

# Petrogenesis of Dunite Xenoliths from Koolau Volcano, Oahu, Hawaii: Implications for Hawaiian Volcanism

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## ABSTRACT

Ultramafic xenoliths from Koolau Volcano on the island of Oahu, Hawaii, are divided into spinel lherzolite, pyroxenite, and dunite suites. On the basis of a study of the petrography and mineral compositions of 43 spinel lherzolites, 12 pyroxenites, and 20 dunites, the following characteristics of the dunites in relation to the other nodule types and to Hawaiian lavas emerge. (1) The forsterite content of olivines in the Koolau dunites ( $FO_{82.6}$ – $FO_{89.7}$ ) overlap those of Hawaiian tholeiitic and alkalic lavas and are generally lower than those in abyssal lherzolites and dunites and in Koolau spinel lherzolites. (2) Most of the dunites contain no orthopyroxene, all except two contain chrome spinel, and a few contain interstitial plagioclase and clinopyroxene. (3) Chrome spinels from the Koolau dunites are distinctly higher in  $Cr/(Cr + Al)$ , lower in  $Mg/(Mg + Fe^{2+})$ , and higher in  $TiO_2$  than those from abyssal basalts and peridotites. Chrome spinels in the dunites correspond closely in composition to chrome spinels in Hawaiian tholeiitic and alkalic lavas. (4) The abundance of dunite relative to other nodule types decreases outward from the central part of the volcano. The dunites are interpreted not as residues of partial fusion of the mantle but as crystal accumulations stored at shallow depths beneath the central part of Koolau Volcano and derived from picritic magmas parental to the shield-building tholeiitic lavas.

## INTRODUCTION

A generalized sequence of events for Hawaiian volcanoes consists of (1) a main shield-building stage in which voluminous tholeiitic lavas are produced, (2) a mature stage characterized by alkalic lavas and less frequent eruptions, (3) a long period of erosion, and (4) a final stage of activity marked by 'post-erosional' nephelinitic as well as alkalic lavas. Also, recent evidence suggests that an initial alkalic stage of activity precedes the main shield-building tholeiitic stage (Moore *et al.*, 1982; Frey & Clague, 1983). Many volcanoes do not display all of these stages, but the relative time sequence of the various stages is consistent throughout the Hawaiian Islands.

Koolau Volcano on the eastern part of the island of Oahu displays lavas associated with the main tholeiitic stage (Koolau and Kailua Volcanics) and the final nephelinitic stage (Honolulu Volcanics). Lavas associated with the mature alkalic stage have not been found (Stearns & Vaksvik, 1935, 1938; Jackson & Wright, 1970). Vents of the Honolulu Volcanics are scattered widely over the shield and many of them contain ultramafic xenoliths. Thus, xenoliths from the Koolau shield present an excellent opportunity to characterize the

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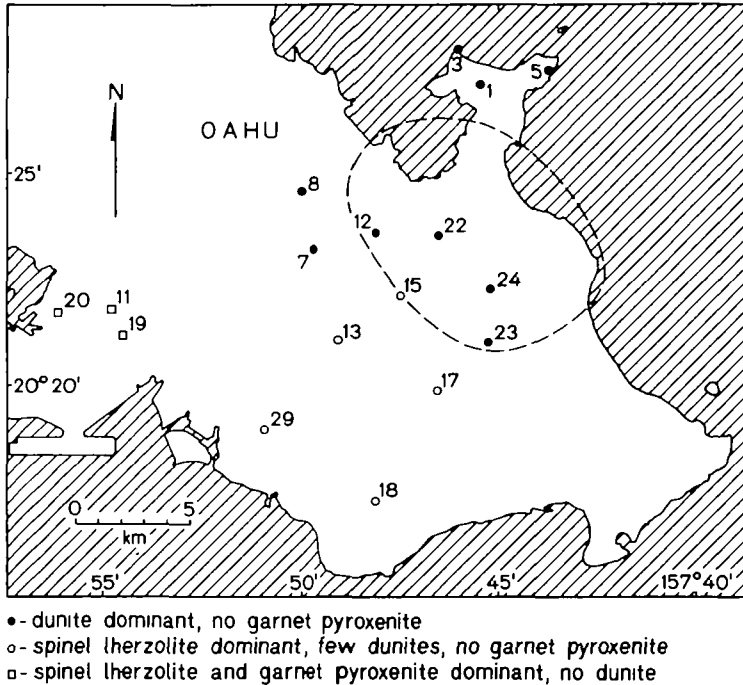


FIG. 1. Outline of eastern Oahu showing xenolith-bearing vents of Honolulu Volcanics (see Table 1 for names of vents). Vent locations are after Stearns (1939). The dashed line outlines the approximate location of the Koolau summit caldera (after Macdonald & Abbott, 1970, fig. 174). See text for a discussion of one dunite xenolith from Salt Lake that is considered to be genetically related to the spinel lherzolite suite.

post-eruptive oceanic lithosphere beneath a Hawaiian volcano (Sen, 1983). Jackson & Wright (1970) showed that the Koolau xenoliths can be divided into dunite, spinel lherzolite, and pyroxenite (including websterite, garnet clinopyroxenite and garnet peridotite) suites and are concentrically zoned by rock type around the central caldera. However, the lava hosts for the xenoliths are not concentrically zoned by composition (Clague & Frey, 1982). Only very limited information has been published on mineral compositions in these xenoliths. This paper is the first of two concerning the petrography, mineral compositions, and petrogenesis of the Koolau xenoliths. In this paper we will be concerned mainly with the dunites. For purposes of comparison, we will also show mineral composition data for the lherzolite and pyroxenite xenoliths, but detailed consideration of these other two suites is left for the second paper.

#### SAMPLE LOCATIONS

Fig. 1 shows locations of the 17 vents and vent groups on the Koolau Volcano that contain ultramafic xenoliths. We have obtained mineral composition data on dunites from six of these vents, Hawaiiiloa, the Pali group, Kaau, Nuuanu (Luakaha), Kalihi, and Ulupau. Table 1 identifies the vents numbered on Fig. 1. Two numbering systems are currently in use for the vents of the Honolulu Volcanics; we follow the one used by Macdonald & Abbott (1970).

## ANALYTICAL METHOD

All analyses were carried out with an automated electron microprobe (ARL model EMX-SM with a Tracor-Northern TN-2000 automation system) at the University of Texas at Dallas. Analytical conditions were 15 kv accelerating potential, 0.15  $\mu$ A beam current, and about 1  $\mu$ m beam diameter. Olivines were analysed against an olivine standard (for Mg, Fe and Si), pyroxene standard (for Ca and Ti), and a garnet standard (for Mn). The wavelength-dispersive spectrometric (WDS) method was used for all olivine analyses. Other phases were analysed against MgO, Al<sub>2</sub>O<sub>3</sub>, TiO<sub>2</sub>, chromite (for Fe), MgCr<sub>2</sub>O<sub>4</sub> (for Cr), wollastonite (for Ca and Si), orthoclase (for K) and Tiburon albite (for Na) standards. Most analyses were done by the WDS method, and the remainder by a combined WDS and energy dispersive method. The WDS method was always used for Fe and Na. Matrix corrections were carried out by the Bence-Albee procedure.

## PETROGRAPHY

In hand specimen, the dunite xenoliths are generally more angular than the lherzolite and pyroxenite suite xenoliths. The dunites have a light green (unaltered olivine) to dirty yellow (partially altered olivine) colour in hand specimen.

All but two of the dunite xenoliths (77PA-11 and 67NUUA-89-90) contain chrome spinel (Table 2). These two samples may contain spinel that was missed in the particular sections cut. The dunite spinels are much darker than the brown spinels of the spinel lherzolites and the green spinels of the pyroxenites. The size and shape of spinels are highly variable. Commonly, they occur interstitially and have amoeboid to subhedral rhomboid shapes. However, euhedral to subhedral spinels also occur as poikilitic inclusions in olivine. Modal abundances of these spinels range from only trace amounts to about 4 per cent.

Olivine in the dunites ranges from small (0.025  $\times$  0.02 mm) equant recrystallized grains to medium grained (1.5  $\times$  1 mm) anhedral porphyroclasts. The dunites typically are highly deformed (see also Jackson & Wright, 1970), and the overall texture varies from porphyroclastic to allotriomorphic granular (Pike & Schwarzmann, 1977). Well-defined triple-point junctions are present in a few places. Deformation lamellae are very common and are more conspicuously developed in the porphyroclastic rocks than in the equigranular rocks. Trains of CO<sub>2</sub> bubbles (Roedder, 1965) occur along healed fracture surfaces in olivine porphyroclasts, a feature most common in the Kalihi dunites. Most recrystallized olivine grains are devoid of CO<sub>2</sub> bubbles.

Interstitial material composed of clinopyroxene, orthopyroxene, plagioclase, and glass is common in the Kalihi dunites, but in dunites from the other vents, the only interstitial mineral other than spinel is clinopyroxene. One exception is 66ULUP-1 which contains amphibole. The Kalihi dunites are notably permeated by glass from the enclosing basalt. The total amount of all minerals other than olivine in the dunites is always less than 5 per cent.

## MINERAL COMPOSITIONS

*Olivine*

Olivine compositions in the dunites range from Fo<sub>82.6</sub> to Fo<sub>89.7</sub> (Table 3 and Fig. 2). Jackson & Wright (1970) used the term 'two-olivine dunite' to refer to the textural heterogeneity of many of these deformed dunites. We have found no compositional difference between the two textural types of olivine. Dunite olivines from the Kalihi vent are slightly more iron-rich (Fo<sub>82.6</sub> to Fo<sub>85</sub>) than those from the other vents (Fo<sub>83.4</sub> to Fo<sub>89.7</sub>).

TABLE 1  
*Honolulu Series xenolith-bearing vents (vent numbers after Macdonald & Abbott, 1970, table 18)*

Vent	Name	Vent	Name
1	Hawaiiiloa	17	Kaau
3	Pyramid Rock	18	Mauumae
5	Ulupau	19	Salt Lake group
7	Kalihi*	20	Makalapa
8	Haiku	22	Castle
11	Aliamanu group	23	Maunawili
12	Kaneohe group	24	Training School
13	Nuuanu (Luakaha)	29	Punchbowl
15	Pali group		

\* Numbers for Kalihi and Kamaikai (vent 28) were interchanged in fig. 250 of Macdonald & Abbott (1970).

TABLE 2  
*Mineralogy of studied dunites*

	<i>ol</i>	<i>opx</i>	<i>cpx</i>	<i>sp</i>	<i>plag</i>	<i>amph</i>
Hawaiiiloa (vent 1)						
69KANS-1 (114929/1) <sup>a</sup>	× (83.4) <sup>b</sup>	—	—	×	—	—
Ulupau (vent 5)						
66ULUP-23 (114753/9)	× (84.7)	—	—	×	—	—
66ULUP-1 (114753/1)	× (84.6)	—	—	×	—	×
70ULUP-2 (114953/2)	× (89.7)	—	—	p	—	—
77ULUP-1	× (87.2)	—	—	×	—	—
77ULUP-2	× (87.2)	—	—	×	—	—
77ULUP-3	× (86.8)	—	—	×	—	—
77ULUP-4	× (84.4)	—	×	×	—	—
Kalihi (vent 7)						
77KALI-1	× (82.6)	×	×	×	×	—
77KALI-1D	× (85.0)	—	—	×	—	—
77KALI-2	× (82.6)	—	p	×	—	—
77KALI-3	× (82.7)	—	×	×	×	—
77KALI-4	× (84.7)	—	×	×	—	—
77KALI-8	× (82.6)	—	×	×	×	—
77KALI-9	× (84.9)	—	p	×	—	—
Nuuanu (vent 13)						
67NUUA-89-90 (114765/17)	× (85.0)	p	—	—	—	—
Pali (vent group 15)						
77PA2-11	× (85.4)	—	—	—	—	—
77PA1-42D	× (87.9)	—	p	×	—	—
Kaau (vent 17)						
77KAPS-13	× (87.0)	—	p	×	—	—
77KAPS-27	× (89.4)	—	—	×	—	—

\* All specimens with the prefix '77' are from the collection of D. C. Presnall at the University of Texas at Dallas. Five specimens are from the collection of E. D. Jackson, now housed at the Smithsonian Institution, Washington, D.C. These samples are designated by the number assigned by Jackson, with the corresponding Smithsonian number in parentheses.

<sup>b</sup> × = mineral analysed, p = mineral present but not analysed. Numbers in parentheses are mole per cent forsterite. For all samples, the modal proportion of olivine is greater than 95 per cent.

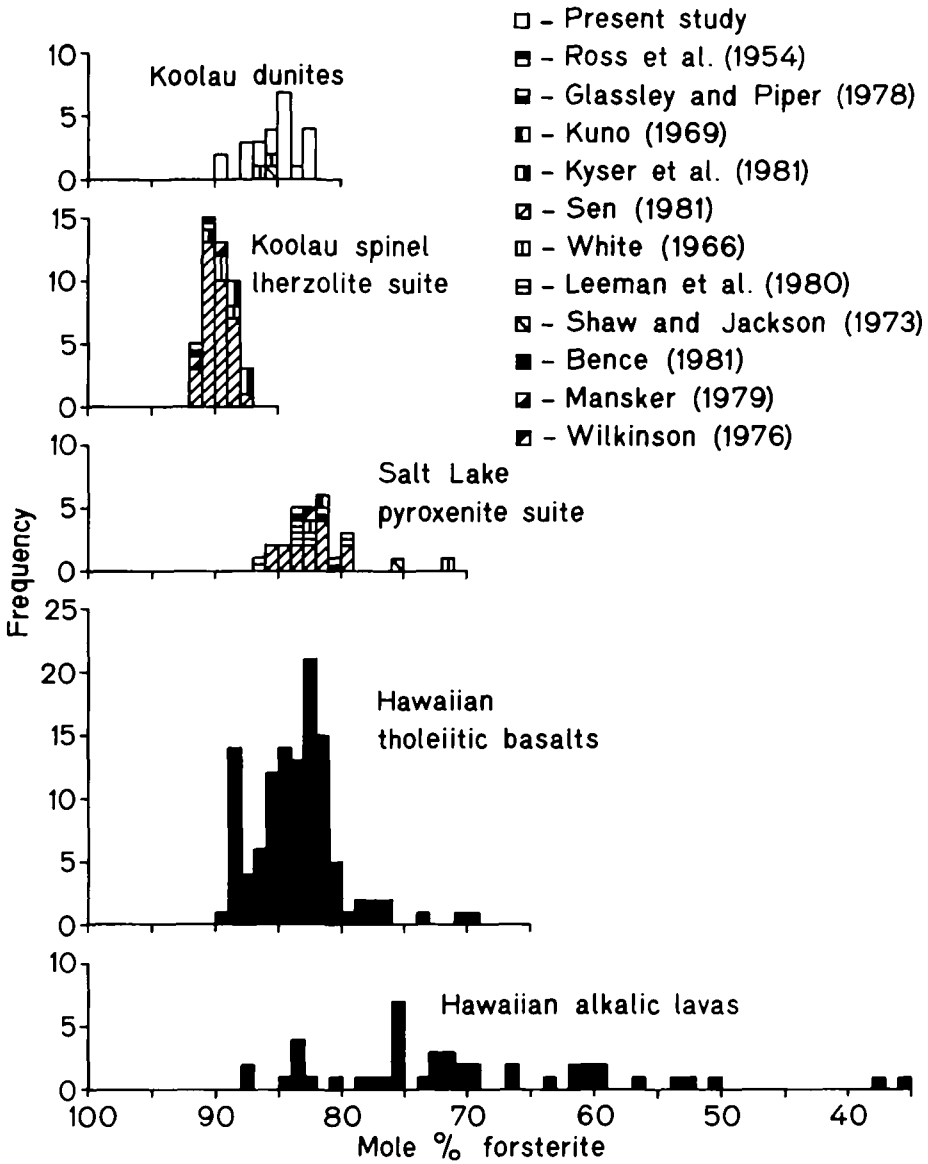


FIG. 2. Histograms of olivine compositions from various Hawaiian rock suites.

*Spinel*

Spinel compositions (Table 4 and Figs. 3 and 4) are heterogeneous even in a single thin section. This variation indicates that at least some spinels are out of equilibrium with other minerals in the rock. Disequilibrium in spinel compositions in single rock specimens has also been noted by many others (for example Roeder *et al.*, 1979; Henry & Medaris, 1980). The  $Fe_2O_3$  values were calculated from an assumed  $R^{2+}R_2^{3+}O_4$  formula after calculating  $TiO_2$  and  $SiO_2$  as  $Fe^{2+}R^{4+}O_4$ .

TABLE 3  
Olivine compositions

No. of grains <sup>b</sup>	Hawaiiia (1)			Ulupau (5)					Kalihī (7)	
	69KANS-1	66ULUP-23	66ULUP-1	70ULUP-2	77ULUP-1	77ULUP-2	77ULUP-3	77ULUP-4	77KALI-1	77KALI-ID
	5	5	3	5	1 <sup>c</sup>	1 <sup>c</sup>	1 <sup>c</sup>	1 <sup>c</sup>	5	5
SiO <sub>2</sub>	38.95	40.43	40.03	40.43	41.17	40.54	41.18	41.30	39.14	40.38
TiO <sub>2</sub>	n.d.	n.d.	0.00	0.00	n.d.	n.d.	n.d.	n.d.	0.00	0.00
Cr <sub>2</sub> O <sub>3</sub>	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.05	n.d.
FeO <sup>r</sup>	15.67	14.67	14.73	9.84	12.31	12.60	12.76	14.98	16.14	14.26
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	44.43	45.63	45.61	48.86	47.32	48.14	47.26	45.55	43.77	45.77
CaO	0.05	0.00	0.04	0.10	0.07	0.00	0.05	0.04	0.20	0.13
Total	99.10	100.73	100.41	99.23	100.87	101.28	101.25	101.87	99.30	100.54
	Cations per 4 oxygens									
Si	0.990	1.004	0.998	0.998	1.009	0.993	1.007	1.013	0.995	1.003
Ti	—	—	—	—	—	—	—	—	—	—
Cr	—	—	—	—	—	—	—	—	0.001	—
Fe	0.333	0.305	0.307	0.203	0.252	0.258	0.261	0.307	0.343	0.296
Mn	—	—	—	—	—	—	—	—	—	—
Mg	1.684	1.688	1.695	1.798	1.728	1.757	1.723	1.666	1.659	1.695
Ca	0.001	—	0.001	0.003	0.002	0.000	0.001	0.001	0.005	0.003
Total	3.008	2.997	3.001	3.002	2.991	3.007	2.993	2.987	3.003	2.997
Mole % Fo	83.4	84.7	84.6	89.7	87.2	87.2	86.8	84.4	82.6	85.0

	Kalihi (7)					Nuuanu (13)	Pali (15)		Kaau (17)	
	77KALI-2	77KALI-3	77KALI-4	77KALI-8	77KALI-9	67NUUA-89-90 <sup>a</sup>	77PA2-11	77PA1-42D	77KAPS-13	77KAPS-27
No. of grains <sup>b</sup>	5	5	5	5	5	1	5	5	5	5
SiO <sub>2</sub>	39.56	39.84	39.96	39.24	40.29	40.41	39.05	40.68	39.64	40.11
TiO <sub>2</sub>	0.00	0.02	0.00	0.00	0.00	n.d.	0.04	0.00	0.08	0.03
Cr <sub>2</sub> O <sub>3</sub>	n.d.	0.03	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
FeO <sup>c</sup>	16.46	16.34	14.43	16.45	14.34	14.25	14.05	11.54	12.13	10.33
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.13	n.d.	0.08	0.04
MgO	43.87	44.50	45.48	43.71	45.76	45.36	46.03	47.70	47.40	48.94
CaO	0.00	0.20	0.15	0.00	0.10	0.05	0.00	0.10	0.43	0.05
Total	99.89	100.93	100.02	99.40	100.49	100.07	99.30	100.02	99.76	99.50
<i>Cations per 4 oxygens</i>										
Si	1.000	0.996	0.999	0.997	1.002	1.008	0.985	1.003	0.986	0.991
Ti	—	0.001	—	—	—	—	0.001	—	0.001	0.001
Cr	—	0.001	—	—	—	—	—	—	—	—
Fe	0.348	0.342	0.302	0.350	0.298	0.297	0.296	0.238	0.252	0.213
Mn	—	—	—	—	—	—	0.003	—	0.002	0.001
Mg	1.653	1.659	1.696	1.656	1.696	1.686	1.730	1.753	1.758	1.802
Ca	—	0.005	0.004	—	0.003	0.001	—	0.003	0.011	0.001
Total	3.001	3.004	3.001	3.003	2.999	2.992	3.015	2.997	3.010	3.009
Mole % Fo	82.6	82.7	84.7	82.6	84.9	85.0	85.4	87.9	87.0	89.4

<sup>a</sup> This sample consists of two physically distinct xenoliths (dunite and harzburgite) in the same host nephelinite sample. The composition listed here is for the olivine in the dunite xenolith.

<sup>b</sup> The number of spot analyses averaged for each column is 3 analyses per grain (core, rim, intermediate) times the number of grains, except where noted. Compositional zoning was not found.

<sup>c</sup> Single spot analysis.

<sup>d</sup> Total Fe as FeO.

n.d. = not determined.

### Other minerals

Compositions of interstitial clinopyroxene (Table 5) in the dunites are similar to the compositions of clinopyroxene in the pyroxenite and spinel lherzolite suites from the Koolau shield except that  $\text{Al}_2\text{O}_3$  and  $\text{Na}_2\text{O}$  contents in the dunite clinopyroxenes (2.1–2.7 per cent  $\text{Al}_2\text{O}_3$  and < 0.4 per cent  $\text{Na}_2\text{O}$ ) are lower than in the lherzolite and pyroxenite clinopyroxenes (2.6–8.6 per cent  $\text{Al}_2\text{O}_3$  and 0.4–2.5 per cent  $\text{Na}_2\text{O}$ ).

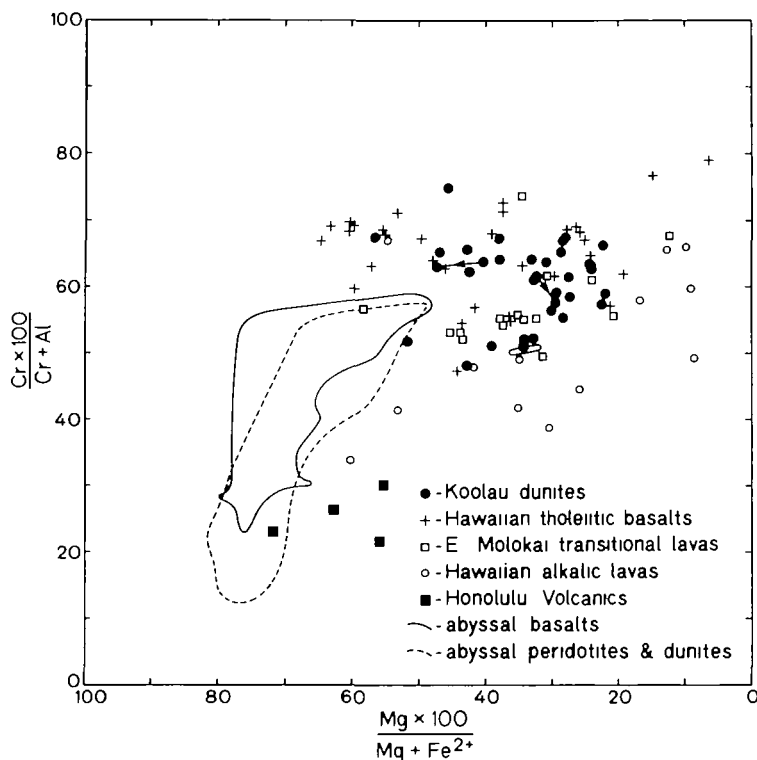


FIG. 3. Spinel compositions. Koolau dunites are from Table 4. Connected points are core and rim analyses of zoned grains with the arrow pointing toward the rim composition. Other data sources are as follows. Hawaiian tholeiitic basalts from Clague *et al.* (1980), Bence (1981, table 1.2.6.6, with recalculated structural formulas), Evans & Wright (1972), and R. T. Helz (unpublished data); E. Molokai transitional lavas selected from Beeson (1976); Hawaiian alkalic lavas from Clague *et al.* (1980) and Bence (1981, table 1.2.6.6); abyssal basalts from Hodges & Papike (1976), Ayuso *et al.* (1976), Sigurdsson & Schilling (1976), Graham *et al.* (1978), Frey *et al.* (1974), and O'Donnell & Presnall (1980); abyssal peridotites and dunites from Dick & Bullen (1984). The cigar-shaped outline at a  $\text{Cr} \times 100/(\text{Cr} + \text{Al})$  value of about 50 and a  $\text{Mg} \times 100/(\text{Mg} + \text{Fe}^{2+})$  value of about 34 encloses two abyssal basalt spinel compositions separated from the main field. Note that the horizontal axis is expanded by a factor of two relative to most plots of this type.

Because in some places interstitial clinopyroxenes in the Kalihi dunites are in contact with glass that has penetrated the xenolith from the host basalt, it might be supposed that some of these clinopyroxenes are the result of crystallization from the penetrating melt. White (1966, table 4) reported a partial analysis of a clinopyroxene from the melilite nephelinite host at Kalihi with a composition of  $\text{En}_{38}\text{Fs}_{12}\text{Wo}_{50}$ . Also, Mansker (1979) analysed a large number of clinopyroxenes from lavas of the Kalihi and Nuuanu vents and found a very similar average composition of  $\text{En}_{36}\text{Fs}_{13}\text{Wo}_{51}$ . The fact that these compositions are much higher in Ca and Fe/Mg than the clinopyroxenes we have analysed (Table 5) suggests that the dunite



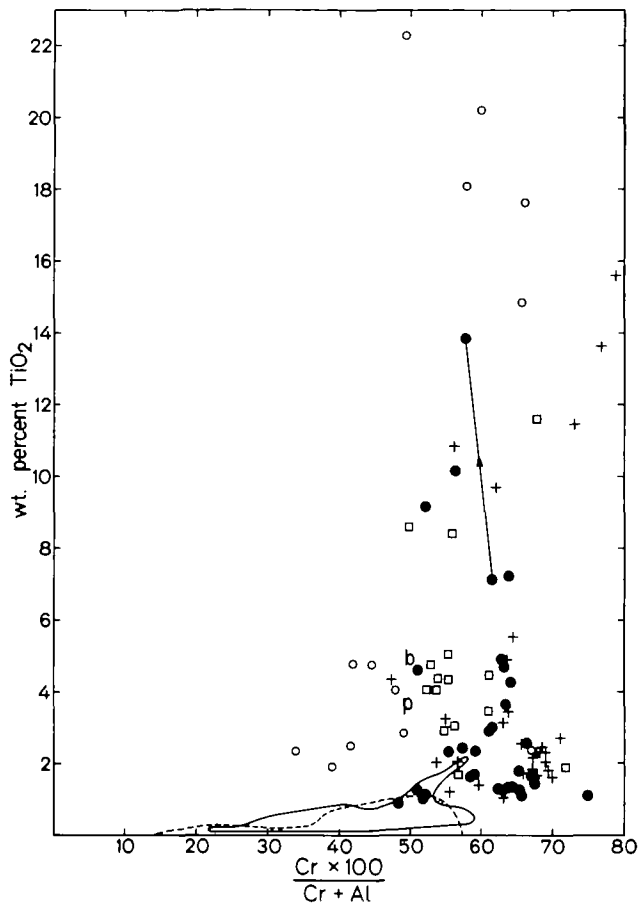


FIG. 4. Spinel compositions. Symbols, lines, and data sources are as in Fig. 3. The symbols p and b are spinels from an abyssal peridotite and abyssal basalt, respectively, that lie outside the main fields for these rock types.

clinopyroxenes are part of the primary mineral assemblage and unrelated to contamination effects from the host lava.

Further support for this conclusion comes from a comparison of  $(Mg \times 100)/(Mg + Fe)$  ratios (*mg* values) of coexisting clinopyroxene and olivine in the dunites. Clinopyroxene is unquestionably part of the primary mineral assemblage in lherzolite and pyroxenite xenoliths from Koolau, and in 31 samples of these xenoliths the *mg* number of coexisting olivine and clinopyroxene is always equal within  $\pm 4$  mole per cent (Sen, 1981). Such a correlation would be expected for equilibrated olivines and clinopyroxenes. With one exception (77KALI-8, which shows a deviation of 6 mole per cent), this same correlation holds also for the Koolau dunites that contain clinopyroxene (Tables 3 and 5), although the *mg* value for clinopyroxene is always slightly greater than that of the coexisting olivine. In contrast, clinopyroxene produced by contamination from the much more iron-rich host basalt would be expected to have somewhat lower *mg* values than the coexisting olivine. Thus, we conclude that the dunite clinopyroxenes are part of the primary mineralogy of the xenoliths and are not products of contamination from the host basalt.

Orthopyroxene occurs interstitially in one dunite from Kalihi and one from Nuuanu (Table 2). The orthopyroxene of 77KALI-1 is less aluminous (1.10 per cent  $Al_2O_3$ , Table 5)

TABLE 4  
Spinel compositions

No. of grains*	<i>Hawaiiiloa (1)</i>		<i>Ulupau (5)</i>							<i>Kalihi (7)</i>		
	<i>69KANS-1</i>		<i>66ULUP-23</i>		<i>66ULUP-1</i>	<i>77ULUP-1</i>		<i>77ULUP-2</i>	<i>77ULUP-3</i>	<i>77ULUP-4</i>	<i>77KALI-1</i>	
	1	1	1	1	3	1	1	1	1	1	1	
SiO <sub>2</sub>	n.d.	n.d.	0·00	0·09	0·13	0·00	0·00	0·00	0·00	0·00	0·00	0·00
TiO <sub>2</sub>	3·00	2·89	1·70	1·62	2·42	1·44	1·35	1·83	1·80	2·58	4·59	
Al <sub>2</sub> O <sub>3</sub>	12·26	12·49	12·80	13·08	15·51	12·23	14·15	12·64	12·57	11·85	18·60	
Cr <sub>2</sub> O <sub>3</sub>	29·19	29·23	27·43	27·45	31·02	37·81	37·81	38·21	35·17	34·70	28·96	
FeO <sup>b</sup>	48·30	46·16	51·14	49·56	46·55	40·91	38·72	40·70	43·19	45·25	38·79	
MnO	0·33	0·33	0·19	0·20	0·19	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	
MgO	5·80	6·95	4·42	5·52	4·75	5·70	6·84	5·88	5·91	4·64	9·16	
CaO	n.d.	n.d.	0·41	0·45	0·38	0·33	0·32	0·33	0·33	0·32	0·00	
Total	98·88	98·05	98·09	97·97	100·95	98·42	99·19	99·59	98·97	99·34	100·10	
Fe <sub>2</sub> O <sub>3</sub>	23·07	23·02	25·71	25·83	18·85	16·61	15·60	15·82	18·84	18·18	14·87	
FeO	27·54	25·45	28·01	26·32	29·59	25·97	24·69	26·47	26·24	28·89	25·41	
New total	101·19	100·36	100·67	100·56	102·84	100·08	100·75	101·18	100·86	101·16	101·59	
<i>Cations per 32 oxygens</i>												
Si	—	—	—	0·024	0·034	—	—	—	—	—	—	
Ti	0·608	0·584	0·349	0·329	0·478	0·294	0·269	0·368	0·364	0·526	0·878	
Al	3·892	3·958	4·114	4·169	4·803	3·911	4·422	3·989	3·983	3·790	5·575	
Cr	6·216	6·215	5·914	5·869	6·445	8·111	7·927	8·088	7·477	7·445	5·823	
Fe <sup>3+</sup>	4·676	4·658	5·275	5·255	3·727	3·391	3·112	3·186	3·812	3·712	2·846	
Fe <sup>2+</sup>	6·203	5·723	6·388	5·952	6·502	5·892	5·474	5·927	5·900	6·556	5·405	
Mn	0·075	0·075	0·044	0·046	0·042	—	—	—	—	—	—	
Mg	2·329	2·786	1·797	2·225	1·861	2·306	2·704	2·347	2·369	1·877	3·473	
Ca	—	—	0·120	0·130	0·107	0·096	0·091	0·095	0·095	0·093	—	

<i>Kalihi (7)</i>												
No. of grains <sup>a</sup>	<i>77KALI-1D</i>		<i>77KALI-2</i>				<i>77KALI-3</i>		<i>77KALI-4</i>		<i>77KALI-8</i>	
	1	1	1	3	1		5	1	1	1	1	1
					core	rim						
SiO <sub>2</sub>	0-12	0-17	0-79	0-64	0-62	0-75	0-00	0-00	0-12	0-08	0-72	1-06
TiO <sub>2</sub>	1-01	4-27	4-91	10-15	7-11	13-84	4-69	7-22	2-34	1-25	9-15	3-66
Al <sub>2</sub> O <sub>3</sub>	18-67	10-07	8-12	8-52	8-42	8-39	7-78	8-03	13-97	19-37	10-05	7-70
Cr <sub>2</sub> O <sub>3</sub>	29-94	26-79	20-45	16-40	20-06	17-07	19-91	21-03	30-19	29-96	16-39	19-90
FeO <sup>b</sup>	38-75	47-66	57-13	53-87	54-90	50-72	61-21	55-09	45-63	41-06	53-24	59-79
MnO	0-14	0-17	n.d.	n.d.	n.d.	n.d.	0-13	0-12	0-17	0-17	n.d.	n.d.
MgO	11-27	8-27	4-95	7-54	7-77	8-05	5-26	7-18	6-16	7-27	8-25	5-29
CaO	0-33	0-39	0-00	0-14	0-00	0-13	0-00	0-00	0-39	0-30	0-00	0-00
Total	100-23	97-79	96-35	97-26	98-88	98-95	98-98	98-67	98-97	99-46	97-80	97-40
Fe <sub>2</sub> O <sub>3</sub>	22-22	26-10	29-52	25-15	28-95	18-21	35-11	29-21	21-24	18-02	25-89	33-54
FeO	18-75	24-17	30-57	31-24	28-86	34-34	29-62	28-81	26-51	24-84	29-95	29-61
New total	102-46	100-41	99-31	99-78	101-78	100-77	102-50	101-60	101-10	101-27	100-39	100-76
<i>Cations per 32 oxygens</i>												
Si	0-030	0-046	0-222	0-175	0-167	0-202	—	—	0-032	0-021	0-194	0-294
Ti	0-189	0-864	1-038	2-088	1-438	2-803	0-967	1-474	0-469	0-242	1-851	0-764
Al	5-488	3-194	2-690	2-747	2-669	2-664	2-513	2-569	4-385	5-880	3-186	2-518
Cr	5-903	5-700	4-545	3-547	4-265	3-635	4-314	4-514	6-357	6-101	3-486	4-365
Fe <sup>3+</sup>	4-170	5-286	6-245	5-179	5-857	3-690	7-240	5-968	4-257	3-493	5-240	7-002
Fe <sup>2+</sup>	3-912	5-441	7-186	7-147	6-490	7-735	6-787	6-540	5-905	5-351	6-736	6-870
Mn	0-030	0-039	—	—	—	—	0-030	0-028	0-038	0-037	—	—
Mg	4-190	3-318	2-074	3-075	3-115	3-233	2-149	2-906	2-446	2-792	3-308	2-188
Ca	0-088	0-112	—	0-041	—	0-038	—	—	0-111	0-083	—	—

Table 4 continues on p. 208

TABLE 4 (cont.)

No. of grains <sup>a</sup>	<i>Kalihi (7)</i>			<i>Pah (15)</i>	<i>Kaau (17)</i>					
	<i>77KALI-9</i>			<i>77PAI-42D</i>	<i>77KAPS-13</i>			<i>77KAPS-27</i>		
	1	1	1	1	6	1	1		3	1
							core	rim		
SiO <sub>2</sub>	0.52	0.63	0.08	0.00	0.02	0.00	0.03	0.02	0.08	0.08
TiO <sub>2</sub>	2.31	1.11	0.88	1.10	1.53	1.30	1.32	1.27	1.10	1.27
Al <sub>2</sub> O <sub>3</sub>	16.73	20.04	23.14	9.61	12.46	14.99	14.34	14.97	9.14	11.30
Cr <sub>2</sub> O <sub>3</sub>	30.92	32.59	32.13	42.69	38.74	36.98	37.46	37.94	26.03	31.67
FeO <sup>b</sup>	43.43	38.12	34.29	34.39	33.97	37.10	38.12	35.10	53.68	44.01
MnO	n.d.	n.d.	n.d.	0.35	0.14	0.19	0.17	0.17	0.13	0.16
MgO	6.24	7.47	9.71	9.21	12.10	8.90	8.48	10.07	8.81	9.75
CaO	0.00	0.20	0.00	0.00	0.34	0.35	0.35	0.30	0.25	0.26
Total	100.15	100.16	100.95	97.25	99.30	99.81	100.27	99.84	99.22	98.50
Fe <sub>2</sub> O <sub>3</sub>	16.86	13.83	12.31	16.40	19.43	17.21	17.49	16.89	36.25	27.03
FeO	28.26	25.68	23.22	19.53	16.49	21.62	22.39	19.90	21.07	19.69
New total	101.84	101.55	102.18	98.89	101.25	101.53	102.02	101.53	102.85	101.21
<i>Cations per 32 oxygens</i>										
Si	0.136	0.161	0.197	—	0.005	—	0.008	0.005	0.021	0.021
Ti	0.453	0.213	0.163	0.224	0.296	0.253	0.257	0.245	0.219	0.251
Al	5.141	6.025	6.728	3.068	3.773	4.573	4.381	4.531	2.852	3.507
Cr	6.374	6.573	6.267	9.142	7.869	7.569	7.678	7.704	5.448	6.593
Fe <sup>3+</sup>	3.307	2.654	2.284	3.342	3.756	3.352	3.411	3.264	7.220	5.355
Fe <sup>2+</sup>	6.163	5.478	4.790	4.425	3.542	4.680	4.853	4.275	4.663	4.336
Mn	—	—	—	0.080	0.030	0.042	0.037	0.037	0.029	0.036
Mg	2.426	2.841	3.571	3.719	4.634	3.435	3.277	3.856	3.477	3.827
Ca	—	0.055	—	—	0.094	0.097	0.097	0.083	0.071	0.073

<sup>a</sup> For xenoliths with more than one spinel composition, the analysis based on an average of several grains is the most frequently found.

<sup>b</sup> Total Fe as FeO.

n.d. = not detected.

TABLE 5  
Pyroxene compositions<sup>a</sup>

	<i>Ulupau (5)</i>		<i>Kalihi (7)</i>			
	<i>77ULUP-4</i>	<i>77KALI-1</i>		<i>77KALI-3</i>	<i>77KALI-4</i>	<i>77KALI-8</i>
	<i>cpx<sup>b</sup></i>	<i>cpx</i>	<i>opx<sup>c</sup></i>	<i>cpx</i>	<i>cpx</i>	<i>cpx</i>
SiO <sub>2</sub>	53.36	52.99	54.66	53.56	52.49	52.58
TiO <sub>2</sub>	0.09	0.23	0.12	0.64	0.63	0.65
Al <sub>2</sub> O <sub>3</sub>	1.37	1.99	1.10	2.16	2.67	2.09
Cr <sub>2</sub> O <sub>3</sub>	0.39	0.47	0.15	0.20	0.72	0.40
FeO <sup>d</sup>	4.33	6.28	10.58	5.20	3.58	3.99
MnO	n.d.	0.11	0.10	0.04	0.05	n.d.
MgO	17.31	19.72	31.03	16.41	16.16	16.86
CaO	21.74	18.75	0.77	21.51	23.06	22.69
Na <sub>2</sub> O	0.54	0.03	0.01	0.43	0.39	0.19
K <sub>2</sub> O	n.d.	0.00	0.00	n.d.	n.d.	0.00
Total	99.13	100.57	98.52	100.15	99.75	99.45
<i>Cations per 6 oxygens</i>						
Si	1.964	1.924	1.956	1.954	1.924	1.932
Al <sup>IV</sup>	0.036	0.076	0.044	0.046	0.076	0.068
Al <sup>VI</sup>	0.024	0.009	0.003	0.047	0.039	0.023
Ti	0.002	0.006	0.003	0.018	0.017	0.018
Cr	0.011	0.013	0.004	0.006	0.021	0.012
Mg	0.950	1.067	1.656	0.893	0.883	0.924
Fe <sup>2+</sup>	0.133	0.191	0.317	0.159	0.110	0.123
Mn	—	0.003	0.003	0.001	0.002	—
Ca	0.857	0.729	0.030	0.841	0.096	0.893
Na	0.039	0.002	0.001	0.030	0.028	0.014
Total	4.017	4.022	4.016	3.994	4.005	4.006
En	48.9	53.7	82.7	47.2	46.5	47.6
Fs	6.9	9.6	15.8	8.4	5.8	6.3
Wo	44.2	36.7	1.5	44.4	47.7	46.1

<sup>a</sup> Analyses based on average of 3 spots on a single grain, except where noted.

<sup>b</sup> Single spot analysis.

<sup>c</sup> Broad beam analysis of a single grain containing fine cpx exsolution lamellae.

<sup>d</sup> Total Fe as FeO.

n.d. = not determined.

than the orthopyroxenes of the pyroxenite and spinel lherzolite suite xenoliths (2.4–8.9 per cent). In terms of Fe/Mg, this orthopyroxene is in the compositional range of those from the pyroxenite suite but is more Fe-rich than the spinel lherzolite orthopyroxenes. Because the host lava is nephelinitic (Clague & Frey, 1982), this orthopyroxene is considered to be an integral part of the xenolith.

Plagioclase occurs interstitially in several of the Kalihi dunites (Tables 2 and 6). It is clearly not a product of contamination from the host basalt because the host is a nephelinite that contains no plagioclase. The plagioclase compositions for all three dunites analysed are fairly similar (An<sub>63</sub>–An<sub>67</sub>).

Amphibole (Table 7) occurs interstitially in one dunite (66ULUP-1). Its composition lies in the pargasite field (Leake, 1978).

TABLE 6  
*Plagioclase compositions from Kalihi (vent 7)*

	77KALI-1	77KALI-3	77KALI-8
	1,1,1 <sup>*</sup>	8,3,1,1	2
SiO <sub>2</sub>	51.63	53.38	51.63
TiO <sub>2</sub>	n.d.	0.00	0.00
Al <sub>2</sub> O <sub>3</sub>	30.80	29.97	30.38
FeO	n.d.	0.00	0.00
MgO	n.d.	0.00	0.00
CaO	12.57	12.69	13.91
Na <sub>2</sub> O	3.89	4.21	3.59
K <sub>2</sub> O	0.26	0.20	0.19
Total	98.73	100.45	99.70
<i>Cations per 32 oxygens</i>			
Si	9.475	9.621	9.415
Al	6.571	6.366	6.530
Na	1.384	1.471	1.269
Ca	2.472	2.451	2.718
K	0.061	0.046	0.044
Total	19.963	19.955	19.977
mole %			
Ab	35.3	37.1	31.5
An	63.1	61.8	67.4
Or	1.6	1.1	1.1

\* Notation indicates one spot analysed on each of 3 grains. Compositional zoning was not found, and the analyses are averages of all the spots analysed.

#### DISTRIBUTION OF XENOLITH TYPES

Jackson & Wright (1970) showed that the ultramafic xenoliths in the Honolulu Volcanics are compositionally zoned with respect to the old Koolau tholeiitic shield through which the Honolulu Volcanics passed. Dunites predominate in the region of the central caldera, lherzolites predominate at greater distances from the caldera, and pyroxenites are restricted to a zone farthest from the caldera on the apron of the shield. We have found general agreement with this distribution of xenolith rock-types, except for two revisions. Jackson & Wright reported garnet-bearing xenoliths at Kaau, but despite a careful search of flows from this vent, we have found no such xenoliths. In the absence of garnet-bearing samples, the abundance distribution of xenolith rock-types at Kaau becomes very similar to that of the Pali Group, so we have reclassified Kaau as a member of Jackson & Wright's 'intermediate zone' in which lherzolite is dominant over dunite (Fig. 1).

The second revision concerns the report by Jackson & Wright (1970) of dunite xenoliths at the Salt Lake and Aliamanu vent groups. We have no dunites in our own collection from these localities, and only one has been found in the Jackson collection (sample no. 69SAL-111). This sample has an equigranular mosaic texture with few olivine deformation lamellae, and has a mode of 95 per cent olivine, 3 per cent orthopyroxene, 2 per cent clinopyroxene, and a trace of spinel. The olivine composition is Fo<sub>91</sub> and the spinel has a Mg × 100/(Mg + Fe<sup>2+</sup>) ratio of 76, a Cr × 100/(Cr + Al) ratio of 29, and TiO<sub>2</sub> content of 0.43

TABLE 7  
*Amphibole composition in 66ULUP-1 (Ulupau, vent 5)<sup>a</sup>*

		<i>Cations per 23 oxygens</i>	
SiO <sub>2</sub>	42.96	Si	6.175
TiO <sub>2</sub>	3.20	Al <sup>IV</sup>	1.825
Al <sub>2</sub> O <sub>3</sub>	12.58	ΣT	8.000
Cr <sub>2</sub> O <sub>3</sub>	1.33		
FeO <sup>b</sup>	6.70	Al <sup>VI</sup>	0.307
MgO	16.05	Ti	0.346
CaO	11.69	Fe <sup>2+</sup>	0.806
Na <sub>2</sub> O	2.83	Cr	0.151
K <sub>2</sub> O	1.03	Mg	3.390
Total	98.37	ΣC	5.000
		Mg	0.049
		Ca	1.800
		Na	0.151
		ΣB	2.000
		Na	0.638
		K	0.189
		ΣA	0.827

<sup>a</sup> Analysis based on average of 3 spots on a single grain.

<sup>b</sup> Total Fe as FeO.

wt. per cent. Except for its larger proportion of olivine, 69SAL-111 is identical in its mineralogy, mineral compositions, and texture to members of the spinel lherzolite suite. The olivine and spinel compositions in 69SAL-111 are both quite different from those in the other dunites reported here (Figs. 2, 3, and 4). Thus, we consider this sample to be genetically related to samples in the spinel lherzolite suite and have excluded it from the dunites reported here.

When the Kaau vent is reclassified, garnet-bearing xenoliths are restricted to a small area around the Salt Lake vent group (Fig. 1). The validity of Jackson & Wright's contour separating garnet-bearing from garnet-free xenoliths, which they drew roughly concentric with the central caldera, then becomes doubtful. However, the association of dunites with the central part of the volcano is quite strong and is made even stronger by the apparent absence of genetically related dunites from the Salt Lake and Aliamanu vent groups located far out on the apron of the shield.

#### ORIGIN OF DUNITES

The dunite xenoliths could be either accumulations of olivine from crystallization of basaltic magma or residues of partial fusion. These processes are both generally believed to occur at mid-ocean ridges, and the dunites could possibly be samples of the lower oceanic crust or depleted mantle immediately below the crust through which the Honolulu Volcanics passed. Alternatively, the dunites may record crystallization or fusion processes related to the formation of Koolau Volcano.

White (1966) suggested that Hawaiian dunite xenoliths are crystal accumulations from basaltic magma. Jackson & Wright (1970) were impressed by the deformed textures of the dunites and the concentration of this xenolith rock type around the central caldera of Koolau Volcano. Because of the absence of cumulate textures, Jackson & Wright argued that the

dunites were not produced by crystallization from a basaltic magma. They embraced the alternative that the dunites are residues of partial fusion and argued that the concentration of dunites around the central caldera required the partial fusion event to be associated with formation of Koolau Volcano.

We agree with Jackson & Wright (1970) that the association of the dunites with the central caldera is a strong argument for a genetic connection with the volcano-forming process. Compositions of olivines and spinels in the dunites reinforce this conclusion. The compositions of chrome spinel in the dunites lie in the composition range of spinels from Hawaiian tholeiitic, transitional, and alkalic lavas, whereas spinels from abyssal basalts, peridotites, and dunites have distinctly higher  $Mg/(Mg + Fe^{2+})$  ratios and lower  $Cr/(Cr + Al)$  ratios (Fig. 3). Also,  $TiO_2$  in spinels from abyssal basalts and peridotites is, with a very few exceptions, less than about 1 per cent (Dick & Bullen, 1984). Spinels from the Koolau dunites lie within the field of spinels from Hawaiian tholeiitic, transitional, and alkalic lavas at  $TiO_2$  values greater than about 1 per cent (Fig. 4). Olivine compositions of the Koolau dunites range from  $FO_{82.6}$  to  $FO_{89.7}$  (Table 3), which is essentially identical to the composition range of olivines from Hawaiian tholeiitic basalts (Fig. 2). In contrast, Hamlyn & Bonatti (1980) reported two abyssal dunites with olivine compositions of  $FO_{90}$  and  $FO_{89.7}$  and eight abyssal harzburgites and spinel lherzolites with compositions ranging from  $FO_{89}$  to  $FO_{90}$ . Sinton (1979) found olivine compositions in two peridotites from DSDP leg 45 to be  $FO_{90}$  and  $FO_{91.1}$ . Dick & Fisher (1984) reported the range of olivine compositions in 65 abyssal peridotites from 23 separate locations to be  $FO_{90}$  to  $FO_{91.6}$ . In the Samail ophiolite, the basalt cumulus dunite has an olivine composition of  $FO_{89}$  to  $FO_{90.5}$ , and olivine in the harzburgite tectonite beneath the layered sequence has a composition of  $FO_{90.5}$  to  $FO_{90.8}$  (Pallister & Hopson, 1981). All of these data consistently indicate that olivine compositions from dunites and spinel lherzolites in the lower oceanic crust and uppermost mantle have slightly higher and considerably more uniform forsterite contents than olivine from the Koolau dunites. Thus, both the olivine and spinel composition data strongly support the conclusion that the Koolau dunites were produced by processes that formed the Koolau Volcano and are not fragments of typical oceanic upper mantle or lower crust.

We are left with two alternatives. The dunites are either residues from production of the Koolau lavas or crystal accumulations from these lavas. Jackson & Wright's (1970) proposal that the dunites are residues of partial fusion encounters some severe difficulties. First, the high  $TiO_2$  contents (about 1–14 wt. per cent, Fig. 4) of spinels in the Koolau dunites are incompatible with the very low  $TiO_2$  contents in spinels from residual peridotites (Dick & Bullen, 1984). Another problem is that most of the dunites are free of enstatite. To produce such a residue by partial fusion, the amount of melting must be high, which would result in very magnesian olivines. The dunite olivines are low in magnesium ( $FO_{82.6}$  to  $FO_{89.7}$ ) and in particular are lower in magnesium than olivines in spinel lherzolite xenoliths from the Honolulu Volcanics (Fig. 2). Because the lherzolites retain their enstatite, they must represent mantle material that either has not been melted or has been partially melted to a lesser degree than that of a dunite residue. Thus, if the dunites are related to the spinel lherzolites by a partial fusion process, the dunite olivines would be more magnesian than those from the lherzolites, not less, as is observed.

A similar conclusion results from a comparison of the dunites with various estimates of undepleted mantle peridotite. For example, the pyrolite model of Ringwood (1975) has an olivine composition of  $FO_{90}$ , and Carter (1970) argued that undepleted mantle contains olivine with a composition of  $FO_{86}$  to  $FO_{88}$ . All of the dunite olivines are more iron-rich than Ringwood's pyrolite, and most are richer in iron than Carter's undepleted mantle. Thus, the dunites as a group are not a viable residue from either of these starting materials.



Jackson & Wright (1970) argued that the dunites are residues from an undepleted mantle similar to the garnet lherzolite 66SAL-1 from Salt Lake (see also Leeman *et al.*, 1980). The olivine composition ( $Fo_{84}$ ; Shaw & Jackson, 1973) in this iron-rich xenolith falls about in the middle of the composition range of olivine from the Salt Lake pyroxenite suite (Fig. 2), and in other respects the mineral chemistry and texture of 66SAL-1 are typical of the pyroxenites. An unusually large amount of olivine in 66SAL-1 is the only feature that distinguishes it from the pyroxenites. Thus, it has been called a garnet peridotite, but genetically it is part of the pyroxenite suite. Olivine compositions in the pyroxenite-suite xenoliths range from  $Fo_{72}$  to  $Fo_{87}$  (Fig. 2), a composition range too iron-rich for undepleted mantle material. Another feature of the pyroxenites is that they intrude and sometimes enclose spinel lherzolite xenoliths, which suggests that the pyroxenites are not undepleted mantle but instead have a magmatic origin. Most other investigators have concluded that the pyroxenites are either precipitates from alkalic magmas or crystallized magmas (for example, see Kuno, 1969; Herzberg, 1978; Frey, 1980). Even if all of these arguments against an undepleted mantle origin for 66SAL-1 are rejected, one is still left with the fact that many of the dunite xenoliths have olivines more enriched in iron than  $Fo_{84}$  and thus could not represent residues from fusion of 66SAL-1.

Limits on the minimum pressure of partial fusion in the mantle beneath Hawaiian volcanoes and the maximum pressure of formation of the dunites result in further difficulties for the residue hypothesis. On the island of Hawaii, seismic events associated with movement of magma beneath Kilauea Volcano extend to a depth of at least 60 km (Eaton & Murata, 1960) and suggest a deep source region for the tholeiitic magmas. Leeman *et al.* (1980) presented rare earth element data for Hawaiian tholeiites from six Hawaiian volcanoes including Koolau. They concluded that garnet is required in the source region, which would imply a minimum depth to this region of about 60 km (Presnall & Helsley, 1982). Hofmann *et al.* (1984) agreed with this conclusion for Kilauea Volcano. Although Feigenson *et al.* (1983) concluded that the source region for tholeiites from Kohala Volcano contains little or no garnet, Lanphere & Frey (in press) argued that garnet is necessary. Chen & Frey (1983) modelled the trace element concentrations of tholeiites from Haleakala Volcano with a garnet peridotite source. Thus, the available data consistently indicate a source region for Hawaiian tholeiitic magmas at depths greater than about 60 km. Phase equilibrium constraints place a minimum depth of origin for the Hawaiian alkalic lavas at about 37–40 km (Presnall *et al.*, 1978; Presnall & Hoover, 1984). The occurrence of small amounts of plagioclase and pyroxene in several dunites from the Kalihi vent places a maximum on the depth of origin of the Koolau dunites at approximately 30 km, the maximum depth for the stability of plagioclase lherzolite (Green & Hibberson, 1976; Presnall *et al.*, 1979). Also, Roedder (1983) has shown that the entrapment pressure of  $CO_2$  inclusions in olivine from dunite xenoliths at Loihi Seamount, just off the southeastern coast of the island of Hawaii, is only 2.2–4.7 kb. These dunite xenoliths are identical to the Koolau dunites in their olivine compositions, spinel compositions, and deformed textures (D. A. Clague, personal communication). Thus, we conclude that the Koolau dunites formed at depths shallower than the source regions for both tholeiitic and alkalic magmas in Hawaii, and therefore are not residues from partial fusion. On the basis of our earlier abstract concerning these dunites (Sen & Presnall, 1980), Wright (1984, p. 1248) now agrees with this conclusion.

The only alternative is that the dunites are crystal accumulations from magmas of Koolau Volcano. Fig. 3 suggests that chrome spinels from Hawaiian post-erosional lavas are generally lower in  $Cr/(Cr+Al)$  at comparable  $Mg/(Mg+Fe^{2+})$  ratios than those from tholeiitic, transitional, and alkalic lavas. A large set of unpublished data of D. A. Clague (personal communication, 1984) for the island of Niihau confirms this difference. The dunite

spinel fall in the composition range for tholeiitic, transitional, and alkalic lavas (Figs. 3 and 4) and are clearly distinct from spinels in the host Honolulu Volcanics as well as those in other Hawaiian post-erosional lavas. Also, the distributions of olivine compositions in the dunites and in tholeiitic basalts coincide exactly (Fig. 2). Olivine compositions from alkalic lavas have a wider range of forsterite contents but overlap almost all of the composition range of the dunite olivines. Thus, the compositions of both the spinels and olivines in the dunites are consistent with formation of these xenoliths as crystal accumulations from typical Hawaiian tholeiitic, transitional, or alkalic lavas, but not from nephelinitic post-erosional lavas.

Strange *et al.* (1965) modelled the gravity high centred over Koolau Volcano as a shallow high-density plug having a diameter of 16 km and extending to a depth of 13 km below sea level within a thickened crust. The diameter of their plug is consistent with the size of the central zone of xenolith localities dominated by dunite (Fig. 1). Their model density contrast from 3.2 g/cm within the plug to 2.8–2.9 g/cm outside the plug is consistent with a concentration of dunite in the crust beneath the central caldera.

The characteristic deformed textures and absence of any relict cumulate textures was used by Jackson & Wright (1970) as evidence against a cumulate origin. However, we contend that the compositional arguments outweigh the textural evidence. Kirby & Green (1980) argued that dunite xenoliths in lavas of Hualalai Volcano on the island of Hawaii were deformed immediately before incorporation into the host magma. If deformation of the Koolau dunites also occurred just prior to eruption of the host lavas, the original cumulate textures could have been destroyed. Helz (1983 and personal communication) has found dunite clots at Kilauea Volcano, Hawaii, that are essentially identical to the Koolau dunites in their deformed textures, olivine compositions, and spinel compositions. She suggested (Helz, 1983) that these clots are accumulations from tholeiitic magmas beneath Kilauea.

Compositional differences ( $\text{SiO}_2$ ,  $\text{Al}_2\text{O}_3$ ,  $\text{TiO}_2$ ,  $\text{CaO}$ ) between clinopyroxenes in alkalic lavas and those in tholeiites have been documented from a wide variety of tectonic settings (Kushiro, 1960; LeBas, 1962; Nisbet & Pearce, 1977; Leterrier *et al.*, 1982). Also Bence (1981) has shown, specifically for Hawaiian clinopyroxenes, that those in alkalic lavas are higher in the wollastonite component than those in tholeiites. We have used these chemical criteria, with partial success, to determine whether the Koolau dunites that contain clinopyroxene were crystallized from alkalic or tholeiitic magmas. When compared to clinopyroxene compositions from Hawaiian tholeiitic and alkalic lavas (Fodor *et al.*, 1975; Beeson, 1976; Clague *et al.*, 1980; Bence, 1981), one of the dunite clinopyroxenes in Table 5 (77KALI-1) appears unambiguously to be formed from a Hawaiian tholeiitic magma on the basis of its high  $\text{SiO}_2$  and low  $\text{TiO}_2$ ,  $\text{Al}_2\text{O}_3$ , and wollastonite content. The other dunite clinopyroxenes in Table 5 have  $\text{SiO}_2$ ,  $\text{TiO}_2$ , and  $\text{Al}_2\text{O}_3$  contents also consistent with those of Hawaiian tholeiitic clinopyroxenes. However, these same clinopyroxenes have wollastonite contents within the range spanned by Hawaiian alkalic clinopyroxenes and higher than that of any clinopyroxene from a Hawaiian tholeiite. Thus, these clinopyroxenes are neither clearly alkalic nor clearly tholeiitic.

On the basis of the clinopyroxene from 77KALI-1, we conclude that at least some of the dunites are associated with the tholeiitic eruptive stage. Our data do not exclude participation of alkalic magmas in the production of some of the dunites, but alkalic lavas immediately following formation of the main tholeiitic shield have not been found at Koolau. Thus, if alkalic magmas are involved, they would have to be associated with an assumed early eruptive stage now deeply buried beneath the tholeiitic lavas and below the current level of erosion. In view of the discovery of early alkalic lavas at Loihi Seamount, this possibility cannot be ruled out. However, Loihi lies on the flank of Kilauea about 1000 m below sea level and is still a small volcano. Thus, the volume of early alkalic lavas at Loihi is small, and

the traditional view that tholeiitic lavas constitute most of the bulk of Hawaiian volcanoes (Macdonald, 1963) is probably still correct. Because of (1) these relative volume relationships, (2) the inference from gravity data of a large plug of dunite in the central part of Koolau Volcano, and (3) the pronounced trend of olivine control typical of Hawaiian tholeiitic magmas, we favour the view that all or at least most of the dunite xenoliths are associated with the tholeiitic stage of Koolau Volcano. Such a large amount of olivine crystallization is consistent with the contention of Jackson & Wright (1970) and Wright (1984) that the parental magmas for Hawaiian tholeiites are picritic.

Lanphere (1983) has argued that dunite xenoliths from Loihi Seamount probably cannot be produced from tholeiitic magmas because the xenoliths are found in alkalic lavas that are generally older than the tholeiites. However, the tholeiitic or alkalic origin of the Loihi dunites is probably best left open for the present because the dunites could be genetically related to Kilauea Volcano, through which the Loihi magmas passed.

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