

Primitive magmas and source characteristics of the Hawaiian plume: petrology and geochemistry of shield picrites

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Abstract

A suite of tholeiitic picrites from eight of the younger (<2 Ma) Hawaiian shield volcanoes provides new information about the compositions of primitive magmas and source components in the Hawaiian plume. Olivine and bulk rock compositions show that parental melts at Hawaiian volcanoes have at least 13–17% MgO and ~10% Al₂O₃. The picrites have bulk compositions ranging from 14 to 30% MgO, and although most of these lavas have accumulated olivine + spinel, several have compositions that may approximate primitive melts. Olivine and spinel compositions show that the phenocrysts are closely related to the melt fraction of these lavas and are not accidental xenocrysts. Diverse isotopic compositions (Pb, Os, Sr, Nd) of these picrites require multiple sources in the Hawaiian plume, but key trace element characteristics (La, Nb abundances normalized to 16% MgO, Sm/Nd, Lu/Hf, La/Yb, Zr/Nb) are consistent with variable degrees of melting of a common, garnet-bearing source for all of the volcanoes except Koolau. The trace element composition of this Hawaiian pyrolite plume source can be modelled as an incipiently depleted, nearly primitive mantle that has lost a very small melt fraction, but a more complex origin may be more realistic. The Koolau picrites are exceptional in having anomalously low Nb and Ti contents, and high Zr/Nb ratios that fall off the melting arrays defined by the other picrites, indicating a distinctive source component that is also expressed in major element and isotopic compositions. The nearly constant Sr/Pb, Sr/Y, and Ba/Th ratios of these isotopically variable picrites are inconsistent with formation of the plume source either by bulk recycling of oceanic crust into the mantle, or by addition of dacitic melts from entrained eclogite to plume-derived basaltic magmas. Alternatively, the Hawaiian plume may consist of variably depleted mantle that was enriched by small-degree melts, possibly during subduction or entrainment of lithospheric mantle. Radiogenic ¹⁸⁶Os/¹⁸⁸Os isotopic compositions of these picrites are consistent with transport of this material to the deep lower mantle and addition of a small amount of outer core to the plume source. © 1999 Elsevier Science B.V. All rights reserved.

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1. Introduction

Hawaiian volcanoes track the largest, hottest, and one of the longest lived mantle plumes currently active on Earth. Despite intensive study, debate continues over such fundamental issues as the compositions of primitive magmas that supply Hawaiian volcanoes, processes of magma generation, and the scale and origin of compositional heterogeneities within the plume. Moderately evolved tholeiites with 7–9% MgO dominate the subaerial exposures of these volcanoes, and the compositions of these lavas provide important constraints on the nature of the Hawaiian plume [1,2]. Recent studies have shown, however, that the primary, source-related features of even moderately evolved ocean island basalts (OIB) can be obscured by processes such as assimilation of wallrock and oceanic lithosphere, polybaric fractionation, mixing in shallow reservoirs, and interaction with seawater and hydrothermal fluids [3,4].

Picritic shield-building tholeiites potentially provide greater insights into the compositions of primitive magmas feeding Hawaiian volcanoes. Although uncommon subaerially, picrites are abundant on the submarine flanks of Hawaiian volcanoes where deeper levels of the magma system may have been tapped [5,6]. Complicating such an approach is the fact that many Hawaiian picrites are demonstrably not primitive magmas, but formed by accumulation of olivine into evolved melts [7,8]. The primitive characteristics of such picrites may be compromised, e.g., through fractionation of plagioclase and clinopyroxene prior to eruption, and mixing between primitive and evolved magmas. Not all Hawaiian picrites have been compromised, however, and examples with primitive magmatic characteristics have been recognized [5,6,9].

This paper describes a suite of picritic lavas from eight Hawaiian volcanoes that spans 2 million years of plume activity and includes the major source components recognized in Hawaiian tholeiites [10–13]. Although most of these lavas have accumulated variable amounts of olivine and spinel, several have bulk compositions that may approximate primitive melts with 14–17% MgO. These picrites provide insights into the composition of primitive Hawaiian magmas, and the relative roles of melting and source characteristics in creating the shield lavas. We also

evaluate proposals calling for recycled oceanic crust in the Hawaiian plume [13–15], and conclude that these models are inconsistent with the trace element and isotopic characteristics of primitive Hawaiian magmas. Alternatively, the plume may consist of moderately depleted mantle that was variably re-fertilized, possibly during subduction or entrainment of the lithosphere. Transport of this material into the deep lower mantle is indicated by the coupled enrichments of ^{187}Os and ^{186}Os in these picrites, which are consistent with addition of a small amount of outer core to the plume source [16].

2. Samples and analytical procedures

This study includes dredge samples from the submarine flanks of Kilauea, Loihi, Mauna Loa, Mauna Kea, Hualalai, and Kohala (see [17] for glass, volatile element, and stable isotopic compositions, and [18] for rare gas analyses of most of these samples). Also included are four subaerial flows from the SW rift of Mauna Loa, including the 1868 eruption and three prehistoric (>40 ka) flows from near the base of the Kahuku section (MLKAH) [5], the 1840 east rift eruption of Kilauea, two subaerial flows from Koolau, and one subaerial flow from Kahoolawe. One of the Koolau picrites (KOO-17a) has been studied previously [11,19]; the other (KOO-CF) was collected near Waipahu on the west flank of the volcano. The sample from Kahoolawe (KW-25) was also studied by Leeman et al. [20]. The Loihi picrite LO-02-04 was notably richer in olivine in the interior of the flow, so two samples were prepared, one from the phenocryst-rich interior of the flow (LO-02-04C) and one from the exterior portion (LO-02-04F). This difference in phenocryst content is reflected in the MgO, Ni, Co, and Cr contents of these samples.

Polished thin sections were prepared and modal abundances of phenocrysts and matrix obtained by point counting (Table 1). Mineral compositions were determined by electron microprobe at Macquarie University and the University of Hawaii. For the bulk rock samples, interior fragments of the flows were broken into small chips using a hydraulic press, washed in deionized water, and reduced to powder using an agate mortar and pestle, and an agate ring mill. Major and trace element compositions (Table 2)

Table 1

Modes of Hawaiian picritic basalts based on 500* or 1000 points/sample; vesicle-free

Sample	MgO (wt%)	Olivine				Cpx mph	Plag mph	Matrix
		ph	dph	rph	mph			
LO-02-01*	27.4	20.0	5.8	0.2	14.2	0.6	–	59.2
LO-02-02*	24.4	25.8	2.0	0.2	13.2	–	–	58.8
LO-02-04*	29.6	22.8	7.0	4.0	9.2	–	–	57.0
KIL-1-7	16.0	15.6	0.3	2.0	1.8	–	–	80.3
KIL-1-10	18.2	16.4	2.6	1.4	4.8	–	–	74.8
KIL-1-18	13.8	12.1	0.2	0.4	1.2	–	–	86.1
KIL-2-1	20.7	18.4	4.2	1.8	5.2	–	–	70.4
KIL 1840	14.3	13.9	0.9	4.9	1.5	0.5	–	78.3
ML-2-50*	19.9	12.4	<0.1	–	10.2	–	–	77.4
ML1868*	21.5	18.8	1.8	–	8.8	–	–	70.6
MLKAH-1*	21.7	16.4	6.2	0.6	11.6	–	–	65.2
MLKAH-2* ^a	16.5	15.8	3.0	–	6.0	1.0	0.8	73.4
MK-1-6	17.2	8.6	9.0	2.0	9.6	–	–	70.8
H-5	14.9	9.0	4.8	3.1	2.1	0.1	–	80.9
H-11	13.7	11.6	1.9	3.6	2.0	–	–	80.9
H-23	25.8	25.2	11.1	5.6	1.4	–	0.1	56.6
KOH-1-28	20.5	9.6	1.6	2.4	7.6	–	–	78.4
KOO-CF	15.4	14.4	0.6	0.7	1.4	–	–	82.9
KOO-17a	20.9	14.5	12.5	–	1.1	–	0.1	71.8

Abbreviations: ph = phenocryst; dph = deformed phenocryst; rph = resorbed phenocryst; mph = microphenocryst; cpx = clinopyroxene; plag = plagioclase; gdms = groundmass.

^a Orthopyroxene-bearing (~0.2 vol% mph).

were determined by XRF at Macquarie University [21] and by solution ICP–MS at the Australian National University [22]. Pb isotopic compositions (Table 2) were determined by thermal ionization mass spectrometry at the ANU using a ²⁰⁷Pb–²⁰⁴Pb double spike to correct for fractionation. Sr, Nd, and Os isotopic compositions for a subset of these lavas, and additional information on procedures, blanks, and standard determinations for the Pb isotopic analyses are given by Bennett et al. [12] and Brandon et al. [16].

All of the samples are petrographically fresh except for KOO-17a which shows evidence for sub-aerial weathering in thin section and in its low alkali and U contents (Table 2). The Loihi picrites are notable for the remarkably high water contents (up to 1.5 wt% H₂O) trapped in the quenched glassy matrix and melt inclusions from these lavas [23], accounting for the low totals in the major element analyses of these rocks (Table 2). These high water contents apparently reflect incorporation of a high-*T* fluid or brine into the melt during an early stage of magmatic evolution [23]. These Loihi picrites are

also unusually rich in immiscible sulfides and they have elevated Pb concentrations, but alkali element contents (e.g., Ba/Rb and Rb/Cs ratios), Pb isotopic compositions, and other trace element characteristics of these lavas are similar to those of more normal Loihi tholeiites [24–26]. The sample from Mauna Kea (MK-1-6) has somewhat lower Sr contents compared to the other picrites, consistent with the conclusion that most of studied picrites from Mauna Kea represent accumulation of olivine into relatively evolved melts that have lost plagioclase [8].

3. Petrography and mineral compositions

The picrites are rapidly cooled lavas with phenocrysts and microphenocrysts of olivine and chromian spinel suspended in a glassy or fine-grained groundmass (Table 1). Inclusions of spinel and quenched melt are common in the olivine phenocrysts. Plagioclase and clinopyroxene phenocrysts such as those in some compromised picritic lavas from Kilauea [7], Mauna Kea [8], and Mauna Loa

Table 2

Major element, trace element, and Pb isotopic compositions of Hawaiian picrites

	Loihi				Kilauea					Mauna Kea		Hualalai	
	LO-02-01	LO-02-02	LO-02-04F	LO-02-04C	KIL-1-7	KIL-1-10	KIL-1-18	KIL-2-1	KIL 1840	MK-1-6	H-5	H-11	H-23
Major elements (wt%)													
SiO ₂	44.39	44.63	44.61	43.97	48.96	48.77	49.35	47.98	49.32	49.21	49.70	49.93	46.01
TiO ₂	1.22	1.40	1.43	1.08	1.89	1.77	2.03	1.71	2.11	2.04	1.67	1.74	1.08
Al ₂ O ₃	6.14	7.13	7.21	5.51	10.45	9.90	11.22	9.12	11.38	9.83	11.09	11.46	7.37
FeO-t	11.81	11.95	11.77	12.11	11.14	11.17	11.33	11.31	11.22	11.46	11.12	10.98	11.57
MnO	0.18	0.19	0.16	0.19	0.17	0.17	0.18	0.16	0.18	0.18	0.18	0.17	0.17
MgO	27.43	24.21	23.35	29.59	15.97	18.24	13.80	20.66	14.27	17.24	14.93	13.67	25.80
CaO	5.53	6.33	6.48	4.95	8.53	8.12	9.14	7.39	9.19	7.29	8.71	9.02	5.69
Na ₂ O	1.20	1.34	1.25	1.11	1.52	1.45	1.64	1.30	1.79	1.76	1.71	1.86	1.19
K ₂ O	0.32	0.40	0.39	0.27	0.34	0.30	0.39	0.30	0.37	0.39	0.26	0.27	0.19
P ₂ O ₅	0.12	0.14	0.14	0.09	0.16	0.15	0.17	0.15	0.19	0.22	0.15	0.16	0.10
Total	98.34	97.72	96.79	98.87	99.13	100.04	99.25	100.08	100.02	99.62	99.52	99.26	99.17
Trace elements (ppm)													
Cr xrf	2154	2032	2005	2300	930		864	1319	959	1150	902	818	1843
Cu xrf	84	103	100	75	71	96	78	64	79	64	79	81	49
Ni xrf	1395	1219	1158	1558	758	1045	565	1087	583	827	715	617	1391
Zn xrf	93	95	94	93	89	105	92	93	91	98	89	88	86
Li ICP-MS	2.7	3.0	2.9	2.8	3.5	3.9	3.5	3.8	3.6	4.7	4.0	4.1	2.9
Be	0.48	0.66	0.60	0.49	0.74	0.63	0.78	0.62	0.67	0.87	0.63	0.66	0.38
Sc	17.2	21.1	21.2	16.9	27.9	24.4	29.2	22.5	25.4	26.1	28.0	28.8	20.0
V	162	183	191	145	243	228	261	211	251	241	263	280	182
Co	105	99	97	113	71	81	67	86	64	72	72	70	114
Ga	9.8	10.6	11.6	8.9	15.1	15.5	16.2	14.5	17.0	15.7	16.0	16.5	10.6
Rb	5.8	6.5	6.9	4.9	5.8	5.5	6.3	5.4	7.0	6.9	4.3	4.5	3.1
Sr	171	190	204	146	245	236	260	219	291	211	223	233	150
Y	13.5	15.1	15.7	10.5	22.0	21.0	22.7	19.8	24.7	26.6	21.5	22.3	15.0
Zr	85	92	97	71	127	122	135	117	146	157	108	112	74
Nb	9.3	11.0	10.8	7.9	11.0	10.7	12.0	10.3	12.1	11.9	8.1	8.5	5.3
Mo	0.47	0.53	0.53	0.40	0.54	0.63	0.64	0.56	0.63	0.71	0.58	0.50	0.29
Cs	0.062	0.077	0.078	0.072	0.071	0.054	0.078	0.096	0.065	0.075	0.048	0.044	0.035
Ba	70	75	82	63	79	73	86	72	95	88	53	56	33
La	7.22	7.73	8.43	6.49	9.05	8.59	9.94	8.27	9.68	10.32	6.88	7.19	4.33
Ce	16.6	18.1	19.7	14.5	22.4	21.3	25.0	20.3	24.2	26.2	17.3	18.1	10.9
Pr	2.15	2.37	2.55	1.95	3.09	2.98	3.43	2.82	3.36	3.70	2.43	2.53	1.54
Nd	10.2	11.1	11.9	9.1	15.6	14.5	17.2	14.1	17.0	18.1	12.1	12.6	7.7
Sm	2.44	2.69	2.94	2.19	4.02	3.76	4.43	3.77	4.52	4.92	3.42	3.52	2.16
Eu	0.82	0.88	0.97	0.72	1.40	1.31	1.52	1.26	1.57	1.61	1.21	1.25	0.77
Gd	2.68	2.93	3.16	2.36	4.36	4.14	4.94	4.07	5.01	5.53	3.96	4.12	2.61
Tb	0.43	0.47	0.51	0.37	0.72	0.67	0.77	0.65	0.79	0.90	0.65	0.67	0.43
Dy	2.30	2.47	2.73	2.06	3.93	3.71	4.34	3.59	4.40	4.91	3.72	3.87	2.45
Ho	0.45	0.49	0.52	0.39	0.77	0.72	0.84	0.69	0.85	0.97	0.73	0.76	0.48
Er	1.12	1.24	1.33	1.00	1.91	1.81	2.14	1.74	2.09	2.43	1.89	1.96	1.25
Yb	0.93	1.03	1.08	0.82	1.58	1.49	1.71	1.41	1.71	1.99	1.56	1.65	1.04
Lu	0.13	0.14	0.15	0.11	0.22	0.21	0.23	0.20	0.24	0.28	0.22	0.23	0.15
Hf	1.96	2.16	2.26	1.72	3.20	2.92	3.37	2.82	3.37	3.91	2.53	2.61	1.63
Pb	0.75	0.86	0.91	0.69	0.74	0.68	0.80	0.71	0.88	0.88	0.67	0.63	0.44
Th	0.58	0.64	0.66	0.52	0.71	0.65	0.75	0.61	0.73	0.81	0.46	0.47	0.29
U	0.17	0.17	0.20	0.15	0.22	0.21	0.23	0.19	0.23	0.26	0.15	0.16	0.10
Pb isotopic compositions													
²⁰⁶ Pb/ ²⁰⁴ Pb	18.340	18.331	18.355			18.464	18.465	18.426	18.453	18.396	18.233	18.237	18.195
²⁰⁷ Pb/ ²⁰⁴ Pb	15.464	15.459	15.477			15.471	15.467	15.463	15.470	15.462	15.455	15.458	15.453
²⁰⁸ Pb/ ²⁰⁴ Pb	38.137	38.123	38.177			38.088	38.088	38.073	38.086	38.052	38.005	38.013	37.952

Table 2 (continued)

	Mauna Loa					Kohala	Kaoolawe	Koolau	
	ML-2-50	MLKAH-1	MLKAH-2	MLKAH-3	ML1868	KOH-1-28	KW-25	KOO-CF	KOO-17a
Major elements (wt%)									
SiO ₂	48.44	47.81	49.78	49.43	47.96	48.64	49.69	51.03	47.62
TiO ₂	1.40	1.53	1.59	1.57	1.37	1.41	1.84	1.70	1.38
Al ₂ O ₃	9.46	8.83	10.90	10.73	8.88	9.30	10.75	10.80	9.77
FeO-t	11.26	11.47	10.69	10.72	11.37	10.97	11.87	10.87	11.39
MnO	0.17	0.18	0.17	0.16	0.18	0.16	0.17	0.14	0.17
MgO	19.92	21.66	16.46	16.90	21.48	20.52	15.27	15.37	20.89
CaO	7.38	6.68	8.42	8.26	6.82	7.21	8.38	6.71	6.36
Na ₂ O	1.54	1.47	1.61	1.57	1.46	1.30	1.83	1.96	1.50
K ₂ O	0.22	0.25	0.22	0.24	0.26	0.26	0.34	0.47	0.09
P ₂ O ₅	0.12	0.15	0.14	0.14	0.14	0.12	0.19	0.21	0.17
Total	99.91	100.03	99.98	99.72	99.90	99.89	100.33	99.26	99.34
Trace elements (ppm)									
Cr xrf	1095	1148	827	819	1451	1188	853	844	834
Cu xrf	63	29	56	64	62	60	120	72	64
Ni xrf	1073	1122	863	872	1033	1020	635	924	1089
Zn xrf	87	89	86	84	90	85	102	97	99
Li ICP-MS	3.7	3.3	4.1	3.6	3.4	3.4	4.1	4.8	4.5
Be	0.52	0.47	0.61	0.49	0.53	0.54	na	1.06	0.67
Sc	24.8	19.8	28.9	21.8	21.4	23.9	27.4	19.0	19.1
V	206	200	207	206	180	226	273	196	175
Co	81	86	64	69	81	92	67	94	83
Ga	13.2	13.6	14.5	15.2	12.2	13.1	15.4	16.7	13.1
Rb	3.4	4.0	3.3	3.5	4.3	3.9	4.9	7.6	0.3
Sr	183	193	203	217	206	199	254	343	273
Y	18.9	21.8	19.6	21.7	17.6	19.3	21.8	20.2	17.9
Zr	93	115	98	111	95	103	116	132	110
Nb	6.0	7.8	6.3	7.0	6.3	7.0	8.6	8.2	6.2
Mo	0.39	0.42	0.38	0.39	0.42	0.44	na	0.57	0.27
Cs	0.035	0.036	0.032	0.032	0.043	0.037	0.041	0.088	0.005
Ba	45	50	47	47	59	49	144	95	38
La	5.45	6.42	6.02	6.04	6.02	6.26	7.49	10.08	8.03
Ce	14.1	16.6	16.0	15.6	15.4	15.7	17.7	23.8	20.0
Pr	2.03	2.31	2.36	2.25	2.19	2.21	2.53	3.36	2.81
Nd	10.4	11.7	11.9	11.7	11.0	11.0	12.8	16.8	14.0
Sm	3.01	3.30	3.48	3.39	3.05	3.00	3.58	4.35	3.44
Eu	1.06	1.13	1.20	1.20	1.05	1.03	1.31	1.48	1.17
Gd	3.53	3.88	3.86	3.97	3.40	3.47	3.87	4.49	3.67
Tb	0.58	0.62	0.66	0.65	0.55	0.56	0.62	0.68	0.57
Dy	3.32	3.54	3.74	3.65	3.10	3.21	3.56	3.70	3.18
Ho	0.65	0.70	0.74	0.72	0.61	0.62	0.74	0.69	0.61
Er	1.70	1.79	1.88	1.83	1.56	1.60	1.98	1.71	1.55
Yb	1.41	1.46	1.55	1.53	1.26	1.35	1.56	1.31	1.28
Lu	0.20	0.21	0.21	0.22	0.18	0.19	0.23	0.19	0.18
Hf	2.34	2.59	2.50	2.57	2.28	2.27	2.70	3.10	2.62
Pb	0.56	0.65	0.62	0.59	0.64	0.62	0.70	1.03	0.81
Th	0.38	0.46	0.41	0.41	0.40	0.39	0.45	0.60	0.44
U	0.14	0.13	0.13	0.12	0.14	0.13	0.14	0.17	0.04
Pb isotopic compositions									
²⁰⁶ Pb/ ²⁰⁴ Pb						18.072	18.047	17.843	17.871
²⁰⁷ Pb/ ²⁰⁴ Pb						15.445	15.428	15.439	15.444
²⁰⁸ Pb/ ²⁰⁴ Pb						37.792	37.786	37.774	37.870

FeO-t = total Fe; major element and Pb isotopic composition of KW-25 from [20].

[5,9] are absent from the picrites described here, although microphenocrysts of these minerals are present in a few of the lavas. In only two cases do these minerals account for >1 vol% and both of these lavas erupted subaerially and cooled slowly.

The vast majority of olivine phenocrysts are euhedral and normally zoned with a restricted range of core compositions around Fo_{87–90} (Fig. 1). CaO contents of olivines are typical of magmatic values (0.15–0.41 wt%) and they show consistent differences between the volcanoes that are related to melt composition (Fig. 2). All of the picrites contain at least some mildly deformed olivines such as those described in other olivine-rich Hawaiian lavas [5,9,27–29]. No systematic compositional differences between the deformed and undeformed phenocrysts are found, and quenched melt inclusions are common in both. All of these characteristics lead us to agree with previous interpretations that the olivine phenocrysts and melt matrix of each sample are closely related, and that the strained olivines do not represent mantle xenocrysts or unrelated cumulates from different volcanoes [5,9,28,29]. The excellent correlation between Os and Pb isotopic compositions of these picrites [12,16] is consistent with this, as Os would be concentrated in the phenocrysts whereas Pb would be primarily in the melt.

Chromian spinels also have primitive compositions. Cores of spinel grains have a strong mode at Mg# (atomic Mg/Mg + Fe²⁺) = 55–65, and Cr# (atomic Cr/Cr + Al) = 65–70 (Fig. 3). Many of these spinels have a brown translucent color in thin section, which is consistent with the primitive composition. A general feature of spinels in Hawaiian tholeiites, including those from these picrites, is their low Al₂O₃ (typically 10–20 wt%) and high Cr#, similar to spinels in basaltic komatiites and intermediate between those in MORB and boninites (Fig. 3). Spinel compositions in these picrites, therefore, provide additional support for the conclusion based on olivine compositions that the phenocrysts in these lavas bear a close genetic relationship to the melt and are not xenocrysts from the MORB oceanic lithosphere. Spinel compositions also provide an opportunity to evaluate the *f*O₂ under which they crystallized. Fe³⁺/ΣFe ratios of spinels in the picrites, calculated from electron microprobe analyses based

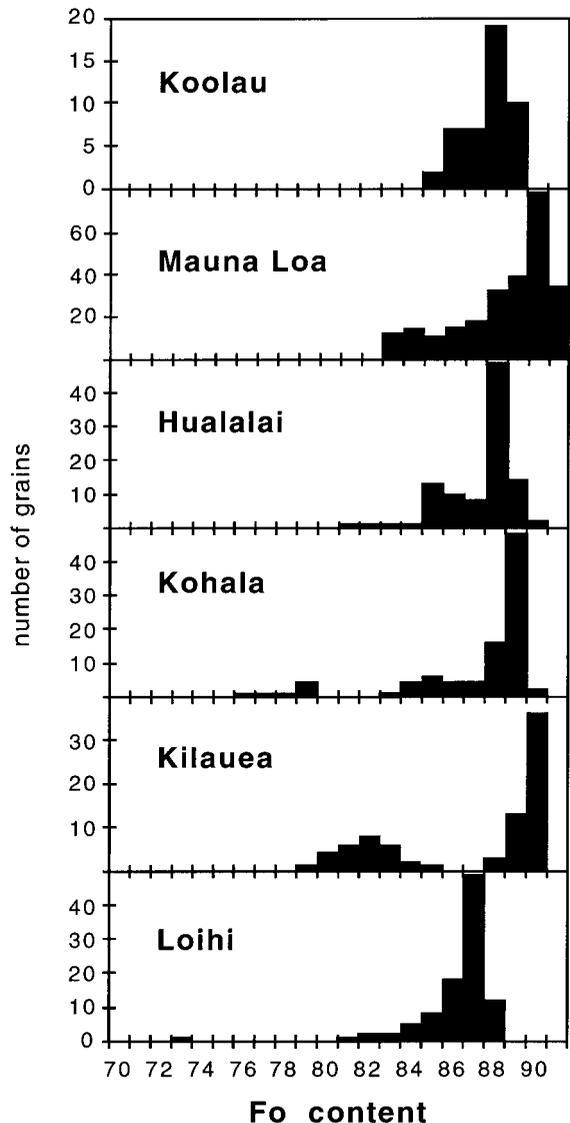


Fig. 1. Histogram of olivine phenocryst compositions in Hawaiian picrites. The abundance of Fo_{87–89} olivines shows that parental magmas with least 13 wt% MgO are common to the shield-building stage of Hawaiian volcanoes. The most magnesian olivines crystallized from parental magmas with at least 17% MgO.

on stoichiometry, have a strong mode at 0.30 ± 0.05 . These compositions indicate an *f*O₂ near QFM or Ni–NiO [30], with no apparent difference between the volcanoes.

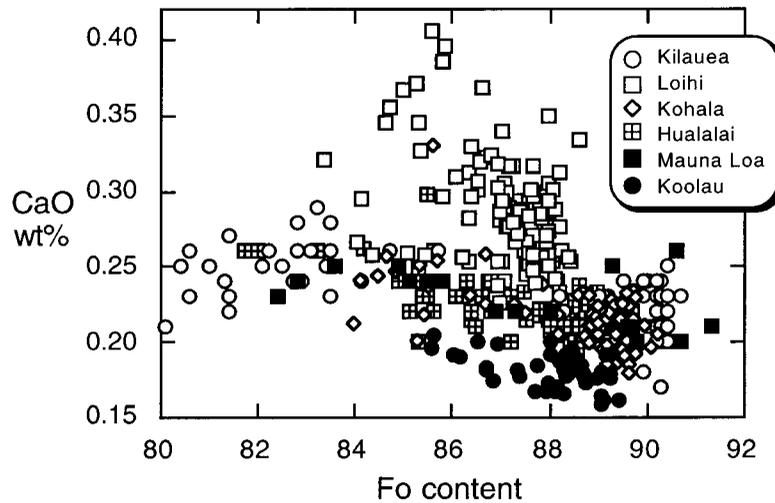


Fig. 2. CaO contents of normally zoned olivine phenocrysts in Hawaiian picrites correlates with melt composition. The high CaO contents of olivines in the Loihi picrites corresponds to high CaO contents in the melt and high Ca/Al ratios in the bulk rock (Table 2). In contrast, picrites from Koolau have melts with low CaO contents and low Ca/Al ratios, and this is reflected in the low CaO concentrations of their olivines. Picrites from Kilauea, Hualalai, Kohala, and Mauna Loa have intermediate compositions. The correlation between Ca content of the melt and the olivines shows that the phenocrysts and melt matrix of these lavas are closely related and that the olivines are not unrelated xenocrysts. Some reversely zoned grains $<F_{o80}$ (Fig. 1) not shown for clarity.

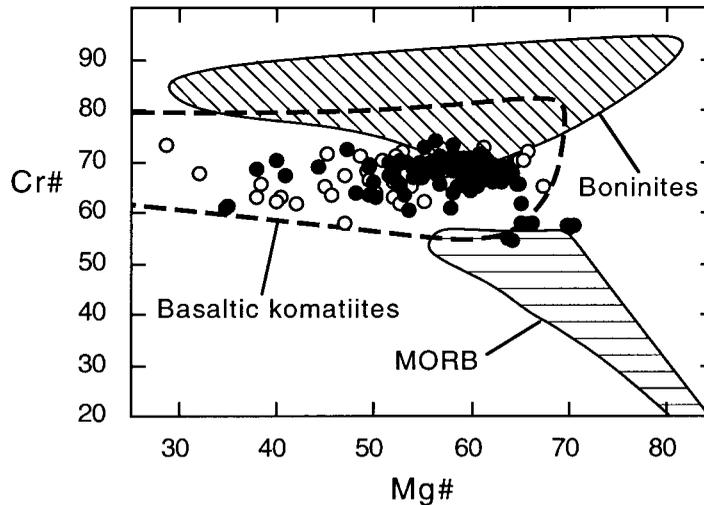


Fig. 3. Chromian spinel compositions in the picrites from this study (filled circles) and other picritic and moderately evolved tholeiites from Kilauea and Mauna Loa (open circles, data sources cited in the text) compared to representative compositions from MORB [58], boninites [59], and basaltic komatiites [60].

4. Major element compositions of primitive Hawaiian melts

The picrites contain 14–30 wt% MgO, with MgO contents and bulk rock Mg# strongly correlated with

the proportion of modal olivine (Tables 1 and 2). Similar forsterite contents of the olivine phenocrysts (Fig. 1) over a range of bulk rock MgO and Mg# (Table 2) suggests that many of these picrites have accumulated olivine and do not represent primary

liquid compositions. Nonetheless, useful constraints on the compositions of primitive magmas feeding Hawaiian volcanoes can be derived from their mineralogy and major element compositions if the effects of olivine accumulation can be accommodated.

High forsterite contents of euhedral, undeformed olivine phenocrysts in the picrites show that their parental melts had MgO contents significantly greater than those of the moderately evolved tholeiites that dominate the subaerial exposures of Hawaiian shields. Olivines with Fo_{87-91} cores dominate the phenocryst populations of the picrites (Fig. 1) but such primitive compositions are rare in subaerial Hawaiian tholeiites that erupted from the summit reservoir (typically Fo_{77-84} [6]). Melts in equilibrium with the forsteritic olivines in the picrites would have 13–17% MgO (assuming $\text{Fe}^{2+}/\Sigma\text{Fe} = 0.9$; $K_d = 0.3$), and they would have 1 atm liquidus temperatures of $\sim 1400^\circ\text{C}$. The discovery of picritic glass near the base of Kilauea provides direct evidence that some Hawaiian melts had at least 15 wt% MgO [7], and such melts appear to be a common occurrence on the submarine flanks of all Hawaiian volcanoes. It is perhaps not surprising that such high MgO liquids are seldom erupted on the subaerial shields, as crystallization of only 2% of Fo_{90} olivine

would reduce the MgO content of the melt by 1 wt%, and intracrystalline diffusion would rapidly re-equilibrate the compositions of olivine phenocrysts in the presence of more evolved melt. Thus, the observed olivine compositions probably provide minimum estimates for the MgO content of Hawaiian primary magmas.

Bulk rock compositions also provide constraints on the characteristics of primitive Hawaiian melts. Log–log plots of an incompatible element (Al_2O_3) vs. a compatible element (MgO) can distinguish compositions related by fractional crystallization vs. accumulation, thereby providing an estimate of parental melt compositions [6]. On such a plot (Fig. 4), Hawaiian tholeiites with <16% MgO fall along a linear trend indicating that they formed by fractional crystallization, whereas more magnesian picrites curve away from the fractionation trend indicating accumulation of olivine. Parental melt compositions with $\sim 16\%$ MgO, $\sim 10\%$ Al_2O_3 , and 500 ppm Ni (not shown) are indicated by the break in slope between the fractionation and accumulation trends (Fig. 4). Although all of the picrites studied here contain abundant phenocrysts, several have 14–17% MgO, and so preserve bulk compositions that may approximate primitive melts (Table 2).

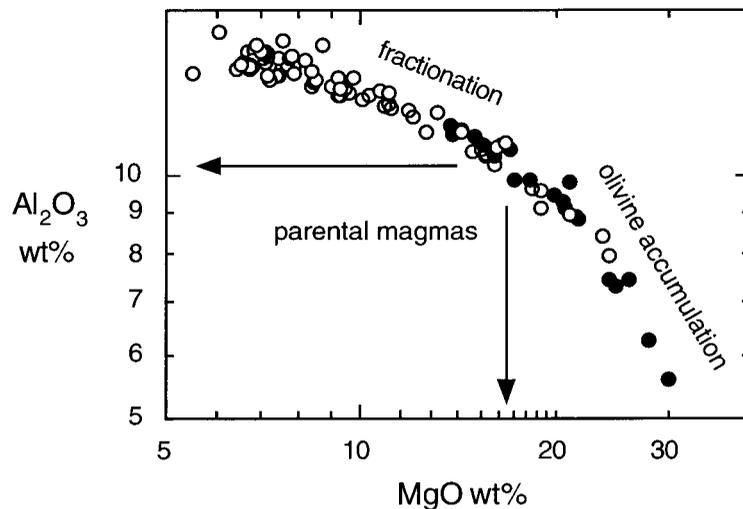


Fig. 4. Log–log plot of MgO vs. Al_2O_3 abundances in Hawaiian picrites (filled circles) and moderately evolved tholeiites from Mauna Loa and Kilauea (open circles). Lavas with less than 15–16% MgO fall along a linear trend indicating that they are related by fractional crystallization, whereas more magnesian picrites curve away from the fractionation trend, indicating accumulation of olivine. The break in slope between the fractionation and accumulation trends indicates parental melts with 15–16% MgO and 10% Al_2O_3 . Data from Table 2 and sources cited in the text.

Densities of 2.81 to 2.94 g/cm³ were calculated for the picrites using the modal and compositional data (Tables 1 and 2) and the glass chemistry ([17] and unpublished data) to estimate eruption temperatures [31]. In contrast, typical subaerial tholeiites have densities of 2.65 to 2.70 g/cm³ [32]. This density contrast between the picrites and the more evolved shield-building tholeiites shows that Hawaiian volcanoes are stratified, with the high-density picritic magmas erupting preferentially on the flanks of the volcanoes and more evolved, lower-density tholeiites erupting more frequently through the summit reservoir.

5. Incompatible trace elements: melting and mixing

5.1. Melting trends and plume source composition

The primitive nature of some Hawaiian picrites affords an opportunity to evaluate the melting processes and source characteristics responsible for generating these lavas. Here we present melting models that account for key incompatible trace element characteristics of these picrites. In order to approximate primary melt compositions, the measured abundances of incompatible trace elements (Table 2) were corrected to 16% MgO by adding or subtracting Fo₈₈ olivine. Ratios of incompatible trace elements also provide important constraints without needing to correct for olivine accumulation.

Incompatible trace element abundances and ratios that should be sensitive to variations in the melting process are well correlated among the picrites, and fall along calculated melting trends (Figs. 5–7). La concentrations, the shape of the REE pattern (La/Yb, Sm/Nd), and the Lu/Hf ratios of these picrites can be reproduced by 4–10% accumulated incremental melting [33] of a common source composed of 86.5% olivine + opx, 10% cpx, and 3.5% garnet, and with slightly depleted relative abundances of the most incompatible elements (Hawaiian pyrolite, Appendix A). The La/Yb, Sm/Nd, and Lu/Hf compositions of the picrites suggest increasing degrees of melting in the sequence: Loihi < Koolau < Kilauea < Mauna Kea < Kohala ≤ Hualalai ≤ Kahoolawe ≤ Mauna Loa (Fig. 5). Nb contents (normalized to 16% MgO) and Zr/Nb ratios of the picrites are com-

patible with this sequence except that the Koolau lavas have distinctively lower Nb concentrations and higher Zr/Nb ratios that fall off the melting arrays defined by the other picrites (Figs. 6 and 7). Although forward melting models are inherently non-unique and subject to significant uncertainties in distribution coefficients, conditions of melting, and style of melt aggregation, certain conclusions regarding the plume source appear to be robust.

It is well known that trace element characteristics of Hawaiian tholeiites require residual garnet in their source region [1,2,24,25,34,35], and this is also a feature of the melting model presented here. Although residual garnet is required, however, the source mineralogy implied by this model has a higher cpx/garnet ratio than typically found in garnet lherzolites. More garnet-rich sources do not reproduce the trends of the picrite compositions as they generate melts with more fractionated REE patterns (higher La/Yb, lower Sm/Nd) and lower Lu/Hf at a given degree of melting or La content (Figs. 5 and 6). This is not simply indicating a refractory garnet lherzolite source, as xenolith compositions and melting experiments show that cpx is the first phase to be exhausted during melting of garnet lherzolite, producing a residue with a low cpx/garnet [36,37]. Alternatively, the high cpx/garnet of the Hawaiian plume source inferred here may be indicating a polybaric melting regime which begins in the garnet field and continues to shallower depths near the base of the oceanic lithosphere above the plume (~70 km). The mineralogy of this Hawaiian pyrolite source can be modelled as a mixture of 65% moderately depleted spinel lherzolite (90% olivine + opx, 10% cpx) and 35% garnet lherzolite (80% olivine + opx, 10% cpx, 10% garnet), although the exact proportions depend somewhat on the distribution coefficients used in the melting calculation (Appendix A). A mixed spinel + garnet lherzolite source has been inferred previously from inversion of Hawaiian REE compositions [35,38] and from major element compositions of Hawaiian tholeiites [39]. Metasomatized spinel–garnet peridotites with mineral proportions similar to those of our inferred plume source have been reported from Malaita [40].

Despite their isotopic diversity [12,16], key incompatible trace element characteristics (La/Yb, Sm/Nd, Lu/Hf) of the picrites can be derived from

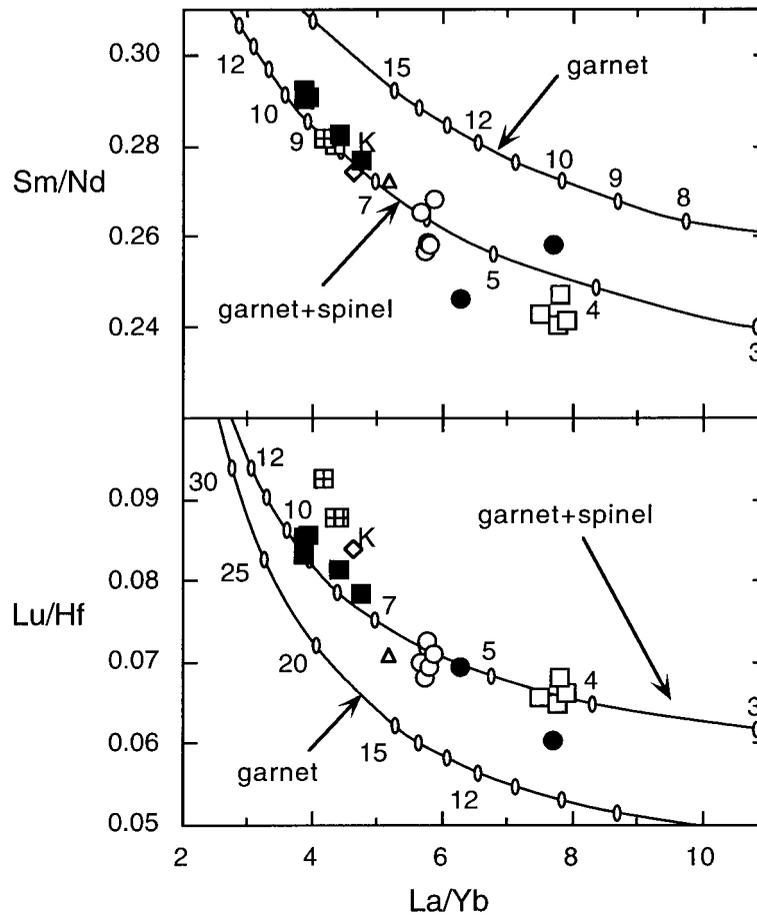


Fig. 5. La/Yb vs. Sm/Nd and Lu/Hf compositions of the Hawaiian picrites are closely matched by 4–10% accumulated incremental melting of a common source with a cpx/garnet ratio higher than that of typical garnet lherzolites. Such a source can be modelled as a mixture of spinel and garnet lherzolite (Appendix A). More garnet-rich sources produce melts that do not correspond to the picrite compositions. Numbers adjacent to knots on the melting trends indicate the degree of melting. Symbols indicate different volcanoes (see legend in Fig. 6).

a common source composition (Fig. 5). With the exception of the lavas from Koolau, La and Nb contents (normalized to 16% MgO), and Zr/Nb ratios of the picrites also fall along a single melting array (Figs. 6 and 7), with Kilauea and Loihi representing one extreme produced by relatively low degrees of melting, and Mauna Loa, Kahoolawe, and Hualalai at the other extreme, indicating higher degrees of melting. Incompatible trace element contents in the Hawaiian pyrolite source are similar to those of primitive mantle, but better fits to the picrite arrays are obtained if the source is slightly depleted in the most incompatible elements (e.g., La, Nb) relative to the primitive mantle (Appendix A). Although ratios of highly

incompatible elements such as Nb/U and Ba/Th clearly preclude pristine primitive mantle as a major source component in the Hawaiian plume [15], melting models for shield-building tholeiites frequently propose source compositions that mimic primitive mantle abundances for the REE and other incompatible trace elements [2,25,34,41,42]. The incipiently depleted composition of the Hawaiian pyrolite source can be modelled as the residue after removal of a small (0.25%) melt fraction from a primitive spinel lherzolite, although more complex processes such as metasomatic enrichment or refertilization of previously depleted mantle might produce a similar composition.

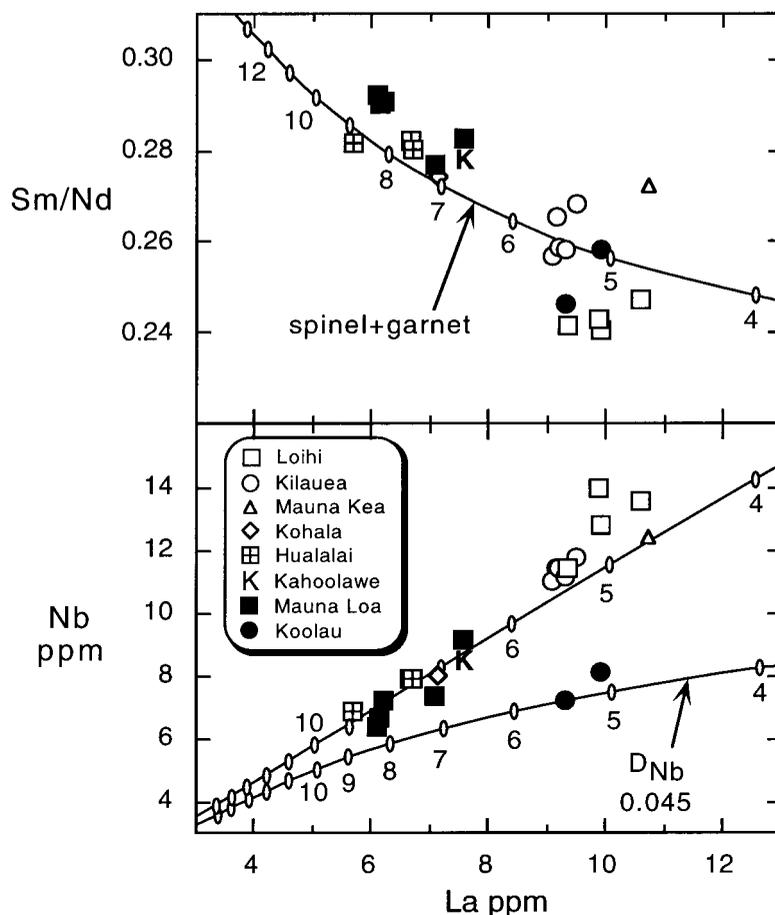


Fig. 6. La vs. Nb abundances (corrected to 16% MgO) and Sm/Nd ratios of the picrites form arrays that can be reproduced by melting of a Hawaiian pyrolite source (Appendix A). The lower Nb contents of Koolau picrites can be generated by arbitrarily increasing the bulk distribution coefficient to 0.045, possibly indicating a distinctive mineralogy in the Koolau plume component.

Picrites from Koolau, have lower Fe, Ti, Ca, and Nb contents at a given MgO, resulting in low Ca/Al and elevated Zr/Nb ratios (also La/Nb, Sr/Nb, etc.) (Table 2). Nb contents and Zr/Nb ratios of the Koolau picrites fall off the melting arrays defined by the other picrites (Figs. 6 and 7). These are general characteristics of shield lavas from Koolau [11], and must be indicating a distinctive source composition, as also attested by their isotopic compositions [12,13,19]. The higher Zr/Nb and lower Ca/Al ratios of Koolau tholeiites may be indicating a more depleted source composition, although this is not evident in their La/Yb and Sm/Nd ratios (Fig. 5). Alternatively, Koolau tholeiites may have experienced more extensive reaction with harzbur-

gite [42], or the Koolau plume component may have a unique mineralogy, resulting in higher bulk distribution coefficients for Nb and Ti (Figs. 6 and 7). The characteristics of the Koolau plume component must be capable of generating the unique combination of major element, trace element, and isotopic compositions of Koolau tholeiites.

5.2. Source component mixing correlates with degree of melting

Trace element and radiogenic isotopic compositions of Hawaiian tholeiites show that the plume contains at least two and probably three mantle components with distinct long-term histories [12–

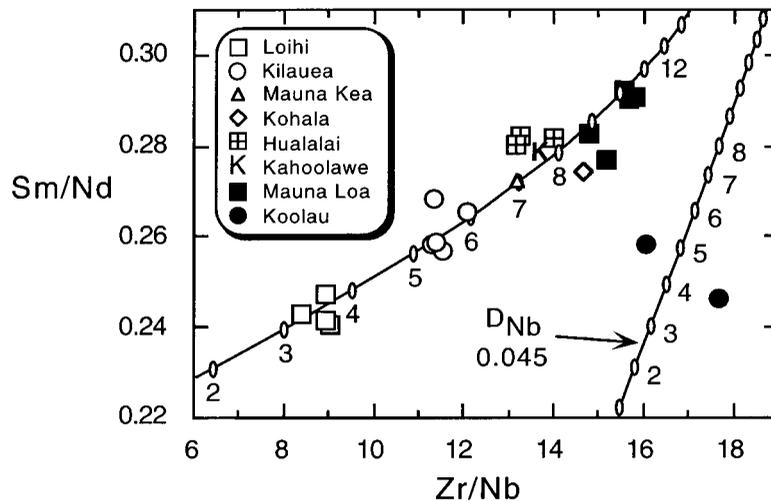


Fig. 7. Zr/Nb vs. Sm/Nd compositions of all of the picrites fall along a single melting array except for those from Koolau. The higher Zr/Nb ratio of Koolau picrites reflects their low Nb contents, which may be indicating a more depleted source component or a higher bulk distribution coefficient for Nb.

14,19,20,43]. The picrites studied here span the range of isotopic compositions in Hawaiian tholeiites [12]. Tholeiites from Kilauea, Loihi, and Koolau represent compositional end-members, while those from Mauna Loa, Hualalai, and Kohala have intermediate characteristics. Spatial and temporal variations in the relative proportions of these components in Hawaiian magmas may reflect large-scale plume geometry [2,44], but this remains speculative, and the origins of the diverse isotopic components in Hawaiian magmas are controversial [12–14,16]. Certain major element and trace element ratios of the picrites that are insensitive to olivine accumulation (e.g., Ca/Al, Zr/Nb) correlate with isotopic compositions, raising the possibility that at least some of the major element and trace element characteristics of Hawaiian tholeiites reflect long-term characteristics of source components in the plume [11,13].

The melting models presented here suggest that those picrites representing isotopic end-members (Loihi, Koolau, Kilauea) formed by smaller degrees of melting than those with intermediate compositions (Kohala, Hualalai, Kahoolawe and Mauna Loa) (Figs. 5–7). These relations suggest that at least some of the correlations between major element, trace element, and isotopic compositions in Hawaiian tholeiites may be controlled by variations in the degree of melting within the plume, such that

smaller melt fractions preferentially sample the compositional end-members, whereas greater extents of melting sample larger volumes of mantle and blend the diverse source components together more effectively.

6. Recycled oceanic crust in the Hawaiian plume?

Ocean island basalts have isotopic compositions that require an ‘enriched’ source component relative to MORB, and this enrichment is canonically ascribed to recycling of continental or oceanic crustal material into the OIB source regions [44]. Three ‘enriched’ mantle components have been recognized in OIB suites (EM1, EM2, HIMU) [45]. EM2 and HIMU components are characterized by isotopic compositions indicating long-term enrichments of Rb/Sr and U/Pb, respectively, which are reasonably explained by crustal recycling [14,45]. EM1, on the other hand, is characterized by Sr, Nd, and Pb isotopic compositions similar to those expected for primitive mantle, and for this component the case for crustal recycling is not so clear [12,45]. Hawaiian basalts trend toward the EM1 composition, which is best expressed at Koolau, Lanai, and Kahoolawe [11–13,19,20]. Various forms of crustal recycling have been invoked to explain the range of major

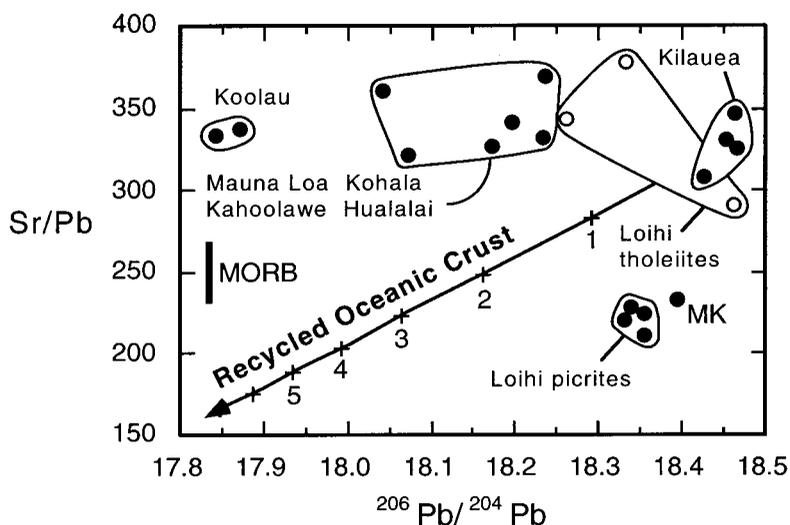


Fig. 8. Recycling 2 Ga oceanic crust (MORB + 10% pelagic sediments) into the Hawaiian plume should produce a strong correlation between Sr/Pb and Pb isotopic compositions, in contrast to the nearly constant Sr/Pb ratios actually observed in primitive Hawaiian lavas. The low Sr/Pb ratios of Loihi picrites (filled circles) are due to the unusually high Pb contents of these picrites compared to other Loihi tholeiites (open circles), possibly reflecting assimilation of a hydrothermal fluid into the picrite parental magma [23]. The low Sr/Pb of the Mauna Kea picrite (MK-1-6) probably reflects cryptic plagioclase fractionation [8]. Black bar indicates the Sr/Pb of MORB. Ticks indicate the fraction of recycled oceanic crust. Model compositions [12,14,61]: Recycled Oceanic Crust = 90% 2 Ga altered MORB plus 10% pelagic sediment. 2 Ga MORB: $^{206}\text{Pb}/^{204}\text{Pb}$ 22.7, Pb 0.25 ppm, $^{87}\text{Sr}/^{86}\text{Sr}$ 0.7029, Sr 95 ppm. Pelagic sediment: $^{206}\text{Pb}/^{204}\text{Pb}$ 16.7, Pb 15 ppm, $^{87}\text{Sr}/^{86}\text{Sr}$ 0.7338, Sr 330 ppm. Kilauea plume source: $^{206}\text{Pb}/^{204}\text{Pb}$ 18.48, Pb 0.075 ppm, $^{87}\text{Sr}/^{86}\text{Sr}$ 0.7035, Sr 25 ppm.

element, trace element, and isotopic compositions of Hawaiian tholeiites, including bulk mixing of plume-type mantle with 2 Ga oceanic crust [14], generation of silicic melts from eclogite pods within the plume [13], and recycling of oceanic gabbros [15]. Here we test these models against the compositions of the Hawaiian picrites.

Perhaps the strongest evidence favoring recycled oceanic crust in the Hawaiian plume is the radiogenic $^{187}\text{Os}/^{188}\text{Os}$ isotopic compositions of Hawaiian basalts [12,14]. Mixing ~25% of 2 Ga oceanic crust (90% basalt +10% pelagic sediment) into the mantle is necessary to account for the trend toward radiogenic ^{187}Os and unradiogenic Pb isotopic compositions that characterizes Hawaiian tholeiites [12,14]. Although such a model can account for the end-member isotopic characteristics of lavas from Koolau, it has difficulty reproducing the well-defined linear correlations between $^{187}\text{Os}/^{188}\text{Os}$ and Pb isotopic compositions of these picrites [12]. This model also predicts much lower Sr/Pb concentration ratios than are observed in the Hawaiian picrites, and a

strong correlation between Sr/Pb ratios and Pb isotopic compositions that contrasts with the nearly constant Sr/Pb ratios observed in the picrites (Fig. 8). In contrast to the low Sr/Pb ratios predicted by this model, primitive Hawaiian tholeiites actually have Sr/Pb ratios higher than typical N-MORB (Fig. 8). Note that the low Sr/Pb ratios of the Mauna Kea picrite (MK-1-6) probably reflects loss of Sr due to cryptic plagioclase fractionation [8], whereas the Loihi picrites are enriched in Pb (Table 2), possibly due to a contribution from a hydrothermal fluid [23].

Mixing of recycled oceanic crust into the mantle also encounters problems accounting for the major element compositions of Hawaiian lavas. Adding up to 25% of a basaltic crustal component to a depleted mantle similar in bulk composition to the MORB source region would produce a hyperfertile mantle source rich in garnet and cpx, which would generate silica-undersaturated basalts with high FeO and TiO_2 contents at a given MgO [46]. These characteristics are opposite to those of Koolau tholeiites, which have quartz-normative compositions, and low FeO

and TiO_2 contents relative to basalts from Kilauea and Loihi [11,13].

As an alternative to bulk mixing of recycled oceanic crust into the mantle, Hauri [13] proposed that the range of major element, trace element, and isotopic compositions observed in Hawaiian tholeiites could be explained by mixing between dacitic and basaltic melts, with the dacite being derived from recycled eclogite or pyroxenite within the plume. Addition of up to 20% of the dacitic melt with basaltic magmas similar to those from Loihi was proposed to explain the compositional characteristics of Koolau tholeiites [13]. Given the possible modifications of recycled oceanic crust during subduction and storage in the deep mantle, such a process is difficult to model directly, but insights into possible compositional characteristics of silicic melts produced from eclogite pods within the plume might be obtained from adakites, which are thought to represent melts of subducted slabs, and eclogite-derived tonalites [47,48].

Adakites and many tonalites are characterized by high Sr/Y, reflecting the presence of residual garnet in their eclogitic source [47,48]. Adding only a few percent of such a melt to a magma similar in composition to the Loihi or Kilauea picrites should produce dramatically higher Sr/Y ratios, in contrast to the nearly constant Sr/Y ratios of the picrites (Fig. 9).

The slightly elevated Sr/Y ratios of the Koolau picrites might be explained by addition of $\leq 10\%$ of an eclogite-derived tonalitic or dacitic melt component, but this cannot be a general process in the Hawaiian plume (Fig. 9). The fact that the REE and HFSE compositions of most of these picrites can be reproduced by variable degrees of melting of a common source (Figs. 5–7) also argues against mixing of compositionally disparate melts as the principal process controlling the compositional characteristics of these lavas.

Hofmann and Jochum [15] cited positive Sr anomalies, and depletions of Th and U relative to Ba and La in Hawaiian lavas as evidence that the plume contains a significant amount of plagioclase-enriched oceanic gabbro. Ba/Th ratios of Hawaiian tholeiites are greater than those of MORB or the primitive mantle, requiring a distinctive source composition in the Hawaiian plume [15]. Most of the picrites have nearly constant Ba/Th, with some variation in the subaerial lavas from Koolau, Kahoolawe, and Mauna Loa (Fig. 10). KOO-17a has a low Ba content and a low Ba/Th reflecting alkali loss during subaerial weathering (Table 2). In contrast, KOO-CF, KW-25, and the 1868 picrite from Mauna Loa all have somewhat higher Ba/Th compared to the other picrites, with KW-25 having an extreme value (Ba/Th = 320). The reason for these high Ba/Th values is

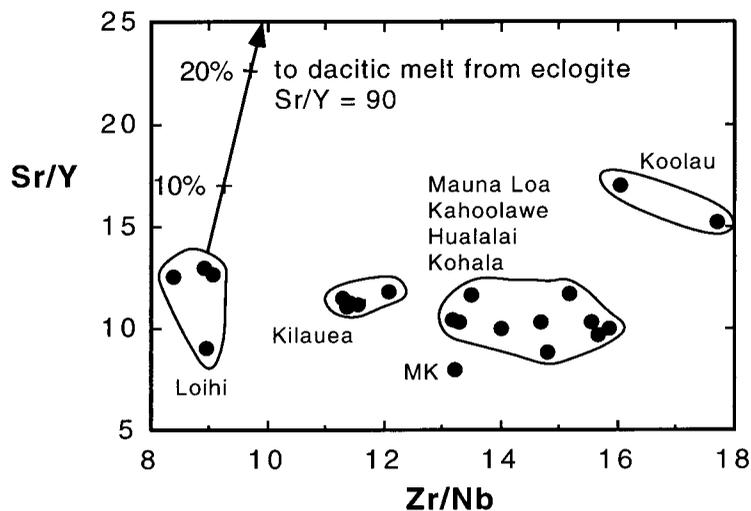


Fig. 9. Dacitic melts derived from eclogite pods within the Hawaiian plume would have high Sr/Y ratios, possibly analogous to the compositions of adakites and tonalites [46,47]. Even a few percent of such a component in Hawaiian magmas should lead to dramatically higher Sr/Y ratios, in contrast to the nearly constant ratios actually observed in these lavas.

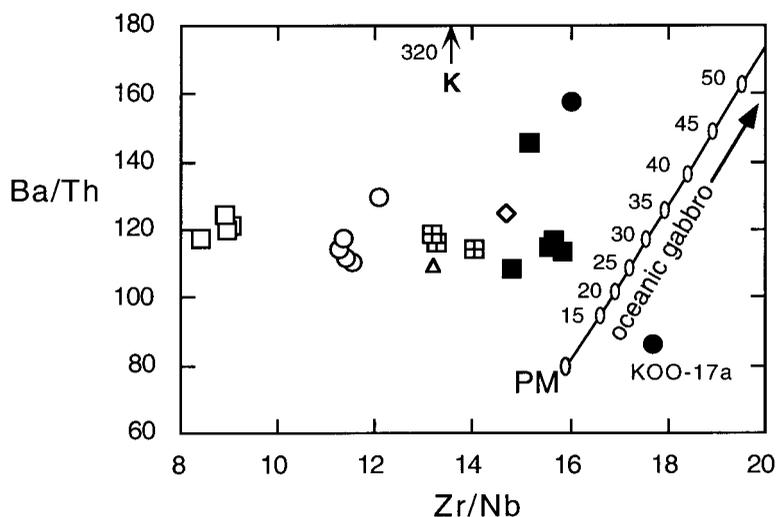


Fig. 10. Ba/Th ratios of Hawaiian lavas should be sensitive to the proportion of plagioclase-enriched oceanic lower crust in the plume source. The curve indicates mixing between a primitive mantle composition and average oceanic gabbro as described in the text. The limited variation of Ba/Th in the Hawaiian picrites would imply similar amounts of lower crustal gabbros in the various plume components, making it difficult to generate the observed isotopic diversity of these lavas. Legend same as Fig. 6.

unclear, but there does not appear to be a simple correlation of Ba/Th with indicators of source characteristics such as Pb isotopic composition or Zr/Nb in Hawaiian lavas (Fig. 10). Mafic cumulates from the Oman ophiolite show that the oceanic lower crust is likely to have high Ba/Th ratios (average 8.5 ppm Ba, 0.007 Th, Ba/Th = 1177 [49]), but the small range of Ba/Th ratios in Hawaiian lavas would imply similar amounts of this material in the various source components of the plume (Fig. 10), which is inconsistent with their diverse isotopic compositions. Therefore, the range of isotopic and trace element compositions in Hawaiian shield-building lavas is unlikely to result from variable amounts of recycled oceanic lower crust in the plume.

7. Source components in the Hawaiian plume

Despite their isotopic diversity, key incompatible trace element characteristics of the picrites can be reproduced by variable degrees of melting of a common source, suggesting that the different mantle source components in the Hawaiian plume have generally similar trace element compositions. This reinforces previous conclusions based on Os–Pb isotopic compositions that the isotopically diverse plume

components have similar Os/Pb concentration ratios [12]. The origins of these plume components are not well understood. Although recycling of oceanic crust can explain some of the isotopic characteristics of Hawaiian lavas, bulk mixing of oceanic crust into the mantle encounters significant difficulties in explaining the compositions of primitive Hawaiian tholeiites (Figs. 8–10). Alternatively, the Hawaiian plume may have formed through enrichment or refertilization of previously depleted mantle, possibly by small-degree melts. Such melts could dominate the trace element budget of the plume source without disturbing the major element compositions. Trace element characteristics that mimic those of primitive mantle, and decoupling of isotopic and trace element compositions are frequently observed in metasomatized continental lithospheric mantle [50], and similar processes may be equally important in the oceanic lithosphere although considerably fewer data are available [1,40,51]. A link between the EM1 mantle component and metasomatized continental lithosphere has been suggested previously [45,52], although it is unclear whether this represents accretion of plumes beneath the continents, or entrainment of continental lithosphere into plume source regions.

The main objection to a lithospheric mantle component in the Hawaiian plume has been the radio-

genic $^{187}\text{Os}/^{188}\text{Os}$ isotopic compositions of Hawaiian basalts [12,14]. Refractory lithospheric mantle generally has a low Re/Os, producing subchondritic Os isotopic compositions that contrast with the radiogenic compositions of Hawaiian tholeiites. This problem may have been resolved with the discovery of coupled enrichments in ^{187}Os and ^{186}Os in Hawaiian basalts that are consistent with addition of a small amount of fractionated outer core to the plume source [16]. As an alternative to recycling oceanic crust into the mantle, the Hawaiian plume may contain a component of refertilized or metasomatized lithosphere (continental or oceanic) that has been entrained into the deep lower mantle. Enrichment of previously depleted mantle, possibly by small-degree melts, may have occurred either within the lithosphere prior to entrainment, or through melting of physically associated slabs or entrained pyroxenites, and reaction of these melts with adjacent peridotite [53]. Ringwood [54] envisioned such a process as an essential step in the creation of plume source regions. In this scenario, compositional characteristics of mafic layers in the mantle may be transferred to the surrounding peridotite via melt–rock reactions without bulk mixing.

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cantly improved the manuscript. This is GEMOC publication 152 and SOEST contribution 4751. [CLJ]

Appendix A

Mantle compositions and distribution coefficients for the melting model discussed in the text [37,55–57]. The Hawaiian pyrolite source composition (columns 2 and 8) provides the best fit to the melting arrays of the picrites.

	1	2	3	4	5	6	7	8
Nb	0.74	0.59	0.005	0.050	0.050	0.0095	0.0140	0.0111
La	0.70	0.50	0.001	0.055	0.020	0.0064	0.0083	0.0071
Nd	1.36	1.23	0.007	0.18	0.50	0.0243	0.0736	0.0416
Zr	11.8	10.9	0.015	0.16	0.85	0.0295	0.1130	0.0587
Hf	0.31	0.29	0.015	0.30	0.80	0.0435	0.1220	0.0710
Sm	0.44	0.42	0.010	0.33	0.70	0.0420	0.1110	0.0662
Yb	0.49	0.47	0.030	0.45	6.5	0.0720	0.7190	0.2985
Lu	0.073	0.071	0.050	0.42	7.0	0.0870	0.7820	0.3303

1 = primitive mantle composition (ppm); 2 = Hawaiian pyrolite composition (ppm); 3 = distribution coefficients for combined olivine + opx; 4 = distribution coefficients for cpx; 5 = distribution coefficients for garnet; 6 = bulk distribution coefficients for spinel lherzolite (90% olivine + opx, 10% cpx); 7 = bulk distribution coefficients for garnet lherzolite (80% olivine + opx, 10% cpx, 10% garnet); 8 = bulk distribution coefficients for Hawaiian pyrolite (86.5% olivine + opx, 10% cpx, 3.5% garnet).

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