30. MINERAL COMPOSITIONS IN LEG 144 LAVAS AND ULTRAMAFIC XENOLITHS: THE ROLES OF CUMULATES AND CARBONATITE METASOMATISM IN MAGMA PETROGENESIS¹

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ABSTRACT

Severe alteration of many of the Cretaceous-aged hotspot lavas recovered by drilling during Leg 144 of the Ocean Drilling Program made geochemical analysis of minerals an attractive alternative to whole-rock analysis. Igneous rock samples were recovered from Limalok, Lo-En, Wodejebato, MIT, and Takuyo-Daisan guyots. Unaltered groundmass grains, microlites, and phenocrysts of olivine, plagioclase, and clinopyroxene in representative lavas were analyzed by electron microprobe for major elements, Cr, and Ni. These data show that the lavas recovered during Leg 144 are moderately to very silica-undersaturated and belong to the alkalic series or, in some cases, to the basanitic series as identified in Hawaii; exceptions are lavas from Lo-En and Takuyo-Daisan guyots, which are more transitional in nature.

Ion microprobe analyses for rare earth elements, Sr, Cr, Y, V, Ti, and Zr on selected microlites and phenocrysts of clinopyroxene show that (1) microphenocrysts contain considerably higher abundances of incompatible elements than phenocrysts from the same lava because of syn- and post-eruption fractionation; (2) the basanitic lavas from Limalok Guyot sampled a different, and probably more enriched, source material than the basanitic lavas from MIT Guyot; (3) the transitional and alkalic series lavas from Lo-En, MIT, and Takuyo-Daisan guyots contain phenocrysts with similar concentrations of incompatible trace elements, despite differences in silica-saturation and source composition; and (4) relative to other lavas, most of the Wodejebato Guyot alkalic series lavas contain phenocrysts with higher trace element concentrations and higher La/Yb, probably because of differences in depth of crystallization and/or source composition.

Ultramafic xenoliths were recovered in lavas from Limalok and MIT guyots. Electron probe analyses of clinopyroxene and orthopyroxene demonstrate that both Type 1 (mantle residue) and Type 2 (cumulate) xenoliths were recovered from each guyot. The Type 2 xenoliths from Limalok Guyot are the cumulate products of differentiation of alkalic liquids. Those from MIT Guyot are the cumulate products of fractionation of tholeitic liquids. Ion microprobe analyses show that carbonatite metasomatism of mantle residue (probably oceanic lithosphere) by a plume process observed throughout the SOPITA region and in the Kerguelen Islands is not only ubiquitous in space across the Southern Hemisphere, but ubiquitous in time as early as the mid-Cretaceous (MIT Guyot is 128 Ma).

INTRODUCTION

Ocean Drilling Program (ODP) Leg 144 was the second "atolls and guyots" leg, sister to Leg 143. The objective was to drill Limalok, Lo-En, and Wodejebato guyots in the Marshall Seamount Chain, as well as MIT Guyot (an isolated feature far northwest of the Marshall Islands) and Takuyo-Daisan Guyot in the Japanese Seamount Group. These guyots are drowned atolls built on submerging volcanic islands formed during the Cretaceous in the southern Pacific Ocean. The principle objectives for recovery of igneous basement were to compare lava types and igneous geochemistry of the drilled guyots and to compare the igneous geochemistry of lavas recovered from these features with modern lavas in the South Pacific Isotopic and Thermal Anomaly (SOPITA) (Staudigel et al., 1991) or Dupal region (Hart, 1984).

The SOPITA region is an area of considerable hotspot volcanism, and at least five hotspots are currently or recently active in the SOPITA: Marquesan, Society, Samoan, Pitcairn, and Cook-Austral (see Koppers et al., this volume, for a location map). Basalts erupted at these hotspots display unusual, and often extreme, isotopic and incompatible element ratios and incompatible element concentrations when compared with lavas erupted at other hotspots around the globe (e.g., Vidal et al., 1984; Wright and White, 1986/1987; Staudigel et al., 1991; Chauvel et al., 1992). The SOPITA appears to have been active since the Cretaceous as short-lived volcanic chains with lavas geochemically similar to SOPITA lavas cover much of the western Pacific basin (Staudigel et al., 1991).

Igneous rock was recovered from nine sites on five guyots drilled during Leg 144. Drilling penetrated 48.5 m of basanitic lava flows and volcaniclastic breccias in Hole 871C on Limalok Guyot and 57.1 m of differentiated transitional lava flows in Hole 872B on Lo-En Guyot. Drilling at each of five sites on Wodejebato Guyot recovered primitive to somewhat fractionated alkalic series basalts. In Hole 873A, volcanic debris-flow breccias and 29.2 m of basalt were recovered. In Hole 874B, 15.9 m of ankaramitic basalt were penetrated. In Hole 875C, 6.7 m of highly altered and fractured basalt were drilled. In Hole 876A, 9.0 m of basalt were penetrated. In Hole 877A, 1.0 m of altered basalt was recovered. On MIT Guyot (Hole 878A), drilling penetrated 187.5 m of "basement" rocks. The recovered igneous rocks include volcanogenic sandstone, lithic tuff, basalt breccia, and alkalic and basanitic series lavas. On Takuyo-Daisan Guyot (Hole 879A), drilling penetrated 34.4 m of a complex, intercalated mixture of clinopyroxene-plagioclase phyric basalt and volcanic breccia, which appears to be a peperite (intrusion of lava into soft, wet sediment).

Some of these lavas are relatively unaltered, but many are altered to claystone (which retains relict igneous texture), and their wholerock compositions do not reflect magmatic compositions. However, unaltered clinopyroxene is ubiquitous and unaltered plagioclase is common. The major element compositions of these phases can be used to characterize, evaluate, and compare *all* of these lavas, regardless of the extent of whole-rock alteration. In some cases, the phenocryst compositions support the conclusions derived from petrographic and whole-rock geochemical analyses. But in other cases, evaluation of phenocryst compositions provide the definitive characterization of a lava or a suite of lavas. For example, groundmass clinopyroxene compositions demonstrate that the Lo-En Guyot lavas are transitional rather than truly alkalic, as petrographic and whole-rock major element analyses have indicated.

 ¹ Haggerty, J.A., Premoli Silva, I., Rack, F., and McNutt, M.K. (Eds.), 1995. Proc. ODP, Sci. Results, 144: College Station, TX (Ocean Drilling Program).
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Trace element concentrations and ratios in clinopyroxene can demonstrate fundamental differences in mantle sources of lavas because trace element concentrations in clinopyroxene have some direct, although not simple, relationships to the trace element concentrations in the lava from which it crystallizes. These differences may be difficult to establish from the whole-rock geochemistry because of the effects of alteration. Furthermore, trace element analyses of carefully chosen clinopyroxene grains can be used to evaluate the extent of syn- and post-eruption differentiation of a liquid.

The first-order objective of the research I present in this paper is to compare the geochemistry of Cretaceous-aged, western Pacific hotspot lavas with SOPITA and Hawaiian lavas through an evaluation of their mineral chemistry. Because distinct geochemical differences are seen between SOPITA and other hotspot lavas (Hawaii is chosen as a Northern Hemisphere baseline because so much data is available), these differences should be apparent in the mineral chemistry. If these differences in mineral chemistry can be established, the relationships of even very altered western Pacific lavas to SOPITA lavas can be evaluated. However, at this time very few trace element data of phenocrysts from either Hawaiian lavas or SOPITA lavas exist. Thus, publication of the data presented in this paper is only the first step toward this larger objective.

Ultramafic xenoliths were recovered in lavas from Limalok and MIT guyots. Chemical analyses of xenoliths provide information about the underlying lithosphere, about plume chemistry and processes, and about magmas that cannot be sampled at the surface. The silicate phases of xenoliths recovered by drilling during Leg 144 were analyzed for major and trace element concentrations. These results are discussed in this paper ("Ultramafic Xenoliths"), along with the results of analyses of phenocrysts, microphenocrysts, and microlites ("Cogenetic Minerals").

ANALYTICAL TECHNIQUES

Compositions of silicate minerals for major elements, Ni, and Cr were obtained using a Cameca CAMEBAX Microbeam electron probe at Scripps Institution of Oceanography (SIO). Analyses were performed on polished 30-µm thin sections. Thin sections made aboard ship were repolished at SIO, and the author's shore-based thin sections were made at SIO (Table 1). All of the latter are 1 in. diameter rounds because this was the only size that could be introduced into the ion microprobe. Data reduction included routine atomic number, absorption, and fluorescence (ZAF) corrections. However, analyses were further normalized to analyses of standard minerals (USNM Kakanui Augite, USNM San Carlos Olivine, USNM Lake County Plagioclase) obtained as unknowns. Selected analyses of olivine, clinopyroxene, and feldspar minerals appear in Tables 2–4, respectively. All of the electron probe data are on the CD-ROM (back pocket).

Ion microprobe analyses of clinopyroxene grains were made at Woods Hole Oceanographic Institution on a Cameca 3f. Rare earth elements analyzed were La, Ce, Nd, Sm, Eu, Dy, Er, and Yb. Other elements analyzed were Zr, Y, Sr, V, Ti, and Cr. These data are presented in Table 5. Analytical uncertainty is estimated at 20%–30%for Eu and 15% for all other elements. Reflected-light photomicrographs were used during ion probe analyses to locate spots previously analyzed by electron probe. Thus, each ion probe analysis has a corresponding electron probe analysis less than 50 µm away. These specific locations are denoted as CPX A, CPX B, and so on for each xenolith ("CPX" = clinopyroxene).

COGENETIC MINERALS

Major Element Chemistry

Olivine

The only Leg 144 lavas with unaltered olivine phenocrysts are the ankaramitic lavas of Wodejebato Guyot (Hole 874B) and one alkalic basalt of MIT Guyot (igneous Unit 15, Hole 878A). Forsterite contents of unaltered olivine phenocrysts from these lavas range from 81% to 86% (Table 2), corresponding to liquid Mg#s of 0.62 to 0.65 (calculations after Roeder and Emslie, 1970). Liquids as magnesian as these are not typical of Leg 144 lavas. Specifically, the clinopyroxene compositions of the fresh-olivine-bearing basalts are more primitive than the majority of Leg 144 lavas.

Clinopyroxene

The interpretation of magma characteristics from clinopyroxene compositions is a complex task that requires distinguishing the effects of the pressure of crystallization, magma chemistry, and magma temperature, all of which exert strong controls on clinopyroxene chemistry (e.g., Le Bas, 1962; Mori and Green, 1976; Wells, 1977; Herzberg, 1978; Gill, 1981; Davies et al., 1989). A further complication arises because most clinopyroxene compositions used in evaluating the nature of the crystallizing liquid were wet chemical analyses of mineral separates (e.g., Le Bas, 1962, figs. 1, 2, 5). A wet chemical analysis probably represents the average clinopyroxene composition crystallized from a liquid. When electron microprobe analyses are evaluated, more scatter is expected because both clinopyroxene phenocryst and groundmass compositions from a single lava vary widely when analyzed in the microscopic detail made possible by the electron microprobe.

I use three diagrams to characterize the Leg 144 lavas from major element chemistry of clinopyroxene. A plot of TiO_2 vs. Mg# (Fig. 1) shows variations in the relative degree of differentiation among lavas and may provide some insight into the tholeiitic or alkalic nature. However, there is considerable overlap in the clinopyroxene compositions from Hawaiian tholeiitic and alkalic lavas (Fig. 1A). By itself, this diagram is not diagnostic of the silica-saturation of a magma.

The second diagram is a plot of Al(z) vs. TiO2 after Le Bas (1962) (Fig. 2). $[Al(z) = Al^{IV} \times 100/2;$ specifically, Al(z) is the percentage of tetrahedral sites occupied by Al.] The nonalkaline/alkaline and alkaline/peralkaline boundaries on this diagram were determined from groundmass compositions (Le Bas, 1962). In general, the Le Bas (1962) nonalkaline and alkaline categories roughly correspond to tholeiitic and alkalic as defined for hotspot lavas. But characterization and interpretation of the Leg 144 lavas are facilitated by plotting both phenocryst and groundmass compositions. Figure 2 shows the fields of Hawaiian tholeiitic and alkalic phenocryst compositions to assist interpretation of Leg 144 phenocryst data. Two effects from magma chemistry are observed on this plot. First, the silica-saturation of a magma controls the amount of tetrahedral aluminum substitution. Titanium positively correlates with Al(z) because of the cation charge deficit in the tetrahedral sites (Al3+ vs. Si4+). Second, differentiation drives clinopyroxene compositions to lower Al(z) and higher Ti for tholeiitic liquids and to higher Al(z) and higher Ti for alkalic liquids (Fig. 2B). As with Figure 1, this diagram is not uniquely diagnostic for ocean-island tholeiitic and alkalic series lavas, but it does appear to be diagnostic for basanitic series lavas that have clinopyroxene grains consistently containing greater than 10% tetrahedral substitution by Al (Al(z) > 10) (Fig. 2A, D).

The third diagram, also from Le Bas (1962), is a Fe+Mn, Mg, Ca+Na ternary (Fig. 3). This is essentially the classic Fs-En-Wo pyroxene ternary, but the addition of Na and Mn better separates the nonalkaline, alkaline, and peralkaline fields. This diagram is very useful for mafic (normalized Fe+Mn < 20%) groundmass compositions. Most phenocryst compositions from a given lava type will fall into the respective field, but some will scatter across the nearest boundary (Le Bas, 1962). Despite the scatter resulting from electron microprobe analyses, groundmass clinopyroxene compositions from the Hawaiian tholeiitic, alkalic, and nephelinitic series show only a slight overlap (Fig. 3A), indicating that this diagram is a useful diagnostic tool for hotspot lavas.

Table 1. Minerals analyzed by electron and/or ion microprobe.

| Core section | | Inneous | Range of unit | Grou | ndmass | P | henocr | ysts - | > | Kenoliths | |
|--------------------------------|------------------------|---------------------|--|------|--------|------------|--------|--------|-----|-----------|----|
| interval (cm) | Source | unit | interval [cm]) | CPX | PLAG | CPX | OL | PLAG | CPX | OPX | OL |
| imalok Guyot: | | | | | | | | | | | |
| 44-871C- 35R-1, 10–12 | Author | Unit I | 144-871C- 35R-1, 6-58 | | E | | | | E,I | | |
| 35R-1, 14-16 | Author | Unit 1 | 35R-1, 6-58 | | | E | | | | | |
| 35R-1, 20-23A | Author | Unit I | 35R-1, 6-58 | | | | | | E,I | г | |
| 35R-1, 20-23B 35R-1, 37-30 | Author | Unit 1 | 35R-1, 6-58 | EI | | E | | | EI | E | |
| 38R-1, 29-32 | Author | Unit 17 | 38R-1, 18 to 38R-4, 17 | E,I | E | E,1 | | | | | |
| 38R-3, 28-30 | Author | Unit 17 | 38R-1, 18 to 38R-4, 17 | E | | E | | | | | |
| 38R-6, 52-55 40R-1, 141-143 | Shipboard Shipboard | Unit 19 Unit 22B | 38R-5, 30 to 38R-6, 135 40R-1, 140 to 40R-2, 53 | | E | | | Е | | | |
| .o-En Guvot: | omproduo | Chill Lab | | | | | | | | | |
| 44-872B- | | | 144-872B- | | | | | 1000 | | | |
| 5R-1, 76-78 | Author | Unit 6 | 4R-1, 33 to 5R-2, 6 | E | Е | EI | | E | | | |
| 6R-1, 75-80 | Shipboard | Unit 11 | 6R-1, 74–100 | L | E | 1.,1 | | E | | | |
| 7R-5, 93-95 | Author | Unit 14B | 7R-4, 56 to 7R-6, 43 | | E | E | | E | | | |
| 7R-7, 23–25 | Author | Unit 15B | 7R-6, 135 to 7R-7, 83 | E | E | | | | | | |
| 8R-2, 131-133 9R-2, 92-94 | Author | Unit 16B | 8R-2, 18 to 9R-1, 5 9R-2, 81 to 9R-3, 90 | E | | Е | | E | | | |
| 9R-5, 107-109 | Author | Unit 18B | 9R-4, 0 to 9R-6, 60 | | | E,I | | - | | | |
| 9R-6, 14-15 | Author | Unit 18B | 9R-4, 0 to 9R-6, 60 | | - | E | | | | | |
| 9R-6, 41-42 | Author | Unit 18B | 9R-4, 0 to 9R-6, 60 | | E | | | | | | |
| Vodejebato Guyot 44-873A- | : | | 144-873A- | | | | | | | | |
| 15R-1, 23-26 | Shipboard | Unit 4 | 15R-1, 2-54 | | E | | | 1:57 | | | |
| 17R-1, 22-28 | Shipboard | Unit 5 | 17R-1, 0-39 | | | E | | E | | | |
| 18R-1, 30-33 18R-1 94-95 | Author | Unit 8 | 18R-1, 0-133 18R-1, 0-133 | E | F | E,I F I | | | | | |
| 144 9740 | Aution | Onit 0 | 144 9740 | | 1 | 1.1,1 | | | | | |
| 23R-1, 23-25 | Author | Unit 1 | 22R-4. 0 to 24R-4, 130 | | | E | | | | | |
| 24R-1, 12-16 | Author | Unit 1 | 22R-4, 0 to 24R-4, 130 | E | | E,I | E | 1125 | | | |
| 24R-4, 11-15 | Author | Unit 1 | 22R-4, 0 to 24R-4, 130 | | Е | E,I | E | Е | | | |
| 24K-4, 102-100 | Author | Unit I | 22K-4, 0 to 24K-4, 150 | | | E | | | | | |
| 144-875C- | A SOLUTION OF | 110420 | 144-875C- | F | E | | | | | | |
| 15M-1, 1–4 15M-1, 19–21 | Author | Unit I Unit I | 14M-1, 65 to 15M-1, 77 | E | E | | | | | | |
| | 1104.0010120 | 1.0000 | | | | | | | | | |
| 144-876A- 16R-1 1-3 | Author | Unit 1B | 144-876A- 15R-1 56 to 16R-1 35 | E | F | F | | | | | |
| 16R-1, 9-11 | Author | Unit 1B | 15R-1, 56 to 16R-1, 35 | | | Ĕ | | | | | |
| 17R-1, 56-61 | Author | Unit 3 | 17R-1, 40-96 | | E | E,I | | | | | |
| 144-877A- | | | 144-877A- | | 1.22 | 10.00 | | | | | |
| 20R-5, 22-24 | Author | Unit 1 | 20R-4, 98 to 20R-5, 35 | | Е | E,I | | E | | | |
| MIT Guyot: | | | 144 070 4 | | | | | | | | |
| 78R-2 44-46 | Author | Unit 1 | 78R-1 44 to 78R-2 88 | | F | | | E | | | |
| 80R-6, 101-104 | Author | Unit 8 | 80R-2, 74 to 80R-6, 120 | E | | | | 10 | | | |
| 81R-2, 37-40 | Author | Unit 10 | 81R-1, 6 to 81R-3, 103 | E | E | | | | E,1 | | |
| 81R-5, 22-25 81R-5, 60, 62 | Shipboard | Unit 11 | 81R-3, 103 to 82R-1, 0 81R-3, 103 to 82R-1, 0 | E | Е | F | | | | | |
| 83R-1, 6-8 | Author | Unit 13 | 83R-2, 0 to 83R-4, 43 | Ē | E | Ĕ,I | | E | | | |
| 84R-4, 111-113 | Author | Unit 15 | 84R-1, 19 to 85R-1, 0 | E | | - | E | | | | |
| 88R-3, 42-44 | Author | Unit 20 | 88R-1, 0 to 88R-3, 97 | E | E | E | F | | | | |
| 89R-2, 87-89 | Author | Unit 21 | 89R-1, 0 to 90R-2, 70 | E | E | E | Б | | | | |
| 89R-2, 101-103 | Author | Unit 21 | 89R-1, 0 to 90R-2, 70 | Ē | | E | | | | | |
| 89R-2, 133-137 | Author | Unit 21 | 89R-1, 0 to 90R-2, 70 | | | E | | | | | |
| 89R-3, 113-115 | Author | Unit 21 | 89R-1, 0 to 90R-2, 70 90R-4, 145 to 92R-1, 0 | | | E | | | E1 | | |
| 90R-6, 63-64 | Author | Unit 23 | 90R-4, 145 to 92R-1, 0 | | | | E | | E,I | Е | |
| 91R-1, 88-89 | Author | Unit 23 | 90R-4, 145 to 92R-1, 0 | | | E | | | E,1 | | |
| 91R-3, 79-81 | Author | Unit 23 | 90R-4, 145 to 92R-1, 0 | | F | | | | | E | 1 |
| 91R-5, 94-97 91R-5, 47-54 | Author | Unit 23 | 90R-4, 145 to 92R-1, 0 90R-4, 145 to 92R-1, 0 | | E | | | | E | | 1 |
| 92R-5, 41-43 | Shipboard | Unit 27 | 92R-4, 101 to 92R-5, 86 | | E | | | | 50 | | |
| 93R-2, 111-114 | Author | Unit 30 | 93R-2, 63 to 94R-1, 58 | E | | E | | | | | |
| 95R-6, 33-34 97R-1, 46, 50 | Author | Unit 32 | 95R-5, 83 to 95R-6, 58 | | | E | | | | | |
| 97R-2, 113-115 | Author | Unit 34 | 97R-1, 19 to 98R-1, 0 | | | E | | | | | |
| 97R-2, 128-130 | Author | Unit 34 | 97R-1, 19 to 98R-1, 0 | E | - | | | E | | | |
| 97R-3, 5–10 | Author | Unit 34 | 97R-1, 19 to 98R-1, 0 | | E | | | | | | |
| Takuyo-Daisan G | uyot: | | 144-8794- | | | | | | | | |
| 22R-1, 133-135 | Author | Unit 1B | 21R-1, 65 to 24R-1, 106 | | | E,I | | | | | |
| 24R-1, 85-87 | Author | Unit 1B | 21R-1, 65 to 24R-1, 106 | E | E | E | | E | | | |
| | | | | | | | | | | | |

Notes: CPX = clinopyroxene, PLAG = plagioclase, OL = olivine, and OPX = orthopyroxene. "E" implies electron probe analyses of the phase were done, and "I" implies ion probe analyses of the phase were done.

Table 2. Representative olivine analyses.

| Hole: | 874B | 874B | 874B | 874B | 874B | 874B | 874A | 874A | 874A | 874A | 874A | 874A | 874A |
|--------------------------------|---------------|------------|------------|------------|------------|------------|------------|------------|------------|------------|----------|----------|----------|
| Core, section: | 24R-1 | 24R-1 | *24R-4 | 24R-4 | 24R-4 | 24R-4 | 84R-4 | 84R-4 | 90R-6 | 90R-6 | 91R-5 | 91R-5 | 91R-5 |
| Interval (cm): | 12-16 | 12-16 | 12-16 | 11-15 | 11-15 | 11-15 | 111-113 | 111-113 | 63-64 | 63-64 | 47-54 | 47-54 | 47-54 |
| Description: | Core of small | Core of | Small | Small | Large | Small | Small | Small | Small | Small | Dunite | Dunite | Dunite |
| 0 10 10 | phenocryst | phenocryst | phenocryst | phenocryst | phenocryst | phenocryst | phenocryst | phenocryst | phenocryst | phenocryst | xenolith | xenolith | xenolith |
| Analysis: | OL 5 | OL 6 | OL 2 | OL 7 | OL 18 | OL 23 | OL 2 | OL 6 | OL 5 | OL 6 | OL 2 | OL 3 | OL 7 |
| Analysis no.: | 1 | 2 | 3 | 5 | 8 | 10 | 11 | 12 | 13 | 14 | 15 | 16 | 17 |
| Element (ppm) | c. | | | | | | | | | | | | |
| SiO | 39.60 | 39.08 | 38.64 | 38.71 | 38.84 | 40.15 | 39.26 | 39.24 | 39.26 | 37.73 | 39.39 | 38.76 | 36.97 |
| TiO ₂ | 0.05 | 0.02 | 0.00 | 0.00 | 0.02 | 0.00 | 0.02 | 0.03 | 0.01 | 0.02 | 0.03 | 0.02 | 0.04 |
| Al ₂ Ô ₂ | 0.04 | 0.03 | 0.04 | 0.03 | 0.05 | 0.04 | 0.07 | 0.05 | 0.05 | 0.03 | 0.01 | 0.00 | 0.03 |
| Cr ₂ O ₃ | 0.04 | 0.04 | 0.02 | 0.01 | 0.04 | 0.03 | 0.03 | 0.04 | 0.00 | 0.01 | 0.00 | 0.00 | 0.02 |
| FeO | 15.17 | 14.82 | 18.00 | 18.52 | 14.69 | 13.76 | 14.70 | 14.83 | 15.79 | 26.11 | 14.05 | 19.44 | 26.82 |
| MnO | 0.19 | 0.19 | 0.24 | 0.26 | 0.18 | 0.20 | 0.22 | 0.22 | 0.22 | 0.36 | 0.19 | 0.36 | 0.39 |
| MgO | 44.77 | 44.80 | 42.94 | 42.56 | 46.15 | 46.50 | 45.37 | 45.26 | 44.89 | 37.53 | 46.24 | 42.33 | 36.81 |
| NiO | 0.21 | 0.18 | 0.18 | 0.18 | 0.15 | 0.19 | 0.23 | 0.22 | 0.32 | 0.07 | 0.40 | 0.27 | 0.09 |
| CaO | 0.24 | 0.30 | 0.31 | 0.31 | 0.22 | 0.34 | 0.27 | 0.28 | 0.15 | 0.20 | 0.10 | 0.08 | 0.18 |
| Total | 100.31 | 99.46 | 100.37 | 100.58 | 100.34 | 101.21 | 100.17 | 100.17 | 100.69 | 102.06 | 100.41 | 101.26 | 101.35 |
| Per four oxyge | ns: | | | | | | | | | | | | |
| Si | 0.994 | 0.989 | 0.984 | 0.986 | 0.974 | 0.992 | 0.986 | 0.986 | 0.986 | 0.981 | 0.984 | 0.984 | 0.973 |
| Ti | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 |
| Al | 0.001 | 0.000 | 0.001 | 0.000 | 0.001 | 0.001 | 0.002 | 0.001 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 |
| Cr | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 |
| Fe ⁺² | 0.318 | 0.314 | 0.383 | 0.394 | 0.308 | 0.284 | 0.309 | 0.312 | 0.332 | 0.567 | 0.294 | 0.413 | 0.590 |
| Mn | 0.004 | 0.004 | 0.005 | 0.006 | 0.004 | 0.004 | 0.005 | 0.005 | 0.005 | 0.008 | 0.004 | 0.008 | 0.009 |
| Mg | 1.675 | 1.690 | 1.630 | 1.615 | 1.726 | 1.712 | 1.698 | 1.695 | 1.680 | 1.454 | 1.722 | 1.602 | 1.444 |
| Ni | 0.004 | 0.004 | 0.004 | 0.004 | 0.003 | 0.004 | 0.005 | 0.004 | 0.006 | 0.001 | 0.008 | 0.006 | 0.002 |
| Ca | 0.006 | 0.008 | 0.008 | 0.008 | 0.006 | 0.009 | 0.007 | 0.008 | 0.004 | 0.006 | 0.003 | 0.002 | 0.005 |
| Total | 3.002 | 3.009 | 3.015 | 3.013 | 3.022 | 3.006 | 3.012 | 3.011 | 3.013 | 3.017 | 3.015 | 3.015 | 3.023 |
| Fo | 84.0 | 84.3 | 81.0 | 80.4 | 84.8 | 85.8 | 84.6 | 84.5 | 83.5 | 71.9 | 85.4 | 79.5 | 71.0 |

Note: Analysis numbers refer to complete data table on CD-ROM (back pocket).

Only a part of this table is reproduced here. The entire table appears on the CD-ROM (back-pocket).

| | | | | Limalo | k Guyot | | | |
|----------------------------|--|--------------|--|--|--|--|---|--------------|
| Hole: | 871C | 871C | 871C | 871C | 871C | 871C | 871C | 871C |
| Lore, section: | 35K-1 | 35K-1 | 30K-1 | 35K-1 23.25 | 37 30 | 20_32 | 20_32 | 28-30 |
| Description: | Alkalic cumulate xenolith grain | Phenocryst | (Xenolith A) Mantle residue xenolith grain | (Xenolith B) Mantle residue xenolith grain | Alkalic cumulate xenolith grain | Core of 1.0-cm xenocryst; CPX 4 6 7 | Rim of 1.0-cm xenocryst; CPX 4, 5, 7 | Phenocryst |
| | gram | | gram | gram | gram | 8, A, B | 8, A, B | |
| Analysis: Analysis no.: | CPX 1 1 | CPX 14 14 | CPX B 21 | OPX 7 28 | CPX 9 31 | CPX 5 39 | CPX 6 40 | CPX 10 51 |
| Element (ppm): | | | | | | | | |
| SiO ₂ | 51.74 | 44.89 | 51.93 | 55.45 | 49.17 | 43.14 | 46.23 | 47.15 |
| TiO ₂ | 1.09 | 4.10 | 0.14 | 0.01 | 1.28 | 3.58 | 2.22 | 2.26 |
| Al_2O_3 | 4.60 | 8.79 | 6.64 | 3.78 | 8.49 | 11.70 | 9.13 | 6.42 |
| Cr_2O_3 | 0.31 | 0.03 | 1.30 | 0.46 | 0.02 | 0.00 | 0.41 | 0.05 |
| FeO | 5.68 | 7.12 | 2.62 | 5.99 | 5.98 | 7.65 | 6.08 | 7.01 |
| MnO | 0.09 | 0.03 | 0.09 | 0.10 | 0.06 | 0.05 | 0.19 | 0.05 |
| MgO | 18.56 | 13.00 | 16.21 | 32.99 | 13.82 | 11.36 | 13.33 | 13.74 |
| NiO | 0.02 | 0.05 | 0.04 | 0.11 | 0.00 | 0.01 | 0.08 | 0.01 |
| CaO | 18.23 | 21.94 | 19.77 | 1.18 | 20.51 | 21.97 | 22.14 | 22.23 |
| Na2O | 0.60 | 0.33 | 1.17 | 0.18 | 1.10 | 0.81 | 0.59 | 0.37 |
| Total | 100.91 | 100.28 | 99.91 | 100.25 | 100.43 | 100.27 | 100.40 | 99.28 |
| Per six oxygens: | | | | | | | | |
| Si | 1.867 | 1.676 | 1.875 | 1.913 | 1.799 | 1.617 | 1.713 | 1.772 |
| Ti | 0.030 | 0.115 | 0.004 | 0.000 | 0.035 | 0,101 | 0.062 | 0.064 |
| Al | 0.196 | 0.387 | 0.283 | 0.154 | 0.366 | 0.517 | 0.399 | 0.284 |
| Cr | 0.009 | 0.000 | 0.037 | 0.013 | 0.000 | 0.000 | 0.012 | 0.001 |
| Fe ⁺² | 0.171 | 0.222 | 0.079 | 0.173 | 0.183 | 0.240 | 0.188 | 0.220 |
| Mn | 0.002 | 0.000 | 0.003 | 0.003 | 0.002 | 0.002 | 0.006 | 0.002 |
| Mg | 0.999 | 0.723 | 0.872 | 1.696 | 0.754 | 0.635 | 0.736 | 0.770 |
| Ni | 0.000 | 0.002 | 0.001 | 0.003 | 0.000 | 0.000 | 0.002 | 0.000 |
| Ca | 0.705 | 0.877 | 0.765 | 0.044 | 0.804 | 0.882 | 0.879 | 0.895 |
| Na | 0.042 | 0.024 | 0.082 | 0.012 | 0.078 | 0.059 | 0.042 | 0.027 |
| Total | 4.022 | 4.028 | 4.000 | 4.010 | 4.021 | 4.053 | 4.039 | 4.035 |
| En | 53.26 | 39.68 | 50.83 | 88.66 | 43.31 | 36.14 | 40.82 | 40.85 |
| Fs | 9.14 | 12.19 | 4.61 | 9.04 | 10.51 | 13.66 | 10.43 | 11.67 |
| Wo | 37.60 | 48.13 | 44.56 | 2.30 | 46.18 | 50.20 | 48.75 | 47.48 |

Table 3. Representative clinopyroxene and orthopyroxene analyses.

Note: Analysis numbers refer to clinopyroxene data tables, by guyot, on the CD-ROM (back pocket).

Only a part of this table is reproduced here. The entire table appears on the CD-ROM (back-pocket).



Figure 1. TiO₂ vs. Mg/(Mg+Fe) of clinopyroxene phenocrysts. Mg and Fe calculated on six oxygens. All Fe in the clinopyroxene analyses are assumed to be Fe²⁺. Fields on each of Figures 1A–E are as follows: solid line = Hawaiian tholeiitic phenocrysts, and dashed line = Hawaiian alkalic phenocrysts. Hawaiian fields are drawn from data in Fodor et al. (1975), Clague et al. (1980), Garcia et al. (1982, 1986), Ho and Garcia (1988), Frey et al. (1990), and Chen et al. (1990).



Figure 2. Al(z) vs. TiO_2 of clinopyroxene phenocrysts. $Al(z) = Al^{IV} \times 100/2$. Al(z) and field boundaries from Le Bas (1962) for groundmass compositions. Fields on each of Figures 2A–E as on Figure 1. Field boundaries do not represent absolute delineations for phenocrysts. Diagonal lines without arrows (on each of Fig. 2A–E) are divisions between nonalkaline/alkaline and alkaline/peralkaline fields as established by Le Bas (1962). Lines with arrows (on Fig. 2B only) are toward increasing differentiation for nonalkaline and alkaline/peralkaline lavas in the respective divisions.



Figure 3. Fe+Mn, Mg, Ca+Na ternary diagram. Values are calculated from structural formulas based on six oxygens. Dashed lines separate nonalkaline/alkaline (lower line) and alkaline/peralkaline (upper line) boundaries after Le Bas (1962). All data are from groundmass analyses. Hawaiian data from sources listed in Figure 1 caption.

Limalok Guyot Lavas

Petrographic and major element analyses of the lavas recovered in Hole 871C at Limalok Guyot unequivocally place these lavas in the basanitic series (Christie et al., this volume). This strongly alkalic nature is reflected in the clinopyroxene chemistry. The TiO₂ contents of phenocrysts range from 1.6 to 4.6 wt% and are negatively correlated with Mg#, which ranges from 0.74 to 0.82 (Fig. 1A; Table 3). The compositions of groundmass grains vary more widely, but most lie within the same range as the phenocryst compositions. Most of these phenocryst and groundmass compositions have greater TiO₂ for a given Mg# than do Hawaiian alkalic series compositions (Fig. 1A). Despite the mafic nature of the lavas, most of the clinopyroxene compositions plot in the peralkaline field on the Al(z) vs. TiO₂ diagram of Le Bas (1962) and many lie at higher TiO₂ and Al(z) than the Hawaiian alkalic series compositions (Fig. 2A). This suggests that the Limalok Guyot lavas are more alkalic than the Hawaiian alkalic series lavas.

This suggestion is supported by the range of Limalok Guyot groundmass clinopyroxene compositions plotted on Figure 3B. As might be expected of basanitic series lavas, the groundmass clinopyroxene compositions of Limalok Guyot plot between (and overlap into each of) the fields of Hawaiian alkalic and nephelenitic series groundmass compositions across the alkaline/peralkaline boundary of the Fe+Mn, Mg, Ca+Na ternary (Fig. 3A, B).

Lo-En Guyot Lavas

Hole 872B lavas have been interpreted as belonging to a moderately differentiated alkalic series (Christie et al., this volume). Consistent with this interpretation, clinopyroxene compositions display moderate TiO₂ contents (0.6%-1.3%) and a tight, negative correlation, usually interpreted as being fractionation controlled, on the TiO₂ vs. Mg# diagram of Figure 1B.

In a plot of Al(z) vs. TiO₂, Lo-En Guyot clinopyroxene compositions straddle the nonalkaline/alkaline boundary of Le Bas (1962) (cf. Fig. 2A, B) and generally lie within the field of Hawaiian alkalic series lavas. However, several of the groundmass compositions lie at lower Al(z) and TiO₂ wt% (Fig. 2B) than do the phenocryst compositions, defining a differentiation trend like that observed by Le Bas (1962) for nonalkaline lavas. However, other groundmass compositions plot at higher Al(z) and TiO_2 wt% than do the phenocryst compositions, suggesting a differentiation trend like that observed by Le Bas (1962) for alkaline and peralkaline magmas.

Petrographic observations, trace element geochemistry (Christie et al., this volume), and the tight phenocryst correlation on Figure 1B all suggest that the Lo-En Guyot lavas belong to a single series (i.e., they were derived from essentially identical parental liquids). Thus, results on Figure 2B are difficult to interpret. However, the cluster of Lo-En Guyot groundmass compositions on the Fe+Mn, Mg, Ca+Na ternary (Fig. 3B) straddle the nonalkaline/alkaline boundary, suggesting an obvious interpretation: the Lo-En Guyot lavas are transitional between tholeiitic and alkalic. These lavas plot in the alkalic region of the SiO₂ vs. Na₂O+K₂O diagram (Christie et al., this volume); however, they lie very close to the tholeiitic/alkalic boundary and, as these lavas are somewhat altered, their Na and K abundances must be viewed as suspect. Thus, their position on the silica-alkalis diagram cannot be viewed as strongly diagnostic, especially as unaltered clinopyroxene compositions support an interpretation of their being transitional.

Wodejebato Guyot Lavas

Clinopyroxene is present in the altered, alkalic series lavas of each of the five sites (Sites 873-877) drilled on Wodejebato Guyot. Clinopyroxene compositions from the ankaramitic lavas of Site 874 are clearly more mafic, but phenocryst data from all sites lie along a well-defined fractionation trend in a plot of TiO₂ vs. Mg# (Fig. 1C), suggesting that the same lava series is present at all sites (see also Christie et al., this volume). The only exceptions are two low-Ti clinopyroxene phenocrysts from Sites 874 and 876. The Site 874 analysis is of a rim of a large, euhedral, zoned phenocryst; the Site 876 analysis is of a small, euhedral phenocryst. Neither of these grains is obviously a xenocryst. Therefore, it is difficult to explain why these two analyses plot below the overall Wodejebato Guyot trend on Figure 1C. On the Al(z) vs. TiO2 diagram (Fig. 2C), the Wodejebato Guyot clinopyroxene phenocryst compositions cover a long, narrow swath across the nonalkaline/alkaline boundary and do not completely fall into either of the Hawaiian fields. Note that this long, narrow swath is clearly a differentiation trend (from the TiO2 vs. Mg# relationship on Fig. 1C) paralleling the direction of alkaline and peralkaline magma differentiation shown on Figure 2B. Phenocrysts with compositions that plot below the Hawaiian alkalic series field crystallized from the mafic ankaramitic lavas of Hole 874B. As these lavas are more mafic than any of the Hawaiian alkalic series lavas and as this trend parallels the alkaline/peralkaline differentiation trend of Le Bas (1962), the interpretation that all Wodejebato Guyot lavas are alkalic is consistent with these compositions.

The Wodejebato Guyot groundmass clinopyroxene compositions cover a range similar to those from Lo-En Guyot on Figure 3B, although only one Wodejebato Guyot analysis falls significantly below the nonalkaline/alkaline boundary. Therefore, I conclude that the Wodejebato Guyot lavas are mildly alkalic because most of the clinopyroxene analyses plot in the alkaline field on Figure 3B and display an alkaline differentiation trend on Figure 2C. Those compositions that lie below the Hawaiian alkalic series field on Figure 2C are simply more mafic than any Hawaiian compositions reported in the literature (Fig. 1C) and do not indicate that the Wodejebato Guyot lavas are tholeiitic or transitional.

MIT Guyot Lavas

The lavas recovered from Site 878 (MIT Guyot) have been interpreted as subaerial basanitic and moderately fractionated alkalic series flows (Premoli Silva, Haggerty, Rack, et al., 1993; Christie et al., this volume). As with the Limalok Guyot lavas, the MIT Guyot lavas identified as basanitic have clinopyroxene grains with > 2 wt% TiO₂ (Figs. 1D, 2D), and most compositions have > 10% tetrahedral aluminum substitution and thus plot in the peralkaline field of Figure 2D. Also, the MIT Guyot analyses plot across the alkaline/peralkaline boundary of Figure 3, as do the Limalok Guyot basanitic groundmass clinopyroxene compositions.

The MIT Guyot lavas identified as belonging to the alkalic series are moderately fractionated. The clinopyroxene Mg#s range from 0.82 to 0.73, and there is little correlation with TiO₂ (Fig. 1D). Phenocryst compositions plot across both the nonalkaline/alkaline and alkaline/peralkaline boundaries on Figure 2D, but most fall in the alkaline field. The groundmass compositions straddle the nonalkaline/alkaline boundary on Figure 3C, but no analyses fall very far into the nonalkaline field. Collectively, these diagrams suggest that the nonbasanitic MIT Guyot lavas are correctly identified by Christie et al. (this volume) as essentially alkalic series lavas. However, some may be transitional in nature, and, probably, they were not derived from a single parental magma type. As there are long erosional episodes recorded in the igneous basement of Hole 878A, the latter is not a surprising result.

Takuyo-Daisan Guyot Lavas

A peperite formed by eruption of clinopyroxene-plagioclase-phyric basalt into unconsolidated volcaniclastic sediment was the only igneous unit recovered at Site 879. Clinopyroxene Mg#s range from 0.77 to 0.81, with low Al and Ti abundances (Figs. 1E, 2E). The single groundmass composition falls well below the nonalkaline/alkaline boundary on Figure 3B and even below the Lo-En Guyot transitional series field. The clinopyroxene compositions suggest that Hole 879A lava is a moderately fractionated transitional or tholeiitic basalt.

Comparisons Among Clinopyroxene Results of Individual Guyots

The basanitic lavas of Limalok and MIT guyots crystallized clinopyroxene grains of similar and rather remarkable chemistry (Table 3). These compositions are distinct from alkalic series lavas, having higher TiO₂ (usually > 2%) and higher Al (typically with > 10% tetrahedral aluminum substitution) (Figs. 1–3). Furthermore, the groundmass clinopyroxene compositions of the basanitic lavas from each guyot straddle the alkaline/peralkaline boundary on Figure 3. These compositions are clearly quite distinct from those of other Leg 144 lavas and may be diagnostic characteristics of clinopyroxene compositions from basanitic series lavas.

Alkalic series lavas were recovered from both MIT and Wodejebato guyots. Some lavas from Wodejebato Guyot have unusually mafic compositions, even when compared to a large set of Hawaiian lavas. The Wodejebato lavas appear to represent a large range of differentiation of a single parental magma type. Conversely, the MIT Guyot alkalic series lavas are all moderately fractionated and appear to have been derived from several distinct parental magma types (or possibly several distinct magma chamber depths; these two effects cannot be separated).

If only the phenocryst compositions are examined, Lo-En Guyot clinopyroxene grains form the tightest cluster of values on a plot of TiO_2 vs. Mg# (Fig. 1B), indicating that the sequence of erupted lavas sampled by drilling represents a narrow range of differentiation of a single parental magma type. This parental magma type was clearly distinct from the types represented by the Wodejebato and MIT guyot lavas. The clinopyroxene compositions indicate that the Lo-En Guyot lavas are transitional between alkalic and tholeiitic. Specifically, the phenocryst compositions have low TiO_2 (Figs. 1B, 2B), and the ground-mass compositions plot across the nonalkaline/alkaline boundary of Figure 3B.

The single lava from Takuyo-Daisan Guyot contains clinopyroxene phenocrysts and groundmass grains with compositions that are similar to those of Lo-En Guyot lavas. This lava may be transitional in nature, like the Lo-En Guyot lavas, but petrographic evidence (Christie et al., this volume) and plagioclase compositions (discussed below) cannot preclude a tholeiitic nature.

Feldspar

Feldspar compositions are diagnostic of the degrees of differentiation and silica-saturation of a lava. Detailed criteria were established for Hawaiian lavas by Keil et al. (1972); these are summarized here. Simply stated, tholeiitic lavas do not contain feldspar phenocrysts, microphenocrysts, or groundmass laths with Or > 1.1, and no tholeiitic feldspar compositions plot in the two-feldspar region or in the anorthoclase or sanidine regions of the An-Ab-Or ternary (Fig. 4A). Alkalic and "nephelinic" (equivalent to basanitic; Keil et al., 1972) lavas contain at least some groundmass laths and interstitial material (crystalline or glassy with feldspar compositions) of anorthoclase or sanidine compositions. Some alkalic lavas contain feldspar grains with compositions in the two-feldspar region (Fig. 4A).

As presented in Table 4 (and on the CD-ROM), definitions of phenocryst and microphenocryst are identical to those of Keil et al. (1972), but definitions of groundmass and interstitial are not equivalent. Specifically, I do not have an interstitial category, although "poikilitic in groundmass" is interstitial material as Keil et al. (1972) defined it. However, most analyses of groundmass grains (Table 4 and CD-ROM) are of laths because interstitial feldspar grains and glassy material in Leg 144 lavas have been altered to clay and chlorite minerals. Therefore, the criteria of Keil et al. (1972) may be used with little attention to differences in definitions of grain types.

Limalok Guyot Lavas

The basanitic lavas from Site 871 do not contain plagioclase phenocrysts. All compositions plotted on Figure 4A are of microlites and poikilitic groundmass grains. Values range from $An_{61.9}$ to $An_{25.2}$ and Or increases smoothly with decreasing An. That these lavas are alkalic is supported by an analysis of a microphenocryst with Or = 8.69 (Table 4, analysis number 12).

Despite thorough exploration, no feldspar of sanidine or anorthoclase composition was found during analysis. Either potassic feldspar did not crystallize in the groundmass or, as seems likely from petrographic observations, it completely altered to clay minerals. Whole-

MINERAL COMPOSITIONS OF LAVAS AND XENOLITHS



Figure 4. Orthoclase-albite-anorthite feldspar ternary diagrams. Fields on Figure 4A are from Deer et al. (1966).

rock K_2O contents are unusually low for basanitic lavas and low relative to the abundances of trace elements of similar incompatibility, suggesting significant losses by alteration (Christie et al., this volume).

Lo-En Guyot Lavas

The moderately differentiated transitional lavas recovered from Lo-En Guyot contain plagioclase and alkali feldspar phenocrysts with compositions that vary from $An_{82.5}$ to $An_{9.5}$ and from $Or_{0.3}$ to $Or_{48.3}$,

with a smooth trend toward increasing Or with decreasing An. Groundmass grains, primarily microlites, range from $An_{61.3}$ to $An_{11.2}$ and several of the analyses plot outside of the fields of equilibrium feldspar compositions (Fig. 4B) in the two-feldspar field. Both the high-Or phenocrysts and the microlites with two-feldspar compositions attest to the alkalic nature of these lavas. However, by their very nature, transitional lavas have some characteristics in common with tholeiitic lavas (e.g., Hawaiian transitional lavas often have mineral

Table 4. Representative plagioclase analyses.

| Hole: Core, section: Interval (cm): Description: | 871C 35R-1 14-16 Poikilitis in | 871C 38R-1 29-32 Polkilitic in | 871C 38R-6 52-55 Poikilitic in | 871C 40R-1 141-143 Micro | 872B 5R-1 76-78 Pim of large | 872B 5R-1 76-78 | 872B 5R-1 76-78 | 872B 6R-1 75-80 Ground- | 872B 6R-1 75-80 Micro- | 872b 7R-5 93–95 Ground- |
|---|---|---|---|-----------------------------------|---------------------------------------|-----------------------|-----------------------|----------------------------------|---------------------------------|----------------------------------|
| Description | groundmass | groundmass | groundmass | phenocryst | phenocryst; PL 1.3 | phenocryst; PL 1.2 | mass | mass | phenocryst (50 um) | mass |
| Analysis: Analysis no.: | PL 16 2 | PL 11 5 | PL 2 8 | PL 3 12 | PL 2 14 | PL 3 15 | PL 4 16 | PL 6 19 | PL 12 24 | PL 11 29 |
| Element (ppm): | | | | | | | | | | |
| SiO ₂ | 52.36 | 62.00 | 52.54 | 59.38 | 47.96 | 47.95 | 54.79 | 51.82 | 62.69 | 53.06 |
| TiO ₂ | 0.31 | 0.11 | 0.07 | 0.09 | 0 | 0.09 | 0.11 | 0.18 | 0.78 | 0.18 |
| Al ₂ Õ ₃ | 28.17 | 22.36 | 30.40 | 24.31 | 33.31 | 33.82 | 27.88 | 28.33 | 20.18 | 29.23 |
| FeO | 1.27 | 0.82 | 0.37 | 0.41 | 0.42 | 0.44 | 1.39 | 1.96 | 1.78 | 0.94 |
| MnO | 0.01 | 0.01 | 0.01 | 0.00 | 0.01 | 0.03 | 0.01 | 0.01 | 0.05 | 0.01 |
| MgO | 1.89 | 1.01 | 0.03 | 0.46 | 0.14 | 0.08 | 0.51 | 0.89 | 0.19 | 0.10 |
| CaO | 10.83 | 4.08 | 12.68 | 5.96 | 17.2 | 17.31 | 10.75 | 12.33 | 2.04 | 12.54 |
| Na ₂ O | 4.51 | 7.36 | 4.15 | 7.14 | 2.13 | 1.99 | 5.03 | 4.13 | 6.48 | 4.57 |
| K ₂ Ô | 0.30 | 2.24 | 0.25 | 1.51 | 0.09 | 0.06 | 0.7 | 0.27 | 6.48 | 0.32 |
| Total | 99.65 | 99.99 | 100.50 | 99.26 | 101.26 | 101.77 | 101.17 | 99.92 | 100.67 | 100.95 |
| Per 32 oxygens: | | | | | | | | | | |
| Si | 9.572 | 11.094 | 9.496 | 10.725 | 8.724 | 8.676 | 9.861 | 9.511 | 11.357 | 9.589 |
| AL | 6.069 | 4.716 | 6.475 | 5.175 | 7.141 | 7.212 | 5.914 | 6.128 | 4.309 | 6.225 |
| Fe | 0.194 | 0.123 | 0.056 | 0.062 | 0.064 | 0.067 | 0.209 | 0.301 | 0.270 | 0.142 |
| Mn | 0.002 | 0.002 | 0.002 | 0.000 | 0.002 | 0.005 | 0.002 | 0.002 | 0.008 | 0.002 |
| Mg | 0.515 | 0.269 | 0.008 | 0.124 | 0.038 | 0.022 | 0.137 | 0.244 | 0.051 | 0.027 |
| Ca | 2.121 | 0.782 | 2.455 | 1.153 | 3.352 | 3.356 | 2.073 | 2.425 | 0.396 | 2.428 |
| Na | 1.599 | 2.553 | 1.454 | 2.500 | 0.751 | 0.698 | 1.755 | 1.470 | 2.276 | 1.601 |
| K | 0.070 | 0.511 | 0.058 | 0.348 | 0.021 | 0.014 | 0.161 | 0.063 | 1.498 | 0.074 |
| Total | 20.185 | 20.065 | 20.013 | 20.099 | 20.092 | 20.061 | 20.126 | 20.167 | 20.270 | 20.112 |
| An | 55.97 | 20.33 | 61.89 | 28.82 | 81.28 | 82.50 | 51.97 | 61.27 | 9.50 | 59.18 |
| Ab | 42.18 | 66.38 | 36.66 | 62.48 | 18.21 | 17.16 | 44.00 | 37.14 | 54.59 | 39.03 |
| Or | 1.85 | 13.29 | 1.45 | 8.69 | 0.51 | 0.34 | 4.03 | 1.60 | 35.92 | 1.80 |

Note: Analysis numbers refer to complete data table on CD-ROM (back pocket).

Only a part of this table is reproduced here. The entire table appears on the CD-ROM (back-pocket).

assemblages identical to Hawaiian tholeiitic lavas (Macdonald and Katsura, 1964) and some characteristics in common with alkalic lavas. It is not unusual for feldspar chemistry to reflect the alkalic characteristics of transitional lavas (e.g., Keil et al., 1972).

Wodejebato Guyot Lavas

Plagioclase compositions from the alkalic series lavas of Sites 873-877 (Wodejebato Guyot) show a broad variability in degree of differentiation, similar to that shown by the clinopyroxene compositions. Site 874 lavas do not contain plagioclase phenocrysts, but the microlite compositions are quite anorthitic (An_{78.3} to An_{64.5}). These indicate crystallization from a relatively primitive liquid (Fig. 4C). The Site 877 lava is similarly undifferentiated, with phenocryst compositions ranging from An_{87.7} to An_{68.8} and microlite compositions ranging from An_{67.7} to An_{63.3}. Site 873 phenocrysts range from An_{68.1} to An60.0 and microlites range from An56.9 to An35.2. This lava is more fractionated than lavas from Sites 874 and 877. The Site 875 compositions, from An59.4 to An40.3, indicate that the liquid is moderately fractionated, similar to the Site 873 lava. Site 876 compositions cover a broader range (An_{67.9} to An_{21.8}) and include two analyses that plot in the two-feldspar field. Note that only groundmass grains were analyzed from the Sites 875 and 876 lavas.

Feldspar compositions that demonstrate the alkalic nature of the Wodejebato Guyot lavas were analyzed for lavas from four of the five holes. Hole 873A lavas have phenocrysts and microphenocrysts with Or > 1.1. Holes 875C and 876A lavas have groundmass laths with Or > 1.1. The Hole 877A lava has similar groundmass and phenocryst rim compositions; and, although none of the analyses has Or > 1.1, similar phenocryst rim and groundmass compositions are in themselves evidence of crystallization from an alkalic liquid (Keil et al., 1972). No analyses for the Hole 874B ankaramitic lavas demonstrate an alkalic nature for these lavas, but the groundmass material is so altered that few analyses were done. Furthermore, the very lack of plagioclase phenocrysts when clinopyroxene phenocrysts occur in

such huge abundances suggests that plagioclase was not yet on the liquidus, a crystallization sequence characteristic of alkalic liquids.

MIT Guyot Lavas

The moderately differentiated, alkalic series lavas recovered from MIT Guyot contain plagioclase phenocrysts with compositions ranging from An_{81.8} to An_{58.9} (Fig. 4D). This apparently small range for such a petrographically diverse suite of lavas (see Premoli Silva, Haggerty, Rack, et al., 1993) probably indicates that the lavas underwent similar degrees of differentiation. Of the lavas identified as alkalic series, analyses of groundmass laths with Or > 1.1 exist for igneous Units 1, 13, and 23, the lava of igneous Unit 27 crystallized groundmass grains of anorthoclase composition, and the lava of igneous Unit 34 crystallized microphenocrysts with Or > 1.1 (see Table 4 and CD-ROM). From the criteria of Keil et al. (1972), it is clear that those MIT Guyot units identified as alkalic series lavas did crystallize feldspar of distinctly alkalic character.

Those MIT Guyot lavas identified as basanitic crystallized groundmass feldspar grains with compositions that range from $An_{67.1}$ to $An_{9.6}$ and from $Or_{1.0}$ to $Or_{45.4}$. These basanitic lavas contained few plagioclase phenocrysts, but all analyses have compositions of Or > 1.1 and one analysis plots in the two-feldspar field of the An-Ab-Or ternary at $Or_{35.3}$. These lavas are clearly alkalic in nature.

Takuyo-Daisan Guyot Lavas

The clinopyroxene-plagioclase-phyric basalt recovered from Takuyo-Daisan Guyot contains plagioclase phenocrysts with compositions ranging from An_{80.7} to An_{53.4} (Fig. 4E). The only two analyzed groundmass grains are within the range of the phenocryst compositions, An_{56.3} and An_{54.8}. One analysis of a phenocryst rim is as potassic (Or_{2.4}) as the groundmass laths (Or_{2.3-3.8}), suggesting (on meager evidence) that this lava cannot be tholeiitic. The hypothesis that this lava is transitional is not precluded by the feldspar data.

Summary of Leg 144 Feldspar Results

Feldspar analyses of the Leg 144 lavas demonstrate (merely suggest in the case of Takuyo-Daisan Guyot) that all of the lavas are alkalic or transitional in nature. The occurrences of phenocrysts, microphenocrysts, and groundmass laths with Or > 1.1, of groundmass grains with anorthoclase, sanidine, or two-feldspar compositions, and the phenocryst rims and groundmass grains with similar Or components support this conclusion.

Trace Element Chemistry

Ion microprobe data were collected on 4 microphenocrysts and 23 phenocrysts of clinopyroxene representing all Leg 144 sites except Sites 875 and 880. These data are presented in Table 5 and Figures 5–7.

Phenocrysts

The locations of the 23 clinopyroxene phenocryst analyses were carefully selected on the basis of major element data. Evaluation of the phenocryst data is difficult because many factors control the abundances of trace elements in clinopyroxene. First, both the chemistry of the mantle source and the degree of partial melt affect the trace element abundances in magma and, obviously, the abundance of an element in a magma has some affect on the abundance of that element in clinopyroxene crystallized from the magma. Second, as a magma differentiates, the abundances of incompatible trace elements increase and this is reflected in the clinopyroxene compositions. Third, both major element chemistry of a magma and the depth of crystallization of a phenocryst control the clinopyroxene/liquid partition coefficients of incompatible elements. (Specifically, this has been demonstrated for rare earth elements [REE] and Y; we may reasonably assume that other incompatible elements are similarly affected [Hack et al., in press].) These various effects are difficult and sometimes impossible to separate. Consequently, the following discussion is a very simplified interpretation of these data.

The data plotted by guyot on chondrite-normalized spider diagrams (Fig. 5) show several interesting features.

1. Phenocrysts from basanitic lavas, as analyzed on thin section Samples 144-871C-35R-1, 37–39 cm, 144-871C-38R-1, 29–32 cm, and 144-878A-89R-2, 42–43 cm, contain greater abundances of incompatible elements than phenocrysts that crystallized from transitional and alkalic series lavas. The separation between basanitic phenocrysts and alkalic series phenocrysts occurs at approximately ten times chondritic values for the light rare earth elements (LREE) (Fig. 5). This, in part, reflects the lower degree of partial melting commonly postulated for basanitic series lavas when compared to alkalic series lavas.

A clear distinction exists between the Limalok and MIT guyot basanitic phenocrysts. The Limalok Guyot phenocrysts have higher LREE and lower heavy rare earth elements (HREE) than the MIT Guyot basanitic phenocrysts (cf. Fig. 5A, D). This difference is also illustrated on a plot of La/Yb vs. Ti (Fig. 6), which shows that the Limalok phenocrysts have higher La/Yb values. Assuming that all the basanitic lavas were derived from equivalent degrees of partial melt, and that all microphenocrysts were crystallized during or after eruption (1 bar of pressure), differences in pressure of crystallization and of degree of partial melt cannot explain this difference. Furthermore, this difference is clearly not caused by fractionation because the Mg#s of the phenocrysts in these lavas are essentially equivalent. Therefore, it appears that the source of the Limalok Guyot basanitic lavas was more enriched than the source of the MIT Guyot basanitic lavas. This is verified by whole-rock chemistry (Christie et al., this volume) and by isotopic ratios (Koppers et al., this volume). Specifically, the Limalok Guyot basanitic lavas have higher Nb/Zr ratios and higher 87Sr/86Sr.

2. The transitional and alkalic series phenocrysts from Lo-En, MIT (thin section Sample 144-878A-23R-1, 6–8 cm), and Takuyo-Daisan guyots have very similar abundances of REE and other trace elements and display convex-upward patterns with slight, positive Ti anomalies (Fig. 5B, D, E). This similarity is rather remarkable because (1) presumably the MIT Guyot lavas were derived from smaller degrees of partial melting than the others; (2) isotopic data indicate that lavas from each of these guyots were derived from distinct mantle sources; and (3) there is no reason to suppose that the depths of crystallization were equivalent, especially as these volcanoes are of quite different sizes (Premoli Silva, Haggerty, Rack, et al., 1993).

3. The alkalic series clinopyroxene phenocrysts from Wodejebato Guyot display a wide range of trace element abundances. Phenocrysts in Sample 144-874B-24R-1, 12-16 cm, have as low or lower abundances of trace elements than do phenocrysts of the transitional lavas of Lo-En Guyot (cf. Fig. 5B, C; see Fig. 6). This may be explained, at least in part, by differences in degrees of differentiation as the Lo-En Guyot lavas are moderately fractionated and the Hole 874B lavas are relatively primitive (cf. Fig. 1B, C), by differences in source enrichment, or by differences in pressure of crystallization. However, no significant differences between the Lo-En Guyot lava source and the Wodejebato Guyot lava source are suggested by the whole-rock and isotopic data (Christie et al. and Koppers et al., both in this volume). In fact, Lo-En Guyot lavas have, on average, slightly lower 87Sr/86Sr and 143Nd/144Nd ratios and lower Pb isotopic ratios (Koppers et al., this volume). This suggests that the Lo-En Guyot mantle source had lower abundances of incompatible trace elements than did the source of the Wodejebato Guyot lavas. The reasons for the low trace element abundances in clinopyroxene phenocrysts of Sample 144-874B-24R-1, 12-16 cm, cannot be clearly established.

However, the low trace element abundances in clinopyroxene phenocrysts of Sample 144-874B-24R-1, 12-16 cm, are unique among the Wodejebato Guyot samples (Fig. 6). Analyses of other clinopyroxene phenocrysts show very high abundances of trace elements in comparison with phenocrysts of other Leg 144 transitional and alkalic series lavas (Figs. 5, 6). Samples 144-874B-24R-4, 11-15 cm, and 144-877A-20R-5, 22-24 cm, are relatively primitive (but slightly more differentiated than the lava sampled in Sample 144-874B-24R-1, 12-16 cm). Other Wodejebato lavas are only moderately fractionated. Therefore, fractionation cannot explain why most of the Wodejebato phenocrysts have higher abundances of trace elements than do phenocrysts from the Lo-En, MIT, and Takuyo-Daisan guyot lavas. With the exception of Sample 144-874B-24R-1, 12-16 cm, the Wodejebato Guyot clinopyroxene phenocrysts also have higher La/Yb values than do other Leg 144 transitional and alkalic lavas (Fig. 6). Again, it is difficult to evaluate the significance of the high trace element abundances and high La/Yb in the Wodejebato phenocrysts, but differences in pressure of crystallization, degree of partial melting, and mantle source chemistry are the most likely explanations.

The ankaramitic lavas recovered from Site 874 were classified as a single igneous unit (a single flow) by D.M. Christie and J.J. Dieu on board ship (Premoli Silva, Haggerty, Rack, et al., 1993). There was no evidence (i.e., modal changes, brecciated flow tops) for indicating more than one flow. However, it is clear from the ion microprobe data (cf. Samples 144-874B-24R-1, 12–16 cm, and 144-874B-24R-4, 11–15 cm, in Table 5; see also Figs. 5C, 6) and from the oxide chemistry (Gee, this volume) that at least two flows were recovered. The upper of these lavas, represented by Sample 144-874B-24R-1, 12–16 cm, was apparently derived from a less enriched source than the sources of the lower lava and the Lo-En, MIT, and Takuyo-Daisan transitional and alkalic lavas.

Microphenocrysts

Four clinopyroxene microphenocrysts were analyzed for REE and other trace elements to understand better the fractionation processes affecting the liquid compositions of Leg 144 lavas before and during eruption (Table 5). All four are well-formed equant crystals less than 50 μ m in diameter. On a diagram of La vs. Yb (Fig. 7), the analyses from Sample 144-878A-89R-2, 42–43 cm, provide an integrated picture of the processes that led to the elevated trace element concentra-

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Figure 5. Ion microprobe analyses of clinopyroxene phenocrysts. Chondrite-normalizing values from Evensen et al. (1978). Elements are plotted with increasing incompatibility to the left.

tions in the microphenocrysts. As crystallization progressed from the formation of phenocryst rims to microphenocryst cores to microphenocryst rims, there was a large increase in the abundances of both La and Yb, reflecting extensive fractionation of the liquid as the lava erupted and chilled. Within the limits of variability of clinopyroxene chemistry, fractionation of clinopyroxene or olivine cannot be distinguished on Figure 7, but petrographic evidence suggests that subequal proportions of these two phases were crystallizing as the lava chilled. Using reasonable partition coefficients (Table 6), the calculated extent of fractionation from crystallization of the phenocryst rim to crystallization of the microphenocryst rim is roughly 65%. (Note: There is no reason, beyond the plausibility of this explanation for the trend observed in Fig. 7, to indicate that the rim of the phenocryst crystallized before the crystallization of the microphenocrysts.) The tie line on Figure 7 from phenocryst rim to microphenocryst for Sample 144-871C-38R-1, 29-32 cm, closely parallels the tie-line sequence for Sample 144-878A-89R-2, 42-43 cm. Again, petrographic evidence suggests that subequal proportions of clinopyroxene and olivine were fractionating from the liquid. Roughly 50% total fractionation between crystallization of the rim and the microphenocryst is indicated. Ion microprobe analysis was not performed on a phenocryst from igneous Unit 1 of Site 871. The composition of a microphenocryst from this lava is plotted (Sample 144-871C-35R-1, 37–39 cm).

Large oscillations in trace element concentrations in single phenocrysts have been observed: (1) Shimizu and le Roex (1986) attributed oscillations in augite phenocrysts in alkaline basalts from Gough Island as arising from repeated magma injections into the chamber; and (2) Shimizu (1990) suggested that oscillations in augite phenocrysts in an alkaline basalt from Papua New Guinea may be caused by nonequilibrium kinetics of reaction-transport of the liquid around the growing crystals. These oscillations may be common in phenocrysts (Shimizu, 1990), and, therefore, the core-to-rim decrease in both La and Yb in Sample 144-871C-38R-1, 29–32 cm, and in La in Sample 144-878A-89R-2, 42–43 cm, is probably not unusual but merely the result of too few analyses of a complex system.



Clinopyroxene- and olivine-controlled fractionation trends are shown. Analyses labeled "core" and "rim" are phenocryst cores and rims; "micro-core" and "micro-rim" are microphenocryst cores and rims; "microphenocryst" represents analyses of cores of very small grains. Tie lines connect analyses of the same lava.

Comparisons of Figures 8A and 9A with Figures 8B and 9B, respectively, show that the clinopyroxene in the Type 1 xenolith from MIT Guyot has higher Mg#s, somewhat lower TiO₂ (although clinopyroxene compositions from each guyot have so little Ti that analytical error is more significant than differences between the clinopyroxene compositions), and significantly lower tetrahedral Al (also overall lower Al₂O₃; see Table 3) than the clinopyroxene in the Type 1 xenolith from MIT Guyot represents more severely melted material than the Type 1 xenoliths from Limalok Guyot.

A goal inherent in the analysis of Type 2 xenoliths is to characterize the liquids that differentiated, forming the cumulate material. Evaluation of mineral data from cumulates is very difficult because depth of crystallization and rate of cooling can drastically affect the major element chemistry of all phases. For example, the Salt Lake Crater cumulate xenoliths (Hawaiian hotspot; Sen, 1988) and the Ua Huka cumulate xenoliths (Marquesas hotspot; J.J. Dieu and J.H. Natland, unpubl. data, 1993) probably crystallized from very similar liquids (i.e., many lavas of the Honolulu Volcanics [the magmas known to have formed the Salt Lake Crater pyroxenites] [Sen, 1988]) and are essentially identical to basanitic and alkalic series lavas on Ua Huka Island (thought to be the magmas that crystallized to form the Ua Huka cumulates [J.J. Dieu and J.H. Natland, unpubl. data, 1993]) (cf. Clague and Frey, 1982; Diraison, 1991). Yet there are very significant differences in the clinopyroxene compositions of each xenolith locality. Salt Lake Crater clinopyroxene grains have lower TiO2 at an equivalent Mg# (Fig. 8) and only overlap the lower half of the Al(z) range (Fig. 9) observed for Ua Huka clinopyroxene grains. Aluminum preferentially moves out of the tetrahedral sites and into the octahedral sites in clinopyroxene under high pressures during solid-solid

Figure 6. Diagram of La/Yb vs. Ti (ppm) for ion microprobe analyses of phenocrysts. Dashed lines connect analyses of basanitic lavas. Solid lines connect analyses from the two Hole 874B ankaramitic samples.

Hole 879A

Hole 878A alkalic series

Hole 878A basanitic series

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ULTRAMAFIC XENOLITHS

Ultramafic xenoliths recovered from Sites 871 and 878 (Limalok and MIT guyots, respectively) include two types: Type 1 xenoliths interpreted to be samples of mantle residue and Type 2 xenoliths interpreted to be samples of cumulate material (after Frey and Prinz, 1978). Major element analyses by electron microprobe of unaltered mineral phases are presented in Tables 2 and 3. Trace element analyses by ion microprobe of representative clinopyroxene grains are presented in Table 5.

Major Element Chemistry

Clinopyroxene

The clearest distinction between clinopyroxene compositions of Type 1 and 2 xenoliths is in their TiO_2 contents (Figs. 8, 9). Type 1 materials have experienced severe melt extraction that has depleted them in Fe, Ti, and other "incompatible" elements (see discussion of Type 1 xenoliths under "Trace Element Chemistry" this section). Clinopyroxene grains in Type 1 xenoliths from both Limalok and MIT guyots have Mg#s that cluster in a small range (0.895–0.93) and TiO₂ values of < 0.20%. This is typical of mantle residue materials sampled by hotspot volcanism (e.g., Sen, 1988; Hauri et al., 1993; J.J. Dieu and J.H. Natland, unpubl. data, 1993). Type 1 analyses from both guyots plot near the field of Samoan mantle residue drawn on Figure 8.

| Hole: Core, section: Interval (cm): Description: | 871C 35R-1 10–12 Alkalic cumulate xenolith grain | 871C 35R-1 10-12 Alkalic cumulate xenolith grain | 871C 35R-1 20–23A Mantle residue xenolith grain | 871C 35R-1 20–23A Mantle residue xenolith grain | 871 35R 37– Alka cumu xeno gra | C 871C t-1 35R-1 39 37-39 alic Ground- late mass lith in | 871C 38R-1 29-32 Core of 1.0-cm xenocryst; CPX B | 871C 38R-1 29–32 Rim of 1.0-cm xenocryst; CPX A | 871C 38R-1 29–32 Ground- mass | 872B 5R-3 114-115 Core of 6-mm, round, spongy | 872B 9R-5 107–109 Core of large, spongy xenocryst; see CPX 11 | 872B 9R-5 107–109 Core of large spongy xenocryst; see CPX 11 | 873A 18R-1 30–33 c, Core of small phenocryst; see CPX 3, 4 analyses |
|--|--|---|---|---|--|---|--|---|--|---|---|--|---|
| Analysis: Analysis no.: | CPX A | CPX B | CPX A | СРХ В 4 | CPX 5 | CA CPX B | CPX A | CPX B 8 | CPX C 9 | CPX A 10 | analysis CPX A 11 | CPX B 12 | CPX A 13 |
| Element (ppm): Ti V Cr Sr Y Zr | 6640 300 1010 194 15.3 140 | 7240 310 1010 200 15.6 150 | 700 230 5300 210 7.9 1.3 | 720 240 5560 300 8.4 5.7 | 22790 370 2790 124 19. 240 | 10410 370 97 223 9 19.8 120 | 26320 450 190 145 28.9 230 | 10030 210 1720 88 13.3 140 | 22210 460 107 135 35.5 450 | 4680 310 5480 44.5 9.3 28.1 | 6260 320 5190 48.2 9.9 36.0 | 6080 310 5620 47.5 10.4 33.8 | 12110 490 1940 63.0 21.8 117 |
| La Ce Nd Sm Eu Dy Er Yb | 31.1 73.9 36.4 7.52 2.31 4.08 2.36 1.53 | 9.5 24.4 19.7 4.04 1.42 2.29 1.26 0.83 | 37.7 48.8 6.0 0.49 0.23 1.17 0.95 0.72 | 41.0 65.2 13.7 1.30 0.51 1.41 1.10 0.81 | 57 125 65 14 4 8 4 2 | .7 18.2 .5 48.1 .6 29.3 .1 6.63 .72 2.77 .07 4.47 .39 2.24 .87 1.69 | 9.37 31.5 31.0 9.31 3.13 6.72 4.3.41 0.2.51 | 8.04 24.9 23.8 7.23 2.27 5.22 2.69 2.09 | 15.6 40.3 33.2 8.65 2.81 6.32 3.04 2.34 | 0.82 2.72 3.91 1.50 0.61 1.52 0.60 0.61 | $\begin{array}{c} 1.31 \\ 4.51 \\ 5.37 \\ 1.62 \\ 0.79 \\ 1.94 \\ 0.84 \\ 0.64 \end{array}$ | 1.39 4.42 6.06 1.87 0.70 1.71 0.90 0.74 | 2.33 8.06 10.8 3.38 1.56 3.42 1.50 1.31 |
| Hole: Core, section: | 874B 24R-1 | 874B 24R-1 | 8741 24R- | B -1 | 874B 24R-4 | 874B 24R-4 | 876A 17R-1 | 877A 20R-5 | 877A 20R-5 | 878A 23R-1 | 878 23R | A -1 | 878A 81R-2 |
| Interval (cm): Description: Analysis: Analysis no.: | 12–16 Core of small phenocryst CPX A 15 | 12–16 Near edge of 1.0-cm, euhedral spongy phenocryst; CPX C CPX B 16 | 12–1 Core of I euhedral, phenoc CPX CPX 17 | 6 .0-cm, spongy ryst; B C | 11–15 L-shaped, skeletal phenocryst CPX A 18 | 11–15 5.0-mm resorbed xenocryst with plagioclase inclusions CPX B 19 | 56-61 Small crystal in clot; see CPX 3 analyses CPX A 20 | 22-24 Core of large phenocryst CPX A 21 | 22-24 Core of large phenocryst CPX B 22 | 6-8 Rim of resorbed xenocryst; CPX B CPX A 23 | 6– Core resort xenocr CPX CPX 24 | of bed yst; A B | 37–40 Tholeiitic cumulate xenolith grain CPX A 25 |
| Element (ppm): Ti V Cr Sr Y Zr | 8230 380 6270 43.7 13.5 52.6 | 2910 200 6570 26.0 6.0 11.1 | 3660 230 6300 34. 7. 16. | 6 0 8 | 7940 430 5500 45.7 14.1 65.4 | 10430 550 2690 41.2 20.8 71.2 | 6810 330 2710 53.2 11.1 53.0 | 4680 300 1890 31.1 8.2 23.8 | 5960 360 3850 34.3 11.0 32.7 | 8900 330 3250 42.6 14.1 46.3 | 7190 270 2160 41 12 30 | .5 .1 .4 | 203 79.5 700 46.7 0.6 12.2 |
| La Ce Nd Sm Eu Dy Er Yb | 0.87 3.26 4.25 1.60 0.61 1.81 0.82 0.74 | 0.41 1.29 2.31 1.07 0.37 0.59 0.25 0.39 | 0. 1. 1. 0. 0. 0. 0. 0. 0. 0. | 47 50 99 80 37 96 56 38 | 3.01 10.7 9.17 3.03 1.09 2.80 1.65 1.05 | 3.20 11.4 9.91 3.20 1.39 3.99 1.74 1.37 | 3.29 10.7 8.24 3.00 0.95 2.05 1.21 1.04 | 2.62 8.76 7.14 2.99 1.11 2.62 1.16 1.05 | 2.06 6.40 5.84 2.66 0.83 2.37 1.10 0.97 | 1.34 5.17 5.73 2.77 1.14 2.77 1.33 1.13 | 1 7 3 1 3 1 1 | .87 .11 .00 .04 .10 .30 .65 .27 | $\begin{array}{c} 0.31 \\ 0.74 \\ 0.59 \\ 0.38 \\ 0.14 \\ 0.42 \\ 0.49 \\ 0.43 \end{array}$ |

| 879A 22R-1 133-135 Near edge of resorbed grain attached to plagioclase phenocryst, CPX A: see CPX 1, 2 analyses CPX 1, 2 analyses | 8790 8790 2450 13.9 44.9 41.97 41.97 41.97 41.97 41.97 2.32 2.32 2.32 2.32 2.05 1.06 0.79 |
|---|---|
| 879A 22R-1 133-135 Core of resorbed grain attached to plagioclase phenocryst. CPX 8: see CPX 1, 2 analyses CPX 1, 2 analyses 36 | 6390 550 1690 42.3 25.4 12.8 25.4 1.30 5.07 5.07 5.07 5.07 5.07 5.07 5.07 5.0 |
| 878A 91R-1 88-89 Mantle residue xenocryst grain CPX A 35 | 1880 60.2 54.6 77.0 170 170 125 93.8 28.8 28.8 28.8 28.8 28.8 28.8 28.8 2 |
| 878A 90R-6 63-64 Mantle residue xenolith grain CPX B 34 | $\begin{array}{c} 3830\\ 3830\\ 110\\ 110\\ 493\\ 493\\ 493\\ 493\\ 493\\ 30.8\\ 30.8\\ 33.3\\ 33.3\\ 33.3\\ 7.77\\ 7.77\\ 7.77\\ 14.3\\ 11.7\end{array}$ |
| 878A 90R-6 63-64 Mantle residue residue xenolith grain CPX A 33 | 300 103 46.7 46.7 390 32.5 32.5 116 32.5 36.5 36.5 36.5 116 11.8 |
| 878A 90R-5 1111-116 Tholeitic cumulate cumulate crain CPX B 32 | 7950 400 400 69.1 13.5 46.9 46.9 4.84 1.57 13.5 13.5 4.84 1.67 1.57 1.24 |
| 878A 90R-5 1111-116 Tuholeitic Tuholeitic xenolith grain CPX A 31 | 12340 390 3470 19.4 68.9 68.9 3.12 9.22 8.59 3.14 1.07 1.10 |
| 878A 878-2 42-43 Micro- phenocryst CPXD 30 | 10430 1265 1265 1265 186.0 86.2 86.2 150.0 150.0 27.3 26.4 10.1 3.56 7.17 7.83 26.4 2.96 |
| 878A 89R-2 42-43 Micro- phenocryst CPX C 29 | 8080 214 23.7 23.5 70.5 54.0 54.0 54.0 36.9 36.9 3.46 3.46 3.46 |
| 878A 89R-2 42-43 Core of 2-mm 2-mm CPX A CPX A 28 28 | 14090 320 37.1 97.1 195.0 195.0 8.55 21.6 2.65 5.80 5.80 5.80 5.80 5.80 5.80 5.80 5.8 |
| 878A 89R-2 42-43 Rim of 2-mm 2-mm phenocryst; CPX B CPX A 27 | 11760 300 300 96.4 16.1 76.7 76.7 25.0 9.18 9.18 9.18 9.18 3.49 2.76 |
| 878A 81R-2 37-40 Tholeittic cumulate xenolith grain CPX B 26 | $\begin{array}{c} 590\\ 5104\\ 7.7\\ 7.7\\ 2.9\\ 4.8\\ 0.27\\ 0.24\\ 0.71\\ 0.24\\ 0.43\\ 0.43\\ 0.24\\ 0.24\\ 0.24\\ 0.24\end{array}$ |
| Hole: Core, section: Interval (em): Description: Analysis: Analysis no.: | Element (ppm): Ti Cr Cr Cr Cr Cr Cr Cr Cr Cr Sr Yr Dy Dy Dy Fr Tr Yb |

Table 5 (continued)

equilibration (e.g., this may affect cumulate grains but not phenocrysts) (Le Bas, 1962). Because silica must diffuse into clinopyroxene for this to occur and, thus, the tetrahedral charge deficit is diminishing, octahedral titanium must be diffusing or exsolving from the clinopyroxene. The Salt Lake Crater pyroxenite xenoliths are thought to have crystallized at or near the base of the Hawaiian lithosphere, at pressures of 25 kbar or less (Sen, 1988). The Ua Huka cumulate xenoliths must have crystallized at much shallower depths, certainly <10 kbar, because they contain plagioclase and fluid inclusions (i.e., vesicles) (J.J. Dieu and J.H. Natland, unpubl. data, 1993). This difference in the depths of equilibration apparently created the differences we observe in the clinopyroxene compositions (Figs. 8, 9).

The clinopyroxene compositions of the Type 2 xenoliths from MIT Guyot are distinctly different from those from Limalok Guyot. On average, the MIT Guyot clinopyroxene compositions have lower TiO_2 and Al_2O_3 [reflected in Al(z)] at equivalent Mg# (Figs. 8, 9). This suggests, assuming that they were crystallized at similar pressures, that the MIT Guyot cumulates crystallized from a different magma type (tholeiitic?) than did the Limalok Guyot cumulates (alkalic?).

I argue that, in fact, the MIT Guyot cumulates did crystallize from tholeiitic liquids on the following two lines of evidence: (1) almost all of the clinopyroxene compositions plot at lower Al(z) than the Salt Lake Crater and Ua Huka clinopyroxene compositions (Fig. 9B), and it seems more likely that the MIT Guyot cumulates are of tholeiitic origin than that they crystallized at depths much greater than 75 km; and (2) the MIT Guyot cumulates contain orthopyroxene (discussed below). The thickness of the lithosphere at the time MIT Guyot was an active volcano is unknown, but it seems unlikely that it was much greater than 75 km (little lithosphere in the ocean basins is), and the MIT Guyot clinopyroxene grains also have much lower concentrations of octahedral Al than do the Salt Lake Crater clinopyroxene grains. The latter demonstrates that the low Al(z) of the MIT Guyot clinopyroxene is not the result of the same process that affected the Salt Lake Crater clinopyroxene.

It seems reasonable to hypothesize that the Limalok cumulate xenoliths were derived from alkalic liquids similar to those that differentiated to form the Ua Huka and Salt Lake Crater cumulates (Figs. 8A, 9A).

Orthopyroxene

Unaltered orthopyroxene was analyzed in three xenoliths from thin section Samples 144-871C-35R-1, 20-23 cm (Xenolith A), 144-878A-90R-6, 63-64 cm, and 144-878A-91R-3, 79-81 cm. Representative analyses are presented in Table 3. The orthopyroxene analyses of the Limalok Guyot sample (Site 871) have a mantle residue signature with moderate Al₂O₃ (almost 4%) and a high En component (>88%). These compositions are typical of hotspot mantle residue (e.g., Sen, 1988; Hauri et al., 1993; J.J. Dieu and J.H. Natland, unpubl. data, 1993) and can be compared with an orthopyroxene analysis from a Samoan mantle residue xenolith (Table 7). The orthopyroxene grains analyzed in two MIT Guyot xenoliths (Site 878) have very different compositions from the Limalok Guyot orthopyroxene grains. The abundances of Al₂O₃ and the En components of the MIT Guyot orthopyroxene grains are much lower (<2% and <82%, respectively). The En values are typical of orthopyroxene compositions from the Salt Lake Crater pyroxenites; however, as observed for the clinopyroxene chemistry, the MIT Guyot orthopyroxene grains have lower Al relative to the Salt Lake Crater grains (cf. data in Tables 3, 7). The presence of orthopyroxene in the MIT Guyot cumulate xenoliths supports the hypothesis of tholeiitic origins for the MIT Guyot Type 2 xenoliths. Specifically, alkalic cumulates are rarely observed to have orthopyroxene (D.A. Clague, pers. comm., 1994), as predicted by experimental research that finds orthopyroxene on the liquidus of alkalic liquids only in very small regions of pressure-temperature space (Takahashi, 1979/80, 1980). At least much of the orthopyroxene in the Salt Lake Crater xenoliths exsolved from clinopyroxene (Sen, 1988). Thus, the presence of orthopyroxene in this one set of



Figure 8. TiO₂ vs. Mg/(Mg+Fe) of clinopyroxene grains in xenoliths. A. Limalok Guyot xenoliths. B. MIT Guyot xenoliths. Fields drawn from data in Sen (1988) for Salt Lake Crater xenoliths and from unpublished data of J.J. Dieu and J.H. Natland (1993) for Ua Huka xenoliths.

alkalic cumulate xenoliths does not undermine the hypothesis that the MIT Guyot cumulates were derived from tholeiitic liquids.

Trace Element Chemistry

Type 1 Xenoliths

As previously discussed in the "Major Element Chemistry" section (this chapter), Type 1 (mantle residue) xenoliths are presumed to



Figure 9. Al(z) vs. TiO_2 of clinopyroxene grains in xenoliths. **A.** Limalok Guyot xenoliths. **B.** MIT Guyot xenoliths. Boundary lines from Le Bas (1962). Fields from sources listed in Figure 8 caption.

TiO2 %

be pieces of a mantle source that was partially melted before being sampled as a xenolith. Such xenoliths erupted at an oceanic hotspot are often samples of oceanic lithosphere (e.g., Sen, 1988; K.A. Farley, unpubl. data, 1994).

This extreme amount of partial melting should be reflected in low incompatible trace element abundances. On an extended REE diagram (Fig. 10A), data for a clinopyroxene in a Type 1 xenolith from Tahiti display the low abundances and pattern representative of mantle residue. However, xenoliths from the Southern Hemisphere oceanic hotspots seldom display these expected patterns. In many cases, some process subsequent to melting has elevated the abundances of

MINERAL COMPOSITIONS OF LAVAS AND XENOLITHS

Table 6. Constants used in percolation model.

| Partition coefficients | CPX | OPX | OL | SP | Equilibrium melt values (CN) |
|---------------------------------|---------------------|---------------------|---------------------|---------------------|------------------------------------|
| La | 0.045 | 0.0015 | 0.0006 | 0.0003 | 3500 |
| Ce | 0.075 | 0.0026 | 0.003 | 0.0005 | 1600 |
| Sr | 0.095 | 0.001 | 0.0028 | 0.0006 | 570 |
| Nd | 0.1 | 0.009 | 0.003 | 0.0008 | 330 |
| Zr | 0.15 | 0.015 | 0.003 | 0.003 | 20 |
| Sm | 0.2 | 0.012 | 0.003 | 0.0009 | 80 |
| Eu | 0.2 | 0.012 | 0.003 | 0.0009 | 60 |
| Ti | 0.34 | 0.1 | 0.01 | 0.005 | 5 |
| Dy | 0.35 | 0.054 | 0.006 | 0.0015 | 24 |
| Y | 0.37 | 0.0481 | 0.007 | 0.0025 | 3.4 |
| Er | 0.37 | 0.076 | 0.008 | 0.003 | 14 |
| Yb | 0.37 | 0.11 | 0.009 | 0.0045 | 8 |
| Diffusion coefficient (cm/a) | 1.00 ⁻¹³ | 1.00 ⁻¹³ | 1.00 ⁻¹² | 1.00 ⁻¹³ | |
| Modal abundance (%) | 1 | 15 | 82 | 2 | |
| Grain size (cm) | 0.2 | 1.5 | 2.5 | 0.1 | |

Notes: Porosity = 1%; percolation distance = 100 m; percolation speed = 1 cm/a. The equilibrium melt values are chondrite normalized (CN). Partition coefficients are within the ranges specified by Arth (1976).

the large ion lithophile elements (LILE) without obviously modifying the modal abundances and major element chemistry of the residue. A process with this signature is called "cryptic metasomatism" (Dawson, 1984; Zindler and Jagoutz, 1988). Extended REE patterns like those plotted on Figure 10 (B, D, E, F) from Limalok Guyot, Savai'i Island (Samoa), MIT Guyot, and Tubuai Island (Cook-Australs), respectively, are typical of Type 1 clinopyroxene patterns from the Samoan, Marquesan, Cook-Austral, and Society hotspot chains (Hauri et al., 1993; J.J. Dieu and J.H. Natland, unpubl. data, 1993) and from the Kerguelen plume (Hassler et al., 1993).

Two distinct extended REE patterns of Type 1 clinopyroxene are observed from the Southern Hemisphere hotspots. The first type of pattern is "spoon-shaped" (Sen et al., 1993) with flat and low (less than ten times chondrite) HREE abundances and strong upward flexure with increasing incompatibility in the LREE. Middle rare earth elements (MREE) may or may not have begun to exhibit increasing abundances. Often the high field strength elements (HFSE), such as Zr and Ti, are at or so close to their post-melting abundances that they create strong negative anomalies in the chondrite-normalized patterns. The clinopyroxene analyses plotted on Figure 10 (B, D) are typical of this type of pattern.

This type of pattern can be modeled as developing from a lowporosity melt percolation process that chromatographically fractionates elements from one another (Navon and Stolper, 1987; Bodinier et al., 1990). As with cation exchange resin in a laboratory column, the basic process is one of speed of transport dependent on an element's proportion of time spent bonded to the mantle residue vs. time spent moving with the melt. The more incompatible elements move at a speed closer to that of the melt front because they spend less time in mantle minerals. Thus, the LREE will arrive at a certain point (later sampled as a xenolith) above the percolation starting point before the MREE and HREE. Diffusion of these early-arriving elements into the residual mantle will cryptically metasomatize the material. For this fragment of the mantle to be sampled, a much larger melt fraction en route to the surface must mechanically fracture into the area and entrain samples of a material that has interacted only with the leading edges of a percolating melt.

The model presented in Figure 10C was calculated using the Fortran percolation program of Bodinier et al. (1990), as modified for Macintosh by E. Takazawa (unpubl. data, 1992). The partition coefficients and other constants used for this model are presented in Table 6. The alteration of the xenolith in thin section Sample 144-871C-35R-1, 23–25 cm (Xenolith A), precluded modal analysis and grain-size estimation. These were taken from a Tahitian xenolith that I think

Table 7. Comparative orthopyroxene analyses.

| | SAV 1-5A | |
|--------------------------------|----------|--------|
| | OPX/1 | 33-H |
| SiO2 | 52.79 | 54.04 |
| TiO | 0.00 | 0.28 |
| Al2Õ3 | 1.35 | 4.15 |
| Cr ₂ O ₃ | 0.67 | 0.00 |
| FeO | 6.52 | 11.07 |
| MnO | 0.19 | ND |
| MgO | 38.40 | 29.50 |
| NiO | 0.00 | ND |
| CaO | 0.90 | 0.97 |
| Na ₂ O | 0.00 | 0.00 |
| Total | 100.82 | 100.01 |
| En | 89.94 | 80.99 |
| Fs | 8.53 | 17.08 |
| Wo | 1.53 | 1.93 |

Note: SAV 1-5A OPX/1 = Samoan mantle residue (data from J.J. Dieu, unpubl. data, 1993). 33-H = Pyroxenite cumulate (data from Sen, 1988). ND = no data.

represents equivalently melted mantle. The initial composition of the percolating melt is plotted on Figure 10H. The early percolation stages of the model are a fairly good reproduction of the data from Sample 144-871C-35R-1, 23–25 cm (Xenolith A). The clinopyroxene pattern that would represent final equilibration with the percolating liquid is plotted on Figure 11C.

The second type of Type 1 clinopyroxene pattern (Fig. 10E, F) is typically quite smooth and slightly convex-upward; it also displays very high abundances of REE [e.g., 40-100 La(n)] with strong negative HFSE anomalies. Different xenoliths from a locality on Savai'i, Western Samoa, contain clinopyroxene with both types of patterns (J.J. Dieu and J.H. Natland, unpubl. data, 1993). Xenoliths with clinopyroxene that have this type of pattern were also cryptically metasomatized (i.e., no hydrous phases were added and there are no observable changes in major element chemistry). These two lines of evidence-the occurrence of both patterns from a single xenolith locality, sampling otherwise identical material, and both patterns created by a process best described as cryptic metasomatism-suggest that this second type of pattern is also created by percolation. In this latter case, percolation occurred for a sufficient time period for the less incompatible elements to arrive and equilibrate before the material was sampled as a xenolith. Basically, these clinopyroxene grains were in equilibrium with a very enriched melt; percolation is invoked as a reasonable, but not required, explanation.

Melts in equilibrium with the clinopyroxene analyses from MIT Guyot Type 1 xenoliths (Fig. 10E) were calculated using the partition coefficients in Table 6 and are plotted, chondrite-normalized, on Figure 10G. These calculated melts and the estimated melt composition used in the modeling of the Limalok Guyot Type 1 clinopyroxene data show unusually high REE abundances and slight Zr and large Ti negative anomalies. These characteristics preclude the metasomatizing melts being basaltic liquids or their differentiates. Actual erupted liquids that most closely resemble these percolating melts are carbonatites (cf. Fig. 10G, H). This was also the conclusion of Hauri et al. (1993) for Samoan and Cook-Austral xenoliths and Hassler et al. (1993) for Kerguelen xenoliths.

Type 2 Xenoliths

Clinopyroxene grains from two Type 2 xenoliths, one from Limalok Guyot and one from MIT Guyot, were analyzed by ion microprobe for rare earth and other trace element concentrations. These data are presented in Table 7. Hereafter in this section, these xenoliths are referred to as the "Limalok xenolith" and the "MIT xenolith."

Figure 11A is a chondrite-normalized extended REE diagram. Two analyses from each of the Limalok and MIT xenoliths are plotted. There are some small anomalies, both positive and negative, in the trace element abundances relative to the REE abundances.



Figure 10. Chondrite-normalizing values from Evensen et al. (1978). A. Ion microprobe data of clinopyroxene in a Tahitian Type 1 xenolith. B. Ion microprobe data of Limalok Guyot Type 1 clinopyroxene. C. Modeled clinopyroxene patterns. Steps at 500, 1500, 2500, and 3500 yr. Constants in Table 6. D. Ion microprobe data of Samoan Type 1 clinopyroxene (from J.J. Dieu and J.H. Natland, unpubl. data, 1993). E. Ion microprobe data of MIT Guyot Type 1 clinopyroxene (from Hauri et al., 1993). G. Melts calculated to be in equilibrium with MIT Guyot Type 1 clinopyroxene and the melt used for the percolation model in Figure 11B. H. Shaded field encloses data for erupted carbonatites reported by Nelson et al. (1988).

Small, random-direction anomalies most likely represent heterogeneities within the clinopyroxene grains. Rare earth and trace elements are analyzed at separate times on distinct spots about 10 µm apart, and heterogeneity in many of the crystals can be measured "downhole" on a submicron scale during an analysis.

The clinopyroxene grains in the Limalok xenolith contain greater abundances of the LREE and of Sr and Zr than do the clinopyroxene grains in the MIT xenolith, and, overall, REE patterns of the Limalok xenolith clinopyroxene grains are steeper (i.e., have greater La/Yb) than those of the MIT xenolith clinopyroxene grains. These differences reflect a fundamental difference in the types of liquids that differentiated to form the cumulate material sampled by the xenoliths. The Limalok cumulate-forming liquids were much more enriched in LREE, Sr, Zr and, by inference, other highly incompatible elements. This agrees with the evidence of the major element chemistry of the clinopyroxene, which suggests crystallization from alkalic liquids.

Figure 11 (B, C, D) present sets of clinopyroxene data from Type 2 xenoliths from Ua Huka Island in the Marquesas Chain, Tahiti Island in the Society Chain, and the Island of Hawaii, respectively. Each of these suites of xenoliths has been interpreted as samples of alkalic cumulates (Chen et al., 1992; J.J. Dieu and J.H. Natland, unpubl. data, 1993). It is clear from the total range of alkalic cumulate



Figure 10 (continued).

clinopyroxene patterns presented in Figure 11 (B, C, D) that identification of the MIT xenolith as a tholeiitic cumulate on the basis of this evidence is not possible. Presumably, clinopyroxene grains in tholeiitic cumulates also show a very wide range of rare earth and trace elements, but the data are not yet available for comparison with the MIT xenolith data.

The clinopyroxene grain analyzed from Sample 144-871C-35R-1, 10-12 cm (CPX A), contains much greater abundances of LREE than I have previously observed in cumulate clinopyroxene grains from the Samoan, Society, or Marquesan chains. Because the identification of this xenolith as Type 2 is firmly established on arguments of petrography (no orthopyroxene) and major element chemistry, this analysis provides a new upper limit to the range of LREE abundances observed in clinopyroxene from alkalic cumulates. The range, as presented in Figure 11, shows a variance in La abundances across 2.5 orders of magnitude. Mafic alkalic series lavas that erupt at hotspots do not exhibit a range anywhere near this great; they vary, perhaps at most, by a factor of 3 in La concentration. Obviously, a process other than simple equilibrium phenocryst growth is controlling the incompatible trace element abundances in cumulate clinopyroxenes. Langmuir (1989) hypothesized and modeled a process he called "in situ fractional crystallization," which described the effects of fractionation on an interstitial liquid as post-cumulus growth occurs during cumulate solidification. Stated simply, the results of this type of fractionation are liquids with only slightly modified major element chemistry but with very elevated abundances of incompatible elements. These pockets of interstitial liquid may crystallize clinopyroxene with Mg#s similar to the original cumulus cores, but with high and variable abundances of LREE and other incompatible elements. Cumulates formed in shallow, rapidly cooling magma chambers appear to preserve these heterogeneities on a grain- or xenolith-length scale because subsolidus diffusion of incompatible elements in clinopyroxene is much too slow to be effective at erasing them (J.J. Dieu and J.H. Natland, unpubl. data, 1993).

CONCLUSIONS

Lavas recovered from Wodejebato Guyot, Sites 874 and 877, and igneous Unit 15 of Site 878 on MIT Guyot are relatively primitive compositions. The liquids had Mg#s that ranged from 0.60 to 0.65, based on calculations of olivine chemistry and comparisons of clinopyroxene and plagioclase chemistry.

Lavas from Limalok and MIT guyots identified as basanitic contain clinopyroxene phenocrysts and groundmass grains with rather remarkable compositions characterized by high TiO₂ (usually > 2%; Fig. 1), high Al₂O₃ (up to 12%; Table 3), and high tetrahedral aluminum (Al(z) > 10). These characteristics were not observed in clinopyroxene from the alkalic series lavas. Thus, identifications by wholerock chemistry and petrography of Limalok and some MIT lavas as basanitic are strongly supported by mineral chemistry.

High La/Yb ratios of phenocrysts in the Limalok Guyot basanitic lavas indicate that these were created from a more enriched source than were the MIT Guyot basanitic lavas. Whole-rock trace element data and isotopic data support this conclusion.

The La/Yb ratios and chondrite-normalized abundances of rare earth and trace elements in phenocrysts from alkalic series lavas from Lo-En, MIT, and Takuyo-Daisan guyots are similar, despite apparent differences in source chemistry (Koppers et al., this volume), in the (probable) degrees of partial melting from which each lava suite was derived, and in the degrees of differentiation. This coincidence illustrates the difficulties inherent in working with phenocryst chemistry.

The La/Yb ratios and chondrite-normalized abundances of rare earth and trace elements in phenocrysts indicate that the ankaramite recovered from Hole 874B consists of at least two flows apparently derived from sources of distinctly different enrichment. The less enriched source was less enriched than the source materials of the transitional and alkalic series lavas from Lo-En, MIT, and Takuyo-Daisan guyots. The phenocrysts of the ankaramitic lava with more enriched signatures have higher abundances of trace elements and higher La/Yb than do phenocrysts in the Lo-En, MIT, and Takuyo-Daisan transitional or alkalic series lavas for any of several reasons.

Incompatible element abundances in phenocryst rims and microphenocryst cores and rims show that fractionation of these liquids was extensive (50%–65%) just before and during eruption.

Major and trace element chemistry of clinopyroxene from Type 2 xenoliths from Limalok Guyot indicate that these are most likely cumulate materials that fractionated from alkalic liquids. Presumably, given the high abundances of rare earth and trace elements, these alkalic liquids were related to earlier volcanism on Limalok Guyot (before the eruption of the basanitic-series that entrained the xenolith) and were not mid-ocean-ridge related.

Major element chemistry of clinopyroxene and the presence of orthopyroxene, coupled with an apparent lack of plagioclase, indicate that the MIT Guyot Type 2 xenoliths are cumulates of tholeiitic liquids with hotspot characteristics rather than abyssal characteristics. This is the single indication from all drilling results of Leg 144 that these volcanic edifices may be built largely of tholeiitic magma and that transitional, alkalic, and basanitic series lavas form only a carapace.

The high abundances of incompatible elements in the clinopyroxene grains in Type 1 xenoliths from MIT and Limalok guyots suggest that cryptic metasomatism occurred. The extended REE patterns from Limalok Guyot Type 1 clinopyroxene grains can be modeled as developing from diffusion of the early-arriving elements in a percolating carbonatite-like liquid. Sampling of these xenoliths occurred before equilibration with the later-arriving elements could occur. The extended REE patterns from MIT Guyot Type 1 clinopyroxene grains



Figure 11. Analyses of clinopyroxene grains from cumulate xenoliths. **A.** Limalok and MIT Guyot Type 2 xenoliths. **B.** Ua Huka Island alkalic cumulate xenoliths (from J.J. Dieu and J.H. Natland, unpubl. data, 1993). **C.** Tahitian alkalic cumulate xenoliths (from J.J. Dieu and J.H. Natland, unpubl. data, 1993). **D.** Hualalai (Hawaii) alkalic cumulate xenoliths (from Chen et al., 1992). All patterns plotted in Figure 11 (A–C) are ion microprobe analyses. Patterns plotted in Figure 11D are INAA analyses of mineral separates.

appear to be fully equilibrated with a carbonatite-like melt. Xenoliths with these types of extended REE patterns have been identified from the Marquesas and Society chains (J.J. Dieu and J.H. Natland, unpubl. data, 1993), from the Samoan and Cook-Austral chains (Hauri et al., 1993), and from the Kerguelen Islands (Hassler et al., 1993). This plume process of carbonatitic liquids percolating into previously depleted peridotite appears to be a process that is unique to the enriched Southern Hemisphere mantle sources. It is a process with a long history in the SOPITA region—at least 128 Ma of history.

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