JOURNAL OF PETROLOGY | VOLUME 44 | NUMBER 12 | PAGES 2313–2339 | 2003 | DOI: 10.1093/petrology/egg079

A Petrologic Perspective of Kīlauea Volcano's Summit Magma Reservoir

MICHAEL O. GARCIA[1]*, AARON J. PIETRUSZKA[1]† AND J. MICHAEL RHODES[2]

[1]DEPARTMENT OF GEOLOGY AND GEOPHYSICS, UNIVERSITY OF HAWAI'I, HONOLULU, HI 96822, USA

[2]DEPARTMENT OF GEOSCIENCES, UNIVERSITY OF MASSACHUSETTS, AMHERST, MA 01003, USA

RECEIVED OCTOBER 23, 2002; ACCEPTED JUNE 19, 2003

Two hundred years of magmatic history are documented by the lavas and tephra sampled from Kīlauea's historical summit eruptions. This paper presents detailed petrographic and geochemical data for a comprehensive suite of samples erupted within or near Kīlauea's summit caldera since the 17th century. Our results elucidate the range of magmatic processes that operate within the volcano's summit magma reservoir and document two compositional trends that span nearly the entire known range for the volcano. Prior to the 1924 summit crater collapse, a trend of increasing incompatible element and CaO and decreasing SiO_2 abundances (at a constant MgO) prevailed. Thereafter, the trend reversed direction and has persisted for the rest of the 20th century, including during the current Pu'u `Ō`ō eruption. The rapid and systematic nature of these temporal geochemical variations indicates that the summit reservoir is a single, relatively small body rather than a plexus of dikes and sills. Olivine fractionation is the dominant petrologic process within this reservoir. Petrographic observations and olivine and whole-rock geochemical data suggest that the summit reservoir has a crown of aphyric, more evolved, low-density magma. Differentiation within this crown involving clinopyroxene and plagioclase is more extensive than previously recognized in Kīlauea summit lavas. The effects of crystal fractionation are superimposed upon an evolving hybrid magma composition produced by mixing new, mantle-derived magmas with more fractionated reservoir magma. Frequent eruptions of these hybrid reservoir magmas document the rapid variation in parental magma composition. These compositional variations correlate with magma supply rate; both are thought to be influenced by the degree of melting of small-scale source heterogeneities within the Hawaiian plume. However, Kīlauea's source compositions and partial-melting processes have varied only within a narrow range over the past 350 kyr.

KEY WORDS: *basalt; geochemistry; magma reservoir; Kīlauea Volcano; Hawai'i*

INTRODUCTION

The magmatic processes within active volcanoes are best delineated by monitoring the chemical and physical evolution of their lavas in space and time. The active Hawaiian volcanoes, Mauna Loa and Kīlauea, are ideal natural laboratories for detailed case studies of basaltic volcanism because of their relatively easy access, frequent eruptions, continuous monitoring by the US Geological Survey's (USGS) Hawaiian Volcano Observatory, and prior study (e.g. Brigham, 1909; Macdonald *et al.*, 1983; Klein *et al.*, 1987; Tilling & Dvorak, 1993; Barnard, 1995). From a petrological perspective, these volcanoes are particularly instructive because their lavas display rapid geochemical variations that are thought to result from short-term changes in the source composition and melting conditions within the Hawaiian plume (e.g. Wright, 1971; Wright & Fiske, 1971; Hofmann *et al.*, 1984; Casadevall & Dzurisin, 1987; Tilling *et al.*, 1987; Rhodes *et al.*, 1989; Kurz *et al.*, 1995; Rhodes & Hart, 1995; Pietruszka & Garcia, 1999*a*; Garcia *et al.*, 2000; DePaolo *et al.*, 2001; Pietruszka *et al.*, 2001). These mantle-derived geochemical signatures are overprinted by crystal fractionation, magma mixing, and assimilation (e.g. Wright, 1971; Wright & Fiske, 1971; Wilkinson & Hensel, 1988; Rhodes, 1995; Garcia *et al.*, 1998). To unravel these complexities and examine such issues as magma residence time (Pietruszka & Garcia, 1999*b*; Cooper *et al.*, 2001), it is necessary to

---

*Corresponding author. Telephone: 808-956-6641.
E-mail: mogarcia@hawaii.edu
†Present address: Department of Geological Sciences, San Diego State University, San Diego, CA 92182, USA.

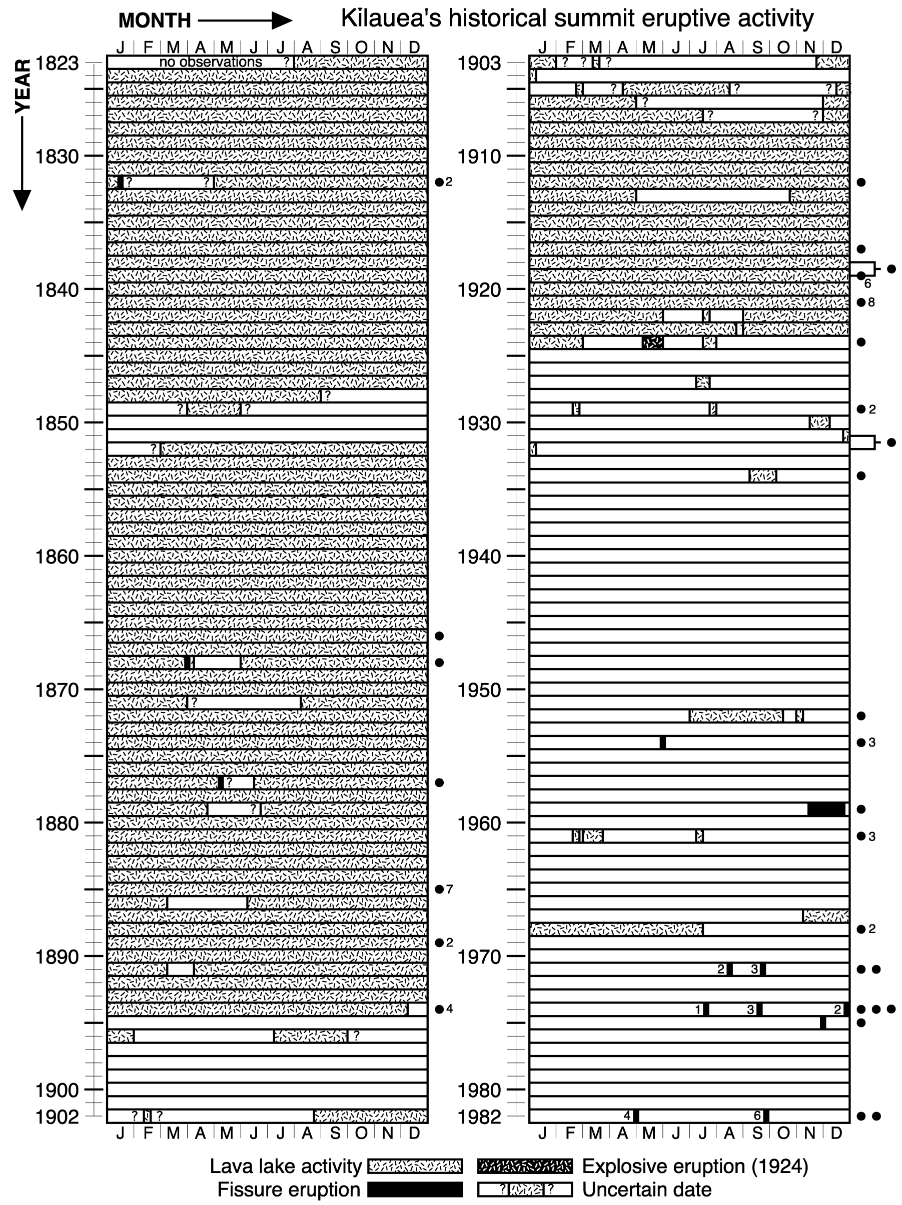

Fig. 1. Graphic summary of the historical eruptive activity at the summit of Kīlauea Volcano from the first written observations in August 1823 to the most recent summit eruption in September 1982. Eruptive activity has been dominated by lava lake activity (light stippled pattern). Some summit fissure eruptions (black bars) did not form an active lava lake because of their location and/or short duration. Explosive eruptions occurred in 1924 and during the prehistoric period including ~1790. The small dots to the right of the calendar indicate the eruptions or eruptive periods sampled for this study. The number next to the dot or eruption indicates how many samples were examined. This figure was modified from Macdonald (1955) based on descriptions from Dana (1889), Brigham (1909), Fiske *et al.* (1987) and Bevens *et al.* (1988), and updated for the 1955–1982 period using Macdonald *et al.* (1983) and references cited therein.

systematically examine a volcano's lava composition and mineralogy.

The ~200-year record of historical Kīlauea eruptions (Fig. 1) presents an excellent opportunity for a detailed petrologic investigation because the volcano's basic magmatic processes are well known (e.g. Tilling & Dvorak, 1993). Petrologic and geochemical studies have shown that the dominant compositional variation of Kīlauea lavas, particularly for those erupted at the summit of the volcano, is a consequence of olivine accumulation and fractionation (Powers, 1955; Wright, 1971; Wright & Fiske, 1971). In contrast, rift-zone magmas may undergo more extensive fractionation involving pyroxene and plagioclase during storage and cooling prior to eruption (Wright & Fiske, 1971), although the most primitive and many of its picritic lavas were erupted along Kīlauea's east rift zone (e.g. Garcia *et al.*, 1989; Clague *et al.*, 1991, 1995). More subtle compositional variations, a mild increase in K_2O, TiO_2 and CaO contents (at a constant MgO value), were noted for prehistoric and older historical lavas compared with younger historical lavas (Wright, 1971). These variations cannot be attributed to olivine control, and thus are thought to result from the influx of compositionally distinct magma batches into the summit magma reservoir (Wright, 1971; Wright & Fiske, 1971). Variations of incompatible trace element and Pb, Sr, and Nd isotopic ratios were also detected in Kīlauea's historical summit lavas, which were attributed to variations in the degree of partial melting of small-scale compositional heterogeneities within the Hawaiian mantle plume (Pietruszka & Garcia, 1999*a*; Pietruszka *et al.*, 2001).

In this study, we report petrologic data [X-ray fluorescence (XRF) whole-rock major and trace element chemistry, electron microprobe analyses of olivine and glass, and modal abundances of minerals] from a comprehensive suite of historical Kīlauea summit lavas and tephra. These data are used to investigate the temporal and spatial variations in the magmatic processes within the summit magma reservoir (e.g. magma mixing, crystal fractionation and crustal assimilation) from the 17th century to the most recent summit eruption in September 1982 for this classic shield volcano. Our goals are to refine our previous model for the structure and size of Kīlauea's summit magma reservoir (Pietruszka & Garcia, 1999*b*), to determine the nature of the temporal and spatial variations in magmatic processes at the volcano's summit, and to link the variations in trace element and isotope geochemistry of historical Kīlauea lavas (Pietruszka & Garcia, 1999*a*; Pietruszka *et al.*, 2001) with the petrologic evolution of the volcano.

Table 1: Historical Kīlauea summit geologic events

Date	Event
late 17th century	Explosions and circumferential fissure lava eruption
$\sim$1790	Violent, phreatomagmatic explosions and summit collapse
1820–1823(?)	Fire-fountain eruption within caldera
1823–1924	Nearly continuous lava lake activity
1832	$>$100 m caldera collapse and caldera rim eruption
1840	$\sim$100 m caldera collapse and large east rift zone eruption
1868	M7·8 earthquake and 90–180 m caldera collapse
1924	$\sim$300 m collapse, violent phreatic explosions and caldera-rim eruption
1924–1982	Infrequent, short-term lava eruptions ($<$1 day to few months)
1983–present	Partial draining of summit magma reservoir during Puʻu ʻŌʻō eruption

Sources: Bingham (1909); Macdonald *et al.* (1983); Dvorak & Dzurisin (1993); D. Swanson (personal communication, 2003).

SUMMARY OF HISTORICAL SUMMIT ERUPTIONS

Kīlauea's historical eruptive activity includes prolonged periods of quiescent effusion, interrupted by major collapses of the caldera floor ($\geq$100 m), and rare violent explosions (Table 1). The most dramatic and deadly historical eruptive event occurred in $\sim$1790 when violent explosions killed $\sim$80 Hawaiians. This eruption was thought to have involved four major explosive phases from vents within the caldera (McPhie *et al.*, 1990). Recent fieldwork and new ^{14}C dating indicate that some of these presumed $\sim$1790 phases are actually prehistoric (dating back to AD $\sim$1500; D. Swanson, personal communication, 2001). The next known event, between 1820 and 1823, involved high lava fountaining within the caldera, producing up to 2·8 m thick reticulite deposits (Sharp *et al.*, 1987). The first written record of Kīlauea eruptive activity, for August 1823, describes vigorous eruptive activity from many vents in an $\sim$240 m deep caldera (Brigham, 1909). Summit lava-lake activity continued until January 1832 (Fig. 1), when an eruption between the caldera and Kīlauea Iki crater (Fig. 2) was followed by draining of the summit lava lake and a $>$100 m collapse of the caldera floor (Brigham, 1909). Lava-lake activity resumed and by 1840, the caldera was half full when it collapsed $\sim$100 m accompanying a major east rift zone eruption. Subsequent summit eruptive activity was centered at Halemaʻumaʻu Crater, where a languid lava lake persisted until 1868 (Fig. 1), when Hawaii's largest historical earthquake, estimated at M $\sim$7·9 (Wyss & Koyanagi, 1992), shook the Kīlauea region. Almost simultaneously, two-thirds of the caldera floor collapsed some 90–180 m and a lava flow erupted into Kīlauea Iki crater from the nearby caldera rim. Vigorous activity resumed in Halemaʻumaʻu and continued until 1894 (Fig. 1), when the lake drained. The next 13 years saw only episodic eruptions (Brigham, 1909). Nearly continuous lava-lake activity resumed at Halemaʻumaʻu in 1907 with major overflows and fissure eruptions from 1918

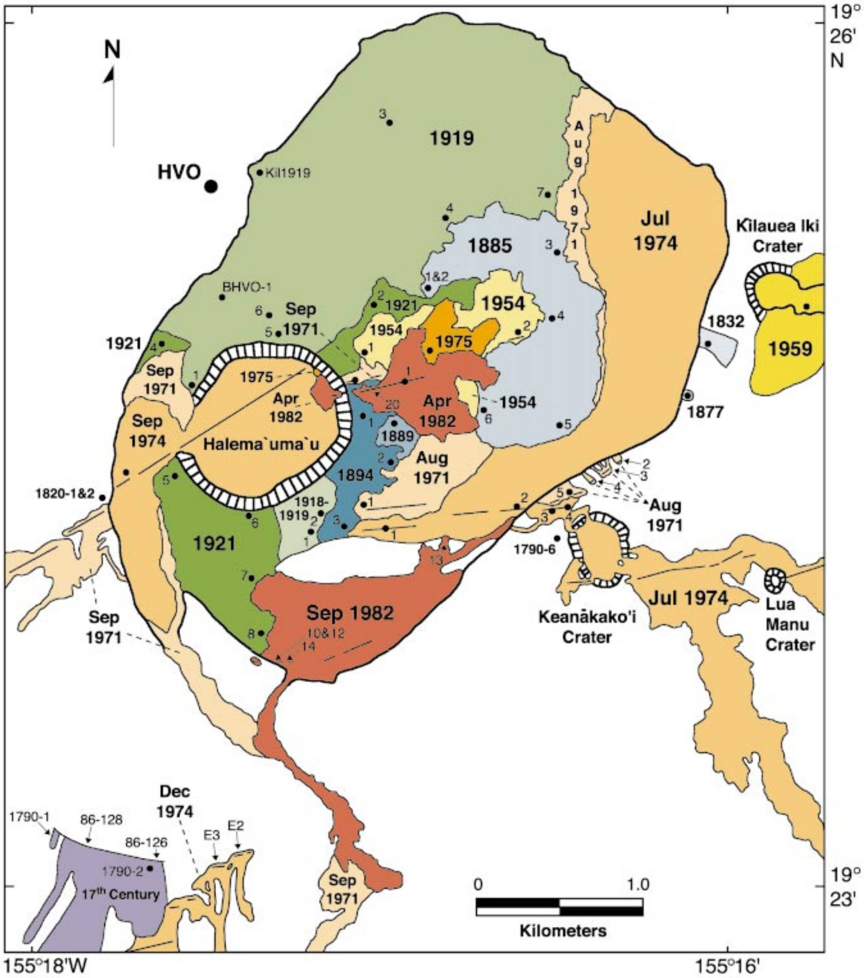

Fig. 2. Geologic map of the historical eruptions and 17th century lava flows within the summit area of Kīlauea Volcano (after Neal & Lockwood, 2003). The colors indicate the relative age of the flows (purple and blue for the older flows, and gold and red for the younger flows). Most of the unlabeled areas adjacent to the caldera are covered with tephra from 17th, 18th and 19th century explosions. Locations of samples studied (Tables 4 and 5) are shown by dots except where they have been covered by later flows (e.g. 1868 in Kīlauea Iki). Eruptive fissures are marked by fine lines. Hatched areas around craters are cliff sections of older lavas. HVO, location of the Hawaiian Volcano Observatory.

to 1921, which covered large portions of the caldera floor (Fig. 2).

In 1924, the lava lake in Halemaʻumaʻu drained, its floor and walls collapsed, and numerous violent phreatic explosions occurred (Jaggar, 1947). Subsequent summit eruptions were sporadic, with seven within Halemaʻumaʻu lasting a few days to 1 month between 1924 and 1934 (Finch, 1940; Jaggar, 1947). Halemaʻumaʻu remained quiet for the next 17 years; subsequent eruptions in 1952, 1954, 1961, 1967–1968, 1971, 1974, 1975, and 1982 were mostly short-lived from fissures in and near the crater (Fig. 1; Macdonald *et al.*, 1983). Short summit eruptions also occurred outside the Halemaʻumaʻu area in 1959, 1971, 1974, and 1982 (Fig. 1). The 1959 eruption in Kīlauea Iki Crater was unusual for its high lava fountains (up to ~600 m) and its picritic lavas (Richter *et al.*, 1970). There have been no summit eruptions since 1982.

Major eruptions have also occurred along the subaerial portion of Kīlauea's east and SW rift zones. The east rift zone had major fissure eruptions in 1790, 1840 and 1955 (Holcomb, 1987). Since 1955, it has been the site of frequent eruptions and intrusions including the current ~20 years long, Pu`u `Ō`ō eruption (e.g. Klein *et al.*, 1987; Wolfe *et al.*, 1987; Garcia *et al.*, 1996*b*, 2000). Eruptions on the SW rift have been less frequent, with only two major (in 1823 and 1919) and three minor eruptions in 1868, 1971, 1974 (Macdonald *et al.* 1983). The well-documented record of historical activity at Kīlauea provides an excellent opportunity to gain insights into processes within its summit magma reservoir.

SAMPLING

Halema'uma'u Crater and adjacent areas were the primary targets for our sampling of historical Kīlauea summit eruptions. We collected from all exposed surface flows and tephra deposits (see Fig. 2 for sample locations and Fig. 1 for age distribution). Additional samples were obtained from the collections of the Smithsonian Institution, US Geological Survey, and the University of Hawaii for some eruptions whose products are now covered by younger deposits. Samples are numbered here according to their date of eruption, unless they were numbered in a previous study. We collected samples from deposits mapped as from the 1790 eruption (McPhie *et al.*, 1990), which are now ^{14}C dated as middle to late 17th century (D. Swanson, personal communication, 2001). They include juvenile tephra from layer 6 of McPhie *et al.* (1990), and lava and spatter from vents along a circumferential fracture ~1·4 km SW of the caldera. Two large clasts of reticulite (~10 cm long) were taken from the 1820 s 'golden pumice' deposit (e.g. Sharp *et al.*, 1987). Two samples of the 1832 lava flow were collected between the caldera and the Kīlauea Iki pit crater. This crater was the site of two other historical eruptions in 1868 and 1959, which were also sampled. The oldest sample obtained from the caldera floor was erupted in 1866. A lava erupted in 1877 on the eastern caldera rim was also examined.

Our samples of eruptions between 1885 and 1921 are from overflows or fissures extending from Halema'uma'u crater onto the caldera floor. Multiple samples were taken from the larger flows (1885, 1894, 1919 and 1921) to determine intra-flow compositional variations. Pele's hair from eruptions in 1912 and 1923 collected by T. Jaggar was obtained for this study. Deposits from these eruptions are now buried. None of the products from the brief and small eruptions during the 30 year period following the 1924 summit explosions are now exposed. However, we obtained samples from the 1929, 1931, 1934 and 1952 eruptions from institutional collections. At least one sample was obtained from each post-1952 summit eruption.

PETROGRAPHY

The basic mineralogy of Kīlauea basalts has been discussed in many papers dating back to Dana (1889); however, few of these studies provide modal data, especially for rocks erupted before 1955. New modes were determined for 52 Kīlauea historical lavas to: (1) establish what minerals were present in the magma reservoir; (2) assess how these minerals might affect whole-rock compositions; (3) evaluate temporal and spatial variations in rock type. All of the historical summit lavas appear fresh in thin section. Many contain brown glass and are strongly vesicular (>20 vol. %); others have a fine-grained, dark gray matrix consisting of plagioclase, clinopyroxene and opaque minerals (mostly Ti-magnetite), with or without olivine. Although olivine is commonly thought to be absent in the groundmass of tholeiitic lavas (e.g. McBirney, 1993), it is ubiquitous in the more MgO-rich (>7 wt %) Kīlauea summit lavas.

Historical Kīlauea summit lavas are overwhelmingly weakly olivine-phyric (~85% have 1–3 vol. % phenocrysts; Table 2). More porphyritic lavas were erupted in 1820 from Halema'uma'u when eruption rates were high, and in 1959, August 1971 and December 1974 from vents on the margin of the caldera (Table 2). Phenocrysts (>0·5 mm across) in these lavas are almost always small (<2 mm across) except in lavas from the December 1974 (Kolysko-Rose, 1999) and 1959 eruptions (Helz, 1987). The small and sparse nature of phenocrysts indicates that whole-rock analyses of most Kīlauea historical summit lavas are essentially representative of magmatic compositions.

Olivine is the most common mineral in historical Kīlauea summit lavas. It usually occurs as equant euhedral crystals but may also be present as skeletal or strongly elongate crystals. Deformed and resorbed olivines (xenocrysts) are rare or absent in most of these lavas (Table 2). Inclusions of glass and euhedral chromite are common in these olivine crystals. Thus, olivine and chromite were probably the only minerals growing in the magma prior to eruption, except for many of the 1885–1921 lavas, which contain other minerals.

Clinopyroxene and plagioclase phenocrysts are absent or rare (≤0·4 vol. %) in historical Kīlauea summit lavas. When present, they are small (≤1 mm in width), subhedral to euhedral without signs of resorption or deformation. Both these minerals are more common as microphenocrysts (usually 0·1–0·3 mm across), which may have grown during and just after eruption of their host lava. Lavas erupted before 1877 and after 1924 contain rare or no clinopyroxene and plagioclase microphenocrysts, except those from 1968 (Table 2). Open-textured, small (<1 cm across) glomerocrystic clots and small (<1 cm across) gabbroic and pyroxenitic xenoliths occur in some summit lavas but are common only in those erupted during periods of long-lived lava lake activity, especially during the period from 1885 to 1894 (Table 2).

ANALYTICAL METHODS

Olivine and glass compositions were determined using the University of Hawaii, five-spectrometer, Cameca SX-50 electron microprobe using SAMx automation. For olivine analyses, the operating conditions were a

Table 2: Petrography, modes and calculated magmatic densities of representative Kīlauea historical summit lavas

Eruption date	Sample	Whole-rock MgO (wt %)	Olivine			Clinopyroxene		Plagioclase		Groundmass	Glomerocrysts	Density
			Phen	Mph	Xeno	Phen	Mph	Phen	Mph			
18th century	1790-6	9·4	1·3	2·4	<0·1	0·0	0·0	0·0	0·0	96·3	none	2·610
18th century	1790-1	10·6	1·8	2·9	0·0	0·0	0·0	0·0	0·0	95·3	none	2·629
1820–1823	1820-1	12·2	3·8	2·9	<0·1	0·0	0·0	0·0	0·0	93·3	none	2·645
1832	1832	7·6	0·7	1·6	<0·1	0·0	<0·1	0·0	0·0	97·7	none	2·600
1868	1868	6·7	0·6	1·2	0·0	0·0	<0·1	0·0	<0·1	98·2	none	2·576
1885	1885-1	6·3	0·4	0·4	0·0	<0·1	1·8	0·0	0·4	97·0	common	2·556
1885	1885-2	6·0	<0·1	0·3	0·0	<0·1	2·0	0·0	3·9	93·8	common*	2·564
1885	1885-3	6·8	<0·1	0·3	0·0	<0·1	2·7	0·0	3·8	93·2	common*	2·564
1885	1885-4	7·2	0·3	0·6	0·0	<0·1	4·3	<0·1	3·6	91·2	common	2·572
1885	1885-5	6·2	0·0	0·2	0·0	0·0	1·2	0·0	1·2	97·4	common	2·549
1885	1885-6	6·6	0·2	0·4	0·0	0·0	0·4	0·0	1·0	98·0	common	2·559
1889	1889	7·4	0·9	2·3	0·1	0·2	1·8	0·0	4·5	90·2	common*	2·596
1894	1894-1	6·5	<0·1	0·6	0·0	0·0	0·2	0·0	0·6	98·6	common	2·576
1894	1894-2	6·6	<0·1	0·8	0·0	0·0	0·3	0·0	5·5	93·9	common*	2·570
1894	1894-3	6·6	<0·1	0·4	0·0	0·0	0·2	0·0	0·8	98·6	rare	2·555
1918–1919	1918-1919	9·5	2·6	3·1	<0·1	<0·1	1·4	<0·1	<0·1	92·9	common	2·625
1919	1919-1	6·8	<0·1	0·5	0·0	<0·1	0·3	0·0	0·3	98·9	common	2·560
1919	1919-3	6·9	0·6	1·2	0·2	0·4	1·8	0·0	<0·1	95·8	common	2·576
1919	1919-4	7·1	0·0	0·5	0·0	0·0	0·8	0·0	0·5	98·2	rare	2·563
1921	1921-4	7·5	0·3	0·7	0·0	0·0	0·5	0·0	<0·1	98·5	rare	2·571
1921	1921-1	8·4	0·3	1·3	0·0	0·0	3·2	0·0	<0·1	95·2	rare*	2·591
1921	1921-2	8·7	0·8	1·0	0·2	0·6	3·4	0·0	0·0	94·0	common	2·588
1921	1921-5	10·6	1·4	2·7	0·0	0·0	2·1	0·0	<0·1	93·8	rare*	2·621
1954	1954-1	7·0	0·0	0·2	<0·1	0·0	0·4	0·0	0·0	99·4	none	2·554
1954	1954-2	7·2	0·0	<0·1	0·0	0·0	<0·1	0·0	0·0	99·9	none	2·559
1959	1959	17·7	15·4	1·2	3·0	0·0	0·0	0·0	0·0	80·4	rare	2·754
1961	K61-22	7·6	<0·1	1·5	0·0	0·0	0·0	0·0	0·0	98·5	none	2·569
1968	HM68-15	7·5	0·0	1·3	0·0	0·0	1·0	0·0	2·4	95·3	common*	2·571
Aug. 1971	1971A-1	7·3	0·0	0·2	0·0	0·0	0·0	0·0	0·0	99·8	none	2·594
Aug. 1971	1971A-2	10·1	2·0	0·2	0·0	0·0	0·0	0·0	0·0	97·8	none	2·603
Sep. 1971	1971S	7·3	0·3	1·0	0·0	0·0	0·0	0·0	0·0	98·7	none	2·570
Jul. 1974	1974 J	7·5	<0·1	0·4	0·0	0·0	0·0	0·0	0·0	99·6	none	2·565
Sep. 1974	1974S	7·2	0·0	0·1	0·0	0·0	<0·1	0·0	0·0	99·9	none	2·557
Dec. 1974	1974D-E3	13·0	15·6	5·7	<0·1	0·0	0·0	0·0	0·0	74·6	none	2·768
1975	1975	7·2	0·0	<0·1	0·0	0·0	<0·1	0·0	0·0	99·9	none	2·559
Apr. 1982	1982A-20	7·3	<0·1	0·9	0·0	0·0	0·2	0·0	0·3	98·5	none	2·569
Sep. 1982	1982S-10	7·5	0·2	0·6	0·0	0·0	0·0	0·0	0·0	99·2	none	2·568
Sep. 1982	1982S-12	7·2	0·0	0·8	0·0	0·0	0·1	0·0	0·2	98·9	none	2·565
Sep. 1982	1982S-13	8·7	<0·1	8·3	0·0	0·0	0·0	0·0	0·0	91·7	none	2·643

*Contains small (<1 cm across) gabbroic and clinopyroxenitic xenoliths.

Modes are in volume % and are based on 500–1000 counts/sample. Phenocrysts (phen) are >0·5 mm in diameter; microphenocrysts (mph) are 0·1–0·5 mm across; olivine xenocrysts (xeno) have planar discontinuities. Densities (g/cm^3) were calculated using the program of Niu & Batiza (1991) using observed modes and whole-rock and mineral compositions.

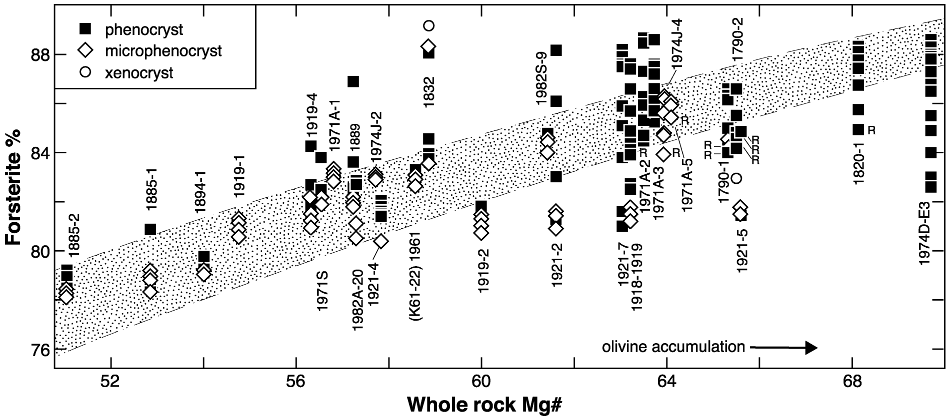

Fig. 3. Olivine core composition (forsterite %) compared with whole-rock Mg number {[(Mg/Fe^{2+} + Mg) × 100], with the Fe^{2+} set to 90% of the total iron} in representative Kīlauea historical summit lavas. Xenocrysts are deformed or resorbed crystals. The broad stippled band shows the range of equilibrium olivine compositions for a given Mg number based on Roeder & Emslie (1970) and Rhodes *et al.* (1979). Rocks with olivines that plot below this band have probably accumulated olivine that grew in more differentiated magma. Rocks with olivines that plot above the band may reflect 'delayed' olivine fractionation (e.g. Maaloe *et al.*, 1988). All crystals are normally zoned or unzoned except those labeled with an 'R', which are reversely zoned. The 2σ errors are smaller than the symbol size. Representative analyses given in Table 3.

minimum spot size (~1 μm), 15 kV and 20 nA beam current. For glasses, a broader beam (~15 μm) and lower current (10 nA) was used to minimize degradation of the sample. Na was analyzed first in the glasses for only 40 s to minimize its loss. Peak counting times for other elements were 65–75 s for glass and 60 s for olivine. The raw data were corrected using ZAF-PAP procedures (e.g. Reed, 1993). The reported analyses are an average of at least three spots in the core of each olivine; rims also were analyzed to check for compositional zoning. The glass data are an average of 5–20 spot analyses.

Whole-rock XRF analyses were made for major and trace (Nb, Rb, Zr, Sr, Y, V, Zn, Ni, Cr) elements at the University of Massachusetts. The methods and the analytical precision for the XRF analyses have been given by Rhodes (1996).

OLIVINE CHEMISTRY

Compositions were determined for olivine phenocrysts and microphenocrysts from a wide range of Kīlauea historical summit lava compositions (5·9–13·0 wt % MgO). Olivines were previously examined in the high-MgO 1959 eruption products and were found to have core compositions of forsterite (Fo) 83·5–88·7 (Helz, 1987). New olivine analyses were made of other Kīlauea summit lavas to ascertain whether their olivines were in equilibrium with their host magma. Although olivine can re-equilibrate rapidly at magmatic temperatures (e.g. Roeder & Emslie, 1970; Ulmer, 1989), analyses of olivine in lavas can provide evidence of magma mixing, crystal accumulation and other magmatic processes (e.g. Nakamura, 1995; Garcia, 1996).

The olivine core compositions in historical Kīlauea summit lavas range from Fo 78 to 89 (Table 3) and, in general, Fo values increase with the whole-rock Mg number (Fig. 3). Most olivines are normally zoned or unzoned, which is typical of Kīlauea (Garcia *et al.*, 1992; Clague *et al.*, 1995) and tholeiitic lavas from other Hawaiian volcanoes (e.g. Garcia, 1996). However, all of the summit lavas with Mg number ≥63·5 contain two populations of olivine; a normally zoned, high-Fo group (Fo 86–89) and reversely zoned, lower forsterite olivine (Fo 83–85%). The presence of reversely zoned olivine in these lavas is indicative of magma mixing. The reverse zoning occurs only at crystal rims (outer 10–20 μm), suggesting that the mixing occurred during or shortly before eruption. These lavas were erupted from fissures on the margin of the caldera, except for the 1820 tephra.

Most of the olivines were in Fe/Mg equilibrium with their host rock when erupted (Fig. 3), which is significant given the wide range in rock Mg number (51–68). Some of the lavas with Mg numbers 58–63 have euhedral olivines with Fo 88 ± 1, which is 1–4% Fo too high for their bulk composition. These undeformed, high-Fo olivines have moderate CaO contents (0·20–0·35 wt %) suggesting they were not formed in the mantle (e.g. Stormer, 1973; Jurewicz & Watson, 1988). Instead, these olivines probably formed early in the crystallization history of their host magmas, and their presence reflects incomplete crystal fractionation (Maaloe *et al.*, 1988). These high-Fo olivines indicate that Kīlauea summit lavas are probably derived from parental magmas with Mg numbers of at least 69 ± 1,

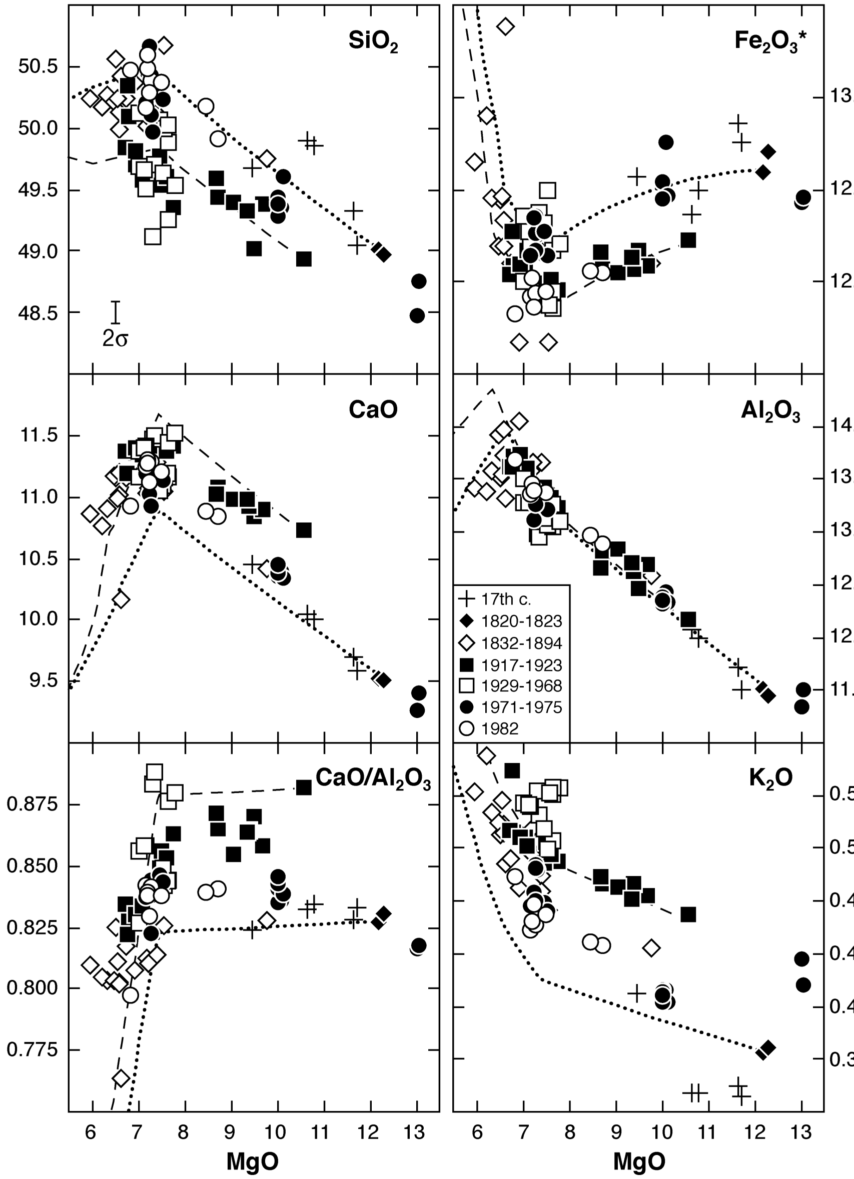

Fig. 4. MgO variation diagrams for Kīlauea historical summit lavas. Samples are divided into seven age groups based on compositional similarity, although not all samples from an age group are similar. The dotted and dashed lines are the liquid lines of descent determined from MELTS modeling (see text for details). The similarity in these model lines to the observed curved or kinked trends for recent Kīlauea summit lavas should be noted; this indicates that crystal fractionation has played an important role in controlling compositional variation. A 2σ error bar is shown for SiO$_2$; the errors for the other oxides are smaller than symbol size. Fe$_2$O$_3$* is total iron; data from Table 4.

which is consistent with the discovery of Kīlauea submarine glasses with Mg numbers of 69–70 (~15 wt % MgO; Clague *et al.*, 1991). Sample 1832-1 contains rare deformed olivines, which yielded the highest measured Fo content (89·2). This lava was erupted from the same area outside the caldera as the 1959 lavas (Fig. 2), which also contain abundant high-forsterite olivine xenocrysts (up to 88·7% Fo; Helz, 1987). The xenocrysts in these lavas may be derived from deformed cumulates from a prehistoric magma reservoir that is thought to underlie this area (Holcomb, 1987).

WHOLE-ROCK GEOCHEMISTRY

Major elements

XRF major-element analyses were made of 82 rocks from Kīlauea summit eruptions; 77 are from 32 separate historical eruptions and five are from mid- to late

Table 3: Representative microprobe analyses of olivine cores from Kīlauea historical summit lavas arranged in order of increasing rock Mg number

Sample:	1885-2		1894-1			1971S (Sep)			K61-22 (1961)			1832		
Rock Mg no.:	51·0		54·0			56·5			58·4			58·8		
Size (mm):	0·6	0·2	0·7	0·5	0·3	0·9	0·6	0·3	0·7	0·3	0·2	0·8	0·4	0·2
SiO_2	39·0	38·7	39·0	38·7	38·7	39·7	39·6	39·6	39·7	39·8	39·8	40·4	39·9	39·4
FeO	19·4	20·1	19·2	19·4	19·3	15·4	16·4	16·6	15·7	15·9	16·1	11·55	14·7	15·9
NiO	0·16	0·16	0·21	0·21	0·21	0·27	0·20	0·18	0·29	0·29	0·27	0·36	0·25	0·21
MgO	41·4	40·6	41·7	41·6	41·6	44·7	43·7	43·6	44·0	43·85	43·8	47·65	45·1	44·45
CaO	0·30	0·31	0·29	0·28	1·29	0·23	0·27	0·27	0·26	0·30	0·30	0·24	0·27	0·27
Total	100·16	99·77	100·4	100·19	100·1	100·3	100·17	100·25	99·95	100·18	100·27	100·2	100·22	100·23
Fo%	79·2	78·3	79·5	79·3	79·3	83·8	82·6	82·5	83·3	83·0	82·9	88·0	84·6	83·3

Sample:	1921-2			1921-7			1918-1919				1971A-2 (Aug.)		
Rock Mg no.:	61·4			63·1			63·2				63·5		
Size (mm)	1·3	0·6	0·3	2·4	2·0	1·8	1·2	0·9	0·8	0·6	0·8	0·6	0·4
SiO_2	40·35	39·45	39·3	40·05	40·25	39·05	40·5	40·3	39·7	39·5	40·0	39·7	39·8
FeO	11·3	17·4	17·6	12·0	11·47	17·1	12·0	12·85	14·8	16·35	11·2	13·3	14·6
NiO	0·39	0·27	0·29	—	—	—	0·36	0·27	0·26	0·25	0·38	0·27	0·26
MgO	48·0	42·8	42·7	47·5	47·5	43·2	47·3	46·6	44·8	43·55	48·1	46·4	45·2
CaO	0·25	0·31	0·29	0·27	0·27	0·27	0·25	0·31	0·32	0·29	0·24	0·24	0·26
Total	100·29	100·23	100·18	99·82	99·49	99·62	100·39	100·33	99·76	99·94	99·92	99·91	100·12
Fo%	88·3	81·4	81·2	87·6	88·2	81·9	87·6	86·6	84·3	82·5	88·4	86·1	84·7*

Sample:	1974 J				1790-1 (17th century)			1820-1			1974D-E3		
Rock Mg no.:	63·9				65·4			68·0			69·7		
Size (mm)	0·4	0·5	0·4	0·2	1·1	0·6	0·6	1·4	0·8	0·6	1·5	1·2	0·8
SiO_2	39·7	39·4	39·65	39·8	40·2	39·9	39·9	40·4	40·1	39·85	40·25	39·9	39·5
FeO	13·65	15·3	14·45	13·1	13·0	14·4	15·1	11·1	12·6	14·2	11·35	12·45	15·2
NiO	—	—	—	—	0·28	0·25	0·24	0·35	0·32	0·29	0·39	0·32	0·28
MgO	46·0	45·0	45·7	46·5	46·45	45·5	44·9	48·4	46·8	45·3	47·95	46·85	44·7
CaO	0·27	0·28	0·28	0·25	0·24	0·27	0·25	0·23	0·22	0·22	0·23	0·26	0·25
Total	99·35	99·7	99·8	99·4	100·17	100·32	100·39	100·48	100·04	99·86	100·17	99·78	99·93
Fo%	85·7	84·0*	84·9	86·4	86·4	84·9	84·1*	88·6	86·9	85·0*	88·3	87·0	84·0

*Reversely zoned.

Rock Mg number is $[Mg/(Fe^{2+} + Mg)] \times 100$ assuming 90% of the total iron is Fe^{2+}; size refers to diameter of olivine crystal; oxides are in wt %; forsterite (Fo).

17th century eruptions (Table 4). Major-element analyses for lavas from many of these eruptions have been reported previously (see Wright, 1971). We report here the first analyses for the 17th century extra-caldera flows, the 1820, 1929, 1934 and 1982 Halema'uma'u eruptions, and 1982 caldera margin lavas.

The Kīlauea historical summit lavas display a wide range in MgO content (6·0–17·8 wt %). This range is nearly identical to that reported for the Hilina basalts (5·8–18·0 wt %; Fig. 4), which are Kīlauea's oldest exposed lavas (>23 ka) and may span ~75 kyr of its history (Easton, 1987; Chen *et al.*, 1996). Kīlauea lavas

Table 4: XRF whole-rock analyses of Kīlauea historical summit rocks

Date	Sample	SiO_2	TiO_2	Al_2O_3	$Fe_2O_3^*$	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Total	Y	Sr	Rb	Nb	Zr	Zn	Ni	Cr	V
17th century	1790-6	49·67	2·289	12·68	12·57	0·18	9·45	10·45	2·26	0·410	0·218	100·18	23·8	296	6·1	12·7	147	123	215	524	272
17th century	1790-1	49·90	2·224	12·07	12·37	0·17	10·64	10·05	1·96	0·334	0·202	99·92	23·1	269	5·5	12·8	136	117	258	838	260
17th century	1790-2	49·86	2·225	12·00	12·50	0·17	10·80	10·01	2·07	0·334	0·190	100·16	22·4	271	5·6	12·8	136	116	265	851	257
17th century	86-126	49·05	2·140	11·51	12·76	0·17	11·71	9·59	1·94	0·332	0·201	99·40	22·6	263	5·7	12·2	138	120	279	813	268
17th century	86-128	49·33	2·174	11·71	12·86	0·17	11·65	9·70	1·96	0·339	0·199	100·09	20·6	261	5·7	11·6	135	118	318	848	248
1820–1823	1820-1	49·01	2·104	11·51	12·59	0·17	12·19	9·52	2·08	0·365	0·207	99·75	21·6	277	5·7	11·9	141	117	398	904	259
1820–1823	1820-2	48·96	2·125	11·44	12·70	0·19	12·31	9·50	1·73	0·368	0·210	99·53	21·2	273	6·3	11·3	132	113	390	917	252
1832	1832	50·66	2·499	13·37	11·67	0·17	7·58	11·04	2·35	0·472	0·252	100·06	24·6	338	8·0	15·2	163	107	134	372	286
1832	K1832-1	49·74	2·357	12·58	12·10	0·16	9·79	10·41	2·16	0·444	0·240	99·98	23·8	331	7·3	14·4	160	116	236	592	285
1866	1866	50·56	2·796	13·52	12·44	0·18	6·53	11·15	2·28	0·530	0·288	100·27									
1868	1868	50·42	2·757	13·31	13·38	0·18	6·66	10·16	2·66	0·507	0·277	100·31	27·2	335	8·8	16·4	187	137	112	154	316
1877	K1877-1	50·36	2·513	14·05	11·67	0·16	6·95	11·34	2·38	0·489	0·254	100·17	25·0	359	8·7	16·4	170	113	96	169	293
1885	1885-1	50·26	2·826	13·57	12·45	0·17	6·34	10·90	2·32	0·547	0·290	99·68	26·7	379	10·1	18·9	190	118	88	182	286
1885	1885-2	50·23	2·904	13·41	12·65	0·18	6·00	10·85	2·65	0·562	0·299	99·74	27·4	379	10·4	19·5	197	125	78	151	296
1885	1885-3	50·24	2·677	13·64	12·10	0·17	6·76	11·15	2·28	0·511	0·273	99·79	25·1	373	9·2	17·4	178	114	106	259	282
1885	1885-4	50·01	2·639	13·57	12·24	0·17	7·21	11·02	2·05	0·516	0·269	99·69	25·4	367	8·9	19·0	175	119	138	303	285
1885	1885-5	50·16	2·965	13·37	12·90	0·18	6·24	10·76	2·44	0·590	0·306	99·91	28·0	373	10·6	21·7	196	128	93	167	309
1885	1885-6	49·98	2·738	13·72	12·33	0·17	6·62	11·01	2·23	0·538	0·279	99·62	26·5	378	9·4	18·4	181	116	111	229	288
1889	1889	50·11	2·601	13·65	12·15	0·17	7·42	11·10	2·41	0·499	0·265	100·38	24·2	354	8·9	16·7	162	113	133	355	284
1889	K1889	50·31	2·543	13·41	11·92	0·17	7·43	11·28	2·23	0·487	0·254	100·03	24·5	356	8·0	16·1	165	107	140	469	283
1894	1894-1	50·21	2·762	13·91	12·19	0·17	6·51	11·17	2·32	0·539	0·282	100·06	26·2	383	9·5	18·9	179	108	105	220	280
1894	1894-2	50·12	2·695	13·96	12·19	0·17	6·62	11·19	2·28	0·531	0·273	100·03	25·3	379	9·6	18·7	175	116	115	249	284
1894	1894-3	50·24	2·815	13·54	12·46	0·17	6·59	10·98	2·48	0·556	0·284	100·12	26·5	373	9·4	19·8	184	121	112	260	300
1894	K1894-1	50·08	2·679	13·65	12·01	0·17	7·23	11·06	2·34	0·535	0·270	100·02	24·9	380	9·1	18·7	178	112	145	411	277
1917	1917	49·35	2·662	13·23	11·96	0·19	7·75	11·42	2·11	0·510	0·265	99·45									
1918–1919	1918-1919	49·02	2·570	12·47	12·17	0·17	9·51	10·84	2·04	0·486	0·258	99·55	22·6	369	8·7	17·6	173	108	253	572	276
1918–1919	1918-1919-2	49·38	2·566	12·70	12·09	0·18	9·68	10·90	2·06	0·484	0·259	100·30	22·7	360	8·4	17·2	164	111	251	605	282
1919	1919-1	50·09	2·798	13·73	12·08	0·17	6·80	11·37	2·30	0·532	0·283	100·15	25·2	396	9·1	19·5	192	115	99	246	302
1919	1919-3	49·70	2·750	13·63	12·09	0·17	6·95	11·36	2·19	0·529	0·273	99·64	25·3	396	9·2	20·1	189	117	121	311	298
1919	1919-4	49·63	2·748	13·61	12·08	0·17	7·07	11·34	2·27	0·529	0·273	99·72	25·3	396	9·1	20·9	188	118	126	298	298
1919	1919-5	49·84	2·793	13·62	12·04	0·17	6·74	11·37	2·18	0·534	0·285	99·57	24·7	391	8·9	18·8	180	113	96	262	292
1919	1919-6	49·77	2·676	13·43	11·90	0·17	7·45	11·42	2·52	0·509	0·273	100·12	23·8	381	8·7	17·9	172	110	125	342	289
1919	Kil1919	49·81	2·763	13·73	12·10	0·17	6·93	11·40	2·50	0·528	0·274	100·21	25·1	398	9·0	19·5	187	110	106	268	288
1919	BHVO-1	49·58	2·740	13·61	12·17	0·17	7·09	11·35	2·40	0·522	0·272	99·90	24·6	388	9·2	19·4	191	113	126	284	285

Date	Sample	SiO_2	TiO_2	Al_2O_3	Fe_2O_3*	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Total	Y	Sr	Rb	Nb	Zr	Zn	Ni	Cr	V
1921	1921-2	49·43	2·638	12·80	12·08	0·17	8·74	11·08	2·09	0·493	0·263	99·79	23·6	374	8·8	18·3	178	110	203	529	290
1921	1921-4	49·53	2·721	13·33	11·99	0·17	7·49	11·41	2·18	0·525	0·269	99·62	24·6	389	8·9	20·6	182	113	153	445	297
1921	1921-5	48·94	2·469	12·17	12·23	0·17	10·59	10·73	2·05	0·470	0·245	100·06	22·3	350	8·2	17·4	163	111	320	778	276
1921	1921-6	49·59	2·677	12·66	12·16	0·17	8·68	11·03	1·91	0·498	0·258	99·63	22·7	369	8·6	17·8	169	108	202	518	287
1921	1921-7	49·34	2·586	12·63	12·07	0·18	9·40	10·92	1·90	0·493	0·263	99·78	22·9	364	8·6	17·3	164	110	233	559	277
1921	1921-8	49·40	2·615	12·85	12·05	0·18	9·05	10·98	2·05	0·491	0·263	99·93	22·8	361	8·3	17·4	165	108	224	570	283
1921	K1921-1	49·33	2·548	12·71	12·13	0·17	9·37	10·98	2·13	0·481	0·250	100·10	22·9	366	8·0	18·1	168	91	251	608	278
1921	K1921-2	49·61	2·703	13·33	12·01	0·17	7·61	11·37	2·29	0·515	0·270	99·88	24·5	383	9·1	20·1	185	121	153	474	303
Aug. 1923	TAJ-9	50·36	2·944	13·62	12·23	0·18	6·75	11·19	2·32	0·579	0·304	100·48									
Jul. 1929	TAJ-10	49·54	2·740	13·10	12·17	0·18	7·78	11·53	2·11	0·567	0·28	100·00									
1929	1929	49·12	2·705	12·99	12·27	0·19	7·31	11·47	2·36	0·563	0·273	99·25									
1931	1931HM	49·71	2·730	12·95	12·37	0·18	7·36	11·50	2·06	0·546	0·278	99·68									
1934	1934	49·25	2·638	13·05	12·17	0·19	7·65	11·44	2·10	0·526	0·261	99·28									
1952	KL1952S	49·69	2·721	13·51	12·36	0·17	7·03	11·17	2·40	0·554	0·272	99·88	24·7	383	10·5	22·0	183	122	127	280	322
1954	1954-1	50·12	2·714	13·28	12·00	0·17	7·01	11·37	2·31	0·555	0·269	99·80	24·6	381	10·1	21·2	180	113	117	312	313
1954	1954-2	49·51	2·727	13·30	12·22	0·17	7·17	11·41	2·18	0·552	0·274	99·51	24·5	382	10·3	21·7	183	119	124	293	313
1954	K1954-1	49·66	2·713	13·28	12·27	0·17	7·13	11·40	2·25	0·553	0·270	99·70	25·2	386	9·8	21·5	180	123	119	315	309
1959	1959	46·64	1·966	9·61	12·77	0·17	17·73	8·49	1·85	0·407	0·200	99·83	17·9	280	7·9	15·5	131	113	794	1247	235
24 Feb. 1961	PA209-08	50·03	2·760	13·26	11·80	0·17	7·62	11·20	2·03	0·570	0·290	99·73	25·6	381	10·3	19·3	180	115	129	334	314
10 Jul. 1961	K61-3	50·00	2·737	13·26	11·81	0·17	7·55	11·17	2·23	0·562	0·288	99·78	25·5	380	10·2	18·5	179	114	122	321	305
10 Jul. 1961	K61-22	49·89	2·716	13·23	11·95	0·17	7·64	11·17	2·57	0·561	0·284	100·18	25·9	380	10·0	20·1	184	114	146	345	298
Feb. 1968	HM68-12	49·63	2·639	13·07	12·50	0·16	7·54	11·10	2·20	0·519	0·270	99·63	25·0	383	9·0	17·1	179	119	116	342	295
Jul. 1968	HM68-15	50·07	2·635	13·14	12·32	0·17	7·47	11·05	2·50	0·535	0·276	100·17	24·9	374	8·8	17·5	182	117	125	326	295
Aug. 1971	1971A-1	49·97	2·585	13·26	12·25	0·17	7·31	11·15	2·31	0·504	0·267	99·78	24·8	366	9·2	16·6	171	116	105	307	289
Aug. 1971	1971A-2	49·35	2·235	12·44	12·76	0·18	10·08	10·40	2·21	0·413	0·222	100·29	22·5	313	6·8	13·8	143	118	218	600	273
Aug. 1971	1971A-3	49·28	2·240	12·38	12·54	0·18	10·02	10·34	1·91	0·413	0·226	99·53	22·0	313	7·4	12·7	140	116	216	612	271
Aug. 1971	1971A-4	49·39	2·236	12·32	12·54	0·18	10·02	10·36	2·07	0·411	0·227	99·75	22·2	310	7·1	12·9	139	116	216	607	265
Aug. 1971	1971A-5	49·60	2·224	12·33	12·47	0·18	10·12	10·34	1·92	0·403	0·223	99·81	22·2	303	6·7	12·3	137	114	231	617	268
Sep. 1971	1971S	50·16	2·591	13·27	12·36	0·17	7·29	11·20	2·39	0·508	0·267	100·21	24·6	362	9·0	17·3	173	117	111	313	285
Sep. 1971	K1971-1	50·11	2·562	13·28	12·26	0·17	7·29	11·13	2·25	0·505	0·270	99·83	25·1	367	9·0	16·9	172	122	109	317	298
Jul. 1974	1974 J	50·36	2·525	13·22	12·27	0·17	7·48	11·19	2·33	0·479	0·255	100·28	24·3	352	8·5	15·6	164	118	110	331	296
Jul. 1974	1974 J-2	50·23	2·514	13·21	12·14	0·18	7·55	11·14	2·16	0·473	0·258	99·86	24·1	353	8·2	14·7	158	111	106	317	279

Table 4: continued

Date	Sample	SiO_2	TiO_2	Al_2O_3	Fe_2O_3*	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Total	Y	Sr	Rb	Nb	Zr	Zn	Ni	Cr	V
Jul. 1974	1974 J-3	49·44	2·236	12·32	12·44	0·18	10·00	10·38	1·95	0·404	0·222	99·57	22·1	311	7·0	12·3	136	110	201	592	261
Jul. 1974	1974 J-4	49·38	2·225	12·36	12·45	0·18	10·03	10·45	2·01	0·408	0·224	99·72	22·4	306	7·1	12·5	137	114	223	608	269
Sep. 1974	1974S	50·66	2·553	13·12	12·35	0·17	7·24	11·02	2·35	0·487	0·258	100·21	24·6	348	8·0	15·7	168	115	115	324	293
Sep. 1974	K1974-1	50·43	2·495	13·27	12·17	0·17	7·28	10·92	2·30	0·480	0·250	99·77	25·3	351	8·0	16·1	166	120	108	325	292
Dec. 1974	1974D-E2	48·48	2·192	11·34	12·43	0·17	13·00	9·26	2·02	0·436	0·219	99·55	21·2	292	7·4	12·7	136	112	454	879	252
Dec. 1974	1974D-E3	48·76	2·206	11·50	12·46	0·18	13·04	9·40	2·02	0·417	0·221	100·20	21·3	296	7·7	12·5	137	112	455	837	251
1975	1975	50·21	2·508	13·36	12·14	0·17	7·16	11·19	2·39	0·476	0·252	99·86	25·5	350	8·4	16·0	165	117	110	299	287
Apr. 1982	1982A-1	50·17	2·472	13·36	11·92	0·17	7·15	11·25	2·22	0·458	0·246	99·42	25·8	351	8·1	16·2	160	116	121	296	301
Apr. 1982	1982A-20	50·38	2·482	13·42	11·94	0·17	7·27	11·29	2·36	0·462	0·245	100·02	25·5	351	8·2	16·1	158	116	114	290	298
Apr. 1982	K1982-1	50·48	2·488	13·46	12·02	0·17	7·21	11·30	2·29	0·465	0·250	100·13	25·6	351	8·0	15·4	164	117	111	325	285
Apr. 1982	K1982-3	50·59	2·477	13·46	12·02	0·17	7·21	11·28	2·27	0·464	0·250	100·19	25·0	349	7·9	15·5	163	119	114	295	292
Sep. 1982	1982S-10	50·37	2·500	13·37	11·95	0·17	7·50	11·20	2·21	0·470	0·250	99·99	24·9	349	8·0	15·7	161	116	131	343	302
Sep. 1982	1982S-12	50·29	2·513	13·40	11·86	0·16	7·25	11·12	2·26	0·478	0·253	99·58	25·1	352	8·5	15·7	159	117	124	312	297
Sep. 1982	1982S-13	49·91	2·384	12·89	12·05	0·17	8·72	10·84	2·19	0·446	0·237	99·84	24·1	338	7·3	15·0	151	121	189	451	293
Sep. 1982	1982S-14	50·47	2·609	13·69	11·83	0·16	6·84	10·92	2·37	0·498	0·268	99·66	26·6	361	8·1	16·5	171	132	111	230	303
Sep. 1982	K1982-2	50·18	2·395	12·97	12·06	0·17	8·46	10·89	2·26	0·449	0·240	100·07	24·4	339	7·4	15·0	159	118	163	444	287

*Fe_2O_3 is total iron.

Oxides are in wt %; trace elements are in ppm. The 17th century eruption dates for the oldest samples are based on ^{14}C dating (D. Swanson, personal communication, 2001). Analyses of rock standards Kil1919 and BHVO-1 are listed for comparison with published values (see Rhodes, 1996). No trace element data were determined for very small samples (<10 g).

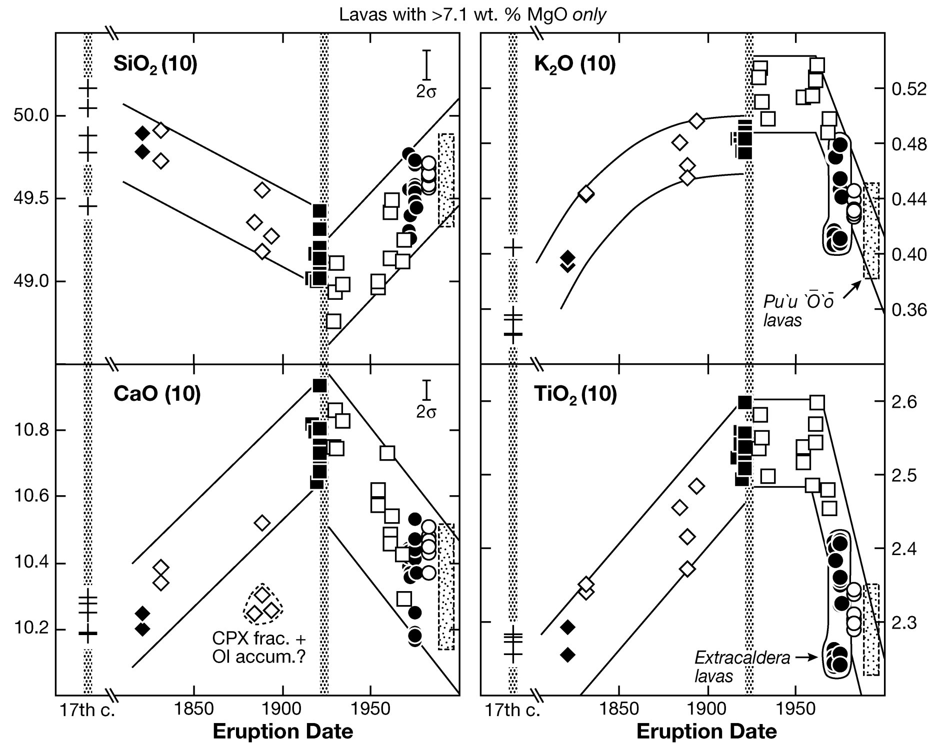

Fig. 5. Temporal major element variation for samples from Kīlauea historical summit eruptions and Pu`u `Ō`ō (stippled block). Only samples with MgO contents $>$7·1 wt % were considered in this plot to minimize the effects of fractionation beyond olivine control. Compositions were normalized to 10 wt % MgO by addition or subtraction of equilibrium olivine in small steps (0·5 mol %). Some of the late 19th century lavas have distinctly low CaO, which may be related to clinopyroxene fractionation and olivine accumulation in these rocks. The more MgO-rich extra-caldera lavas ($\sim$10 wt %) have somewhat lower concentrations of incompatible elements and CaO. The stippled bars are for periods of explosive eruptions during the late 17th century and in 1924. (Note break in scale before 1820.)

with MgO concentrations $<$6·8 wt % MgO are considered to have differentiated beyond olivine-only control and none were thought to be present at the summit (Wright, 1971; Wright & Fiske, 1971). Eleven of our summit samples (mainly from 1885 and 1894 eruptions), however, have MgO concentrations below 6·8 wt % (Table 4). The lower CaO/Al_2O_3 ($<$0·81) for these rocks compared with more MgO-rich samples (Fig. 4) and the presence of rare phenocrysts and abundant microphenocrysts of clinopyroxene or plagioclase (Table 2) confirm that these lavas differentiated beyond olivine control. Nevertheless, we concur with previous studies (e.g. Wright, 1971) that olivine is the dominant mineral crystallizing in Kīlauea summit lavas and that major-element variations are best shown on MgO diagrams.

K_2O (Fig. 4), TiO_2, and P_2O_5 are incompatible in historical Kīlauea summit lavas, forming curved trends when plotted against MgO. These oxides display considerable variation at the same MgO value compared with their analytical error (Fig. 4). The CaO and Fe_2O_3 data show broad trends with a bend at $\sim$7·0 wt % MgO. In contrast, the Al_2O_3 data define one relatively tight trend to $\sim$6·5 wt % MgO, where Al_2O_3 decreases slightly possibly reflecting plagioclase fractionation. The occurrence of clinopyroxene fractionation in summit lavas is suggested by the distinctly lower CaO/Al_2O_3 for lavas with MgO $\leq$7·0 wt % (Fig. 4). However, there is considerable variation in CaO/Al_2O_3 even among olivine-only lavas with MgO $>$7·5 wt %, pointing to possible variations in parental magma composition (Fig. 4). Similar variations have been observed for other suites of Hawaiian tholeiitic lavas (e.g. Frey & Rhodes, 1993).

Kīlauea historical summit lavas vary systematically in composition with age (Fig. 5) when lavas with MgO $>$7·1 wt % (those that have not experienced significant fractionation beyond olivine control) are normalized to 10 wt % MgO by addition or subtraction of equilibrium composition olivine in small steps (0·5 mol %). Incompatible element abundances and CaO increase and SiO_2 decreases from the 19th century to 1923 ending with the 1924 summit collapse and explosions. Lavas erupted from 1929 to 1961 show little change but subsequent summit lavas (1968–1982) and those from the current eruption at Pu`u `Ō`ō have lower incompatible element and CaO contents and higher SiO_2 (Fig. 5), indicating a reversal in compositional variation. It is important to note that this variation is not related to alteration (a common problem in older Hawaiian tholeiites; e.g. Lipman *et al.*, 1990) because the summit lavas have K_2O/P_2O_5 ratios of 1·6–2·0, which is characteristic of unaltered Kīlauea lavas (e.g. Wright, 1971; Garcia *et al.*, 2000), and they appear unaltered in thin section.

Many of the eruptions centered in the Halema'uma'u crater area (Fig. 1) had limited compositional variations (from 0·1 wt % MgO for the 1 day April 1982 eruption to 0·7 wt % MgO for the voluminous eruptions during 1919; Table 4). In contrast, there was a large compositional variation (3·1 wt % MgO) for the 1921 eruptions. Eruptions on the flanks of the caldera, even those of short duration, typically have large compositional variations (e.g. 2·7 wt % MgO for the <1 day August 1971 eruption). The origins of these two distinct compositional trends for individual eruptions are discussed below.

Trace elements

Historical Kīlauea summit lavas exhibit substantial trace element variations (Table 4). Enrichment factors (total compositional variation) for incompatible elements in these lavas range from 1·9 for Rb and Nb (and K), and ~1·5 for Sr and Zr, to ~1·3 for Y and V. These factors are somewhat less than observed for Hilina basalts with the same MgO range (3·4–1·9; Chen *et al.*, 1996), although the largest Hilina variations (3·4 for Rb and 2·8 for K) are related to alteration. Variations for alteration-resistant elements are also larger for Hilina lavas (e.g. 2·5 for Nb and 1·8 for Y). However, ratios of these elements (Nb/Y) yield identical, although somewhat offset ranges (0·61–0·97 for Hilina vs 0·53–0·89 for historical summit lavas) for data collected in the same laboratory (Chen *et al.*, 1996; Table 4). Thus, the extent of partial melting and source composition variations were comparable for these two rock groups, which is surprising because of the very different age ranges represented by these basalts (~75 ka vs 300 years).

Rb, Nb, Zr and Sr data for the historical summit lavas define broad trends well beyond analytical error on K variation plots (Fig. 6). As with the major element data, the variations can be subdivided into age groups, with both older and younger lavas having lower concentrations of incompatible elements at a given K, relative to intermediate-age lavas (Fig. 6). The slight kink in the Sr vs K plot for 1885 lavas (Fig. 6) supports our interpretation, based on modal and Al_2O_3 evidence, for plagioclase fractionation in the more evolved lavas (<6·5 wt %; Table 2; Fig. 4). Some of the other lavas have relatively low Sr contents (1868, 1959, December 1974), which may be related to mixing with fractionated magma.

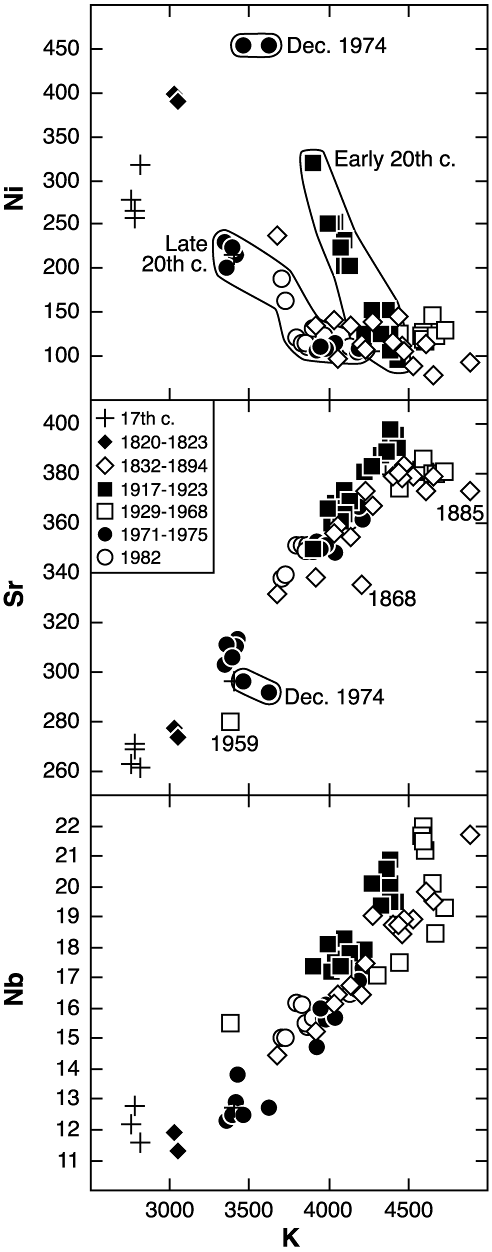

Fig. 6. Potassium variation diagrams for trace elements in Kīlauea historical summit lavas. The age groupings are the same as in Fig. 4. (Note the small kink in the Sr trend for the samples with the highest potassium concentration and that several samples fall off the main compositional trend.) Ni and Nb concentrations range widely for a given K value and can be subdivided into at least two trends (early and late 20th century). The 2σ errors are smaller than the symbol size.

Ratios of incompatible trace elements (e.g. Nb/Y) show systematic variations among the lavas, increasing in value from 1820 to 1959 and then decreasing during the rest of the 20th century including during the

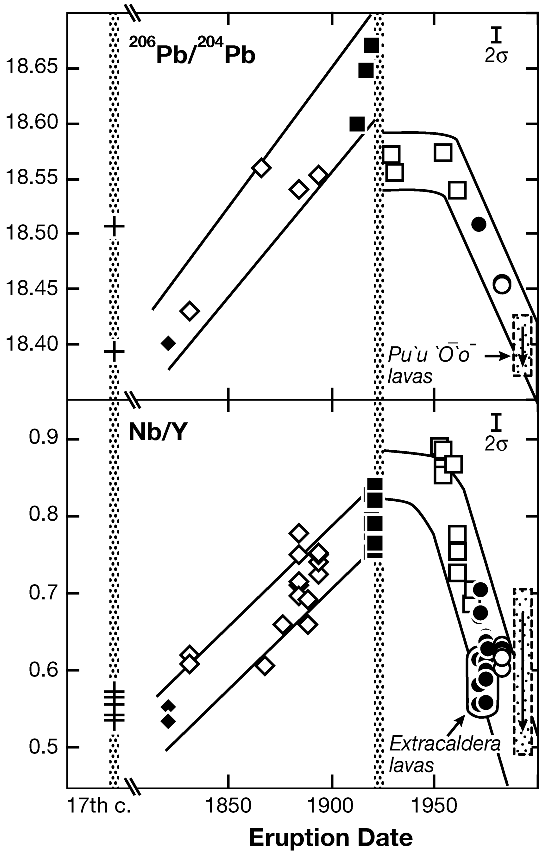

Fig. 7. Temporal variation in Nb/Y and $^{206}Pb/^{204}Pb$ ratios for samples from Kīlauea historical summit eruptions and Pu`u `Ō`ō (stippled block). Same symbols as in Fig. 4. Patterned vertical bars indicate major Kīlauea explosive eruptions in late 17th century and 1924. No XRF trace element data are available for our 1929–1934 samples because insufficient material (<10 g) was available. Bounding lines on the Nb/Y plot are for ICP-MS data from Pietruszka & Garcia (1999*a*). The change in compositional variation following the 1924 explosive event, which is continuing during the Pu`u `Ō`ō eruption (arrow indicates temporal variation trend), should be noted. There is a break in the time axis separating the late 17th century and the 1820 samples. The 2σ error bars are given in the upper right corner of each plot. Data sources: Nb/Y—Table 4 and Garcia *et al.* (1996*b*, 2000) for Pu`u `Ō`ō; $^{206}Pb/^{204}Pb$ ratios—Garcia *et al.* (1996*b*, 2000) and Pietruszka & Garcia (1999*a*).

Pu`u `Ō`ō eruption (Fig. 7). Similar temporal variations were observed for inductively coupled plasma mass spectrometry (ICP-MS) determined trace elements and Pb isotopic ratios (Pietruszka & Garcia, 1999*a*), but the trend reversal occurs after the 1924 collapse of the summit crater (Fig. 7). Unfortunately, we have no XRF trace element data for the 1929–1934 samples because of their very small size. However, our XRF database is more extensive for post-1934 eruptions and these data reveal that the August 1971 and July 1974 high-MgO lavas erupted from caldera-rim vents (Fig. 2) have lower Nb/Y ($<0{\cdot}60$) than coeval lavas from intracaldera vents (Fig. 7).

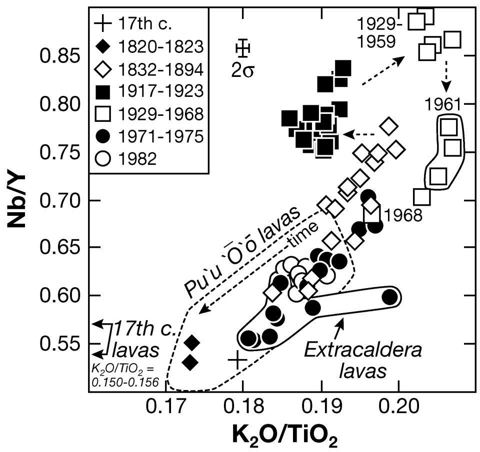

Fig. 8. Incompatible element ratio–ratio plot for Kīlauea historical summit lavas. The 17th century samples plot off the diagram to the lower left. The deviations from the overall positive trend for these ratios after the 1894 and the 1959 eruptions should be noted. Pu`u `Ō`ō lavas also show a progressive decrease in these ratios with time, as shown by arrow (data from Garcia *et al.*, 1996*b*, 2000). The arrows pointing from the 1832–1894 data to the 1917–1921 symbols and from these data to the 1929–1959 symbols highlight the short-term shifts in summit lava composition. The 2σ error bars are given for reference.

Another interesting observation is that the composition of the 1959 Kīlauea Iki eruption lava plots within the distinctive field for 1952 and 1954 Halema'uma'u eruptions (Figs 7 and 8). Previous studies had suggested that one component of the 1959 lava composition bypassed the summit reservoir (Helz, 1987) and the other component was of the same composition as 1929–1934 reservoir lava (Wright & Fiske, 1971). Our whole-rock major and trace element results indicate that the 1959 lavas are chemically identical to those for reservoir lavas from the 1950 s (Figs 7 and 8) except for their high olivine contents. In contrast, subsequently erupted lavas are compositionally distinct.

Ni (Fig. 6) and Cr are compatible in the historical Kīlauea summit lavas and display considerable variations when plotted against a highly incompatible element such as K. These variations can be subdivided into subparallel age trends as observed for the major elements (Figs 5 and 6). Lavas from the current Pu`u `Ō`ō eruption also show a temporal Ni–K trend with a progressive decrease in K for the younger lavas (Garcia *et al.*, 2000). Zn concentrations in summit lavas are nearly constant for all but two samples (115 ± 15 ppm; Table 4).

GLASS GEOCHEMISTRY

Kīlauea summit lavas usually have glassy crusts. Thirty-three glassy crusts from 19 separate eruptions were analyzed for major elements by electron

Table 5: Electron microprobe analyses of Kīlauea historical summit glasses

Date	Sample	*n*	SiO_2	TiO_2	Al_2O_3	FeO^*	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	Total	*T* (°C)
17th century	1790-6	18	51·24	2·47	13·85	10·96	0·17	6·94	11·34	2·26	0·43	0·25	99·91	1154
1820–1823	1820-1	5	50·95	2·27	12·90	11·23	0·18	9·29	10·66	2·08	0·40	0·22	100·18	1201
1868	1868	10	50·73	2·78	13·08	11·79	0·17	6·28	10·58	2·45	0·53	0·30	98·69	1140
1885	1885-5	10	50·65	3·35	12·98	12·40	0·17	5·81	10·44	2·51	0·65	0·34	99·30	1131
1894	1894-3	10	50·44	3·07	13·37	11·98	0·16	6·31	11·05	2·37	0·61	0·32	99·68	1141
1912	95-TAJ-3	5	50·76	2·93	13·74	11·22	—	6·83	11·31	2·38	0·57	0·25	99·99	1151
1917	1917	5	50·54	2·69	13·94	10·75	0·18	7·08	11·72	2·26	0·51	0·29	99·96	1156
1919	1919-3	6	50·03	3·05	13·44	11·62	0·16	6·60	11·41	2·33	0·60	0·33	99·57	1147
1919	1919-4	10	50·45	3·01	13·55	11·21	0·17	6·55	11·35	2·38	0·56	0·29	99·52	1146
1919	1919-5	5	49·94	3·24	13·55	12·00	0·18	6·41	10·99	2·42	0·60	0·26	99·60	1143
1919	1919-6	5	50·21	3·20	13·48	12·12	0·20	6·27	11·01	2·46	0·62	0·30	99·88	1140
1919	1919-7	5	50·20	2·89	13·98	11·26	0·17	6·88	11·59	2·33	0·52	0·25	100·07	1152
1921	1921-4	14	50·32	2·99	13·92	11·23	0·17	6·66	11·43	2·51	0·57	0·28	100·08	1148
1921	1921-5	5	50·45	3·00	13·80	11·35	0·17	6·40	11·40	2·32	0·53	0·26	99·68	1143
1921	1921-8	5	50·00	2·91	14·17	11·07	0·20	6·78	11·45	2·38	0·55	0·26	99·76	1150
1929	1929	7	50·26	2·94	13·56	11·61	0·18	6·72	11·31	2·33	0·60	0·29	99·80	1149
1934	1934	5	50·42	2·68	13·61	11·09	0·17	7·29	11·70	2·24	0·53	0·28	100·00	1161
1959	1959	10	50·50	2·81	13·77	10·61	0·16	7·10	12·16	2·30	0·53	0·29	100·23	1157
1961	K61-22	10	50·49	2·77	13·33	10·50	0·18	7·07	11·49	2·39	0·57	0·29	99·08	1156
Aug. 1971	1971A-3	5	50·52	2·50	13·80	11·40	0·17	7·32	11·45	2·29	0·45	0·20	100·11	1161
Aug. 1971	1971A-4	5	50·42	2·45	13·55	11·40	0·17	7·94	11·29	2·26	0·44	0·22	100·14	1174
Aug. 1971	1971A-5	5	49·48	2·33	12·98	11·53	0·19	9·30	10·76	2·14	0·41	0·20	99·32	1201
Sep. 1971	1971S	10	50·94	2·79	14·10	11·22	0·17	6·53	11·12	2·39	0·51	0·29	100·06	1145
Jul. 1974	1974 J	10	50·75	2·69	13·92	10·92	0·18	6·40	11·25	2·35	0·51	0·21	99·18	1143
Jul. 1974	1974 J-2	5	50·54	2·77	13·91	11·75	0·20	6·46	11·07	2·40	0·53	0·26	99·88	1144
Jul. 1974	1974 J-3	5	50·51	2·52	14·20	11·27	0·13	6·36	11·38	2·35	0·45	0·20	99·36	1142
Jul. 1974	1974 J-4	5	50·24	2·42	13·57	11·30	0·15	7·84	11·23	2·21	0·44	0·19	99·59	1172
Sep. 1974	1974S	10	50·91	2·60	14·10	10·77	0·17	6·76	11·32	2·33	0·48	0·24	99·68	1150
1975	1975	10	51·01	2·57	14·08	10·73	0·18	6·89	11·42	2·33	0·47	0·23	99·91	1152
Sep. 1982	1982S-10	20	51·43	2·76	13·77	11·22	0·17	6·41	11·12	2·43	0·53	0·27	100·11	1143

*FeO is total iron.

Oxides are in wt %; *n*, number of replicate analyses. Glass temperatures were estimated using the MgO geothermometer of Helz & Thornber (1987). Samples 1921-7 and 95-TAJ-3 were analyzed only by microprobe because of limited sample volume.

microprobe (Table 5) to better understand the liquid lines of descent for Kīlauea lavas. Glasses from lavas with bulk MgO contents <8·0 wt % typically have only slightly lower MgO content than the whole rock. This similarity in composition reflects the combined effects of sparse olivine phenocrysts in most of these samples and rapid quenching. Among the samples with higher MgO, only the 1820 tephra and August 1971 lava from caldera-rim vents have glasses with high MgO contents (9·3 wt %). The other MgO-rich lavas have lower MgO glasses reflecting either accumulation of olivine (e.g. 1959) or extensive cooling prior to eruption. Estimates of glass-quenching temperatures range from ~1130 to 1200°C based on the glass MgO geothermometer for Kīlauea compositions (Helz & Thornber, 1987). The relatively high temperature (>1150°C) for many of these glasses reflects sample collection at or near its vent (Fig. 2).

Crystallization of clinopyroxene and plagioclase in Kīlauea summit glasses occurs at ~7·0–7·2 and ~6·4–6·8 wt % MgO, respectively, based on inflections of CaO and Al_2O_3 trends on MgO-variation diagrams (Fig. 9) and the presence of these minerals as microphenocrysts (Table 2). This crystallization sequence is consistent with previous petrographic observations of Kīlauea lava lake glasses (e.g. Wright & Peck, 1978), experimental work on Kīlauea lavas (e.g. Helz & Thornber, 1987), and results from

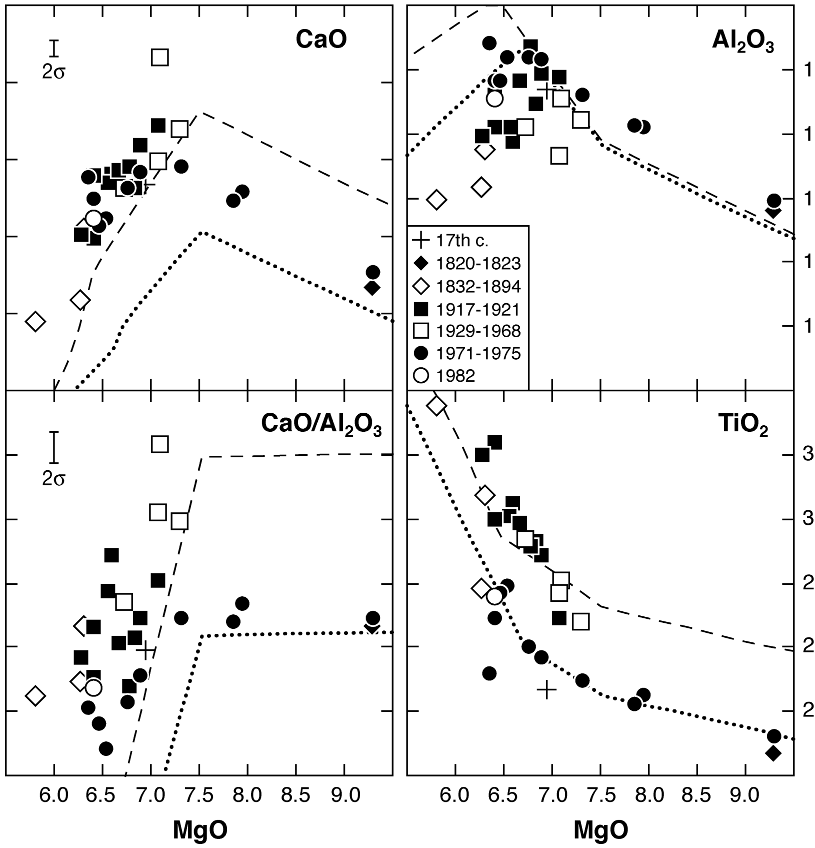

Fig. 9. MgO variation diagrams for glasses from Kīlauea historical summit lavas. The dotted and dashed lines are the liquid lines of descent determined from MELTS modeling (see text for details). The similarity of these lines to the observed curved or kinked trends for recent Kīlauea summit lava glasses should be noted. This similarity supports the inference from whole-rock data that crystal fractionation has played an important role in controlling compositional variation in Kīlauea summit lavas. The 2σ error bars are shown for CaO and CaO/Al_2O_3; the errors for the other oxides are about the size of the symbols.

calculations using the MELTS program (Fig. 9) of Ghiorso & Sack (1995).

Kīlauea summit glasses show temporal groups on MgO variation plots (Fig. 9) with the older and younger groups having similar composition, like the whole-rock analyses (Fig. 4). These data indicate that Kīlauea historical magmas had at least two distinct liquid lines of descent.

TEMPORAL AND SPATIAL VARIATIONS

Mineralogy

Substantial mineralogical variation has occurred in Kīlauea lavas during the last 100 kyr (Table 6). The ~23–100 ka Hilina basalts are predominantly olivine-phyric (Easton & Garcia, 1980). Historical summit lavas are mostly weakly olivine-phyric to aphyric lavas (Tables 2 and 6). This overall mineralogical variation to less porphyritic lavas is the opposite of those reported by Macdonald (1949) and Easton & Garcia (1980). If lavas from the rift zones are included in this comparison, the overall trend towards less olivine-phyric lava remains, because the voluminous Pu`u `Ō`ō eruption (which is comparable in lava production with the summit from 1820 to 1982; ~2 km^3) has to date only produced weakly olivine-phyric lava (e.g. Garcia *et al.*, 2000). However, this eruption is thought to be supplied only partly from the summit reservoir (Garcia *et al.*, 1996*b*).

The nature of Kīlauea's recent mineralogical change is revealed in the walls of its caldera. All but the lowermost of the 20 flows overlying the ~2·1 ka Uwekahuna Ash are moderately to strongly olivine-phyric (>5 vol. %) with high MgO contents (9–20 wt %; Casadevall & Dzurisin, 1987). In contrast, the subsequent 30 prehistoric flows are all aphyric or weakly olivine-phyric basalts (<2 vol. %) with moderate MgO contents (6·3–8·0 wt %; Casadevall & Dzurisin, 1987). With few exceptions and only from vents outside the caldera (1959 and December 1974;

Table 6: Mineralogical variations in lavas erupted from Kīlauea's summit area

	Basalt (<5 vol. % olivine)	Olivine basalt (5–15 vol. % olivine)	Picritic basalt (>15 vol. % olivine)
Historical (1790–1982)	90	6	4
Prehistoric (>210 yr to 2·8 ka)	70	20	10
Hilina Basalts (∼23–100 ka)	39	59	2

Data sources: Hilina Basalts—Easton & Garcia (1980) for lavas exposed in an ∼300 m thick section south of the caldera (these lavas are thought to be from overflows of the caldera); prehistoric lavas—Casadevall & Dzurisin (1987) and Casadevall (unpublished data, 1997) for lavas in an ∼125 m thick section from northern caldera wall; historical lavas—Table 1.

Table 2), this tendency to erupt weakly porphyritic to aphyric basalt has persisted during historical summit eruptions. However, historical eruptions of olivine-rich basalts are more common along Kīlauea's rift zones (e.g. 1790, 1840, 1921, 1960, 1968 and 1969–1974). Therefore, the dominance of weakly porphyritic lavas at Kīlauea's summit during the last ∼2000 years probably reflects processes in the summit magma reservoir.

A spatial mineralogical variation is also evident for Kīlauea's 20th century summit lavas. Lavas erupted in or from fissures extending from Halema'uma'u crater since the 1924 summit collapse are weakly olivine-phyric or aphyric (Table 2). In contrast, many of the lavas erupted during the last half of the 20th century from vents on the fringes of the caldera contain more olivine. These more olivine-rich lavas follow the overall temporal geochemical trend for summit lavas (Fig. 6), so they presumably must have been derived from the same magma reservoir.

Several models have been proposed to explain mineralogical variations for Hawaiian shield volcanoes. A stratified magma reservoir tapped at different levels was invoked by Macdonald (1942) to explain the eruption of olivine-rich magmas on the flanks of these volcanoes. An updated version of this model (Ryan, 1988) advocates that lower density (∼2·7 g/cm^3), weakly olivine-phyric magmas are able to ascend into the summit reservoir, whereas the higher density (∼2·8–2·9 g/cm^3), olivine-rich magma is stored at depth and periodically supplied to the rift zones. This model is consistent with the observation that olivine-rich lavas are preferentially erupted along the submarine rift zones of Hawaiian shield volcanoes (Garcia *et al.*, 1989, 1995).

Alternatively (or in addition), during periods of high-magma supply olivine may be flushed from the base of the summit magma reservoir to produce olivine-rich lavas (e.g. Rhodes, 1995). If so, this would imply a general decrease in magma supply to Kīlauea during the last few thousand years. However, there is no evidence for a long-term decrease in Kīlauea's magma supply or of a major change in magma composition during the last 350 kyr (Quane *et al.*, 2000), as would be expected if magma supply rates (and extents of melting) had changed significantly. Furthermore, olivine-rich subaerial lavas were erupted along Kīlauea's rift zones throughout most of the historical period (1840–1974) despite wide variations in the magma-supply rate (0·01–0·18 km^3/yr; Dvorak & Dzurisin, 1993; Cayol *et al.*, 2000). Olivine-rich lavas are more common at the summit of Kīlauea during the periods of higher magma supply (1790–1820 and 1959–present), so fluctuations in magma-supply rate may play a role in influencing temporal variations in mineralogy. However, except those from the 1959 eruption (Helz, 1987), olivines in Kīlauea lavas are overwhelming undeformed, euhedral and were probably grew in their host magma before eruption (Fig. 3). Thus, periods of higher magma supply probably allow less time for crustal-level differentiation and favor eruption of olivine-rich magma.

Long-term geochemical trends

Historical Kīlauea summit lavas display two distinct geochemical variation trends with a change in direction following the 1924 summit collapse (Fig. 5). This reversed trend is continuing during the current Pu'u 'Ō'ō eruption (Figs 5 and 7). Systematic compositional variations are also observed for lavas from other Hawaiian shield volcanoes (e.g. Mauna Loa; Rhodes & Hart, 1995, and Mauna Kea; Albarède, 1996; Yang *et al.*, 1996; Blichert-Toft *et al.*, 2003) and from other frequently active ocean island volcanoes (e.g. Piton de la Fournaise; Albarède & Tamagnan, 1988). The geochemical variations at Kīlauea and other volcanoes must be related to small-scale source heterogeneities because they are accompanied by rapid Pb, Sr, and Nd isotope ratio variations (e.g. Fig. 7). Systematic variations in the proportions of source components in Hawaiian magmas appear to be fundamental in the magma genesis for these volcanoes. It is our interpretation that Kīlauea's two distinct compositional variation trends (Fig. 7) are predominantly controlled by fluctuations in the composition of parental magmas delivered from the mantle rather than the result of magma mixing in the summit reservoir over periods of decades to centuries (Pietruszka & Garcia, 1999*a*).

Kīlauea's current trend of decreasing incompatible element abundance typifies the 'common' melting trend of Reiners (2002), which was identified for relatively primitive basalts from eruptions and eruptive sequences in a variety of tectonic settings including the Pu`u `Ō`ō eruption. However, Kīlauea's current summit trend does not show decrease in MgO (Garcia *et al.*, 2000) as predicted for the common trend from mixing melts produced by different degrees of melting of two compositionally distinct sources with different solidi and melt productivities (peridotite and pyroxenite; Reiners, 2002). Another complication for the two-source common melting model is the need for three source components to explain the compositional and isotopic data for Kīlauea's historical lavas (Garcia *et al.*, 2000). The nature of these components remains enigmatic but the lack of a systematic MgO variation (Fig. 4) and the relatively low Os isotopic and SiO_2 contents of Kīlauea lavas (Hauri, 1996) argue that neither source component can be pyroxenite as proposed by Reiners (2002). Instead, the source components are probably all peridotites based on the inferred high MgO content of Kīlauea's parental magmas (e.g. >16 wt %; Clague *et al.*, 1995). The 19th to early 20th century summit trend with increasing incompatible elements (Figs 5 and 7) is opposite to that predicted by the common melting trend.

Historical lavas (1843–1984) from neighboring Mauna Loa volcano display systematic variations in ratios of incompatible elements, although not in major elements (Rhodes & Hart, 1995). The chemical variations of lavas from these two volcanoes are out of phase with each other and they have distinct Pb, Sr and Nd isotopic ratios (e.g. Rhodes & Hart, 1995; Garcia *et al.*, 2000). Thus, historically the lavas from these volcanoes are chemically distinct and are not being supplied by the same mantle conduit, as was suggested based on earthquake locations that appeared to merge at ~60 km below these two volcanoes (Klein, 1982) and by the simultaneous inflation of both volcanoes in May 2002 (Miklius & Cervelli, 2003). However, the strong overlap of some Kilauea lavas erupted from ~0·4 to 2 ka with Mauna Loa compositions led Rhodes *et al.* (1989) to propose that Mauna Loa magma moved laterally at shallow levels within the crust and mixed with Kīlauea magma during that period. There is no evidence of historical magmatic communication between the two volcanoes.

Short-term compositional trends

Significant compositional variation occurred within several brief Kīlauea historical summit eruptions. Lavas became less MgO-rich during the 14 h September 1982 eruption (8·7–6·8 wt %; Table 4), the 10 h August 1971 eruption (11·5–7·3 wt %; Duffield *et al.*, 1982) and the 2·5 day July 1974 eruption (11·0–6·7 wt %; Lockwood *et al.*, 1999), all of which occurred on the margins of the caldera. The lava MgO content decreased progressively with time during the September 1982 eruption. In contrast, MgO variation for the 1971 and 1974 eruptions was apparently spatially controlled. High-MgO lavas (10·0–11·5 wt %) were erupted on the caldera rim, whereas lower-MgO lavas (6·7–7·6 wt %) were erupted from fissures on the caldera floor (Duffield *et al.*, 1982; Lockwood *et al.*, 1999). It is unclear whether these two eruptions had a temporal variation in composition because no lavas were collected inside the caldera during the early part of each eruption when the MgO-rich lavas were erupted on the caldera rim. The early caldera floor lavas were buried by later flows. Overflows of Halema'uma'u lava lake during the early months of 1921 also record large MgO variation (7·5–10·6 wt %). The volumes of magma erupted during these four eruptions are relatively small ($\sim 6 \times 10^6$ m^3 for 1921; 10×10^6 m^3 for August 1971; $6{\cdot}5 \times 10^6$ m^3 for July 1974; 3×10^6 m^3 for September 1982; Duffield *et al.*, 1982; Macdonald *et al.*, 1983; Lockwood *et al.*, 1999), compared with the nearly homogeneous (7·4–7·6 wt % MgO) and much larger ($80{\cdot}3 \times 10^6$ m^3) 1967–1968 eruption in Halema'uma'u crater (Wright, 1971; Macdonald *et al.*, 1983).

Olivines from Kīlauea historical lavas demonstrate the complexities in accounting for the short-term MgO variations. Olivines from the nearly aphyric to weakly porphyritic September 1982, and the lower-MgO 1971 and 1974 lavas, are normally zoned and forsterite contents are in equilibrium with their whole-rock compositions (Fig. 3). In contrast, the MgO-rich August 1971 and July 1974 lavas have lower Fo content crystals with narrow, reversely zoned rims and higher forsterite crystals with normal zoning, which is indicative of mixing of lower and higher MgO content magmas during these eruptions. The variably olivine-phyric 1921 lavas (1·0–4·1 vol. %; Table 1) have a wide range in olivine composition (Fo 81–88), although most are Fo 81–82 despite the wide range in whole-rock Mg number (Fig. 3). The wide compositional range for 1921 lavas cannot be explained solely by variable amounts of accumulated olivine because insufficient olivine is observed (3·1 vol. % total variation) compared with the amount required by mass balance calculations (~8 vol. %).

The high-Fo olivines in lavas from both the overflows of Halema'uma'u crater from 1918 to 1921 and caldera rim eruptions in September 1982, August 1971 and July 1974 require the presence of MgO-rich magmas within the summit reservoir. The progressive compositional change during the September 1982 eruption, with olivines in equilibrium with the host rock, suggests gradual compositional zonation within the reservoir. The decrease in MgO during this eruption may reflect the tapping the flanks of a vertically zoned body, as proposed by Macdonald (1942). Utilizing this model, the sharp compositional changes during the August 1971 and July 1974 eruptions are the result of tapping separate, compositionally distinct parts of the reservoir. The implications of these results for the nature of Kīlauea's summit magma reservoir are discussed below.

PROCESSES WITHIN THE SUMMIT MAGMA RESERVOIR

The detailed sampling of the output of Kīlauea's summit magma reservoir for the past few hundred years allows us to examine the nature, and determine the relative importance, of various petrologic processes that controlled lava composition. Below, we examine magma mixing, crystal fractionation, and assimilation processes in Kīlauea's summit magma reservoir as recorded in its historical lavas.

Magma mixing vs crystal fractionation

Magma mixing is an important process for many volcanoes including Kīlauea (e.g. Wright & Fiske, 1971). The progressive compositional variation of Kīlauea historical summit lavas (Fig. 7) is undoubtedly a consequence of mixing new magma, supplied nearly continuously from the mantle (Tilling & Dvorak, 1993), with magma resident in its summit reservoir. The presence of reversely zoned olivine in some mafic lavas (Fig. 3) reaffirms that mixing occurs in the summit magma reservoir. The coherence of the geochemical variation for Kīlauea summit lavas indicates that new magma is mixed relatively rapidly (<10 years) with magma in the reservoir. For example, lavas erupted during the same year (1982) in different parts of the summit area are compositionally similar (i.e. lie along the same liquid line of descent; Fig. 4) but those erupted in the same area 10 or more years apart are compositionally distinct (e.g. 1960s vs 1970s; Fig. 4). Thus, magma mixing is an important crustal process controlling the overall compositional variation of Kīlauea summit lavas. The kinked and curved major element trends (Fig. 4) are not indicative of magma mixing; rather, they suggest that crystal fractionation has played an important role in controlling the composition of the lavas erupted at Kīlauea.

MELTS modeling (Ghiorso & Sack, 1995) was undertaken to better understand crystal fractionation processes at Kīlauea and to calculate liquid lines of descent for comparison with the observed compositional trends. Two parental magma compositions that span the range of observed compositions (1820–1821 and 1921–1925; Table 4, Fig. 4) were used for modeling. The modeling conditions were fractional crystallization, oxygen fugacity of QFM − 0·2 (where QFM is the quartz–fayalite–magnetite buffer) (within the range observed for Kīlauea; Moore & Ault, 1965), pressure of 0·5 kbar (consistent with a shallow magma reservoir) and water contents of 0·2–0·35 wt % for the parental magmas (somewhat below expected water contents for these parental magmas, 0·48–61 wt %; Wallace & Anderson, 1998). Higher water values delayed the start of plagioclase crystallization beyond its observed point of crystallization (e.g. 0·5 wt % H_2O delayed plagioclase crystallization to ~6·1 wt % MgO vs observed 6·4–6·8 wt % MgO; Figs 4 and 9). This delay may be an artifact of the approximations in MELTS algorithms and not necessarily an indication of the water content of Kīlauea parental magmas.

The crystal fractionation modeling results demonstrate that the large compositional variations observed for several short-lived Kīlauea eruptions (e.g. August 1971, July 1974, and September 1982) and the overall variation for various time segments parallel the calculated liquid lines of descent (Fig. 4). Thus, the observed curved and kinked variation trends are probably related to crystal fractionation rather than to magma mixing, which would produce linear trends. However, the effects of crystal fractionation are superimposed on compositions produced by magma mixing. Olivine crystallization is the dominant control for much of the compositional variation, although many summit lavas also show the effects of clinopyroxene and plagioclase fractionation (Fig. 4). Fractionation of plagioclase and clinopyroxene is especially important during periods of low magma supply (e.g. late 19th to early 20th century; Fig. 10). For example, lavas erupted during 1885–1894 exhibit the greatest extent of fractionation (i.e. contain more plagioclase and clinopyroxene crystals and have lower MgO contents; Fig. 10). During periods of higher eruption rate (e.g. early 19th century), more primitive magmas are erupted as evidenced by the higher forsterite content of olivines and the virtual absence of plagioclase and clinopyroxene in these lavas (Fig. 10; Table 2). No other mineral is observed as a phenocryst or microphenocryst in Kīlauea historical summit lavas (Table 2) or inferred from compositional data (Fig. 4, Table 4) to have crystallized in its summit reservoir. In contrast, some Kīlauea rift zone lavas exhibit petrographic and chemical evidence of orthopyroxene and iron–titanium oxide crystallization (Wright & Fiske, 1971; Ho & Garcia, 1988).

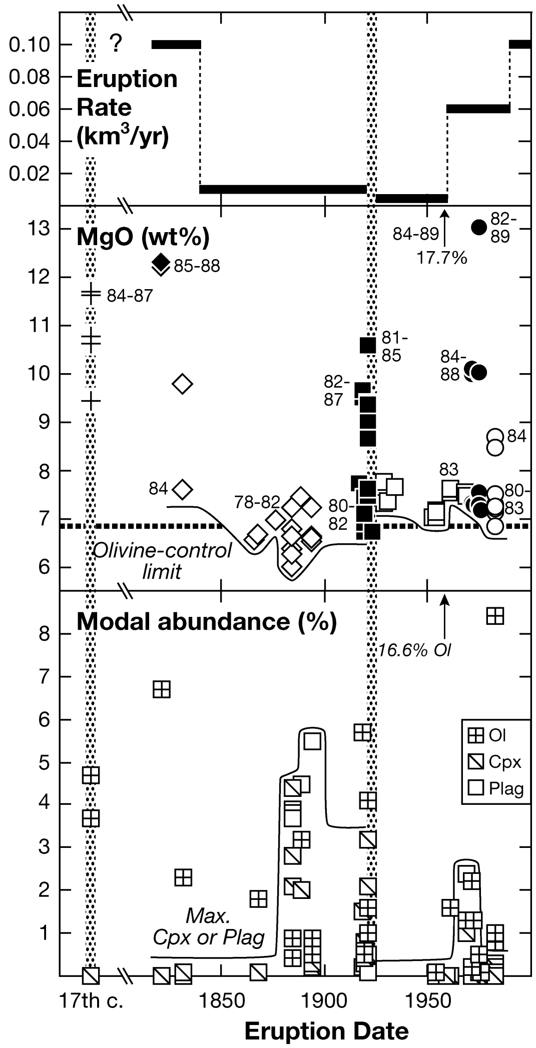

Fig. 10. Variations in Kīlauea's eruption rate, whole-rock MgO content and mineral abundances in historical summit lavas. The large range in values on each plot should be noted. The 1959 picritic lava plots off scale for MgO and modal abundance. Lavas erupted during periods of lower eruption rates tend to have low MgO contents and higher abundances of clinopyroxene (cpx) and plagioclase (plag). These samples also tend to have olivine cores with lower forsterite contents (ranges given next to symbol on the MgO plot). The 2σ errors for modal abundances and MgO contents are smaller than the symbol size; data are from Tables 2 and 4. Eruption rate estimates are from Pietruszka & Garcia (1999*b*). Patterned vertical bars indicate major Kīlauea explosive eruptions in the late 17th century and 1924.

Accumulation of crystals is another potentially important crustal process that can modify lava composition (e.g. Bowen, 1928). Samples from the 1959 and Dec. 1974 Kīlauea summit eruptions have abundant olivine phenocrysts (~15 vol. %) with forsterite contents mostly too low to be in equilibrium with their bulk composition (Helz, 1987; Fig. 3). These picritic basalts probably accumulated significant amounts of olivine. In contrast, most Kīlauea historical summit lavas are weakly olivine-phyric ($\leq$3 vol. %), although they may have MgO contents of 9–10 wt % (Table 2). Thus, among Kīlauea historical summit lavas only the bulk compositions of the picritic basalts have been significantly modified by the mineral accumulation. Therefore, the compositions of the other rocks are representative of pre-eruptive magmatic compositions.

Assimilation

Crustal contamination is another potentially important process affecting the composition of Kīlauea (e.g. Kyser *et al.*, 1982; Hemond *et al.*, 1994) and other oceanic island lavas (e.g. Eiler *et al.*, 1996; Thirlwall *et al.*, 1997). Oxygen isotope analyses demonstrate that crustal assimilation has affected some Kīlauea historical lavas, including those from the current Pu`u `Ō`ō eruption (Garcia *et al.*, 1998). Early lavas from this eruption have low matrix oxygen isotope values ($\delta^{18}O$ = 4·6–5·0‰), which are out of equilibrium with oxygen in coexisting olivine (Garcia *et al.*, 1988). Later lavas have higher matrix oxygen isotope values (5·0–5·25‰) and are in equilibrium with coexisting olivine. The values for these later lavas are identical to those for summit lavas erupted after the 1924 collapse of Halema'uma'u crater (Garcia *et al.*, 1996*a*) but are below normal mantle values (5·5–6·0‰; Ito *et al.*, 1987). In contrast, pre-1924 summit lavas have oxygen isotope values of 5·5 ± 0·1‰ (Kyser *et al.*, 1982; Garcia *et al.*, 1996*a*). The lower oxygen isotope values for post-1924 lavas are not accompanied by any major compositional change (Figs 5 and 6; Table 4), although there are minor differences in Pb, Sr, and Nd isotopes for post-1924 lavas (Pietruszka & Garcia, 1999*a*).

The 1924 collapse of Halema'uma'u caused the crater to grow markedly in size (increasing its depth from 100 to 400 m and volume by ~0·2 km³; Dvorak & Dzurisin, 1993). Stoping of older Kīlauea lavas into the summit magma reservoir probably accompanied this collapse (Pietruszka & Garcia, 1999*a*). The minor compositional and isotopic differences of lavas erupted just after the 1924 collapse can be explained by mixing 3·5% of a melt formed by 5% melting of the stoped blocks with 46% resident reservoir magma and 50·5% new, MgO-rich magma (Pietruszka & Garcia, 1999*a*). Thus, crustal assimilation may affect some Kīlauea summit magmas, especially following a major summit collapse. Our sample suite for the 19th century is too sparse to evaluate whether the summit collapses in 1832, 1840 and 1868 were accompanied by significant compositional variation. However, the impact of the contamination associated with the 1924 collapse on the overall geochemistry of summit lavas appears to be minor. Crustal contamination also had little effect on the composition of Pu`u `Ō`ō lavas (Garcia *et al.*, 1998).

In summary, magma mixing and crystal fractionation are both important magmatic processes controlling compositional variation of Kīlauea historical summit lavas. The effects of crystal fractionation process are superimposed on a changing resident magma composition caused by the mixing of compositionally distinct parental magmas. Catastrophic collapse of Halema'uma'u crater in 1924 may have caused assimilation of Kīlauea rocks but this had a minor effect on the composition of subsequent lavas. The 1924 collapse, however, heralded a much larger magmatic event; the start of a new compositional trend that has continued for over 75 years.

GEOMETRY OF KĪLAUEA'S SUMMIT MAGMA RESERVOIR

The shape and size of Kīlauea's summit reservoir have fundamental implications for how the volcano's magmatic plumbing system operates. Estimates for these features have varied greatly. One model (e.g. Tilling & Dvorak, 1993) based on ground deformation data suggests that the summit reservoir is a plexus of interconnected dikes and sills located 2–6 km beneath the summit (Fiske & Kinoshita, 1969; Decker, 1987). Alternatively, modeling of the same and more recent ground deformation data indicates that the reservoir is a single spherical body (Yang *et al.*, 1992). Estimates of the size of the summit reservoir have varied from 0·08 to 40 km^3 (Klein, 1982; Decker, 1987; Klein *et al.*, 1987). Unfortunately, seismic studies to date have been unable to resolve these discrepancies (e.g. Rowan & Clayton, 1993).

The coherence of the major and trace element whole-rock data (especially the intracaldera lavas; Fig. 6) and isotopic ratios with time for Kīlauea summit historical lavas requires that magma in the summit reservoir mixes efficiently (Pietruszka & Garcia, 1999*a*). Efficient mixing on a time scale of less than a decade in unlikely in a plexus of dikes and sills. Thus, the geochemical data suggest that Kīlauea's summit reservoir is a well-mixed single body (Fig. 11). Our schematic drawing of Kīlauea's summit reservoir utilizes the single spherical body model of Yang *et al.* (1992). This body underlies the southern margin of the caldera near Halema'uma'u (Klein *et al.*, 1987), the site of frequent summit eruptions (Fig. 11), and is thought to be supplied at its base by new mantle-derived magma (Tilling & Dvovak, 1993).

The size of the summit reservoir was estimated using ICP-MS trace element data (Nb/Y) for historical summit lavas and residence time analysis (Pietruszka & Garcia, 1999*b*). For the period from 1820 to just before the collapse of Halema'uma'u in 1924, the reservoir volume was estimated to have been ~2·3–4·6 km^3; from 1961 to 1982, the size was 1·7–2·7 km^3. The range in values reflects the choice of parental magma composition; the smaller values are based on observed historical compositions and the larger values reflect end-member Kīlauea compositions (Pietruszka & Garcia, 1999*b*).

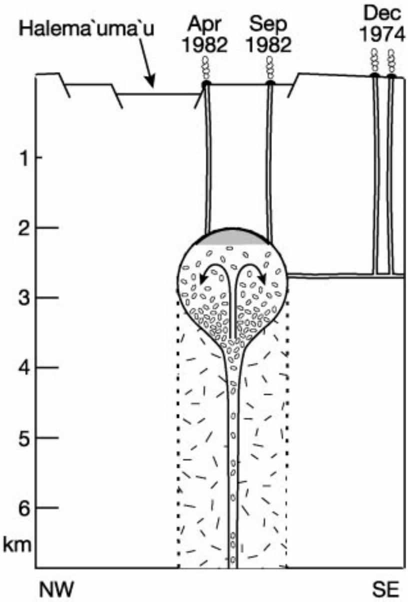

Fig. 11. Schematic cross-section of Kīlauea's late 20th century summit magma reservoir. The dates across the top of the figure show the locations of recent eruptions. Vertical bars represent dikes that fed these eruptions. The reservoir's depth is based on seismic data (Klein *et al.*, 1987). The upper part of magma reservoir (shaded area) contains aphyric, lower-density, moderate MgO (7·0–7·5 wt %) magma. The underlying reservoir contains denser, olivine-phyric magma (blocky symbols represent olivine crystals). Arrows symbolize the relatively rapid mixing of new, mantle-fed magma with reservoir magma. The patterned area below the chamber represents Kīlauea's intrusive core. No vertical exaggeration.

Eruption rates and lava compositions changed significantly from 1820 to 1954 (Figs 5, 7, 8 and 10). Thus, we decided to re-examine the implications of these changes on the size of the reservoir for this period using a trace element ratio (Ba/Sm) that appears to be more sensitive to source compositional variations (Fig. 12) and the residence time analysis model of Albarède (1993). Additional details on this modeling, including a minor correction, have been given by Pietruszka & Garcia (1999*b*). The Ba/Sm variation of the lavas is linear from 1820 to 1894. If we assume that the parental magma being added to the reservoir had 10 wt % MgO and a Ba/Sm ratio comparable with that of the 1894 lava, the reservoir volume is 2·6 km^3, consistent with the results of Pietruszka & Garcia (1999*b*). Varying the MgO content of the input magma from 6·6 to 15 wt % causes only a minor change in volume (±0·2 km^3). However, if Kīlauea's

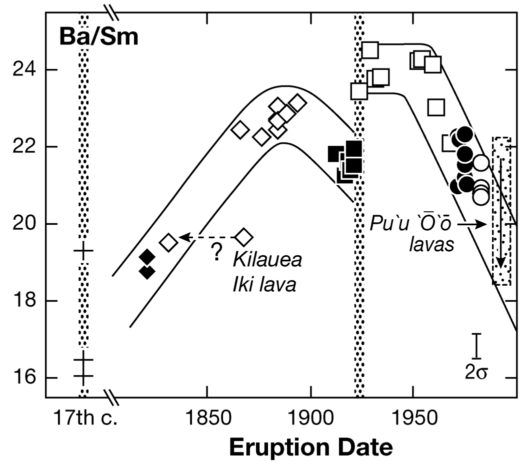

Fig. 12. Temporal variation in Ba/Sm in Kīlauea historical summit and Pu`u `Ō`ō eruption lavas (stippled block). The gray vertical bars mark ∼17th century explosions and the 1924 Halema'uma'u collapse. The decrease in Ba/Sm for lavas erupted just before the 1924 collapse should be noted (see text for discussion). The Kīlauea Iki 1868 lava falls off the coherent temporal trend for Kilauea lavas. This evolved lava may have been related to early 19th century magma as shown by arrow. Symbols are the same as in Fig. 4. Data from Pietruszka & Garcia (1999*a*) and Garcia *et al.* (2000).

highest Ba/Sm is used for the parent magma (from Hilina Basalts; Chen *et al.*, 1996), the size of the reservoir nearly doubles (up to 5·1 km^3 depending on the MgO value of the input magma). This is somewhat larger than the upper value reported by Pietruszka & Garcia (1999*b*). From 1894 to 1921, Ba/Sm decreased (Fig. 12) unlike Nb/Y (Fig. 7). Magma residence time calculations based on this shift to lower Ba/Sm and the lower lava eruption rates for this period (Fig. 10) suggest a dramatic decrease in reservoir size during this period to ∼0·3 km^3 using an average 1920s era parental magma composition or a decrease to ∼0·9 km^3 using Kīlauea's lowest Ba/Sm (1790–1). Based on a shift to higher Ba/Sm values after 1924 (Fig. 12) and the extremely low eruption rates to 1954 (Fig. 10), the calculations suggest a further decrease in reservoir size to $<0{\cdot}1$ km^3 using the observed or end-member ratios. Thus, regardless of the input parent magma composition, there appears to have been a dramatic decrease in the size of Kīlauea's summit reservoir from 1820 to 1954.

Kīlauea had multiple major summit collapses during the period from 1820 to 1954 (Table 1). It is our interpretation that the changes in lava composition and the summit collapses are consequences of a decrease in magma supply rate to the summit magma reservoir (Fig. 10). This decrease in magma supply is thought to be a direct consequence of the gradual decrease in the melt fraction during the 19th to early 20th century at Kilauea (Pietruszka & Garcia, 1999*a*). Subsequently, there was a 17 year gap in Kīlauea eruptive activity, which may have been a period of magma reservoir inflation. The reservoir continued to inflate during a period of increased magma supply and is thought to have reached a size of ∼2 km^3 by 1982, before it began to shrink ∼10% during the Pu`u `Ō`ō eruption (Denlinger, 1997). Interestingly, the late 20th century is inferred to be a period in which the degree of partial melting increased by a factor of ∼2 (Pietruszka & Garcia, 1999*a*). Overall, these variations in the melt fraction are thought to result from differences in the fertility of the source region (Pietruszka *et al.*, 2001).

The main mass of Kīlauea's summit reservoir is compositionally zoned. This interpretation is based on the extremely wide range in whole-rock MgO content for historically erupted Kīlauea lava (6·0–17·8 wt %; Table 4) and in the systematic compositional variation during some historical eruptions. A single, compositionally zoned reservoir model is supported by the presence of euhedral, undeformed high forsterite olivines (Fo 87–88) in more evolved lavas (e.g. 1832 and 1889 eruptions; Fig. 3), which indicates mixing between more primitive magma (from the lower part of the reservoir freshly fed from the mantle conduit) and more evolved magma in the upper part of the reservoir. This primitive component was tapped during three caldera rim eruptions in the 1970 s, which produced MgO-rich lavas with somewhat lower Nb/Y ($<0{\cdot}60$) than coeval intracaldera lavas. The summit reservoir is also zoned in temperature based on glass compositions from the 1970 s eruptions, which indicate at least 40°C for individual eruptions and as much as 60°C for eruptions a few months apart (Table 5).

A crown of lower density [2·56–2·58 g/cm^3; densities calculated using the procedures of Niu & Batiza (1991)] magma overlies the summit reservoir beneath Halema'uma'u (Fig. 11). This interpretation is based on the restricted overall compositional range of lavas erupted in and around Halema'uma'u since its 1924 catastrophic collapse (total MgO range is 7·0–7·7 wt %; Table 4), the presence of olivines in equilibrium with their host rock composition (Fig. 3), and the higher density of the more MgO-rich lavas (2·60–2·78 g/cm^3). Furthermore, the compositional variation for lavas from individual eruptions is remarkably small (e.g. 7·4–7·6 wt % MgO for the large, 80·3 × 10^6 m^3, 1967–1968 eruption; Wright, 1971; Macdonald *et al.*, 1983) compared with the much smaller volume eruptions on the caldera margins (e.g. 6·7–11·0 wt % for 6·5 × 10^6 m^3 of lava during July 1974; Lockwood *et al.*, 1999). The limited compositional variation of these lavas is probably related to two processes: (1) an increase in the enthalpy of crystallization, especially as clinopyroxene begins to crystallize at ∼7·2 wt % MgO, as pointed out by Ghiorso (1995) for basaltic magmas; (2) nearly continuous influx of new magma (Dzurisin *et al.*, 1984). During periods of low magma supply and small, infrequent eruptions, the volume of the crown probably increases and the magma becomes more differentiated (as evidenced by the lavas from the 1885 and 1894 eruptions; Fig. 4). In contrast, during periods of high magma supply rates, more MgO-rich magmas are erupted from the Halema'uma'u crater area (e.g. 1820 eruption).

One enigmatic aspect of the Kīlauea's magmatic plumbing system is the connection between the summit reservoir and the rift zones. Shallow-level (1–3 km deep) earthquake locations (e.g. Klein *et al.*, 1987) have been used to infer that dikes from the shallow portions of the summit reservoir fed the rift zones (Dzurisin *et al.*, 1984). However, at least some intrusions into the more active east rift zone seem to bypass the summit reservoir, based on seismic and geochemical evidence (Ryan, 1988; Garcia *et al.*, 1996*b*). It has also been reported that samples of compositionally distinct magma batches are erupted along the east rift zone before they appear at the summit (Wright & Fiske, 1971), although this is not true for the Pu`u `Ō`ō eruption (Garcia *et al.*, 1992). However, the summit reservoir has shrunk by ~10% during this eruption (Denlinger, 1997), so there must be a connection between the reservoir and the east rift zone. More work is needed on Kīlauea's other rift zone eruptions to gain a better understanding of the interplay between its summit reservoir and rift zones now that we have a geochemical baseline for summit eruptions.

COMPOSITION OF KĪLAUEA PARENTAL MAGMAS: IMPLICATIONS FOR MANTLE MELTING PROCESS AND SOURCE HETEROGENEITY

The minimum MgO content of the parental magmas for historical Kīlauea lavas can be inferred from the compositions of their most forsteritic euhedral, undeformed olivine phenocrysts (e.g. Wilkinson & Hensel, 1988). Over a long time span (1820–1974), Kīlauea has erupted lavas with high forsterite content olivine phenocrysts (88–89%; Fig. 3). Similar olivines have been found in Kīlauea historical rift zone lavas (e.g. Nicholls & Stout, 1988; Wilkinson & Hensel, 1988). Using the experimentally determined equilibrium relationship between olivine and whole-rock compositions (Roeder & Emslie, 1970; Ulmer, 1989), the MgO content of the parental magmas for Kīlauea summit lavas was calculated to be at least 13–14 wt % (assuming 90% of the total iron is Fe^{2+}). This interpretation is consistent with the discovery of submarine Kīlauea glasses with up to 15 wt % MgO (Clague *et al.*, 1991) and estimates for the minimum MgO content for Kīlauea rift zone parental magmas (e.g. Nicholls & Stout, 1988; Wilkinson & Hensel, 1988; Clague *et al.*, 1995). It also demonstrates that Kīlauea's mantle conduit supplies relatively primitive magma to both the summit reservoir and its rift zones.

The rate of compositional change of Kīlauea parental magmas has varied during the last ~200 years, with more rapid changes during the latter part of the 20th century (Figs 5 and 7). This rapid change is continuing during the current Pu`u `Ō`ō eruption, which produced ~50–75% of the summit historical compositional variation during its first 15 years, although only ~20% of the summit Pb isotopic variation. Roughly equal amounts of lava were produced historically at the summit and from the current eruption (~2 km^3, adjusted to dense rock; Dvorak & Dzurisin, 1993; Garcia *et al.*, 2000). Modeling of trace element and isotopic data indicates that the extent of partial melting for Kīlauea magmas has varied by a factor of ~2 during historical times and that the proportions of its source components have varied markedly (Pietruszka & Garcia, 1999*a*). The change in source proportions correlates with large variations in magma supply (<0·01 to 0·18 km^3/yr; Dvorak & Dzurisin, 1993; Cayol *et al.*, 2000) with the lowest supply rates related to the depleted source during the early 20th century (Pietruszka & Garcia, 1999*a*). Although Kīlauea is likely to have experienced endogenous growth (e.g. Francis *et al.*, 1993) during the last 200 years, the strong link between magma supply rate and source composition suggests that melting processes are the dominant factor controlling the compositional variation of historical summit lavas.

Despite the rapid historical variations, Kīlauea's overall compositional variation appears to be limited. Lavas from a 1·7 km deep drill core into the volcano's east rift zone show roughly the same compositional variation as historical Kīlauea lavas (Quane *et al.*, 2000). These drill core lavas date back to ~350 ka (Quane *et al.*, 2000). Thus, although the extent of mantle melting is variable (Pietruszka & Garcia, 1999*a*), Kīlauea has shown no long-term trend in compositional variation.

CONCLUSIONS

Historical Kīlauea summit lavas record magmatic processes within the summit magma reservoir during the past 200 years. These lavas reveal well-defined systematic compositional variations that imply the summit magma reservoir is a single, well-mixed body. The restricted range in mineralogy and geochemistry of lavas erupted in and near Halema'uma'u Crater since

1924 suggests that a crown of relatively low-density (2·56–2·58 g/cm^3), aphyric magma overlies more crystal-rich, denser (2·60–2·78 g/cm^3) magma in the main mass of the reservoir. Magma in this crown undergoes more extensive crystal fractionation, especially during periods of lower magma supply, than previously envisioned. The compositions of glasses from some eruptions indicate temperature gradients of at least 40–60°C within the summit reservoir. The effects of crystal fractionation are superimposed on the hybrid magmas produced by mixing of new mantle-derived magma of varying composition with preexisting magma residing within the summit reservoir. Underlying the crown is more olivine-rich magma that is occasionally tapped by eruptions on the caldera margin. Crustal contamination apparently occurred following the 1924 collapse of the summit, although the overall geochemical effects were minor. Variations in parental magma composition for Kīlauea summit lavas correlate with lava eruption rate and indicate significant changes in its source composition and extent of partial melting over the last 200 years. However, the ranges in source composition and extents of partial melting have varied little at Kīlauea for 350 kyr.

ACKNOWLEDGEMENTS

Many people helped make this study possible. Our warmest aloha to Tom Casadevall for sharing unpublished modes for summit prehistoric lavas, Sorena Sorensen for lending critical samples from the Smithsonian collection, Tina Neal for sharing her formerly unpublished Kīlauea summit map, J. Rayray, J. Parker and Kelly Kolysko for sample preparation, Nancy Baker for collecting some of the summit samples, Rachel Konishi, Joann Romano, Kate Bridges, Kristina Garcia and Joe Horrell for help in the field, Pete Dawson for able maintenance of the XRF laboratory at the University of Massachusetts, and Mike Vollinger for unwavering expert XRF analyses. Thanks go to Amy Gaffney, Robert Tilling and Pete Reiners for helpful comments on the manuscript for this paper. This study was supported by an NSF grant (EAR00-01123). This is SOEST contribution 6209.

REFERENCES

Albarède, F. (1993). Residence time analysis of geochemical fluctuations in volcanic series. *Geochimica et Cosmochimica Acta* **57**, 615–621.

Albarède, F. (1996). High-resolution geochemical stratigraphy of Mauna Kea flows from the Hawaii Scientific Drilling Project core. *Journal of Geophysical Research* **101**, 11841–11853.

Albarède, F. & Tamagnan, V. (1988). Modeling the recent geochemical evolution of Piton de la Fournaise volcano, Reunion island, 1931–1986. *Journal of Petrology* **29**, 997–1030.

Barnard, W. M. (1995). Mauna Loa volcano: historical eruptions, exploration, and observations (1779–1910). *Geophysical Monograph, American Geophysical Union* **92**, 1–19.

Bevens, D., Takahashi, T. J. & Wright, T. L. (eds) (1988). *The Early Serial Publications of the Hawaiian Volcano Observatory. Hawaii Natural History Association* **1–3**, 1224 pp.

Blichert-Toft, J., Weis, D., Maerschalk, C. & Albarede, F. (2003). Hawaiian hotspot dynamics as inferred from the Hf and Pb isotopic evolution of Mauna Kea volcano. *Geochemistry, Geophysics, Geosystems* **4**, 2002GC000340.

Bowen, N. L. (1928). *The Evolution of Igneous Rocks*. Princeton, NJ: Princeton University Press, pp. 133–172.

Brigham, W. T. (1909). *The Volcanoes of Kilauea and Mauna Loa on the Island of Hawaii, their Variously Recorded History to the Present Time. Memoirs of the Bernice Pauahi Bishop Museum 2.* Honolulu, HI: Bishop Museum Press. (Reprinted 1974, Kraus Reprint Company, Millwood, NY, 608 pp.)

Casadevall, T. J. & Dzurisin, D. (1987). Stratigraphy and petrology of the Uwekahuna Bluff Section, Kilauea Caldera. *US Geological Survey, Professional Papers* **1350**, 351–375.

Cayol, V., Dieterich, J. H., Okamura, A. T. & Miklius, A. (2000). High magma storage rates before the 1983 eruption of Kilauea, Hawaii. *Science* **288**, 2343–2346.

Chen, C.-Y., Frey, F. A., Rhodes, J. M. & Easton, R. M. (1996). Temporal geochemical evolution of Kilauea Volcano: comparison of Hilina and Puna Basalt. *Geophysical Monograph, American Geophysical Union* **95**, 161–181.

Clague, D. A., Moore, J. G., Dixon, J. E. & Friesen, W. B. (1995). Petrology of submarine lavas from Kilauea's Puna Ridge, Hawaii. *Journal of Petrology* **36**, 299–349.

Clague, D. A., Weber, W. S. & Dixon, J. E. (1991). Picritic glasses from Hawaii. *Nature* **353**, 553–556.

Cooper, K. M., Reid, M. R., Murrell, M. T. & Clague, D. A. (2001). Crystal and magma residence at Kilauea volcano, Hawaii. *Earth and Planetary Science Letters* **184**, 703–718.

Dana, E. S. (1889). Contributions to the petrography of the Sandwich Island. *American Journal of Science* **37**, 441–467.

Decker, R. W. (1987). Dynamics of Hawaiian volcanoes: an overview. *US Geological Survey, Professional Papers* **1350**, 997–1018.

Denlinger, R. P. (1997). A dynamic balance between magma supply and eruption rate at Kilauea volcano, Hawaii. *Journal of Geophysical Research* **102**, 18091–18100.

DePaolo, D. J., Bryce J. G., Dodson, A., Shuster, D. L. & Kennedy, B. M. (2001). Isotopic evolution of Mauna Loa and the chemical structure of the Hawaiian plume. *Geochemistry, Geophysics, Geosystems* **3**, 2000GC000139.

Duffield, W. A., Christiansen, R. L., Koyanagi, R. Y. & Peterson, D. W. (1982). Storage, migration, and eruption of magma at Kilauea Volcano, Hawaii, 1971–1972. *Journal of Volcanology and Geothermal Research* **13**, 273–307.

Dvorak, J. J. & Dzurisin, D. (1993). Variations in magma supply rate at Kilauea Volcano, Hawaii. *Journal of Geophysical Research* **98**, 22255–22268.

Dzurisin, D., Koyanagi, R. Y. & English, T. T. (1984). Magma supply and storage at Kilauea Volcano, Hawaii, 1956–1983. *Journal of Volcanology and Geothermal Research* **21**, 177–206.

Easton, R. M. (1987). Stratigraphy of Kilauea Volcano. *US Geological Survey, Professional Papers* **1350**, 243–260.

Easton, R. M. & Garcia, M. O. (1980). Petrology of the Hilina Formation, Kilauea Volcano, Hawaii. *Bulletin Volcanologique* **43**, 657–673.

Eiler, J. M., Farley, K. A., Valley, J. W., Hofmann, A. W. & Stolper, E. M. (1996). Oxygen isotope constraints on the sources of Hawaiian volcanism. *Earth and Planetary Science Letters* **144**, 453–468.

Finch, R. H. (1940). Engulfment at Kilauea Volcano. In: Fiske, R. S., Simkin, T. & Nielsen, E. A. (eds) *The Volcano Letter, 470*. Washington, DC: Smithsonian Institution Press, pp. 1–4.

Fiske, R. S. & Kinoshita, W. T. (1969). Inflation of Kilauea Volcano prior to its 1967–1968 eruption. *Science* **165**, 341–349.

Fiske, R. S., Simkin, T. & Nielsen, E. A. (eds) (1987). *The Volcano Letter*. Washington, DC: Smithsonian Institution Press, 530 pp.

Francis, P., Oppenheimer, C. & Stevenson, D. (1993). Endogenous growth of persistently active volcanoes. *Nature* **366**, 554–557.

Frey, F. A. & Rhodes, J. M. (1993). Intershield geochemical differences among Hawaiian volcanoes: implications for source compositions, melting process and magma ascent paths. *Philosophical Transactions of the Royal Society of London* **342**, 121–136.

Garcia, M. O. (1996). Petrography and olivine and glass chemistry of lavas from the Hawaii Scientific Drilling Project. *Journal of Geophysical Research* **101**, 11701–11713.

Garcia, M. O., Muenow, D., Aggrey, K. & O'Neil, J. (1989). Major element, volatile and stable isotope geochemistry of Hawaiian submarine tholeiitic glasses. *Journal of Geophysical Research* **94**, 10525–10538.

Garcia, M. O., Rhodes, J. M., Ho, R. A., Ulrich, G. & Wolfe, E. W. (1992). Petrology of lavas from episodes 2–47 of the Pu`u `Ō`ō eruption of Kilauea Volcano, Hawaii: evaluation of magmatic processes. *Bulletin of Volcanology* **55**, 1–16.

Garcia, M. O., Hulsebosch, T. & Rhodes, J. (1995). Olivine-rich submarine basalts from the southwest rift zone of Mauna Loa Volcano: implications for magmatic processes and geochemical evolution. *Geophysical Monograph, American Geophysical Union* **92**, 219–239.

Garcia, M. O., Ito, E., Eiler, J. & Pietruszka, A. (1996*a*). Crustal contamination of lavas from Kilauea Volcano magmas revealed by oxygen isotope analysis of glass and olivine from the Pu`u `Ō`ō eruption lavas. *EOS Transactions, American Geophysical Union* **77**(46), F798.

Garcia, M. O., Rhodes, J. M., Trusdell, F. A. & Pietruszka, A. P. (1996*b*). Petrology of lavas from the Pu`u `Ō`ō eruption of Kilauea Volcano: III. The Kupaianaha episode (1986–1992). *Bulletin of Volcanology* **58**, 359–379.

Garcia, M. O., Ito, E., Eiler, J. M. & Pietruszka, A. J. (1998). Crystal contamination of Kilauea Volcano magmas revealed by oxygen isotope analyses of glass and olivine from Pu`u `Ō`ō eruption lavas. *Journal of Petrology* **39**, 803–817.

Garcia, M. O., Pietruszka, A. J., Rhodes, J. M. & Swanson, K. (2000). Magmatic processes during the prolonged Pu`u `Ō`ō eruption of Kilauea Volcano, Hawaii. *Journal of Petrology* **41**, 967–990.

Ghiorso, M. S. (1995). Thermodynamic models of igneous processes. *Annual Review of Earth and Planetary Sciences* **25**, 221–241.

Ghiorso, M. S. & Sack, R. O. (1995). Chemical mass transfer in magmatic systems IV. A revised and internally consistent thermodynamic model for the interpolation and extrapolation of liquid–solid equilibria in magmatic systems at elevated temperatures and pressures. *Contributions to Mineralogy and Petrology* **119**, 197–212.

Hauri, E. H. (1996). Major-element variability in the Hawaiian mantle plume. *Nature* **382**, 415–419.

Helz, R. T. (1987). Differentiation behavior of Kilauea Iki lava lake, Kilauea Volcano, Hawaii: an overview of past and current work. *Geochemical Society, Special Publication* **1**, 241–258.

Helz, R. T. & Thornber, C. R. (1987). Geothermometry of Kilauea Iki lava lake, Hawaii. *Bulletin of Volcanology* **49**, 651–668.

Hemond, C., Hofmann, A. W., Heusser, G., Condomines, M., Raczek, I. & Rhodes, J. M. (1994). U–Th–Ra systematics in Kilauea and Mauna Loa tholeiites. *Chemical Geology* **116**, 163–180.

Ho, R. & Garcia, M. O. (1988). Origin of differentiated lavas at Kilauea Volcano, Hawaii: implications from the 1955 eruption. *Bulletin of Volcanology* **50**, 35–46.

Hofmann, A. W., Feigenson, M. D. & 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* **88**, 24–35.

Holcomb, R. T. (1987). Eruptive history and long-term behavior of Kilauea Volcano. *US Geological Survey, Professional Papers* **1350**, 261–350.

Ito, E., White, W. M., Gopel, C. (1987). The O, Sr, Nd, and Pb isotopic geochemistry of MORB. *Chemical Geology* **62**, 157–176.

Jaggar, T. A. (1947). *Origin and Evolution of Craters*. *Geological Society of America, Memoirs* **121**, 508 pp.

Jurewicz, A. & Watson, E. B. (1988). Cations in olivine, Part 1, Calcium. *Contributions to Mineralogy and Petrology* **99**, 176–185.

Klein, F. W. (1982). Earthquakes at Loihi submarine volcano and the Hawaiian hotspot. *Journal of Geophysical Research* **87**, 7719–7726.

Klein, F. W., Koyanagi, R. Y., Nakata, J. S. & Tanigawa, W. R. (1987). The seismicity of Kilauea's magma system. *US Geological Survey, Professional Papers* **1350**, 1019–1185.

Kolysko-Rose, K. W. (1999). Fireworks to usher in the New Year: the 31 December 1974 eruption at Kilauea Volcano, Hawaii. B.Sc. thesis, University of Hawaii, 21 pp.

Kurz, M. D., Kenna, T. C., Kammer, D. P., Rhodes, J. M. & Garcia, M. O. (1995). Isotopic evolution of Mauna Loa Volcano: a view from the submarine southwest rift zone. *Geophysical Monograph, American Geophysical Union* **92**, 289–306.

Kyser, T. K., O'Neil, J. R. & Carmichael, I. S. E. (1982). Genetic relations among basic lavas and ultramafic nodules: evidence from oxygen isotope compositions. *Contributions to Mineralogy and Petrology* **81**, 88–102.

Lipman, P. W., Rhodes, J. M. & Dalrymple, G. B. (1990). The Ninole Basalt—implications for the structural evolution of Mauna Loa Volcano, Hawaii. *Bulletin of Volcanology* **53**, 1–19.

Lockwood, R. I., Tilling, R. I., Klein, F., Okamura, A. T. & Peterson, D. W. (1999). *Magma Migration and Resupply during the 1974 Summit Eruptions of Kilauea, Volcano, Hawaii*. *US Geological Survey, Professional Papers* **1613**, 37 pp.

Maaloe, S., Pedersen, R. B. & James, D. (1988). Delayed fractionation of basaltic lavas. *Contributions to Mineralogy and Petrology* **98**, 401–407.

Macdonald, G. A. (1942). Petrography of Maui. In: *Geology and Ground-water Resources of the Island of Maui, Hawaii*. Honolulu, HI: Advertiser Publishing Co., pp. 275–334.

Macdonald, G. A. (1949). *Petrography of the Island of Hawaii*. *US Geological Survey, Professional Papers* **214-D**, 96 pp.

Macdonald, G. A. (1955). *Catologue of Active Volcanoes of the World: Part III Hawaiian Islands*. Naples: International Volcanological Association, 37 pp.

Macdonald, G. A., Abbott, A. T. & Peterson, F. L. (1983). *Volcanoes in the Sea: the Geology of Hawaii*. Honolulu, HI: University of Hawaii Press.

McBirney, A. R. (1993). *Igneous Petrology*. Boston, MA: Jones and Bartlett.

McPhie, J., Walker, G. P. L. & 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* **52**, 334–354.

Miklius, A & Cervelli, P. (2003). Interaction between Kilauea and Mauna Loa. *Nature* **421**, 229.

Moore, J. G. & Ault, W. U. (1965). Historic littoral cones in Hawaii. *Pacific Science* **19**, 3–11.

Nakamura, M. (1995). Residence time and crystallization history of nickeliferous olivine phenocrysts from the northern Yatsugatake volcanoes, Central Japan. *Journal of Volcanology and Geothermal Research* **66**, 81–100.

Neal, C. A. & Lockwood, J. P. (2003). *Geologic Map of the Summit Area of Kilauea Volcano, Hawaii. US Geological Survey Map* **I-2759**.

Nicholls, J. & Stout, M. Z. (1988). Picritic melts in Kilauea—evidence from the 1967–1968 Halemaumau and Hiiaka eruptions. *Journal of Petrology* **29**, 1031–1057.

Niu, Y. & Batiza, R. (1991). DENSCAL: a program for calculating densities of silicate melts and mantle minerals as a function of pressure, temperature, and composition in melting range. *Computers and Geosciences* **17**, 679–687.

Pietruszka, A. P. & Garcia, M. O. (1999*a*). A rapid fluctuation in the mantle source and melting history of Kilauea Volcano inferred from the geochemistry of it historical summit lavas (1790–1982). *Journal of Petrology* **40**, 1321–1342.

Pietruszka, A. J. & Garcia, M. O. (1999*b*). The size and shape of Kilauea Volcano's summit magma storage reservoir: a geochemical probe. *Earth and Planetary Science Letters* **167**, 311–320.

Pietruszka, A. P., Rubin, K. H. & 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* **186**, 15–31.

Powers, H. (1955). Composition and origin of basaltic magma of the Hawaiian Islands. *Geochimica et Cosmochimica Acta* **7**, 77–107.

Quane, S. L., Garcia, M. O., Guillou, H. & Hulsebosch, T. P. (2000). Magmatic history of the east rift zone of Kilauea Volcano, Hawaii based on drill core from SOH 1. *Journal of Volcanology and Geothermal Research* **102**, 319–338.

Reed, S. J. B. (1993). *Electron Microprobe Analysis*. Cambridge: Cambridge University Press, 326 pp.

Reiners, P. W. (2002). Temporal–compositional trends in intraplate basalt eruptions: implications for mantle heterogeneity and melting processes. *Geochemistry, Geophysics, Geosystems* **3**, 10.1029/2001 GC000250.

Rhodes, J. M. (1995). The 1852 and 1868 Mauna Loa Picrite eruptions: clues to parental magma compositions and the magmatic plumbing system. *Geophysical Monograph, American Geophysical Union* **92**, 241–262.

Rhodes, J. M. (1996). Geochemical stratigraphy of lava flows sampled by the Hawaii Scientific Drilling Project. *Journal of Geophysical Research* **101**, 11729–11746.

Rhodes, J. M. & Hart, S. R. (1995). Episodic trace element and isotopic variations in historical Mauna Loa lavas: implications for magma and plume dynamics. *Geophysical Monograph, American Geophysical Union* **92**, 263–286.

Rhodes, J. M., Dungan, M. A., Blanchard, D. P. & Long, P. E. (1979). Magma mixing at mid-ocean ridges: evidence from basalts drilled near 22 N on the Mid-Atlantic Ridge. *Tectonophysics* **55**, 36–61.

Rhodes, J. M., Wenz, K. P., Neal, C. A., Sparks, J. W. & Lockwood, J. P. (1989). Geochemical evidence for invasion of Kilauea's plumbing system by Mauna Loa magmas. *Nature* **337**, 257–260.

Richter, D. H., Eaton, J. P., Murata, K. J., Ault, W. U. & Krivoy, H. L. (1970). *Chronology Narrative of the 1959–1960 Eruption of Kilauea Volcano, Hawaii. US Geological Survey, Professional Papers* **537-E**, 71 pp.

Roeder, P. L. & Emslie, R. F. (1970). Olivine–liquid equilibrium. *Contributions to Mineralogy and Petrology* **29**, 275–289.

Rowan, L. R. & Clayton, R. W. (1993). The three-dimensional structure of Kilauea Volcano, Hawaii, from travel time tomography. *Journal of Geophysical Research* **98**, 4355–4375.

Ryan, M. P. (1988). The mechanics and three-dimensional internal structure of active magmatic systems: Kilauea Volcano, Hawaii. *Journal of Geophysical Research* **93**, 4213–4248.

Sharp, R. P., Dzurisin, D. & Malin, M. C. (1987). An early 19th century reticulite pumice from Kilauea Volcano. *US Geological Survey, Professional Papers* **1350**, 395–404.

Stormer, J. (1973). Calcium zoning in olivine and its relationship to silica activity and pressure. *Geochimica et Cosmochimica Acta* **27**, 1815–1821.

Thirlwall, M. F., Jenkins, C., Vroon, P. Z. & Mattey, D. P. (1997). Crustal interaction during construction of ocean islands. Pb–Sr–Nd–O isotope geochemistry of shield basalts of Gran Canaria, Canary Islands. *Chemical Geology* **135**, 233–262.

Tilling, R. I. & Dvorak, J. J. (1993). Anatomy of a basaltic volcano. *Nature* **363**, 125–133.

Tilling, R. I., Wright, T. L. & Millard, H. T., Jr (1987). Trace-element chemistry of Kilauea and Mauna Loa lava in space and time: a reconnaissance. *US Geological Survey, Professional Papers* **1350**, 641–689.

Ulmer, P. (1989). The dependence of the Fe^{2+}–Mg cation-partitioning between olivine and basaltic liquid on pressure, temperature and composition. *Contributions to Mineralogy and Petrology* **101**, 261–273.

Wallace, P. J. & 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* **59**, 327–344.

Wilkinson, J. F. G. & Hensel, H. D. (1988). The petrology of some picrites from Mauna Loa and Kilauea volcanoes, Hawaii. *Contributions to Mineralogy and Petrology* **98**, 326–345.

Wolfe, E. W., Garcia, M. O., Jackson, D. B., Koyanagi, R. Y., Neal, C. A. & Okamura, A. T. (1987). The Pu'u 'Ō'ō eruption of Kilauea volcano, episodes 1 through 20, January 3, 1983, to June 8, 1984. *US Geological Survey, Professional Papers* **1350**, 471–508.

Wright, T. L. (1971). Chemistry of Kilauea and Mauna Loa in space and time. *US Geological Survey, Professional Papers* **735**, 1–40.

Wright, T. L. & Fiske, R. S. (1971). Origin of differentiated and hybrid lavas of Kilauea Volcano, Hawaii. *Journal of Petrology* **12**, 1–65.

Wright, T. L. & Peck, D. L. (1978). *Crystallization and Differentiation of the Alae Magma, Alae Lava Lake, Hawaii. US Geological Survey, Professional Papers* **935-C**, 20 pp.

Wyss, M. & Koyanagi, R. Y. (1992). *Isoseismal Maps, Macroseismic Epicenters and Estimated Magnitudes of Historic Earthquakes in the Hawaiian Islands. US Geological Survey Bulletin* **2006**, 93 pp.

Yang, H. J., Frey, F. A., Rhodes, J. M. & Garcia, M. O. (1996). Evolution of Mauna Kea volcano: inferences from lava compositions recovered in the Hawaii Scientific Drilling Project. *Journal of Geophysical Research* **101**, 11747–11767.

Yang, X., Davis, P. M., Delaney, P. T. & Okamura, A. T. (1992). Geodetic analysis of dike intrusion and motion of the magma reservoir beneath the summit of Kilauea Volcano, Hawaii: 1970–1985. *Journal of Geophysical Research* **97**, 3305–3324.