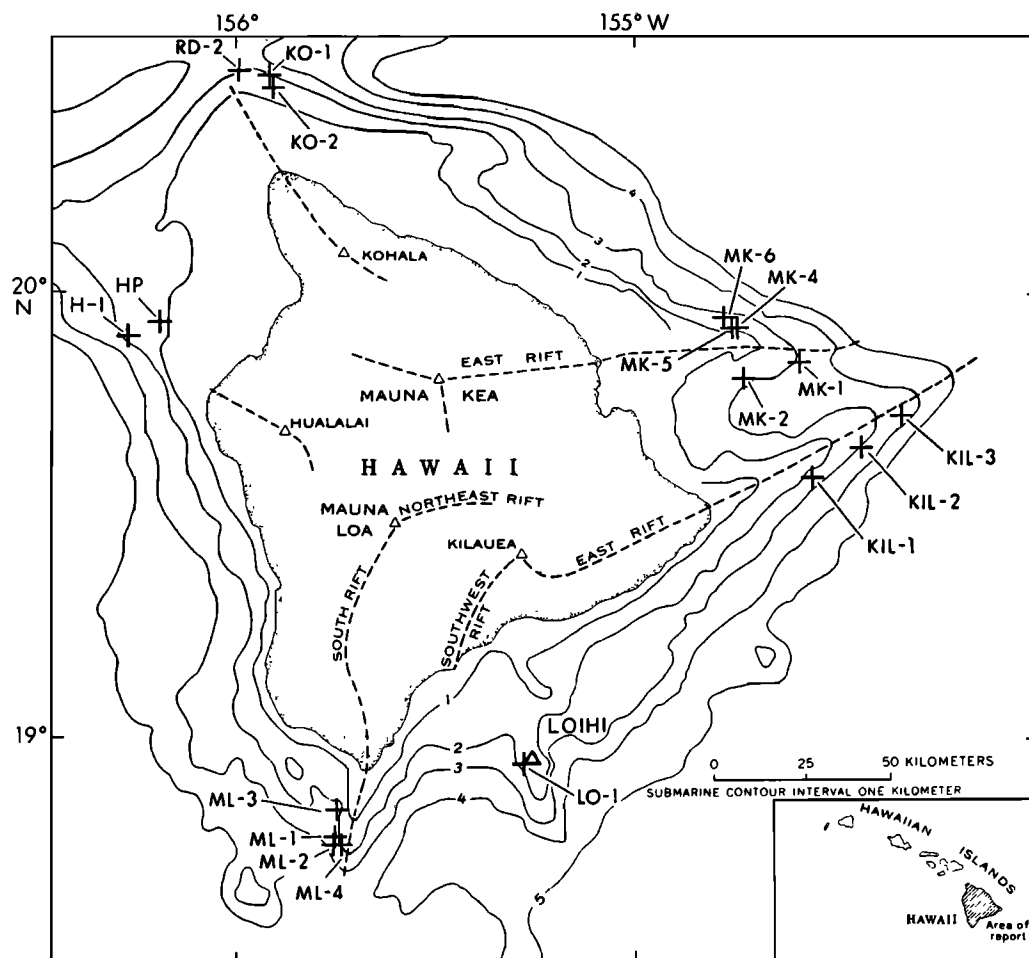


Fig. 1. Map of the island of Hawaii showing the five volcanoes that form the island and Loihi Seamount (the summit of each volcano is shown by a triangle). The rift zones of the volcanoes are shown by dashed lines. Dredge haul locations are shown by the plus symbol for each volcano. The bathymetric contour interval is 1 km.

were used because alkalic lavas mantle the tholeiitic shields on most Hawaiian volcanoes. Those tholeiites that are exposed are commonly altered (especially the contents of alkali metals) due to groundwater percolation [Chen and Frey, 1985].

In this paper, we use major elements, volatile abundances (H_2O , CO_2 , Cl, F, S), and hydrogen and oxygen isotope data of glasses to further constrain the sources for Hawaiian tholeiitic lavas. Separate studies are in progress on the trace element and radiogenic isotopes [Garcia et al., 1986a, Gurriet et al., 1988] and rare gases [Lupton and Garcia, 1986] in these lavas.

Samples

Most of the samples used in this study were obtained from dredges we made during a 1983 R/V KANA KEOKI cruise around the island of Hawaii. Rocks with abundant fresh glass were obtained from the deep submarine extensions of the major rift zones of four of the five volcanoes comprising the island of Hawaii (except Kohala)

and Loihi Seamount. We dredged at various locations (Figure 1) along these rift zones at water depths ranging from 1900 to 3700 m in order to sample as diverse an assortment of lava compositions and ages as possible. The depth of recovery of the samples is greater than the depth they were erupted at because the island of Hawaii is sinking rapidly (~ 2.5 mm/y [Moore, 1987]). Kohala has sunk approximately 1 km; Mauna Kea about 400-500 m; and Kilauea ~ 100 m [Moore, 1987]. Our dredging concentrated on Mauna Loa and Mauna Kea since little fresh, unaltered pillow rim glass was available from these volcanoes. We also obtained glassy samples from Kohala (RD-2) and Hualalai (HP) that were taken during engineering surveys of these volcanoes.

Pillow basalts with well-developed glass rinds (>0.5 cm) were the focus for our study. For each dredge haul, we determined petrographic groups and selected representative samples for microprobe analyses; 41 distinct groups have been defined. A major element analysis for a representative sample from each group is given in Table 1.

TABLE 1. Microprobe Analyses of Glasses from Mauna Loa, Mauna Kea, Kilauea, Hualalai, Kohala and Loihi Volcanoes, Hawaii

Sample Number	Major Elements										Total	Modes				Total
	SiO ₂	TiO ₂	Al ₂ O ₃	FeO*	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	Volatiles		Ol	Cpx	Plag		
Loihi																
LO-1-4	48.69	2.82	13.44	11.82	6.81	11.68	2.54	0.54	0.32	0.87	99.53	4.6	0.1	0.0	4.7	
LO-1-6	49.24	2.67	13.61	12.01	6.70	11.79	2.57	0.44	0.32	0.87	100.22	3.5	0.2	0.4	4.1	
LO-1-9	49.27	2.79	13.96	12.23	6.28	11.41	2.61	0.45	0.31	0.92	100.23	3.0	0.2	0.1	3.3	
Kilauea																
KIL-1-BR	51.50	2.57	13.71	10.67	6.40	11.20	2.30	0.43	0.26	0.73	99.77	12.2	0.0	0.0	12.2	
KIL-1-1	51.02	2.56	13.64	10.62	6.47	11.26	2.29	0.44	0.28	0.74	99.32	10.4	0.0	0.0	10.4	
KIL-1-4	51.23	2.58	13.89	10.58	6.41	11.13	2.24	0.44	0.27	0.77	99.54	3.6	0.1	0.0	3.7	
KIL-1-5	50.78	2.86	13.95	11.02	6.20	10.89	2.36	0.49	0.31	0.87	99.73	1.4	0.4	1.4	3.2	
KIL-1-9	51.07	2.42	13.57	10.52	7.23	11.27	2.20	0.42	0.24	0.80	99.74	15.0	0.0	0.0	15.0	
KIL-2-1	51.72	2.66	14.10	10.11	6.37	11.07	2.28	0.46	0.30	0.64	99.71	29.6	0.1	0.0	29.7	
KIL-2-8	51.76	2.73	14.02	10.17	6.34	10.98	2.28	0.47	0.29	0.78	99.83	28.6	0.1	0.0	28.7	
KIL-2-9	51.38	2.64	13.73	10.11	6.64	11.15	2.41	0.47	0.26	0.73	99.52	28.6	0.0	0.0	28.6	
KIL-3-9	51.71	3.25	13.94	10.64	5.77	9.95	2.56	0.60	0.34	0.83	99.59	27.8	0.8	0.0	28.6	
KIL-3-19	51.75	3.33	13.98	11.30	5.85	10.02	2.49	0.55	0.36	0.91	100.54	16.0	0.1	0.2	16.3	
Mauna Loa																
ML-1-1	52.03	2.16	13.89	9.69	7.05	10.99	2.28	0.33	0.24	0.65	99.31	29.4	0.4	0.0	29.8	
ML-1-3	52.45	2.27	13.95	10.52	6.82	10.82	2.30	0.36	0.28	0.32	100.09	13.4	0.6	1.8	15.8	
ML-1-10	52.38	2.83	13.58	11.05	6.01	9.86	2.49	0.54	0.33	0.52	99.58	11.4	2.2	1.0	14.6	
ML-1-11	52.46	2.20	14.21	9.91	6.39	11.50	2.28	0.34	0.25	0.49	100.04	12.8	0.0	0.0	12.8	
ML-1-12	52.43	2.15	14.09	10.00	6.70	11.12	2.32	0.35	0.23	0.61	100.00	20.6	0.2	0.1	20.9	
ML-2-1	52.24	2.62	13.84	11.04	6.32	10.76	2.46	0.49	0.30	0.44	100.19	12.4	0.8	1.4	14.6	
ML-2-3	52.65	2.26	14.10	10.06	6.87	10.88	2.37	0.36	0.27	0.36	100.18	14.2	0.8	2.2	17.2	
ML-2-8	52.04	2.01	13.94	9.91	7.65	10.88	2.23	0.32	0.23	0.42	99.63	23.4	0.0	0.0	23.4	
ML-3-30	52.07	2.32	14.53	10.37	6.30	10.59	2.32	0.37	0.26	0.72	99.57	24.0	0.6	0.0	24.6	
ML-4-2	51.90	2.65	13.58	11.41	5.98	10.25	2.36	0.49	0.34	0.40	99.35	11.8	1.0	1.6	14.4	
ML-4-10	52.04	2.01	13.89	10.35	7.63	10.82	2.26	0.33	0.26	0.41	100.00	22.4	0.0	0.0	22.4	
ML-4-11	51.95	2.67	13.80	11.38	6.05	10.40	2.44	0.48	0.31	0.40	99.88	6.4	1.0	1.2	8.6	
Mauna Kea																
MK-1-2	52.58	2.83	13.50	10.41	6.58	10.29	2.49	0.58	0.36	0.73	100.36	13.8	0.0	0.0	13.8	
MK-1-8	52.49	3.18	13.66	10.87	5.26	9.22	2.87	0.75	0.41	0.92	99.63	16.4	0.6	0.8	17.8	
MK-1-10	52.58	2.79	13.72	10.28	6.70	10.41	2.45	0.51	0.35	0.62	100.41	16.8	0.0	0.0	16.8	
MK-2-1	51.00	2.95	13.61	11.92	6.31	10.64	2.36	0.45	0.32	0.61	100.17	14.0	0.0	0.2	14.2	
MK-4-1	51.39	2.41	13.97	10.50	6.62	10.86	2.33	0.35	0.31	0.41	99.15	14.8	0.0	0.1	14.9	
MK-5-2	51.71	3.25	13.25	12.31	5.54	9.87	2.52	0.60	0.37	0.64	100.08	5.6	0.0	0.1	5.7	
MK-5-5	51.58	2.50	13.88	10.81	6.64	11.00	2.37	0.40	0.30	0.37	99.92	14.0	0.0	0.0	14.0	
MK-5-25	51.84	2.49	13.51	11.07	6.53	11.10	2.42	0.37	0.29	0.35	99.97	15.4	0.0	0.0	15.4	
MK-6-6	51.98	2.69	13.69	10.79	6.49	10.90	2.37	0.42	0.33	0.36	100.02	22.0	0.2	0.4	22.6	
MK-6-18	51.84	2.44	14.13	10.87	6.68	10.70	2.28	0.38	0.29	0.40	100.01	5.2	0.0	0.0	5.2	
Hualalai																
H-1-5	52.18	2.12	14.14	10.49	6.71	11.06	2.33	0.34	0.25	0.28	99.89	16.2	0.0	0.0	16.2	
H-1-7	51.92	2.14	13.95	10.67	7.27	10.98	2.27	0.31	0.26	0.39	100.14	11.0	0.0	0.0	11.0	
H-P	52.03	2.30	14.26	10.10	6.70	10.91	2.40	0.37	0.26	0.39	99.72	34.4	0.0	0.1	34.5	
Kohala																
KO-Camp	51.39	2.54	13.76	11.26	6.36	10.88	2.37	0.48	0.28	0.07*	99.39	4.8	0.0	0.0	4.8	
KO-2	50.45	3.40	13.53	12.86	5.90	10.31	2.71	0.55	0.43	ND	100.14	28.0	0.0	0.0	28.0	
KO1-17	51.60	2.18	13.99	11.35	6.00	11.40	2.60	0.45	0.33	ND	99.90	5.0	0.0	0.0	5.0	

ND denotes not determined. Ol denotes olivine; cpx, clinopyroxene; plag, plagioclase.

*Degassed value.

Analytical Methods

The samples analyzed in this study were brown, translucent, glass fragments (0.5-1.0 mm), which were hand-picked using a binocular microscope. Extreme care was taken to reject glass fragments with phenocrysts, vesicles,

incipient crystallization or visible signs of alteration. The grains were ultrasonically cleaned in 0.3 N HCl and distilled water and dried in an oven at 120°C for 24 hours [Byers et al., 1983]. Microprobe thin sections were made across the pillow rinds into the interior of each sample to characterize their petrography.

Five-hundred-count modes were made on each sample. Only phenocryst (>0.5 mm) data are reported in Table 1.

Major element analyses were made on multiple glass fragments from each glassy rim using an electron microprobe (see Garcia et al. [1986b] for a summary of methods used). Each reported analysis (Table 1) is an average of 10 analyses. The Smithsonian Institution glass standard VG-A99 [Jarosewich et al., 1979] was used as an internal control.

Volatile contents of the glasses were determined using high-temperature mass spectrometry (see Byers et al. [1986] for methods). Typically, a 40-50 mg sample is placed in an alumina Knudsen cell assembly and heated at 5°C/min to 1250°C under a vacuum of 10^{-7} to 10^{-8} torr. Degassing of the samples is continuously monitored by a computer-controlled quadrupole mass spectrometer which rapidly scans the spectrum from 2 to 100 amu. Mass peaks and their corresponding ion-current signals and cell temperatures are measured and stored on magnetic tape for data reduction and analysis. Minimum signals corresponding to concentrations of 1 ppm for Cl, F and S and 5 ppm for H₂O and carbon-containing volatiles can be discerned from background contributions. Mass pyrograms (computer plots of ion intensity versus temperature) are generated from the reduced data. These provide quantitative information on volatile content of each sample since the area under each curve is proportional to the total amount of that volatile released within any given temperature interval. After degassing, the samples are reweighed for total weight loss.

Oxygen isotope analyses were made using a ClF₃ reagent [Borthwick and Harmon, 1982]. After degassing at 150°C, water was released from the glasses at 1200°C and reacted with uranium metal to produce hydrogen for isotopic analysis [Bigeleison et al., 1952]. Replicate analyses were made of each sample and the data are precise to ± 0.1 and ± 1.0 per mil for $\delta^{18}\text{O}$ and δD , respectively.

Results and Discussion

Petrography

All of the analyzed samples have well-defined glassy rims (brown glass in thin section, 0.5-10 mm in thickness). The glassy rims are weakly vesicular (<5%, usually less than 2%). Olivine is the dominant phenocryst and is present in all of the samples. The range in olivine phenocryst content is 1.2 to 34.4 vol % (Table 1). Thirty-one of the 41 samples analyzed have more than 10 vol % olivine phenocrysts. Plagioclase and augite phenocrysts are rare (<1 vol %) in these lavas (except some of those from Mauna Loa with 1-2 vol %) but they are common matrix phases in pillow interiors. Cr-spinel is present in many olivine phenocrysts. Orthopyroxene phenocrysts occur only in sample ML-1-7.

The abundance of picritic lavas (>15 vol % olivine phenocrysts) in the submarine rift collections (40-60% for the volcanoes with >10 samples) is in sharp contrast to the subaerial sections of these volcanoes (3-6% for Kilauea and 15-20% for Mauna Loa [Macdonald, 1949]).

The higher percentage of picritic lavas on the submarine rifts of these Hawaiian volcanoes may be related to their higher density. Ryan [1988] suggested that magmas rise to levels of neutral buoyancy. Picritic magma would tend to remain in the deeper portions of the volcanoes and to be intruded at greater depths (8-10 km within the rift zones [Ryan, 1988]). The abundance of these high density lavas (2.65-2.70 g/cm³ for 16-22.5 wt % MgO versus 2.59 g/cm³ for 7.1 wt % MgO; calculated using the method of Bottinga and Weill [1970] with observed H₂O contents included) on the lower portions of the rift zones may be the result of preferential intrusion of denser, more MgO-rich magmas into the deeper portions of the rift zone that may feed the submarine portions of the rift zones [see Holcomb, 1987].

Major Elements

Microprobe analyses were made of glasses from the outer pillow rim of all the glassy lavas recovered. They are all tholeiitic in composition (Figure 2) and quartz-normative (except for the Loihi glasses which have high contents of normative-hypersthene, 11-16%). This is in contrast to the wide variety of rock types recovered from Loihi (tholeiite to basanitoid [Moore et al., 1982]). Thus if alkalic lavas are erupted during the early stages of all Hawaiian volcanoes [Clague, 1987a], they are probably completely covered by subsequent tholeiitic lavas. Furthermore, as noted by Clague [1987b] for Hualalai, alkalic lavas were not recovered from the volcanoes that have alkalic caps (i.e., Mauna Kea, Hualalai and Kohala). Thus the vents for late-stage alkalic volcanism must be concentrated nearer the summit of these volcanoes. This indicates that the rift zones which are so effective in trans-

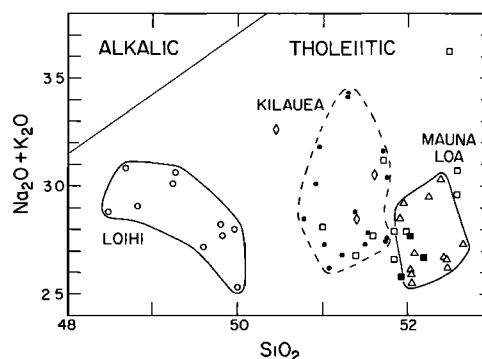


Fig. 2. Alkalies (Na₂O + K₂O wt %) versus SiO₂ (wt %) in Hawaiian tholeiitic glasses. Analyses are in Table 1 with additional tholeiitic glass analyses from Byers et al. [1985] for Kilauea and Loihi. The boundary separating alkalic from tholeiitic compositions is from Macdonald and Katsura [1964]. Symbols for volcanoes: Loihi, open circle; Kilauea, solid circle; Mauna Loa, triangle; Mauna Kea, open square; Hualalai, solid square; Kohala, diamond.

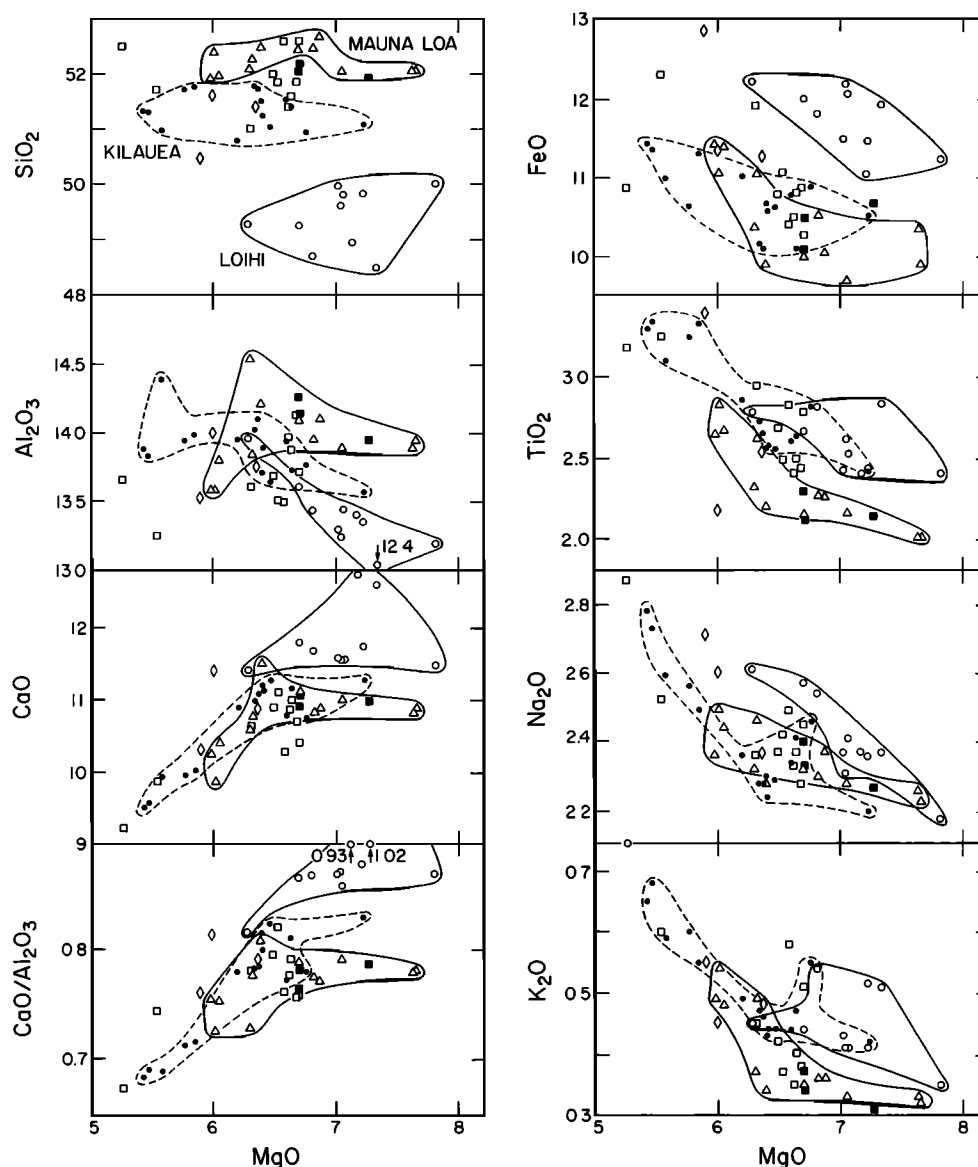


Fig. 3. MgO variation diagrams for Hawaiian tholeiitic glasses. Data and symbols as in Figure 2. Fields are drawn for Loihi, Kilauea and Mauna Loa glasses.

porting tholeiitic magmas tens of kilometers from the summit reservoir must become plugged during the waning stages of growth of a Hawaiian volcano. This is probably due to the decreased eruption frequency and volume of lava erupted during this stage (for example, one eruption/1900 years and $4 \times 10^4 \text{ m}^3$ of lava/year for Kohala during its alkalic cap stage versus one eruption every few years and $30 \times 10^6 \text{ m}^3$ of lava/year for Kilauea during its shield-building stage [Spengler and García, 1988; Dzurisin et al., 1984]).

The major element analyses are plotted against MgO in Figure 3. Above about 6.5 wt % MgO the glasses contain only olivine with minor Cr-spinel. Below that value, augite \pm plagioclase join olivine as phenocrysts. This is well shown on the CaO versus MgO plot. The dominance of augite over plagioclase is shown on the $\text{CaO}/\text{Al}_2\text{O}_3$ versus MgO plot. Comparisons between

glasses from the different volcanoes are made below only for samples with >6.5 wt % MgO because olivine fractionation has only a minor effect on the trend of the element concentrations.

Studies of lavas from historic eruptions of Hawaiian volcanoes have shown systematic compositional differences between adjacent volcanoes and even substantial intravolcano variations [e.g., Wright, 1971; Tilling et al., 1987]. These studies used whole rocks, which may mask the magmatic compositions due to accumulation of phenocrysts. In this study, only glasses were used. There are some differences in the compositions of glasses from different volcanoes. For example, Loihi glasses are lower in Si and Al and higher in Ti, Fe, Ca, Na and K compared to glasses from other Hawaiian volcanoes at the same MgO level (Figure 3). Mauna Loa and Hualalai glasses are just the

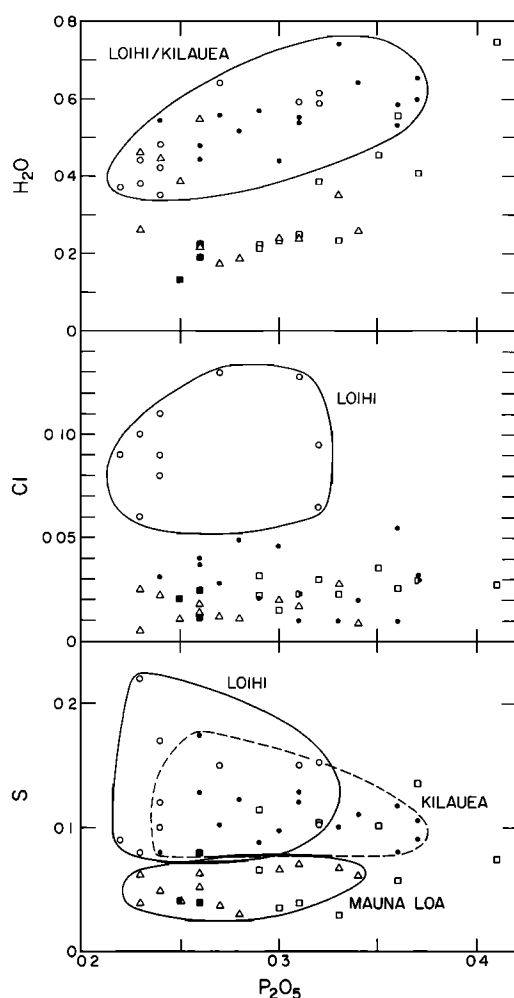


Fig. 4. P_2O_5 variation diagrams for H_2O , Cl and S in Hawaiian tholeiitic glasses. Symbols as in Figure 2.

opposite. Kilauea and Kohala glasses are intermediate. Mauna Kea glasses span a broad range from Mauna Loa-like to Kilauea-like (Figure 3).

Volatiles

Inter-volcano variations. All of the glasses except those from Kohala Volcano contain moderate to high volatile contents (0.28 to 0.92 wt % total volatiles; Table 2) indicating they were erupted in at least moderate ocean depths [see Moore, 1970; Killingley and Muenow, 1975]. The Kohala glass is distinct because it contains very low H_2O , CO_2 and S contents (<0.10 wt % total volatiles) which are characteristic of shallow marine or subaerial eruptions. Although the Kohala sample was dredged from a water depth of 1400–2600 m, it may not have been in place. Moore and Campbell [1987] have recently shown that Kohala has subsided about 1 km in the last 500 to 600 thousand years so it is not surprising that this sample was degassed to some extent.

As is the case for incompatible major elements, Loihi and Kilauea glasses have consistently higher volatile contents than most

Mauna Kea, Hualalai and Mauna Loa glasses of similar P_2O_5 content (Figure 4). P_2O_5 content is used as an indicator of degree of fractionation. (Note: The figures for volatiles also include data for tholeiitic glasses from Kilauea and Loihi volcanoes from Byers et al. [1985]). The volatile rich nature of Kilauea and Loihi glasses is principally a consequence of their higher H_2O and Cl contents (Figure 4). Schilling et al. [1980] first drew attention to the Cl-enriched nature of some mid-ocean ridge basalts (MORBs) near the Icelandic hot spot and referred to such enrichment as a "hot spot" effect. Loihi glasses have distinctly higher Cl abundances than other Hawaiian glasses (nearly an order of magnitude higher on the average compared to most Mauna Kea and Mauna Loa glasses; Figure 4). Thus, the Loihi glasses may have a greater percentage of the enriched component in their source. This is consistent with the $^3He/^4He$ -ratio data for Hawaiian lavas [Kurz et al., 1983]. Some Kilauea glasses also have high Cl contents. The higher Cl contents of Loihi and some Kilauea glasses may be diagnostic of young Hawaiian volcanoes indicating a temporal variation in the composition of Hawaiian tholeiites.

Another major difference in intervolcano volatile abundances is in the H_2O content of the glasses (Figure 4). A H_2O -rich group (H_2O/P_2O_5 ratio >1.4) contains all the Loihi and Kilauea glasses, four from Mauna Loa and two from Mauna Kea while a H_2O -poor field contains all the glasses from Hualalai and most of those from Mauna Kea (8 of 10) and Mauna Loa (8 of 12). Thus, in addition to the "hot spot" effect as evidenced by Cl enrichments, Kilauea and Loihi glasses also are H_2O -rich (as noted for other hot spots by Schilling et al. [1983] and Poreda et al. [1986]).

Sulfur contents of Loihi and Kilauea glasses are also distinctly higher than those of Mauna Loa glasses (Figure 4). This distinction persists when the glasses are compared at the same FeO content (Figure 5).

Carbon abundances in the Hawaiian glasses are highly variable and do not vary systematically with extent of fractionation. They are also higher than what would be expected on the basis of solubilities at the hydrostatic pressures corresponding to the water depths from which the samples were recovered. Stolper and Holloway

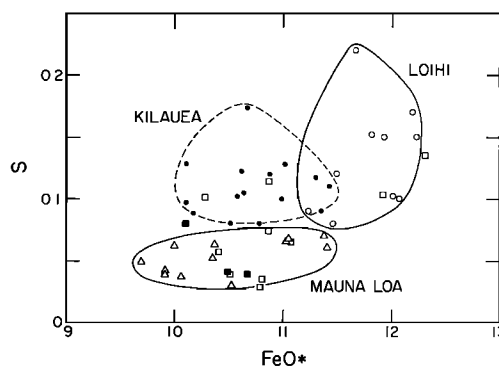


Fig. 5. Total FeO (wt %) versus S (wt %) for Hawaiian tholeiitic glasses. Symbols as in Figure 2.

TABLE 2. Volatile Abundances (wt %) in Hawaii Submarine Pillow Rim Glasses

Sample	Mg#	Total Volatiles	H ₂ O	CO ₂	S	Cl	F	CO	CH
<u>Loihi</u>									
LO-1-4	53.3	0.874	0.588	0.033	0.152	0.095	0.006	N.D.	N.D.
LO-1-6	52.5	0.869	0.613	0.066	0.102	0.065	0.006	0.013	0.004
LO-1-9	50.4	0.923	0.591	0.045	0.150	0.128	0.009	N.D.	N.D.
<u>Kilauea</u>									
KIL-1-BR	54.3	0.729	0.442	0.051	0.174	0.040	0.009	0.013	N.D.
KIL-1-1	54.7	0.745	0.516	0.046	0.122	0.049	0.012	N.D.	N.D.
KIL-1-4	54.5	0.766	0.557	0.054	0.102	0.028	0.008	0.017	N.D.
KIL-1-5	52.7	0.866	0.551	0.145	0.128	0.023	0.009	0.012	N.D.
KIL-1-9	57.6	0.804	0.542	0.136	0.080	0.031	0.006	0.006	0.003
KIL-2-1	55.5	0.642	0.437	0.047	0.097	0.046	0.015	N.D.	N.D.
KIL-2-8	55.2	0.782	0.568	0.097	0.088	0.021	0.008	N.D.	N.D.
KIL-2-9	56.5	0.727	0.478	0.075	0.128	0.037	0.009	N.D.	N.D.
KIL-3-9	51.9	0.827	0.597	0.084	0.105	0.032	0.009	N.D.	N.D.
KIL-3-19	50.6	0.906	0.585	0.139	0.117	0.055	0.010	N.D.	N.D.
<u>Mauna Loa</u>									
ML-1-1	59.0	0.646	0.445	0.124	0.049	0.022	0.006	N.D.	N.D.
ML-1-3	56.2	0.322	0.188	0.091	0.030	0.011	0.002	N.D.	N.D.
ML-1-10	51.9	0.520	0.350	0.067	0.067	0.028	0.008	N.D.	N.D.
ML-1-11	56.1	0.494	0.386	0.051	0.042	0.011	0.004	N.D.	N.D.
ML-1-12	57.0	0.607	0.461	0.069	0.062	0.005	0.004	0.006	N.D.
ML-2-1	53.1	0.443	0.240	0.109	0.066	0.020	0.008	N.D.	N.D.
ML-2-3	57.5	0.356	0.175	0.129	0.037	0.012	0.003	N.D.	N.D.
ML-2-8	60.5	0.423	0.262	0.094	0.039	0.025	0.003	N.D.	N.D.
ML-3-30	54.6	0.721	0.548	0.087	0.063	0.014	0.009	N.D.	N.D.
ML-4-2	50.9	0.403	0.259	0.058	0.061	0.009	0.003	0.013	N.D.
ML-4-10	59.3	0.414	0.217	0.125	0.052	0.018	0.002	N.D.	N.D.
ML-4-11	51.3	0.405	0.239	0.072	0.070	0.017	0.007	N.D.	N.D.
<u>Mauna Kea</u>									
MK-1-2	55.6	0.733	0.555	0.071	0.057	0.026	0.013	0.011	N.D.
MK-1-8	48.9	0.925	0.745	0.068	0.074	0.028	0.007	N.D.	0.003
MK-1-10	56.3	0.619	0.454	0.071	0.101	0.036	0.008	N.D.	N.D.
MK-2-1	51.2	0.612	0.385	0.089	0.103	0.030	0.005	N.D.	N.D.
MK-4-1	55.5	0.412	0.251	0.093	0.039	0.023	0.006	N.D.	N.D.
MK-5-2	47.1	0.642	0.407	0.058	0.135	0.030	0.009	0.003	N.D.
MK-5-5	54.9	0.371	0.233	0.080	0.035	0.015	0.006	N.D.	0.002
MK-5-25	53.9	0.353	0.214	0.077	0.066	0.023	0.006	N.D.	N.D.
MK-6-6	54.4	0.358	0.234	0.061	0.029	0.023	0.011	N.D.	N.D.
MK-6-18	54.9	0.403	0.221	0.087	0.114	0.032	0.006	N.D.	N.D.
<u>Hualalai</u>									
H-1-5	55.9	0.282	0.133	0.079	0.041	0.021	0.002	0.006	N.D.
H-1-7	57.4	0.390	0.226	0.092	0.039	0.012	0.002	0.007	0.012
H-P	56.8	0.393	0.191	0.081	0.080	0.025	0.002	0.005	0.009
<u>Kohala</u>									
KO-CAMP*	52.8	0.071	0.030	0.008	0.021	0.010	0.002	N.D.	N.D.

Mg# = $[(\text{Mg})/(\text{Mg}+0.9\text{Fe}^{\text{T}})] \times 100$; N.D. denotes none detected; *degassed.

[1988] suggest that the common occurrence of CO₂ contents in excess of experimentally determined solubilities is due to incomplete degassing of magmas that ascend rapidly. Although variable magma ascent rates may permit variable extents of CO₂ degassing and hence yield measured abundances not necessarily representative of magma source concentrations, this will not similarly affect the other volatiles since they are all far more soluble (see, for example, Basaltic Volcanism Study Project [1981]). The CO₂ contents reported in Table 2 are magmatic in

origin. They were obtained from the glasses at temperatures above 600°C. Carbon isotope studies have shown that CO₂ released at temperatures >600°C have a mantle signature ($\delta^{13}\text{C}$ between -4 and -8 relative to PDB). Carbon released at temperatures lower than 600°C is isotopically light ($\delta^{13}\text{C}$ approximately -20 to -26), typical of surficial organic carbon contamination (see, for example, Des Marais and Moore [1984] and Byers et al. [1986]).

Although the absolute abundance of CO₂ in the glass may vary with the measuring technique,

there are differences in CO_2 abundances among glasses from the different volcanoes. For example, the average CO_2 abundances in glasses from four volcanoes (Mauna Kea, Mauna Loa, Kilauea and Loihi) increase progressively with decreasing age: 0.074, 0.086, 0.092 and 0.101 wt %, respectively. Similarly, the range of values also increases from 0.058–0.093% for Mauna Kea to 0.033–0.170% for Loihi. Assuming that these glasses have sampled lavas progressively younger in the tholeiitic shield-building stage of these volcanoes (Mauna Kea through Loihi), these data indicate that as the volcano moves off the "hot spot", CO_2 abundances decrease and the range of values narrows. The decrease in CO_2 with age of the volcano may be related to variations in the degree of partial melting and differences in source composition and depth of storage and/or residence time of the magma in the crust (i.e., older volcanoes have longer-lived crustal magma chambers where the magmas lose CO_2).

Reduced carbon species (CO and CH_4) are present in small to trace amounts in some of the tholeiitic glasses especially from Kilauea and Hualalai volcanoes (Table 2). Our previous study reported for the first time in Hawaiian lavas small amounts of CO and/or CH_4 in glasses from alkalic basalts from Loihi but none were detected in the tholeiites from either Kilauea or Loihi [Byers et al., 1985]. Our new data show that reduced carbon species also occur in tholeiitic glasses from all of the volcanoes studied although they are detected in only 40% of the samples. There is no apparent correlation of abundance or presence of reduced-carbon species and any other geochemical parameter. The presence of reduced carbon may be related to storage time of the magma in the crust (i.e., longer storage favors oxidation of carbon [Mathez, 1984]).

Ratio:ratio plots using incompatible elements and a common denominator have been used to show the presence of one or more sources or magmas [Hart and Allegre, 1984]. The data for Hawaiian glasses are remarkably colinear and overlap on such a plot (Figure 6). The overlap for Mauna Loa with Kilauea is particularly striking because previous studies using subaerial samples have shown them to be geochemically distinct (for example, La/Sm ratio [Leeman et al.,

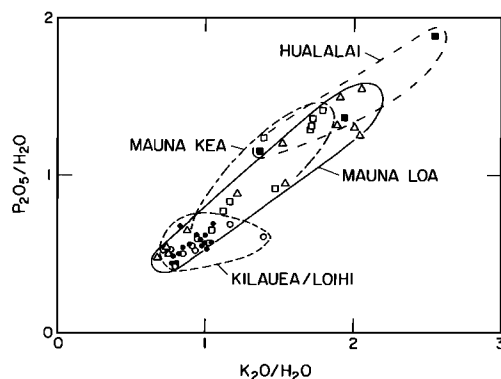


Fig. 6. $\text{K}_2\text{O}/\text{H}_2\text{O}$ versus $\text{P}_2\text{O}_5/\text{H}_2\text{O}$ plot for Hawaiian glasses. See text for explanation.

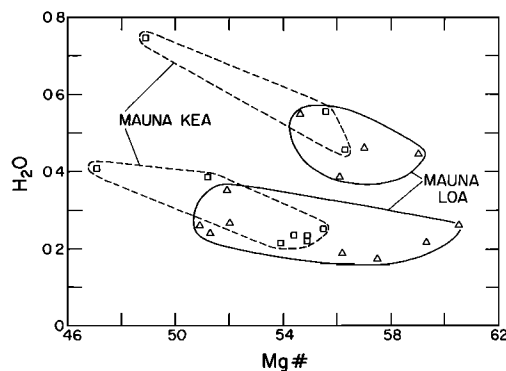


Fig. 7. Mg\# versus H_2O (wt %) for Mauna Loa and Mauna Kea glasses showing similar H_2O -enriched and -depleted trends.

1980]). Our interpretation of these data is that tholeiitic lavas from Hawaii volcanoes are derived by mixing of magmas from two distinct sources in variable proportions. This is consistent with new isotopic data on Hawaiian tholeiites [e.g., West et al., 1987]. Loihi and Kilauea lavas represent one end-member; Hualalai and some of the Mauna Kea and Mauna Loa lavas the other end-member. An apparent temporal variation in the composition of the lavas is present.

Intra-volcano variations. On plots of H_2O versus Mg\# [$100 \times \text{Mg}/(\text{Mg} + 0.9\text{Fe}^{\text{T}})$] there are two distinct and overlapping trends for glasses from Mauna Kea and Mauna Loa (Figure 7). For Loihi and Kilauea glasses, there is only one trend (Figure 8), which is also the same as the trend for the H_2O -rich group of Mauna Loa and Mauna Kea glasses. All Loihi glasses, most of the Kilauea glasses and the H_2O -rich Mauna Loa and Mauna Kea glasses are Cl-rich compared to the H_2O -poor samples. All of the Loihi, Kilauea and Mauna Kea H_2O -rich samples have high S contents, but none of the Mauna Loa glasses do. Thus there are two dominant volatile content groups with some overlap. The younger volcanoes (Loihi and Kilauea) have only the volatile-rich group, whereas the older volcanoes have two volatile groups.

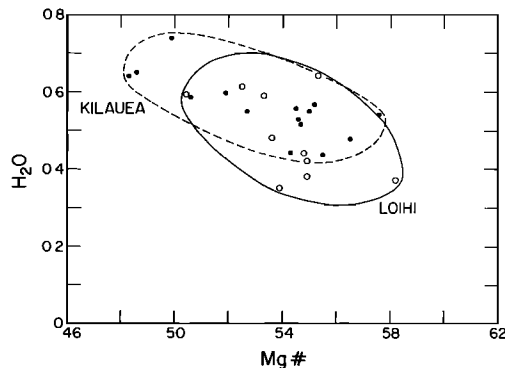


Fig. 8. Mg\# versus H_2O (wt %) for Kilauea and Loihi glasses showing one common H_2O -enriched trend.

The different volatile groupings are not related to degassing. The abundance of CO_2 , the least soluble volatile, is independent of the grouping (i.e., there are CO_2 -rich and -poor samples in each group). Also, K_2O contents vary similarly (except for Mauna Loa which has only one group). The large variation in H_2O for Mauna Loa and Mauna Kea may be a consequence of sampling lavas from a longer period of development. Lavas from early in the evolution of a volcano are H_2O -rich (for example, Loihi and Kilauea). Later lavas may be H_2O -poor. This variation may reflect the progressive depletion in the mantle source of incompatible elements during the growth of Hawaiian volcanoes.

Oxygen Isotopes

This study presents the first systematic oxygen isotope study of Hawaiian tholeiitic glasses. In a reconnaissance study of basalts from many different tectonic settings, Kyser et al. [1982] showed that Hawaiian tholeiites have whole rock $\delta^{18}\text{O}$ values that are generally lower than those of MORB (average values: 5.4 versus 5.7). Our study utilizes glasses from selected samples that span the mineralogical and compositional range from each volcano. The results for these samples confirm that Hawaiian tholeiites have relatively low $\delta^{18}\text{O}$ values of 4.6 to 5.6 (Table 3). The average value of ~5.1 is even lower than that reported by Kyser et al. [1982]. Samples from Kilauea, Loihi and Mauna Kea have consistently low values (4.9 to 5.3) while those from Hualalai are uniformly higher (5.5). In contrast, a large variation was found for samples from Mauna Loa (4.6 to 5.5) and Kohala (4.8 to 5.6).

Low- $\delta^{18}\text{O}$ basalts are not unique to Hawaii. Since the initial discovery of low- $\delta^{18}\text{O}$ basalts on Iceland [Muehlenbachs et al., 1974], their origin has been hotly debated [see Sheppard, 1986; Taylor, 1986]. The most common explanation for low- $\delta^{18}\text{O}$ values is assimilation of hydrothermally altered crust. Larson and Taylor [1986] suggested that emplacement in a rift zone tectonic setting may be an important factor in determining when this phenomenon takes place. In such environments there are numerous fractures that could allow meteoric waters to penetrate deep into the crust. Ascending magmas might assimilate such hydrothermally altered material to produce low- $\delta^{18}\text{O}$ lavas [Hattori and Muehlenbachs, 1982; Hemond et al., 1988]. The assimilation model for Iceland is supported by the correlation of evolved lavas and low values of $\delta^{18}\text{O}$ [Hemond et al., 1988].

The $\delta^{18}\text{O}$ values of Hawaiian glasses do not correlate with either the mineralogy or MgO content of the samples. However there is an inverse correlation of $\delta^{18}\text{O}$ with H_2O content and $\text{K}_2\text{O}/\text{P}_2\text{O}_5$ (Figure 9). The samples with high H_2O content and $\text{K}_2\text{O}/\text{P}_2\text{O}_5$ ratio have low $\delta^{18}\text{O}$ values. Note the good correlation of H_2O versus $\delta^{18}\text{O}$ for the Mauna Loa glasses which are the best studied group. The Kilauea and Loihi and some of the Mauna Kea and Mauna Loa samples have a high H_2O content and $\text{K}_2\text{O}/\text{P}_2\text{O}_5$ ratio which we attribute to a greater portion of a plume or hot spot component in their source. These correlations may be indicative that relatively

TABLE 3. Oxygen and Hydrogen Isotopic Values of Hawaiian Pillow Rim Glasses

Sample	$\delta^{18}\text{O}$	δD
<u>Loihi</u>		
LO-1-4	5.0	-84
LO-1-6	5.1	-69
LO-1-9	5.0	NA
<u>Kilauea</u>		
Kil-1-4	4.9	-61
Kil-1-9	4.9	-74
Kil-2-1	5.1	-62
Kil-2-8	5.0	-64
Kil-3-9	5.0	NA
<u>Mauna Loa</u>		
ML-1-1	5.0	-69
ML-1-3*	5.4	...
ML-1-7	5.5	...
ML-1-10	5.1	-69
ML-1-11	5.0	-80
ML-1-12	4.9	-83
ML-2-1*	4.7	...
ML-2-3*	5.5	...
ML-2-8*	5.1	...
ML-3-30*	4.6	...
ML-4-10*	5.3	...
ML-4-11*	5.3	...
<u>Mauna Kea</u>		
MK-1-2	5.0	-78
MK-1-8	5.0	-65
MK-5-2	5.3	-88
MK-5-5	4.9	-62
MK-6-6	5.1	NA
<u>Hualalai</u>		
H-P*	5.5	...
H1-5*	5.5	...
H1-7*	5.5	...
<u>Kohala</u>		
KO-Camp*	4.8	...
KO-2*	5.6	...
KO-1-17*	5.0	...

NA denotes not analyzed.

*Sample contains low H_2O . H isotopic analysis was unsuccessful or not attempted because insufficient glass was available.

low $\delta^{18}\text{O}$ values (~5.0-5.1) are a feature of the hot spot component.

The assimilation mechanism is probably not a viable explanation for the low- $\delta^{18}\text{O}$ Hawaiian glasses because many of the low- $\delta^{18}\text{O}$ samples also have pristine, "magmatic" δD values of -83 to -72 [Sheppard and Epstein, 1970; Boettcher and O'Neil, 1980] (Figure 10). Assimilation of substantial amounts of hydrothermally altered rock is required to modify the $\delta^{18}\text{O}$ value and this might (depending on the temperature of the assimilated rock) drastically raise the δD value [Kyser and O'Neil, 1984].

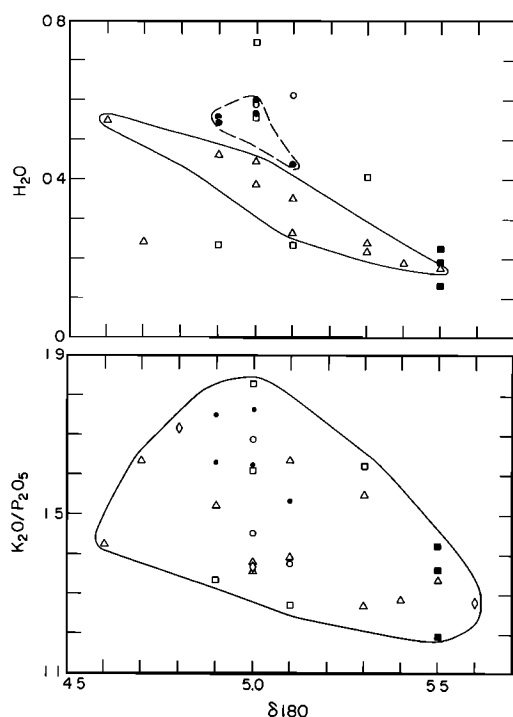


Fig. 9. Values of $\delta^{18}\text{O}$ versus $\text{K}_2\text{O}/\text{P}_2\text{O}_5$ ratio and H_2O (wt %) for some Hawaiian tholeiitic glasses. Symbols as in Figure 2. Fields show trend in the data.

As an alternative to the assimilation scenario these relatively low $\delta^{18}\text{O}$ values may be characteristic of the mantle source. Mantle-derived lherzolite xenoliths in Hawaiian posterosional lavas have $\delta^{18}\text{O}$ values of 5.6 to 6.2, values too high to be the source for most Hawaiian tholeiitic basalts unless there is a significant oxygen isotope fractionation on melting to produce the basalt [Kyser et al., 1982]. These xenoliths are from the lithosphere and are probably related to MORB generation. They are probably not a source for Hawaiian tholeiitic magmas [e.g., West et al., 1987]. There is evidence in other areas for mantle material with relatively low $\delta^{18}\text{O}$ values. Recently Harmon et al. [1986] reported $\delta^{18}\text{O}$ values of 4.9 to 5.7 for Mongolian lherzolite xenoliths. They proposed that these xenoliths were derived from a high-temperature diapir, a model which is similar to that proposed for the Hawaiian hot spot [see Clague and Dalrymple, 1987]. Harmon et al. [1986] also suggested that low $\delta^{18}\text{O}$ values may be representative of the large majority of the upper mantle; this is supported by oxygen isotope studies of Beni Bouchera, Lherz, and Lanzo high-temperature peridotites [Javoy, 1980]. Furthermore, relatively low $\delta^{18}\text{O}$ values (5.2–5.4) were recently reported for four basalts from another hot spot, Ascension Island [Weis et al., 1987]. Thus, the low $\delta^{18}\text{O}$ values of Hawaiian tholeiites may be characteristic of the plume source for Hawaiian magmas. This hypothesis is supported by the enrichment of H_2O , Cl, S and other

incompatible elements (interpreted to be characteristic of the plume source) in the glasses with low $\delta^{18}\text{O}$ values (Figure 9).

From a large number of oxygen isotope analyses of basalts, ultramafic massifs, mantle nodules, lunar rocks and ordinary chondrites [e.g., Taylor, 1968; Clayton, 1976; Javoy, 1980; Kyser et al., 1981], it would appear that there was a large and relatively uniform "primordial" reservoir of oxygen with a $\delta^{18}\text{O}$ value of about 6.0. The most abundant mantle-derived rock on Earth is MORB whose $\delta^{18}\text{O}$ values are exceedingly uniform at 5.7 ± 0.3 . The source of MORB, however, is clearly not primordial as it is highly depleted in incompatible elements. In fact the chemical and oxygen isotope composition of the MORB source should complement that of the continental crust. The continental crust contains abundant $\delta^{18}\text{O}$ -rich materials and has an average $\delta^{18}\text{O} > 7.5$ [e.g., Shieh and Schwarcz, 1977]. As pointed out by Taylor and Sheppard [1986], material balance considerations suggest that either the MORB reservoir is much larger than the continental crust, or there must be a significant reservoir of low $\delta^{18}\text{O}$ material elsewhere in the lower crust or mantle. We propose that this reservoir of low $\delta^{18}\text{O}$ material is the plume source.

This explanation may also be applicable for some Icelandic basalts. They have characteristically low $\delta^{18}\text{O}$ values < 5.5 [Muehlenbachs et al., 1974; Hattori and Muehlenbachs, 1982; Condomines et al., 1983; Hemond et al., 1988]. Although assimilation of hydrothermally altered rocks explains basalts with $\delta^{18}\text{O}$ values < 4.5 , others with values of 4.6 to 5.5 may simply reflect a component from a plume source. Many Icelandic lavas have high $^3\text{He}/^4\text{He}$ ratios which is considered characteristic of a plume source and not of hydrothermally altered rocks [e.g., Kurz et al., 1983; Condomines et al., 1983; Poreda et al., 1986]. Deuterium analyses of the Icelandic basalts might resolve the assimilation versus mantle source origin for these low $\delta^{18}\text{O}$ basalts.

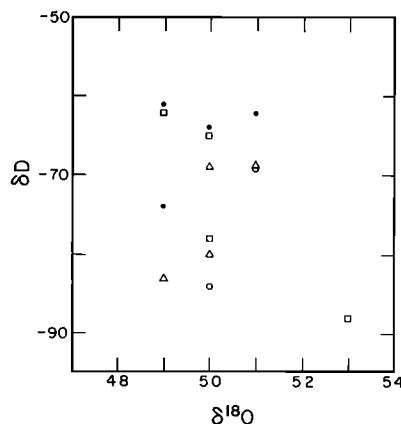


Fig. 10. Values of $\delta^{18}\text{O}$ versus hydrogen isotope ratio δD for some Hawaiian tholeiitic glasses. Symbols as in Figure 2. No apparent trend in data.

Hydrogen Isotopes

The Hawaiian glasses analyzed in this study have δD values from -88 to -61 (Table 3). For reference, δD values of MORB range from -84 to -60 [Kyser and O'Neil, 1984] and for hydrous mantle minerals they are usually -80 to -50 [Sheppard and Epstein, 1970; Boettcher and O'Neil, 1980; Poreda et al., 1986]. Previously reported values for Hawaiian submarine glasses (mostly from Kilauea lavas) range from -74 to -33 [Kyser and O'Neil, 1984; Rison and Craig, 1983]. Kyser and O'Neil [1984] noted a positive correlation between δD and H_2O content for Kilauea and related it to direct seawater addition to the magma. A similar trend was noted for glasses from the Mid-Atlantic Ridge just north and south of Iceland [Poreda et al., 1986]. However, they also found a positive correlation between these parameters and the La/Sm ratio which they interpreted to be related to the addition of an enriched mantle source component (either from a primary undegassed mantle or recycled altered crust [Poreda et al., 1986]).

No correlation is evident between H_2O and δD values in our analyses of Hawaiian tholeiitic glasses (Figure 11) nor is there a correlation with an enrichment ratio (for example, K_2O/P_2O_5). Thus the H_2O in these glasses is probably of magmatic origin. We concur with earlier, hydrogen isotope reconnaissance work [Craig and Lupton, 1976; Rison and Craig, 1983; Kyser and O'Neil, 1984] that magmatic δD values for many Hawaiian tholeiites are similar to those of MORB.

Geochemical Evolution of Hawaiian Volcanoes

The cause of the systematic major and volatile element differences between the young Loihi lavas and the lavas from the older volcanoes is not simply related to distinct sources. For example, the $^{87}Sr/^{86}Sr$ ratios of the tholeiitic lavas from Loihi, Kilauea and Mauna Loa volcanoes overlap (Loihi, 0.7035-0.7037 [Staudigel et al., 1984]; Kilauea, 0.7035-0.7037 [Hofmann et al., 1984; Stille et al., 1986; Mauna Loa, 0.7036-0.7038, Stille et al., 1986; Gurriet et al., 1988]). Furthermore,

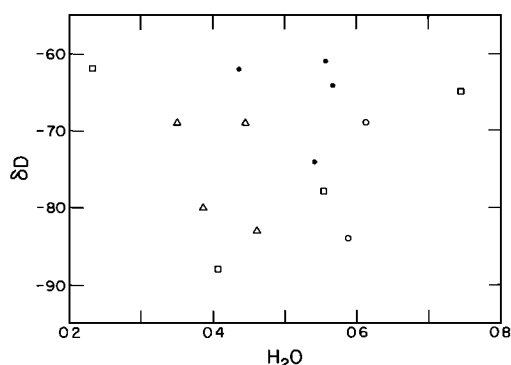


Fig. 11. H_2O (wt %) versus δD for some Hawaiian tholeiites. Symbols as in Figure 2.

all of the lavas have a significant plume component based on $^3He/^4He$ ratios although it is greater in the Loihi lavas (Loihi, 24-30 x atmosphere [Kurz et al., 1983]; Kilauea, 14-17 x atmosphere [Lupton and Garcia, 1986]; Mauna Loa, 14-16 x atmosphere [Lupton and Garcia, 1986]). Thus there may be other explanations for the major and volatile element variations.

Some studies of major elements in MORB have assumed a relatively homogeneous source and have related compositional variations to different degrees of partial melting or the depth of partial melting [e.g., Presnall and Hoover, 1984; Elthon and Scarfe, 1984]. Seismic evidence indicates that the conduits for Loihi, Kilauea and Mauna Loa (the active Hawaiian tholeiitic volcanoes) all extend to at least 50 km [Klein et al., 1987]. Below this depth, earthquakes are rare and difficult to locate. Rare earth element (REE) data for lavas from these volcanoes require garnet in their source [Leeman et al., 1980; Frey and Clague, 1983]. Thus, their magmas are derived at moderate to high pressures. Although we cannot place more precise limits on the depth of melt segregation at this time, it is unlikely that it plays a significant role in controlling compositional variation in Hawaiian tholeiites because garnet is in the source for the magmas for each volcano. (Note: Feigenson et al. [1983] argued that Kohala lavas could be generated without garnet in the source.)

The observed compositional variation from Loihi to Mauna Loa may be related to an evolutionary process within Hawaiian volcanoes. Loihi is the youngest and smallest (~2 km high) Hawaiian volcano; Mauna Loa is a fully mature (>9 km in thickness) volcano. Kilauea is intermediate. The systematic major and volatile element variations follow this sequence. The incompatible elements (Fe, Ti, Na, K and volatiles) and Ca become successively more depleted in the lavas from the older volcanoes. Such a variation may be related to the progressive melting and depletion of the "source" for each volcano. The decrease of incompatible elements with age of the volcano may be a consequence of the extensive melting needed to produce a Hawaiian shield volcano. For example, the volume of Mauna Loa, uncorrected for subsidence, is 42.5×10^3 km³ [Bargar and Jackson, 1974]. Moore [1987] suggested that subsidence has hidden 50% of the volcano. In any event, the true volume of a typical Hawaiian volcano is enormous. The decrease in Ca with age may indicate that clinopyroxene, an early melting phase in lherzolites [Takahashi and Kushiro, 1983], and/or garnet have been progressively depleted from the source.

Conclusions

The excellent suite of fresh submarine glasses that we collected from the volcanoes of the island of Hawaii has allowed us to document a remarkable evolutionary trend among Hawaiian tholeiites. Early tholeiitic lavas are enriched in Ca, Fe, Ti, Na, and K and depleted in Si and Al. Later lavas have the opposite trend. Volatile contents (H_2O , Cl, S and CO_2) are

distinctly higher in glasses from young Hawaiian volcanoes than from older volcanoes. This variation may be related to the progressive depletion of the source for Hawaiian volcanoes during their formation.

Oxygen isotope ratios of Hawaiian glasses are lower than those reported for MORBs (averages: 5.1 versus 5.7). This may be a distinct feature of hot spot lavas. The hydrogen isotope ratios of the Hawaiian samples fall broadly within the ranges of MORB and hydrous mantle minerals. Thus the H₂O contents of these glasses are probably of magmatic origin.

In concert with the progressive depletion of the "source" during the life of a Hawaiian shield volcano, there may also be a progressive change in the proportions of the two source components for Hawaiian tholeiites. Loihi and Kilauea lavas define an end-member on plots of Pb and Sr isotope ratios; Mauna Loa and Hualalai lavas define another (for volcanoes from the island of Hawaii). This would imply that the plume component has relatively low $\delta^{18}\text{O}$ (~-5.0) and moderate $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7034-0.7035) and $^{206}\text{Pb}/^{204}\text{Pb}$ (18.6-18.7) ratios and that the nonplume component has higher values for these isotope ratios. This eliminates the ~100 Ma MORB lithosphere under Hawaiian volcanoes as the nonplume component because its Sr isotope ratio is too low (~0.7030) to explain the observed variation (0.7035-0.7046).

This evolutionary model for Hawaiian tholeiites is an initial attempt to evaluate the evolution of the most voluminous stage in the development of Hawaiian shield volcanoes. It provides a framework to evaluate the evolution of other major volcanoes. However, it assumes that each Hawaiian volcano follows the same evolutionary sequence. This assumption may be tested by a 2-km drill hole into Mauna Kea that is being planned for the Deep Observation and Sampling of the Earth's Continental Crust program.

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