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Differentiation behavior of Kilauea Iki lava lake, Kilauea Volcano, Hawaii: An overview of past and current work

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Abstract—The 1959 eruption of Kilauea Volcano, which produced the only picritic lava at Kilauea's summit in historic time, formed Kilauea Iki lava lake, a large pond of lava over 100 m deep. Both this unusual eruption and the lava lake have been the subject of many studies.

Petrologic investigations of subsolidus and partially molten drill core samples from Kilauea Iki carried out to date include studies of zoning or re-equilibration patterns, or both, in olivines and chromites, and determination of the quenching temperature of the core using a variety of geothermometers. In addition, detailed study of interstitial glass compositions shows that (1) the basaltic liquids in the lake are capable of fractionating to rhyolitic liquid compositions and (2) the succession of liquid compositions found is significantly affected by the amount of olivine present. At low olivine contents, glasses are relatively Fe-rich and low in SiO_2, Al_2O_3, alkalies and P_2O_5, whereas glasses in olivine-rich rocks show the opposite effects. The reason for the effect is that as olivine re-equilibrates with the liquid, it takes up FeO, so the liquids at a given MgO content become progressively lower in FeO and higher in SiO_2, Al_2O_3 the alkalies and P_2O_5.

Processes occurring within Kilauea Iki lava lake include (1) redistribution of olivine phenocrysts by gravitative settling; (2) migraton and loss of vapor bubbles (vesicles); (3) redistribution of olivine + augite during convection; (4) migration of low-density liquid from the bottom of the lake to the top via diapiric melt transfer; (5) formation of pipe-like vuggy olivine-rich bodies by a second, lower-temperature diapiric process; (6) formation of segregation veins, sill-like internal differentiates of ferrobasaltic composition; and (7) formation of minor veinlets of andesitic to rhyolitic composition by flowage of melt into cracks that form as the partially molten rock cools and fractures. Quantification of the physical parameters controlling these magmatic processes may eventually make it possible to apply results from Kilauea Iki to other, larger mafic magma chambers.

INTRODUCTION

SEVERAL TIMES in recent decades, lava erupted at Kilauea Volcano has ponded in one or another of the pit craters that occur along the upper east rift zone of Kilauea volcano (Figure 1). The resulting lava lakes are natural laboratories for the study of the processes by which basaltic lava cools, crystallizes and differentiates. Historic lava lakes investigated periodically include the 1963 Alae lava lake (PECK *et al.*, 1966; WRIGHT and PECK, 1978) and the 1965 Makaopuhi lava lake (WRIGHT *et al.*, 1968; WRIGHT and OKAMURA, 1977). In addition, the prehistoric Makaopuhi lava lake was investigated by J. G. Moore and B. W. Evans (MOORE and EVANS, 1967; EVANS and MOORE, 1968). An earlier summary of lava lake studies was given by WRIGHT *et al.* (1976).

Kilauea Iki lava lake, formed by the 1959 eruption of Kilauea is the biggest of the historic lava lakes and by far the most thoroughly studied. Unlike the 1963 Alae and 1965 Makaopuhi lakes, which have been buried by the 1969–1974 lavas from Mauna Ulu (SWANSON *et al.*, 1972, 1979; TILLING *et al.*, 1987), Kilauea Iki has not been the scene of any more recent eruptive activity, and so has remained accessible from 1959 to the present (1986). Its size and accessibility, and the unusual character of the eruption which fed it have made it an object of interest to petrologists and geophysicists for over two decades. The purpose of this paper is to review previous work and to present some of the work in progress on both the 1959 eruption and lava lake.

THE 1959 ERUPTION

The spectacular 1959 summit eruption of Kilauea Volcano was a benchmark in modern volcanology because it was the first major Kilauean eruption for which most of the array of modern volcano-monitoring techniques were used. Descriptions of the relationship between earthquake activity, summit deformation, and the 1959 and 1960 eruptions of Kilauea, by EATON (1962) and EATON and MURATA (1960), gave an unprecedentedly detailed picture of Kilauea's magmatic plumbing. More recently, EATON *et al.* (1987) have interpreted the tilt records in greater detail, showing the pattern of exchange of magma between Kilauea Iki lava lake and the subsurface reservoir during the 1959 eruption, and documenting the arrival of new magma into the system through the first half of the eruption.

A day-by-day account of the 1959 eruption is given by RICHTER *et al.* (1970). It lasted from November 14 to December 20, formed a prominent cinder cone and pumice blanket as well as Kilauea Iki lava lake, and included seventeen phases of high fountaining and lava output, separated by periods of relative quiescence. Temperatures as high as 1215°C, the highest in Kilauea's history, were re-

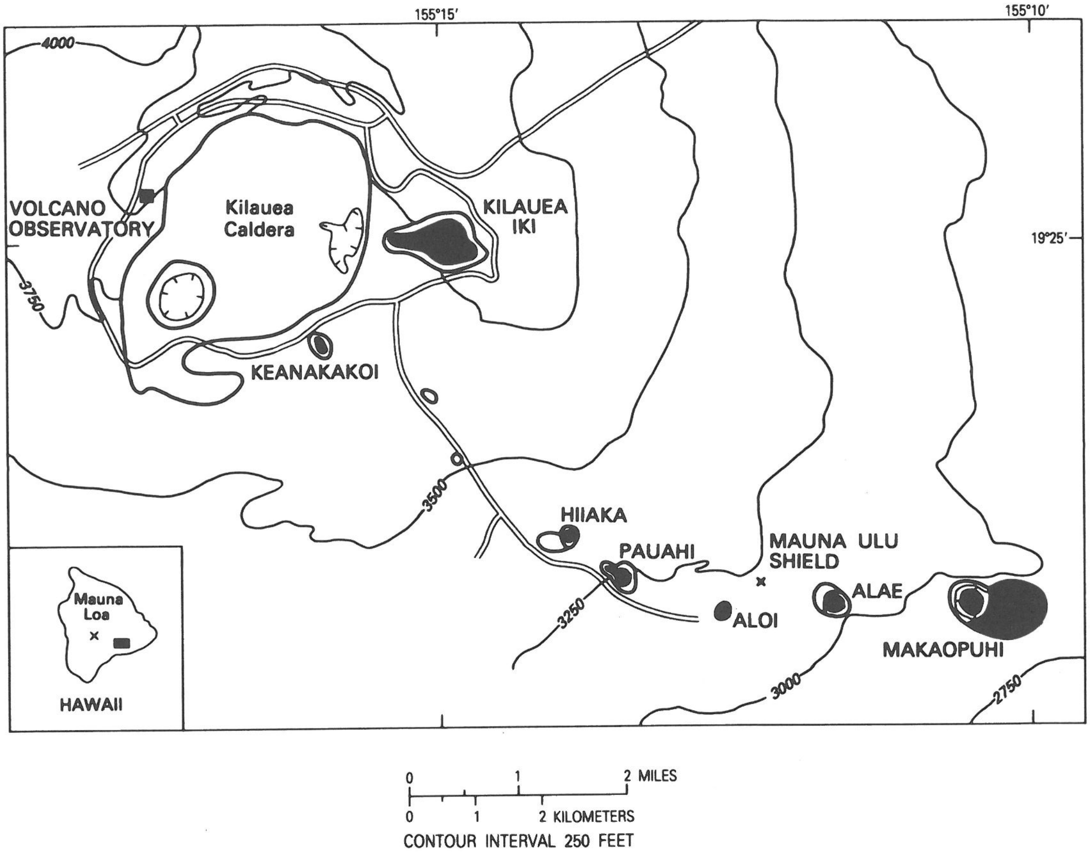

FIG. 1. Index map of the summit area of Kilauea Volcano. Pit craters containing ponded flows (= lava lakes) are labelled by name. Of these, only Kilauea Iki (formed in 1959), Alae (1963) and Makaopuhi (1965) have been studied in any detail. The prehistoric Makaopuhi lava lake, shown in the stippled pattern, has also been studied. The craters of Aloi, Alae and the west pit of Makaopuhi are now completely filled by lavas of the 1969–74 Mauna Ulu shield.

corded, and fountain heights of up to 1900 feet (580 m), the highest of any historic summit eruption, were observed. Repeated crustal foundering and growth of new crust continued for three days after the eruption ended; it was December 23, 1959 before the crust stabilized and studies of the lava lake could begin (RICHTER *et al.*, 1970; AULT *et al.*, 1961).

Early work on the 1959 lavas included studies of their chemistry (MURATA and RICHTER, 1966a; MACDONALD and KATSURA, 1961) and of their petrology (RICHTER and MURATA, 1966). Trace-element data for a suite of 1959 samples, presented by GUNN (1971), showed clearly the effects of the varying olivine content of the eruption pumices. In addition, the gases emitted during and shortly after the eruption were analyzed (HEALD *et al.*, 1963; MURATA, 1966). Processes documented as occurring during the eruption include olivine settling (MURATA and RICHTER, 1966b) and magma mixing (WRIGHT, 1973). WADGE (1981) analyzed the variation in magma-discharge rate as the 1959 eruption proceeded and concluded that it was compatible with progressive degassing of a closed system, leading to higher effusion rates as the magma became progressively cooler and less gas-rich, and hence more viscous and non-Newtonian.

The 1959 lava was picritic in character, the only picritic lava erupted at or near Kilauea's summit in historic times. WRIGHT (1973) estimated the average MgO content of the eruption pumices to be 15.43 percent by weight. As would be expected from their chemistry, the eruption samples contain abundant (17–20 weight percent, on the average) olivine phenocrysts, plus minor chromite, in a matrix of vesicular brown glass. The chromite, which occurs almost exclusively as inclusions in olivine, was analyzed by EVANS and WRIGHT (1972). Compositional data on olivine and glass compositions were reported by LEEMAN and SCHEIDEGGER (1977) and by HARRIS and ANDERSON (1983). The latter study also presents data on the H_2O, CO_2, and S content of glasses included in olivine phenocrysts.

More recently, the olivine phenocryst population has been re-examined and classified on the basis of crystal morphology, characteristic inclusions, etc. by SCHWINDINGER *et al.* (1983), SCHWINDINGER (1986), and by HELZ (1983, 1987a), who also presents analytical data for olivines and glasses in the pumice samples and for olivines in the lake. The results on the pumice samples corroborate the inferences of EATON *et al.* (1987) that new material was entering the erupting system through phase 9 of the eruption. In addition, HELZ (1987a) documents the occurrence in the lava lake of dunitic inclusions having deformed or annealed textures, or both, like those of mantle-derived xenoliths (see *e.g.*, KIRBY and GREEN, 1980), and of rare olivine megacrysts (up to 2 cm long). The presence of these materials, plus the unusually high eruption temperatures and fountain heights observed during the 1959 eruption, have been explained by a model in which one of the two mixing components of the eruption (the S-1 component of WRIGHT, 1973) is hypothesized to have come directly from the mantle, without being stored for a significant time in Kilauea's shallow summit reservoir. In this model, the juvenile component originated at 45–60 km depth, where it was liberated from the mantle during the August 14–19 earthquake swarm (EATON, 1962). It reached the surface November 14. It thus had an average ascent velocity of 0.6–0.8 cm/sec, just sufficient to offset the settling velocity of, and hence to entrain, the largest (1 × 2 cm) olivine grains and dunitic aggregates found in the 1959 lavas.

PREVIOUS LAVA LAKE STUDIES

In early 1960, shortly after the crust stabilized, a network of leveling nails (shown in Figure 2a) was established and drilling of Kilauea Iki lava lake began in April 1960. The lake was drilled in 1960, 1961, and 1962, by Hawaii Volcano Observatory (HVO) staff and also by workers from Lawrence Radiation Laboratories. The resulting drill core was described and analyzed by RICHTER and MOORE (1966). Downhole temperature measurements showed temperatures of 1060–1106°C at the crust-melt interface, as defined by drilling (RAWSON, 1960; AULT *et al.*, 1961, 1962; RICHTER and MOORE, 1966). Gas from one of the early drill holes, was collected and analyzed by HEALD *et al.* (1963), and determined to be dominantly $SO_2 + H_2O$.

The lake was drilled by the U.S. Geological Survey (U.S.G.S.) in 1967 and 1975, whereas drillings in 1976, 1979, and 1981 were cooperative efforts between the U.S. Geological Survey (U.S.G.S.) and Sandia National Laboratories. Figures 2a and 2b show the location in plan view and in cross-section of the 1967–1981 drill holes. Observations on this drilling experience, including discussions of (1) the significance of the crust-melt interface, (2) backfilling of the holes by melt and crystal-liquid mushes, and (3) degassing of the lake with time are presented in WRIGHT *et al.* (1976) and HELZ and WRIGHT (1983). Petrographic logs of all of the core are given in HELZ *et al.* (1984) and HELZ and WRIGHT (1983).

The open drill holes have been used for a variety of downhole geophysical studies and measurements. ZABLOCKI and TILLING (1976) determined the Curie point of the cooling basalt in situ. Sandia personnel conducted a number of heat-extraction and other experiments in the 1981 drill holes (HARDEE *et al.*, 1981). In addition, since 1967, temperature profiles have been measured in most of the drill holes. Most of these profiles are still unpublished, but examples can be found in ZABLOCKI and TILLING (1976; profiles from holes KI75-1 and KI75-2), in HERMANCE and COLP (1982; profile from hole KI76-1), and in HARDEE *et al.* (1981, profile from KI81-1). The latter profile is anomalous in shape, probably because another hole (KI81-5) was being drilled only 5 m away at the time the profile was taken. Generalized thermal modeling of the lake has been done in one dimension (PECK *et al.*, 1977) and two dimensions (RYAN, 1979). The only attempt to date to work with and interpret the shapes of actual temperature profiles is that of HARDEE (1980).

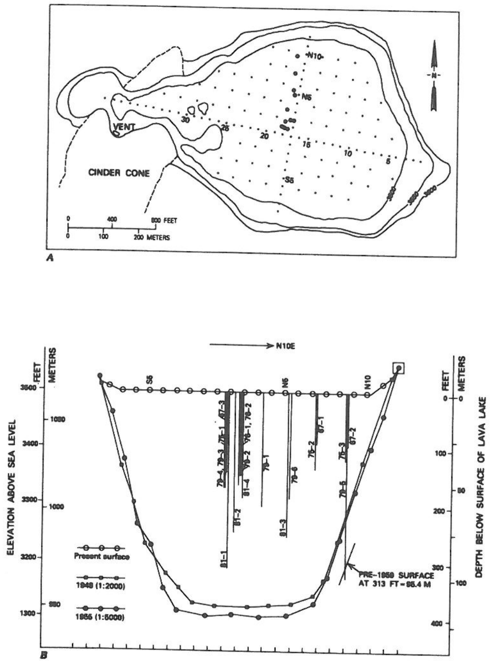

FIG. 2. (A) Plan view of the post-1959 surface of Kilauea Iki lava lake. The large dots indicate the location of drilling sites occupied from 1967 to 1981. The network of levelling stations is shown by the small dots. Individual levelling stations along the principal N-S line of stations have been labelled in Figures 2a and 2b to facilitate comparison. (B) Cross-section of Kilauea Iki lava lake, taken along the N-S line of closely-spaced levelling stations shown in Figure 2a. The present surface of the lava lake and two pre-eruption profiles are shown. The actual location of the lake bottom below hole KI79-5 is significantly different from either, as indicated; possible causes are discussed in HELZ (1980). Vertical exaggeration is 4:1. The drill holes, most of which lie along a line 100 feet west of the section given in Figure 2b, are shown projected onto this cross-section. Several of the drill hole locations have been drilled more than once, in order to sample the same section of crust as cooling proceeded. These clusters of closely spaced drill holes are shown schematically in this figure; the spacing between holes within clusters is not to scale.

During the 1970's the lake was the object of many geophysical investigations. CHOUET (1979) and RYAN (1979) studied the seismic activity associated with crack formation in the lake. ZABLOCKI (1976), FLANIGAN and ZABLOCKI (1977) and SMITH *et al.* (1977) used a variety of electromagnetic sounding methods to locate the boundaries of the lens of melt as it existed in the mid-seventies. A summary of the geoelectric structure of the lava lake is given by ANDERSON (1987). CHOUET and AKI (1981) used both active and passive seismic techniques to characterize the lake. HERMANCE and COLP (1982) attempted a synthesis of these various geophysical studies of the lava lake. The lateral extent of the lens of melt, as defined electrically and seismically, agreed well with that inferred from the drilling results and the leveling data (HVO unpublished data). The thickness of the lens of melt was less well defined, with estimates ranging from <10 m (based on seismic data) to >30 m (from the electromagnetic sounding experiments). Both the drilling results (HELZ *et al.,* 1984) and early petrographic reports (HELZ, 1980) suggested the greater thickness was correct, and this was subsequently confirmed by the results of the 1981 drilling (HELZ and WRIGHT, 1983). The lens of melt, where "melt" is defined as a crystal-liquid mush too fluid to drill into, was 30–40 m thick in 1976. The crystallinity or vesicle content, or both, of the lens of melt at that time were nevertheless sufficiently high for the molten lens to transmit shear waves, as reported by CHOUET and AKI (1981). This result suggests that the delineation of partially molten zones in the crust or upper mantle by seismic methods will be a very difficult problem.

Most other studies of the lake have focused on the drill core, using the observed bulk compositions, mineral assemblages and phase compositions to delineate the differentiation history of the lava lake, a task which is still in progress. HELZ (1980) presented a summary of the gross petrographic variations of the 1967–1979 core, including a detailed discussion of the pattern of occurrence of segregation veins in the lake. LUTH and GERLACH (1980) and LUTH *et al.* (1981) characterized the phase compositions and assemblages of selected 1979 and 1981 core samples. In addition, HELZ and THORNBER (1981) determined the 1-atmosphere melting relations of two core samples. Analysis of the resulting melts showed that CaO and MgO in the melt varied linearly with temperature, suggesting that the concentration of these oxides in glass found in quenched partially molten drill core could be used as empirical geothermometers, to assign quenching temperatures to all glassy core from the lake. Lastly, JOHNSON (1980) determined the porosity and permeability of a suite of samples from hole KI76-1.

CURRENT WORK

Repeated drilling of Kilauea Iki lava lake (in 1960, 1961, 1962, 1975, 1976, 1979 and 1981) has provided a suite of drill cores that record in detail the crystallization and differentiation of this small, self-roofed magma chamber. Throughout most of the period 1960–1981, drilling was stopped at the base of the upper crust, that is, wherever the drillers encountered material too fluid to drill into. Most holes, therefore, sample only the upper crust, the exceptions being hole KI79-5, which passed through the entire lake near its north edge (see Figure 2B) and several of the 1981 holes (KI81-1, 2 and 3) which passed the thermal maximum in the lake (see Figure 2B; also HELZ and THORNBER, 1987). Core recovery rates in 1960–62 were quite low, and varied from 70% to 98% for the 1967–1975 holes (HELZ *et al.,* 1984). In all subsequent drillings (1976–1981), use of diamond drill bits resulted in 100% recovery of core in all holes (HELZ and WRIGHT, 1983; HELZ *et al.,* 1984).

The drill core is very fresh, except for local oxidation at temperatures of 100–950°C. Much of the core was partially molten prior to being quenched by the water used to cool the bit. Only core quenched from just above the incoming of the first oxide phase (ilmenite or ferro-pseudobrookite) shows any quench-phase crystallization, in the form of black fuzzy rims on plagioclase or augite.

The available core has been extensively sampled (>400 samples) for petrography and chemical work, and the bulk compositions of 143 samples have been determined gravimetrically, by H. R. Kirschenbaum and J. W. Marinenko, of the U.S.G.S. The trace element characteristics of the lava lake are being investigated by M. M. Lindstrom, who has analyzed a suite of 55 of these same, analyzed samples, by using neutron activation techniques. The whole-rock analyses will not be summarized here, but some of the following discussion draws on this extensive body of largely unpublished data. In all of the work summarized below, including that done by workers other than the present author, this same group of samples has been used when possible, to facilitate comparison among the various studies.

Crystalline assemblages and phase compositions

Examination of the drill core has led to recognition of three texturally had chemically distinct rock types in the lake (not counting the complicated textures, produced by annealing and small-scale filterpressing, seen in foundered crust). Most of the lake is olivine-phyric basalt, with normal igneous textures, referred to in this paper and others (*e.g.,* HELZ, 1980) as "matrix rock." (Photomicrographs of typical matrix rock are shown in HELZ, 1980, 1987a,b). The principal macroscopic variation in the matrix rock is variation in the amount of phenocrystic olivine, which ranges from 1–3 volume percent (MANGAN and HELZ, 1985) to >40 volume percent.

In addition, the upper crust of the lake, especially the interval between 18 and 56 m below the surface, contains 6–10% by volume segregation veins. These veins are coarse-grained, diabase-textured micropegmatites (see photomicrographs in RICHTER and MOORE, 1966; HELZ, 1987b), and are irregular to sill-like in form. They occur in all the Hawaiian lava lakes (WRIGHT *et al.,* 1976), but are particularly numerous and well-developed in Kilauea Iki (HELZ, 1980). Chemically, the segregation veins are olivine-poor ferrobasalt. Their range of composition in Kilauea Iki does not overlap that of the olivine-phyric matrix rock from which they are derived (see HELZ, 1987b; also Figure 3).

The third rock type consists of the vertical olivine–rich bodies (vorbs) described and illustrated in HELZ (1980, 1987b). These irregular pipe-like bodies, found principally between 40 and 58 m below the lake surface, extend vertically up the side of the core, and are locally traceable for a meter or more. They are enriched in iron-rich olivine and contain an excess of differentiated, segregation-vein-like liquid relative to normal olivine–bearing matrix rock of similar MgO content (HELZ, 1987b). Much of the textural contrast between these cross-cutting bodies and the adjacent matrix results from the fact that the groundmass phases (olivine + augite + plagioclase) in the vorb mostly occur adjacent to olivine, so that the main pools of melt are free of crystals. This textural rearrangement has been interpreted as resulting from differential vertical flow of crystals, liquid and vesicles with the vorb (HELZ, 1980).

These textural variants are not mineralogically distinct, the bulk compositional differences between them being accommodated by variations in the amount of the different phases present. In the matrix rock and vorbs, olivine is the only phenocryst phase. The groundmass of the olivine-phyric basalt consists of olivine, tiny (0.1 mm or less) augite grains and lathy plagioclase, with or without Fe–Ti oxides, pigeonite or hypersthene, apatite, sulfide and glass depending on the quenching temperature of the core. Cristobalite rosettes occur in vesicles or diktytaxitic cavities, especially in olivine–poor holocrystalline samples; they have not been observed to coexist with glass in any olivine–bearing sample. No discrete alkali feldspar phase has been observed.

The segregation veins are aphyric. They contain only minor, late–crystallizing olivine and always contain pigeonite rather than hypersthene, but are not otherwise mineralogically distinct from the matrix rock. Rarely, cristobalite has been observed crystallizing from glass in segregation veins quenched from just above the solidus.

The amount and petrographic and chemical character of the olivine phenocrysts vary, as described in HELZ (1980, 1987a). The olivine phenocrysts (typically 0.5–8 mm long) characteristically contain inclusions of chromite. The composition of the chromite and its compositional changes with time from 1959 until quenched during drilling have been documented by SCOWEN (1986; see also SCOWEN *et al.*, 1986). She found that chromite continued to re–equilibrate with the host olivine and with the liquid outside the olivine, probably by diffusion of ions through the olivine. With the passage of time since the eruption, Mg, Al and Cr in chromite decreased, while Ti, Fe^{2+} and Fe^{3+} increased, the effect being greater for core from progressively deeper in the lake. The results have implications for the study of olivine-chromite assemblages in other mafic and ultramafic bodies.

The pyroxenes in Kilauea Iki are similar in composition to pyroxenes from other Hawaiian lavas (see analyses in MURATA and RICHTER, 1966a; also in WRIGHT and PECK, 1978). They are of interest because of the opportunity to determine rather precisely (1) the bulk compositional controls on the occurrence of pigeonite *vs* hypersthene, both of which occur in the lake; (2) the composition of all coexisting phases, including that of the melt at the point where the low–Ca pyroxene appears (see *e.g.*, the glass in Table 1, Column 5); and (3) the stoichiometry of the two reactons at that point.

Another noteworthy feature of the pyroxenes is the tendency of hypersthene to occur as oikocrysts. The monoclinic pyroxenes rarely exhibit this habit: only one augite oikocryst has been found in the lake to date. In contrast, most hypersthene in the lake is oikocrystic, occurring in plates up to a few millimeters in length. The hypersthene encloses plagioclase laths, the grain size and random orientation of which are identical to those of plagioclase laths outside the oikocrysts. The augite and groundmass olivine, which occur uniformly throughout the matrix, are almost completely absent within the oikocrysts. The resorption of olivine by hypersthene is not surprising; the reaction relationship olivine + liquid $\rightarrow$ hypersthene is familiar to all petrologists. There is not a recognized reaction relationship between augite and hypersthene, however, and yet the augite (which first crystallized at $\geq 1170°C$) has been resorbed by the oikocrysts, which begin to grow at $\sim 1090°C$. Why and how this resorption takes place is not yet clear. These observations suggest, however, that the inclusion population in pyroxene oikocrysts in layered intrusives is not a reliable indicator of the presence or absence of cumulus minerals other than plagioclase.

Plagioclase in the lake varies in composition from $An_{78.2}Ab_{21.2}Or_{0.6}$ to $An_{8.7}Ab_{69.8}Or_{21.5}$, though this range is not found within a single sample. The range of plagioclase compositions found within each sample increases from 5 mol percent An at 1140°C to 30–35 mol percent An in near-solidus samples. Maximum zoning observed within single grains is typically 10 mol percent An in plagioclase in olivine-phyric rock; in segregation veins and other, more differentiated samples, An content may vary by as much as 25% within a single crystal. The plagioclase zonation patterns in the lava lake are of interest because of their bearing on the development and preservation of zoning in plagioclase in other mafic rocks.

The phase assemblages exhibited by the Fe–Ti oxides are somewhat unusual because of the existence of three oxide phases (ilmenite, ferropseu-

Table 1. Representative electron microprobe analyses of interstitial glasses from Kilauea Iki lava lake.

Sample no.	1 KI81-1-224.4	2 KI75-1-145.1	3 KI75-1-143.8	4 KI75-1-141.8	5 KI81-1-299.9	6 KI67-3-83.8	7 KI79-1-167.6	8 KI75-1-130	9 KI79-5-193.3
No. points analyzed	14	7	19	13	6	7		18	10
SiO_2	51.4	50.7	51.8	51.7	54.2	55.0	60.6	71.4	68.9
TiO_2	3.89	4.14	5.08	5.40	3.98	3.11	2.17	1.00	1.52
Al_2O_3	13.5	13.0	12.6	12.1	13.2	12.6	14.0	13.8	14.6
Cr_2O_3	.02	.01	.03	.00	.00	.00	.00	.00	.00
FeO^a	10.3	11.8	12.1	12.7	10.6	13.2	7.51	2.71	2.45
MnO	.12	.14	.14	.18	.14	.16	.06	.02	.01
MgO	6.26	5.88	4.99	4.52	4.33	3.18	2.21	.42	.59
CaO	10.1	9.84	8.82	8.48	7.67	6.98	4.53	.94	1.01
Na_2O	2.87	2.57	2.67	3.12	3.36	2.71	3.99	3.65	3.77
K_2O	.78	.88	1.03	1.06	1.66	1.62	3.05	5.59	5.85
P_2O_5	.31	.40	.43	.39	.62	.71	1.45	.12	.22
Sum	99.6	99.3	99.7	99.6	99.8	99.2	99.6	99.7	98.9
Comments:	Glass at thermal maximum	Glass below crust/melt interface	Glass in typical matrix	Glass just above oxide-in	Glass just below opx-in	Glass below pigeonite-in, at magnetite-in	Glass in typical matrix	Last glass, olivine basalt (MgO = 10%)	Last glass, olivine–rich basalt (MgO = 21%)
Chemical distinction:	Most Mg-rich, low-Fe melt series	Most Mg-rich, mid-Fe melt series	Mid-Fe melt series	Most Ti-rich, mid-Fe melt series	Low-Fe melt series	Most Fe-rich, high-Fe melt series	Most P-rich, mid-Fe melt series	Most Si-rich, Mid-Fe melt series	Most Si-rich, low-Fe melt series
Quenching temperature:									
T_{MgO} (°C)^b	1141	1134	1114	1102	1099	1076	1053	—	—
T_{CaO} (°C)^b	1138	1129	1114	1110	1096	1080	1046	986	990

[a] Total iron as FeO.

[b] T_{MgO} indicates a temperature estimate based on MgO content of the glass; T_{CaO} indicates a temperature estimate based on CaO content.

dobrookite and magnetite) in a variety of assemblages and crystallization sequences. In olivine–poor core, the first oxide to crystallize is ilmenite (at 1105°–1110°C) followed by magnetite (at 1080°–1090°C). As the olivine content of the rock increases, however, the Fe/Ti ratio of the final, near-solidus oxide assemblage decreases, passing from ilmenite + magnetite, to ilmenite-dominant assemblages with minor ferropseudobrookite and magnetite, to ferropseudobrookite-dominant assemblages where groundmass spinel is absent, and only a trace of ilmenite is found.

The magnetite-ilmenite assemblage is a well-known geothermometer (BUDDINGTON and LINDSLEY, 1964) and has been used as such for the lake samples. HELZ and THORNBER (1987) compared the results of oxide geothermometry with glass geothermometry and downhole temperature measurements, and concluded that oxide and glass geothermometry gave essentially identical results, over the temperature range (980°–1090°C) where glass, magnetite and ilmenite coexist.

Melt compositions

Among the phases present in the lake, the quenched melt, or glass, present in the core varies the most in composition. The range of melt compositions present and their significance has been determined via (1) melting experiments on selected core samples (HELZ and THORNBER, 1981, 1987) and (2) electron microprobe analysis of glasses in situ in drill core and bulk chemical analysis of naturally produced liquid segregations and of artificial melt segregations ("oozes") that flowed into drill holes and were subsequently recovered. The latter data were presented briefly in HELZ (1984) and will be discussed in more detail here.

Melting experiments, at 1 atmosphere and at f_{O_2} conditions of the nickel-bunsenite buffer, have been performed on sample KI75-1-143.8 (MgO = 12.15 weight percent) and on KI67-3-83.8 (MgO = 7.54 weight percent). Details of the experimental procedures are given in HELZ and THORNBER (1987). The resulting phase relations are given in Table 2. These results agree closely with those of THOMPSON and TILLEY (1969), who used 1959 eruption samples, rather than core from the lava lake, as starting materials.

The most useful feature of these experimental results is that the MgO and CaO contents of the melts were observed to vary linearly with temperature, as follows: the MgO content of glasses coexisting with olivine is linear from 1050° to 1250°C, whereas the CaO content of glasses coexisting with olivine [or (orthopyroxene) + augite + plagioclase] is linear from 1000° to 1160°C. These results have therefore been used to assign quenching temperatures to every piece of olivine–bearing glassy core analyzed (see Table 1, for example). The shapes of the resulting profiles generally resemble those predicted by theoretical thermal modeling; the minor deviations from predicted shapes are also interpretable (see discussion in HELZ and THORNBER, 1987).

Table 2. Temperatures (°C) of first occurrence of phases in 1-atmosphere melting experiments (from HELZ and THORNBER, 1987).

Phase	K167-3-83.8	K175-1-143.8
Olivine	1169	1260
Augite	1160 ± 10	1172 ± 8
Plagioclase	1160 ± 10	1163 ± 2
Ilmenite or pseudobrookite	1100 ± 10	1106 ± 6
Magnetite	1080 ± 10	1090 ± 10

Figure 3 shows all bulk analyses and electron microprobe analyses from the 1967–1981 drill core for which MgO < 11 weight percent. Most analyses of the olivine-phyric samples lie at higher bulk MgO contents. They were omitted to better display the full range of compositions seen in the interstitial glasses. In addition, representative analyses of a variety of interstitial glasses are presented in Tables 1 and 3.

Inspection of Figure 3 shows the resulting suite of microprobe analyses to be internally coherent and consistent with the wet chemical analyses of segregation veins and oozes. The analyses also provide a very complete determination of the line of descent of the liquid in olivine tholeiite. The smoothness and tightness of the variation of most oxides testify to the internal coherence of the data set. The fact that compositions of high-temperature (1100°–1140°C) glasses largely overlap wet-chemical analyses of oozes and segregation veins (samples with MgO = 2.61 to 5.96 weight percent, see also Figure 8, below), suggests that the glass compositions are of acceptable accuracy. Given that, and the geothermometer scale implicit in the glass compositions (shown along the top of the individual plots in Figure 3), one can determine the sequence of crystallization of phase in the lake, and the temperature at which they first appear, by inspection, as follows:

(1) The first silicate phase to crystallize is olivine. The liquidus temperature of individual samples varies from 1170°C (at MgO = 7.5 weight percent) to $\gg$1200°C, depending on the olivine content of

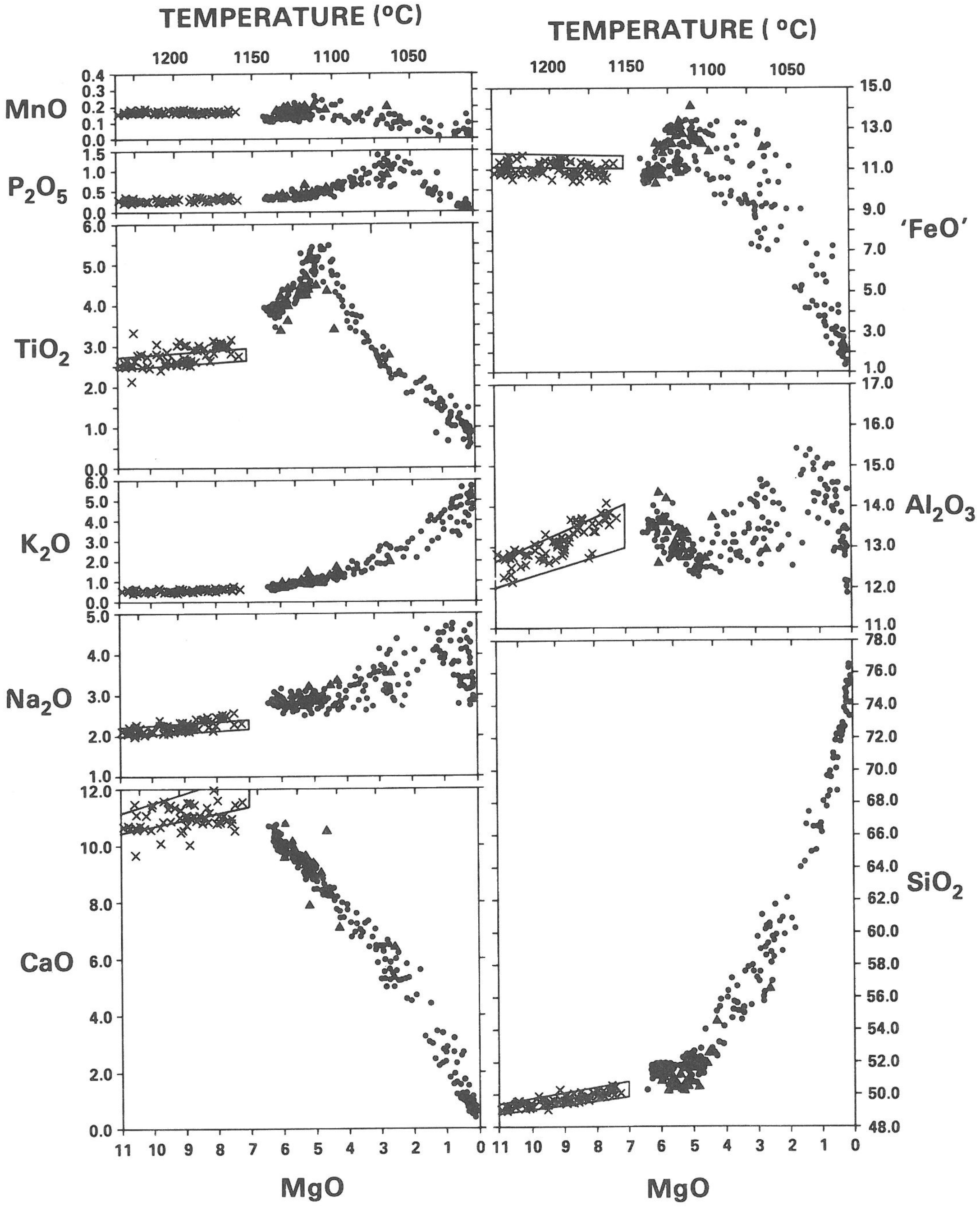

FIG. 3. Composition of all analyzed bulk samples and interstitial glasses from Kilauea Iki lava lake with MgO ≤ 11 weight percent, plotted against their MgO content. All quantities are in weight percent, with FeO = total iron as FeO. Wet chemical analyses of olivine–phyric rocks are indicated by crosses; those of segregation veins and oozes are shown by solid triangles. Electron microprobe analyses of interstitial glass are shown as small black dots. The temperature scale at the top gives the quenching temperature of each glass, corresponding to its MgO content, based on the calibration of HELZ and THORNBER (1987). Rectangular field at MgO = 7–11 weight percent outlines the field of composition of low-MgO eruption samples.

Table 3. Additional glasses from Kilauea Iki laval lake.

Sample no.	1 KI81-3-185.5	2 KI81-3-185.5	3 KI79-1-158	4 KI79-1-158	5 KI79-1-158	6 KI79-1-158	7 KI79-1-158
No. points analyzed	8	10	2	2	2	2	14
SiO_2	65.2	62.9	43.1	70.7	40.1	72.1	73.5
TiO_2	1.90	1.70	6.80	1.30	6.48	1.04	.59
Al_2O_3	14.4	12.9	4.56	11.1	4.11	10.9	12.7
Cr_2O_3	.00	.00	.00	.00	.00	.00	.00
FeO	5.05	9.28	26.6	5.56	29.9	5.51	2.69
MnO	.07	.12	.32	.04	.40	.06	.02
MgO	1.39	1.62	5.24	.94	3.99	.60	.19
CaO	2.87	4.60	9.81	1.97	9.59	1.68	.92
Na_2O	4.22	3.57	1.35	(2.66)[a]	1.32	(2.76)[a]	4.24
K_2O	3.73	2.46	.62	4.22	.48	3.65	4.41
P_2O_5	.58	.80	1.94	.12	3.05	.12	.08
Sum	99.4	100.0	100.3	98.6	99.4	98.4	99.4
Comments:	Glass in olivine-phyric matrix	Glass in segregation vein	Coexisting immiscible liquids included in plagioclase in a segregation vein				Glass outside plagioclase, same thin section
	Both in same thin section, but not coexisting						

[a] Total iron as FeO.

[b] Parentheses indicate probable loss of Na_2O during analysis.

the rock. [The most magnesian glass found in the eruption samples was quenched from 1216°C (HELZ, 1981a), corresponding to MgO = 10 weight percent. Samples with higher MgO contents were not erupted as liquids.]

(2) Augite begins to crystallize between 1175° and 1160°C (see CaO *vs* MgO, Figure 3). No drill core this hot has been recovered, so the CaO peak is not defined very precisely.

(3) Plagioclase begins to crystallize between 1165° and 1145°C (Al_2O_3 *vs* MgO). Again, no core was recovered from this temperature range, so the Al_2O_3 peak is located only approximately.

(4) A TiO_2-rich Fe–Ti oxide (either ilmenite or ferropseudobrookite) begins to crystallize at 1110 ± 5°C (see TiO_2 *vs* MgO). The TiO_2 peak is very sharp in glasses from any individual drill hole; the slight broadening of the peak in Figure 3 probably results from the fact that the two different oxide phases come in at slightly different temperatures. Note that the peak TiO_2 content (~5.45 weight percent, see column 4, Table 1) is the same regardless of which oxide is crystallizing.

(5) A broad peak in FeO reflects the first appearance of hypersthene or pigeonite, ± magnetite at 1090 ± 10°C, as well as the incoming of the first Fe–Ti oxide at 1110 ± 5°C.

(6) Apatite begins to crystallize at 1065 ± 10°C (see P_2O_5 *vs* MgO).

As for the other major oxides, MnO follows TiO_2, whereas SiO_2 and K_2O increase monotonically in the liquid. This increase reflects the absence of K-feldspar and the scarcity of coexisting cristobalite + glass, noted above. The Na_2O variation is problematical. It may drop off slightly at extremely low MgO contents, but most of the decline seen in Figure 3 is more likely the result of loss of Na_2O under the electron microprobe beam.

This liquid descent line does not reflect either perfect equilibrium or perfect fractional crystallization. Rather it is a hybrid, especially in the high-temperature olivine-phyric samples. Olivine re-equilibrates with the liquid (HELZ, 1987a) as do chromite (SCOWEN, 1986) and the Fe–Ti oxides (HELZ, unpublished data; HELZ and THORNBER, 1987). At the other extreme, plagioclase is markedly zoned, as noted above. The pyroxenes are intermediate between olivine and plagioclase in the degree of compositional variability exhibited.

As temperature decreases in the olivine-phyric samples, at some point all the remaining glass becomes isolated from the olivine, and the liquid compositions produced by subsequent crystallization approach a perfect fractional crystallization trend. The temperature at which this occurs lies between 1050° and 1080°C. The characteristic endpoint liquids in olivine-phyric rock (see, for example columns 7 and 8 in Table 1) are thus produced by extreme in situ fractional crystallization, and are not in equilibrium with olivine.

In the segregation veins, by contrast, the pyroxenes as well as plagioclase exhibit strong zoning, so that liquids in the veins should approach a perfect fractional crystallization trend from the start. This kind of hybrid fractionation trend should not be peculiar to the lava lake, but may occur in other magma chambers as well.

It is obvious from Figure 3 that the concentration of some oxides inthe melt is controlled quite closely by the crystalline assemblage. Thus olivine ± low–Ca pyroxene controls MgO, ilmenite or ferropseudobrookite controls TiO_2, and the assemblage olivine + augite + plagioclase constrains CaO. Other oxides, notably FeO and Al_2O_3, appear to be much more variable.

The wide spread in FeO contents results from the re–equilibration of olivine with the liquid as crystallization proceeds (HELZ, 1984). As the olivine grows more Fe-rich as a result of this in-situ re-equilibration, it takes iron from the adjacent interstitial melt. The more olivine present initially, the more severely depleted in iron the liquid becomes, at a given MgO content. Core samples from the 1967 drill holes, which bottomed in the olivine–depleted zone, thus all contain interstitial glasses that are relatively Fe-rich (see *e.g.*, Column 6, Table 1). Samples from the partially molten zones of KI79-5 and from the 1981 drill holes, which come from the most olivine–rich parts of the lake, contain glasses that are very Fe-depleted. Core samples from the 1975, 1976 and the other 1979 holes contains glasses of intermediate iron content. This effect is clearly displayed in Figure 4, where liquid descent lines for three individual holes [KI67-3, KI75-1 and KI81-1, from the southern–most cluster in the center of the lake (Figure 2a,b)] are shown separately.

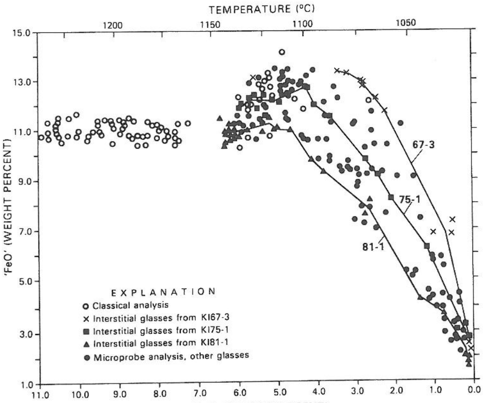

FIG. 4. FeO *vs* MgO, for the same samples as in Figure 3. Glasses from three drill holes are distinguished, illustrating the succession of liquid compositions for samples from within individual holes. Temperature scale as in Figure 3.

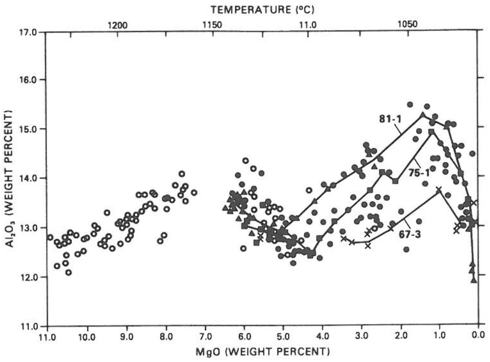

FIG. 5. Al_2O_3 *vs* MgO, for the same samples as in Figure 3. Symbols for samples from holes KI67-3, KI75-1 and KI81-1 as in Figure 4.

The effect on FeO in the glass of varying olivine content can also be seen wherever a segregation vein cuts an olivine–rich host. An extreme example of this is shown by the analyses in Columns 1 and 2 of Table 3, for two melts from a single thin section, quenched from 1025°–1040°C. The glass in the segregation vein contains 9.28 weight percent FeO, while that in the adjacent wall rock, a few millimeters away, contains 5.05 weight percent FeO. This difference exists because the melt composition is controlled by *local* bulk composition, even though a significant amount of melt is present.

The variation in FeO in the melt imposed by re-equilibration of olivine is compensated for by increases in all the other, relatively unbuffered, oxides (SiO_2, Al_2O_3, Na_2O, K_2O and P_2O_5). The apparent scatter in Al_2O_3 *vs* MgO (Figure 5) thus resolves into a set of nested curves, tightly defined for any one drill hole, but offset from each other so as to produce a wide smear when all data are superimposed.

Figure 6 shows SiO_2 *vs* MgO variation for different drill holes: at a given MgO content and hence

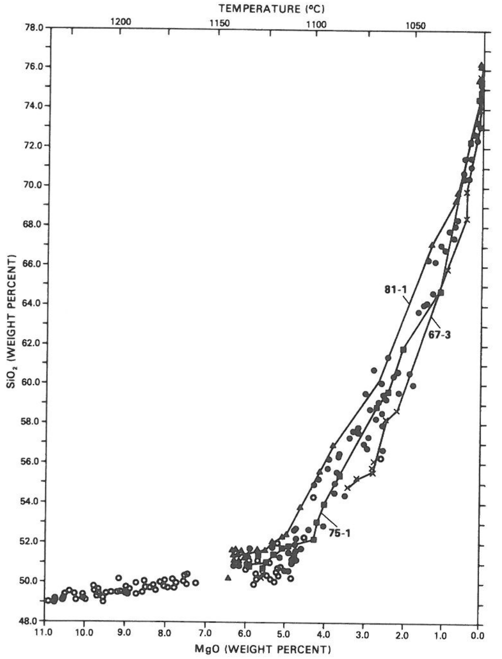

FIG. 6. SiO_2 *vs* MgO, for the same samples as in figure 3. Symbols for samples from holes KI67-3, KI75-1 and KI81-1 as in Figure 4.

temperature, the melts in the most olivine-rich rocks are the richest in SiO_2 and alkalies, as well as Al_2O_3. These effects, like the FeO effect, can also be seen in melts from the segregation vein/matrix pairs within a single thin section (see again columns 1 and 2, Table 3). The *maximum* SiO_2 enrichment reached by the interstitial glass in olivine-phyric rock decreases as olivine content increases, of course; this can be seen by comparing columns 8 and 9 in Table 1.

The uppermost levels of SiO_2 enrichment seen in Figures 3 and 6 are attained by rest liquids in the segregation veins (see, *e.g.*, column 7, Table 3). As segregation veins are all olivine–poor, the various liquid descent lines tend to converge in this range. Because the liquid that makes up the bulk vein was derived in part from the adjacent rock, analysis of the interstitial liquids within the segregation vein show that they too reflect the olivine content of the adjacent host rock. Accordingly, peak SiO_2 contents ($\geq$76 weight percent SiO_2) are found in the residual liquids in segregation veins in olivine–rich host rock. Thus, initial differences in liquid composition, established by re-equilibration of olivine at 1100–1150°C propagate through all subsequent crystallization/fractionation events, to the last drop of liquid.

These varying liquid descent lines affect the nature of the later crystalline phases, of course. Orthopyroxene is the low-Ca pyroxene in olivine-rich samples, whereas pigeonite is found in olivine-poor core and in the segregation veins. The effects on the Fe–Ti oxide assemblages have already been noted above.

One feature of these glass data worth noting is the temperature at which compositional divergence starts. In the olivine-phyric rock, it clearly begins at 1125°–1140°C. Liquids in segregation veins start to diverge from those in the host at temperatures of 1070°C in most core, by 1080°–85°C in some of the deepest veins, where the compositional contrast with the host rock is largest. In both cases, the melt does not occur in isolated pools, but remains continuous. At the higher temperatures, the crystal network is barely self-supporting, and yet melt compositions start to reflect the local bulk compositional environment. This observation implies that interstitial liquids in other bodies, such as layered intrusives, may also start to differ between adjacent layers at relatively high melt contents, unless circumstances favor circulation of such melts out into the main chamber. This behavior also raises the possibility that one could get more than one group of lavas, significantly different in composition, from a single magma chamber, not because of varying source chemistry or degree of contamination, but by tapping different parts of a chamber undergoing crystal-liquid fractionation.

Liquid immiscibility is not a significant petrogenetic process in Kilauea Iki lava lake. It occurs only at a very late stage of crystallization, and then principally in melts included within plagioclase. Plagioclases in segregation veins, which are coarse grained and may contain large melt inclusions, locally contain spectacular examples of coexisting liquids, however. Two pairs of such coexisting liquids are given in Table 3 (Columns 3–4, 5–6). The segregation vein in which they occur was quenched from 985°C. The extreme compositions of the mafic liquids are unlike any known basalt in Hawaii.

PROCESSES ACTIVE IN KILAUEA IKI

One of the principal objectives of the Hawaiian lava lake studies has been to characterize, as completely as possible, the physical processes responsible for producing the chemical variations observed in the lakes. A summary of the processes active in Kilauea Iki, Alae and Makaopuni was given in WRIGHT *et al.* (1976). An updated summary is given in WRIGHT and HELZ (1987). What follows is an

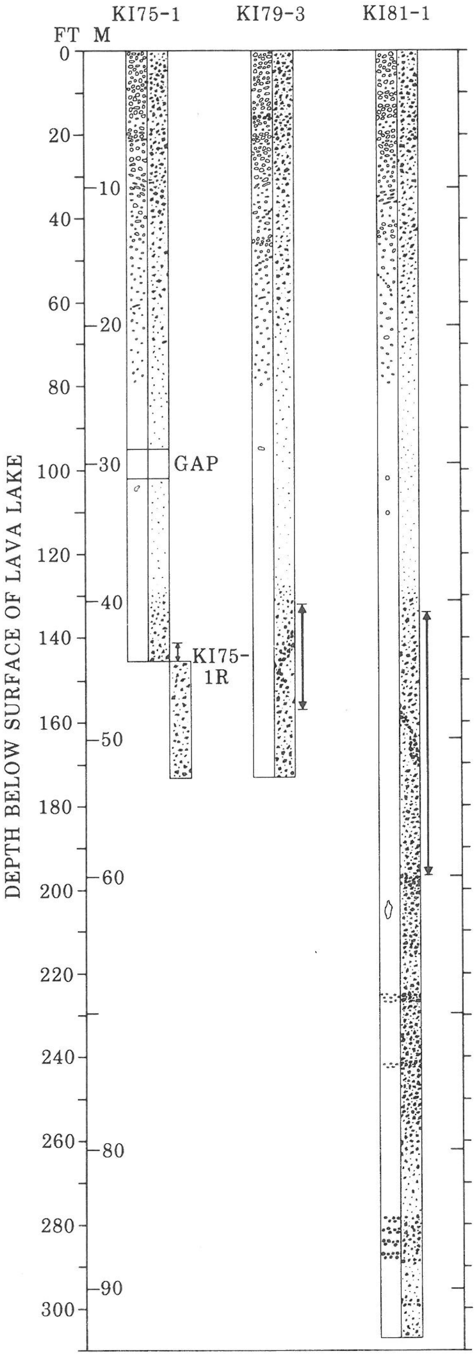

FIG. 7. Schematic representation of the distribution of vesicles (left-hand columns) and olivine phenocrysts (right-hand columns) with depth, for three drill holes from the center of Kilauea Iki lava lake. Hole KI75-1R is an extension of hole KI75-1, drilled in December 1978 (HELZ and WRIGHT, 1983). Symbols convey qualitative information on the size, shape and clustering patterns of vesicles and olivine crystals, as well as on their relative amounts. Arrows beside the columns indicate the depth range in which vorbs are particularly well-developed in each drill hole. Very deep vesicular layers at 85–90 m in KI81-1 are filled with melt: this interval is all foundered crust (HELZ and WRIGHT, 1983).

expanded version of the latter summary, as it applies to Kilauea Iki lava lake.

Olivine settling and compaction

Gravitative settling of coarser phenocrystic olivines has occurred in Kilauea Iki, producing a zone between 10 and 40 m depth in which olivine content is <3% and olivines >3 mm^2 are essentially absent (MANGAN and HELZ, 1985). The process did not become efficient until 1963 (HELZ, 1980). The final distribution of olivines in the central part of the lake is illustrated semi-quantitatively in Figure 7. Some size sorting is observed, so that average grain size and amount of phenocrysts are positively correlated. On plots of SiO_2 *vs* MgO and Al_2O_3 *vs* MgO (Figure 3) it can be seen that most of the olivine-bearing samples define a trend parallel to the original field of compositions of the eruption pumices, a field that can be extrapolated to an olivine composition of Fo_{86}. Thus olivine redistribution was achieved while the olivine phenocrysts retained their original compositions (average Fo_{86}), that is, before significant cooling of the lake had occurred (HELZ, 1987b). This conclusion implies that olivine redistribution took place at $T \geq 1180°C$ (HELZ, 1987b; HELZ and THORNBER, 1987).

It has been observed (HELZ, 1980; WRIGHT *et al.*, 1976) that settling of olivine in this and other lava lakes was not nearly as efficient as would be predicted from Stokes' law calculations. The decrease in olivine size and content observed between 10 and 20 m depth correlates with a decrease in vesicularity in the same range, suggesting that early settling of olivine was inhibited by the presence of vesicles (HELZ, 1980). No actual flotation of olivine crystals by individual vesicles seems to be involved (MANGAN, unpublished data). Rather the idea is that the presence of vesicles gives the melt the properties of a Bingham plastic rather than a Newtonian fluid, as was inferred by SHAW *et al.* (1968) for the 1965 Makaopuhi lava lake. If so, the gradual decrease in olivine size and content with depth in the 10–20 m range would reflect a gradual decrease in the yield strength of the Bingham-body melt as vesicle content decreased. M. Mangan has quantified the size and distribution of olivine and vesicles in the core, in order to verify this model (MANGAN and HELZ, 1985).

A related problem is the diffuseness of the cumulate zone, in which olivine content never exceeds 50% and is usually <40%, with the olivines rarely touching each other. HELZ (1980) suggested that migrating olivines carried with them a stagnant boundary layer of melt which slowed their settling and inhibited compaction. Such "skins" of melt have not been observed in lake samples, but some olivine crystals in the eruption pumices do exhibit boundary layers, detectable because of their contrasting composition (HELZ, 1987a).

Vesicle movement and retention

The distribution of vesicles in the lake, shown semiquantitatively in Figure 7, has been investigated quantitatively by M. Mangan (MANGAN and HELZ, 1985). The vesicles are of interest because they demonstrate the existence of a gas phase in the lake, and allow us to monitor the degassing history of the lake. There is some evidence (HELZ and WRIGHT, 1983) that loss of original gases was complete before 1979. Nevertheless, bubble-driven processes continued to be active in the lake, and late-stage vesicle accumulation at the final position of the thermal maximum was observed in the 1981 drill holes (see Figure 7, vesicles at 70–75 m depth). This and other lines of evidence suggest that the lake may have picked up meteoric water, probably through cracks in the sides and bottom of the lake.

The other reason for studying vesicles is that they are extremely mobile in the lake because of their low density, and their ability to coalesce and to deform as they pass through a crystal framework. Hence they interact with other processes, triggering some and inhibiting others. Olivine settling may have been greatly slowed by the flux of vesicles rising from below. Large-scale convective overturn of the lake appears to have been virtually eliminated, in part because of vesicle gradients. The vesicle contents involved are mostly <10% (MANGAN and HELZ, 1985). Evaluating the effects of such moderate vesicle contents on the various processes that occur in the lake should be relevant to processes in other subvolcanic magma chambers as well.

Convective redistribution of crystals

Within the olivine–depleted zone, small (1–2 mm) clots of augite microphenocrysts are present in core samples from near the edge of the lake, and largely absent from samples from the middle. This differential concentration of augite raises the CaO content of the core from 10.5 weight percent in the middle to 11.6 weight percent CaO at the edge. The small olivine phenocrysts present show a similar pattern. The implied lateral redistribution of crystals seems most readiy explicable by convection within the olivine–depleted zone. Flow differentiation by convection was a major process in the 1965 Makaopuhi lava lake (WRIGHT and OKAMURA, 1977), and affected plagioclase distribution as well as olivine + augite in that lake. In Kilauea Iki, by contrast, flow differentiation of this sort was a minor process, confined to temperatures above the incoming of plagioclase and below the incoming of augite and probably confined to the most olivine–depleted part of the lake (20–40 m) (HELZ, 1987b).

Diapiric melt transfer

"Diapiric melt transfer" refers to a process whereby relatively low-density melt from within the olivine–rich mush at the base of the lens of melt extricates itself from that mush, rises through the main lens of melt, to pool and mingle with the melt at the base of the upper crust. This process, documented in HELZ (1986, 1987b), involves a compositionally unique melt present at 1145°–1160°C, at and just below the incoming of plagioclase. Not coincidentally, this melt is lower in density than any other along the liquid descent line at temperatures >1100°C.

The process has been hypothesized to occur because it was observed (HELZ, 1980, 1987b) that core from the lake shows strong chemical zonation with depth, in which the upper part of the lake is enriched in TiO_2 and incompatible elements, and depleted in FeO and CaO, whereas deeper parts of the lake show the opposite signature. Details of the pattern of chemical variation, and the very small grain size of augite and plagioclase in the lake, make it very unlikely that this zonation was produced by large–scale downward movement of olivine + augite + plagioclase crystals (HELZ, 1987b). Earlier, with only the upper, enriched zone available for analysis, HELZ (1980) tentatively ascribed the enrichment to filterpressing of liquid formed during partial remelting of foundered crust, deep in the lake. Recovery and analysis of deeper core has led to our present recognition, first that this process does not require foundered crust, and second that it has affected the chemistry of the lake at virtually all levels. The only more pervasive process was olivine settling.

The melt extraction process is very efficient. The depleted zone at 56–78 m has lost 21–42% liquid. Some aspects of it are still not well understood, however. For example, we do not know the minimum value of the density inversion in the melt column required for the process to occur. From knowledge of the rate at which the lake has crystallized, gained from repeated drilling of the lake, we can infer that diapiric melt transfer started in 1960 and ceased around 1971.

Formation of vertical olivine–rich bodies

The core between 40 and 58 m depth contains the curious vertical olivine–rich bodies (vorbs) described above. They are hypothesized to form as streams of vesicles ± melt rise from within the lower crust, interact with the olivine–rich core of the lake and emplace themselves in the base of the upper crust. The vesicles are hypothesized to entrain first melt, then coarse olivines, from the lower part of the lake. As the plumes rise into the cooler material they begin to segregate, the vesicles + melt eventually detaching themselves from the trail of coarse olivines. These olivine–rich bodies contain excess iron-rich olivine (Fo_{77-79}) and differentiated liquid, relative to the adjacent host rock, so they are active at lower temperatures than is the diapiric transfer process (HELZ, 1987b). Analogous vuggy, vertical bodies also occur in the interval between 22–40 m, but they are more difficult to spot, as they lack coarse olivines (HELZ, 1980). The melt- and vesicle–rich top of the vorbs, where present, looks like a segregation vein, which led HELZ (1980) to suggest that segregation veins might be related to these bodies in some way.

Formation of segregation veins

Segregation veins, as discussed earlier, are coarse-grained diabasic sills formed by processes internal to the lake. Such veins are more abundant in Kilauea Iki than in the other lava lakes. The most important subgroup forms an apparently continuous set of sills, concentrated between 18 and 56 m depth, which make up 5–10% by volume of the upper crust of the lake (HELZ, 1980). Bulk compositions of the segregation veins correspond to liquids produced at 1135°–1105°C, estimated by using the calibration of HELZ and THORNBER (1987), as shown in Figure 8. In Kilauea Iki, they are more magnesian and presumably form at higher temperatures than segregation veins in Alae, Makaopuhi, and the prehistoric Makaopuhi lava lakes. The process by which they form is some sort of filterpressing. Temperature profiles through partially molten crust surrounding segregation veins have anomalous shapes above the veins (HELZ and THORNBER, 1987) suggesting that the melt in the segregation vein was hotter than the overlying crust and hence was derived from deeper in the lake. Samples with the chemical signature indicative of loss of segregation-vein liquid are found at almost all levels in the lake, except for the 56–78 m interval. This clearly implies that some of the material in any given segregation vein may be derived from the crust near the vein. The presence of such samples at depths below 78 m, however, suggests that some of the material now present as segregation veins was transported, via the vorbs, from below 78 m to the 18–56 m depth range, (see HELZ, 1987b for a discussion of the variation of core chemistry with depth) though the full details of this process remain to be worked out.

Formation of late differentiates

Locally the core contains late fractures or large vesicles partly or completely filled by highly differentiated liquids, with liquidus temperatures of 1060°–1000°C. Figure 8 shows the compositions of some of these minor late differentiates plotted on a magnesia variation diagram, along with the bulk compositions of segregation veins and oozes ($MgO = 4–6$ weight percent) and the olivine-phyric matrix rock ($MgO \geq 7.5$ weight percent). The fractures are produced by brittle failure of the lake as it shrinks during crystallization and cooling. The compositional gap between about 4.2 and 3.0 weight percent MgO is probably real. This gap corresponds to the temperature range over which the segregations have already formed within the crust, but are not yet rigid enough to break. So long as the crust contains a few slushy layers, stress cannot build to the breaking point. Once the veins have "set," pervasive fracturing becomes possible. It is probably not coincidental that the highest-temperature late differentiates have the composition of the liquid recovered from a half-crystallized, barely drillable segregation (c.f., Figure 8 and the ooze analysis in Table 3, Col. 5, HELZ, 1980).

The bulk compositions of these tiny bodies are andesitic to rhyolitic ($SiO_2 = 56–70$ weight percent; $MgO = 3.0–0.5$ weight percent). Their existence demonstrates the mobility of interstitial liquid even when very little (<10 percent) melt is present and confining pressures are low (HELZ, 1984). The driving mechanism is inferred to be gas filterpressing (ANDERSON *et al.*, 1984). In deeper, larger intrusives, the process of segregation might be different, but the following observations are still relevant. First, basalt can fractionate to produce dacitic and rhyolitic liquids. Second, when partially molten rock fractures, the interstitial melt can move, even at very high melt viscosities.

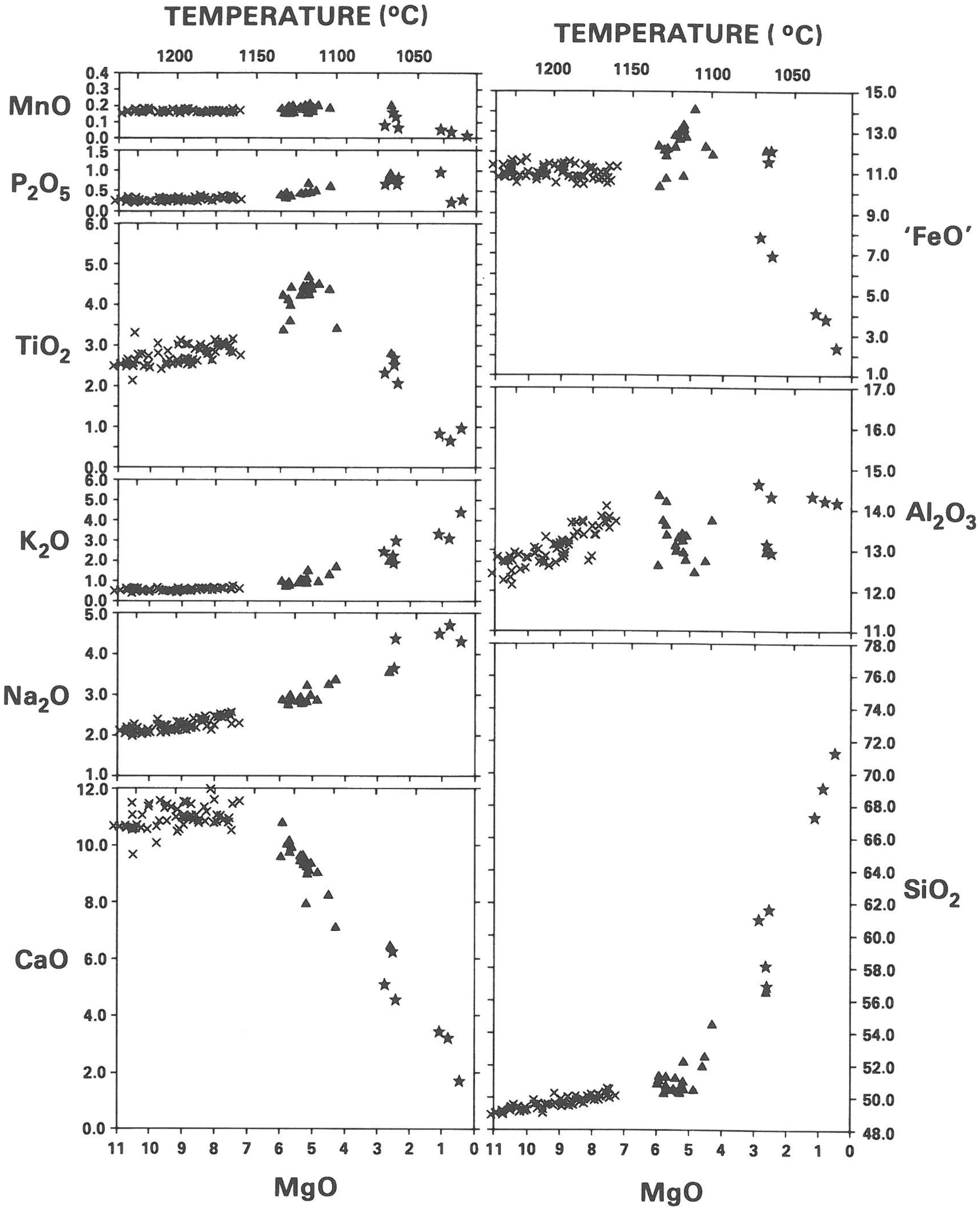

FIG. 8. Bulk analyses of differentiated samples from Kilauea Iki lava lake, plotted against MgO. All quantities in weight percent. Wet chemical analyses of olivine-phyric rock (crosses) and of segregation veins and oozes (triangle) as in Figure 3. Bulk compositions shown by stars were obtained by microprobe analyses of fused glasses and/or by rastered-beam analyses of the differentiated veinlets in polished section. Where both methods were used they gave comparable results.

SUMMARY COMMENTS

This overview of studies, past and present, of Kilauea Iki lava lake should give some feeling for the range of studies possible and for their potential petrologic applications. The work does not exhaust the possibilities. Other aspects of the lava lake's behavior (trace element geochemistry and partitioning, detailed thermal modeling, for example) are still in the earliest stages or have not been done at all.

The most important feature of Kilauea Iki is that it is not a fossil system. In deciphering the history of the lava lake, for example, it has been invaluable to have drill core that records many different stages of the lake's crystallization history. The main reason we can state with confidence that the diapiric melt transfer process began in 1960 and ended about 1971, or that the vorbs began to form by at least 1967 and ceased in 1979 is precisely because we know directly from the drilling results how the crust grew with time, and hence how the position of particular isotherms varied with time. The fact that the melt diapirism and the vorbs started and stopped independently in turn confirms that these two diapiric processes operated independently of each other, and hence that from 1967–1971, there were two different populations of diapirs migrating from bottom to top within the lens of melt.

The simple availability of any partially molten core is extremely important because it enables us to "capture" processes in progress. Recognition of the vorbs, in particular of their textures and significance, would have been much more difficult if only subsolidus core were available for study. Similarly, the fact that we have drilled into segregation veins at all stages of their formation (HELZ, 1980; HELZ and WRIGHT, 1983) tells us where and at what temperature they form.

Another valuable set of constraints has been provided by the detailed field observations on the 1959 eruption, and by the availability of samples collected day by day as the eruption proceeded. Information on the composition and petrography of the material fed into the lake, plus knowledge of the sequence in which it was erupted, permits us to conclude unequivocally, for example, that the present distribution of olivine phenocrysts in the lake is not inherited from the eruption. Furthermore, the observation that olivine-phyric samples lie on Fo_{86} control lines in certain projections, just as the eruption pumices do, constrains the time and temperature at which olivine settling took place. The latter information is one of several pieces of data that rule out three-phase crystal settling of olivine + augite + plagioclase in the lake, which in turn led to the recognition of the process of diapiric melt transfer (HELZ, 1987b).

Petrologic features of the lake that have emerged as being critical to the occurrence of particular processes discussed above include (1) melt density gradients (2) the presence and abundance of vesicles and (3) the exact crystallinity of the partially molten parts of the lake. The latter has exerted a major control on all processes. As noted in HELZ (1987b) the presence of 17–20% initial olivine in Kilauea Iki *vs* <5% in the 1965 and prehistoric Makaopuhi lava lakes was sufficient to curtail lake-wide convection, with flow differentiation, as a major fractionation process. Also, because of the higher crystallinity in Kilauea Iki, other processes (such as segregation-vein formation) occur at higher temperature in Kilauea Iki than in the other lakes; hence the composition of the segregation veins is different. Drilling behavior is also affected: the crust-melt interface temperature in Kilauea Iki varies as a function of olivine content and has almost always been higher (HELZ and WRIGHT, 1983) than the 1075°C typically observed in Alae and Makaopuhi (WRIGHT *et al.,* 1976). There is no unique crust-melt interface temperature, therefore. What is important is the percent crystallinity, and to a lesser extent, the shape of the crystals present (HELZ, 1980). Quantification of the physical processes and the *physical* parameters that control them in the lava lake, may eventually make it possible to use our knowledge of the lava lakes to decipher magmatic processes within other, larger magma chambers.

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