Characteristics of Hawaiian Volcanoes
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Chapter 6

Petrologic Insights into Basaltic Volcanism at Historically Active Hawaiian Volcanoes

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Abstract

Study of the petrology of Hawaiian volcanoes, in particular the historically active volcanoes on the Island of Hawaiʻi, has long been of worldwide scientific interest. When Dr. Thomas A. Jaggar, Jr., established the Hawaiian Volcano Observatory (HVO) in 1912, detailed observations on basaltic activity at Kīlauea and Mauna Loa volcanoes increased dramatically. The period from 1912 to 1958 saw a gradual increase in the collection and analysis of samples from the historical eruptions of Kīlauea and Mauna Loa and development of the concepts needed to evaluate them. In a classic 1955 paper, Howard Powers introduced the concepts of magnesia variation diagrams, to display basaltic compositions, and olivine-control lines, to distinguish between possibly comagmatic and clearly distinct basaltic lineages. In particular, he and others recognized that Kīlauea and Mauna Loa basalts must have different sources.

Subsequent years saw a great increase in petrologic data, as the development of the electron microprobe made it possible to routinely monitor glass and mineral compositions, in addition to bulk rock compositions. We now have 100 years' worth of glass compositions for Kīlauea summit eruptions, which, together with expanding databases on prehistoric tephras, provide important constraints on the nature of Kīlauea's summit reservoir. A series of chemically distinctive eruptions in the 1950s and 1960s facilitated evaluation of magma mixing and transport processes at Kīlauea. At Mauna Loa, lava compositions are distinctive only at the trace element level, suggesting that its summit reservoir is better mixed than Kīlauea's. Most summit lavas at both volcanoes, however, lie on olivine control lines having the same olivine composition ($Fo_{86–87}$). Study of the ongoing East Rift Zone eruption at Kīlauea has further illuminated the complexity of magma storage, resupply, and mixing along this very active rift zone.

Studies of active and closed-system lava lakes have been part of HVO's efforts since Jaggar's unique descriptions of the Halemaʻumaʻu lava lake that existed before 1924. Detailed study of closed-system bodies, including the 1959 Kīlauea Iki, 1963 ʻAlae, and 1965 and prehistoric Makaopuhi lava lakes and the Uēkahuna laccolith, have allowed recognition and quantification of processes of basalt differentiation. Specific topics reviewed herein include the occurrence of segregation veins and related structures, overall cooling history, and patterns of crystallization and reequilibration of olivine in various lava lakes.

In recent decades, study of the submarine slopes of the Island of Hawaiʻi and of Lōʻihi Seamount has revolutionized our understanding of the early history of Hawaiian volcanoes. Observations of Lōʻihi lavas first established the existence of an early alkalic stage in the evolution of Hawaiian volcanoes. Stages of volcanic development from inception to tholeiitic shield building can be observed in Kīlauea's submarine and subaerial sections. One distinctive feature of submarine volcanics at Kīlauea, Mauna Loa, and Hualālai is that picritic lavas are more abundant than in subaerial eruptions. Also, olivine compositions of submarine lavas are more magnesian, ranging from Fo_{88} at Kīlauea to Fo_{89} at Hualālai. The most magnesian glasses known from Kīlauea (MgO=14.7–15.0 weight percent) were found along the submarine part of Kīlauea's East Rift Zone.

Contributions to our knowledge of the nature of the mantle source(s) of Hawaiian basalts are reviewed briefly, although this is a topic where debate is ongoing. Finally, our accumulated petrologic observations impose constraints on the nature of the summit reservoirs at Kīlauea and Mauna Loa, specifically whether the summit chamber has been continuous or segmented during past decades.

Introduction

The youngest volcanoes of Hawaiʻi have attracted volcanologists to the Island of Hawaiʻi for almost two centuries because, first, they erupt frequently, being among the most active volcanoes in the world; second, Hawaiian eruptions over the past 200 years have usually been effusions of basaltic lava,

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rather than violent explosions, so they can be approached and observed in reasonable safety; and finally, Hawai‘i’s geographic isolation from all continents makes these volcanoes exemplars for basaltic volcanism, uncomplicated by interactions with continental crust. This chapter reviews our petrologic knowledge of four volcanoes: Kīlauea, Mauna Loa, Hualālai (which last erupted in 1801), and Lō‘ihi, a submarine shield volcano off the southeast coast of the Island of Hawai‘i (fig. 1), all of which have erupted since 1790. Among them, they exhibit three of the main stages of Hawaiian volcanism: the preshield alkalic stage (Lō‘ihi), the tholeiitic shield-building stage (Kīlauea and Mauna Loa), and the postshield alkalic stage (Hualālai), as described in Macdonald and Katsura (1964), further developed by Clague and Dixon (2000) and summarized in Clague and Sherrod (this volume, chap. 3). Of these, Hualālai, Mauna Loa, and Lō‘ihi lie on the geochemically defined Loa line (as shown by Jackson and others, 1972, expanding on an observation of Dana, 1849), whereas Kīlauea lies on the Kea line (fig. 1). Volcanoes on these two lines differ in isotopic compositions (Tatsumoto, 1978), as well as in the character of their postshield alkalic caps (Wright and Helz, 1987), so the four active volcanoes exhibit most of the range of behavior of older Hawaiian volcanoes.

Scientists from the Hawaiian Volcano Observatory (HVO) and their collaborators have pursued a wide range of petrologic and chemical studies at these volcanoes. These studies have elucidated processes of basalt crystallization and fractionation, magma mixing, and transport in the volcanic plumbing system, both subaerial and submarine. Lastly, Hawai‘i’s erupted materials, especially from the currently active volcanoes, continue to serve as a test-bed for studies of various trace-element and isotopic systems, allowing scientists to evaluate their behavior during generation, transport, degassing, fractionation, and mixing of basalt, free from contamination by crustal and most sedimentary rocks.

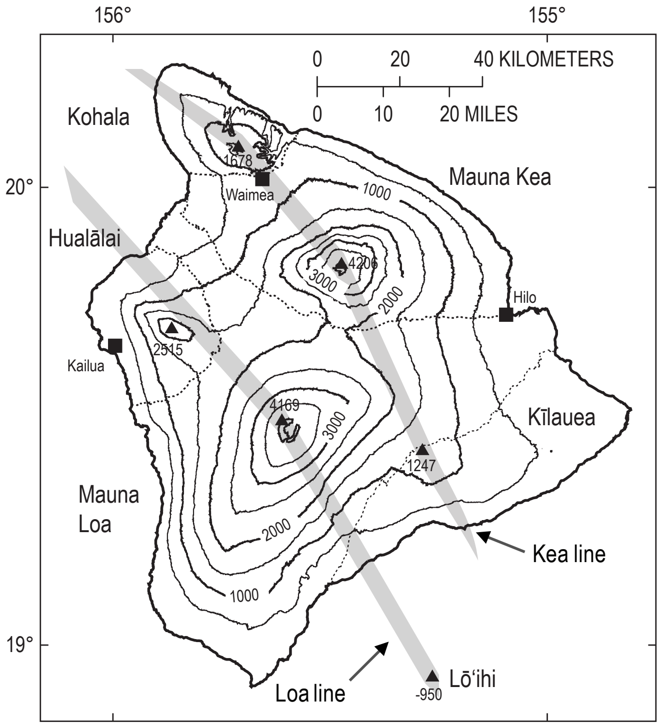

Figure 1. Index map of the Island of Hawai‘i, showing the five volcanoes that make up the island (delineated by dotted lines; triangles mark summits) and the active submarine volcano Lō‘ihi to the south. After Peterson and Moore (1987). Gray bands show trajectories of the Loa and Kea geochemical lines. Elevations in meters; contour interval 500 m.

The topics reviewed herein include (1) petrologic work at HVO from 1912 to 1958, which focused on eruptions at Kīlauea and Mauna Loa, (2) petrologic studies of subaerial activity at Kīlauea and Mauna Loa, especially summit activity from 1959 to the present, (3) Kīlauea’s ongoing East Rift Zone eruption, in its 30th year as of the time of this writing, (4) selected topics on Kīlauea lava lakes, and (5) the petrology of the submarine sections of these four historically active volcanoes. The final section will consider what this work, in particular that of HVO scientists, has contributed to our understanding of primary tholeiitic magma compositions and their origin, the origin of preshield lavas, and the nature of the summit magma chambers of Kīlauea and Mauna Loa.

Early History of Petrology at HVO (1912–58)

Many of the earliest scientific studies of Kīlauea and Mauna Loa were petrologic in nature, including analyses of lava flows and descriptions of their mineralogy and texture, carried out by scientific visitors to Hawai‘i (see reviews in Peterson and Moore, 1987; Wright and Helz, 1987). The 1912 founding of HVO, located on the rim of Kīlauea Caldera, provided greater continuity of observation, and the work of Thomas Jaggar and others associated with HVO greatly increased the extent and quality of observations of eruptive behavior and degassing activity. These early reports have been compiled in Bevens and others (1988) and Fiske and others (1987) to make them more accessible to modern researchers. The collection and analysis of lava samples was not the main focus of the early efforts, but the scientific community continued to collect, analyze, and slowly build up a body of petrographic and chemical data on Kīlauea and Mauna Loa basalts.

Jaggar’s observations on the active lava lake in Halema‘uma‘u Crater remain unique, because the behavior observed from before 1912 until 1924 was only partly replicated in subsequent Halema‘uma‘u eruptions of 1952 (Macdonald, 1952, 1955) and 1967–68 (Fiske and Kinoshita, 1969; Kinoshita and others, 1969). Jaggar (1917a, 1917b) described the filling, circulation patterns, and draining of the lava lake in Halema‘uma‘u and established its depth at that time as 14 m. He also obtained a temperature profile through the active lava lake (Jaggar, 1917a). Lastly, he offered a thorough description and evaluation of the 1924 phreatic eruption and noted the occurrence of incandescent intrusions in the walls of Halemaʻumaʻu, which had been truncated during its widening (Jaggar and Finch, 1924).

The gradual refilling of Halemaʻumaʻu after its 1924 collapse was documented by Jaggar in detail in *The Volcano Letter* (Fiske and others, 1987). Petrologic sampling of this activity was restricted to the collection of small spatter samples; only one sample, taken in 1931 (as reported in *The Volcano Letter,* no. 288; Jaggar, 1932), was big enough to analyze by the methods then available. During this period, eruptions at Mauna Loa were more spectacular, with significant activity in 1940 and 1942 (Macdonald and Abbott, 1970) and especially in 1949 and 1950 (Macdonald and Orr, 1950; Finch and Macdonald, 1953), but sampling for analysis was minimal, as can be seen in the database used by Wright (1971).

Petrologic research during this period expanded the base of samples analyzed petrographically and chemically. Lava classification at the time depended on optical characterization of groundmass minerals, especially pyroxenes, so samples that were holocrystalline and as coarse-grained as possible were desirable. Macdonald (1949a), using these techniques, correctly identified the shield-building lavas of Hawaiian volcanoes as tholeiitic in nature. His work on the 1840 eruption of Kīlauea (Macdonald, 1944), in which upper rift olivine-poor flows and lower rift picritic lavas erupted almost simultaneously, was the first to suggest that olivine settling in the conduits of Hawaiian volcanoes might explain such differences in composition; subsequent analysis of the olivine-poor material (Macdonald, 1949b), together with analyses of augite separates from the 1921 Kīlauea lava, suggested that the uprift lava might have lost some pyroxene and plagioclase, as well (Muir and Tilley, 1957). Macdonald and coworkers also provided descriptions of the 1952 Halemaʻumaʻu activity (Macdonald, 1952, 1955) and the brief 1954 summit eruption (Macdonald, 1954; Macdonald and Eaton, 1957). The final eruption at Kīlauea during this period occurred in 1955, the first along the lower East Rift Zone since the 1840 eruption. This eruption was extensively investigated, with 10 samples analyzed for major elements and 8 for selected trace elements, the latter data the first such for Kīlauea (Macdonald and Eaton, 1964). The 1955 lavas were unlike the 1840 lavas, being much more differentiated than those of any previously observed eruption at Kīlauea.

Powers's (1955) paper on the composition and origin of Hawaiian basalts offers the most advanced discussion of Hawaiian petrology of the period. That report introduced the use of magnesia variation diagrams (with MgO replacing SiO_2 on the abscissa) as a superior means of visualizing relationships among lavas where addition or removal of olivine was the dominant process. Their use enabled Powers to recognize that Hawaiian olivine basalts differed from one volcano to the next and, specifically, that the historical lavas of Kīlauea and Mauna Loa were distinct. He introduced the concept of "magma batches" to describe these differences, and the term "olivine-control lines" to describe variations within a magma batch produced by variation in olivine only. Additional analytical and petrographic data on Kīlauea and Mauna Loa lavas presented by Tilley (1960) and Tilley and Scoon (1961) provided further evidence that the historical lavas at these two volcanoes are chemically and petrographically distinct.

Subaerial Activity at Kīlauea and Mauna Loa

Petrologic investigations expanded rapidly as the 1950s drew to a close, stimulated by the spectacular 1959 summit (Kīlauea Iki) and 1960 Puna (Kapoho) eruptions. Developments that have greatly enhanced the subsequent evolution of petrology include the following:

(1) Collection of more samples, especially in real time, through the course of an eruption. This was begun during the 1955 eruption, but such collecting was greatly expanded during the 1959 and 1960 eruptions and has been maintained in all subsequent activity.

(2) Submission of more samples for whole-rock analysis, beginning with the 1959 and 1960 eruptions (Murata and Richter, 1966; Richter and Moore, 1966). Gaps in earlier sampling and analysis were filled in wherever possible, including sampling of young prehistoric flows (Wright, 1971).

(3) Development of microbeam analytical techniques, beginning with the electron microprobe, which have made it possible to obtain chemical analyses of glass and individual mineral phases in volcanic samples. The first significant contributions involving microprobe data were Moore and Evans (1967) and Evans and Moore (1968), on mineral compositions from a suite of samples from the prehistoric Makaopuhi lava lake. Other early microprobe studies looked at silicate and Fe-Ti oxide phenocrysts (Anderson and Wright, 1972) and sulfides (Desborough and others, 1968) in various Kīlauea lavas.

(4) Use of experimental petrology to determine the phase relations of natural basalt samples (for example, Thompson and Tilley, 1969), which supports field observations on the crystallization sequence of lavas. Melting experiments, coupled with microprobe analysis of the resulting phases, led to the use of glass composition (the MgO content of olivine-saturated melts) as an indicator of quenching temperature at both Kīlauea and Mauna Loa. Originally developed to allow temperature estimates for glassy core from Kīlauea Iki lava lake (Helz and Thornber, 1987), this technique has subsequently been extended to other lavas at Kīlauea (Helz and others, 1995) and Mauna Loa (Montierth and others, 1995).

These expanded observations and quantitative analyses of rocks, glasses, and mineral phases have produced advances and further refinement of models for magma fractionation, mixing, transport to eruption sites, and storage.

Melt and Rock Compositions at Kīlauea and Mauna Loa

Before reviewing the range of melt and rock compositions observed for Kīlauea and Mauna Loa, let us look at the overall geography and structure of these two volcanoes (fig. 2). Each volcano has a summit caldera and two prominent rift zones; these features are the focus of almost all eruptive activity. Magma rises from the mantle beneath the summit, where it may enter the summit magma reservoir or move out into either of the two rift zones. In discussing the compositional data, we review Kīlauea summit compositions first; discussion of rift lavas through the 1970s follows, in chronological sequence, followed by a summary of Mauna Loa compositions. The ongoing Puʻu ʻŌʻō-Kupaianaha eruption on Kīlauea's East Rift Zone (January 3, 1983–present) is discussed last, in a separate section below.

Summit Melts at Kīlauea

Figure 3 presents data for selected major elements in melts erupted at the summit of Kīlauea from 1912 to 2010, a period of almost 100 years. The data for the period 1912–34 are newly presented microprobe analyses (appendix table 1) of small spatter samples collected by Jaggar. The results show that melts erupted effusively within Halemaʻumaʻu Crater and on the floor of Kīlauea Caldera had consistently low MgO contents of 6.4–7.6 weight percent over the entire period. MgO content may have decreased slightly with time from 1912 to 1924, but then increased during and after the refilling of Halemaʻumaʻu, following the 1924 phreatic eruption, by approximately 0.9 weight percent MgO over the 86-year period, as shown by the least-squares-fit lines in figure 3. For most other major oxides (for example, CaO and TiO_2, lines shown in fig. 3), the concentrations decrease as MgO

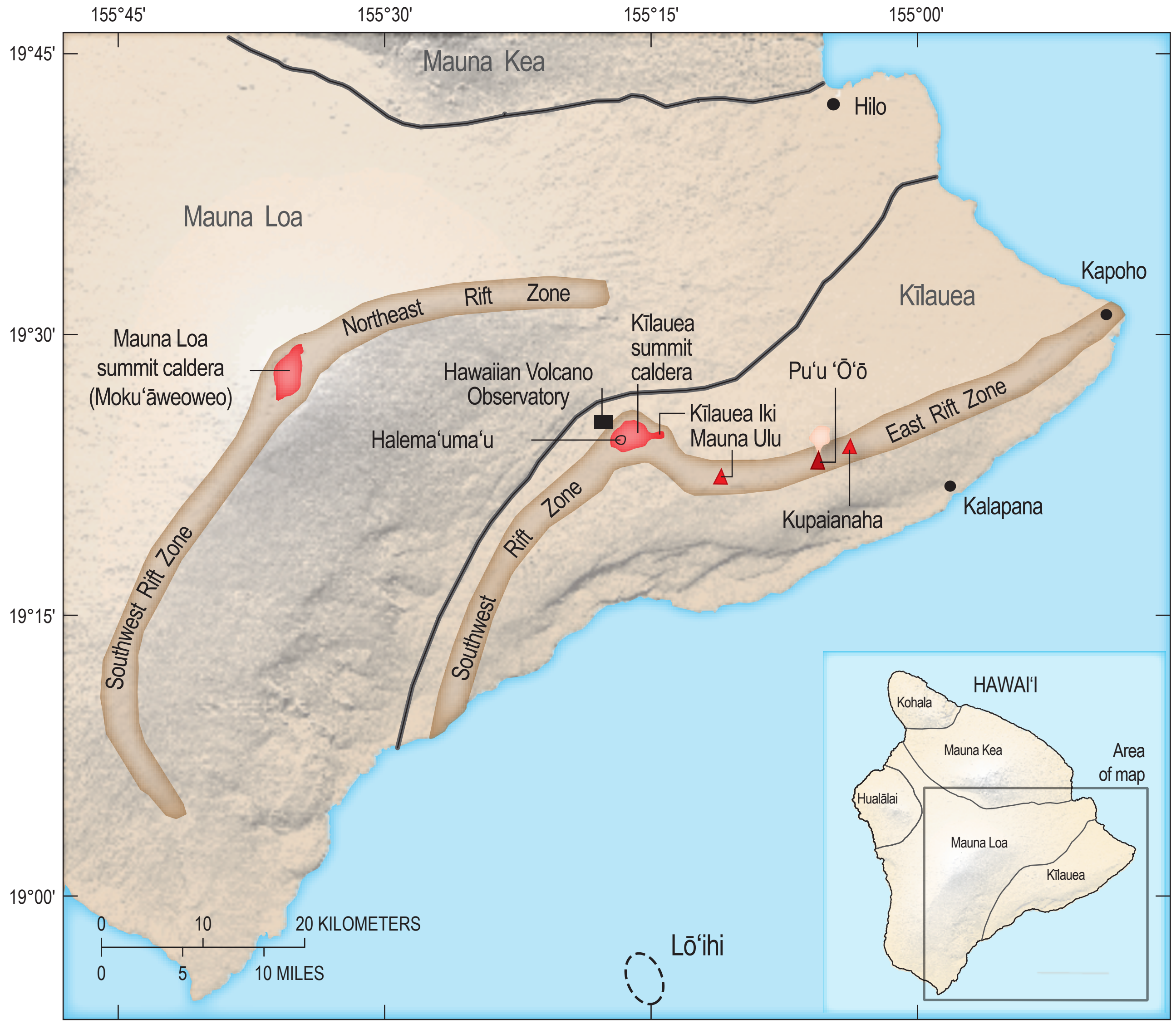

Figure 2. Map showing the summit calderas and rift zones of Kīlauea and Mauna Loa Volcanoes and the sites of the Mauna Ulu and Puʻu ʻŌʻō eruptions on Kīlauea's East Rift Zone. The location of the submarine volcano Lōʻihi is also shown. Towns shown by black dots. After Babb and others (2011).

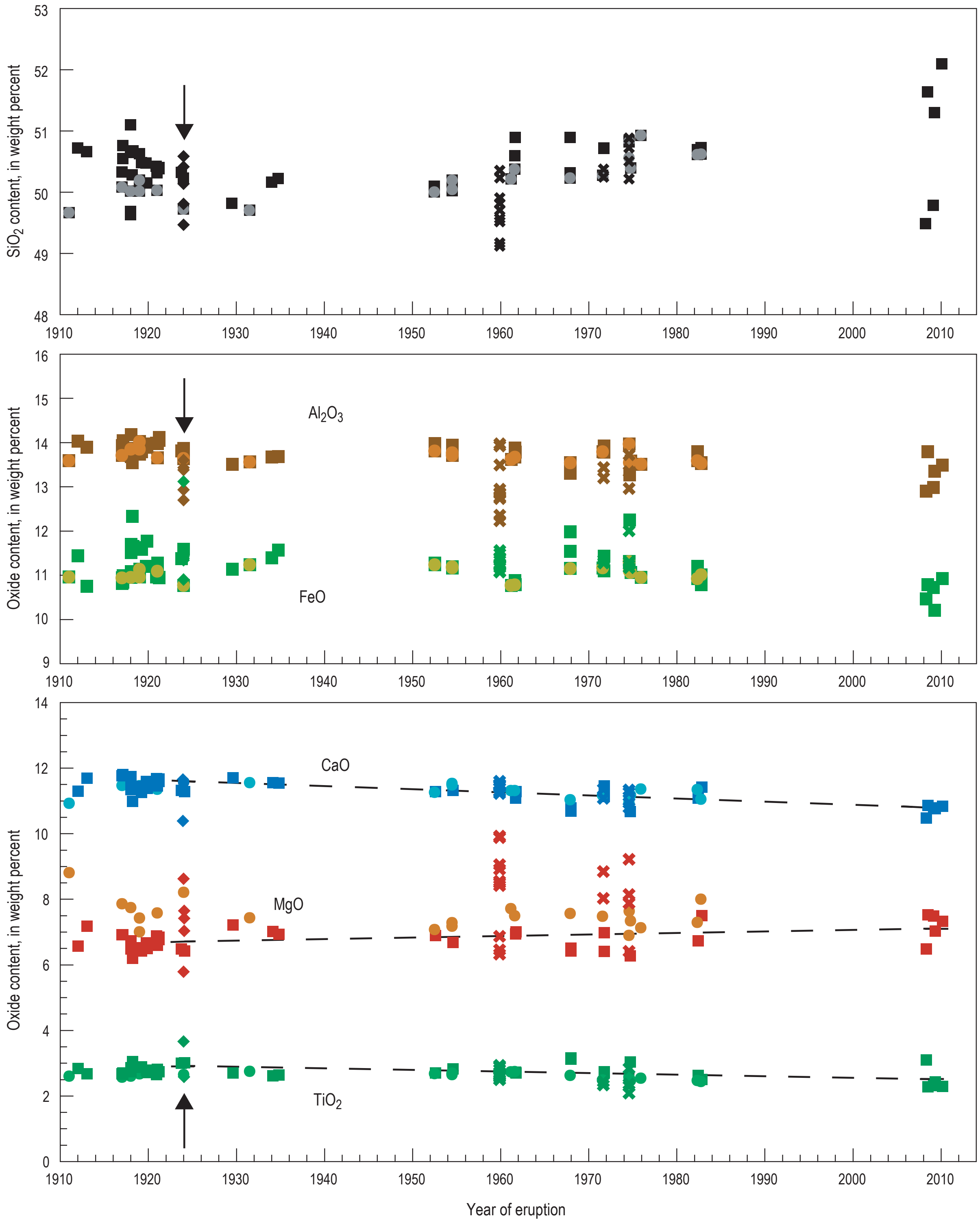

Figure 3. Time plots showing major oxide composition (color-coded by oxide) of intracaldera glasses erupted at the summit of Kīlauea from 1912 to 2010 (rectangles for scoria and spatter, diamonds for 1924 Pele's tears), together with extracaldera glasses from the 1959, 1971, and 1974 eruptions (crosses) and selected whole-rock analyses from 1911 to 1982 (circles). Glass analyses from 1912–34 are in appendix tables 1 (M. Beeson, analyst) and 2 (T. Rose, analyst); most 1952–82 intracaldera glass data are from Helz and others (1995) and Wright and Helz (1996). Analyses showing the range of MgO in the 1959 glasses are from Helz (2009), and those for 1971 and 1974 are in appendix table 3 (R.T. Helz, analyst). Data showing the range in the current (since 2008) summit activity are from Wooten and others (2009) and C.R. Thornber (unpublished data). Whole-rock data for lavas erupted from Halema‘uma‘u or onto the floor of Kīlauea Caldera are from Wright (1971), Basaltic Volcanism Study Project (1981), Duffield and others (1982), Baker (1987), and Lockwood and others (1999). Arrows mark the time of the 1924 phreatic eruption and draining of Halema‘uma‘u. Trend lines are discussed in the text.

increases. Trends for the alkalies and P_2O_5 parallel the variation seen in TiO_2.

By contrast, some extracaldera summit glasses are more magnesian (fig. 3). Glasses from the 1959 eruption, the vents of which lay just outside the caldera near Kīlauea Iki (fig. 2), range from 6.4 to 10.0 weight percent MgO (Helz, 1987a, 2009) and vary widely in other oxides, as well. New data on glasses from the 1971 and 1974 summit eruptions (fig. 3, appendix table 3) show that some extracaldera summit glasses are more magnesian than the intracaldera glasses, even along the same eruptive fissure.

Glasses with MgO contents greater than the intracaldera summit glasses (MgO=6.4–7.6 weight percent) also occur in tephra layers at Kīlauea. The 1924 tephra includes glassy Pele's tears with as much as 8.7 weight percent MgO (appendix table 2). Glasses in older tephra range to even higher MgO contents. The range and relative frequency of glass MgO contents in the 1500–1800 C.E. Keanakāko‘i Tephra Member of the Puna Basalt (Mastin and others, 2004; Swanson and others, 2012; Helz and others, 2014) and of five sections of the 400–1000 C.E. Kulanaokuaiki Tephra (Fiske and others, 2009; Rose and others, 2011; Helz and others, 2014) are shown in figure 4, with the 1959 glasses for comparison. Figure 4 also includes data from the uppermost thick tephra in the Hilina Pali section, referred to by Easton (1987) as the Pāhala Ash (Clague and others, 1995a; Helz and others, 2014), with an estimated age of 25–10 ka. The range and distribution of glass compositions for the Keanakāko‘i section shown are similar to the results of Garcia and others (2011) on another section of the deposit and are broadly similar to the data from the 1959 eruption. The 1959 results are also consistent with those of Harris and Anderson (1983) and Anderson and Brown (1993), who found MgO contents as high as 10.5 weight percent for glass inclusions in olivines from the 1959 eruption. Glasses in the two younger units lie mostly between 7 and 11 weight percent MgO. The Kulanaokuaiki and Pāhala histograms include a greater number of relatively differentiated (6.0–7.0 weight percent MgO) glasses than do the 1959 and Keanakāko‘i plots,

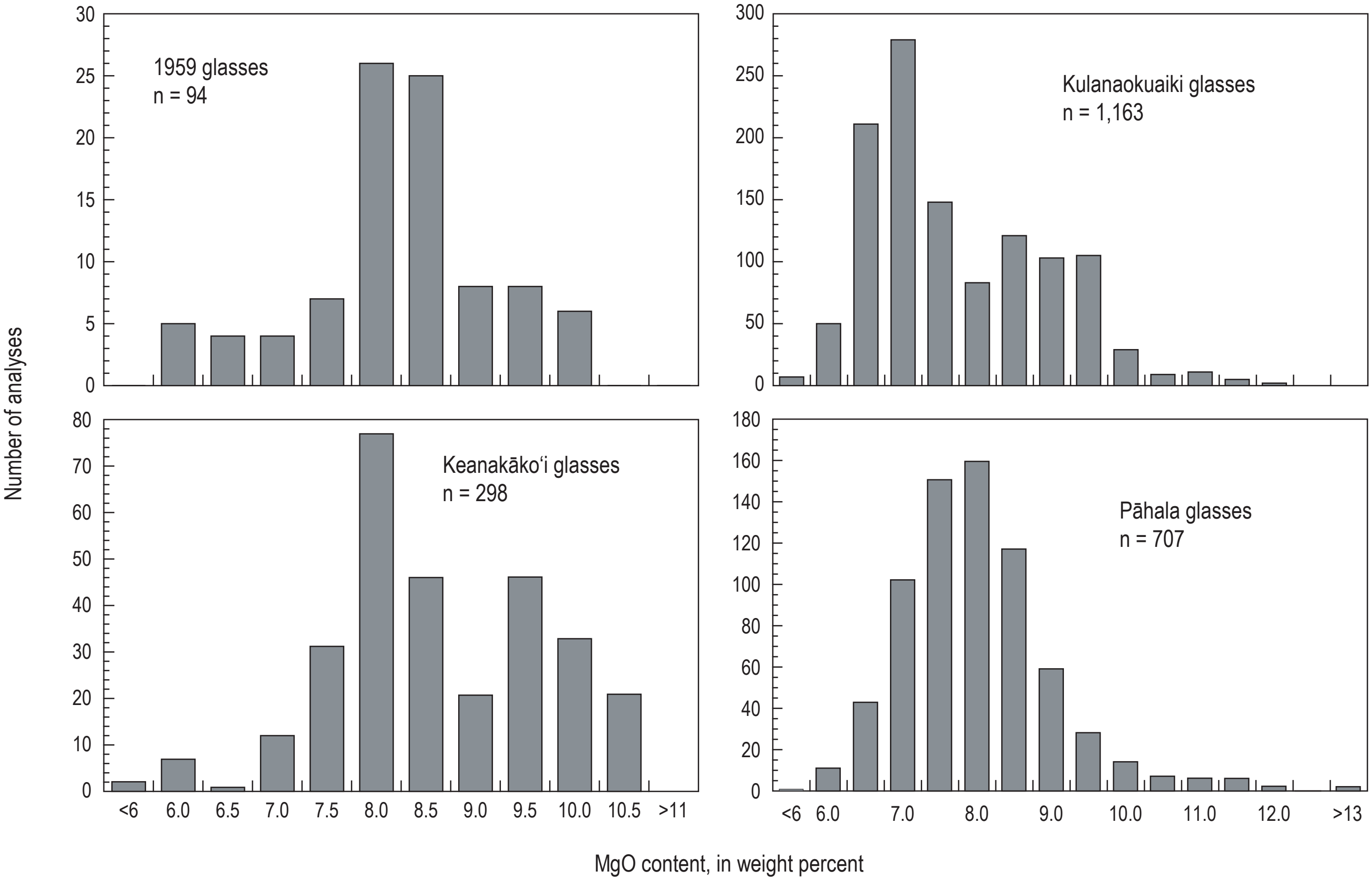

Figure 4. Histograms showing frequency distributions of MgO content in glasses from the 1959 Kīlauea Iki scoria (Helz, 2009; and unpublished data) and various other tephra units. The Keanakāko‘i data are from the Sand Wash section (Mastin and others, 2004; Helz and others, 2014), which does not include the upper units shown in McPhie and others (1990). The Kulanaokuaiki data are from multiple sections (Rose and others, 2011; Helz and others, 2014), and the Pāhala data are from Helz and others (2014). The Kulanaokuaiki and Pāhala tephra contain glasses with MgO>11.0 weight percent, unlike the younger tephras.

but in these older tephra, the distribution has a tail that extends beyond 12 weight percent MgO, higher than observed in the Keanakākoʻi and 1959 distributions.

Summit Lava Compositions Versus Melt Compositions

Figure 3 also includes whole-rock analyses for intracaldera lavas from 1911 through 1982, including data for one eruption (1975) for which no glass data are available. The whole-rock data run about a percent higher in MgO (7.0–8.8 weight percent) than the intracaldera glasses, and the data from 1932 through 1982 show the same slight upward trend in MgO with time as the glasses. Other oxide components vary less, suggesting that the bulk samples differ from the glasses mainly in containing varying amounts of olivine crystals. Lavas erupted at the summit but immediately outside the caldera have much higher bulk MgO contents, correlating with significantly higher olivine phenocryst contents. This pattern can be seen in the 1959 Kīlauea Iki eruption (Murata and Richter, 1966) and also in the 1971 and 1974 summit eruptions, where lavas immediately outside the caldera have MgO contents of 10.2–11.5 weight percent (Duffield and others, 1982; Lockwood and others, 1999). In the 1959 eruption, higher bulk MgO contents correlate strongly with more Mg-rich glasses; new data on samples of 1971 and 1974 scoria (appendix table 3) confirm that this is true for those eruptions, as well.

The data in figure 3 demonstrate that melts differentiated to the point of multiple saturation (where olivine would coprecipitate with augite and perhaps plagioclase) are present in the upper part of Kīlauea's summit magma reservoir over long time intervals. Similar stability in the summit reservoir's melt composition was inferred for the ʻAilāʻau lavas (Clague and others, 1999) on the basis of glasses found as rinds on the ʻAilāʻau flows, which were erupted over a period of at least 50 years, ending in about 1470 C.E. However, in the absence of tephra or scoria samples for ʻAilāʻau, this inference is not entirely conclusive, for two reasons. First, glasses from near-vent scoria or tephra can be more magnesian (and hence quenched from somewhat higher temperatures) than selvages from lava flows, as was seen in samples from the 1984 Mauna Loa eruption (Helz and others, 1995) or during the high-fountaining episodes at Puʻu ʻŌʻō, where flow selvages never preserved the more magnesian melts seen in tephra from fountains (as discussed in Helz and others, 2003). Second, some extracaldera summit glasses, such as those from the 1959, 1971, and 1974 eruptions, have relatively high MgO contents, indicating the availability of more magnesian liquids in the summit region. It is therefore possible that the MgO content of erupted melts for the ʻAilāʻau flows was higher; this seems particularly likely for the picritic[3] ʻAilāʻau lavas.

The occurrence of olivine-rich lavas near Kīlauea's summit is similar to that of the highly magnesian glasses. The 1959 lavas, erupted just outside the caldera, were picritic tholeiites (Murata and Richter, 1966), and the picritic flows within the (precaldera) ʻAilāʻau sequence are near the ʻAilāʻau vents, just to the east of Kīlauea Iki Crater (fig 2). Other historical picritic tholeiites have been erupted lower on the East Rift Zone, including those of the 1840 eruption (Macdonald, 1944) and the 1968 Hiʻiaka eruption (Nicholls and Stout, 1988).

Rift Zone Lavas

Kīlauea has also erupted differentiated lavas, that is, those with bulk compositions produced by multiphase fractionation, along its rift zones. The best studied are the 1955 lavas (Macdonald and Eaton, 1964; Wright and Fiske, 1971), which have bulk MgO contents of 5.04–5.68 weight percent in the early part of the eruption and 6.19–6.69 weight percent in the later lavas. The early 1955 lavas have very complex phenocryst assemblages, including three Fe-Ti oxide phases in addition to olivine, augite, plagioclase, and orthopyroxene (Anderson and Wright, 1972). The 1977 lavas, compositionally similar to the early 1955 lavas (MgO=5.28–5.84 weight percent) are nearly aphyric, containing minor plagioclase plus microphenocrysts of olivine+augite+plagioclase (Moore and others, 1980). The recent discovery of a subsurface body of dacitic magma in the Puna geothermal field (Teplow and others, 2009), similar to the Mauna Kuwale rhyodacite at Waiʻanae on Oʻahu (Bauer and others, 1973), raises the possibility that more differentiated lavas may be buried at Kīlauea.

Mauna Loa Melts and Lavas

The distributions of glass and whole-rock compositions at Mauna Loa are simpler than at Kīlauea. Summit and upper-flank lavas and glasses are fairly differentiated, with the 1975 and 1984 eruptions being representative (whole-rock data in Lockwood and others, 1987, and Rhodes, 1988; glasses from the 1984 eruption in appendix to Helz and others, 1995). More magnesian (and olivine-rich) flows and their more magnesian host glasses are found lower on the slopes, as reported by Riker and others (2009) for the 1859 eruption and by Rhodes (1995) for the picritic phases of the 1852 and 1868 lavas; these have glass or groundmass MgO contents ranging from 7.9 to 9.9 weight percent in whole-rock compositions, with as much as 22.2 weight percent MgO. C.R. Thornber and F.A. Trusdell (unpublished data) observe a similar but more modest trend in bulk compositions of the 1950 lavas, with MgO contents increasing from 6.8 weight percent at the highest-elevation vents to 10 weight percent at the lowermost vents. Lastly, Sparks's (1990) study of the precaldera lavas and intrusives exposed in the walls of Mokuʻāweoweo Caldera at Mauna Loa's summit identified a period of eruption/intrusion of MgO-rich magma (suite B, average Mg=10.9 weight percent)

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[3] Olivine-rich tholeiites with MgO≥15 weight percent are commonly referred to in the Hawaiian literature as picritic. These are, in all cases, tholeiitic, and the various authors consider at least some of the olivine phenocrysts to be cumulate.

in prehistoric time. Lavas with bulk compositions significantly more differentiated than 6.8 weight percent MgO, rare at Kīlauea, are unknown among Mauna Loa subaerial lavas, although one example (MgO content=5.50 weight percent; Wright, 1971) has been found in the shallow submarine section.

Olivine Control at Kīlauea and Mauna Loa

Olivine is the only phenocryst phase in subaerial lavas at Kīlauea, except for rare differentiated lavas, such as those erupted in 1955 (Macdonald and Eaton, 1964; Wright and Fiske, 1971; Anderson and Wright, 1972), and it is also the dominant phenocryst in Mauna Loa lavas (Wright, 1971). Studies of relatively rare olivine-rich lavas have provided much information on variations in the petrographic character and composition of phenocrystic olivine (Helz, 1987a; Wilkinson and Hensel, 1988; Schwindinger and Anderson, 1989; Rhodes, 1995). The most important petrologic consequence of this restricted phenocryst population, however, is that closely related lavas may differ in composition solely by addition or subtraction of olivine. This led to the use of magnesia variation diagrams to display chemical variation in Hawaiian lavas (Powers, 1955), a technique extended by Wright (1971, 1973), who found that addition or removal of olivine ($Fo_{86.0-87.5}$) was the dominant process of compositional fractionation at both Kīlauea and Mauna Loa for all but the most differentiated (bulk MgO<6.8 weight percent) lavas.

Olivine-controlled lava compositions from Mauna Loa are distinct from those at Kīlauea in major elements (Powers, 1955; Tilley, 1960; Wright, 1971; Rhodes, 1988, 1995; Rhodes and Hart, 1995) and in trace elements (Tilling and others, 1987a, 1987b; Rhodes and Hart, 1995); however, magma batches definable by major-element chemistry are lacking (Wright, 1971). By contrast, cross-trend variations within olivine-controlled Kīlauea lavas exist and are coherent in time and space (Wright, 1971), and the distinctive lava groups are referred to as "magma batches." Recognition and tracking of these magma batches at Kīlauea have provided critical insights into the processes of magma mixing and preeruptive shallow transport in that system (Wright and Fiske, 1971).

Magma Batches, Magma Mixing, and Lateral Magma Transport at Kīlauea

Beginning with the 1952 eruption, scientists routinely sought to obtain summit lava samples large enough for bulk analysis. This allowed recognition of the diverse nature of magmas fed into Kīlauea from 1952 to the present. Information on the succession of magma batches erupted at the summit during this time is summarized in table 1.

New magma batches appeared at Kīlauea's summit every 1–7 years from 1952 through 1982. The spacing between these batches is quite short, probably less than the time they existed in the combined summit-rift system, which for several batches is about 10 years (table 1). This relatively rapid succession has implications for the size of the batches and the size and continuity of Kīlauea's summit reservoir. Estimation of storage times is difficult because of the paucity of phenocrysts in most summit lavas. Mangan (1990), however, using crystal-size distribution results for olivines in the 1959 lavas, estimated storage times of ~10 years, somewhat longer than the spacing between batches. This suggests that either storage times are longer than the interval between batches or some olivine is retained in the summit reservoir while successive melts pass through.

Examining the relationship between summit batches and rift eruptions, three different periods can be inferred from table 1, as follows:

(1) From 1952 through 1968, summit eruptions dominated, and the magma batches were defined by a series of distinct summit lava compositions (Wright and Fiske, 1971). The 1952 and 1959E magma batches appeared as components of rift lavas only after they had erupted at the summit, while the 1961 and 1967–68 batches first appeared in the 1960 Puna eruption on the East Rift Zone (near Kapoho; fig. 2) before they were erupted at the summit.

(2) From 1971 through 1975, activity at Kīlauea was dominated by eruptions at Mauna Ulu (fig. 2), though moderately extensive summit eruptions (1971, 1974) also occurred. The 10 magma batches defined for Mauna Ulu (Wright and others, 1975; Wright and Tilling, 1980) differed more subtly from each other than had the magma batches of the earlier period. Summit lavas for the 1971, 1974, and 1975 eruptions were defined in terms of Mauna Ulu variants, as shown in table 1. This is the reverse of the earlier process, in which summit batches were used as the starting point for unraveling magma mixing in rift zone lavas; however, the timing of their eruption (they appeared at the summit after they had appeared at Mauna Ulu) was the same as for all summit batches since 1961. The November 1975 Kalapana earthquake (*M*7.7) appears to have brought this behavior to an end (Holcomb, 1987).

After a hiatus of two years, eruptive activity at Kīlauea resumed with the 1977 eruption (Moore and others, 1980), followed by minor activity in 1979 and 1980, all on the East Rift Zone (Holcomb, 1987).

(3) From 1982 to the present, activity at Kīlauea has been dominated by the ongoing Puʻu ʻŌʻō-Kupaianaha East Rift Zone eruption (fig. 2), reviewed in the next major section. Summit activity has been minimal, consisting, until 2008, of two 1-day eruptions in 1982 (Banks and others, 1982; Helz and others, 1995), so summit magma batches have been essentially undefined. The progressive change in magma partitioning between the summit and East Rift Zone seen in table 1 suggests that that Kīlauea's East Rift Zone has become steadily more accessible to new magma since 1952.

Period 1: Magma Mixing and Transport at Kīlauea, 1952–68

Magma mixing was proposed early on to be a major process at Kīlauea (Wright and Fiske, 1971; Wright, 1973), a

Table 1. Timing of appearance of magma batches at the summit of Kīlauea Volcano from 1952 to the present (2014).

Time of appearance of lava at the summit	Time after previous summit batch	Eruption(s) with summit batch as mixing component	Time between appearances	Comments, references
1952	?	Late 1955 Puna 1960 Puna	3 years after summit 8 years after summit	Wright and Fiske (1971), Helz and Wright (1992) Wright and Fiske (1971), Wright and Helz (1996)
1954	2 years	1959 (also a summit eruption)	5 years after summit	Equivalent to 1959W component as in Wright (1973), Helz (1987a)
1959E	5 years	1960 Puna (latest samples)	3 months after summit	Wright and Fiske (1971), Wright and Helz (1996)
1961	1.5–2 years	1960 Puna	1 year before summit	Wright and Fiske (1971), Wright and Helz (1996)
1967–68	6 years	1960 Puna August 1968, October 1968, February 1969 middle east rift	7–8 years before summit 1 year after summit	Wright and Fiske (1971), Wright and Helz (1996) Wright and others (1975)
August 1971	3–4 years	1969–71 Mauna Ulu	1–2 years before summit	Made up of Mauna Ulu 3 & 5 (results in Wright and Tilling, 1980); analyses in Duffield and others (1982)
July 1974 September 1974	3 years 3 years	1972–74 Mauna Ulu 1972–74 Mauna Ulu	2 years before summit 1–2 years before summit	Equivalent to Mauna Ulu 6 (Wright and Tilling, 1980) Made up of Mauna Ulu 6 & 10 (Wright and Tilling, 1980); analyses in Lockwood and others (1999)
November 1975	1 year	1972–74 Mauna Ulu	2 years before summit	Analysis from Basaltic Volcanism Study Project (1981); Mauna Ulu variant 10 (Wright and Tilling, 1980)
April 1982 September 1982	7 years	Pu‘u ‘Ō‘ō episodes 17–20	2 years after summit	Analyses in Baker (1987); comparison with Pu‘u ‘Ō‘ō data shown in Garcia and Wolfe (1988)
2007–present	25 years	Equivalent to current Pu‘u ‘Ō‘ō lava	Concurrent	Rowe and others (2009)

recognition facilitated by the diverse compositions of the summit batches. These early studies were based largely on whole-rock major-element chemistry and mixing calculations and so were contested as not proven for the late 1955 lavas (Ho and Garcia, 1988) or for any of the 1954–60 lavas (Russell and Stanley, 1990). However, further work—which included petrographic and microprobe analysis, as well as modeling of trace element compositions—has demonstrated conclusively that the late 1955 and middle to late 1960 lavas were mixed magmas (Helz and Wright, 1992; Wright and Helz, 1996).

One outcome of reexamining the late 1955 lavas was recognition that mixing of thermally disparate magmas produces resorption of olivine, even though both magmas are saturated with olivine (Helz and Wright, 1992). Thermal equilibration is achieved much faster than chemical equilibrium, so crystals from the cooler mixing component are resorbed during initial mixing. The widespread occurrence of resorbed olivine phenocrysts in Hawaiian lava flows, noted earlier by Macdonald (1949a) and Powers (1955), and examples of which can be found even in the picritic 1959 eruption (Helz, 1987b), is rarely due to reaction with melt to form pyroxene, but rather is evidence that magma mixing is widespread at Kīlauea and other Hawaiian volcanoes.

Reexamination of the lavas of the 1955 and 1960 eruptions also provides insight into the length of time that petrographic evidence for mixing may be expected to survive. The late 1955 lava flows had mixed phenocryst assemblages, with resorbed and (or) reversely zoned crystals, consistent with very recent magma mixing (Helz and Wright, 1992). The early 1960 lava flows, although identical in composition (Wright and Fiske, 1971), show no such evidence for mixing, containing only a modest amount of microphenocrystic olivine+augite+plagioclase (Wright and Helz, 1996). The 5 years that elapsed between the two eruptions appears to have been sufficient for chemical and thermal equilibrium to be achieved in magmas reinjected by hotter material.

By contrast, the early 1955 lavas (with their highly complex phenocryst assemblages) may have been isolated for a long period. A study by Cooper and others (2001) using ^{230}Th-^{226}Ra dating suggests that the plagioclase phenocrysts in those lavas had been in storage for at least 550 years before eruption. This is much longer than has been inferred for other East Rift Zone magmas, which appear to have crystallized at rates of 1–2 weight percent per year (Wright and Tilling, 1980; Wright and Helz, 1996) and so may not reflect the age of the rest of the lava.

Samples from late 1960 lavas show renewed petrographic and chemical evidence for magma mixing, interpreted as coeval with the eruption (Wright and Helz, 1996), with both major and trace-element data requiring the addition of a succession of four summit batches. The pattern of appearance of the summit batches in the 1955 and 1960 lavas elucidates transport processes within Kīlauea during this time. The 1952, 1961, and 1967–68 summit components followed one after the other in time (see fig. 16 in Wright and Helz, 1996), as if these three batches were transported to the 1960 mixing chamber along the same conduit. The 1959E component, which makes up as much as 27 percent of the latest 1960 lavas, is out of sequence with the 1961 and 1967–68 batches. This very hot, highly phyric component may have followed its own path to Puna (fig. 16 in Wright and Helz, 1996), just as it followed an aberrant path to the surface at the summit (Eaton and Murata, 1960; Eaton and others, 1987; Helz, 1987b). The presence of clusters of mildly deformed olivine crystals (Wright and Helz, 1996) in those 1960 samples that required the 1959E as a mixing component supports the identification of this component, and its relatively deep path out to Puna, because such olivines are unusual in subaerial Kīlauea lavas.

Some late 1960 scoria contain phenocrystic phases (orthopyroxene±ilmenite) like those seen in the early 1955 lavas (Anderson and Wright, 1972), suggesting that there were small pockets of magma similar to the early 1955 compositions that had survived until at least 1960. Such isolated pockets of magma are plausible parents (Helz, 2008) for the dacitic magma discovered in a geothermal drilling project in Puna (Teplow and others, 2009).

Period 2: Magma Batches, Mixing, and Transport at Kīlauea, 1969–75

During the period from 1969 through 1975, new magma batches, defined in terms of Mauna Ulu variants, surfaced along the East Rift Zone before erupting at the summit and the upper Southwest Rift Zone (table 1). The only earlier stored batch recognized was the 1967–68 magma, which contributed to a series of eruptions in the middle East Rift Zone one year after the Halema‘uma‘u activity ceased (Wright and others, 1975). The tiny November 1975 eruption, which occurred during, and perhaps as a consequence of, the Kalapana earthquake (Tilling and others, 1976), most closely resembles Mauna Ulu variant 10 (Wright and Tilling, 1980).

In general, the chemical and petrographic contrasts between Mauna Ulu variants were subtler than between the summit batches of 1952 through 1967–68; for example, variations in trace element contents showed almost continuous, monotonic decreases (Hofmann and others, 1984). The Mauna Ulu lavas contain as much as 19 weight percent MgO (Wright and others, 1975; Wright and Tilling, 1980), consistent with high olivine phenocryst content. In a recent paper, Vinet and Higgins (2010) document the occurrence of highly magnesian and (or) deformed olivine phenocrysts in the Mauna Ulu lavas and suggest that they were entrained from the décollement at the base of Kīlauea's volcanic edifice (shown in their figure 20 as below 11 km depth). However, olivine control for the Mauna Ulu lavas shows that the average olivine composition remains Fo_{86-87}, the same as is found in Kīlauea's summit lavas, so it is not clear that a source region as deep as the base of the décollement under the East Rift Zone is required.

The time span 1969–75 encompassed the two periods of Mauna Ulu activity (Swanson and others, 1979; Tilling and others, 1987c) and the 1971 and 1974 summit eruptions (Duffield and others, 1982; Lockwood and others, 1999). During this period, magma first moved up into the base of the summit reservoir, then out the East Rift Zone to Mauna Ulu, without erupting at the summit. Later, when the upper East Rift Zone became blocked, the summit inflated and the complex 1971 and 1974 summit eruptions occurred (Duffield and others, 1982; Lockwood and others, 1999). In both instances, the summit activity was followed by eruptive activity along the upper Southwest Rift Zone, which otherwise had seen no activity since the 1919–20 eruption (Holcomb, 1987). Most notably, this period saw a major change in volcano monitoring: magma movement, formerly inferred by using chemically defined magma batches (as described above), has since been tracked using newly enhanced geophysical monitoring techniques (see Okubo and others, this volume, chap. 2, and Poland and others, this volume, chap. 5).

Subsequent work at Kīlauea has been largely focused on the ongoing Pu‘u ‘Ō‘ō eruption (January 3, 1983, to the present), which is reviewed in the "Petrologic Overview of Kīlauea's Ongoing Eruption: 1983–2011" section below.

Magma Batches, Magma Mixing, and Magma Transport at Mauna Loa

Magma batches defined at the major-element level are lacking at Mauna Loa (Wright, 1971), and the lavas are unusually uniform within eruptions (Rhodes, 1988). Lavas with phenocrysts of olivine and plagioclase plus various pyroxenes are fairly common (Macdonald, 1949a,b; Rhodes and Hart, 1995), but lavas with complex, obviously mixed, phenocryst assemblages have not been reported. The inference from these observations is that the magma chamber under Mauna Loa is large enough to buffer melt compositions, and that a large part of that volume is crystallizing the three-phase assemblage olivine+pyroxene+plagioclase (Rhodes, 1988; Rhodes and Hart, 1995).

Compositional shifts have been observed in trace elements and isotopic ratios, however, with concentrations of incompatible trace elements decreasing with time (Tilling and others, 1987a,b; Rhodes and Hart, 1995). Early results raised the possibility that the shift to lower concentrations was caused by disruption of the volcano's plumbing by the 1868 earthquake (Tilling and others, 1987a); subsequent work (Rhodes and Hart, 1995) shows the decrease to have begun before 1868, and to have bottomed out in 1880. Levels rebounded somewhat by 1900 and have shown little variation through the 20th century.

In contrast to Kīlauea, post-18th century Mauna Loa eruptions have consistently begun at the summit. The eruptions have then either continued immediately as rift zone eruptions (along the Northeast or Southwest Rift Zone, or both; see fig. 2) or less commonly from vents on the north flank or resumed as a rift zone eruption following a hiatus (in which case a time delay of 1–2 years is typical), as summarized in Lockwood and Lipman (1987). This contrasts with post-1960 behavior at Kīlauea, where magma batches consistently appear first along the East Rift Zone and then at the summit, and only rarely and subsequently appear along the volcano's Southwest Rift Zone.

Petrologic Overview of Kīlauea's Ongoing Eruption, 1983–2011

The Kīlauea East Rift Zone has been erupting since 1983 and continues to do so at the time of this writing. This eruptive era was ushered in by summit eruptions in April and September 1982. Since March 2008, eruptive activity has expanded to include simultaneous summit and rift zone eruptions.

In January 1983, magma erupted through a downrift sequence of fissure vents extending 7.5 km from the west side of Nāpau Crater, past the eventual Puʻu ʻŌʻō vent, and nearly to the eventual Kupaianaha vent (fig. 5). The details of eruption onset and early activity, including the high-fountaining episodes, are presented in Wolfe and others (1987, 1988). More recently, Heliker and others (2003) provided summaries of HVO's monitoring-related research during the first 20 years of the Puʻu ʻŌʻō-Kupaianaha eruption. Included in that volume, and briefly reviewed here, is a chapter that interprets the petrology and geochemistry of an extensive suite of well-quenched lava samples collected from East Rift Zone vents between January 1983 and October 2001 (Thornber, 2003; data in Thornber and others, 2003a).

Yet another 10 years of petrologic data gathering, during 2001–11, have further enlightened our perspectives on the fundamental traits of near-continuous magma recharge and eruption associated with the shield-building stages of Hawaiian volcanism. The petrologic details of simultaneously erupted summit and rift zone lava provide opposing perspectives on magma complexities prior to rift eruption. Combined with geophysical and behavioral observations,

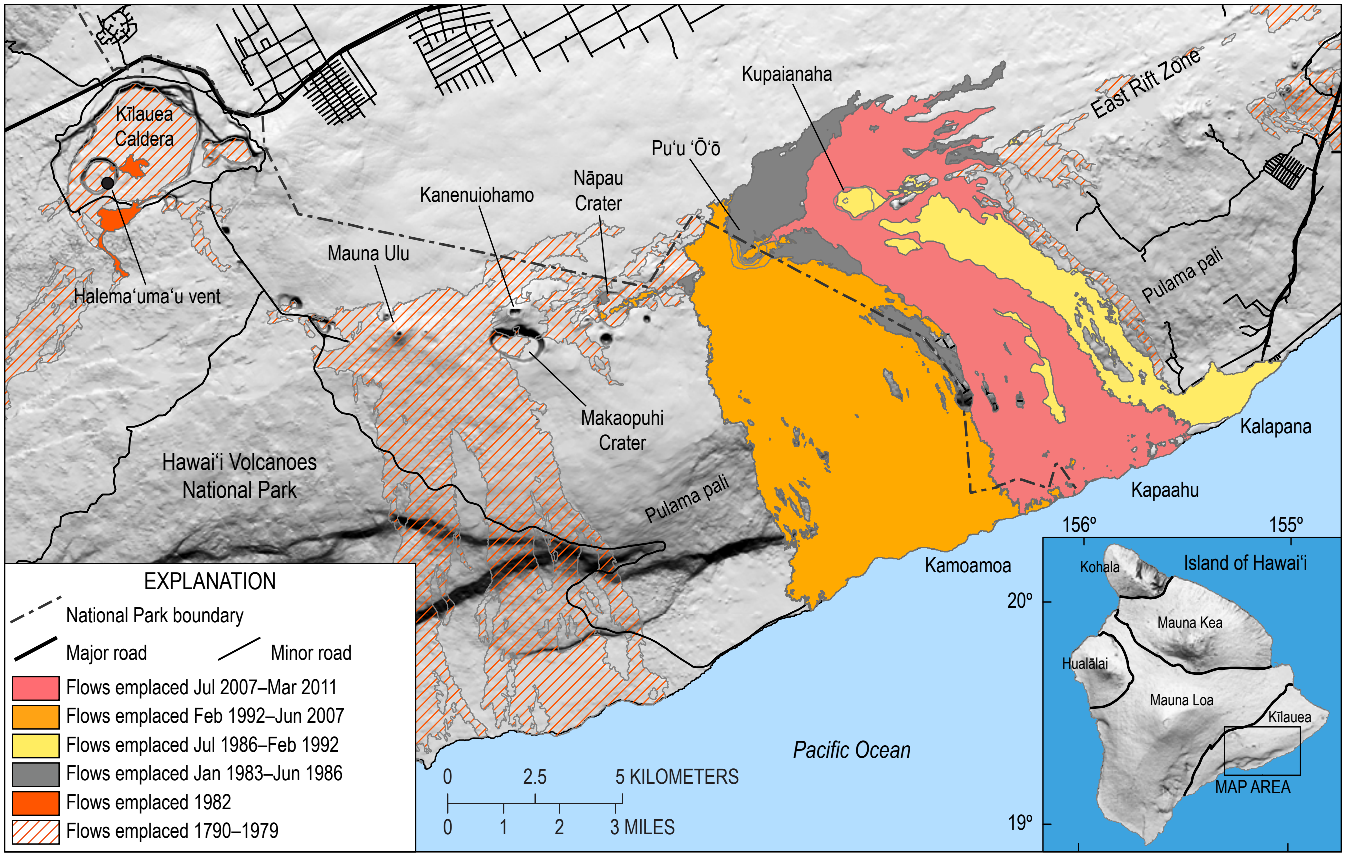


Figure 5. Map showing the extent of lava flows of the ongoing Kīlauea East Rift Zone eruption (from January 1983 through March 2011).

petrologic monitoring at both early and later stages of the rift zone eruption has led to better understanding of how and where magma is transported, stored, erupted, and recharged within Kīlauea's edifice.

An overview of changes in eruption petrology is best facilitated with reference to the temporal variation of an MgO index. The weight percent magnesia (MgO) in bulk lava and matrix glasses tracks fluctuations in the eruption temperature and the character of magma tapped throughout that eruption (fig. 6). The lowest MgO values were recorded during the early stages of the eruption and again during (and shortly after) the brief Nāpau Crater eruption in January 1997. In both cases, petrologic studies confirm magma mixing shortly before eruption between cooler rift-stored magma pockets and hotter olivine-phryic magma at shallow depths (Garcia and others, 1989, 1992; Thornber and others, 2003b). Otherwise, as detailed by Thornber (2003), sparsely olivine-phyric lava was erupted from well-established vents at Puʻu ʻŌʻō and Kupaianaha between 1985 and 2001. During those periods of prolonged and steady eruption, the consistent MgO thermometry difference of 30 ºC between bulk lava and matrix glasses is explained by equilibrium olivine growth with cooling during summit-to-rift transit (Thornber, 2001).

Thornber (2003) suggested that the persistent overall limits to the range of MgO cycles of steady-state eruption products during the period 1985–2001 reflect a geostatic balance between two distinct magmatic components that persist within the shallow volcanic edifice during prolonged intervals of recharge and eruption. One of them is a shallow recharge component with an upper MgO limit of 10 weight percent. The other end member, at 6.8 to 7.0 weight percent MgO, is the most common of Kīlauea eruption compositions. This is a low-pressure multitectic composition for olivine tholeiite magma at which clinopyroxene and plagioclase crystallize, together with olivine (Helz and Thornber, 1987), and is therefore a thermodynamically favored open-system reservoir component that serves as a magmatic buffer during recharge (Thornber, 2003). The observed limits of cyclic MgO variation of this eruption (fig. 6) are within the slightly larger range of historical intracaldera summit lavas and define a persistent temperature range of olivine-saturated end-member magmas, which is presumably regulated by recharge of the shallow magmatic plumbing system during prolonged shield-building eruptions.

Chemical Signature of Recently Erupted Kīlauea Magma

As with all eruptive sequences of Hawaiian olivine tholeiites, bulk-lava MgO variation diagrams for major and trace elements provide a first-order means of assessing

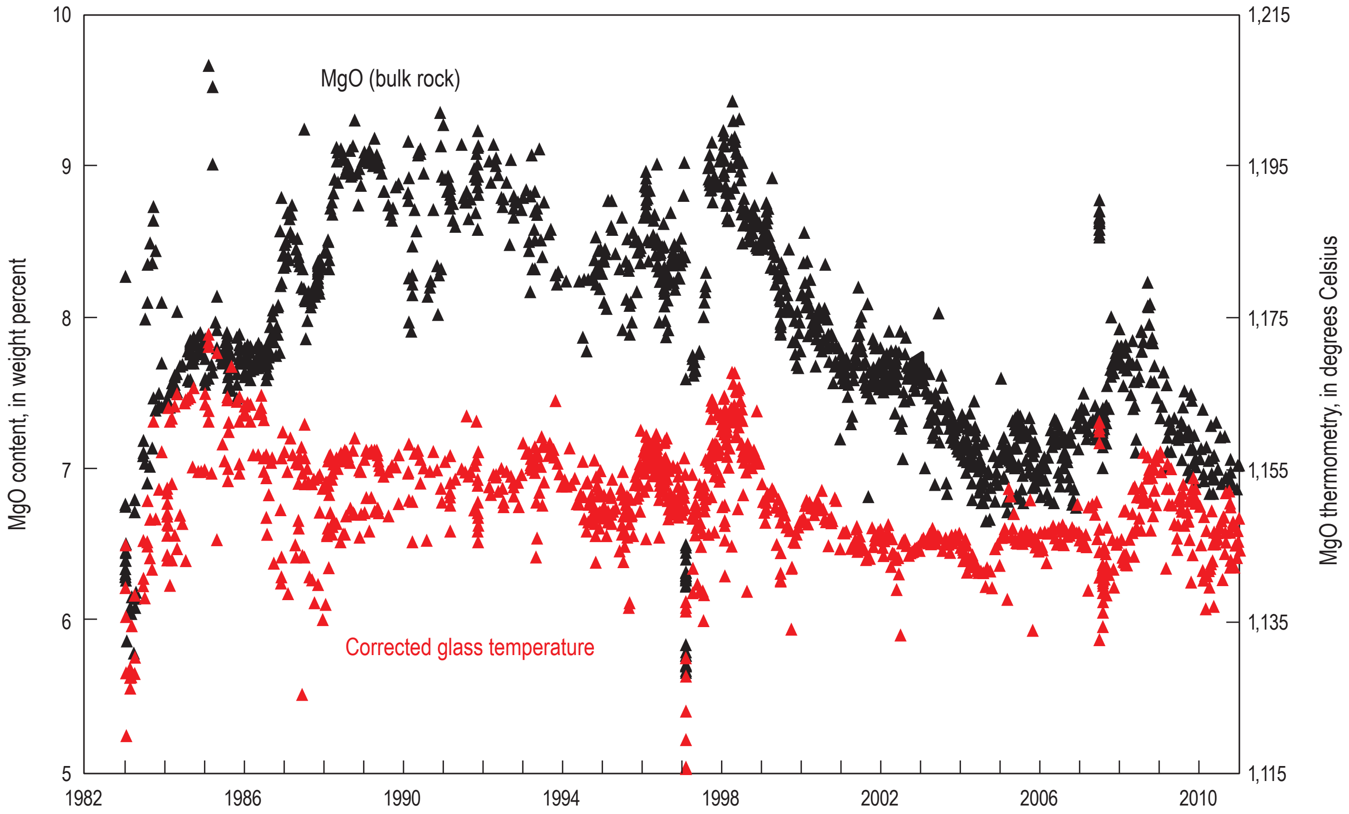


Figure 6. Time plot showing temporal variation from 1983 to 2010 of MgO weight percent (left axis) and corresponding MgO glass geothermometer values (Helz and Thornber, 1987) (right axis) for bulk lava samples (black triangles) and average matrix glasses (red triangles). The latter are vent-corrected if sampled away from the vent by adding 0.9 ºC or 0.05 weight percent MgO glass per kilometer of tube distance (Thornber, 2001).

magmatic conditions, such as low-pressure fractionation, mixing, or assimilation. Short-term cycles of temperature, MgO, and incompatible element concentrations are broadly consistent with repeated cycles of olivine fractionation from melts of ~10 to ~7 weight percent MgO. As exemplified by CaO variations (fig. 7), olivine incompatible elements increase with progressive olivine fractionation, delineating an olivine-control trend or an olivine-saturated liquid-line-of-descent (LLD). The LLD for steady-state eruption products has shifted progressively, from 1985 to 1995, toward lower concentrations of incompatible elements.

The trend of decreasing incompatible elements over time is apparent in ratios of highly to moderately incompatible elements, such as La/Yb, in olivine-normalized trace element patterns for successive intervals of steady-state eruption (Thornber, 2003) and is consistent with subtle variations of Sr, Nd, and Pb isotopes reported by Thornber and others (2003c), Garcia and others (1996, 2000), and Pietruszka and Garcia (1999a). This long-term decrease of incompatible-element ratios has been proposed by Garcia and others (1996, 2000) to represent changes in mantle-source conditions. In contrast, Thornber (2003) makes a case for a top-down rather than a bottom-up mechanism, attributing the change in chemical signature to progressive flushing of diminishing proportions of pre-1983 summit magma (maintained at multitectic conditions) by magma derived from a chemically uniform mantle source. Long-term summit deflation since 1983 is coincident with the long-term geochemical trends seen in subsequent products of continuous eruption, suggesting that progressive summit-reservoir depletion may be responsible. In this scenario, magma derived from a uniform mantle source has apparently flushed older resident magma from the shallow edifice and could now completely occupy the shallow magmatic plumbing system. This idea was further supported when new magma began erupting at the summit in March 2008.

Post-2001 Eruption Petrology: Unexpected Changes Associated with an Increase in Magma Supply

Between October 2000 and April 2001, the petrology of steady-state Puʻu ʻŌʻō-Kupaianaha lava underwent a fundamental and long-term change following the steady decline in bulk MgO from a 1998 high of 9.5 down to around 7.5 weight percent (fig. 6) and amidst the onset of the first prolonged summit inflation in more than 20 years (Poland and others, 2012). For the first time in more than 20 years, the steady-state rift zone eruption was no longer tapping an olivine-only shallow magma source. The preeruptive condition changed to a more complex comagmatic mixture of hotter olivine-only magma and a cooler magma with clinopyroxene (±olivine, ±plagioclase; Thornber and others, 2007). All post-2001 low-temperature lava is characterized by a disequilibrium low-pressure phenocryst assemblage. Olivine ($Fo_{80.5–81.5}$) coexists with phenocrysts of lower temperature clinopyroxene (±plagioclase, ±Fe-rich olivine). Mixing between hotter and cooler magmas is texturally documented by complex pyroxene zoning and resorption, along with olivine overgrowths of resorbed pyroxene and plagioclase grains.

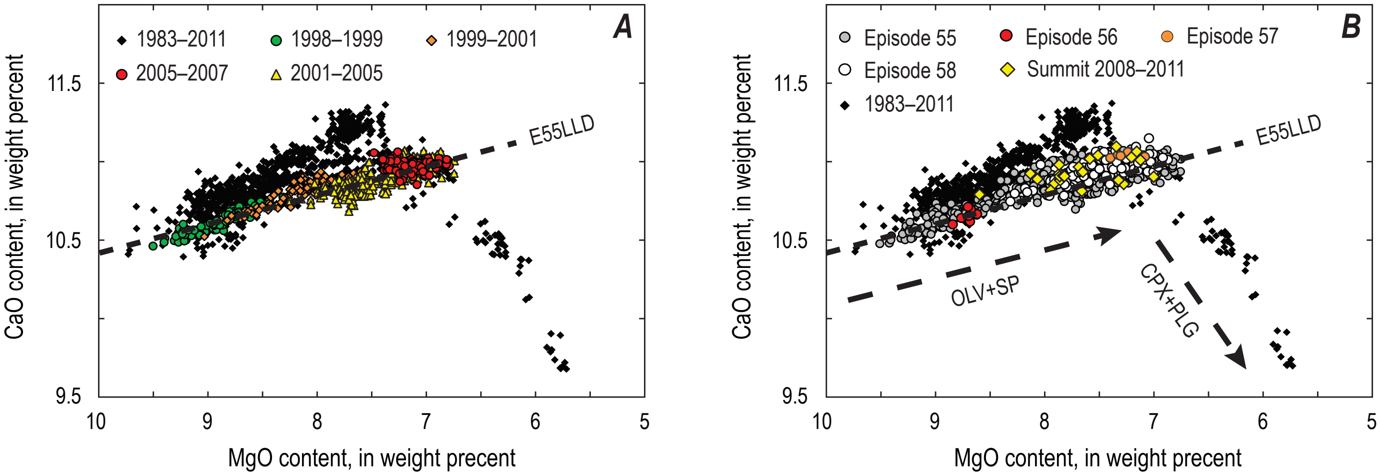


Figure 7. Graphs of CaO versus MgO contents in Kīlauea lava erupted from 1983 through 2011. *A*, Analyses from selected time periods from 1998 to 2005, highlighted in color to emphasize the episode 55 trend of decreasing MgO with time. *B*, Analyses from selected time periods 1997 to 2007, highlighted to contrast episode 56, episode 57, and episode 58 lava compositions, along with 2008–11 summit juvenile tephra compositions. E55LLD is episode 55 liquid-line-of-descent, as shown in figure 2 of Thornber (2003), which shows episode 48–55 changes and provides additional explanatory text. The arrows indicate the change in magma compositions resulting from low-pressure fractional crystallization of (1) olivine and spinel (OLV+SP) at high temperatures (from recharge magma with up to 10 weight percent MgO) and (2) clinopyroxene and plagioclase (CPX+PLG) at lower temperatures (from isolated magma with ~6.8 weight percent MgO). The fulcrum in this plot is the 5-phase multitectic point for shallow Kīlauean magma that occupies cooler zones within the actively recharged reservoir. During prolonged recharge and eruption, variations along the olivine-controlled liquid-line-of-descent (LLD) with time are affected by proportional mixing of these perpetually present components.

The prolonged decline in the MgO index bottomed out in 2004 at 6.8 weight percent. This value corresponds to the low-pressure multitectic composition for Kīlauea magma and is the lowest sustainable MgO value during periods of continuous recharge and eruption (Thornber, 2003). The bulk MgO of olivine-clinopyroxene-phyric lava from vents at Puʻu ʻŌʻō continued to hover around 7.0 to 7.3 weight percent until the June 19, 2007, eruption at Kanenuiohamo (episode 56, the Father's Day activity), a few kilometers uprift of Puʻu ʻŌʻō, of the hottest and most primitive olivine-only lava seen since 1998. This occurred at the culmination of the 4-year summit inflationary cycle and provided a glimpse of the relatively primitive component that is perpetually present during magmatic recharge of the shallow edifice. The relatively primitive magma erupted at Kanenuiohamo lies along the same LLD as all post-1995 lava (see episode 56 in fig. 7*B*) and is a petrologic match for the hotter, olivine-phyric end member of simple binary mixing between recharge magma and stagnant crystal-saturated magma.

When the East Rift Zone eruption revived downrift of Puʻu ʻŌʻō one month after the Kanenuiohamo eruption, the July 2007 fissures once again issued low-MgO, olivine-clinopyroxene-phyric lava. MgO of rift zone lava increased steadily to 8.0 weight percent 5 months after the onset of Kīlauea's summit eruption in March 2008. Increasing eruption temperature with time accompanied compositional migration back up the olivine-controlled LLD (fig. 7*B*) and, by August 2008, the preeruptive condition retransitioned to olivine-only for another 10.5 months. Meanwhile, bulk MgO began to decline steadily in September 2008 and breached 7.5 weight percent in mid-June 2009 as clinopyroxene phenocrysts began to reappear in rift lava. The MgO decline bottomed out in October 2009 near 7.1 weight percent, where it remained through December 2010.

The hybrid lava erupted on the East Rift Zone from 2001 to 2010 is a mixture of two end-member components within the shallow magmatic regime that feeds the eruption. The high-MgO and high-temperature magma is equivalent to that which was sparsely erupted at Kanenuiohamo in June 2007 (and which is currently being erupted at the summit vent). The cooler multitectic component of this magma mixture persists in pockets or perhaps throughout the active shallow plumbing system; it has the same time-depleted incompatible-element signature (for example, La/Yb, Zr/Y; C.R. Thornber, unpublished data) as the high temperature component, which suggests that this component is maintained in a reservoir (or dike network) open to near-continuous magmatic recharge. Furthermore, the phenocryst phase relations, textures, and compositions, along with sulfur concentrations of melt inclusions, define a dynamic shallow mixing environment that is driven by magmatic recharge into a zone of denser, cooler, degassed, and partially crystalline shallow magma (Thornber and others, 2010).

The long-term petrologic monitoring effort during the East Rift Zone eruption from 1983 to present suggests that such quasi-stagnant magma zones are perpetually flushed during continuous eruption of lower-MgO hybrid lava. Geophysical and gas monitoring data are consistent with an increase in magma supply during the post-2001 interval (Poland and others, 2012). Counterintuitively, the apparent increase of magma supply to the edifice coincides with a pulse of low-MgO hybrid lava through the East Rift Zone that continues to the end of 2010 (fig. 6). This inverse relation suggests that cooler crystal mush zones within the shallow edifice were being flushed more efficiently during the period of increased magma supply.

Summit-to-Rift Magmatic Continuity in the Ongoing Eruption

Throughout the duration of this epic East Rift Zone eruption, physical connectivity between a summit magma reservoir and the rift zone vents has been inferred by correlations of geophysical flexure at the summit to surges and lulls in activity and changes in the MgO index (Thornber, 2001, 2003; Thornber and others, 2003b). Since the onset of summit eruption in 2008, petrologic analysis of lava simultaneously erupted at both ends of the eruptive plumbing system has provided a unique and unequivocal demonstration of magmatic continuity between the shallow summit and throughout the 18-km-long rift zone conduit. Trace-element signatures for glasses, glass inclusions, and bulk lava in 2008 rift zone and summit samples are indistinguishable (Rowe and others, 2009; Thornber and others, 2010). Olivine-phyric lava from the 2008-to-present Halemaʻumaʻu vent lies along the same incompatible-element-depleted LLD as all post-1995 Kīlauea lava (fig. 7*B*). This has significant implications for the overall magmatic condition of the Kīlauea edifice.

Although earthquake and summit deformation patterns have long implied a shallow reservoir of magma beneath Halemaʻumaʻu Crater (see, for example, Poland and others, this volume, chap. 5), its size, age, longevity, and physical characteristics have been topics of considerable speculation (Pietruszka and Garcia, 1999b; Thornber, 2003). Many assumed it would likely comprise the stagnant magma left from past summit eruptions in 1982, 1974, or perhaps even decades earlier.

Our recent petrologic evidence shows that the magma of the long-term East Rift Zone eruption has flushed out older resident summit magma and occupies the entire shallow magmatic plumbing system (Thornber, 2003). Furthermore, the persistent mixed-crystal assemblage in rift zone lava attests to the perpetual presence of magma mush along cooler margins of the active magma pathways and in cavities, cracks, or sills within the shallow volcanic edifice. A zone of mixing between hotter magma that recharges the edifice and cooler magma mush that resides within it may exist in the summit reservoir, adjacent to the East Rift Zone conduit, but it is also possible that preeruptive magma mixing occurs all along the transport path to the East Rift Zone vent. The persistently low-MgO lava must reflect a steady-state condition in which a volumetrically significant open reservoir within the shallow edifice is being perpetually refreshed by recharge magma. This dynamic condition of variable recharge with two simultaneous eruptions at different sites underlines the open-system nature of Kīlauea’s magmatism.

Investigations of Hawaiian Lava Lakes

Investigations of lava lake activity and behavior have been an important component of HVO’s scientific output for the entire century of the observatory’s existence. Two distinct features have been called lava lakes: the first are active lava lakes underlain by a magma column (as at Halema‘uma‘u before 1924 and during 1952, 1967–68 and 2008–present; Mauna Ulu during 1969–74; and Pu‘u ‘Ō‘ō, especially during 1983–86), and the second are closed-system lava lakes, produced when lava ponds in preexisting pit craters and is no longer in contact with its source.

Active Lava Lakes

The most famous Hawaiian lava lake occupied Halema‘uma‘u Crater for most of the 19th and early 20th centuries, dominating the attention of early observers, as well as that of Dr. Jaggar in HVO’s early years (Jaggar, 1917a, 1917b; Bevens and others, 1988). In spite of the hazardous environment, Jaggar collected a small suite of glassy spatter samples discussed earlier but was not able to collect the more viscous material he referred to as “bench lava,” which surrounded the more fluid parts of the lava lake; however, he described seeing incandescent slabs of bench lava spall off the sides, buckling and then disintegrating to glowing talus (Jaggar, 1917b) during draining of the lava lake. His vivid description suggests that this material was similar to the melt-rich mush encountered in various historical lava lakes. The mush (fig. 8) consists of a very tenuous crystal network in a matrix of as much as 45 volume percent melt. Although this mush has been drilled and recovered, bore holes in it collapse within days (Helz and Wright, 1983), much as the bench lava collapsed gradually after withdrawal of the active lava lake.

More recently, significant lava lake activity in Halema‘uma‘u occurred during the 136-day 1952 eruption (Macdonald, 1955) and for 18 months in 1967–68 (Kinoshita and others, 1969). The current summit activity (2008 to present) differs from these earlier eruptions in that the top of the magma column is visible within a small pit crater 160 m in diameter (T. Orr, written commun., 2012) against the wall of Halema‘uma‘u Crater (location shown in fig. 5) rather than occupying much of the crater’s floor. Webcam images of the magma pool show similar magma surface patterns to those observed on the magma column at Mauna Ulu (Duffield, 1972). The relatively small area is similar to the active lava lake that existed at the beginning of HVO’s life, which occupied only a small part of Halema‘uma‘u (Jaggar, 1917a, 1917b).

The style of activity of this lava lake generates mostly spatter, but microbeam techniques now allow extensive analysis of even tiny amounts of material in support of petrologic study. Samples from the 2008–present summit eruption are similar to Halema‘uma‘u scoria and spatter sampled over the past 100 years—they contain few crystals, and both glass and bulk compositions have a limited range of MgO (figs. 3 and 9), which lies at the end of Kīlauea’s olivine control line (as discussed in the previous section).

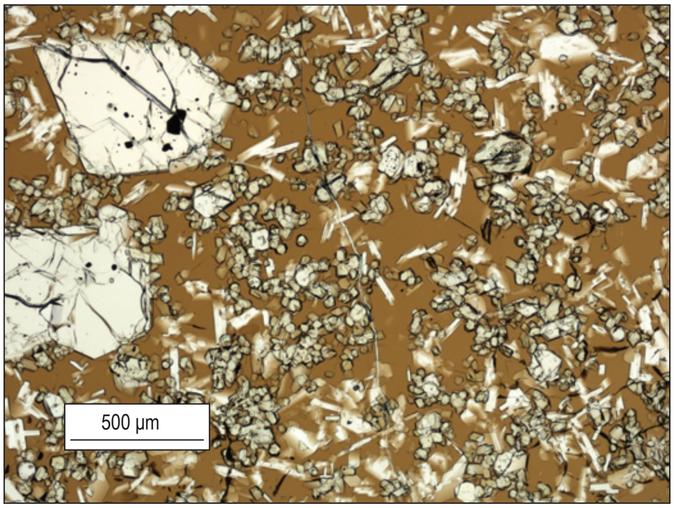


Figure 8. Photomicrograph showing groundmass in partly molten core from near the limit of drillability of the crust of the Kīlauea Iki lava lake. Brown glass makes up 40–45 percent by volume of the core. Crystals include phenocrystic, microphenocrystic, and groundmass olivine, plus groundmass augite (green) and plagioclase. The crystals tend to nucleate on each other, leaving more melt-rich areas between the clusters and chains of crystals.

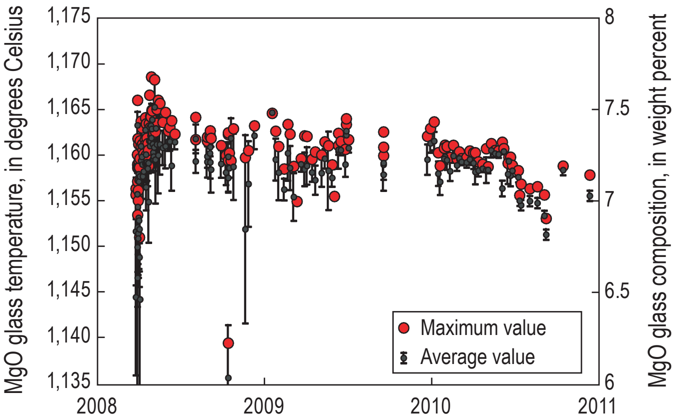


Figure 9. Graph showing variation of MgO content and glass quenching temperature of tephra from the ongoing (as of 2014) Kīlauea summit eruption versus time. Maximum values, shown as red dots, provide the best indication of changing magmatic conditions in the summit reservoir. Average values with standard deviation are indicated with a black dot and vertical bar. Data from Wooten and others (2009) and C.R. Thornber (unpublished data).

Closed-System Lava Lakes and Sills

The second class of lava lakes, not fed from below but formed by lava flowing into preexisting pit craters (as described in Richter and others, 1970; Peck and Kinoshita, 1976; Wright and others, 1968), has received major attention at HVO, and study of these bodies has contributed greatly to our understanding of the crystallization and fractionation behavior of tholeiitic magma. Many lava lakes have formed in the 20th century (fig. 10); three of these, the 1959 Kīlauea Iki, 1963 ‘Alae, and 1965 Makaopuhi lava lakes, have been studied extensively. Older analogues include a prehistoric Makaopuhi lava lake (Moore and Evans, 1967; Evans and Moore, 1968), the Uēkahuna laccolith (Murata and Richter, 1961), a prehistoric lava lake at Mauna Loa that has been recovered in fragments (McCarter and others, 2006), and similar bodies observed in the walls of Moku‘āweoweo (Sparks, 1990). Table 2 summarizes some of the properties and data sets available for the lava lakes and sill at Kīlauea.

Kīlauea Iki, ‘Alae, and the 1965 Makaopuhi lava lakes have been the focus of elaborate field studies, including (1) observations of their filling and surficial subsidence, (2) repeated core drilling, with sample collection and petrologic and chemical analysis, and (3) field determinations of temperature, melt viscosity, oxygen fugacity, seismic, electrical and other geophysical properties. Wright and others (1976) summarized the range of results for all these lakes, and further results for Kīlauea Iki were summarized in Helz (1987b) and Barth and others (1994). The sections below compare selected results from the historical lava lakes with those for the prehistoric Makaopuhi lava lake and the Uēkahuna laccolith.

Controls on Whole-Rock Compositions in Lava Lakes and Sills

Petrologic studies of lava lakes at Kīlauea have documented how tholeiitic basalt differentiates as it cools and crystallizes by tracking changes in bulk composition and interstitial melt composition. The crystallization behavior and liquid lines of descent for the relatively olivine-poor 1965 Makaopuhi and 1963 ‘Alae lava lakes (MgO=7.5–8.2 weight percent; table 2) were described by Wright and Okamura (1977), Peck and others (1966), and Wright and Peck (1978), while the liquid lines of descent and extreme differentiation products of Kīlauea Iki were described in Helz (1987b). Unfortunately, the ‘Alae and 1965 Makaopuhi bodies were covered by lavas of the Mauna Ulu eruption (as was the cliff face through the prehistoric Makaopuhi lava lake), so they have not been studied further.

Figure 11 shows magnesia variation diagrams for selected oxides (SiO_2, Al_2O_3, and CaO) for Kīlauea Iki (MgO=15.5 weight percent; Wright, 1973) and the Uēkahuna laccolith (MgO=16.0 weight percent; Murata and Richter, 1961), both of which are still accessible. The figure also includes analyses from the prehistoric Makaopuhi lava lake (MgO=9.67 weight percent; Moore and Evans, 1967; Gunn, 1971), because the analyzed section includes its

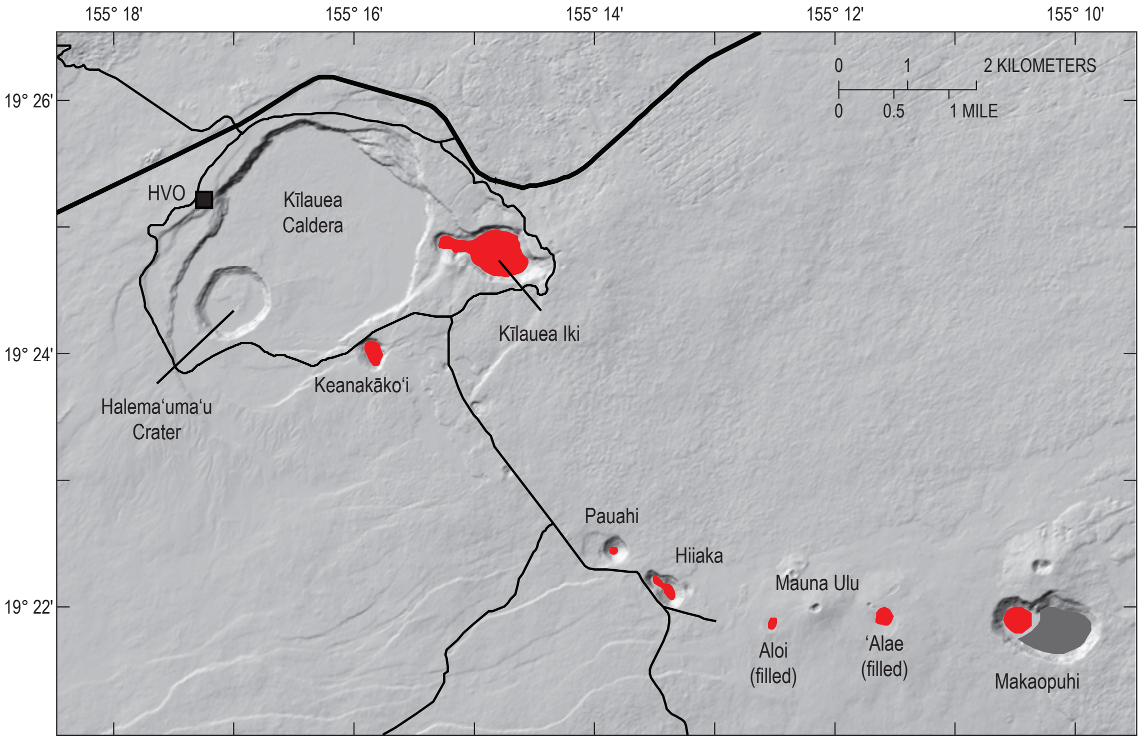


Figure 10. Index map of Kīlauea’s summit and upper East Rift Zone showing locations of recent lava lakes (red), plus the prehistoric Makaopuhi lava lake (gray). Of the lava lakes shown, only ‘Alae (1963), Makaopuhi (1965), Kīlauea Iki (1959), and the prehistoric Makaopuhi lava lakes have been extensively studied.

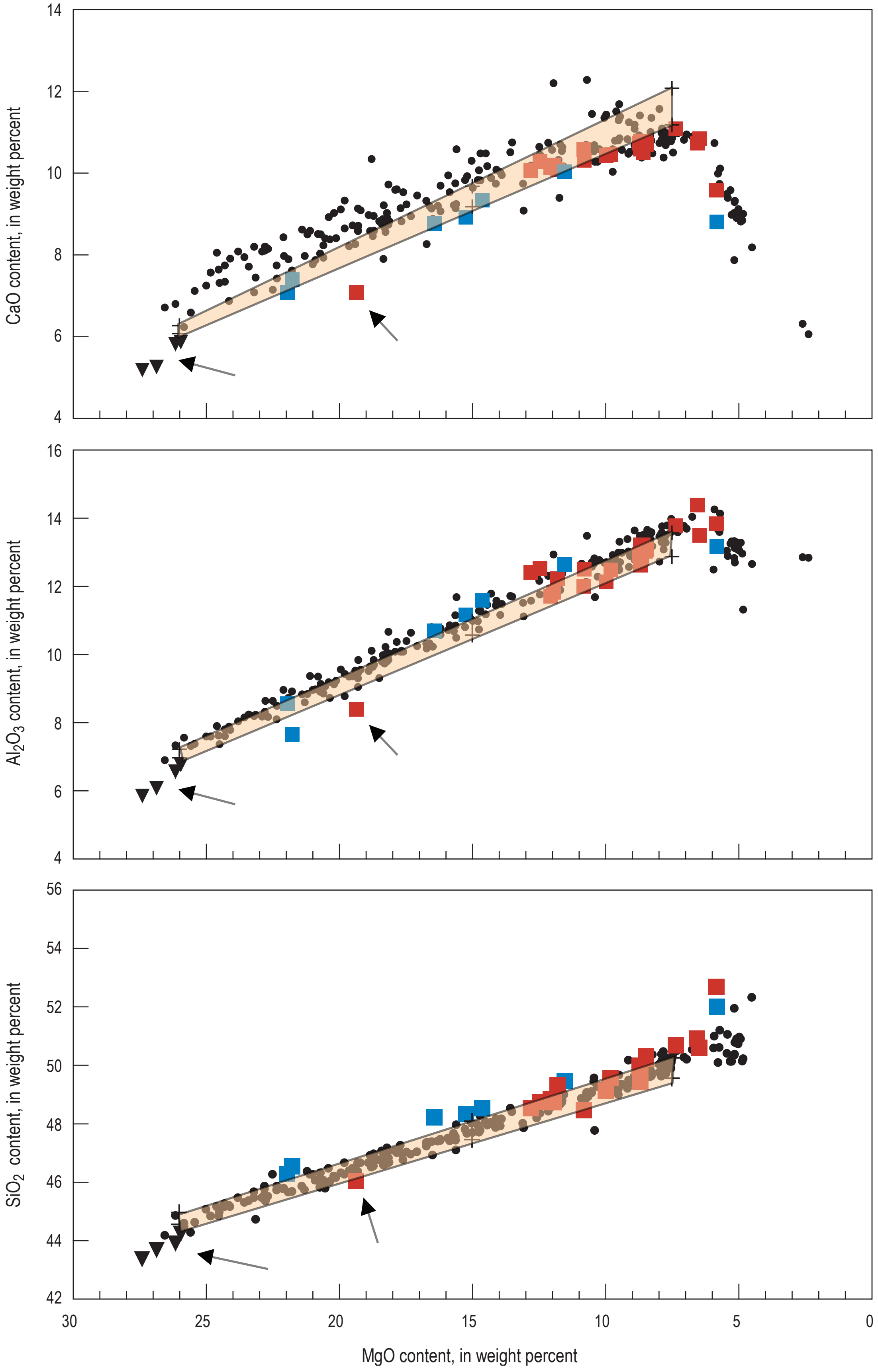
Table 2. Properties and data available for lava lakes and sills at Kīlauea Volcano.

[m, meters; est., estimated; max., maximum; XRF, X-ray fluorescence; INAA, instrumental neutron activation analysis; EDXRF, energy-dispersive X-ray fluorescence; PGE, platinum group elements; px, pyroxenes; Re, rhenium]

	1963 ‘Alae	1965 Makaopuhi	1959 Kīlauea Iki	Prehistoric Makaopuhi	Uēkahuna laccolith
Thickness of body (m)	13.7	83 (est.)	130–135 (est. from 1988 holes)	69 (cliff) to 120 (est.)	8.5 (sampled) to 27 (max.)
Average bulk MgO content (weight percent)	7.6	8.2	15.5	9.7	16
Present accessibility	Covered by Mauna Ulu shield	Covered by Mauna Ulu lavas	Still exposed	Cliff covered by Mauna Ulu lavas	Still exposed
Temperature profiles	1963–67, through body	1965–1969, to 11.6 m depth	1960–1988, to 76 m (center), through body at edge (95.4 m)	None	None
Thermal modeling	Yes	No	Yes	Yes	No
Samples	Drill core* through body	Drill core* to 20 m	Drill core* to 114.6 m, through body’s (95.4 m) north edge	Outcrop samples from cliff (69 m)	Outcrop samples from cliff (8.5 m)
Major element chemistry	33 classical	72 classical	194 classical 64 XRF	13 classical 8 XRF	8 classical
Reference(s) (majors)	Wright and Peck (1978)	Wright and Okamura (1977)	Richter and Moore (1966); Helz and others (1994); Helz and Taggart (2010)	Moore and Evans (1967); Gunn (1971)	Murata and Richter (1961)
Trace element chemistry	14 spectrographic	None	72 INAA + 57 EDXRF; 19 PGE + Re	8 XRF	None
References (traces)	Wright and Peck (1978)	--	Helz (2012); Pitcher and others, (2009)	Gunn (1971)	--
Electron microprobe phase chemistry	One residual glass	3 px, 2 residual glasses, oxides	Olivine (this report), glasses (Helz, 1987b), olivine + chromite (Scowen and others, 1991)	All crystalline phases (Moore and Evans, 1967; Evans and Moore, 1968)	--

[*] Drill core available through the National Museum of Natural History, Smithsonian Institution, Washington, D.C.

Figure 11. Graphs showing whole-rock contents of SiO_2, Al_2O_3, and CaO, plotted against MgO content (all quantities in weight percent), for the prehistoric Makaopuhi lava lake (Moore and Evans, 1967; Gunn, 1971; red squares), and for the Uēkahuna laccolith (Murata and Richter, 1961; blue squares). Also shown for comparison are data from 1967–88 core from Kīlauea Iki lava lake, shown as black dots (Helz and others, 1994; Helz and Taggart, 2010). Rock compositions that contain less than 6 weight percent MgO are segregation veins and dikelets and are present in all three bodies. Arrows indicate vertical olivine-rich bodies (vorbs) in Kīlauea Iki (inverted black triangles) and in the prehistoric Makaopuhi lava lake. The colored field shows the range of compositions of the 1959 eruption samples (Murata and Richter, 1966) and 1960–61 drill core from Kīlauea Iki lava lake (Richter and Moore, 1966). The sides of this field are olivine control lines for $Fo_{86–87}$. See text for further discussion.



olivine-enriched zone, allowing direct comparison with the olivine-rich zones in the picritic bodies. The three diagrams chosen are sufficient to show the effects of crystallization of olivine, augite, and plagioclase on the range of whole-rock compositions observed.

For most sets of samples, olivine control is the dominant source of bulk compositional variation; in each case, this has been attributed to redistribution of phenocrystic olivine within the body (Murata and Richter, 1961; Moore and Evans, 1967; Helz and others, 1989). Data for 1967–88 Kīlauea Iki core samples lie mostly within the colored field that is defined by the compositional ranges of the 1959 eruption samples and early (1960–61) core samples in SiO_2 versus MgO. Those for the two prehistoric bodies lie parallel to the field at slightly higher SiO_2 contents. This distribution shows that, in all cases, the olivine composition was $Fo_{86–87}$ at the time of redistribution. The array of points in Al_2O_3 versus MgO is also parallel to the $Fo_{86–87}$ control lines for all three bodies, showing that plagioclase was not materially involved as a fractionating phase, except for samples with MgO<7 weight percent, as has been widely observed for other Kīlauea lavas (Wright, 1971).

The patterns for CaO versus MgO are more complex: highly magnesian Kīlauea Iki samples are enriched in CaO, relative to the 1959 eruption samples, while the Uēkahuna samples run lower, but parallel to, the $Fo_{86–87}$ control lines. However, samples with 7–11 weight percent MgO, both from Kīlauea Iki and from the prehistoric Makaopuhi lava lake, lie on a slope flatter than olivine control, which gives the appearance of olivine+augite fractionation. Augite does not crystallize until melt MgO=7.2 weight percent (Helz and Thornber, 1987), so the flattening is not produced by fractionation of olivine+augite from liquids having the bulk composition of the samples, but by migration of interstitial liquid from within the olivine-enriched zone to upper parts of the lake (Moore and Evans, 1967; Helz and others, 1989). In Kīlauea Iki, this process clearly occurs after settling of the olivine phenocrysts. The Uēkahuna laccolith does not show signs of this process, probably because it is much thinner (table 2).

The lava lakes and laccolith discussed here contain internal differentiates in addition to the dominant, olivine-phyric matrix rock. Figure 11 includes analyses of segregation veins from Kīlauea Iki and prehistoric Makaopuhi lava lakes, and one analysis of aphanitic dikelets from the Uēkahuna laccolith. All have similar compositions, with bulk MgO contents of 4–6 weight percent and low CaO and Al_2O_3 relative to the olivine-controlled samples. They correspond to melts with liquidus temperatures of 1,100–1,135 °C (Wright and Okamura, 1977; Helz, 1987b; Helz and others, 1989) and have been interpreted as filter-pressed liquids from within a mush crystallizing olivine+augite+plagioclase (Moore and Evans, 1967; Helz, 1980, 1987b; Helz and others, 1989). Similar coarse differentiates were reported by McCarter and others (2006) for a disrupted lava lake erupted as blocks on Mauna Loa.

Kīlauea Iki lava lake also contains vertical, cross-cutting pipes of varying character, enriched in segregation-vein liquid, which have been interpreted as diapir tracks related to the formation of segregation veins. Most abundant are the vertical olivine-rich bodies (“vorbs”) observed in all cores from Kīlauea Iki in the depth range 18–58 m from 1975 onward (Helz, 1980; Helz and others, 1989). Other vertical structures found at and below 78-m depth resemble the melt chimneys of Tait and others (1992) and are interpreted as points of escape of plumes of vesicles+melt from within the lower crust of the lava lake (Helz, 1993; Helz and Taggart, 2010). Both sets of vertical structures, when quenched in a partly molten state, are variably enriched in Fe-rich olivine ($Fo_{78–80}$; Helz and others, 1989) and highly enriched in segregation-vein melt and vesicles relative to the adjacent matrix (fig. 12*A*,*B*,*D*).

Moore and Evans (1967) noted a population of “pipe-like masses, rich in olivine” in the prehistoric Makaopuhi lava lake, one of which is shown in figure 12*C*. The one analysis of such a body is similar to the vorbs in Kīlauea Iki: it is enriched in segregation-vein liquid and in Fe-rich olivine, so (like the vorbs) is low in SiO_2, Al_2O_3, and CaO relative to the matrix rock, although high in MgO, as shown in figure 11. The position of these pipes low in the cliff face and their relatively short (0.15 m) height (Moore and Evans, 1967), however, makes them more like the deeper vertical structures found in Kīlauea Iki (for example, fig. 12*D*). McCarter and others (2006) describe an “open-textured dunite” from the Mauna Loa body as a possible vorb, but its relationship to the rest of the dismembered lava lake is unknown.

Appearance, Distribution, and Abundance of Segregation Veins

All Kīlauea lava lakes contain segregation veins. Though variously described as diabase (Peck and others, 1966), ferrodiabase (Helz, 1980, 1987b), or mafic micropegmatite (Moore and Evans, 1967), they are very consistent in chemical and petrographic character. Figure 13*A* shows a partly molten segregation vein from Kīlauea Iki; it is olivine-free and consists of pyroxenes, plagioclase, Fe-Ti oxides, and brown glass, with all crystalline phases 1–2 orders of magnitude coarser grained than their equivalents in the underlying olivine-phyric matrix. The subhorizontal, relatively sharp lower contact is typical. The 13.4-m thick 1963 ‘Alae lava lake contains one such vein (0.4 m thick, or 3 percent of the lake’s thickness), and several were encountered in drilling the upper 20 m of the 1965 Makaopuhi lava lake. The disaggregated lava lake described by McCarter and others (2006) also contains abundant coarse-grained internal differentiates like the segregation veins described here.

The cliff face through the prehistoric Makaopuhi lava lake and the extensive drilling program in Kīlauea Iki give the most complete picture of the extent of these internal differentiates. Figure 14 shows the array of segregation veins exposed

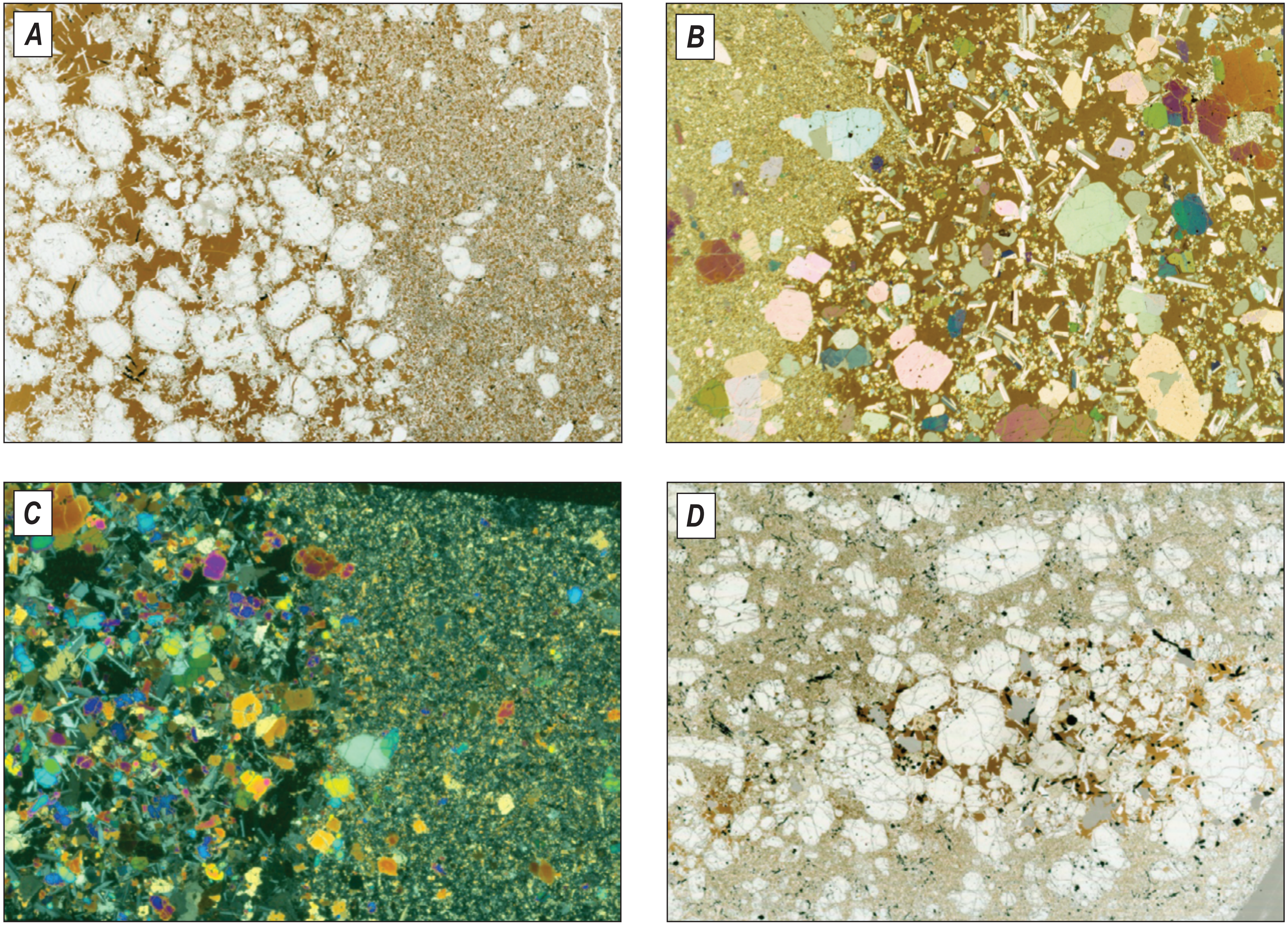


Figure 12. Photomicrographs of drill-core samples from Kīlauea lava lakes. *A*, Contact between vertical olivine-rich body, or "vorb" (left), and normal olivine-phyric matrix (sample KI75-1-141.5, T=1,095 °C) in plane light, Kīlauea Iki. Horizontal cut through vertical contact; field of view is 3.2 cm across. *B*, Irregular vertical contact between melt-rich chimney (right) and normal olivine-phyric matrix (KI81-5-258±1, T=1,130 °C) with nicols partly crossed, Kīlauea Iki. Melt chimney contains coarser plagioclase and more melt and vesicles than matrix. Vertical cut; field of view is 2.2 cm across. *C*, Contact between normal matrix (right) and "pipe-like mass rich in olivine" from the prehistoric Makaopuhi lava lake (MP-200) with crossed nicols. Field of view is 3 cm across. *D*, Small plume enriched in olivine, melt, and vesicles (gray) in normal olivine-phyric matrix (KI88-1-275.1, T=1,100 °C) in plane light, Kīlauea Iki. This is a variant melt chimney. Vertical cut, up is to the right. Field of view is 2.5 cm across.

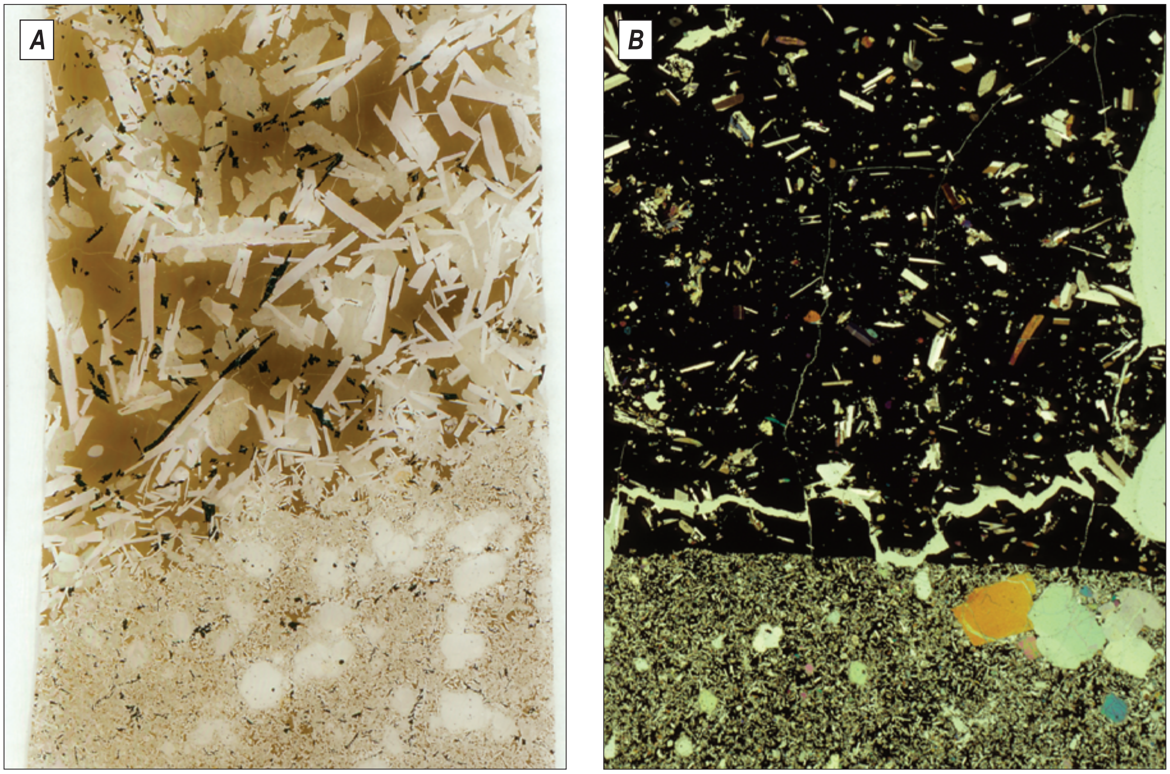


Figure 13. Photomicrographs showing contacts in drill-core samples from Kīlauea Iki lava lake. *A*, Subhorizontal contact of segregation vein with underlying olivine-phyric host in sample KI79-3-164.5 (T=1,068 °C) in plane light. Vertical cut; field of view is 1.7 cm across. *B*, Contact between black, quenched ooze (upper part of image) and olivine-phyric wall rock in borehole KI76-2, recovered at a depth of 141.7 ft (43.2 m), with nicols partly crossed. Vertical cut through core; contact is vertical, up to the right. Field of view is 1.7 cm across.

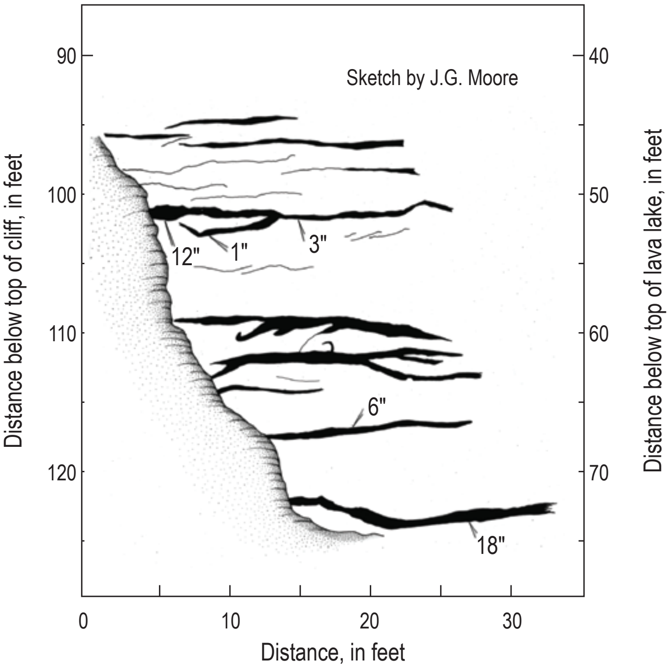


Figure 14. Sketch of segregation veins exposed in the cliff through the prehistoric Makaopuhi lava lake. This sketch was made by J.G. Moore while examining the cliff in a bosun's chair during the summer of 1962. The exposure was covered in 1969, when lava from Mauna Ulu filled the west pit to the level of the top of the cliff. One foot equals 0.3 m; 1 inch (1") equals 0.4 cm.

in a section of the cliff face that cut the prehistoric Makopuhi lava lake. The veins are coarse-grained, with relatively sharp boundaries, and typically show lateral extents 20–100 times greater than their thickness. For comparison, figure 15 shows the depth and thickness of all segregation veins encountered in the 1967–81 Kīlauea Iki drill cores. The overall pattern is similar to that observed in the prehistoric Makaopuhi cliff face. The individual segregation vein with the greatest inferred lateral extent (highlighted in red in fig. 15) is found at a depth of 45.72 m (150 ft) below the lake surface; it occurs in 10 cores and has the same thickness (~0.3–0.4 m) over a lateral distance of more than 150 m.

These stacks of internally produced sills make up an appreciable fraction of the upper crust of the lava lakes. That fraction is 6.2 to 11.5 percent by volume in Kīlauea Iki above ~53 m (fig. 15); the frequency of such sills in the prehistoric Makaopuhi lava lake (fig. 14) is similar. This corresponds to 2.5–4.5 percent of the thickness of the entire body, which compares closely with the 3 percent observed in the much thinner ʻAlae lava lake. Equivalent liquids would make up 45–50 percent by weight of Kīlauea Iki's bulk composition (Wright, 1973), so the amount of liquid segregated into veins is an order of magnitude less abundant than in the bulk composition.

The crosscutting aphanitic dikelets in the north body of the Uēkahuna laccolith have the composition of segregation veins (fig. 11) but differ in grain size and in their relation to

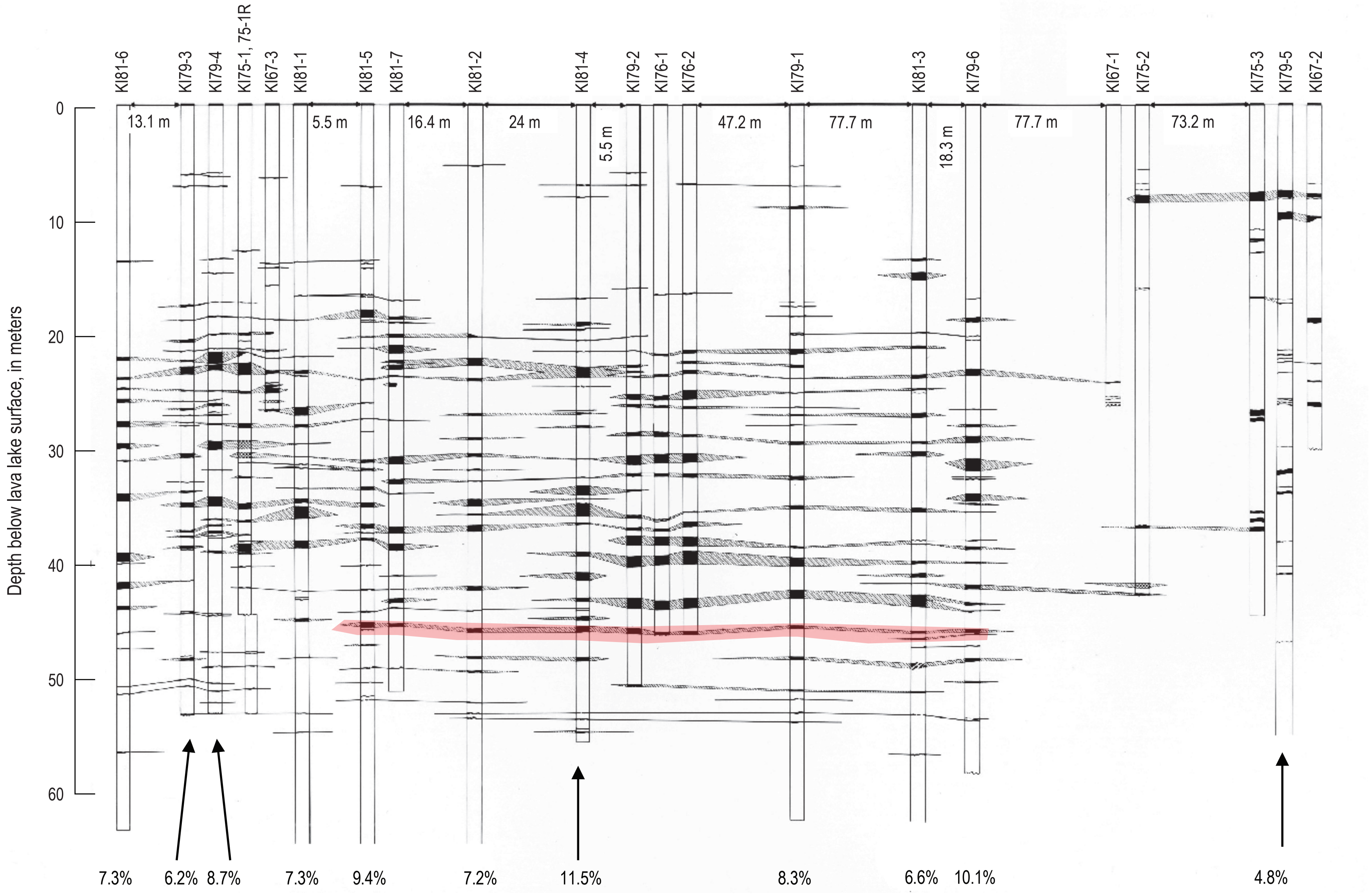


Figure 15. Correlation chart showing the depths and thicknesses of all segregation veins encountered in 1967–81 drill core in Kīlauea Iki lava lake; core identification numbers given at top. The solid black intervals show all veins thicker than 1 cm. Vertical thicknesses are to scale; horizontal distances between drill holes are not to scale, but distances between some cores indicated in meters (m). Dashed lines show correlations of segregation veins between drill holes based on depth. One particularly widespread vein at a depth of 45.7 m (150 ft) is highlighted in orange. Numbers below the columns give the fraction of segregation veins in the core between the surface of the lake and the deepest segregation vein. Other details as in Helz (1980, fig. 7).

the host body (Murata and Richter, 1961). They are not sill-like, and their aphanitic grain size resembles that of oozes recovered in boreholes in Kīlauea Iki (and other lava lakes), an example of which is shown in figure 13*B*. The oozes are crystal-poor melts, usually derived from the mush below the base of the drillable crust; they quench rapidly as they move into open boreholes. A possible origin for the aphanitic dikelets in Uēkahuna is that they represent interstitial melt released from partly crystalline mush during shearing of the host body. Exposure of the sheared face to surface conditions during caldera formation would have quenched the resulting melt stringers to an aphanitic texture. If so, the Uēkahuna body would have resembled the incandescent, truncated intrusives observed by Jaggar and Finch (1924) in the wall of Halemaʻumaʻu after the 1924 eruption. This model would constrain the age of the laccolith to be only slightly older than the age of formation of Kīlauea's current caldera (1470–90 C.E.; Swanson and others, 2012), making it approximately coeval with the ʻAilāʻau activity (Clague and others, 1999).

Thermal History of Lava Lakes

ʻAlae and Kīlauea Iki lava lakes were monitored thermally for most of their crystallization history. Cooling of the prehistoric Makaopuhi lava lake was modeled (Moore and Evans, 1967) using a simple conductive cooling model (Jaeger, 1961). Peck and others (1977) included the effects of rainfall in their modeling of the cooling of ʻAlae lava lake; with this modification, they were able to match observed field measurements of temperature very closely (fig. 16*A*). Ryan (1979) reported results of two-dimensional thermal modeling of Kīlauea Iki, but the grid used was coarse and the results not directly comparable to those reviewed here.

A composite of temperature profiles for the central part of Kīlauea Iki, obtained for 1961 through 1988, is shown in figure 16*B*. Above 100-m depth, the thermal evolution of the lava lake is well documented. Constraints on the curves below 100 m are provided by (1) an estimate of the depth of the lava lake, which is generally about 130 m (Barth and others, 1994), but probably slightly deeper at the center of the lake (Helz, 1993), and (2) the assumption that the lower contact temperature is half of the initial temperature of 1,190 °C, as expected for conductive cooling (Jaeger, 1961). The observed curves compare closely with both observed and model temperature profiles through the much thinner ʻAlae lava lake, which cooled to subsolidus temperatures in about 9 months (Peck and others, 1977; Peck, 1978). Based on extrapolation of these data, the interior of Kīlauea Iki reached subsolidus temperatures in the mid-1990s, giving a cooling time of about 35 years. This solidification time is similar to the 30 years Moore and Evans (1967) estimated for the prehistoric Makaopuhi body to reach the solidus throughout.

The effect of high rainfall at Kīlauea Iki is clearly visible in the upper parts of the temperature profiles. The subsolidus interval of the curve for 1975, a year of unusually

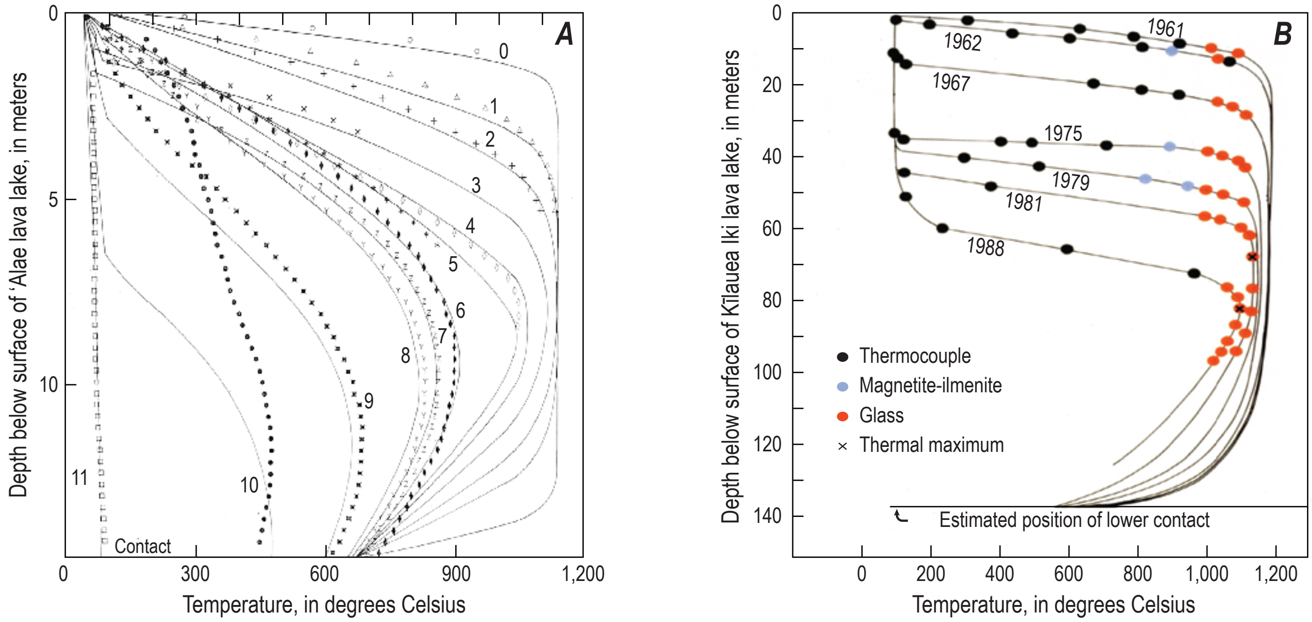


Figure 16. Graphs showing temporal evolution of vertical temperature profiles through Kīlauea lava lakes. *A*, Computed (continuous curves) and observed (symbols) temperature profiles in the 1963 ʻAlae lava lake during the 4 years following its formation, after figure 4 in Peck and others (1977). Data for individual temperature profiles in Peck (1978). *B*, Variation of temperature with depth below the surface of Kīlauea Iki lava lake as a function of time for drill holes near the center of the lake. Temperatures estimated by glass geothermometry (Helz and Thornber, 1987), from coexisting magnetite-ilmenite compositions (Buddington and Lindsley, 1964), or measured by thermocouples lowered down the hole, as indicated. Shapes of curves below ~100 m are constrained by (1) an assumed depth of 135 m (based on results for drill hole KI88-2; see discussion in Helz, 1993) and (2) a lower contact temperature of ~600 °C, based on an initial lake temperature of ~1,190 °C (Helz, 2009).

Table 3. Cooling rates in the partly molten interior region of Kīlauea Iki lava lake.

[Numbers in parentheses are depths (in feet) of the individual samples in figure 16*B* used in calculations. Temperatures are based on glass geothermometry (Helz and Thornber, 1987)]

Time interval (years)	Temperature interval (°C)	Cooling rate
1959 to 1967 (84.5)*	1,190 to 1,090*	12.5 °C/year
1959 to 1975 (145.1)*	1,190 to 1,120*	4.4 °C/year
1959 to 1976 (150)*	1,190 to 1,115*	4.4 °C/year
1959 to 1979 (203.6)*	1,190 to 1,141*	2.4 °C/year
1959 to 1981 (219.8)*	1,190 to 1,140*	2.3 °C/year
1959 to 1988 (266.6)*	1,190 to 1,104*	3.0 °C/year
1979 (203.6) to 1981 (205.4)	1,141 to 1,135	3.0 °C/year
1981 (269.9) to 1988 (266.6)	1,128 to 1,106	3.1 °C/year
1981 (299.9) to 1988 (300)	1,099 to 1,081	2.6 °C/year

* Hottest sample recovered in the year the core was obtained.

high rainfall, is flatter than most, while that of the 1988 curve, monitored during a drought, is steeper. Rainfall did not affect the shapes of curves in ‘Alae, but the modeling showed that it did hasten cooling of that body. Similar modeling for Kīlauea Iki predicted a rate of growth of the upper crust that was much faster than observed (Wright and others, 1976; Peck and others, 1977), with the drillable crust being only 43 m thick in 1975 instead of the predicted 58 m. Possible factors include: (1) condensation of steam and recirculation of water in the thicker upper crust of Kīlauea Iki (Wright and others, 1976; Peck and others, 1977) and (2) extensive transfer of melt and heat upward in Kīlauea Iki (Helz and others, 1989). Whatever the balance of these effects, the rate of growth of the upper crust in Kīlauea Iki was nearly constant between 1967 and 1979, at 2.3–2.4 m per year (Helz, 1980).

Thermal data for Kīlauea Iki show that cores collected at different times and depths have very different cooling histories. For example, partly molten core recovered from 60–90 m deep cooled at 2–3 °C per year, over the period from 1959 to 1979–88, for all time intervals where data are available (table 3). By contrast, the hottest samples recovered in 1967 cooled from 1,190 °C to 1,090 °C over 8 years, at an average rate of 12.5 °C per year. The hottest parts of cores from 1975 and 1976 cooled at rates between these values. Profiles in the upper crust flatten at less than ~1,100 °C (fig. 16*B*); below that temperature, the solidus is reached in one year's time, for a cooling rate of 100+ °C per year. These variations in cooling rate have significantly affected the crystallization and reequilibration of minerals in Kīlauea Iki, especially olivine.

Crystallization and Reequilibration of Olivine in Kīlauea Lava Lakes

The classic study of Moore and Evans (1967) on the prehistoric Makaopuhi lava lake focused on the observed compositions and inferred reequilibration of its olivine phenocrysts. For Kīlauea Iki, Helz (1987b) described the effects of reequilibration of olivine on interstitial melt compositions and groundmass assemblages, while Scowen and others (1991) documented how chromite inclusions in olivine phenocrysts reequilibrated as olivine became more Fe-rich. This section presents olivine data for comparison with the results of Moore and Evans (1967). Because Kīlauea Iki was sampled as it crystallized, and glass composition varies with temperature (Helz and Thornber, 1987), it is possible to determine how olivine has crystallized and reequilibrated as a function of cooling rate and local bulk composition.

Kīlauea Iki contains abundant olivine crystals of all sizes. The phenocrysts were present in the eruption scoria (Helz, 1987a), as were many smaller olivines, but some smaller crystals (<1 mm in length), plus overgrowths on inherited phenocrysts, have grown in situ in the lava lake. Table 4 summarizes data on the compositions of smaller olivines and rims on phenocrysts (hereafter refered to as “in situ” olivines) for 36 samples chosen to show the range of crystallization and reequilibration behavior in Kīlauea Iki. The average composition of in situ olivine decreases linearly as temperature decreases from 1,150 to 1,080 °C (fig. 17) and is always more Fe-rich than the initial average composition ($Fo_{86–87}$).

Table 4. Average composition and range for olivine in samples from Kīlauea Iki lava lake, with glass quenching temperatures (Helz and Thornber, 1987), depth and bulk MgO content of the sample, and identity of coexisting low-Ca pyroxene, if present.

[Sample numbers give year and drill core number (KI67-3), followed by the sample depth in feet (75.0). All samples contain augite and plagioclase; R, replicate analysis; phenocryst core compositions excluded from data except as noted (incl.); mol % Fo, molar percent forsterite; wt %, weight percent; ---, no data; vorb, vertical olivine-rich body; Opx, orthopyroxene]

Sample no.	Average (mol % Fo)	Range (mol % Fo)	Temperature (°C)	Depth (m)	Bulk MgO (wt %)	Any Low-Ca pyroxene?	Comments
A. 1967 samples and subsolidus equivalents							
KI67-3-75.0	61.0	69.5 to 52.2	995	22.86	9.47	Pigeonite	
KI67-3-77.3	61.8	72.1 to 54.5	1,026	23.56	7.90	Pigeonite	
KI67-3-80.7	64.4	72.5 to 56.0	1,057	24.60	7.73	Pigeonite	
KI67-3-83.8	64.1	69.9 to 61.1	1,076	25.54	7.54	Pigeonite	
KI67-3-83.8R	65.8	71.6 to 61.2	1,076	25.54	7.54	Pigeonite	
KI67-3-84.5	69.0	74.5 to 63.4	1,084	25.76	7.69	Pigeonite	
KI75-1-85.5	60.0	64.4 to 52.2	Subsolidus	26.06	8.33	Pigeonite	
KI81-1-86.0	56.8	59.7 to 53.4	Subsolidus	26.21	8.74	Pigeonite	
B. 1975 samples and subsolidus equivalents							
KI75-1-130±	63.9	71.2 to 59.0	985	39.62	9.73	Pigeonite	
KI75-1-133.3	65.4	71.7 to 58.8	1,035	40.63	10.75	Pigeonite	
KI75-1-134.4	65.4	73.4 to 61.6	1,050	40.96	10.90	Pigeonite	
KI75-1-138±	67.4	75.0 to 63.3	1,066	42.1	---	Pigeonite	
KI75-1-139.3	70.1	77.1 to 67.1	1,084	42.46	11.64	Pigeonite	
KI75-1-143.8	75.2	77.8 to 74.5	1,115	43.83	12.15	None	
KI75-1-143.8R	76.1	79.1 to 74.5	1,115	43.83	12.15	None	
KI75-1-144.9	77.2	77.6 to 76.5	1,117	44.16	---	None	
KI75-1-144.9R	77.2	77.3 to 77.0	1,117	44.16	---	none	
KI81-1-145.1	64.8	69.1 to 63.0	Subsolidus	44.23	---	Pigeonite	
KI88-2-138.9	58.6	63.8 to 55.2	Subsolidus	42.34	---	Pigeonite	
C. 1979 samples and subsolidus equivalents							
KI79-3-157.7	67.1	69.6 to 61.4	985	48.07	---	Pigeonite	
KI79-3-157.7R	66.5	72.8 to 61.2	985	48.07	---	Pigeonite	
KI79-3-163.7	67.2	71.1 to 64.4	1,057	49.90	---	Pigeonite	
KI79-3-163.7R	69.1	73.4 to 66.1	1,057	49.90	---	Pigeonite	
KI79-3-166.1	74.1	75.1 to 72.7	1,071	50.63	27.41	Pigeonite?	Vorb
KI79-3-166.1R	75.0	76.5 to 74.1	1,071	50.63	27.41	Pigeonite?	Vorb
KI79-3-172.9	75.7	79.7 to 74.1	1,107	52.70	18.71	None	
KI79-3-172.9R	76.9	79.5 to 74.4	1,107	52.70	18.71	None	
KI79-1-187.4	79.8	80.9 to 79.1	1,130	57.12	19.3	None	
KI79-1-203.6	80.9	81.3 to 80.5	1,141	62.06	22.68	None	Phenocrysts incl.
KI88-2-147.4	65.8	69.2 to 63.1	Subsolidus	44.93	14.8	Pigeonite	
KI88-2-160.5	68.6	74.2 to 65.0	Subsolidus	48.92	17.7	Pigeonite	

Table 4. Average composition and range for olivine in samples from Kīlauea Iki lava lake, with glass quenching temperatures (Helz and Thornber, 1987), depth and bulk MgO content of the sample, and identity of coexisting low-Ca pyroxene, if present.—Continued

[Sample numbers give year and drill core number (KI67-3), followed by the sample depth in feet (75.0). All samples contain augite and plagioclase; R, replicate analysis; phenocryst core compositions excluded from data except as noted (incl.); mol % Fo, molar percent forsterite; wt %, weight percent; ---, no data; vorb, vertical olivine-rich body; Opx, orthopyroxene]

Sample no.	Average (mol % Fo)	Range (mol % Fo)	Temperature (°C)	Depth (m)	Bulk MgO (wt %)	Any Low-Ca pyroxene?	Comments
D. 1981 and 1988 samples and subsolidus equivalents							
KI81-1-181.5	74.7	78.2 to 71.4	1,036	55.32	17.84	Opx	
KI81-1-181.5R	73.8	77.2 to 73.2	1,036	55.32	17.84	Opx	
KI81-1-181.5	78.5	80.2 to 74.7	1,036	55.32	---	None	Dunite inclusion
KI81-1-181.5R	79.2	81.2 to 74.3	1,036	55.32	---	None	Dunite inclusion
KI81-1-186.7	75.8	76.8 to 74.6	1,085	56.91	---	Opx	
KI81-1-205.4	79.9	80.2 to 79.8	1,134	62.61	---	None	Phenocrysts incl.
KI81-1-205.4R	80.1	81.0 to 79.3	1,134	62.61	---	None	Phenocrysts incl.
KI81-1-219.8	80.0	80.0 to 79.8	1,140	67.00	23.58	None	Phenocrysts incl.
KI81-1-219.8R	80.4	80.5 to 80.2	1,140	67.00	23.58	None	Phenocrysts incl.
KI81-1-269.9	77.1	78.0 to 76.8	1,128	82.26	18.79	None	
KI81-1-306.7	75.2	76.5 to 74.1	1,087	93.48	17.47	Opx	
KI81-1-306.7R	74.9	76.5 to 74.2	1,087	93.48	17.47	Opx	
KI88-2-205.2	74.2	76.9 to 72.3	Subsolidus	62.54	19.8	Opx	
KI88-2-266.6	77.4	78.5 to 76.8	1,105	81.26	20.9	Opx	
KI88-2-300	72.4	74.7 to 71.8	1,081	91.4	----	Opx	
KI88-2-310	73.8	75.3 to 73.2	1,066	94.5	22.5	Opx	Foundered crust
KI88-2-336.3	61.4	62.8 to 60.7	~970	102.5	9.49	Pigeonite	Below foundered crust

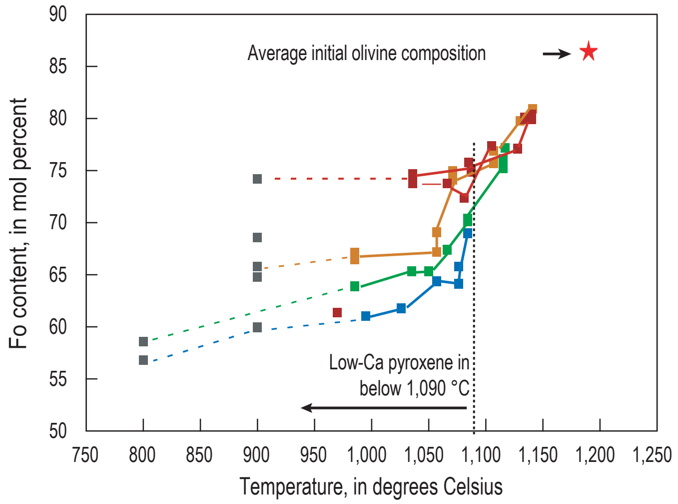


Figure 17. Graph showing average forsterite content (Fo) of in situ olivine versus temperature for core samples from Kīlauea Iki lava lake (data in table 4). The red star shows the average initial composition of olivine in the 1959 eruption samples. Individual points show the average in situ olivine composition for partly molten samples (blue=1967 core, green=1975 core, orange=1979 core, red=1981 and 1988 cores). Subsolidus samples (from 1975, 1981, and 1988 cores, recovered at T~100 °C) are shown in gray and plotted at arbitrary temperatures. Samples plotted at 900 °C are 6–9 years postsolidus; those at 800 °C are 14–21 years postsolidus. The color of the dashed lines shows which molten and subsolidus samples most resemble each other in depth and bulk MgO content. Further discussion in text.

At temperatures below the incoming of the second pyroxene (pigeonite in the less magnesian samples, orthopyroxene in the most olivine-rich core; see table 4), the average forsterite (Fo) content of in situ olivine changes more slowly. Also, the individual curves fan out, moving toward different subsolidus olivine compositions, depending on local bulk composition. That local bulk compositional control is important is shown by the location of the isolated red point in figure 17, which corresponds to the deepest sample analyzed (KI88-2-336.3, with MgO=9.49 weight percent), with an average in situ olivine composition like that in the low-MgO 1967 core. Data for subsolidus samples (plotted arbitrarily at 900 °C or 800 °C), are linked in figure 17 with the curves best matching their depth and bulk MgO content (table 4). The results suggest that the composition of in situ olivine continued to change slowly below the solidus (T=970–1,000 °C; Helz, 1987a).

Figure 18 shows how the compositions of olivine phenocryst cores, in situ olivine, and olivine megacrysts vary with depth in Kīlauea Iki lava lake; the partially molten samples from table 4 are shown in figure 18*A* and the smaller group of subsolidus samples in figure 18*B*. This projection allows comparison with the cooling history of the lava lake (fig. 16*B*), showing clearly that the range of compositions for in situ olivine (buff field) within partly molten samples reflects the average cooling rate. The relatively shallow, more rapidly cooled 1967 and 1975 samples show a wide range of in situ olivine composition. By contrast, the 1981–88 partly molten samples, which have cooled at 2–3 °C per year over 22–29 years (see table 3; see also clustered cooling curves in fig. 16*B*), have narrow ranges of in situ forsterite content. For these hottest, most slowly cooled samples, even phenocryst core compositions fall within the range observed for microphenocrystic and groundmass olivine (table 4), so they are not shown separately in figure 18*A*.

Figure 18 also shows that few olivine crystals retain the Fo$_{86-87}$ composition observed in the eruption samples. In particular, olivine phenocrysts in the olivine-rich zone present between 60 and 90 m in the lake have reequilibrated from Fo$_{86-87}$ to Fo$_{74-82}$, with compositions confined to 79.8–81.3 mol percent Fo in the very hottest samples. Because of this uniformity, no zoning is visible in back-scattered electron (BSE) imagery of these samples (fig. 19*A*). Data from Scowen and others (1991) and Vinet and Higgins (2011) show similar, restricted ranges of olivine compositions for

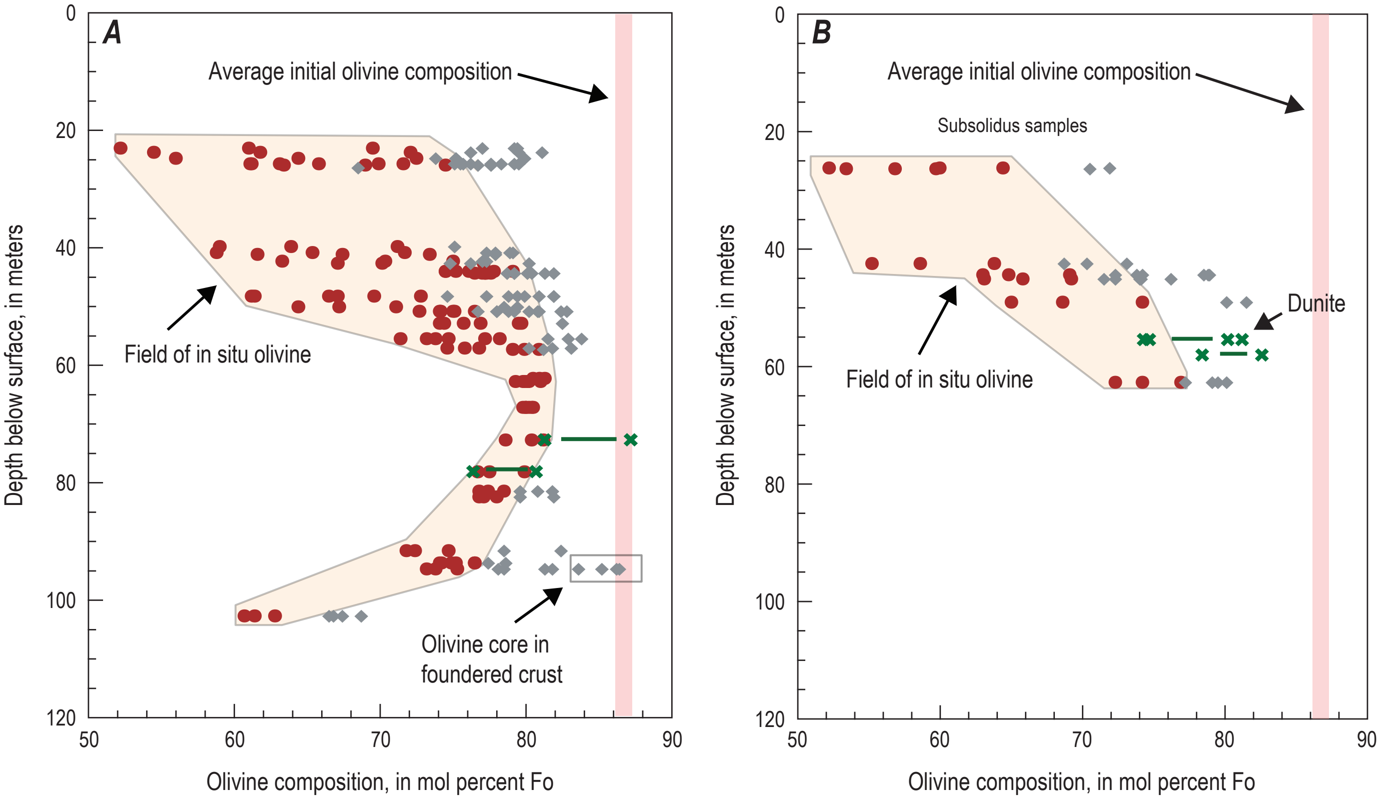


Figure 18. Graphs showing olivine compositions versus depth in Kīlauea Iki lava lake drill cores. *A*, Compositional ranges for all olivine in partly molten 1967 core, plus near- and subsolidus samples with similar bulk MgO content from 1975, 1981, and 1988, plotted against depth. The average and extreme in situ olivine compositions are shown by the red dots, highlighted by orange field. Gray diamonds are compositions of olivine phenocryst cores; green crosses show composition of megacrysts (green lines show ranges). The vertical pink band shows the original average olivine composition (Fo$_{86-87}$) for all Kīlauea Iki drill core. *B*, As in part A, but for subsolidus samples. The range of compositions in the dunite in sample KI81-1-181.5 is included here (same symbol as megacrysts), as it contains no glass within its boundaries.

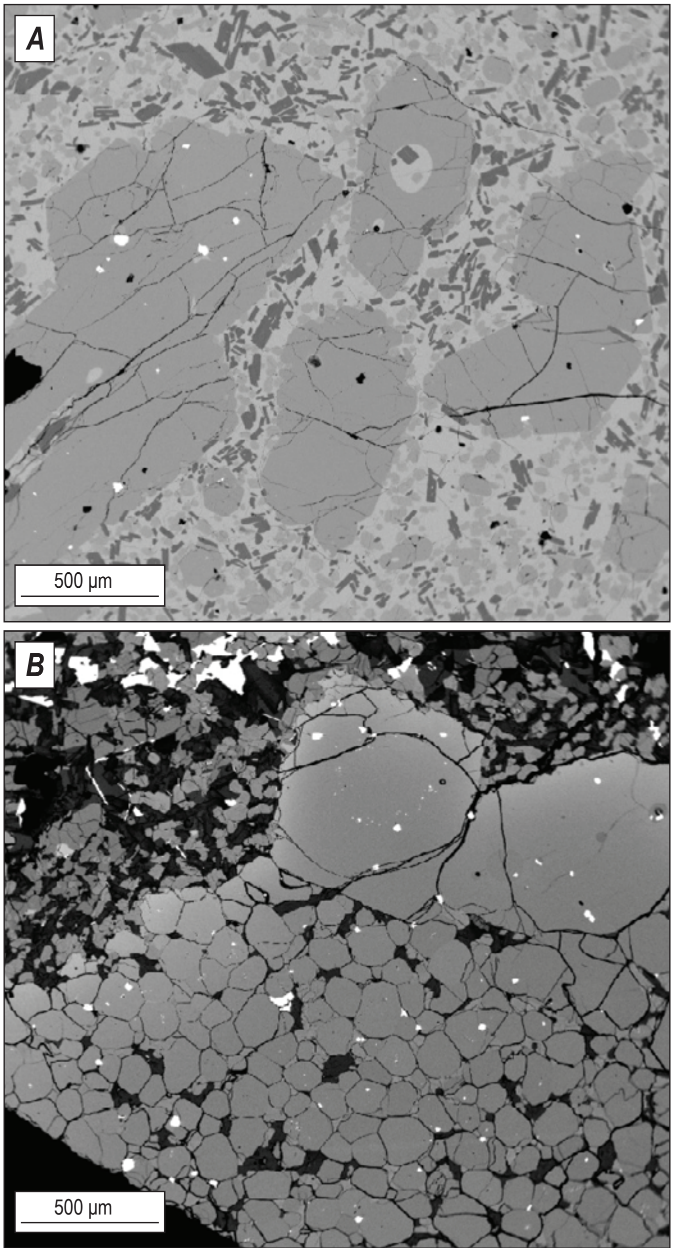


Figure 19. Back-scattered electron (BSE) images of samples from Kīlauea Iki lava lake cores. *A*, BSE image of olivine and groundmass in sample KI79-1-203.4, showing that olivine is virtually unzoned at this level in Kīlauea Iki lava lake. The total compositional range in this sample is 80.5–81.3 percent Fo, in crystals of all sizes. Other phases present include chromite (bright inclusions in olivine), melt (brighter than olivine), augite (small grains, slightly lighter than olivine), and plagioclase (dark laths). *B*, BSE image of the dunite inclusion in sample KI81-1-181.5, showing very slight zoning around the periphery, in contact with groundmass, and no discernible zoning within the dunite. The dunite also contains interstitial chrome spinel, augite, and plagioclase, but no melt.

samples from these depths. Much of this reequilibration took place before 1975, as the olivine in the vorbs, interpreted as derived from the cumulate zone, was already Fo$_{78–80}$ (Helz and others, 1989) in the 1975 core.

Average forsterite content of in situ olivine drops and its range tends to expand as core recovery temperature decreases below ~1,100 °C. Phenocryst core compositions begin to lag the compositional shift of the in situ olivine below 1,130 °C, as can be seen most clearly in the deepest samples (below 90 m) in figure 18*A*. This may reflect decreasing diffusion rates for Mg and Fe in olivine at lower temperatures (Chakraborty, 2010), as well as somewhat faster cooling rates below 1,100 °C.

Comparison of figures 18*A* and *B* show that in situ olivine compositions continue to shift to more Fe-rich compositions, and also that the range of compositions (buff field) becomes narrower. Phenocryst cores tend to lag, so that some samples have developed a second generation of olivine compositional zoning after passing through a stage of near-uniformity (compare KI79-1-203.6 and KI81-1-205.4 to KI88-2-205.2 in table 4).

This second-generation zoning can be seen clearly in the dunite inclusion in sample KI81-1-181.5 (shown in figs. 25.8a and b in Helz, 1987a). Olivine in the dunite is more magnesian, on average, than the in situ olivine (table 4) but less magnesian, than phenocryst cores in KI81-1-181.5. The dunite is now melt-free, so is plotted in figure 18*B*, but the presence of interstitial augite+plagioclase suggests that grain boundaries were formerly coated with melt; the melt films and the fine grain size of the olivine presumably facilitated reequilibration of the dunite, so that it reequilibrated more extensively than the phenocrysts. BSE imagery (fig. 19*B*) shows that the interior of the dunite is very uniform in composition, with the present zoning peripheral to the entire inclusion.

It is clear from the data presented here that, in order for original (1959) Fo$_{86–87}$ compositions to be preserved, the sample must be (1) shallow, hence rapidly cooled, and (or) (2) have unusually large olivine crystals. In deep (60–100 m) samples, few crystals retain 1959 compositions. Two that do (fig.18*A*) are the megacryst in sample KI81-1-238 and the coarsest (3×6 mm) phenocryst in KI88-2-310, which is a sample of foundered crust (Helz, 1993). The preservation of Fo$_{86–87}$ in the megacryst probably reflects its size, while the survival of Fo$_{86}$ in the foundered crust sample may be due to initial rapid cooling of the crust in the upper part of the lake prior to its sinking to its present depth of >95 m, as well as the sample's relatively coarse grain size.

Figure 20 shows compositional traverses and photomicrographs for three megacrysts from Kīlauea Iki. KI76-2-20.5 (not shown in figure 18*B*) is the shallowest (6.2 m) and most rapidly cooled, with a broad core of Fo$_{86–87}$; its very narrow rims track changes in melt composition to the point where the olivine was armored by groundmass phases. The megacryst in KI81-1-238 (also shown in fig. 18*A*) has shoulder regions that retain Fo$_{86-87}$, but the inner core is Fo$_{85}$, possibly because of reequilibration with large melt

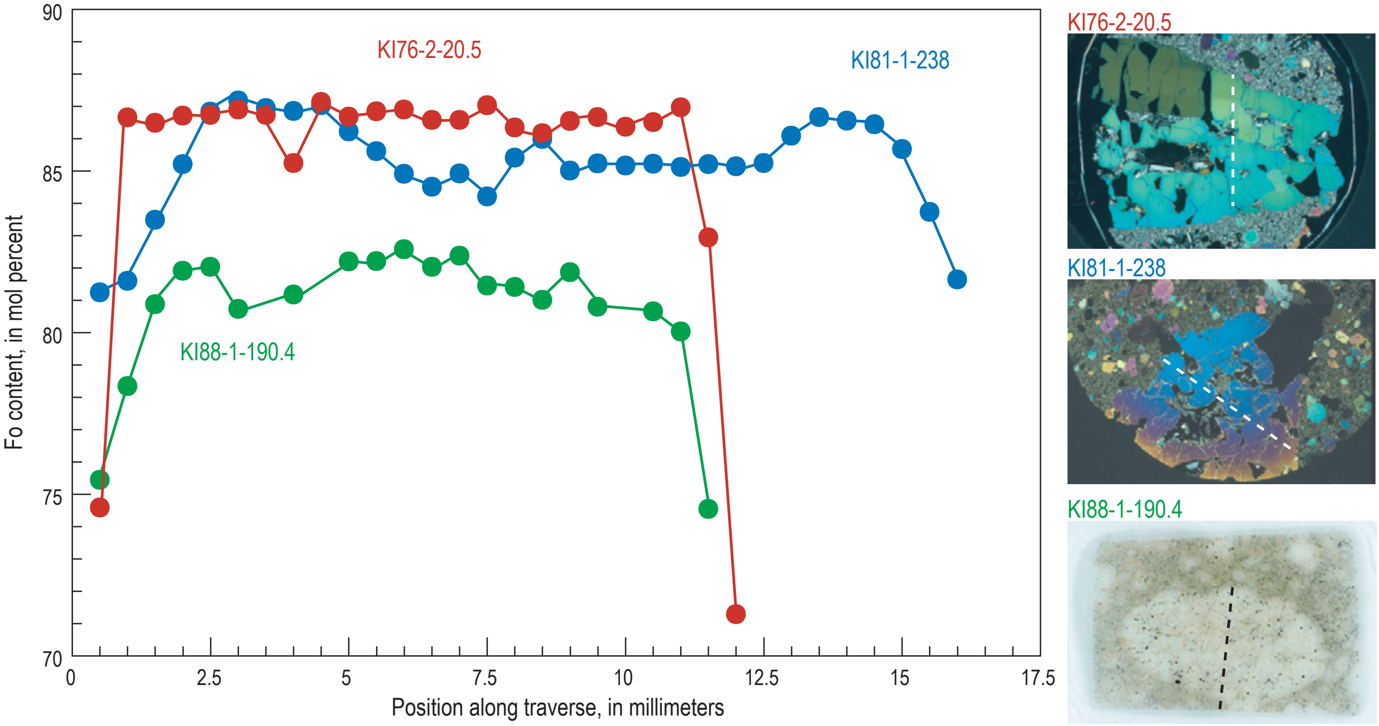


Figure 20. Compositional profiles showing forsterite (Fo) content along traverses in three olivine megacrysts from Kīlauea Iki lava lake. Traverses shown by dashed lines in accompanying photomicrographs. The shallowest megacryst KI76-2-20.5 (red) was cooled fastest; it reached the solidus (~1,000 °C) in 1961, cooled to ~100 °C by 1964, and was recovered in 1976. Sample KI81-1-238 (blue) was partly molten when recovered from the hot zone (T~1,130–1,135 °C) in 1981. Sample KI88-1-190.4 (green) reached the solidus in 1981, cooled to ~100 °C by 1986, and was recovered in 1988.

inclusions. It has wide rims, zoned out to $Fo_{81–82}$, in contact with the interstitial melt, similar to in situ olivine compositions observed in other very hot samples (fig. 18*A*; table 4). The third megacryst (KI88-1-190.4, also shown in figure 18*B*) is similar in composition to the phenocrysts in KI81-1-181.5 (table 4), so it is slightly more Mg-rich than the fine-grained dunite in that same sample (fig. 18*B*), perhaps because it is a single crystal rather than an aggregate. All three, however, retain compositions more magnesian than the in situ olivine in the same samples. By contrast, a fourth megacryst (in sample KI88-1-255), quenched from ~1,050 °C, has reequilibrated to the same composition as the in situ olivine, as shown in figure 18*A*. Evidently, even the largest olivine crystals reequilibrate to groundmass compositions if they are in contact with melt for 29 years.

The pattern of reequilibration of olivine in Kīlauea Iki compares closely with that described by Moore and Evans (1967) in the prehistoric Makaopuhi lava lake, although olivine compositions are more magnesian in Kīlauea Iki. The samples with the most uniform olivine in the prehistoric Makaopuhi lava lake were found within the olivine-rich zone, where phenocrysts reequilibrated to $Fo_{64.3–68.3}$ (Moore and Evans, 1967), with ranges of 1.5–3.3 mol percent Fo, similar to those from the hottest zone within Kīlauea Iki (table 3). Samples higher in the section showed progressively wider ranges of olivine compositions, as observed in Kīlauea Iki (fig. 18*A*,*B*). Similar extensive reequilibration of olivine from chilled margin to the core of the Uēkahuna laccolith was reported by Wilkinson and Hensel (1988).

An important difference between the two lava lakes is that groundmass olivine, ubiquitous in Kīlauea Iki, was found only in the uppermost 6 m of the Makaopuhi lake, being replaced by low-Ca pyroxene at greater depths; however, the occurrence of orthopyroxene oikocrysts is very similar in the two bodies in both location and habit. In Makaopuhi (Evans and Moore, 1968), poikilitic orthopyroxene is found in the olivine cumulate zone (40–67 m below the surface), as it is in Kīlauea Iki (table 4; also Helz, 1987b). The oikocrysts enclose plagioclase and Fe-Ti oxides without displacing them but appear to have completely consumed both olivine and augite as they grew (fig. 21; also Evans and Moore, 1968). In both bodies, the oikocrysts are better developed below the thermal maximum, probably because the cooling rate in the lower crust is lower than in the upper crust (Moore and Evans, 1967).

Summary and Future Directions

Lava lake studies, especially on closed-system lava lakes, have produced a wealth of information about how basaltic systems crystallize and fractionate. The list of fractionation processes observed or inferred is large (Wright and others, 1976; Wright and Okamura, 1977; Helz, 1987b, 2009), as is the range

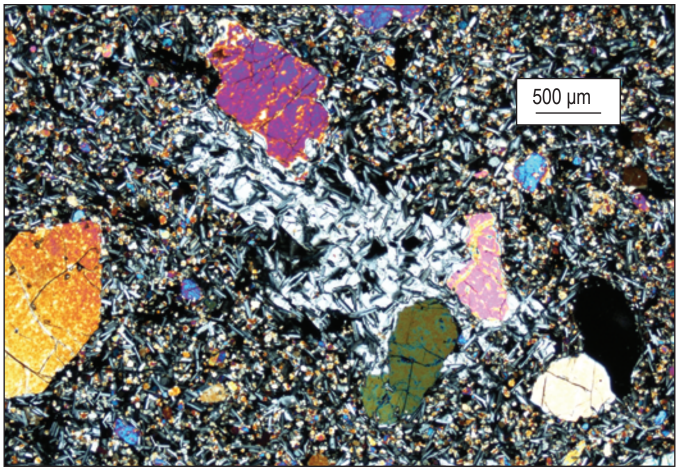


Figure 21. Photomicrograph showing orthopyroxene oikocryst (light gray) extending between several small olivine phenocrysts in sample KI81-1-306.7 (T=1,087 °C) from Kīlauea Iki lava lake. Nicols crossed. Note the near-absence of both olivine and augite inclusions within the area of the oikocryst.

of whole-rock and melt compositions analyzed. These results have been widely applied to (inter alia) the structure of mid-ocean ridge magma chambers (Barth and others, 1994) and the behavior of melts in marginal zones in magma chambers (Tait and Jaupart, 1996).

Results presented here that are relevant to the interpretation of basaltic systems include observations that (1) the actual efficiency of segregation/extraction of differentiated liquids is much lower than might be predicted from theoretical fractionation of bulk compositions; (2) the position of the last interstitial liquids (at Kīlauea Iki they are 80–100 m below the surface; see fig. 16*B*) is much deeper than the position of the most differentiated part of the lake (20–40 m; Helz, 2009), which solidified before 1975; (3) reequilibration of original olivine compositions from Fo_{86-87} to $\leq Fo_{80-81}$ is extensive after 15 years and essentially complete after 30 years of slow cooling; and (4) a shift in cooling rates (from 2–3 °C/year above 1,100 °C to 100°C/year below 1,100 °C) produces a distinct second generation of normal zoning in olivines from the most slowly cooled parts of the lava lake.

The use of samples from Kīlauea lava lakes for studies of trace elements and isotopes is just beginning. Trace element compositions of bulk samples usually mirror processes inferred from major elements (Wright and Peck, 1978; Wright and Helz, 1996; Helz, 2012) in Kīlauea lavas. Study of the platinum-group elements plus rhenium (Re) in Kīlauea Iki (Pitcher and others, 2009) documented their behavior in the low-pressure, high-temperature igneous environment, unmodified by subsequent (including hydrothermal) processes. One result is that Re is progressively depleted in later drill core relative to samples quenched earlier, apparently because of loss of Re to the volatile phase. Recent work by Teng and others (2008, 2011) and Sio and others (2011), on both bulk samples and individual crystals of olivine, has found significant fractionation of Fe and Mg isotopes in olivines from Kīlauea Iki, apparently produced during diffusive reequilibration, whereas Fabbrizio and others (2010) have documented the preservation of phosphorus zoning in otherwise reequilibrated, unzoned olivines from Kīlauea Iki (KI81-1) samples. With expanded use of microbeam analysis for trace elements and isotopes, it should be possible to exploit samples from the various historical lava lakes, which include cores containing melts quenched from temperatures ranging from 980 to 1,140 °C, for many studies in the future.

Other workers have taken advantage of the known cooling history and extensive sample collections from Kīlauea Iki and other lava lakes to investigate the evolution of mineral textures. Cashman (1993) used plagioclase data from the prehistoric Makaopuhi lava lake (Moore and Evans, 1967) to elucidate cooling rates in bodies of crystallizing basalt. Studies of Kīlauea Iki samples include crystal size distribution analysis of olivine (Vinet and Higgins, 2011) and variation in the intercrystalline angle at plagioclase-pyroxene-plagioclase contacts (Holness and others, 2012a,b), also proposed as a means of estimating cooling rates.

Studies of Submarine Hawaiian Lavas

The collection of lava samples from the submarine slopes of Hawai‘i began in the early 1960s (Moore, 1965; Moore and Fiske, 1969). Such submarine erupted samples, quenched under high water pressure, preserve melt compositions as glass and are natural analogues to hand-dipped molten samples collected and quenched during eruptions at Kīlauea and Mauna Loa. These early samples from Kīlauea, Mauna Loa, and Mauna Kea were incorporated into many subsequent studies on the petrology and geology of Hawai‘i. An important attribute of submarine erupted and quenched basaltic glasses is that volatile components that would be lost to the atmosphere during subaerial eruption are largely trapped in solution in the glass and can be analyzed to determine preeruptive volatile contents. One of the most striking discoveries from study of submarine lavas was that Lō‘ihi Seamount is an active Hawaiian volcano, likely to become the next Hawaiian island (Moore and others, 1982). Equally striking was the discovery that Lō‘ihi lavas included alkalic basalts, as well as the expected tholeiitic shield lavas.

Preshield Stage Lavas

Rocks recovered from Lō‘ihi Seamount in 1981 included a range of alkalic lavas, such as basanite, alkalic basalt, and hawaiite (Moore and others, 1982), some of which contained mantle xenoliths (Clague, 1988). Thicker palagonite alteration of the alkalic glasses suggested they were generally older than the tholeiitic lavas and led to the conclusion that Lō‘ihi Seamount, in particular, and Hawaiian and detailed petrologic studies indicate that the composition of Lōʻihi's rocks represents one of the several mantle components currently thought to contribute to generation of Hawaiian magmas (see, for example, Dixon and Clague, 2001; Huang and others, 2009). The discovery of early alkalic magmas constrains melting models, such that growth of Hawaiian volcanoes begins and ends with smaller-degree alkalic melts (Clague, 1987b; Clague and Dalrymple, 1987) separated from more enriched mantle source rocks at greater depth during the pre- and postshield alkalic stages, compared with the larger-degree tholeiitic melts that form the voluminous shield stage, separated at shallower depths from less enriched mantle sources (for example, DePaolo and Stolper, 1996; Xu and others, 2007).

The high $^3He/^4He$ isotopic ratios and other rare-gas compositions of Lōʻihi (and some Kīlauea) lavas (for example, Kurz and others, 1983; Honda and others, 1991, 1993; Trieloff and others, 2000, 2001) indicate an ancient, undegassed mantle source and have been a primary argument for the deep-mantle source of the plume, although this view is not uncontested (see, for example, Anderson, 1998). Neon isotopes have been proposed to have a solar (Honda and others, 1991, 1993, Ballentine and others, 2001) or a meteoritic (Trieloff and others, 2000, 2001) signature, indicating an ancient source undegassed during Earth's history. In contrast, analyses of Lōʻihi lavas for CO_2, H_2O, Cl, and S (Dixon and Clague, 2001) show that the Lōʻihi mantle component is relatively dry and therefore derived from a recycled mantle component that preferentially lost H_2O during subduction. This conclusion seems to contradict that based on rare gas components, but it may indicate that the rare gases are derived mainly from one mantle component, whereas water is mainly derived from another.

The Early Growth of Kīlauea

Lōʻihi Seamount was the only known example of a preshield-stage volcano until 1998 to 2002, when a series of dive programs, using the Japan Agency for Marine Earth Science and Technology (JAMSTEC) remotely operated vehicle *Kaiko* and manned submersible *Shinkai 6500*, explored the deep rift zones and flanks of a number of Hawaiian volcanoes. These dives discovered extensive preshield-stage alkalic products from early Kīlauea (Lipman and others, 2002; Sisson and others, 2002; Sisson, 2003; Coombs and others, 2004, 2006; Calvert and Lanphere, 2006; Kimura and others, 2006; Hanyu and others, 2010), as well as small exposures of alkalic material from early Hualālai (Hammer and others, 2006).

Because of the Hawaiian Volcano Observatory's location on Kīlauea and emphasis on shield-stage Kīlauea studies, the long-term development of Kīlauea is of particular relevance. Early Kīlauea eruptive products are exposed in the submarine region due south of the volcano's summit. This region consists of a south-facing slope that descends steeply from the shoreline to a broad mid-slope "Hilina" bench at ~2,500–3,000 m depth, fronted by another south-facing scarp that descends steeply to the seafloor at ~5,000-m depth. The submarine Hilina bench is elongate northeast-southwest, parallel to the shoreline, with maximum dimensions of 15 by 55 km. It is bounded on the southwest by Papaʻu Seamount and tapers to the northeast, where it merges with the Puna Ridge. Because of its flat top and orientation parallel to the shoreline, the Hilina bench was initially interpreted as a prodigious rotational landslide block that dropped ~10 percent of Kīlauea's subaerial tholeiitic shield to great water depths along normal faults similar to those of the subaerial Hilina Pali (Lipman and others, 1985; Moore and others, 1989; Smith and others, 1999). Initial JAMSTEC dive observations and recovered samples, however, revealed that the frontal scarp exposes bedded sandstones and debris-flow breccias (Lipman and others, 2002), consistent with seismic reflection profiles (Hills and others, 2002) that show the Hilina bench to be a fold and thrust wedge of marine volcaniclastic rocks being extruded from beneath growing Kīlauea as a result of edifice spreading. The upper slope, above the bench and toward the shoreline, is widely mantled by unconsolidated to weakly consolidated glassy sands derived by shattering of shoreline-crossing lava flows, but narrow rock ribs expose underlying in-place pillow lavas incised in landslide headscarps.

The oldest preshield Kīlauea products are exposed along the 3,000–5,000-m deep frontal scarp of the Hilina bench, and Kīlauea's submarine products become progressively less alkaline and younger, overall, both at shallower water depths and northeastward toward the Puna Ridge. Bedded sedimentary rocks of the frontal scarp are mainly volcanic-glass sandstones, with glass grains dominated by degassed, relatively Si-rich, and Na- and Ti-poor tholeiitic compositions, similar to modern products of Mauna Loa, but accompanied by subordinate populations of alkalic glasses, most with elevated S, CO_2, and H_2O concentrations indicative of submarine eruption (Sisson and others, 2002, 2009; Coombs and others, 2006). Similarly, clasts in bedded debris-flow breccias of the frontal scarp are Mauna Loa-type tholeiites accompanied by alkalic clasts, rarely with preserved glass that also has elevated volatile concentrations. Alkalic glasses encompass basanites, nephelinites, hawaiites, alkali basalts, tephriphonolites, phonotephrites, benmoreites, and mugearites (Sisson and others, 2009). Alkalic clasts encompass alkali basalts, hawaiites, and basanites; a probable sill or thick lava flow of phlogopite nephelinite was also discovered. Judging from the wide range of eruption depths interpreted from dissolved volatiles, Kīlauea began as diverse alkalic vents spread across the south flank of the extant and active Mauna Loa shield. The sedimentary rocks of the frontal scarp are composed of fragmentation products shed mainly from shattering of shoreline-crossing Mauna Loa lavas and also from shattering, landsliding, and explosive eruptions of shallow-to-deep marine early Kīlauea vents (Coombs and others, 2006). Minor amounts of low-Si tholeiitic glasses with elevated S concentrations accompany the alkalic and degassed Mauna Loa products, possibly indicating that transitionally tholeiitic magmas appeared early in Kīlauea's tholeiitic glasses with elevated S concentrations accompany the alkalic and degassed Mauna Loa products, possibly indicating that transitionally tholeiitic magmas appeared early in Kīlauea's history. Alkalic samples from the frontal scarp have $^{40}Ar/^{39}Ar$ ages spanning from 280±20 ka to 212±38 ka, including higher precision results on phlogopite from nephelinite of 234±9 and 236±10 ka (Calvert and Lanphere, 2006). These results support an inception age of ~300 ka for Kīlauea.

A narrow rock rib above the midslope bench, directly toward Kīlauea's summit, consists of in-place weakly alkalic and transitional basaltic pillow lavas (Kimura and others, 2006). $^{40}Ar/^{39}Ar$ ages of in-place, weakly alkalic pillow lavas from the upper slope span from 195±48 ka to 138±30 ka (Calvert and Lanphere, 2006), and two transitional basalt clasts shed from the upper slope onto the midslope bench as landslide debris have ages as young as 67±29 ka and 65±28 ka (Hanyu and others, 2010). To the northeast, where the midslope bench gives way to the Puna Ridge, lava samples are tholeiitic, but their glasses are notably lower in Si and higher in Ti and Na than is typical of the subaerial Kīlauea shield. $^{40}Ar/^{39}Ar$ ages measured on these low-Si tholeiites (transitional basalts of Sisson and others, 2002) are imprecise because of low K concentrations, but are 228±114 and 138±115 ka (Calvert and Lanphere, 2006). The location of these low-Si tholeiites toward the Puna Ridge, and the absence of alkalic materials from that region, suggest that they record an intermediate stage in Kīlauea's development commencing at roughly 150–100 ka. At that time, volcanic output increased markedly, initiating growth of the East Rift Zone and the Puna Ridge and ending the period of alkalic magmatism, during which the diversity and degree of undersaturation of alkalic products diminished as the volcano matured.

Subsequently, as volcanic output increased markedly, magma compositions came to be dominated by low-Si tholeiites. Cessation of this low-Si tholeiitic intermediate growth period was probably gradual, giving way to tholeiites typical of Kīlauea's subaerial shield, as is shown by the presence of low-Si tholeiitic lavas among ordinary tholeiites exposed in fault scarps of the lower subaerial south flank, estimated to be close to 25 ka in age (Chen and others, 1996), and the local presence of similar relatively high-Ti and high-K glasses (440–640 C.E.; Fiske and others, 2009) in the Kulanaokuaiki Tephra. Although the record of Kīlauea's development is complicated by posteruptive sedimentation, concurrent deposition of products from Mauna Loa, deformation resulting from island spreading, and incomplete exposures, it is the only Hawaiian volcano where products of all stages, from inception to mature shield stage, are accessible and have been dated by radiometric dating.

Occurrence and Characteristics of Submarine Picritic Lavas

The deep submarine rift zones of Hawaiian volcanoes consist of common to abundant picritic tholeiites, with less common aphyric or sparsely phyric lavas (Moore, 1965; Garcia and others, 1995b). Clague and others (1995b) studied Kīlauea samples reported in Moore (1965), plus more recently acquired samples, and found that all the lavas had fractionated glasses with MgO<7 weight percent, while containing abundant olivine and less abundant augite, plagioclase, and orthopyroxene. Many of the olivine crystals are either more primitive or more evolved than crystals in equilibrium with the host melts; on average, the data parallel an olivine control line of $Fo_{87.9}$, which is more magnesian than the $Fo_{86–87}$ observed in Kīlauea's subaerial lavas. Clague and others (1995b; and unpublished data) find 13 of 47 sampled submarine sites have rocks with whole-rock MgO>15 weight percent, so picritic lavas are 25–30 percent of the sampled flows and much more abundant than picritic lavas on land. The glasses are generally highly fractionated and contain between 4.3 and 7 percent MgO, with a fairly even distribution from 4.8 to 6.5 percent and a few samples with higher and lower values. Another suite of 65 lava rind glasses recovered along the submarine rift (Johnson and others, 2002) includes four with MgO>7 weight percent (one at 7.2 and three from the same dredge at 9.2 weight percent MgO), but almost all are between 4.2 and 6.7 weight percent MgO. Thirty-five other tholeiitic glasses from the Puna Ridge and south flank of Kīlauea range from 5.3–7.1 weight percent MgO (Coombs and others, 2006). The combination of glass, mineral, and whole-rock compositions was interpreted by Clague and others (1995b) as having been formed by a complex series of processes including spinel+olivine and then multiphase (olivine+plagioclase+augite) fractionation, degassing, wall-rock stoping and assimilation (Kent and others, 1999), magma mixing in the crustal reservoir, entrainment of olivine xenocrysts from a hot ductile olivine cumulate body (Clague and Denlinger, 1994), and disruption of gabbroic wall rocks in the rift zone. The rocks may thus be picritic, but the host melts are almost all highly fractionated and have undergone a complex evolution.

On Mauna Loa, 54 percent of 46 samples from the submarine south rift are picritic tholeiites containing >15 percent olivine crystals (Garcia and others, 1995b). The host lavas contain 5.8 to 8.4 weight percent MgO, while olivine compositions vary from Fo_{81} to Fo_{91}. Moore and Clague (1992) report only 7 of 19 whole-rock pillow samples from this rift as having >15 weight percent MgO, and the host glasses of 20 samples range from 4.5 to 8.0 weight percent MgO. Seventeen of these 20 samples have glass MgO between 5.5 and 6.9 weight percent. Wanless and others (2006a,b) report compositions for lavas from radial vents on the submarine west flank of Mauna Loa, but only 1 of 9 samples contains >15 weight percent MgO, with the remainder having whole-rock MgO contents of 6.0–7.6 percent. Morgan and others (2007) report MgO contents of 4.9–7.8 weight percent MgO, with 90 percent between 5.4 and 6.9 weight percent MgO (median of about 6.3 weight percent MgO) in 259 glass grains and pillow rinds from the southwest flank of Mauna Loa. Only 5 of 54 whole-rock analyses contain >12 weight percent MgO, and none have more than the 15 weight percent MgO that would be considered picritic. Thus, although common to abundant picritic lavas occur along Mauna Loa's submarine rift zone, most have host melts that are highly fractionated and have undergone a complex evolution similar to those on Kīlauea's Puna Ridge.

On Lōʻihi Seamount, nine samples from the lower part of its south rift zone (depth 4,976–5,058 m) are picritic tholeiites, with all whole-rock analyses having >20 weight percent MgO (Matveenkov and Sorokhtin, 1998). Tholeiitic samples, too numerous to count, from the summit, upper south rift zone, and east flank are only rarely picritic (see, for example, Moore and others, 1982; Frey and Clague, 1983; Hawkins and Melchior, 1983; Garcia and others, 1993, 1995b, 1998; Clague and others, 2000, 2003; Dixon and Clague, 2001).

The submarine part of Hualālai's northwest rift zone is also characterized by abundant picritic lavas (Clague, 1987b; Moore and Clague, 1992). Fifteen of 22 whole-rock analyses have MgO>15 weight percent, and 11 of those have MgO=20–31.1 weight percent. The host glasses contain 5.9 to 8.0 weight percent MgO (median about 6.8 weight percent), while the bulk compositions define an olivine control line of ~Fo_{89}. Fully one-third of the glasses have >7.2 weight percent MgO, compared with only 9 percent of glasses from Mauna Loa and none from Kīlauea's Puna Ridge. Tholeiitic samples from the submarine west flank of Hualālai (Hammer and others, 2006; Lipman and Coombs, 2006) include picritic lavas, but the glasses in rinds and fragments range from 6.0 to 8.8 weight percent MgO, with a median of about 6.5 weight percent MgO for the picritic pillow lavas from *Kaiko* dives 218 and 219 (Lipman and Coombs, 2006), and a median of about 7.15 weight percent MgO for the glass grains in rocks from *Shinkai* dives 690 and 692 (Hammer and others, 2006).

Although picritic lavas are common to abundant along the submarine rifts of Kīlauea, Mauna Loa, Hualālai, and Lōʻihi, none represent primary or near-primary melt compositions, all indicating, instead, strongly to moderately fractionated melts with variable amounts of added olivine, mostly as deformed xenocrysts. Picritic lavas from these several volcanoes differ in two ways: (1) the ranges of MgO contents of the host melts differ, with Kīlauea having the most fractionated host melts and Hualālai having the least fractionated ones, and (2) the amount of olivine added to these host melts also varies, with Hualālai and Lōʻihi having abundant olivine xenocrysts added to the melts and Kīlauea having the least added olivine.

Glasses and Volatiles in Samples from Kīlauea's Submarine East Rift Zone

Sand grains recovered in a box core north of the Puna Ridge included a spectrum of glasses with MgO contents as high as 14.7–15.0 weight percent (Clague and others, 1991, 1995b; Wagner and others, 1998). These glasses are Kīlauea-like and allow reconstruction of primary melt compositions by adding in small amounts of olivine, with minimal uncertainty due to fractionation, mixing, or assimilation of wall rock.

The primary melts are estimated to contain 13.4 to 18.4 weight percent MgO, depending on their FeO contents; the volatile contents for a primary melt with 16.4 weight percent MgO are estimated to be 1,000 ppm S, 100 ppm Cl, and 0.37 weight percent H_2O (Clague and others, 1995b). The melts have calculated liquidus temperatures of 1,283 °C to 1,384 °C, with an average of 1,346 °C, and very low viscosities (0.3–0.6 Pa·s), so olivine would sink rapidly in these melts. They also have densities slightly greater (2.66–2.68 g/cm^3) than more fractionated melts, so the magma reservoir should be stably stratified until the uppermost magma degasses; if no crystallization occurred, this would increase melt density and could cause turnover in the reservoir (Clague and Dixon, 2000).

Additional high-MgO glass grains with as much as 11 weight percent MgO were found in Kīlauea beach sand, recovered in the Hilo drill core (Beeson and others, 1996), and also in several ash deposits on Kīlauea (as much as 14.5 weight percent MgO; Clague and others, 1995a; Helz and others, 2014). These latter glasses were erupted subaerially or in shallow water and, unlike the deep submarine ones, have lost most of their volatiles. Lastly, extensive sampling along the Puna Ridge (Johnson and others, 2002) recovered glasses with as much as 9.2–9.3 weight percent MgO; these were from the same shallow cone as three pillow fragments from picrites (glass MgO=9.2 weight percent) that were mentioned above.

Measurements of magmatic H_2O and S in Hawaiian submarine lavas (some of the earliest such studies were Moore and Fabbi, 1971; Killingsley and Meunow, 1975; Meunow and others, 1979) yield varying results. Dixon and others (1991), using many of the samples from Kīlauea dredged by Jim Moore in 1962, found variable H_2O and S contents. They attributed the variability to mixing of degassed and undegassed magmas, followed by equilibration of CO_2 in the hybrid at the eruption pressure. The degassed end member required gas loss at very low pressures, perhaps even subaerial conditions. Such recycling of degassed lava back into the magma chamber, where it might mix with undegassed magma, was observed during the 1959 Kīlauea Iki eruption, which had repeated cycles of high fountains and drainback (Richter and others, 1970; Eaton and others, 1987). One component of the 1959 eruption is known to have traveled out Kīlauea's East Rift Zone at least as far as Puna, where it mixed with stored rift lava, and then was erupted in the latest 1960 lavas (Wright and Helz, 1996). It and earlier magmas could well have traveled farther, to erupt along the submarine east rift.

Kīlauea and Mauna Loa: Insights into Primary Magma Compositions and Sources

Because of their isolation from interaction with continental crust, Hawaiian volcanoes, and especially Kīlauea and Mauna Loa, provide a direct look at basaltic magmas and their mantle sources. Similarly, the relative simplicity of each volcanic edifice has allowed a relatively clear look at magmatic plumbing systems. This final section will first summarize petrologic constraints on primary (or parental) magma compositions, the nature of the mantle source(s), and the relationship between compositions of Kīlauea and Mauna Loa, both shield-stage and preshield. Then we will consider the nature of the currently active volcanic plumbing system and review the constraints imposed by petrologic data on magma storage, especially for the summit reservoirs of Kīlauea and Mauna Loa.

Primary Magmas and Sources for Shield-Stage Lavas

One of the earliest papers to grapple with the subject of primary melt compositions, coupled with the need for mantle sources capable of producing large volumes of basaltic melt at closely spaced volcanoes, is Murata's (1970). Given his assumption that primary Hawaiian magma had ~20 weight percent MgO, to be in equilibrium with peridotite, and given the 37±3 km spacing between the volcanoes, he estimated that Mauna Loa would need a source ranging from 60 to 125 km in depth. In view of the limited range of composition of olivine phenocrysts observed at Kīlauea and Mauna Loa, however, he suggested that this primary liquid would rise quickly, crystallizing significant olivine only in the summit reservoir. Mauna Loa basalts have higher SiO_2 contents at a given MgO content (Tilley and Scoon, 1961; Wright, 1971), which led Murata (1970) to suggest that their source was richer in orthopyroxene than that of Kīlauea's lavas. Work of Clague and others (1991, 1995b), discussed above, has documented the existence of glasses with MgO contents as high as 15 weight percent, present as fragments in dredge samples from the submarine part of Kīlauea's East Rift Zone. How this material leaks to the surface is not clear, but, even in trace amounts, it provides direct evidence that melts in Kīlauea's plumbing may be highly magnesian, though not quite as magnesian as Murata (1970) suggested.

Wright (1984) revisited the question of the origin of Hawaiian tholeiite, again with a particular focus on Kīlauea and Mauna Loa. This approach, which involved modeling rare earth elements (REE) as well as major elements, established the need for a complex source, in which the large melt fractions (as much as 42 percent of the hybrid peridotite source) were supplemented by varying amounts of nephelinitic liquid plus amphibole, apatite, and ilmenite. The nephelinite provided the REE signature for residual garnet, a feature of both Kīlauea and Mauna Loa basalts (Hoffman and Wright, 1979; Leeman and others, 1980). Melting and melt segregation were considered to take place over a range of pressures. A dynamic model, with fresh mantle moving beneath the volcanoes, reduced the thickness of the melted zone needed from Murata's (1970) estimate of 65 km to ~27 km (Wright, 1984). In contrast to Murata (1970), Wright estimated the parental compositions of Kīlauea and Mauna Loa basalts as having 13–14 weight percent MgO, with olivine fractionating during rise of magma from the mantle to the base of the summit reservoirs. This is consistent with data of Helz and others (2014), who have found rare subaerial glasses in the Pāhala Ash at Kīlauea with as much as 14.5 weight percent MgO.

More recent models also hypothesize complex mantle sources for Hawaiian shield-stage tholeiites. Hauri (1996) suggested that oceanic crust is a significant component, especially prominent in the Loa-line volcanoes, from Ko'olau onward to Mauna Loa. The model offers no explanation for why this component is so much lower in the Kea-line volcanoes (such as Kīlauea), and the isotopic data from Lō'ihi (a Loa-line volcano) require still further complexities. More recently, Sobolev and others (2005), although acknowledging that the mantle is dominantly peridotite, suggested that the source for Hawaiian tholeiite is secondary pyroxenite produced by reaction between mantle peridotite and recycled oceanic crust. A new study by Putirka and others (2011), however, suggests that the Ni content of olivines in Hawaiian picrites requires a peridotitic source. Their proposed parental melts for Kīlauea and Mauna Loa have MgO=18 weight percent (versus 20 weight percent assumed by Murata, 1970, and 13–14 weight percent assumed by Wright, 1984).

Commonalities among the various models include (1) that parental melts are highly magnesian, although rarely seen at the surface, whether the source is strictly peridotitic or not; (2) that there is much olivine removal as the magmas ascend higher in the volcanic plumbing (resulting in a gravity high associated with the magma conduit; see, for example, Wright and Klein, 2006); and (3) that separate sources (or different mixes of the various components) are clearly needed for volcanoes along the Loa and Kea lines.

Implications of Isotopic Signatures of Preshield Lavas for the Origins of Hawaiian Magmas

To a first-order approximation, the succession of lava types and erupted volumes at Hawaiian volcanoes, from early alkalic to voluminous shield-stage tholeiitic to late alkalic, can be interpreted as an increase and then a decrease in extent of melting as each center passes over some portion of the leading edge, interior, and trailing margin of the upwelling source region (Moore and others, 1982; Clague, 1987a; Clague and Dalrymple, 1987). Isotopic measurements show, however, that the types of materials undergoing melting also change with the stages of volcano development. Isotopes of Pb, Sr, Nd, Hf, and He can be interpreted in terms of four general end-member sources for Hawaiian magmas (see summary in Hanyu and others, 2010). These end members are represented most purely by the early and late products of the volcanic systems, and, notably, the sources for early magmas are not the same as those for the latest, suggesting an isotopically zoned and (or) mixed mantle source.

Inception-stage products are currently well represented by abundant samples and isotopic measurements from Lōʻihi and offshore Kīlauea (Kurz and others, 1983; Staudigel and others, 1984; Stille and others, 1986; Kimura and others, 2006; Hanyu and others, 2010). Most early Kīlauea magmas are distinguished by their high and uniform values of $^{206}Pb/^{204}Pb$, defining what has been named the "Kea" or "Hilina" end-member source component. Lōʻihi's magmas do not have as high $^{206}Pb/^{204}Pb$ but are relatively high in $^{208}Pb/^{204}Pb$ for Hawaiʻi, defining the "Lōʻihi" end-member source component. An important further distinction is that Kea end-member samples from early Kīlauea have moderate values of $^{3}He/^{4}He$ (10–15 times the present atmospheric ratio [Ra]), whereas $^{3}He/^{4}He$ in the Lōʻihi end member is characteristically high (to >30 Ra). A "depleted" end-member component, similar isotopically to many mid-ocean ridge basalts (low $^{87}Sr/^{86}Sr$, high $^{143}Nd/^{144}Nd$), is sampled most purely by low-volume alkalic magmas that erupt well after shield-building, as well as on the crest of the Hawaiian Arch (Chen and Frey, 1983, 1985; West and others, 1987; Frey and others, 2000). The fourth, "Koʻolau," end-member component is represented by anomalously high-SiO_2 basalts or basaltic andesites that erupted late in the shield stages of Koʻolau and Lānaʻi and that have high $^{87}Sr/^{86}Sr$, but low $^{206}Pb/^{204}Pb$ and $^{143}Nd/^{144}Nd$ (Roden and others, 1994).

The progression in isotopic values of Hawaiian magmas gives insights into possible configurations of the region undergoing melting. Inception-stage magmas are dominated either by the Lōʻihi or Kea components, whereas shield-stage tholeiites have isotopic values intermediate between those and the depleted or Koʻolau components. There is also a gross geographic division in which tholeiitic shields on the northeast side of the Hawaiian Island chain have the Kea component as a persistent contributor, while shields on the southwest (Loa) side receive contributions from the Lōʻihi component (Tatsumoto, 1978). The origin of this bilateral division is unknown, but it is consistent with the relative geographic positions of early Kīlauea and Lōʻihi. An apparent paradox is that the inception-stage magmas of Kīlauea and Lōʻihi are chemically heterogeneous, encompassing diverse alkalic and transitional compositions, but tend to be isotopically uniform within each locality, whereas the shield-stage tholeiites can be chemically monotonous but are isotopically quite variable, with contributions from the depleted and Koʻolau components that do not appear in the inception-stage products. Explanations for this paradox have postulated (1) that the Hawaiian magma source is compositionally zoned with the Lōʻihi and Kea components localized to the outer or leading margin, the Ko'olau component concentrated in the interior (Takahashi and Nakajima, 2002), and the depleted component either also in the interior or representing ambient upper mantle melted by the Hawaiian source, (2) that the Lōʻihi and Kea components are readily melted materials widely distributed in the Hawaiian magmatic source region within a matrix of more refractory Koʻolau and possibly depleted components that are only capable of melting in the central region where temperatures are high and upwelling is vigorous (Kimura and others, 2006; Hanyu and others, 2010), or (3) a counterintuitive view that inception-stage components are concentrated in the interior of the upwelling source but appear most purely in inception-stage magmas because those components melt deep and early, whereas marginal, more refractory, components only melt with greater degrees of upwelling where the source has also been dragged "downstream" by motion of the overriding Pacific Plate (DePaolo and Stolper, 1996; DePaolo and others, 2001).

An issue with interpretations that would place the Lōʻihi component, in particular, at the outer edge of the Hawaiian magmatic source is that this component is typified by high $^{3}He/^{4}He$. ^{4}He is an alpha-decay product of heavier radioactive elements, mainly uranium and thorium, whereas ^{3}He is not, so if a region has undergone degassing early in Earth's history, it will have low He/U or He/Th and subsequent radioactive ingrowth of ^{4}He will lead to a low $^{3}He/^{4}He$ value. As an example, depleted mid-ocean ridge basalts have a $^{3}He/^{4}He$ ratio of 8±1 Ra (Hilton and Porcelli, 2005). Conversely, regions that escaped early degassing should have high $^{3}He/^{4}He$ values. It has been interpreted, therefore, that the high $^{3}He/^{4}He$ Lōʻihi component originates from portions of the mantle that escaped degassing, or were less degassed than average, and that this source must reside in the deep mantle (Kurz and others, 1983). Dynamic models of mantle upwelling indicate that deeper materials end up in the interior of an upwelling zone or plume with shallower materials entrained along the margins (Hauri and others, 1994), so a simple expectation would be that the Lōʻihi component would occupy the interior of the ascending Hawaiian mantle plume (DePaolo and Stolper, 1996). The interior of the plume would be the hottest and undergo the greatest decompression, so another expectation is that it would melt to the greatest amount, in which case the Lōʻihi component might dominate the culminating shield-stage, not the inception-stage, of volcanism. One proposed resolution is that the He becomes mobile very early and deep in the melting process, possibly owing to high CO_2 contents (Gerlach and others, 2002), and that the upwelling source is inclined as a result of shear by the overriding Pacific Plate, so that initial melts carrying the He signal from the interior of the source arrive early, supplying inception-stage volcanoes, whereas most of the melting of the rising source takes place at shallower levels and displaced laterally to a position beneath the shield-stage volcanoes (DePaolo and Stolper, 1996; DePaolo and others, 2001). An alternative resolution is the experimental discovery (Parman and others, 2005) that, during partial melting at elevated pressures, He is retained in olivine more readily than is U; therefore, in regions that have undergone high-pressure melting, melt loss may have elevated He/U, resulting in high values of $^{3}He/^{4}He$ over time. Under this interpretation, the high $^{3}He/^{4}He$ of the Lōʻihi component allows for an ancient high-pressure melting event, consistent with the anomalously low H_2O concentrations of Lōʻihi magmas (Dixon and Clague, 2001) and the low He concentrations of high $^{3}He/^{4}He$ magmas worldwide (Anderson, 1998). Although the Lōʻihi source appears to be nonpristine, and diametrically different proposals for the spatial configuration of Hawaiian source components remain unresolved, the general association of high $^3He/^4He$ and ocean-island volcanoes supports the interpretation that their sources are mainly deeper than those that supply the mid-ocean ridge system (Hart and others, 1992).

Petrologic Constraints on Summit Reservoirs at Kīlauea and Mauna Loa

This chapter has reviewed petrologic features of tholeiitic shield-stage magmas stored in, and erupted from, the summit reservoirs of Kīlauea and, to a lesser extent, of Mauna Loa. Here, these observations will be used to constrain a model of the summit reservoir systems, with particular emphasis on Kīlauea. Critical observations include:

(1) Individual magma batches at Kīlauea succeeded each other at intervals of 2–7 years from 1952 to 1982 (table 1), with residence times of the batches within the volcano (where definable) of ~10 years (table 1; see also Wright and Klein, 2010). Pietruszka and Garcia (1999a) and Pietruszka and others (2001) have shown that trace element and isotope compositions also vary for the entire historical period at Kīlauea (1790 to the present), although not in parallel with batches defined by major-element chemistry.

(2) In contrast to the magma batches, the residence time for crystals, as determined by Mangan (1990) for olivine in the 1959 summit eruption, is one decade, and Pietruzka and Garcia (1999b) estimate magma storage times of several decades. The time interval between batches is thus shorter than the lifetime of each batch. Also, olivine may be retained in the reservoir somewhat longer than are the individual melt batches.

(3) Variation in the composition of melts erupted effusively within Kīlauea's caldera has been subtle from 1912 to the present, with melts becoming slightly more magnesian (and, hence, slightly hotter, per Helz and Thornber, 1987) from 1924 on (fig. 3). These melts, which are almost crystal-free and lie at the end of the olivine-control line (~7 weight percent MgO), presumably represent the magma at the top of the summit reservoir.

(4) The bulk composition of olivine in the summit reservoir is $Fo_{86–87}$ at both Kīlauea and Mauna Loa, and all historical summit lavas exhibit olivine control (Wright, 1971; Rhodes and Hart, 1995). The results of Sparks (1990) show olivine control at $Fo_{86.5}$ for young but pre-1790 subaerial Mauna Loa lava, while figure 11 shows that the same is true for Kīlauea's Uēkahuna laccolith. The laccolith, which predates the formation of Kīlauea's caldera at 1470–1490 C.E. (Swanson and others, 2012), provides evidence that olivine from Kīlauea's summit reservoir has also had an average composition of $Fo_{86–87}$ for many centuries.

(5) Recent data document the range of melt compositions that may exist at any one time in the summit reservoir. Extra-caldera summit lavas, such as those of the 1959 Kīlauea Iki eruption and the 1971 and 1974 summit eruptions, contain melts with a much wider range of MgO contents than observed in the intracaldera lavas, as have explosive eruptions, including the 1924 eruption and various prehistoric tephras (fig. 4).

Calculations using the equilibrium constant for the exchange of Fe and Mg between melt and olivine (Roeder and Emslie, 1970), and using the most magnesian melt compositions (9.7–10.4 weight percent MgO) from the 1959 eruption, show that $Fo_{86–87}$ is in equilibrium with those melts when Fe_2O_3 is 15±5 percent of the iron present. This is somewhat higher than fayalite-magnetite-quartz (FMQ) conditions and is a widely observed level for Kīlauea basalts (Carmichael and Ghiorso, 1986) and gases (Gerlach, 1993). The fit suggests that melts similar to the most magnesian 1959 melts, which fall within the upper end of the MgO range of the various tephras, are appropriate to be interstitial to olivine (Fo_{86-87}) at the base of the reservoir. Given the rate at which olivine reequilibrates as interstitial melts evolve (shifting from $Fo_{86–87}$ to $<Fo_{81}$ in 15–30 years in Kīlauea Iki), the fact that olivine composition is constant over centuries requires a continuous flux of MgO-rich melt into the base of the reservoir at both Kīlauea and Mauna Loa.

The Structure of Kīlauea's Summit Reservoir in Light of Petrologic Observations

Kīlauea's very active summit reservoir leaked low-MgO but olivine-controlled lava almost continually through the 19th and 20th centuries; the only significant breaks in summit eruptive activity have been the intervals 1934–52 and 1982–2008. But why does the summit rarely erupt anything other than the limited range of melts illustrated in figure 3? And why is the average olivine composition in more olivine-rich lavas Fo_{86-87}, when we rarely see intracaldera melts with the necessary MgO content? The uppermost and lowermost parts of Kīlauea's summit reservoir have both been consistent in character for several centuries. They are also very different, raising the question of what physical arrangement is consistent with their persistent differences.

Kīlauea's summit reservoir was first envisaged as a plexus of sill-like magma bodies because of the complex pattern of summit inflation that preceded the 1967–68 summit eruption (Fiske and Kinoshita, 1969; Kinoshita and others, 1969). Other studies, depending on geophysical observations (Duffield and others, 1982; Ryan and others, 1983), led to similar models of a segmented magma reservoir, or even multiple discrete but interconnected reservoirs (Poland and others, this volume, chap. 5). Segmentation of the reservoir would allow different levels in the reservoir to contain persistently different melt compositions. It would also allow the distinctive magma batches observed at Kīlauea (table 1) to be preserved as they move through the summit reservoir.

The variation of density of Kīlauea melts with temperature may also be relevant. Figure 22 shows densities of glasses from the 1959 eruption, together with glass densities for selected layers from the Keanakāko‘i and Pāhala tephras, calculated using the data of Lange and Carmichael (1990). The olivine-controlled melts have temperatures of ~1,170–1,280 °C (Helz and Thornber, 1987) and their densities decrease as temperature decreases, as would be expected for a sequence of melts crystallizing olivine (Sparks and others, 1980; Stolper and Walker, 1980; Sparks and Huppert, 1984).

The presence of dissolved water will lower the calculated melt density values considerably (Ochs and Lange, 1999). Clague and others (1995b) have estimated the density of primary Kīlauea melt (MgO=16.5 percent) to be 2.66–2.68 g/cm^3 at 0.37 percent H_2O, as cited above. The data in figure 22 stop at 13 weight percent MgO and T=1,280 °C, so the uppermost densities are not directly comparable. Although the scale on the ordinate in figure 22 would change at such H_2O contents, the slopes for olivine-controlled melts would be steeper, not flattened or reversed (Lange and Carmichael, 1990; Helz, 2009). Also, although stored magma loses CO_2 early, H_2O is largely retained in the melt to very shallow levels (Gerlach and Graeber, 1985; Gerlach and others, 2002), being released in significant quantities only during eruption (for example, Wallace and Anderson, 1998).

Magma consists of crystals and gas bubbles, as well as melt—here, again, the expected gradients (more olivine at the bottom, less at the top; gas bubbles produced during storage either a uniform flux or perhaps slightly increasing toward the top) would result in higher bulk density at the bottom of the reservoir and lower bulk density at the top. The principal source of instability in the magma reservoir (between major summit eruptions) would appear to be the continual input of fresh magma from the mantle.

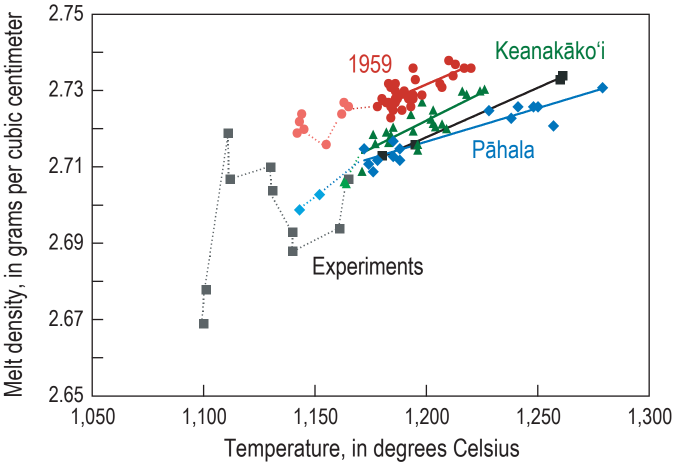


Figure 22. Graph of melt density versus temperature for four sets of glass analyses. Densities calculated using the procedure of Lange and Carmichael (1990), with H_2O and Fe_2O_3 set to zero in the absence of data. Red dots, glasses from 1959 eruption samples (Helz, 2009; and R.T. Helz, unpublished data). Green triangles, glasses from selected layers in the Keanakāko‘i tephra (Helz and others, 2014). Blue diamonds, selected layers in the Pāhala Ash (Helz and others, 2014). Black squares, glasses from Kīlauea Iki sample KI75-1-143.8 experiments (Helz and Thornber, 1987). Darker symbols indicate melts crystallizing only olivine+chromite; lighter symbols indicate melts crystallizing augite±plagioclase±Fe-Ti oxides in addition to olivine. Lines above 1,170 °C are least-squares fits for the melts crystallizing only olivine±chromite; dotted lines indicate subsequent path of density versus temperature for more differentiated samples.

As discussed in Gerlach and others (2002), the presence of CO_2 bubbles is critical to the buoyancy of intruding primary magma when the crystal content of the stored magma is <5 percent. Crystal contents in the deepest part of the reservoir are probably higher than 5 percent, but they are very low at the top, with the crystallinity at intermediate levels being unknown. A further consideration is that the specific molar volume of CO_2 increases rapidly at pressures below 2 kbar (Bottinga and Javoy, 1989), which is roughly the pressure at the base of Kīlauea's summit reservoir (Ryan, 1987); thus, a slow but steady input of primary magma into the lowest part of the reservoir should produce considerable turbulence in the deepest segment of that reservoir. The rapid increase in buoyancy of the CO_2 bubbles once they enter the reservoir, however, should facilitate their separation from the input magma. The resulting bubble plumes might entrain significant crystals+melt as they rise further in the reservoir (Sparks, 1978). The vorbs in Kīlauea Iki formed by a similar process (Helz and others, 1989) and have entrained abundant coarse olivine (fig. 12*A*), but, in the lava lake, the bubble plumes traversed a semicoherent crystal mush of olivine+augite+plagioclase. In the summit reservoir, by contrast, bubble plumes would encounter only olivine±chromite in rapidly diminishing quantities at higher levels. Efficient segregation of bubbles in magma of decreasing crystal content would tend to restrict turbulence/entrainment to the immediate vicinity of the bubble plume. Furthermore, if the summit reservoir is segmented, it is not clear how transfer of turbulence from the segment at the level of primary input to higher segments would occur. One observation that suggests limited disruption of the summit reservoir by the normal level of magma input is that the CO_2 plume apparently carries no melt with it, and has been emitted steadily whether lava is erupting at the summit or not (Gerlach and others, 2002).

To summarize, geophysical evidence favors a segmented summit reservoir, with septa of olivine-rich and (or) partly molten material surrounding melt-rich bodies. A cartoon version of such a reservoir (fig. 23) envisions small, discontinuous but connected bodies of melt lying within the hot aseismic region shown in Ryan (1988). Such a structure is consistent with the observed preservation of distinct magma batches at Kīlauea and the observed persistent differences between the top and the base of the reservoir. The eruption of more magnesian melts just outside the caldera (as in 1959, 1971, and 1974) also argues for some degree of separation or stratification within the reservoir. Internal segregation by weak septa would be supported or enhanced if the melt column at Kīlauea tends to be gravitationally stable. In any case, a segmented reservoir structure would inhibit complete overturn of the summit reservoir under normal conditions.

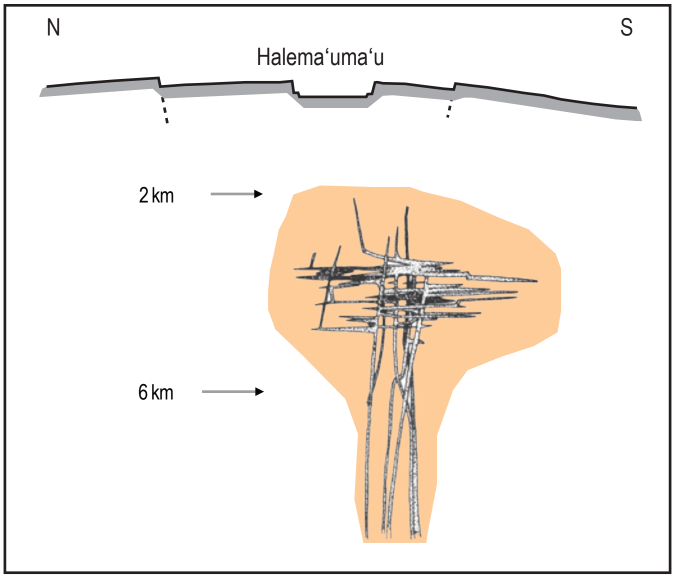


Figure 23. Cartoon showing Kīlauea's summit reservoir in a north-south cross section that intersects neither of the volcano's rift zones, modified from Fiske and Kinoshita (1969). The colored field shows the seismically quiet volume described by Ryan (1988), which, by implication, delineates the summit reservoir as lying between 2 and 6 km below the surface. The magma bodies, as drawn, are complexly segmented and lie within the seismically quiet volume.

The Summit Complex Disrupted

Relatively young, explosive eruptions at Kīlauea's summit include the 1924 eruption, the Keanakākoʻi eruptions (which occurred from 1460 to 1790 C.E.; Swanson and others, 2012), and the Kulanaokuaiki eruptions (400–1000 C.E.; Fiske and others, 2009). These, together with the 1959 summit eruption and the still older Pāhala Ash, show wide ranges of melt compositions (fig. 4).

The Keanakākoʻi eruptions are interpreted by Mastin and others (2004) to have resulted from rapid injection of magma to high levels, presumably overstepping the storage capacity of the summit reservoir. The frequency distributions for all eruptions in figure 4, however, are dominated volumetrically by melts at intermediate MgO contents, which cannot reasonably be explained by mixing between the sparse low-MgO and high-MgO tails of the distributions during fragmentation and rapid eruption. It seems more likely that most melt compositions (excepting those in the high-MgO tails) existed in the reservoir and were entrained during these explosive eruptions. By contrast, the 1924 tephra resulted from collapse of the upper part of the reservoir, with ejection of only traces of juvenile material, which nevertheless included melts with as much as 8.7 percent MgO. The presence of relatively magnesian liquids in the 1959, 1971, and 1974 summit eruptions, as well, suggests that such melts are commonly present in the existing reservoir.

The 1959 Kīlauea Iki Eruption

The 1959 eruption was closely monitored and observed, and many samples of scoria were collected in real time (Murata and Richter, 1966; Richter and others, 1970), thus providing a view into the reservoir during the eruption. One component of the 1959 magma was tracked seismically from below 60 km to the surface (Eaton and Murata, 1960). It intersected Kīlauea's summit reservoir on its north side and interacted with stored magma related to the 1954 summit batch to produce a range of mixed magmas (Wright, 1973; Helz, 1987a). This juvenile (1959E) component was injected into the erupting volume (which was dominated by the stored 1959W component) at several stages during the eruption (Eaton and others, 1987; Helz, 1987a). Unusually high fountain heights (300–580 m; Richter and others, 1970) showed that the eruption was exceptionally gas-rich. Studies of the olivine phenocryst population also suggest that part of the magma bypassed significant storage time (Harris and Anderson, 1983; Helz, 1987a; Anderson and Brown, 1993; Wallace and Anderson, 1998).

The observed succession of scoria erupted in 1959 suggests that the summit reservoir contained discrete magma bodies, in addition to entrainable Fo_{86-87} at its base. The early lavas erupted were dominated by the juvenile (1959E) component, and include some of the hottest magmas of the eruption (9.7–10.4 weight percent MgO, T=1,210–1,220 °C). These were followed abruptly by lavas (erupted on November 20–21 and again on November 26) that contained only 6–9 percent by weight of the juvenile component (Wright, 1973). This group of samples provides a good look at one body of the stored (1959W) magma, which had been in the reservoir for at least five years but was still uniformly hotter and more magnesian (8.5–8.6 weight percent MgO, T=1,185–1,187 °C) than melts in the top of the reservoir.

After a pause in the eruption, renewed activity on November 28–29 produced melts 10–30 °C cooler than those from the previously sampled body of stored (1959W) magma. The source body for these phase 3 lavas must have been separate from the body sampled earlier first, because (unlike any but the two earliest 1959 scoria) the lavas erupted from it were not hybrids (Wright, 1973) and second, because it was too cool to be parental to any of the later (phase 4–17) hybrids, which have glass MgO contents and temperatures similar to, or higher than, the first body of 1959W magma (Helz, 1987a, 2009).

The dominant melts (fig. 4) of the 1959 eruption should be in equilibrium with Fo_{83-84} at the redox conditions inferred for the deepest parts of the reservoir (Fe_2O_3 ~15±5 percent of the iron present, as discussed above). Nevertheless, the overall average for the eruption is Fo_{86-87} (fig. 11), and olivine $<Fo_{85}$ is sparse in the observed phenocryst population (Helz, 1987a). This is consistent with the two stored magma bodies tapped during the eruption being isolated from the deeper olivine-rich zone of the reservoir, suggesting that three distinct levels of the reservoir were tapped during the 1959 eruption.

Evolution of the Summit Reservoirs Through Time

The pattern of eruption at Kīlauea suggests that the distribution and extent of melt-rich bodies in the summit reservoir may change with time. First, the summit reservoir gradually recharged between 1924 and 1968, with high segmentation and gradually expanding magma storage. Later, as eruptive activity at Kīlauea shifted more strongly to the East Rift Zone, magma batches became less distinctive, although the entire system (summit and both rift zones) saw magma storage and activity during 1971 and 1974. Since 1982, however, East Rift Zone activity has dominated. A reasonable consequence of the long-term, ongoing East Rift Zone eruption would be reconfiguration of the summit reservoir to facilitate transport to the eruption site on the East Rift Zone. This, in turn, could lead to atrophy of parts of the reservoir, so that its overall capacity would be less than in the 1960s and 1970s and the only melts in storage would be related to the ongoing East Rift Zone eruption. However, a recent microgravity study by Johnson and others (2010), covering the time span 1975–2008, suggests that magma has been accumulating at a depth of 1 km below the east edge of Halemaʻumaʻu Crater over much of this same period. The summit vent that opened in early 2008 is almost directly over the inferred location of the accumulation zone, and Johnson and others (2010) suggest that this shallow body is the source of the current summit activity. The compositions of tephra from the current summit eruption are identical to coevally erupted material at Puʻu ʻŌʻō (Rowe and others, 2009), suggesting that the extent of cooling en route to the shallow summit reservoir is similar to that which occurs in the main flux of magma out to Puʻu ʻŌʻō.

Mauna Loa has seen only limited activity in the past 60 years, in contrast to earlier periods; nevertheless, the lavas erupted have been extremely uniform. If the magma in its reservoir is crystallizing plagioclase (not true at Kīlauea at present; fig. 9), there may be spontaneous mixing of melts at the top of the reservoir, driven by the reversal in melt density versus temperature seen at that point in basalt crystallization (fig. 22). Another explanation would be that Mauna Loa's mantle source has been more uniform than Kīlauea's over this period.

Long-term variations in trace element and isotopic compositions of basalts at Kīlauea and Mauna Loa shed light on the variability of their magmatic behavior over the past two centuries. Rhodes and Hart (1995) found analytically significant variations in trace element levels of Mauna Loa lavas, even in the absence of magma batches definable by major element chemistry. Pietruzka and Garcia (1999a,b) and Pietruzka and others (2001) show similar variability in trace element and isotopic signatures at Kīlauea, again decoupled from magma batches defined by major elements. This variability has been attributed to changes in the degree of partial melting and (or) variations in exact source composition. At Mauna Loa, shifts consistent with increasing melting and magma supply preceded the 1868 seismic crisis, with incompatible trace element levels rising (and magma supply dropping) afterward (Rhodes and Hart, 1995). At Kīlauea, the major event was the 1924 collapse of the summit magma column (Pietruzka and Garcia, 1999a,b; Pietruzka and others, 2001), with incompatible trace elements rising (and the extent of melting and magma supply decreasing) before the event and recovering afterward. Evidently, the entire magmatic system at both volcanoes is continuous enough that any changes in magma production and (or) supply from the source region have large and geologically rapid impacts on the upper reaches of the volcanic plumbing, as documented recently by Poland and others (2012).

Summary and Conclusions

Study of the youngest Hawaiian volcanoes over the past 100 years has produced a flood of observations relevant to understanding basaltic volcanism, especially, but not exclusively, at shield volcanoes. The overarching, distinctive feature of this effort on the part of Hawaiian Volcano Observatory staff and others is the attempt to observe, document, and quantify basaltic eruptions as completely as possible. Highlights among the results reviewed here include:

(1) Early recognition of magma mixing at Kīlauea in both summit and rift zone lavas, which is now known to be pervasive at this and other volcanoes.

(2) New data that show that the range of melts erupted effusively within Kīlauea Caldera do not represent the entire range of melts in Kīlauea's summit reservoir during the 20th century, as evidenced by the eruption of extracaldera summit lavas with glass MgO contents of 8–9 percent by weight in the 1959, 1971, and 1974 eruptions and in the rare presence of similar material in the 1924 tephra.

(3) Recognition that young, prehistoric tephras at Kīlauea have similar or larger ranges of glass MgO contents. The frequency distributions of glass MgO contents suggest that most of the range of melts sampled during Kīlauea's explosive eruptions existed in the summit reservoir before eruption.

(4) Documentation of the reequilibration of olivine during closed-system crystallization of Kīlauea Iki lava lake, which shows that olivine compositions shifted from Fo_{86-87} to $<Fo_{81}$ in 15–30 years. The Fo_{86-87} control lines widely observed in lavas from both Kīlauea and Mauna Loa therefore require that the lower parts of both summit reservoirs see a continual influx of melt of appropriate composition.

(5) Olivine control lines for subaerial rift zone lavas, whether olivine-poor or picritic, are also Fo_{86-87} at both volcanoes. Most rift zone eruptions thus appear to be fed from no deeper than the middle to lower parts of the summit reservoir (4–6 km depth at Kīlauea).

(6) Olivine control lines for submarine picritic lavas are Fo_{88} at Kīlauea and Fo_{89} at Hualālai, implying that melts below the level of the summit reservoirs, including in the deeper parts of the rift zones, are, on average, more magnesian than the subaerially erupted melts shown in figure 4.

(7) The most magnesian submarine glasses on the submerged part of Kīlauea's East Rift Zone (14.7–15.0 weight percent MgO) are close to inferred primary compositions for Kīlauea and Mauna Loa magmas (14–20 weight percent MgO). However, picritic lavas observed at these and other Hawaiian volcanoes—even though they have similar bulk MgO contents—have been interpreted, without exception, as cumulates and not as quenched ultramafic liquids.

The accumulated observations of the past 100 years are consistent with the view that the magmatic systems of Kīlauea and Mauna Loa are extremely open, locally segmented but well connected, and continually refreshed with new magma from the mantle.

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References Cited

Anderson, A.T., Jr., and Brown, G.G., 1993, CO_2 contents and formation pressures of some Kilauean melt inclusions: American Mineralogist, v. 78, nos. 7–8, p. 794–803, accessed June 4, 2013, at http://www.minsocam.org/ammin/AM78/AM78_794.pdf.

Anderson, A.T., Jr., and Wright, T.L., 1972, Phenocrysts and glass inclusions and their bearing on oxidation and mixing of basaltic magmas, Kilauea volcano, Hawaii: American Mineralogist, v. 57, nos. 1–2, p. 188–216, accessed June 4, 2013, at http://www.minsocam.org/ammin/AM57/AM57_188.pdf.

Anderson, D.L., 1998, The helium paradoxes: Proceedings of the National Academy of Sciences, U.S.A., v. 95, p. 4822–4827, accessed June 4, 2013, at http://www.pnas.org/content/95/9/4822.full.pdf.

Babb, J.L., Kauahikaua, J.P., and Tilling R.I., 2011, The story of the Hawaiian Volcano Observatory—A remarkable first 100 years of tracking eruptions and earthquakes: U.S. Geological Survey General Information Product 135, 60 p. [Also available at http://pubs.usgs.gov/gip/135/.]

Baker, N.A., 1987, A petrologic study of the 1982 summit lavas of Kilauea Volcano, Hawaii: Honolulu, University of Hawaiʻi at Mānoa, Senior Honors thesis, 61 p.

Ballentine, C.J., Porcelli, D., and Weiler, R., 2001, Noble gases in mantle plumes; technical comments [on Trieloff, M., Kunz, J., Clague, D.A., Harrison, D., and Allègre, C.J., 2000, The nature of pristine noble gases in mantle plumes]: Science, v. 291, no. 5512, p. 2269, doi:10.1126/science.291.5512.2269a.

Banks, N.G., Wolfe, E.W., Duggan, T.J., Okamura, A.T., Koyanagi, R.Y., Greenland, L.P., and Jackson, D.B., 1982, Magmatic events at Kilauea Volcano, Hawaii, 1982 [abs.]: Eos (American Geophysical Union Transactions), v. 64, no. 45, supp., p. 901–902, abstract no. V51A–12.

Barth, G.A., Kleinrock, M.C., and Helz, R.T., 1994, The magma body at Kilauea Iki lava lake; potential insights into mid-ocean ridge magma chambers: Journal of Geophysical Research, v. 99, no. B4, p. 7199–7217, doi:10.1029/93JB02804.

Basaltic Volcanism Study Project, 1981, Oceanic intraplate volcanism, *in* Basaltic volcanism on the terrestrial planets: New York, Pergamon Press, p. 161–192.

Bauer, G.R., Fodor, R.V., Husler, J.W., and Keil, K., 1973, Contributions to the mineral chemistry of Hawaiian rocks; III. Composition and mineralogy of a new rhyodacite occurrence on Oahu, Hawaii: Contributions to Mineralogy and Petrology, v. 40, no. 3, p. 183–194, doi:10.1007/BF00371038.

Beeson, M.H., Clague, D.A., and Lockwood, J.P., 1996, Origin and depositional environment of clastic deposits in the Hilo Drill Hole, Hawaii, *in* Results of the Hawaii Scientific Drilling Project 1-km core hole at Hilo, Hawaii: Journal of Geophysical Research, v. 101, no. B5 (special section), p. 11617–11629, doi:10.1029/95JB03703.

Bevens, D., Takahashi, T.J., and Wright, T.L., eds., 1988, The early serial publications of the Hawaiian Volcano Observatory (compiled and reprinted): Hawaii National Park, Hawaii, Hawaiʻi Natural History Association, 3 v., 3,062 p. (v. 1, 565 p.; v. 2, 1,273 p.; v. 3, 1,224 p.)

Bottinga, Y., and Javoy, M., 1989, MORB degassing; evolution of CO_2: Earth and Planetary Science Letters, v. 95, nos. 3–4, p. 215–225, doi:10.1016/0012-821X(89)90098-8.

Buddington, A.F., and Lindsley, D.H., 1964, Iron-titanium oxide minerals and synthetic equivalents: Journal of Petrology, v. 5, no. 2, p. 310–357, doi:10.1093/petrology/5.2.310.

Calvert, A.T., and Lanphere, M.A., 2006, Argon geochronology of Kilauea's early submarine history, *in* Coombs, M.L., Eakins, B.W., and Cervelli, P.F., eds., Growth and collapse of Hawaiian volcanoes: Journal of Volcanology and Geothermal Research, v. 151, nos. 1–3 (special issue), p. 1–18, doi.org/10.1016/j.jvolgeores.2005.07.023.

Carmichael, I.S.E., and Ghiorso, M.S., 1986, Oxidation-reduction relations in basic magma; a case for homogeneous equilibria: Earth and Planetary Science Letters, v. 78, nos. 2–3, p. 200–210, doi:10.1016/0012-821X(86)90061-0.

Cashman, K.V., 1993, Relationship between plagioclase crystallization and cooling rate in basaltic melts: Contributions to Mineralogy and Petrology, v. 113, no. 1, p. 126–142, doi:10.1007/BF00320836.

Chakraborty, S., 2010, Diffusion coefficients in olivine, wadsleyite and ringwoodite, *in* Zhang, Y., and Cherniak, D., eds., Diffusion in minerals and melts: Reviews in Mineralogy and Geochemistry, v. 72, no. 1, p. 603–639, doi:10.2138/rmg.2010.72.13.

Chen, C.-Y., and Frey, F.A., 1983, Origin of Hawaiian tholeiite and alkalic basalt: Nature, v. 302, no. 5911, p. 785–789, doi:10.1038/302785a0.

Chen, C.-Y., and Frey, F.A., 1985, Trace element and isotopic geochemistry of lavas from Haleakala volcano, East Maui, Hawaii; implications for the origin of Hawaiian basalts: Journal of Geophysical Research, v. 90, no. B10, p. 8743–8768, doi:10.1029/JB090iB10p08743.

Chen, C.-Y., Frey, F.A., Rhodes, J.M., and Easton, R.M., 1996, Temporal geochemical evolution of Kilauea Volcano; comparison of Hilina and Puna basalt, *in* Basu, A., and Hart, S., eds., Earth processes; reading the isotopic code: American Geophysical Union Geophysical Monograph 95, p. 161–181, doi:10.1029/GM095p0161.

Clague, D.A., 1987a, Hawaiian alkaline volcanism, *in* Fitton, J.G., and Upton, B.G.J., eds., Alkaline igneous rocks: Geological Society of London Special Publication No. 30, p. 227–252, doi:10.1144/GSL.SP.1987.030.01.10.

Clague, D.A., 1987b, Hawaiian xenolith populations, magma supply rates, and development of magma chambers: Bulletin of Volcanology, v. 49, no. 4, p. 577–587, doi:10.1007/BF01079963.

Clague, D.A., 1988, Petrology of ultramafic xenoliths from Loihi Seamount, Hawaii: Journal of Petrology, v. 29, no. 6, p. 1161–1186, doi:10.1093/petrology/29.6.1161.

Clague, D.A., and Dalrymple, G.B., 1987, The Hawaiian-Emperor volcanic chain; part I. Geologic evolution, chap. 1 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 1, p. 5–54. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Clague, D.A., and Denlinger, R.P., 1994, Role of olivine cumulates in destabilizing the flanks of Hawaiian volcanoes: Bulletin of Volcanology, v. 56, nos. 6–7, p. 425–434, doi:10.1007/BF00302824.

Clague, D.A., and Dixon, J.E., 2000, Extrinsic controls on the evolution of Hawaiian ocean island volcanoes: Geochemistry, Geophysics, Geosystems (G^3), v. 1, no. 4, 1010, 9 p., doi:10.1029/1999GC000023.

Clague, D.A., and Sherrod, D.R., 2014, Growth and degradation of Hawaiian volcanoes, chap. 3 *of* Poland, M.P., Takahashi, T.J., and Landowski, C.M., eds., Characteristics of Hawaiian volcanoes: U.S. Geological Survey Professional Paper 1801 (this volume).

Clague, D.A., Weber, W.S., and Dixon, J.E., 1991, Picritic glasses from Hawaii: Nature, v. 353, no. 6344, p. 553–556, doi:10.1038/353553a0.

Clague, D.A., Beeson, M.H., Denlinger, R.P., and Mastin L.G., 1995a, Ancient ash deposits and calderas at Kilauea Volcano [abs.]: American Geophysical Union, Fall Meeting 1995 Abstracts, abstract no. V22C–03, accessed June 4, 2013, at http://abstractsearch.agu.org/meetings/1995/FM/sections/V/sessions/V22C.html.

Clague, D.A., Moore, J.G., Dixon, J.E., and Friesen, W.G., 1995b, Petrology of submarine lavas from Kilauea's Puna Ridge, Hawaii: Journal of Petrology, v. 36, no. 2, p. 299–349, doi:10.1093/petrology/36.2.299.

Clague, D.A., Hagstrum, J.T., Champion D.E., and Beeson, M.H., 1999, Kīlauea summit overflows; their ages and distribution in the Puna District, Hawai‘i: Bulletin of Volcanology, v. 6, no. 6, p. 363–381, doi:10.1007/s004450050279.

Clague, D.A., Davis, A.S., Bischoff, J.L., Dixon, J.E., and Geyer, R., 2000, Lava bubble-wall fragments formed by submarine hydrovolcanic explosions on Lō‘ihi Seamount and Kīlauea Volcano: Bulletin of Volcanology, v. 61, no. 7, p. 437–449, doi:10.1007/PL00008910.

Clague, D.A., Batiza, R., Head, J.W., III, and Davis, A.S., 2003, Pyroclastic and hydroclastic deposits on Loihi Seamount, Hawaii, *in* White, J.D.L., Smellie, J.L., and Clague, D.A., eds., Explosive subaqueous volcanism: American Geophysical Union Geophysical Monograph 140, p. 73–95, doi:10.1029/140GM05.

Coombs, M.L., Sisson, T.W., and Kimura, J., 2004, Ultra-high chlorine in submarine Kīlauea glasses; evidence for assimilation of brine by magma: Earth and Planetary Science Letters, v. 217, nos. 3–4, p. 297–313, doi:10.1016/S0012-821X(03)00631-9.

Coombs, M., Sisson, T., and Lipman, P., 2006, Growth history of Kilauea inferred from volatile concentrations in submarine-collected basalts, *in* Coombs, M.L., Eakins, B.W., and Cervelli, P.F., eds., Growth and collapse of Hawaiian volcanoes: Journal of Volcanology and Geothermal Research, v. 151, nos. 1–3 (special issue), p. 19–49, doi:10.1016/j.jvolgeores.2005.07.037.

Cooper, K.M., Reid, M.R., Murrell, M.T., and Clague, D.A., 2001, Crystal and magma residence at Kilauea Volcano, Hawaii; ^{230}Th-^{226}Ra dating of the 1955 east rift eruption: Earth and Planetary Science Letters, v. 184, nos. 3–4, p. 703–718, doi:10.1016/S0012-821X(00)00341-1.

Dana, J.D., 1849, Geology, v. 10 *of* Wilkes C., Narrative of the United States Exploring Expedition; during the years 1838, 1839, 1840, 1841, 1842 (under the command of Charles Wilkes, U.S.N.): New York, George P. Putman, 756 p., with a folio atlas of 21 pls.

DePaolo, D.J., and Stolper, E.M., 1996, Models of Hawaiian volcano growth and plume structure; implications of results from the Hawaii Scientific Drilling Project, *in* Results of the Hawaii Scientific Drilling Project 1-km core hole at Hilo, Hawaii: Journal of Geophysical Research, v. 101, no. B5 (special section), p. 11643–11654, doi:10.1029/96JB00070.

DePaolo, D.J., Bryce, J.G., Dodson, A., Shuster, D.L., and Kennedy, B.M., 2001, Isotopic evolution of Mauna Loa and the chemical structure of the Hawaiian plume: Geochemistry, Geophysics, Geosystems (G^3), v. 2, no. 7, 1044, 32 p., doi:10.1029/2000GC000139.

Desborough, G.A., Anderson, A.T., Jr., and Wright, T.L., 1968, Mineralogy of sulfides from certain Hawaiian basalts: Economic Geology, v. 63, no. 6, p. 636–644, doi:10.2113/gsecongeo.63.6.636.

Dixon, J.E., and Clague, D.A., 2001, Volatiles in basaltic glass from Loihi seamount, Hawaii; evidence for a relatively dry plume component: Journal of Petrology, v. 42, no. 3, p. 627–654, doi:10.1093/petrology/42.3.627.

Dixon, J.E., Clague, D.A., and Stolper, E.M., 1991, Degassing history of water, sulfur, and carbon in submarine lavas from Kilauea Volcano, Hawaii: The Journal of Geology, v. 99. no. 3, p. 371–394, doi:10.1086/629501.

Duffield, W.A., 1972, A naturally occurring model of global plate tectonics: Journal of Geophysical Research, v. 77, no. 14, p. 2543–2555, doi:10.1029/JB077i014p02543.

Duffield, W.A., Christiansen, R.L., Koyanagi, R.Y., and Peterson, D.W., 1982, Storage, migration, and eruption of magma at Kilauea Volcano Hawaii, 1971–1972: Journal of Volcanology and Geothermal Research, v. 13, nos. 3–4, p. 273–307, doi:10.1016/0377-0273(82)90054-3.

Easton, R.M., 1987, Stratigraphy of Kilauea Volcano, chap. 11 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 1, p. 243–260. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Eaton, J.P., and Murata, K.J., 1960, How volcanoes grow: Science, v. 132, no. 3432, p. 925–938, doi:10.1126/science.132.3432.925.

Eaton, J.P., Richter, D.H., and Krivoy, H.L., 1987, Cycling of magma between the summit reservoir and Kilauea Iki lava lake during the 1959 eruption of Kilauea Volcano, chap. 48 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 2, p. 1307–1335. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Evans, B.W., and Moore, J.G., 1968, Mineralogy as a function of depth in the prehistoric Makaopuhi tholeiitic lava lake, Hawaii: Contributions to Mineralogy and Petrology, v. 17, no. 2, p. 85–115, doi:10.1007/BF00373204.

Fabbrizio, A., Beckett, J.R., Baker, M.B., and Stolper, E.M., 2010, Phosphorus zoning in olivine of Kilauea Iki lava lake, Hawaii [abs.]: Geophysical Research Abstracts, v. 12, EGU2010-1418-1, p. 1418 (EGU General Assembly 2010 meeting, Vienna, Austria, May 2–7, 2010), accessed June 4, 2013, at http://adsabs.harvard.edu/abs/2010EGUGA..12.1418F.

Finch, R.H., and Macdonald, G.A., 1953, Hawaiian volcanoes during 1950: U.S. Geological Survey Bulletin 996–B, p. 27–89, 2 folded maps in pocket, scale ~1:190,080. [Also available at http://pubs.usgs.gov/bul/0996b/report.pdf.]

Fiske, R.S., and Kinoshita, W.T., 1969, Inflation of Kilauea Volcano prior to its 1967–1968 eruption: Science, v. 165, no. 3891, p. 341–349, doi:10.1126/science.165.3891.341.

Fiske, R.S., Simkin, T., and Nielsen, E.A., eds., 1987, The Volcano Letter: Washington, D.C., Smithsonian Institution Press, n.p. (530 issues, compiled and reprinted; originally published by the Hawaiian Volcano Observatory, 1925–1955).

Fiske, R.S., Rose, T.R., Swanson, D.A., Champion, D.E., and McGeehin, J.P., 2009, Kulanaokuaiki Tephra (ca. A.D. 400–1000); newly recognized evidence for highly explosive eruptions at Kīlauea Volcano, Hawaiʻi: Geological Society of America Bulletin, v. 121, nos. 5–6, p. 712–728, doi:10.1130/B26327.1.

Frey, F.A., and Clague, D.A., 1983, Geochemistry of diverse basalt types from Loihi Seamount, Hawaii; petrogenetic implications, *in* Loihi Seamount; collected papers: Earth and Planetary Science Letters, v. 66, December, p. 337–355, doi:10.1016/0012-821X(83)90150-4.

Frey, F.A., Clague, D.A., Mahoney, J.J., and Sinton, J.M., 2000, Volcanism at the edge of the Hawaiian plume; petrogenesis of submarine alkali lavas from the North Arch Volcanic Field: Journal of Petrology, v. 41, no. 5, p. 667–691, doi:10.1093/petrology/41.5.667.

Garcia, M.O., and Wolfe, E.W., 1988, Petrology of the erupted lava, chap. 3 *of* Wolfe, E.W., ed., The Puu Oo eruption of Kilauea Volcano, Hawaii; episodes 1 through 20, January 3, 1983, through June 8, 1984: U.S. Geological Survey Professional Paper 1463, p. 127–143. [Also available at http://pubs.usgs.gov/pp/1463/report.pdf.]

Garcia, M.O., Ho, R.A., Rhodes, J.M., and Wolfe, E.W., 1989, Petrologic constraints on rift-zone processes; results from episode 1 of the Puu Oo eruption of Kilauea volcano, Hawaii: Bulletin of Volcanology, v. 52, no. 2, p. 81–96, doi:10.1007/BF00301548.

Garcia, M.O., Rhodes, J.M., Wolfe, E.W., Ulrich, G.E., and Ho, R.A., 1992, Petrology of lavas from episodes 2–47 of the Puu Oo eruption of Kilauea Volcano, Hawaii; evaluation of magmatic processes: Bulletin of Volcanology, v. 55, nos. 1–2, p. 1–16, doi:10.1007/BF00301115.

Garcia, M.O., Jorgenson, B.A., Mahoney, J.J., Ito, E., and Irving, A.J., 1993, An evaluation of temporal geochemical evolution of Loihi summit lavas; results from *Alvin* submersible dives: Journal of Geophysical Research, v. 98, no. B1, p. 537–550, doi:10.1029/92JB01707.

Garcia, M.O., Foss, D.J.P., West, H.B., and Mahoney, J.J., 1995a, Geochemical and isotopic evolution of Loihi Volcano, Hawaii: Journal of Petrology, v. 36, no. 6, p. 1647–1674. (Correction to figure 10 published in Journal of Petrology, 1996, v. 37, no. 3, p. 729, doi:10.1093/petrology/37.3.729.)

Garcia, M.O., Hulsebosch, T.P., and Rhodes, J.M., 1995b, Olivine-rich submarine basalts from the southwest rift zone of Mauna Loa Volcano; implications for magmatic processes and geochemical evolution, *in* Rhodes, J.M., and Lockwood, J.P., eds., Mauna Loa revealed; structure, composition, history, and hazards: American Geophysical Union Geophysical Monograph 92, p. 219–239, doi:10.1029/GM092p0219.

Garcia, M.O., Rhodes, J.M., Trusdell, F.A., and Pietruszka, A.J., 1996, Petrology of lavas from the Pu‘u ‘O‘o eruption of Kilauea Volcano; III. The Kupaianaha episode (1986–1992): Bulletin of Volcanology, v. 58, no. 5, p. 359–379, doi:10.1007/s004450050145.

Garcia, M.O., Rubin, K.H., Norman, M.D., Rhodes, J.M., Graham, D.W., Muenow, D.W., and Spencer, K., 1998, Petrology and geochronology of basalt breccia from the 1996 earthquake swarm of Loihi seamount, Hawaii; magmatic history of its 1996 eruption: Bulletin of Volcanology, v. 59, no. 8, p. 577–592, doi:10.1007/s004450050211.

Garcia, M.O., Pietruszka, A.J., Rhodes, J.M., and Swanson, K., 2000, Magmatic processes during the prolonged Pu’u ’O’o eruption of Kilauea Volcano, Hawaii: Journal of Petrology, v. 41, no. 7, p. 967–990, doi:10.1093/petrology/41.7.967.

Garcia, M.O., Caplan-Auerbach, J., De Carlo, E.H., Kurz, M.D., and Becker, N., 2006, Geology, geochemistry and earthquake history of Lō`ihi Seamount, Hawai`i’s youngest volcano: Chemie der Erde, v. 66, no. 2, p. 81–108, doi:10.1016/j.chemer.2005.09.002.

Garcia, M.O., Mucek, A.E., and Swanson D.A., 2011, Geochemistry of glass and olivine from Keanakako‘i tephra at Kilauea Volcano, Hawai‘i [abs.]: American Geophysical Union, Fall Meeting 2011 Abstracts, abstract no. V41A–2480, accessed April 28, 2014, at http://abstractsearch.agu.org/meetings/2011/FM/sections/V/sessions/V41A/abstracts/V41A-2480.html.

Gerlach, T.M., 1993, Oxygen buffering of Kilauea volcanic gases and the oxygen fugacity of Kilauea basalt: Geochimica et Cosmochimica Acta, v. 57, no. 4, p. 795–814, doi:10.1016/0016-7037(93)90169-W.

Gerlach, T.M., and Graeber, E.J., 1985, Volatile budget of Kilauea volcano: Nature, v. 313, no. 6000, p. 273–277, doi:10.1038/313273a0.

Gerlach, T.M., McGee, K.A., Elias, T., Sutton, A.J., and Doukas, M.P., 2002, Carbon dioxide emission rate of Kīlauea Volcano; implications for primary magma and the summit reservoir: Journal of Geophysical Research, v. 107, no. B9, 2189, 15 p., doi:10.1029/2001JB000407.

Gunn, B.M., 1971, Trace element partition during olivine fractionation of Hawaiian basalts: Chemical Geology, v. 8, no. 1, p. 1–13, doi:10.1016/0009-2541(71)90043-X.

Hammer, J.E., Coombs, M.L., Shamberger, P.J., and Kimura, J.-I., 2006, Submarine sliver in North Kona; a window into the early magmatic and growth history of Hualalai Volcano, Hawaii, *in* Coombs, M.L., Eakins, B.W., and Cervelli, P.F., eds., Growth and collapse of Hawaiian volcanoes: Journal of Volcanology and Geothermal Research, v. 151, nos. 1–3 (special issue), p. 157–188, doi:10.1016/j.jvolgeores.2005.07.028.

Hanyu, T., Kimura, J.-I., Katakuse, M., Calvert, A.T., Sisson, T.W., and Nakai, S., 2010, Source materials for inception stage Hawaiian magmas; Pb-He isotope variations for early Kilauea: Geochemistry, Geophysics, Geosystems (G^3), v. 11, Q0AC01, 25 p., doi:10.1029/2009GC002760.

Harris, D.M., and Anderson, A.T., Jr., 1983, Concentrations, sources, and losses of H_2O, CO_2, and S in Kilauean basalt: Geochimica et Cosmochimica Acta, v. 47, no. 6, p. 1139–1150, doi:10.1016/0016-7037(83)90244-2.

Hart, S.R., Hauri, E.H., Oschmann, L.A., and Whitehead, J.A., 1992, Mantle plumes and entrainment; isotopic evidence: Science, v. 256, no. 5056, p. 517–520, doi:10.1126/science.256.5056.517.

Hauri, E.H., 1996, Major-element variability in the Hawaiian mantle plume: Nature, v. 382, no. 6590, p. 415–419, doi:10.1038/382415a0.

Hauri, E.H., Whitehead, J.A., and Hart, S.R., 1994, Fluid dynamic and geochemical aspects of entrainment in mantle plumes: Journal of Geophysical Research, v. 99, no. B12, p. 24275–24300, doi:10.1029/94JB01257.

Hawkins, J., and Melchior, J., 1983, Petrology of basalts from Loihi Seamount, Hawaii, *in* Loihi Seamount; collected papers: Earth and Planetary Science Letters, v. 66, December, p. 356–368, doi:10.1016/0012-821X(83)90151-6.

Heliker, C.C., Swanson, D.A., and Takahashi, T.J., eds., 2003, The Puʻu ʻŌʻō-Kūpaianaha eruption of Kīlauea Volcano, Hawaiʻi; the first twenty years: U.S. Geological Survey Professional Paper 1676, 206 p. [Also available at http://pubs.usgs.gov/pp/pp1676/.]

Helz, R.T., 1980, Crystallization history of Kilauea Iki lava lake as seen in drill core recovered in 1967–1979: Bulletin of Volcanology, v. 43, no. 4, p. 675–701, doi:10.1007/BF02600365.

Helz, R.T., 1987a, Diverse olivine types in lavas of the 1959 eruption of Kilauea Volcano and their bearing on eruption dynamics, chap. 25 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 1, p. 691–722. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Helz, R.T., 1987b, Differentiation behavior of Kilauea Iki lava lake, Kilauea volcano, Hawaii; an overview of past and current work, *in* Mysen, B.O., ed., Magmatic processes; physicochemical principles: The Geochemical Society Special Publication No. 1 (A volume in honor of Hatten S. Yoder, Jr.), p. 241–258.

Helz, R.T., 1993, Drilling report and core logs for the 1988 drilling of Kilauea Iki lava lake, Kilauea Volcano, Hawaii, with summary descriptions of the occurrence of foundered crust and fractures in the drill core: U.S. Geological Survey Open-File Report 93–15, 57 p.

Helz, R.T., 2008, How to produce dacitic melt at Kilauea; evidence from historic Kilauea lava lakes [abs.]: American Geophysical Union, Fall Meeting 2008 Abstracts, abstract no. V23A–2135, accessed April 28, 2014, at http://abstractsearch.agu.org/meetings/2008/FM/sections/V/sessions/V23A/abstracts/V23A-2135.html.

Helz, R.T., 2009, Processes active in mafic magma chambers; the example of Kilauea Iki lava lake, Hawaii: Lithos, v. 111, nos. 1–2, p. 37–46, with electronic supplement, doi:10.1016/j.lithos.2008.11.007.

Helz, R.T., 2012, Trace element analyses of core samples from the 1967–1988 drillings of Kilauea Iki lava lake, Hawaii: U.S. Geological Survey Open-File Report 2012–1050, 46 p. [Also available at http://pubs.usgs.gov/of/2012/1050/OFR2012-1050.pdf.]

Helz, R.T., and Taggart, J.E., Jr., 2010, Whole-rock analyses of core samples from the 1988 drilling of Kilauea Iki lava lake, Hawaii: U.S. Geological Survey Open-File Report 2010–1093, 47 p. [Also available at http://pubs.usgs.gov/of/2010/1093/pdf/ofr2010-1093.pdf.]

Helz, R.T., and Thornber, C.R., 1987, Geothermometry of Kilauea Iki lava lake, Hawaii: Bulletin of Volcanology, v. 49, no. 5, p. 651–668, doi:10.1007/BF01080357.

Helz, R.T., and Wright, T.L., 1983, Drilling report and core logs for the 1981 drilling of Kilauea Iki lava lake (Kilauea Volcano, Hawaii), with comparative notes on earlier (1967–1979) drilling experiences: U.S. Geological Survey Open-File Report 83–326, 66 p. [Also available at http://pubs.usgs.gov/of/1983/0326/report.pdf.]

Helz, R.T., and Wright, T.L., 1992, Differentiation and magma mixing on Kilauea's east rift zone; a further look at the eruptions of 1955 and 1960. Part I. The late 1955 lavas: Bulletin of Volcanology, v. 54, no. 5, p. 361–384, doi:10.1007/BF00312319.

Helz, R.T., Kirschenbaum H., and Marinenko, J.W., 1989, Diapiric transfer of melt in Kilauea Iki lava lake, Hawaii; a quick, efficient process of igneous differentiation: Geological Society of America Bulletin, v. 101, no. 4, p. 578–594, doi:10.1130/0016-7606(1989)101<0578:DTOMIK>2.3.CO;2.

Helz, R.T., Kirschenbaum, H., Marinenko, J.W., and Qian, R., 1994, Whole-rock analyses of core samples from the 1967, 1975, 1979 and 1981 drillings of Kilauea Iki lava lake, Hawaii: U.S. Geological Survey Open-File Report 94–684, 65 p. [Also available at http://pubs.usgs.gov/of/1994/0684/report.pdf.]

Helz, R.T., Banks, N.G., Heliker, C., Neal, C.A., and Wolfe, E.W., 1995, Comparative geothermometry of recent Hawaiian eruptions: Journal of Geophysical Research, v. 100, no. B9, p. 17637–17657, doi:10.1029/95JB01309.

Helz, R.T., Heliker, C., Hon, K., and Mangan, M.T., 2003, Thermal efficiency of lava tubes in the Puʻu ʻŌʻō-Kūpaianaha eruption, *in* Heliker, C., Swanson D.A., and Takahashi, T.J., eds., The Puʻu ʻŌʻō-Kūpaianaha eruption of Kīlauea Volcano, Hawaiʻi; the first 20 years: U.S. Geological Survey Professional Paper 1676, p. 105–120. [Also available at http://pubs.usgs.gov/pp/pp1676/.]

Helz, R.T., Clague, D.A., Mastin, L.G., and Rose, T.R., 2014, Electron microprobe analyses of glasses from Kīlauea Tephra Units, Kīlauea Volcano, Hawaii: U.S. Geological Survey Open-File Report 2014–1090, 24 p., 2 appendixes in separate files, accessed June 4, 2014, at http://dx.doi.org/10.3133/ofr20141090.

Hills, D.J., Morgan, J.K., Moore, G.F., and Leslie, S.C., 2002, Structural variability along the submarine south flank of Kilauea volcano, Hawai‘i, from a multichannel seismic reflection survey, *in* Takahashi, E., Lipman, P.W., Garcia, M.O., Naka, J., and Aramaki, S., eds., Hawaiian volcanoes; deep underwater perspectives: American Geophysical Union Geophysical Monograph 128, p. 105–124, doi:10.1029/GM128p0105.

Hilton, D.R., and Porcelli, D., 2005, Noble gases as mantle tracers, *in* Carlson, R.W., ed., The mantle and core (2d ed.): Oxford, U.K., Elsevier, Treatise on Geochemistry Series, v. 2, p. 277–318, doi:10.1016/B0-08-043751-6/02007-7.

Ho, R.A., and Garcia, M.O., 1988, Origin of differentiated lavas at Kilauea volcano, Hawaii; implications for the 1955 eruption: Bulletin of Volcanology, v. 50, no. 1, p. 35–46, doi:10.1007/BF01047507.

Hofmann, A.W., and Wright, T.L., 1979, Trace element fractionation and the nature of residual phases during tholeiite production in Hawaii, *in* Annual Report of the Director, Department of Terrestrial Magnetism, 1978–1979: Carnegie Institution of Washington Year Book 78, p. 335–340.

Hofmann, A.W., Feigenson, M.D., and Raczek, I., 1984, Case studies on the origin of basalt; III. Petrogenesis of the Mauna Ulu eruption, Kilauea, 1969–1971: Contributions to Mineralogy and Petrology, v. 88, nos. 1–2, p. 24–35, doi:10.1007/BF00371409.

Holcomb, R.T., 1987, Eruptive history and long-term behavior of Kilauea Volcano, chap. 12 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 1, p. 261–350. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Holness, M.B., Humphreys, M.C., Sides, R., Helz, R., and Tegner, C., 2012a, Towards an understanding of disequilibrium dihedral angles in mafic rocks: Journal of Geophysical Research, v. 117, no. B6, B06207, 31 p., doi:10.1029/2011JB008902.

Holness, M.B., Richardson, C., and Helz, R.T., 2012b, Disequilibrium dihedral angles in dolerite sills; a new proxy for cooling rate: Geology, v. 40, no. 9, p. 795–798, doi:10.1130/G33119.1.

Honda, M., McDougall, I., Patterson, D.B., Doulgeris, A., and Clague, D.A., 1991, Possible solar noble-gas component in Hawaiian basalts: Nature, v. 349, no. 6305, p. 149–151, doi:10.1038/349149a0.

Honda, M., McDougall, I., Patterson, D.B., Doulgeris, A., and Clague, D.A., 1993, Noble gases in submarine pillow basalt glasses from Loihi and Kilauea, Hawaii; a solar component in the Earth: Geochimica et Cosmochimica Acta, v. 57, no. 4, p. 859–874, doi:10.1016/0016-7037(93)90174-U.

Huang, S., Abouchami, W., Blichert-Toft, J., Clague, D.A., Cousens, B.L., Frey, F.A., and Humayun, M., 2009, Ancient carbonate sedimentary signature in the Hawaiian plume; evidence from Mahukona volcano, Hawaii: Geochemistry, Geophysics, Geosystems (G^3), v. 10, Q08002, 30 p., doi:10.1029/2009GC002418.

Jackson, E.D., Silver, E.A., and Dalrymple, G.B., 1972, Hawaiian-Emperor chain and its relation to Cenozoic tectonics: Bulletin of the Geological Society of America, v. 83, no. 3, p. 405–430, doi:10.1130/0016-7606(1972)83[601:HCAIRT] 2.0.CO;2.

Jaeger, J.C., 1961, The cooling of irregularly shaped igneous bodies: American Journal of Science, v. 259, no. 10, p. 721–734, doi:10.2475/ajs.259.10.72.

Jaggar, T.A., Jr., 1917a, Thermal gradient of Kilauea lava lake: Journal of the Washington Academy of Sciences, v. 7, no. 13, p. 397–405.

Jaggar, T.A., Jr., 1917b, Volcanologic investigations at Kilauea: American Journal of Science, series 4, v. 44, no. 261, article 16, p. 161–220, 1 pl., doi:10.2475/ajs.s4-44.261.161.

Jaggar, T.A., Jr., 1932, Kilauea report for June, 1932—Volcanology: The Volcano Letter, no. 388, June, p. 1. [Reprinted in Fiske, R.S., Simkin, T., and Nielsen, E., eds., 1987, The Volcano Letter: Washington, D.C., Smithsonian Institution Press, n.p.]

Jaggar, T.A., Jr., and Finch, R.H., 1924, The explosive eruption of Kilauea in Hawaii, 1924: American Journal of Science, series 5, v. 8, no. 47, article 29, p. 353–374, doi:10.2475/ajs.s5-8.47.353.

Johnson, D.J., Eggers, A.A., Bagnardi, M., Battaglia, M., Poland, M.P., and Miklius, A., 2010, Shallow magma accumulation at Kīlauea Volcano, Hawai‘i, revealed by microgravity surveys: Geology, v. 38, no. 12, p. 1139–1142, doi:10.1130/G31323.1.

Johnson, K.T.M., Reynolds, J.R., Vonderhaar, D., Smith, D.K., and Kong, L.S.L., 2002, Petrological systematics of submarine basalt glasses from the Puna Ridge, Hawai‘i; implications for rift zone plumbing and magmatic processes, *in* Takahashi, E., Lipman, P.W., Garcia, M.O, Naka, J., and Aramaki, S., eds., Hawaiian volcanoes; deep underwater perspectives: American Geophysical Union Geophysical Monograph 128, p. 143–159, doi:10.1029/GM128p0143.

Kent, A.J.R., Clague, D.A., Honda, M., Stolper, E.M., Hutcheon, I., and Norman, M.D., 1999, Widespread assimilation of a seawater-derived component at Loihi seamount, Hawaii: Geochimica et Cosmochimica Acta, v. 63, no. 18, p. 2749–2762, doi:10.1016/S0016-7037(99)00215-X.

Killingsley, J.S., and Meunow, D.M., 1975, Volatiles from Hawaiian submarine basalts determined by dynamic high temperature mass spectrometry: Geochimica et Cosmochimica Acta, v. 39, no. 11, p. 1467–1473, doi:10.1016/0016-7037(75)90148-9.

Kimura, J., Sisson, T.W., Nakano, N., Coombs, M.L., and Lipman, P.L., 2006, Isotope geochemistry of early Kilauea magmas from the submarine Hilina bench; the nature of the Hilina mantle component, *in* Coombs, M.L., Eakins, B.W., and Cervelli, P.F., eds., Growth and collapse of Hawaiian volcanoes: Journal of Volcanology and Geothermal Research, v. 151, nos. 1–3 (special issue), p. 51–72, doi:10.1016/j.jvolgeores.2005.07.024.

Kinoshita, W.T., Koyanagi, R.Y., Wright, T.L., and Fiske, R.S., 1969, Kilauea Volcano; the 1967–68 summit eruption: Science, v. 166, no. 3904, p. 459–468, doi:10.1126/science.166.3904.459.

Kurz, M.D., Jenkins, W.J., Hart, S.R., and Clague, D.A., 1983, Helium isotopic variations in volcanic rocks from Loihi Seamount and the Island of Hawaii, *in* Loihi Seamount; collected papers: Earth and Planetary Science Letters, v. 66, December, p. 388–406, doi:10.1016/0012-821X(83)90154-1.

Lange, R.L., and Carmichael, I.S.E., 1990, Thermodynamic properties of silicate liquids with emphasis on density, thermal expansion and compressibility, *in* Nicholls, J., and Russell, J.K., eds., Modern methods of igneous petrology; understanding magmatic processes: Reviews in Mineralogy, v. 24, no. 1, p. 25–64.

Leeman, W.P., Budahn, J.R., Gerlach, D.C., Smith, D.R., and Powell, B.N., 1980, Origin of Hawaiian tholeiites; trace element constraints, *in* Irving, A.J., and Dungan, M.A., eds., The Jackson volume: American Journal of Science, v. 280–A, pt. 2, p. 794–819, accessed June 4, 2013, at http://earth.geology.yale.edu/~ajs/1980/ajs_280A_1.pdf/794.pdf.

Lipman, P.W., and Coombs, M.L., 2006, North Kona slump; a submarine flank failure during the early(?) tholeiitic shield stage of Hualalai Volcano, *in* Coombs, M.L., Eakins, B.W., and Cervelli, P.F., eds., Growth and collapse of Hawaiian volcanoes: Journal of Volcanology and Geothermal Research, v. 151, nos. 1–3 (special issue), p. 189–216, doi:10.1016/j.jvolgeores.2005.07.029.

Lipman, P.W., Lockwood, J.P., Okamura, R.T., Swanson, D.A., and Yamashita, K.M., 1985, Ground deformation associated with the 1975 magnitude-7.2 earthquake and resulting changes in activity of Kilauea volcano, Hawaii: U.S. Geological Survey Professional Paper 1276, 45 p. [Also available at http://pubs.usgs.gov/pp/1276/report.pdf.]

Lipman, P.W., Sisson, T.W., Ui, T., Naka, J., and Smith, J.R., 2002, Ancestral submarine growth of Kīlauea volcano and instability of its south flank, *in* Takahashi, E., Lipman, P.W., Garcia, M.O, Naka, J., and Aramaki, S., eds., Hawaiian volcanoes; deep underwater perspectives: American Geophysical Union Geophysical Monograph 128, p. 161–191, accessed June 4, 2013, at http://www.agu.org/books/gm/v128/GM128p0161/GM128p0161.pdf.

Lockwood, J.P., and Lipman, P.W., 1987, Holocene eruptive activity of Mauna Loa Volcano, chap. 18 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 1, p. 509–535. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Lockwood, J.P., Dvorak, J.J., English, T.T., Koyanagi, R.Y., Okamura, A.T., Summers, M.L., and Tanigawa, W.R., 1987, Mauna Loa 1974–1984; a decade of intrusive and extrusive activity, chap. 19 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 1, p. 537–570. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Lockwood, J.P., Tilling, R.I., Holcomb, R.T., Klein, F., Okamura, A.T., and Peterson, D.W., 1999, Magma migration and resupply during the 1974 summit eruptions of Kīlauea Volcano, Hawai‘i: U.S. Geological Survey Professional Paper 1613, 37 p. [Also available at http://pubs.usgs.gov/pp/pp1613/pp1613.pdf.]

Macdonald, G.A., 1944, The 1840 eruption and crystal differentiation in the Kilauean magma column: American Journal of Science, v. 242, no. 4, p. 177–189, doi:10.2475/ajs.242.4.177.

Macdonald, G.A., 1949a, Petrography of the Island of Hawaii, *in* Shorter contributions to general geology: U.S. Geological Survey Professional Paper 214–D, p. 51–96, folded map, scale 1:316,800. [Also available at http://pubs.usgs.gov/pp/0214d/report.pdf.]

Macdonald, G.A., 1949b, Hawaiian petrographic province: Geological Society of America Bulletin, v. 60, no. 10, p. 1541–1596, doi:10.1130/0016-7606(1949)60[1541:HPP]2.0.CO;2.

Macdonald, G.A., 1952, The 1952 eruption of Kilauea: The Volcano Letter, no. 518, October–December, p. 1–10. (Reprinted in Fiske, R.S., Simkin, T., and Nielsen, E., eds., 1987, The Volcano Letter: Washington, D.C., Smithsonian Institution Press, n.p.)

Macdonald, G.A., 1954, The eruption of Kilauea volcano in May 1954: The Volcano Letter, no. 524, April–June, p. 1–11. (Reprinted in Fiske, R.S., Simkin, T., and Nielsen, E.A., eds., 1987, The Volcano Letter: Washington, D.C., Smithsonian Institution Press, n.p.)

Macdonald, G.A., 1955, The eruption of Kilauea; brief history of activity in Kilauea caldera, *in* Hawaiian volcanoes during 1952: U.S. Geological Survey Bulletin 1021–B, p. 53–105, 14 pls. [Also available at http://pubs.usgs.gov/bul/1021b/report.pdf.]

Macdonald, G.A., and Abbott, A.T., 1970, Volcanoes in the sea; the geology of Hawaii: Honolulu, University of Hawai‘i Press, 441 p.

Macdonald, G.A., and Eaton, J.P., 1957, Hawaiian volcanoes during 1954: U.S. Geological Survey Bulletin 1061–B, p. 17–72, 1 pl. [Also available at http://pubs.usgs.gov/bul/1061b/report.pdf.]

Macdonald, G.A., and Eaton, J.P., 1964, Hawaiian volcanoes during 1955: U.S. Geological Survey Bulletin 1171, 170 p., 5 pls. [Also available at http://pubs.usgs.gov/bul/1171/report.pdf.]

Macdonald, G.A., and Katsura, T., 1964, Chemical composition of Hawaiian lavas: Journal of Petrology, v. 5, no. 1, p. 82–133, doi:10.1093/petrology/5.1.82.

Macdonald, G.A., and Orr, J.B., 1950, The 1949 summit eruption of Mauna Loa, Hawaii: U.S. Geological Survey Bulletin 974–A, 33 p., accessed June 4, 2013, at https://archive.org/stream/1949summiterupti00macd#page/10/mode/2up.

Mangan, M.T., 1990, Crystal size distribution systematics and the determination of magma storage times; the 1959 eruption of Kilauea volcano, Hawaii: Journal of Volcanology and Geothermal Research, v. 44, nos. 3–4, p. 295–302, doi:10.1016/0377-0273(90)90023-9.

Mastin, L.G., Christiansen, R.L., Thornber, C., Lowenstern, J., and Beeson, M., 2004, What makes hydromagmatic eruptions violent? Some insights from the Keanakāko'i Ash, Kīlauea Volcano Hawai'i: Journal of Volcanology and Geothermal Research, v. 137, nos. 1–3, p. 115–31, doi:10.1016/j.jvolgeores.2004.05.015.

Matveenkov, V.V., and Sorokhtin, O.G., 1998, Petrologic peculiarities of the initial stages of development of intraplate volcanism of the Loihi Island (Hawaiian Archipelago): Oceanology, v. 38, no. 5, p. 671–678.

McCarter, R.L., Fodor, R.V., and Trusdell, F., 2006, Perspectives on basaltic magma crystallization and differentiation; lava-lake blocks erupted at Mauna Loa volcano summit, Hawaii: Lithos, v. 90, nos. 3–4, p. 187–213, doi:10.1016/j.lithos.2006.03.005.

McPhie, J., Walker, G.P.L., and Christiansen, R.L., 1990, Phreatomagmatic and phreatic fall and surge deposits from explosions at Kilauea volcano, Hawaii, 1790 A.D.; Keanakakoi Ash Member: Bulletin of Volcanology, v. 52, no. 5, p. 334–354, doi:10.1007/BF00302047.

Meunow, D.W., Graham, D.G., Liu, N.W.K., and Delaney, J.R., 1979, The abundance of volatiles in Hawaiian tholeiitic submarine basalts: Earth and Planetary Science Letters, v. 42, no. 1, p. 71–76, doi:10.1016/0012-821X(79)90191-2.

Montierth, C., Johnston, A.D., and Cashman, K.V., 1995, An empirical glass-composition-based geothermometer for Mauna Loa lavas, *in* Rhodes, J.M., and Lockwood, J.P., eds., Mauna Loa revealed; structure, composition, history, and hazards: American Geophysical Union Geophysical Monograph 92, p. 207–217, doi:10.1029/GM092p0207.

Moore, J.G., 1965, Petrology of deep-sea basalt near Hawaii: American Journal of Science, v. 263, no. 1, p. 40–52, doi:10.2475/ajs.263.1.40.

Moore, J.G., and Clague, D.A., 1992, Volcano growth and evolution of the island of Hawaii: Geological Society of America Bulletin, v. 104, no. 11, p. 1471–1484, doi:10.1130/0016-7606(1992)10.

Moore, J.G., and Evans, B.W., 1967, The role of olivine in the crystallization of the prehistoric Makaopuhi tholeiitic lava lake, Hawaii: Contributions to Mineralogy and Petrology, v. 15, no. 3, p. 202–223, doi:10.1007/BF01185342.

Moore, J.G., and Fabbi, B.P., 1971, An estimate of the juvenile sulfur content of basalt: Contributions to Mineralogy and Petrology, v. 33, no. 2, p. 118–127, doi:10.1007/BF00386110.

Moore, J.G., and Fiske, R.S., 1969, Volcanic substructure inferred from dredge samples and ocean-bottom photographs, Hawaii: Geological Society of America Bulletin, v. 80, no. 7, p. 1191–1202, doi:10.1130/0016-7606(1969)80[1191:VSIFDS] 2.0.CO;2.

Moore, J.G., Clague, D.A., and Normark, W.R., 1982, Diverse basalt types from Loihi seamount, Hawaii: Geology, v. 10, no. 2, p. 88–92, doi:10.1130/0091-7613(1982).

Moore, J.G., Clague, D.A., Holcomb, R.T., Lipman, P.W., Normark, W.R., and Torresan, M.E., 1989, Prodigious submarine landslides on the Hawaiian Ridge: Journal of Geophysical Research, v. 94, no. B12, p. 17465–17484, doi:10.1029/JB094iB12p17465.

Moore, R.B., Helz, R.T., Dzurisin, D., Eaton, G.P., Koyanagi, R.Y., Lipman, P.W., Lockwood, J.P., and Puniwai, G.S., 1980, The 1977 eruption of Kilauea Volcano, Hawaii, *in* McBirney, A.R., ed., Gordon A. Macdonald Memorial Volume: Journal of Volcanology and Geothermal Research, v. 7, nos. 3–4 (special issue), p. 189–210, doi:10.1016/0377-0273(80)90029-3.

Morgan, J.K., Clague, D.A., Borchers, D.C., Davis, A.S., and Milliken, K.L., 2007, Mauna Loa's submarine western flank; landsliding, deep volcanic spreading, and hydrothermal alteration: Geochemistry, Geophysics, Geosystems (G^3), v. 8, Q05002, 42 p., doi:10.1029/2006GC001420.

Muir, I.D., and Tilley, C.E., 1957, Contributions to the petrology of Hawaiian basalts, [part] 1. The picrite-basalts of Kilauea: American Journal of Science, v. 255, no. 4, p. 241–253, doi:10.2475/ajs.255.4.241.

Murata, K.J., 1970, Tholeiitic basalt magmatism of Kilauea and Mauna Loa volcanoes of Hawaii: Die Naturwissenschaften, v. 57, no. 3, p. 108–113, doi:10.1007/BF00600044.

Murata, K.J., and Richter, D.H., 1961, Magmatic differentiation in the Uwekahuna laccolith, Kilauea caldera, Hawaii: Journal of Petrology, v. 2, no. 3, p. 424–437, doi:10.1093/petrology/2.3.424.

Murata, K.J., and Richter, D.H., 1966, Chemistry of the lavas of the 1959–60 eruption of Kilauea Volcano, Hawaii, *in* The 1959–60 eruption of Kilauea volcano, Hawaii: U.S. Geological Survey Professional Paper 537–A, p. A1–A26. [Also available at http://pubs.usgs.gov/pp/0537a/report.pdf.]

Nicholls, J., and Stout, M.Z., 1988, Picritic melts in Kilauea—Evidence from the 1967–68 Halemaumau and Hiiaka eruptions: Journal of Petrology, v. 29, no. 5, p. 1031–1058, doi:10.1093/petrology/29.5.1031.

Ochs, F.A., III, and Lange, R.A., 1999, The density of hydrous magmatic liquids: Science, v. 283, no. 5406, p. 1314–1318, doi:10.1126/science.283.5406.1314.

Okubo, P.G., Nakata, J.S., and Koyanagi, R.Y., 2014, The evolution of seismic monitoring systems at the Hawaiian Volcano Observatory, chap. 2 *of* Poland, M.P., Takahashi, T.J., and Landowski, C.M., eds., Characteristics of Hawaiian volcanoes: U.S. Geological Survey Professional Paper 1801 (this volume).

Parman, S.W., Kurz, M.D., Hart, S.R., and Grove, T.L., 2005, Helium solubility in olivine and implications for high ^{3}He/^{4}He in ocean island basalts: Nature, v. 437, no. 7062, p. 1140–1143, doi:10.1038/nature04215.

Peck, D.L., 1978, Cooling and vesiculation of Alae lava lake, Hawaii, *in* Solidification of Alae lava lake, Hawaii: U.S. Geological Survey Professional Paper 935–B, 59 p. [Also available at http://pubs.usgs.gov/pp/0935b/report.pdf.]

Peck, D.L., and Kinoshita, W.T., 1976, The eruption of August 1963 and the formation of Alae lava lake, Hawaii, *in* Solidification of Alae lava lake, Hawaii: U.S. Geological Survey Professional Paper 935–A, p. A1–A33. [Also available at http://pubs.usgs.gov/pp/0935a/report.pdf.]

Peck. D.L., Wright, T.L., and Moore, J.G., 1966, Crystallization of tholeiitic basalt in Alae lava lake, Hawaii: Bulletin Volcanologique, v. 29, no. 1, p. 629–656, doi:10.1007/BF02597182.

Peck, D.L., Hamilton, M.S., and Shaw, H.R., 1977, Numerical analysis of lava lake cooling models; part II, Application to Alae lava lake, Hawaii: American Journal of Science, v. 277, no. 4, p. 415–437, doi:10.2475/ajs.277.4.415.

Peterson, D.W., and Moore, R.B., 1987, Geologic history and evolution of geologic concepts, Island of Hawaii, chap. 7 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 1, p. 149–189. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Pietruszka, A.J., and Garcia, M.O., 1999a, A rapid fluctuation in the mantle source and melting history of Kilauea Volcano inferred from the geochemistry of its historical summit lavas (1790–1982): Journal of Petrology, v. 40, no. 8, p. 1321–1342, doi:10.1093/petroj/40.8.1321.

Pietruszka, A.J., and Garcia, M.O., 1999b, The size and shape of Kilauea Volcano's summit magma storage reservoir; a geochemical probe: Earth and Planetary Science Letters, v. 167, nos. 3–4, p. 311–320, doi:10.1016/S0012-821X(99)00036-9.

Pietruszka, A.J., Rubin, K.H., and Garcia, M.O., 2001, ^{226}Ra-^{230}Th -^{238}U disequilibria of historical Kilauea lavas (1790–1982) and the dynamics of mantle melting within the Hawaiian plume: Earth and Planetary Science Letters, v. 186, no. 1, p. 15–31, doi:10.1016/S0012-821X(01)00230-8.

Pitcher, L., Helz, R.T., Walker, R.J., and Piccoli, P., 2009, Fractionation of the platinum-group elements and Re during crystallization of basalt in Kilauea Iki lava lake, Hawaii: Chemical Geology, v. 260, nos. 3–4, p. 196–210, doi:10.1016/j.chemgeo.2008.12.022.

Poland, M.P., Miklius, A., Sutton, A.J., and Thornber, C.R., 2012, A mantle-driven surge in magma supply to Kīlauea Volcano during 2003–2007: Nature Geoscience, v. 5, no. 4, p. 295–300 (supplementary material, 16 p.), doi:10.1038/ngeo1426.

Poland, M.P., Miklius, A., and Montgomery-Brown, E.K., 2014, Magma supply, storage, and transport at shield-stage Hawaiian volcanoes, chap. 5 *of* Poland, M.P., Takahashi, T.J., and Landowski, C.M., eds., Characteristics of Hawaiian volcanoes: U.S. Geological Survey Professional Paper 1801 (this volume).

Powers, H.A., 1955, Composition and origin of basaltic magma of the Hawaiian Islands: Geochimica et Cosmochimica Acta, v. 7, nos. 1–2, p. 77–107, doi:10.1016/0016-7037(55)90047-8.

Putirka, E., Ryerson, F.J., Perfit, M., and Ridley, W.I., 2011, Mineralogy and composition of the oceanic mantle: Journal of Petrology, v. 52, no. 2, p. 279–313, doi:10.1093/petrology/egq080.

Rhodes, J.M., 1988, Geochemistry of the 1984 Mauna Loa eruption; implications for magma storage and supply: Journal of Geophysical Research, v. 93, no. B5, p. 4453–4466, doi:10.1007/BF00371276.

Rhodes, J.M., 1995, The 1852 and 1868 Mauna Loa picrite eruptions; clues to parental magma compositions and the magmatic plumbing system, *in* Rhodes, J.M., and Lockwood, J.P., eds., Mauna Loa revealed; structure, composition, history, and hazards: American Geophysical Union Geophysical Monograph 92, p. 241–262, doi:10.1029/GM092p0241.

Rhodes, J.M., and Hart, S.R., 1995, Episodic trace element and isotopic variations in historical Mauna Loa lavas; implications for magma and plume dynamics, *in* Rhodes, J.M., and Lockwood, J.P., eds., Mauna Loa revealed; structure, composition, history, and hazards: American Geophysical Union Geophysical Monograph 92, p. 263–288, doi:10.1029/GM092p0263.

Richter, D.H., and Moore, J.G., 1966, Petrology of the Kilauea Iki lava lake, Hawaii, *in* The 1959–60 eruption of Kilauea volcano, Hawaii: U.S. Geological Survey Professional Paper 537–B, p. B1–B26. [Also available at http://pubs.usgs.gov/pp/0537/report.pdf.]

Richter, D.H., Eaton, J.P., Murata, J., Ault, W.U., and Krivoy, H.L., 1970, Chronological narrative of the 1959–60 eruption of Kilauea Volcano, Hawaii, *in* The 1959–60 eruption of Kilauea volcano, Hawaii: U.S. Geological Survey Professional Paper 537–E, p. E1–E73. [Also available at http://pubs.usgs.gov/pp/0537e/report.pdf.]

Riker, J.M., Cashman, K.V., Kauahikaua, J.P., and Montierth, C.M., 2009, The length of channelized lava flows; insight from the 1959 eruption of Mauna Loa Volcano, Hawai‘i: Journal of Volcanology and Geothermal Research, v. 183, nos. 3–4, p. 139–156, doi:10.1016/j.jvolgeores.2009.03.002.

Roden, M.F., Trull, T., Hart, S.R., and Frey, F.A., 1994, New He, Nd, Pb, and Sr isotopic constraints on the constitution of the Hawaiian plume; results from Koolau Volcano, Oahu, Hawaii, USA: Geochimica et Cosmochimica Acta, v. 58, no. 5, p. 1431–1440, doi:10.1016/0016-7037(94)90547-9.

Roeder, P.L., and Emslie, R.F., 1970, Olivine-liquid equilibrium: Contributions to Mineralogy and Petrology, v. 29, no. 4, p. 275–289, doi:10.1007/BF00371276.

Roeder, P., Gofton, E., and Thornber, C., 2006, Cotectic proportions of olivine and spinel in olivine-tholeiitic basalt and evaluation of pre-eruptive processes: Journal of Petrology, v. 47, no. 5, p. 883–900, doi:10.1093/petrology/egi099.

Rose, T.R., Fiske, R.S., and Swanson, D.A., 2011, High-MgO vitric ash in Upper Kulanaokuaiki Tephra, Kīlauea Volcano, Hawai‘i; a preliminary description [abs.]: American Geophysical Union, Fall Meeting 2011 Abstracts, abstract no. V41A–2481, accessed April 28, 2014, at http://abstractsearch.agu.org/meetings/2011/FM/sections/V/sessions/V41A/abstracts/V41A-2481.html.

Rowe, M.C., Thornber, C.R., and Orr, T.R., 2009, Kīlauea 2008; primitive components and degassing recorded by olivine-hosted melt inclusions and matrix glasses [abs.]: Geochimica et Cosmochimica Acta, v. 73, no. 13, supp. 1 (Goldschmidt 2009 Abstracts), p. A1126.

Russell, J.K., and Stanley, C.R., 1990, Origins of the 1954–1960 lavas, Kilauea Volcano, Hawaii; major element constraints on shallow reservoir magmatic processes: Journal of Geophysical Research, v. 95, no. B4, p. 5021–5047, doi:10.1029/JB095iB04p05021.

Ryan, M.P., 1979, High-temperature mechanical properties of basalt: University Park, Pa., The Pennsylvania State University, Ph.D. dissertation, 417 p.

Ryan, M.P., 1987, Neutral buoyancy and the mechanical evolution of magmatic systems, *in* Mysen, B.O., ed., Magmatic processes; physicochemical principles: Geochemical Society Special Publication No. 1 (A volume in honor of Hatten S. Yoder, Jr.), p. 259–287.

Ryan, M.P., 1988, The mechanics and three-dimensional internal structure of active magmatic systems; Kilauea Volcano, Hawaii: Journal of Geophysical Research, v. 93, no. B5, p. 4213–4248, doi:10.1029/JB093iB05p04213.

Ryan, M.P., Blevins, J.Y.K., Okamura, A.T., and Koyanagi, R.Y., 1983, Magma reservoir subsidence mechanics; theoretical summary and application to Kilauea Volcano, Hawaii: Journal of Geophysical Research, v. 88, no. B5, p. 4147–4181, doi:10.1029/JB088iB05p04147.

Schwindinger, K.R., and Anderson, A.T., Jr., 1989, Synneusis of Kilauea Iki olivines: Contributions to Mineralogy and Petrology, v. 103, no. 2, p. 187–198, doi:10.1007/BF00378504.

Scowen, P.A.H., Roeder, P.L., and Helz, R.T., 1991, Re-equilibration of chromite within Kilauea Iki lava lake, Hawaii: Contributions to Mineralogy and Petrology, v. 107, no. 1, p. 8–20, doi:10.1007/BF00311181.

Sio, C.K., Dauphas, N., Teng, F.-Z., Helz, R.T., and Chaussidon, M., 2011, In-situ Fe-Mg isotopic analysis of zoned olivine [abs.]: Mineralogical Magazine, v. 75, no. 3 (Goldschmidt 2011 Abstracts), p. 1884, accessed June 4, 2013, at http://goldschmidt.info/2011/abstracts/finalPDFs/1884.pdf.

Sisson, T.W., 2003, Native gold in a Hawaiian alkalic magma: Economic Geology, v. 98, no. 3, p. 643–648, doi:10.2113/gsecongeo.98.3.643.

Sisson, T.W., Lipman, P.W., and Naka, J., 2002, Submarine alkalic through tholeiitic shield-stage development of Kīlauea volcano, Hawai‘i, *in* Takahashi, E., Lipman, P.W., Garcia, M.O, Naka, J., and Aramaki, S., eds., Hawaiian volcanoes; deep underwater perspectives: American Geophysical Union Geophysical Monograph 128, p. 193–219, doi:10.1029/GM128p0193.

Sisson, T.W., Kimura, J.-I., and Coombs, M.L., 2009, Basanite-nephelinite suite from early Kilauea; carbonated melts of phlogopite-garnet peridotite at Hawaii’s leading magmatic edge: Contributions to Mineralogy and Petrology, v. 158, no. 6, p. 803–829, doi:10.1007/s00410-009-0411-8.

Smith, J.R., Malahoff, A., and Shor, A.N., 1999, Submarine geology of the Hilina slump and morpho-structural evolution of Kilauea volcano, Hawaii: Journal of Volcanology and Geothermal Research, v. 94, no. 1, p. 59–88, doi:10.1016/S0377-0273(99)00098-0.

Sobolev, A.V., Hofmann, A.W., Soboloev, S.V., and Nikogosian, I.K., 2005, An olivine-free mantle source of Hawaiian shield basalts: Nature, v. 434, no. 7033, p. 590–597, doi:10.1038/nature03411.

Sparks, J.W., 1990, Long-term compositional and eruptive behavior of Mauna Loa Volcano; evidence from prehistoric caldera basalts: Amherst, University of Massachusetts, Ph.D. dissertation, 216 p., paper AAI9110218.

Sparks, R.S.J., 1978, The dynamics of bubble formation and growth in magmas; a review and analysis: Journal of Volcanology and Geothermal Research, v. 3, nos. 1–2, p. 1–37, doi:10.1016/0377-0273(78)90002-1.

Sparks, R.S.J., and Huppert, H.E., 1984, Density changes during the fractional crystallization of basaltic magmas; fluid dynamic implications: Contributions to Mineralogy and Petrology, v. 85, no. 3, p. 300–309, doi:10.1007/BF00378108.

Sparks, R.S.J., Meyer, P., and Sigurdsson, H., 1980, Density variation amongst mid-ocean ridge basalts; implications for magma mixing and the scarcity of primitive lava: Earth and Planetary Science Letters, v. 46, no. 3, p. 419–430, doi:10.1016/0012-821X(80)90055-2.

Staudigel, H., Zindler, A., Hart, S.R., Leslie, T., Chen, C.-Y., and Clague, D., 1984, The isotope systematics of a juvenile intraplate volcano; Pb, Nd, and Sr isotope ratios of basalts from Loihi Seamount, Hawaii: Earth and Planetary Science Letters, v. 69, no. 1, p. 13–29, doi:10.1016/0012-821X(84)90071-2.

Stille, P., Unruh, D.M., and Tatsumoto, M., 1986, Pb, Sr, Nd, and Hf isotopic constraints on the origin of Hawaiian basalts and evidence for a unique mantle source: Geochimica et Cosmochimica Acta, v. 50, no. 10, p. 2303–2319, doi:10.1016/0016-7037(86)90084-0.

Stolper, E., and Walker, D., 1980, Melt density and the average composition of basalt: Contributions to Mineralogy and Petrology, v. 74, no. 1, p. 7–12, doi:10.1007/BF00375484.

Swanson, D.A., Duffield, W.A., Jackson, D.B., and Peterson, D.W., 1979, Chronological narrative of the 1969–71 Mauna Ulu eruption of Kilauea Volcano, Hawaii: U.S. Geological Survey Professional Paper 1056, 55 p., 4 pls. [Also available at http://pubs.usgs.gov/pp/1056/report.pdf.]

Swanson, D.A., Rose, T.R., Fiske, R.S., and McGeehin, J.P., 2012, Keanakāko‘i Tephra produced by 300 years of explosive eruptions following collapse of Kīlauea’s caldera in about 1500 CE: Journal of Volcanology and Geothermal Research, v. 215–216, February 15, p. 8–25, doi:10.1016/j.jvolgeores.2011.11.009.

Tait, S.R., and Jaupart, C., 1996, The production of chemically stratified and adcumulate plutonic igneous rocks: Mineralogical Magazine, v. 60, no. 398, p. 99–114, doi:10.1180/minmag.1996.060.398.07.

Tait, S.R., Jahrling, K., and Jaupart, C., 1992, The planform of compositional convection and chimney formation in a mushy layer: Nature, v. 359, no. 6394, p. 406–408, doi:10.1038/359406a0.

Takahashi, E., and Nakajima, K., 2002, Melting processes in the Hawaiian plume; an experimental study, *in* Takahashi, E., Lipman, P.W., Garcia, M.O., Naka, J., and Aramaki, S., eds., Hawaiian volcanoes; deep underwater perspectives: American Geophysical Union Geophysical Monograph 128, p. 403–418, doi:10.1029/GM128p0403.

Tatsumoto, M., 1978, Isotopic composition of lead in oceanic basalt and its implication to mantle evolution: Earth and Planetary Science Letters, v. 38, no. 1, p. 63–87, doi:10.1016/0012-821X(78)90126-7.

Teng, F.-Z., Dauphas, N., and Helz, R.T., 2008, Iron isotope fractionation during magmatic differentiation in Kilauea Iki lava lake: Science, v. 320, no. 5883, p. 1620–1622, doi:10.1126/science.1157166.

Teng, F.-Z., Dauphas, N., Helz, R.T., Gao, S., and Huang, S., 2011, Diffusion-driven magnesium and iron isotope fractionation in Hawaiian olivine: Earth and Planetary Science Letters, v. 308, nos. 3–4, p. 317–324, doi:10.1016/j.epsl.2011.06.003.

Teplow, W., Marsh, B., Hulen, J., Spielman, P., Kaleikini, M., Fitch, D., and Rickard, W., 2009, Dacite melt at the Puna Geothermal Venture wellfield, Big Island of Hawaii: Geothermal Resources Council Transactions, v. 33, p. 989–994 [Geothermal 2009 Annual Meeting, Reno, Nev., October 4–7, 2009].

Thompson, R.N., and Tilley, C.E., 1969, Melting and crystallisation relations of Kilauean basalts of Hawaii; the lavas of the 1959–60 Kilauea eruption: Earth and Planetary Science Letters, v. 5, p. 469–477, doi:10.1016/S0012-821X(68)80081-0.

Thornber, C.R., 2001, Olivine-liquid relations of lava erupted by Kīlauea Volcano from 1994–1998; implications for shallow magmatic processes associated with the ongoing east-rift-zone eruption: Canadian Mineralogist, v. 39, no. 2, p. 239–266, doi:10.2113/gscanmin.39.2.239.

Thornber, C.R., 2003, Magma-reservoir processes revealed by geochemistry of the Pu‘u ‘Ō‘ō-Kūpaianaha eruption, *in* Heliker, C., Swanson, D.A., and Takahashi, T.J., eds., The Pu‘u ‘Ō‘ō-Kūpaianaha eruption of Kīlauea Volcano, Hawai‘i; the first 20 years: U.S. Geological Survey Professional Paper 1676, p. 121–136. [Also available at http://pubs.usgs.gov/pp/pp1676/.]

Thornber, C.R., Hon, K., Heliker, C., and Sherrod, D.A., 2003a, A compilation of whole-rock and glass major-element geochemistry of Kīlauea Volcano, Hawai‘i, near-vent eruptive products; January 1983 through September 2001: U.S. Geological Survey Open-File Report 03–477, 8 p. [Also available at http://pubs.usgs.gov/of/2003/0477/pdf/of03–477.pdf.]

Thornber, C.R., Heliker, C., Sherrod, D.R., Kauahikaua, J.P., Miklius, A., Okubo, P.G., Trusdell, F.A., Budahn, J.R., Ridley, W.I., and Meeker, G.P., 2003b, Kilauea east rift zone magmatism; an episode 54 perspective: Journal of Petrology, v. 44, no. 9, p. 1525–1559, doi:10.1093/petrology/egg048.

Thornber, C.R., Budahn, J.R., Ridley, W.I., and Unruh, D.M., 2003c, Trace element and Nd, Sr, Pb isotope geochemistry of Kīlauea Volcano, Hawaiʻi, near-vent eruptive products, 1983–2001: U.S. Geological Survey Open-File Report 2003–493, 5 p. [Also available at http://pubs.usgs.gov/of/2003/0493/pdf/of03-493.pdf.]

Thornber, C., Orr, T., Lowers, H., Heliker, C., and Hoblitt, R., 2007, An episode 56 perspective on post-2001 comagmatic mixing along Kilauea's east rift zone [abs.]: American Geophysical Union, Fall Meeting 2007 Abstracts, abstract no. V51H–04, accessed April 28, 2014, at http://abstractsearch.agu.org/meetings/2007/FM/sections/V/sessions/V51H/abstracts/V51H-04.html.

Thornber, C.R., Rowe, M.C., Adams, D.B., and Orr, T.R., 2010, Application of microbeam techniques to identifying and assessing comagmatic mixing between summit and rift eruptions at Kilauea Volcano [abs.]: American Geophysical Union, Fall Meeting 2010 Abstracts, abstract no. V43F–01, accessed April 28, 2014, at http://abstractsearch.agu.org/meetings/2010/FM/sections/V/sessions/V43F/abstracts/V43F-01.html.

Tilley, C.E., 1960, Differentiation of Hawaiian basalts; some variants in lava suites of dated Kilauean eruptions: Journal of Petrology, v. 1, no. 1, p. 47–55, doi:10.1093/petrology/1.1.47.

Tilley, C.E., and Scoon, J.H., 1961, Differentiation of Hawaiian basalts; trends of Mauna Loa and Kilauea historic magmas: American Journal of Science, v. 259, no. 1, p. 60–68, doi:10.2475/ajs.259.1.60.

Tilling, R.I., Koyanagi, R.Y., Lipman, P.W., Lockwood, J.P., Moore, J.G., and Swanson, D.A., 1976, Earthquake and related catastrophic events, island of Hawaii, November 29, 1975; a preliminary report: U.S. Geological Survey Circular 740, 33 p. [Also available at http://pubs.usgs.gov/circ/1976/0740/report.pdf.]

Tilling, R.I., Rhodes, J.M., Sparks, J.W., Lockwood, J.P., and Lipman, P.W., 1987a, Disruption of the Mauna Loa magma system by the 1868 Hawaiian earthquake; geochemical evidence: Science, v. 235, no. 4785, p. 196–199, doi:10.1126/science.235.4785.196.

Tilling, R.I., Wright, T.L., and Millard, H.T., Jr., 1987b, Trace-element chemistry of Kīlauea and Mauna Loa lava in space and time; a reconnaissance, chap. 24 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 1, p. 641–689. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Tilling, R.I., Christiansen, R.L., Duffield, W.A., Endo, E.T., Holcomb, R.T., Koyanagi, R.Y., Peterson, D.W., and Unger, J.D., 1987c, The 1972–1974 Mauna Ulu eruption, Kilauea Volcano; an example of quasi-steady-state magma transfer, chap. 16 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 1, p. 405–469. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Trieloff, M., Kunz, J., Clague, D.A., Harrison, D., and Allègre, C.J., 2000, The nature of pristine noble gases in mantle plumes: Science, v. 288, no. 5468, p. 1036–1038, doi:10.1126/science.288.5468.1036.

Trieloff, M., Kunz, J., Clague, D.A., Harrison, D., and Allègre, C.J., 2001, Reply to comment on "the nature of pristine noble gases in mantle plumes by C.J. Ballentine, D. Porcelli, and R. Weiler: Science, v. 291, no. 5512, p., article 2269a, doi:10.1126/science.291.5512.2269a, accessed June 4, 2013, at www.sciencemag.org/cgi/content/full/291/5512/2269a.

Vinet, N., and Higgins, M.D., 2010, Magma solidification processes beneath Kilauea Volcano, Hawaii; a quantitative textural and geochemical study of the 1969–1974 Mauna Ulu lavas: Journal of Petrology, v. 51, no. 6, p. 1297–1332, doi:10.1093/petrology/egq020.

Vinet, N., and Higgins, M.D., 2011, What can crystal size distributions and olivine compositions tell us about magma solidification processes inside Kilauea Iki lava lake, Hawaii?: Journal of Volcanology and Geothermal Research, v. 208, nos. 3–4, p. 136–162, doi:10.1016/j.jvolgeores.2011.09.006.

Wagner, T.P., Clague, D.A., Hauri, E.H., and Grove, T.L., 1998, Trace-element abundances of high-MgO glasses from Kilauea, Mauna Loa, and Haleakala volcanoes, Hawaii: Contributions to Mineralogy and Petrology, v. 131, no. 1, p. 13–21, doi:10.1007/s004100050375.

Wallace, P.J., and Anderson, A.T., Jr., 1998, Effects of eruption and lava drainback on the H_2O contents of basaltic magmas at Kilauea Volcano: Bulletin of Volcanology, v. 59, no. 5, p. 327–344, doi:10.1007/s004450050195.

Wanless, V.D., Garcia, M.O., Rhodes, J.M., Weis, D., and Norman, M.D., 2006a, Shield-stage alkalic volcanism on Mauna Loa Volcano, Hawaii, *in* Coombs, M.L., Eakins, B.W., and Cervelli, P.F., eds., Growth and collapse of Hawaiian volcanoes: Journal of Volcanology and Geothermal Research, v. 151, nos. 1–3 (special issue), p. 141–155, doi:10.1016/j.jvolgeores.2005.07.027.

Wanless, V.D., Garcia, M.O., Trusdell, F.A., Rhodes, J.M., Norman, M.D., Weis, D., Fornari, D.J., Kurz, M.D., and Guillou, H., 2006b, Submarine radial vents on Mauna Loa Volcano, Hawai`i: Geochemistry, Geophysics, Geosystems (G^3), v. 7, no. 5, Q05001, 28 p., doi:10.1029/2005GC001086.

West, H.B., Gerlach, D.C., Leeman, W.P., and Garcia, M.O., 1987, Isotopic constraints on the origin of Hawaiian lavas from the Maui Volcanic Complex, Hawaii: Nature, v. 330, no. 6145, p. 216–219, doi:10.1038/330216a0.

Wilkinson, J.F.G., and Hensel, H.D., 1988, The petrology of some picrites from Mauna Loa and Kilauea volcanoes, Hawaii: Contributions to Mineralogy and Petrology, v. 98, no. 3, p. 326–245, doi:10.1007/BF00375183.

Wolfe, E.W., Garcia, M.O., Jackson, D.B., Koyanagi, R.Y., Neal, C.A., and Okamura, A.T., 1987, The Puu Oo eruption of Kilauea Volcano, episodes 1–20, January 3, 1983, to June 8, 1984, chap. 17 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 1, p. 471–508. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Wolfe, E.W., Neal, C.R., Banks, N.G., and Duggan, T.J., 1988, Geologic observations and chronology of eruptive events, chap. 1 *of* Wolfe, E.W., ed., The Puu Oo eruption of Kilauea Volcano, Hawaii; episodes 1 through 20, January 3, 1983, through June 8, 1984: U.S. Geological Survey Professional Paper 1463, p. 1–98. [Also available at http://pubs.usgs.gov/pp/1463/report.pdf.]

Wooten, K.M., Thornber, C.R., Orr, T.R., Ellis, J.F., and Trusdell, F.A., 2009, Catalog of tephra samples from Kīlauea's summit eruption, March–December 2008: U.S. Geological Survey Open-File Report 2009–1134, 26 p. and database. [Also available at http://pubs.usgs.gov/of/2009/1134/of2009-1134.pdf.]

Wright, T.L., 1971, Chemistry of Kilauea and Mauna Loa lava in space and time: U.S. Geological Survey Professional Paper 735, 40 p. [Also available at http://pubs.usgs.gov/pp/0735/report.pdf.]

Wright, T.L., 1973, Magma mixing as illustrated by the 1959 eruption, Kilauea Volcano, Hawaii: Geological Society of America Bulletin, v. 84, no. 3, p. 849–858, doi:10.1130/0016-7606(1973).

Wright, T.L., 1984, Origin of Hawaiian tholeiite; a metasomatic model: Journal of Geophysical Research, v. 89, no. B5, p. 3233–3252, doi:10.1029/JB089iB05p03233.

Wright, T.L., and Fiske, R.S., 1971, Origin of the differentiated and hybrid lavas of Kilauea Volcano, Hawaii: Journal of Petrology, v. 12, no. 1, p. 1–65, doi:10.1093/petrology/12.1.1.

Wright, T.L., and Helz, R.T., 1987, Recent advances in Hawaiian petrology and geochemistry, chap. 23 *of* Decker, R.W., Wright, T.L., and Stauffer, P.H., eds., Volcanism in Hawaii: U.S. Geological Survey Professional Paper 1350, v. 1, p. 625–640. [Also available at http://pubs.usgs.gov/pp/1987/1350/.]

Wright, T.L., and Helz, R.T., 1996, Differentiation and magma mixing on Kilauea's east rift zone; a further look at the eruptions of 1955 and 1960. Part II. The 1960 lavas: Bulletin of Volcanology, v. 57, no. 8, p. 602–630, doi:10.1007/s004450050115.

Wright, T.L., and Klein, F.W., 2006, Deep magma transport at Kilauea volcano, Hawaii: Lithos, v. 87, nos. 1–2, p. 50–79, doi:10.1016/j.lithos.2005.05.004.

Wright, T.L., and Klein, F.W., 2010, Magma transport and storage at Kilauea Volcano, HI [abs.]: American Geophysical Union, Fall Meeting 2010 Abstracts, abstract no. V43C–2401, accessed April 28, 2014, at http://abstractsearch.agu.org/meetings/2010/FM/sections/V/sessions/V43C/abstracts/V43C-2401.html.

Wright, T.L., and Okamura, R., 1977, Cooling and crystallization of tholeiitic basalt, 1965 Makaopuhi lava lake, Hawaii: U.S. Geological Survey Professional Paper 1004, 78 p. [Also available at http://pubs.usgs.gov/pp/1004/report.pdf.]

Wright, T.L., and Peck, D.L., 1978, Crystallization and differentiation of the Alae magma, Alae lava lake, Hawaii, *in* Solidification of Alae lava lake, Hawaii: U.S. Geological Survey Professional Paper 935–C, p. C1–C20. [Also available at http://pubs.usgs.gov/pp/0935c/report.pdf.]

Wright, T.L., and Tilling, R.I., 1980, Chemical variations in Kilauea eruptions 1971–1974, *in* Irving, A.J., and Dungan, M.A., eds., The Jackson Volume: American Journal of Science, v. 280–A, pt. 1, p. 777–793, accessed June 4, 2013, at http://earth.geology.yale.edu/~ajs/1980/ajs_280A_1.pdf/777.pdf.

Wright, T.L., Kinoshita, W.T., and Peck, D.L., 1968, March 1965 eruption of Kilauea volcano and the formation of Makaopuhi lava lake: Journal of Geophysical Research, v. 73, no. 10, p. 3181–3205, doi:10.1029/JB073i010p03181.

Wright, T.L., Swanson, D.A., and Duffield, W.A., 1975, Chemical compositions of Kilauea east-rift lava, 1968–1971: Journal of Petrology, v. 16, no. 1, p. 110–133, doi:10.1093/petrology/16.1.110.

Wright, T.L., Peck, D.L., and Shaw, H.R., 1976, Kilauea lava lakes; natural laboratories for study of cooling, crystallization, and differentiation of basaltic magma, *in* Sutton, G.H., Manghnani, M.H., and Moberly, R., eds., The geophysics of the Pacific Ocean basin and its margin: American Geophysical Union Geophysical Monograph 19 (The Woollard volume), pt. 5, no. 32, p. 375–392, doi:10.1029/GM019p0375.

Xu, G., Frey, F.A., Clague, D.A., Abouchami, W., Blichert-Toft, J., Cousens, B., and Weisler, M., 2007, Geochemical characteristics of West Molokai shield- and postshield-stage lavas; constraints on Hawaiian plume models: Geochemistry, Geophysics, Geosystems (G^3), v. 8, Q08G21, 40 p., doi:10.1029/2006GC001554.

Appendix. Microprobe Analyses of Glasses from Kīlauea Volcano

The following tables present microprobe analyses of glasses from Kīlauea Volcano. These include samples collected at Halema‘uma‘u from 1912 through 1934 (table A1), tephra samples from the 1924 summit eruption (table A2), and samples from the summit eruptions of 1954, 1971, and 1974 (table A3).

Table A1. Microprobe analyses of glasses in Halemaʻumaʻu samples collected by T. A. Jaggar from 1912 through 1934.

[Analyses by M.H. Beeson; "Comments" are as written on sample bag labels, per M.H. Beeson; oxide compositions in weight percent]

Sample ID	Date erupted	No. points analyzed	SiO_2	Al_2O_3	FeO	MgO	CaO	Na_2O	K_2O	TiO_2	P_2O_5	MnO	S	Total	Comments
95TAJ-3	1912	20	50.73	14.05	11.46	6.84	11.32	2.43	0.56	2.86	0.28	0.17	0.02	100.72	Pele's hair
95TAJ-4	1913	16	50.67	13.92	10.77	7.46	11.72	2.29	0.49	2.71	0.26	0.17	0.01	100.47	Pele's hair
95TAJ-5	Jan 1917	16	50.77	14.06	11.01	7.19	11.80	2.34	0.51	2.72	0.26	0.16	0.01	100.83	Driblet from Halemaʻumaʻu
95TAJ-23	Jan 1917	16	50.34	13.95	10.83	7.19	11.79	2.29	0.49	2.71	0.27	0.16	0.01	100.03	Overflow
95TAJ-24	Jan 1917	16	50.56	13.86	10.86	7.19	11.83	2.31	0.51	2.69	0.27	0.15	0.01	100.21	Spatter lava Halemaʻumaʻu
95TAJ-2	Mar 16, 1918	16	50.68	13.80	11.67	6.72	11.40	2.40	0.56	2.91	0.28	0.17	0.01	100.60	Lava spatter fragments
95TAJ-15	Jan 1918	13	49.69	13.71	11.53	6.78	11.37	2.35	0.52	2.79	0.27	0.16	0.008	99.17	Spatter caught red-hot
95TAJ-16	Feb 1918	15	50.29	13.57	12.35	6.44	11.01	2.49	0.56	3.07	0.28	0.17	0.005	100.24	Glassy Kīlauea lava collected with stick
95TAJ-21-3	Jan 1918	12	51.11	14.20	11.72	6.99	11.49	2.52	0.54	2.77	0.25	0.18	0.005	101.78	Splash on ledge 8 ft above lake
95TAJ-21-4	Jan 1918	4	49.64	13.76	11.55	6.74	11.52	2.42	0.52	2.80	0.26	0.15	0.007	99.37	Splash on ledge 8 ft above lake
95TAJ-29	1918	16	50.66	13.88	11.10	6.95	11.77	2.36	0.51	2.89	0.29	0.18	0.00	100.59	(no description)
95TAJ-18	Oct 20, 1919	15	50.16	13.92	11.79	6.75	11.41	2.40	0.49	2.74	0.26	0.17	0.008	100.10	Splash over N (W?) bank
95TAJ-22	1919	16	50.64	13.76	11.64	6.78	11.47	2.40	0.55	2.90	0.29	0.17	0.01	100.61	Flow glassy selvage
95TAJ-28	Mar 17, 1919	16	50.49	13.81	11.61	6.68	11.29	2.41	0.56	2.91	0.30	0.17	0.01	100.24	Splash are next to NE cone
95TAJ-34	Oct 6, 1919	16	50.48	13.99	11.22	6.93	11.61	2.35	0.53	2.81	0.27	0.17	0.01	100.37	Flow from Jaggar Lake
95TAJ-6	Nov 1920	16	50.43	14.01	10.98	7.16	11.70	2.34	0.49	2.68	0.26	0.17	0.01	100.23	Pele's hair
95TAJ-7	Mar 2, 1921	16	50.41	14.07	10.97	7.12	11.69	2.38	0.50	2.77	0.28	0.17	0.01	100.37	Glass lapilli etc. Halemaʻumaʻu
95TAJ-19	1921	14	50.32	14.00	11.30	6.86	11.48	2.44	0.52	2.83	0.25	0.15	0.012	100.16	Halemaʻumaʻu
95TAJ-8	Mar 21, 1921	16	50.39	14.14	10.96	7.03	11.61	2.38	0.50	2.77	0.27	0.16	0.01	100.28	Halemaʻumaʻu spatter
95TAJ-9	Aug 28, 1923	16	50.33	13.83	11.40	6.73	11.35	2.44	0.57	3.02	0.30	0.18	0.01	100.16	Pele's hair
95TAJ-32	1924	16	50.24	13.88	11.61	6.66	11.30	2.43	0.58	3.03	0.30	0.17	0.01	100.21	1924 lava
95TAJ-10	Jul 25, 1929	16	49.82	13.53	11.15	7.49	11.72	2.32	0.55	2.72	0.27	0.16	0.01	99.74	Glassy cinders
95TAJ-11	1934	10	50.17	13.69	11.41	7.28	11.57	2.27	0.52	2.63	0.24	0.17	0.013	99.97	Glassy cinders
95TAJ-12	Sept 1934	14	50.23	13.70	11.59	7.19	11.56	2.32	0.52	2.66	0.26	0.17	0.012	100.21	Glassy cinders
95TAJ-13	1868	16	51.16	13.60	12.24	6.27	10.73	2.47	0.49	2.72	0.27	0.18	0.012	100.14	1868 Kīlauea inner walls

Table A2. Microprobe analyses of glasses from tephra of the 1924 summit eruption.

[Analyses by T.R. Rose; oxide compositions in weight percent; T (MgO), glass quenching temperature using Helz and Thornber (1987) calibration]

Sample ID	Glassy Pele's tear 36	Glassy Pele's tear 37	Glassy Pele's tear 38	Glassy shards, with skeletal plagioclase	Partly crystalline Pele's tears
No. tears	1	1	1	11	5
SiO_2	50.15	50.60	49.48	50.43	49.82
TiO_2	2.65	2.60	2.69	2.92	3.61
Al_2O_3	13.41	13.46	12.96	13.63	12.73
FeO	10.93	11.37	11.45	11.41	13.15
MnO	0.14	0.19	0.14	0.16	0.20
MgO	7.39	7.61	8.72	7.02	5.85
CaO	11.59	11.55	11.49	11.70	10.39
Na_2O	2.22	2.27	2.25	2.29	2.43
K_2O	0.52	0.53	0.56	0.54	0.76
P_2O_5	0.26	0.25	0.29	0.26	0.39
Total	99.27	100.42	100.03	100.37	99.33
T (MgO) in degrees Celsius	1,163	1,168	1,191	1,154	1,132

Table A3. Microprobe analyses of glasses from the 1954, 1971, and 1974 summit eruptions.

[Analyses by R.T. Helz; oxide compositions in weight percent; --, no data; T(MgO), glass quenching temperature using Helz and Thornber (1987) calibration; NMNH, Smithsonian National Museum of Natural History]

Sample ID	1954A	1954B	Kil 71-3	Kil 71-5	Kil 71-6	Kil 71-8	Kil-774-2	Kil-774-3	Kil-774-6a	Kil-774-4	Kil 74-v2
No. points	11	13	11	12	10	12	16	12	11	15	19
SiO_2	50.18	49.89	50.25	50.38	50.73	50.72	50.73	50.22	50.84	50.89	50.51
TiO_2	2.86	2.82	2.46	2.36	2.72	2.76	2.39	2.32	3.05	2.86	2.26
Al_2O_3	13.97	13.95	13.45	13.22	13.94	13.80	13.55	12.98	13.29	13.74	13.36
Cr_2O_3	0.01	0.01	0.07	0.07	0.02	0.02	0.07	0.07	0.01	0.02	0.08
FeO	11.17	11.19	11.29	11.36	11.12	11.45	11.19	11.28	12.14	11.90	11.19
MnO	0.19	0.16	0.17	0.18	0.17	0.19	0.18	0.19	0.21	0.20	0.17
MgO	6.68	6.73	8.05	8.87	6.99	6.42	7.90	9.24	6.28	6.43	8.18
CaO	11.29	11.39	11.38	11.10	11.47	11.15	11.36	10.96	10.68	10.82	11.26
Na_2O	2.26	2.29	2.13	2.01	2.25	2.28	2.12	2.09	2.26	2.27	2.11
K_2O	0.56	0.54	0.42	0.41	0.51	0.54	0.42	0.42	0.59	0.53	0.42
P_2O_5	0.26	0.26	0.20	0.21	0.24	0.28	0.21	0.22	0.28	0.27	0.21
Total	99.43	99.23	99.87	100.17	100.16	99.61	100.12	100.57	99.04	100.09	99.75
T(MgO) in degrees Celsius	1148	1149	1177	1193	1154	1145	1173	1203	1140	1143	1179
Eruption date	6/1954	6/1954	8/14/1971	8/14/1971	8/14/1971	8/14/1971	7/19/1974	7/20/1974	7/20/1974	7/19/1974	7/1974
Location	caldera floor	caldera floor	extra-caldera	extra-caldera	caldera floor	caldera floor	northern vents	northern vents	caldera floor	southern vents	northern vents
Whole rock analysis in	Wright, 1971	Wright, 1971	Duffield and others, 1982	Duffield and others, 1982	Duffield and others, 1982	Duffield and others, 1982	Lockwood and others, 1999	Lockwood and others, 1999	Lockwood and others, 1999	Lockwood and others, 1999	--
NMNH sample ID	116859-20	116859-20	116095-3	116095-5	116095-6	116095-8	116125-2	116126-3	116126-7	116125-4	--