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Two magma bodies beneath the summit of Kīlauea Volcano unveiled by isotopically distinct melt deliveries from the mantle

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ABSTRACT

The summit magma storage reservoir of Kīlauea Volcano is one of the most important components of the magmatic plumbing system of this frequently active basaltic shield-building volcano. Here we use new high-precision Pb isotopic analyses of Kīlauea summit lavas—from 1959 to the active Halema'uma'u lava lake—to infer the number, size, and interconnectedness of magma bodies within the volcano's summit reservoir. From 1971 to 1982, the $^{206}Pb/^{204}Pb$ ratios of the lavas define two separate magma mixing trends that correlate with differences in vent location and/or pre-eruptive magma temperature. These relationships, which contrast with a single magma mixing trend for lavas from 1959 to 1968, indicate that Kīlauea summit eruptions since at least 1971 were supplied from two distinct magma bodies. The locations of these magma bodies are inferred to coincide with two major deformation centers identified by geodetic monitoring of the volcano's summit region: (1) the main locus of the summit reservoir ~2–4 km below the southern rim of Kīlauea Caldera and (2) a shallower magma body <2 km below the eastern rim of Halema'uma'u pit crater. Residence time modeling suggests that the total volume of magma within Kīlauea's summit reservoir during the late 20th century (1959–1982) was exceedingly small (~0.1–0.5 km^3). Voluminous Kīlauea eruptions, such as the ongoing, 32-yr old Pu'u 'Ō'ō rift eruption (>4 km^3 of lava erupted), must therefore be sustained by a nearly continuous supply of new melt from the mantle. The model results show that a minimum of four compositionally distinct, mantle-derived magma batches were delivered to the volcano (at least three directly to the summit reservoir) since 1959. These melt inputs correlate with the initiation of energetic (1959 Kīlauea Iki) and/or sustained (1969–1974 Mauna Ulu, 1983-present Pu'u 'Ō'ō and 2008-present Halema'uma'u) eruptions. Thus, Kīlauea's eruptive behavior is partly tied to the delivery of new magma batches from the volcano's source region within the Hawaiian mantle plume.

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1. Introduction

The processing of magma in the crustal reservoirs of basaltic volcanoes blurs the chemical and isotopic signatures of mantle heterogeneity and melt generation (e.g., Maclennan, 2008; Rubin et al., 2009; Vlastélic et al., 2009) and masks the timing of melt delivery from the mantle. Kīlauea Volcano on the Island of Hawai'i—one of the world's most active and best studied basaltic shield-building volcanoes—displays rapid variations in lava chemistry that ultimately result from its mantle source and melting processes (e.g., Greene et al., 2013; Hofmann et al., 1984; Pietruszka and Garcia, 1999a; Pietruszka et al., 2013). Like many other basaltic volcanoes, a shallow magma storage reservoir underlies Kīlauea's summit region (Tilling and Dvorak, 1993), but the geometry (size and shape) of this reservoir (e.g., Poland et al., 2014; Wright and Klein, 2014), and its ability to modulate the mantle-derived fluctuations in the composition of the input magma (e.g., Greene et al., 2013; Pietruszka and Garcia, 1999b; Thornber, 2003; Thornber et al., 2015), is a matter of debate. A full understanding of the geometry of Kīlauea's summit reservoir is key to deciphering the process of melt generation and transport within the heterogeneous Hawaiian mantle plume. This is particularly important for the interpretation of rapid variations in lava chemistry during sustained Kīlauea eruptions, such as the Pu'u 'Ō'ō eruption on the volcano's east rift zone (1983-present). For example, Pu'u 'Ō'ō lavas display short-term fluctuations in their Pb isotope ratios on a time scale of <10 years that may result from

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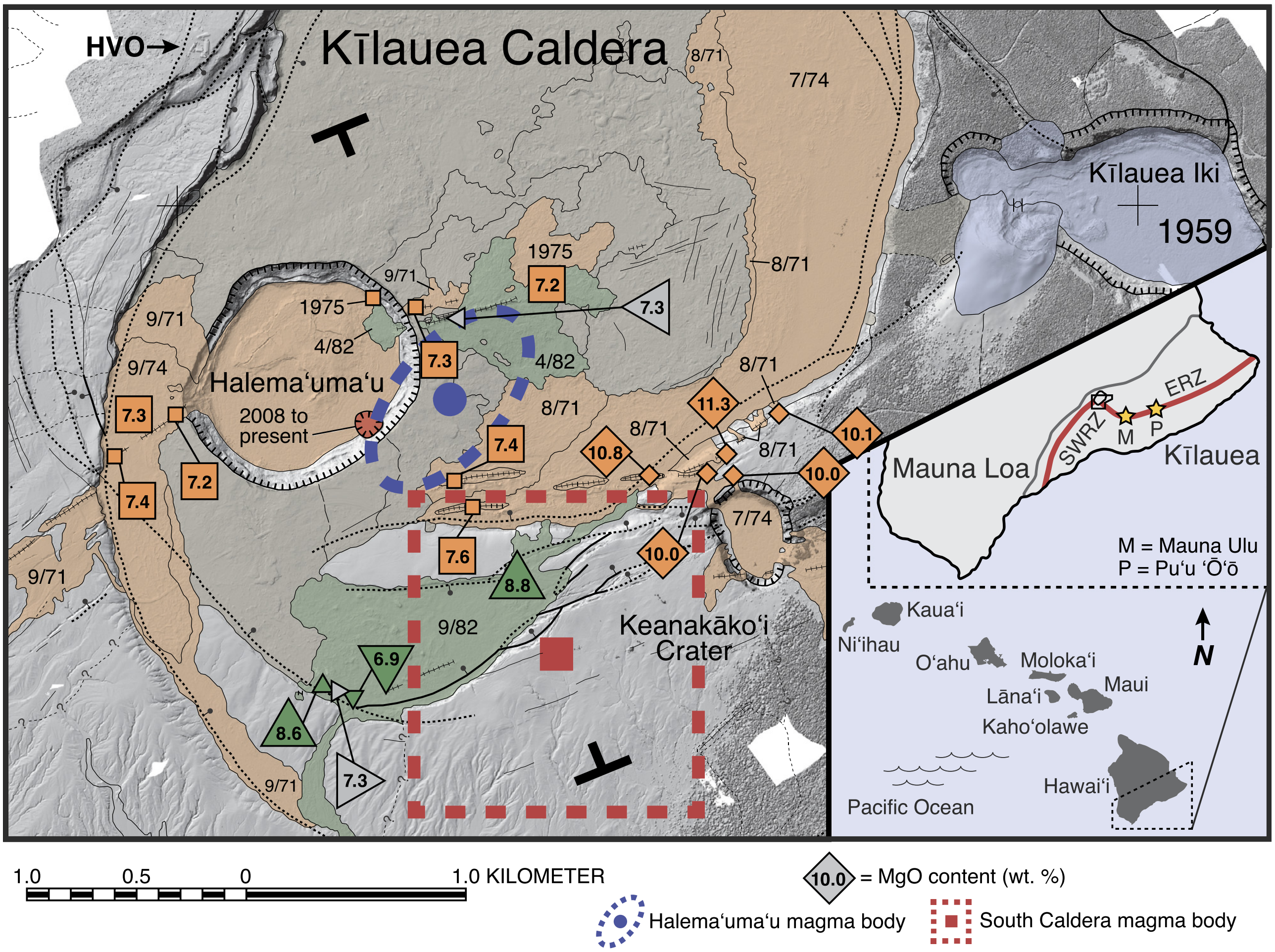

Fig. 1. Map of the summit region of Kīlauea Volcano. The inset map shows the location of Kīlauea Volcano on Island of Hawai‘i, the volcano's summit caldera, and its rift zones to the east (ERZ) and southwest (SWRZ). The main vents of the sustained Mauna Ulu (M) and Pu‘u ‘Ō‘ō (P) eruptions are indicated. Mapped lava flow boundaries, modified from the geologic map of Neal and Lockwood (2003), are superimposed on a Jun. 2009 LiDAR image of Kīlauea's summit region. Thus, the size of the 2008 Halema‘uma‘u eruption vent is shown as of Jun. 2009. As of Jan. 2015, this elliptical vent is ∼160 by 210 m in diameter. The LiDAR survey may be accessed at http://dx.doi.org/10.5069/G9DZ067X. Historical lava flows prior to the 1959 Kīlauea Iki eruption are shown in light brown. Sample locations are shown for the eruptions from 1971 to 1982. The location of each sample signifies its eruption date (except for the 1975 sample from the caldera floor, which now underlies the Apr. 1982 lava). The symbols on the map with MgO contents (in wt%) are keyed to eruption date and/or vent location as shown in Fig. 2. Complete locations for all of the samples and their MgO contents, including literature data sources, are provided in Appendix A. The red square indicates the main locus of Kīlauea's summit reservoir—the "South Caldera" (SC) magma body—based on the observed range (dashed box) of deformation centers. The Halema‘uma‘u magma body is located near the blue circle (dashed oval outlines the range of deformation centers). The "T"s indicate the orientation of the cross section in Fig. 5. The position of the Hawaiian Volcano Observatory (HVO) is shown for reference. The crosses in the center of Kīlauea Iki crater and at the western margin of the caldera mark 19°25′N latitude, and 155°15′W and 155°17′30″W longitude, respectively. (For interpretation of the references to color in this figure caption, the reader is referred to the online version of this article.)

the melting of a fine-scale repeating pattern of filamentous heterogeneities in the mantle source region (Greene et al., 2013). In this view, the mantle-derived changes in $^{206}Pb/^{204}Pb$ are efficiently transmitted from the source to the surface because the parental magmas for the Pu‘u ‘Ō‘ō eruption partially bypass the volcano's summit reservoir, and avoid its buffering effects (e.g., Garcia et al., 1996). In contrast, Thornber (2003) observed temporal fluctuations in the K_2O/TiO_2 ratios of Pu‘u ‘Ō‘ō lavas (superimposed on a long-term decreasing trend) that partially correlate with the deformation of the volcano's summit region. These signatures were interpreted to result from mixing between a compositionally homogeneous Pu‘u ‘Ō‘ō-type mantle-derived input magma (with a relatively low K_2O/TiO_2 ratio) and variably decreasing amounts of stored 1982-era magma (with a relatively high K_2O/TiO_2 ratio) from the volcano's summit reservoir (Thornber, 2003; Thornber et al., 2015).

The basic model for the magmatic plumbing system of Kīlauea Volcano is based on decades of geophysical (e.g., Eaton and Murata, 1960; Poland et al., 2014) and petrologic (e.g., Garcia et al., 2003; Thornber et al., 2003; Wright, 1971; Wright and Fiske, 1971) study. A primary conduit delivers compositionally distinct batches (e.g., Wright and Fiske, 1971; Wright and Tilling, 1980; Wright and Klein, 2014) of primitive mantle-derived melt from a depth of >60 km to a shallow reservoir beneath the volcano's summit, where magma is mixed and stored prior to eruption at the summit or lateral intrusion into the volcano's two rift zones for further storage, mixing, differentiation, and/or eruption. Geophysical studies (e.g., Eaton and Murata, 1960; Fiske and Kinoshita, 1969; Klein et al., 1987; Yang et al., 1992) have long identified the main locus of Kīlauea's summit reservoir ∼2 km southeast of Halema‘uma‘u pit crater (near the southern rim of Kīlauea Caldera; Fig. 1) at a depth of ∼2–4 km. More recent investigations of ground deformation (Anderson et al., 2015; Baker and Amelung, 2012; Cervelli and Miklius, 2003), residual gravity (Johnson et al., 2010), and long-period seismicity (Battaglia et al., 2003) have discovered a shallower (<2 km deep) magma body beneath the eastern rim of Halema‘uma‘u that dates back to at least 1975 (Fig. 1). Thus, Kīlauea's summit reservoir—originally viewed as a plexus of magma-filled dikes and sills (e.g., Fiske and Kinoshita, 1969)—is now thought to comprise two distinct magma bodies (e.g., Poland et al., 2014). Estimates for the volume of each body (based mostly on geophysics) vary widely from ∼0.1–6 km^3 for the shallower Halemaʻumaʻu magma body to ∼13–27 km^3 for the deeper "South Caldera" (SC) magma body (Anderson et al., 2015; Dawson et al., 1999; Johnson, 1992; Poland et al., 2009; Segall et al., 2001), as reinterpreted by Anderson et al. (2015) and Poland et al. (2014). The maximum possible volume of the summit reservoir (based on the size of the aseismic region beneath the volcano's summit; Klein et al., 1987) is ∼40 km^3.

The presence of two magma bodies beneath Kīlauea's summit has not been confirmed with lava chemistry. Modeling of systematic temporal variations in ratios of incompatible elements (e.g., Nb/Y or La/Yb) for Kīlauea lavas erupted over the last 200 years suggested that the volcano's summit reservoir was a single, "spherical" body holding ∼2–3 km^3 of magma (Pietruszka and Garcia, 1999b). In contrast, two stages of degassing beneath Kīlauea's summit—one presumably within the deeper SC magma body and the other marked by shallow equilibration depths (<2 km) that may correspond to the Halemaʻumaʻu magma body—have been proposed based on (1) unusually CO_2-poor gas emissions near the rim of Halemaʻumaʻu in 2009, and (2) low pressures of H_2O–CO_2 equilibrium for olivine-hosted melt inclusions from summit tephra erupted in 2008 and 2010 (Edmonds et al., 2013). However, this evidence is inconclusive because shallow degassing of magma may occur within the conduit that links the active Halemaʻumaʻu lava lake to the summit reservoir (without stalling in a second magma body beneath Halemaʻumaʻu).

A time-series analysis of the high-precision Pb isotopic variations of Kīlauea summit lavas has the potential to (1) refine the primarily geophysics-based view of the geometry of the volcano's summit reservoir and (2) clarify the way it processes the melt delivered from the mantle. Kīlauea summit lavas erupted over the last millennium—most of which are thought to be supplied directly from the summit reservoir (Pietruszka and Garcia, 1999b)—display fluctuations in $^{206}Pb/^{204}Pb$ on time scales of years to centuries (Marske et al., 2007; Pietruszka and Garcia, 1999a). These variations ultimately result from changes in the composition of the parental magma delivered to the volcano from its source within the Hawaiian mantle plume, which is thought to be heterogeneous (e.g., Abouchami et al., 2005; Tatsumoto, 1978; Weis et al., 2011) on a small scale (Greene et al., 2013; Marske et al., 2007, 2008; Pietruszka and Garcia, 1999a; Pietruszka et al., 2013). However, the mantle-derived fluctuations in the Pb isotope ratios of the erupted lavas are likely modulated by shallow magma mixing, depending on the number, size, and interconnectedness of magma bodies beneath the volcano's summit, and the residence time of magma within them (Albarède, 1993; Pietruszka and Garcia, 1999b). In this study, we present new high-precision Pb isotopic analyses of Kīlauea summit lavas—from 1959 to the active Halemaʻumaʻu lava lake—and model the variations in their $^{206}Pb/^{204}Pb$ ratios using a time-dependent mass-balance equation to (1) estimate the residence time and volume of magma within the volcano's summit reservoir and (2) infer the timing of compositionally distinct melt deliveries (i.e., magma batches) from the mantle.

2. Sampling and analytical methods

Thirty-eight Kīlauea lava or juvenile tephra samples were analyzed for high-precision Pb isotope ratios from rock or glass chips using thallium-doping on a Nu Plasma HR or Nu Plasma 1700 multiple-collector inductively coupled plasma mass spectrometer (MC-ICPMS) at San Diego State University. The analyzed samples range in age from the 1959 eruption in Kīlauea Iki Crater to the ongoing 2008 Halemaʻumaʻu eruption (2009–2010), and include recent samples from the ongoing Puʻu ʻŌʻō eruption (2008–2012) for comparison with the latter. The Pb isotopic results and the analytical metadata related to this study, including results for two in-house Hawaiian lava standards (Kil1919, a 1919 summit lava, and Menehune, a quenched Puʻu ʻŌʻō lava from 2006), are reported

Table 1
Pb isotope ratios of historical lavas from Kīlauea Volcano.

Sample	Eruption date	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
1959 Kīlauea Iki eruption				
Iki-58 (S-1)	14-Nov-59	18.5949	15.4839	38.2081
Iki-01 (S-2)	14-Nov-59	18.5866	15.4844	38.2105
Iki-32 (S-22)	17-Dec-59	18.5954	15.4858	38.2094
Iki-32 (S-22)[a]	17-Dec-59	18.5957	15.4858	38.2094
1961 Halemaʻumaʻu eruption				
24-Feb-61	24-Feb-61	18.5623	15.4821	38.1835
K61-03	10-Jul-61	18.5561	15.4816	38.1787
1967–1968 Halemaʻumaʻu eruption				
HM68-12	Feb. 1968	18.5026	15.4777	38.1343
HM68-15	13-Jul-68	18.5075	15.4798	38.1431
August 1971 eruption				
KIL 71-3	14-Aug-71	18.4923	15.4777	38.1268
KIL 71-5	14-Aug-71	18.4913	15.4768	38.1239
KIL 71-6	14-Aug-71	18.5081	15.4775	38.1369
September 1971 eruption				
KL-971-1	24-Sep-71	18.5085	15.4785	38.1393
KL-971-9	24- to 25-Sep-71	18.5069	15.4775	38.1362
July 1974 eruption				
KIL-774-1	19-Jul-74	18.4642	15.4748	38.1033
KIL-774-2	19-Jul-74	18.4655	15.4768	38.1088
KIL-774-3	20-Jul-74	18.4680	15.4773	38.1128
KIL-774-6A	20-Jul-74	18.4929	15.4760	38.1231
September 1974 eruption				
KIL-974-8	19-Sep-74	18.4928	15.4764	38.1250
KIL-375-14A	19-Sep-74	18.4952	15.4787	38.1311
November 1975 eruption				
KIL75-1	29-Nov-75	18.4904	15.4791	38.1306
KIL75-2	29-Nov-75	18.4895	15.4776	38.1240
April 1982 eruption				
1982A-20	30-Apr-82	18.4690	15.4753	38.1076

Table 1 (*continued*)

Sample	Eruption date	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$
September 1982 eruption				
1982S-3	25-Sep-82	18.4553	15.4750	38.0988
1982S-9	26-Sep-82	18.4557	15.4767	38.1038
1982S-12	26-Sep-82	18.4678	15.4755	38.1071
1982S-14	26-Sep-82	18.4752	15.4787	38.1192
1982S-14[a]	26-Sep-82	18.4745	15.4776	38.1159
December 1974 eccentric summit eruption				
SWR-1274-5	31-Dec-74	18.4707	15.4746	38.1088
SWR-375-13	31-Dec-74	18.4771	15.4759	38.1155
SWR-175-12	31-Dec-74	18.4702	15.4770	38.1142
Puʻu ʻŌʻō eruption (1983 to present)				
PO5-Sep-08	5-Sep-08	18.3969	15.4696	38.0562
PO11-Jun-10[b]	11-Jun-10	18.3964	15.4692	38.0533
PO25-Jul-10	25-Jul-10	18.3970	15.4696	38.0554
PO25-Jul-10[a]	25-Jul-10	18.3977	15.4706	38.0584
PO6-Oct-10[b]	6-Oct-10	18.3968	15.4699	38.0562
PO23-Jun-11	23-Jun-11	18.3973	15.4704	38.0579
PO24-Feb-12	24-Feb-12	18.3974	15.4695	38.0541
Halemaʻumaʻu eruption (2008 to present)				
KS09-152H	17-Sep-09	18.3956	15.4716	38.0602
KS10-158A	13- to 15-Jan-10	18.3911	15.4686	38.0501
KS10-158A[a]	13- to 15-Jan-10	18.3915	15.4688	38.0500
KS10-185A	10- to 21-Jun-10	18.3878	15.4653	38.0398
KS10-185A[a]	10- to 21-Jun-10	18.3889	15.4664	38.0429
KS10-191A	23-Jul to 13-Aug-10	18.3872	15.4643	38.0367
In-house standards from Kīlauea Volcano				
Kil1919 ($n = 33$)	1919	18.6552	15.4897	38.2068
$\pm 2\sigma$		0.0027	0.0020	0.0057
$\pm 2\sigma_m$		0.0005	0.0004	0.0010
Kil1919 ($n = 12$)[c]	1919	18.6557	15.4896	38.2069
$\pm 2\sigma$		0.0031	0.0012	0.0041
$\pm 2\sigma_m$		0.0009	0.0003	0.0012
Menehune ($n = 68$)	24-Jun-06	18.4073	15.4714	38.0627
$\pm 2\sigma$		0.0016	0.0018	0.0060

Two in-house standards were analyzed as unknowns during this study: (1) 21 dissolutions of a powdered rock standard (Kil1919) collected from the same 1919 lava flow of Kīlauea Volcano as the BHVO-1 and -2 international rock standards and (2) 37 dissolutions of a new volcanic glass standard (called "Menehune") that was quenched from a Puʻu ʻŌʻō lava flow of Kīlauea Volcano on June 24, 2006. The Menehune standard was analyzed as ~1- to 2-mm sized handpicked glass fragments. The details of the analytical methods were previously described by Marske et al. (2007). Briefly, the Pb isotope ratios were measured using Tl doping (SRM997) to correct for instrumental mass bias on a Nu Plasma HR or Nu Plasma 1700 multiple-collector inductively coupled plasma mass spectrometer (MC-ICPMS) at San Diego State University. Analyses of SRM981 ($n = 312$) as a bracketing standard over the course of this study gave average bias-corrected values ($\pm 2\sigma$) of $^{206}Pb/^{204}Pb = 16.9448 \pm 52$, $^{207}Pb/^{204}Pb = 15.5027 \pm 58$, and $^{208}Pb/^{204}Pb = 36.7349 \pm 198$. The bias-corrected Pb isotope ratios of the samples and in-house standards were normalized based on the daily bias-corrected average for SRM981 relative to the values of Galer and Abouchami (1998): $^{206}Pb/^{204}Pb = 16.9405$, $^{207}Pb/^{204}Pb = 15.4963$, and $^{208}Pb/^{204}Pb = 36.7219$. The reproducibility of the $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios for Kil1919 and Menehune is similar, but the reproducibility of the $^{206}Pb/^{204}Pb$ ratio for Menehune is a factor of ~2 better than Kil1919, suggesting that the Kil1919 standard is slightly heterogeneous. Thus, the Menehune standard provides the best estimate of the $\pm 2\sigma$ reproducibility of the $^{206}Pb/^{204}Pb$ (± 0.0016), $^{207}Pb/^{204}Pb$ (± 0.0018), and $^{208}Pb/^{204}Pb$ (± 0.0060) ratios of the samples. Average in-run $\pm 2\sigma_m$ errors of individual analyses were ± 0.0004 for $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ and ± 0.0011 for $^{208}Pb/^{204}Pb$, and the maximum values were ± 0.0007 for $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ and ± 0.0019 for $^{208}Pb/^{204}Pb$. The total procedural blanks for Pb were <100 pg during this study, which is insignificant compared to the mass of each sample and its Pb concentration.

[a] Replicate analysis from the same dissolution.

[b] U.S. Geological Survey sample names are KE58-2894F (PO11-Jun-10) and KE58-2914F (PO6-Oct-10).

[c] These average MC-ICPMS Pb isotope ratios for Kil1919 from a previous study in the same laboratory by Marske et al. (2007) are identical to the results from this study within $\pm 2\sigma_m$.

in Table 1. Vent locations, methods of collection, and the remaining details of the analytical methods for all of the samples are provided in Appendix A.

3. Lead isotopic variations of Kīlauea summit lavas

Kīlauea summit lavas erupted from 1959 to 2010 display an overall temporal trend of decreasing $^{206}Pb/^{204}Pb$ ratios (Fig. 2a). Between the two most recent summit eruptions in Sep. 1982 and 2008, Puʻu ʻŌʻō rift zone lavas are characterized by short-term fluctuations in $^{206}Pb/^{204}Pb$ (Fig. 2a). Tephra that erupted from Halemaʻumaʻu in 2009 has a $^{206}Pb/^{204}Pb$ ratio similar to contemporaneous Puʻu ʻŌʻō lavas (Fig. 2b), but subsequent tephra from the Halemaʻumaʻu eruption in 2010 trend towards slightly lower $^{206}Pb/^{204}Pb$ ratios (Fig. 2b) and lower $^{208}Pb/^{204}Pb$ at a given $^{206}Pb/^{204}Pb$ than Puʻu ʻŌʻō lavas from 2008–2012 (Fig. 3a). All of the other samples plot along narrow trends on Pb–Pb diagrams (Fig. 3b), which indicates that the Pb isotope ratios of Kīlauea lavas erupted since 1959 have not been significantly affected by crustal contamination with the underlying volcanic edifice or the Pacific oceanic crust. Indeed, assimilation of such isotopically diverse materials would tend to destroy the systematic temporal variations in $^{206}Pb/^{204}Pb$ (Fig. 2a).

In detail, the $^{206}Pb/^{204}Pb$ ratios of Kīlauea summit lavas decreased from the 1959 Kīlauea Iki eruption through a series of Halemaʻumaʻu eruptions in 1961 and 1967–1968 (Fig. 2c). This temporal trend in Pb isotope ratios can be modeled by the progressive mixing of two compositionally distinct magmas within a single body (Fig. 2c). Conversely, lavas from the Aug. 1971 and Jul. 1974 summit eruptions display significant differences in $^{206}Pb/^{204}Pb$ (Fig. 2d) and whole-rock MgO contents (Fig. 4) that correlate with vent location (Fig. 1). These eruptions occurred from fissures that extended from the caldera floor (southeast of Halemaʻumaʻu) toward Keanakākaoi Crater outside the caldera (and beyond for the Jul. 1974 eruption). For each eruption, the easternmost lavas have significantly lower $^{206}Pb/^{204}Pb$ ratios (Fig. 2d) and higher MgO contents than the westernmost lavas (10.0–11.3 vs. 7.4–7.6 wt%). Other lavas that erupted near or within Halemaʻumaʻu in Sep. 1971, Sep. 1974, and 1975 have Pb isotope ratios (Fig. 2d) and MgO contents (7.2–7.4 wt%) similar to the con-

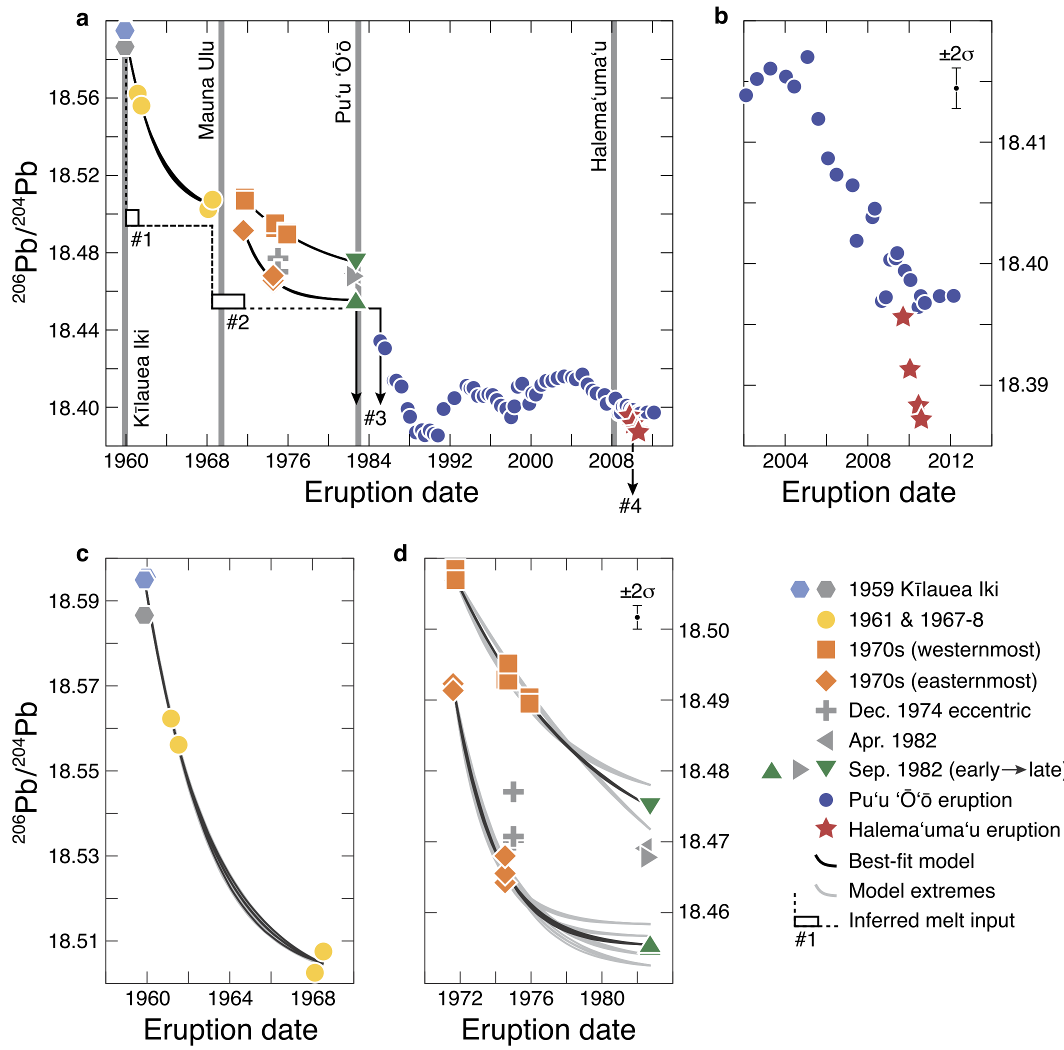

Fig. 2. Variations in the $^{206}Pb/^{204}Pb$ ratios of Kīlauea lavas from 1959 to 2012 with model results. (a) Temporal variations in the $^{206}Pb/^{204}Pb$ ratios of lavas from Kīlauea's summit region (1959–1982 and 2009–2010) and the Pu'u 'Ō'ō eruption (1985–2012) on the volcano's east rift zone (the latter includes literature data; Greene et al., 2013; Marske et al., 2008). The vertical shaded lines mark the initiation of energetic (1959 Kīlauea Iki) or sustained (1969–1974 Mauna Ulu, 1983-present Pu'u 'Ō'ō, and 2008-present Halema'uma'u) eruptions. The composition and timing of mantle-derived melt inputs to Kīlauea Volcano are indicated as white boxes or arrows labeled #1-4. The width of the box or arrows indicates the maximum range in the initial time of melt input, and the height of the box represents the maximum range in its $^{206}Pb/^{204}Pb$ ratio (inferred from the model results). In some cases, one (early Pu'u 'Ō'ō eruption, #3) or both (2008 Halema'uma'u eruption, #4) parameters cannot be estimated from the available data. (b) Close-up of the samples from the 2008 Halema'uma'u and Pu'u 'Ō'ō eruptions. (c) Close-up of the single mixing trend for summit lavas from 1959 to 1968. (d) Close-up of the two mixing trends for summit lavas from 1971 to 1982. Mixing curves were calculated using only the samples shown by colored symbols. Samples shown by gray symbols are thought to have experienced magma mixing outside Kīlauea's summit reservoir (Dec. 1974 and possibly one sample from 1959) or between the two magma bodies that make up the summit reservoir (Apr. 1982 and one sample from Sep. 1982), and thus, were excluded from the model. Best-fit mixing curves are shown in black. The widest permissible range of model parameters were used to calculate the gray mixing curves shown in (c) and (d). A $\pm 2\sigma$ error bar is shown on each plot unless it is smaller than the size of the symbols. (For interpretation of the references to color in this figure caption, the reader is referred to the online version of this article.)

temporaneous westernmost lavas from Aug. 1971 and Jul. 1974 (Fig. 1). Lavas from the Sep. 1982 eruption also display significant variations in Pb isotope ratios that correlate with MgO contents (Fig. 4). Unlike the Aug. 1971 and Jul. 1974 eruptions, the changes during the Sep. 1982 eruption were a function of time, rather than vent location. The earliest lavas from Sep. 1982 had relatively high MgO contents (8.6–8.8 wt%) and low $^{206}Pb/^{204}Pb$ ratios. During this <1-day eruption, the MgO contents of the lavas decreased by ~2 wt% and the $^{206}Pb/^{204}Pb$ ratios of the lavas increased.

The $^{206}Pb/^{204}Pb$ ratios of most Kīlauea summit lavas erupted from 1971 to 1982 define two separate decreasing temporal trends (Fig. 2d). The westernmost lavas from Aug. 1971 and Jul. 1974, the lavas erupted near Halema'uma'u in Sep. 1971, Sep. 1974, and 1975, and the lower MgO, late lavas from Sep. 1982 have higher $^{206}Pb/^{204}Pb$ ratios at a given time (hereafter called the "upper trend") compared to the easternmost lavas from Aug. 1971 and Jul. 1974, and the higher MgO, early lavas from Sep. 1982 (hereafter called the "lower trend"). Each of these trends can be modeled by the progressive mixing of two compositionally distinct magmas within single (but separate) bodies (Fig. 2d). Next, a time-dependent mass-balance mixing model is used to quantitatively determine the residence time and volume of magma within Kīlauea's summit reservoir during the late 20th century for (1) the single Pb isotopic trend from 1959 to 1968, (2) the "upper" trend from 1971 to 1982 with relatively high $^{206}Pb/^{204}Pb$ ratios, and (3) the "lower" trend from 1971 to 1982 with relatively low $^{206}Pb/^{204}Pb$ ratios.

4. The residence time model and magma volumes

4.1. Description of the model

We modified the residence time model of Albarède (1993) for use with isotope ratios as described in Appendix A. This model

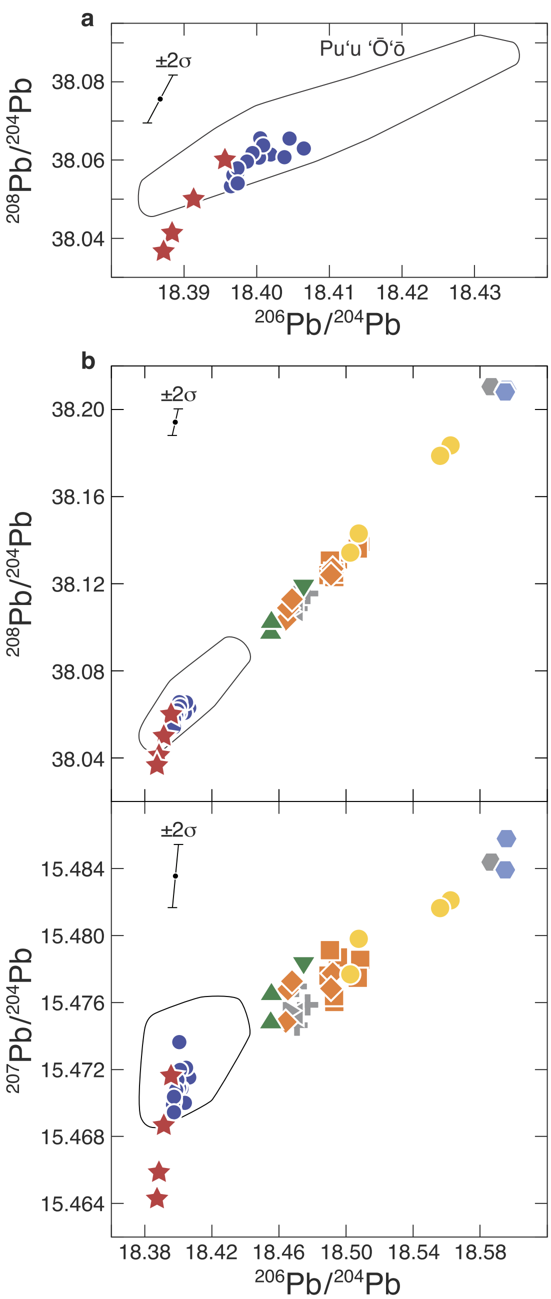

Fig. 3. Plots of $^{206}Pb/^{204}Pb$ vs. $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ for Kīlauea lavas from 1959 to 2012. The Pb isotope ratios of lavas from Kīlauea's summit region (1959–1982 and 2009–2010) and the Pu'u 'Ō'ō eruption (1985–2012) on the volcano's east rift zone (the latter includes literature data; Greene et al., 2013; Marske et al., 2008) are shown. Symbols for Pu'u 'Ō'ō lavas are shown only for samples that erupted within two years of the analyzed samples from the 2008 Halema'uma'u eruption. The field represents the range of older lavas from the Pu'u 'Ō'ō eruption (1985–2006). A $\pm 2\sigma$ error bar is shown on each plot. The other symbols are the same as in Fig. 2. (a) Close-up of lavas from the 2008 Halema'uma'u and Pu'u 'Ō'ō eruptions. (b) Overview of all lavas. (For interpretation of the references to color in this figure caption, the reader is referred to the online version of this article.)

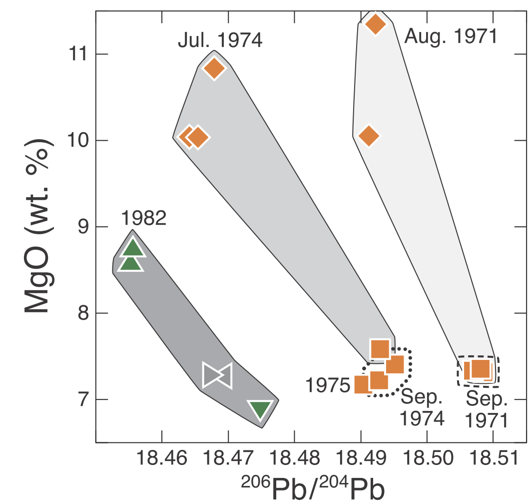

Fig. 4. Relationship between the MgO content and $^{206}Pb/^{204}Pb$ ratios of Kīlauea summit lavas from 1971 to 1982. The fields surround samples from individual eruptions. The other symbols are the same as in Fig. 2. The $\pm 2\sigma$ error bar is smaller than the size of the symbols. The literature data sources for the MgO contents of the lavas are provided in Appendix A. (For interpretation of the references to color in this figure caption, the reader is referred to the online version of this article.)

assumes that the temporal change in the mass of magma in a chamber (M) is controlled by the balance between (1) a constant rate of magma input (Q_{in}) with a constant Pb concentration (C_{in}) and $^{206}Pb/^{204}Pb$ ratio (R_{in}) from the mantle or a deeper chamber, (2) a constant rate of magma output (Q_{out}) to eruptions at the volcano's summit and/or withdrawals to the rift zones, and (3) a constant rate of crystallization (X). If the magma in the chamber is well mixed—a good assumption for Kīlauea (Gerlach et al., 2002)—the final expression used to model the residence time of magma in the chamber is

$$R(t) = \frac{R_0 \frac{C_0}{C_{in}} e^{-\alpha t/\tau} + \frac{R_{in}}{\alpha}(1 - e^{-\alpha t/\tau})}{\frac{C_0}{C_{in}} e^{-\alpha t/\tau} + \frac{1}{\alpha}(1 - e^{-\alpha t/\tau})}, \tag{1}$$

where $R(t)$ is the $^{206}Pb/^{204}Pb$ ratio of the erupted lava at a given time (t), R_0 is the initial $^{206}Pb/^{204}Pb$ ratio of the chamber magma, C_0 is the initial Pb concentration of the chamber magma, and τ is the magma residence time. The effects of crystal fractionation are described by $\alpha = 1 + [f_x \times (D^{Pb} - 1)]$, where D^{Pb} is the bulk partition coefficient of Pb and f_x is the fraction of the input magma flux that is lost to crystallization ($f_x = X/Q_{in}$). Lead is expected to be highly incompatible in olivine and the magma in Kīlauea's summit reservoir is expected to be dominantly olivine controlled, particularly since the mid-20th century (Garcia et al., 2003; Wright, 1971). In this case, $D^{Pb} = 0$ and $\alpha = 1 - f_x$. In order to calculate the volume of magma in the chamber from the definition of magma residence time ($\tau = M/Q_{in}$), an estimate of the magma supply rate (Q_{in}) is required. For simplicity, the densities of the input and chamber magmas are assumed to be equal. In this case, the magma residence time can be expressed in terms of magma volume (V) and a volumetric supply rate (Q^v_{in}). Thus, the final equation used to calculate the volume of magma in the chamber from the magma residence time is $V = \tau \times Q^v_{in}$.

A detailed explanation of the model assumptions, derivation, and results (including sensitivity tests) is provided in Appendix A. Briefly, the modeling was performed to best match the $^{206}Pb/^{204}Pb$ ratios of the lavas for each of the three mixing trends (excluding the lavas used to define the R_0 values) by iteratively varying the residence time (τ) and $^{206}Pb/^{204}Pb$ ratio of the input magma (R_{in}) for a wide range of C_0/C_{in} and α values. This approach is thought to provide estimates of the reasonable extremes in the τ and R_{in} values.

4.2. The 1959–1968 period

The 1959 Kīlauea Iki eruption (Fig. 1) was characterized by a gas- and olivine-rich magma that erupted as unusually hot (up to 1192 °C), energetic lava fountains up to 580 m high. The eruption is thought to have been triggered by a relatively primitive gas-rich melt with ∼10 wt% MgO that rose to a shallow depth and mixed with a cooler, more differentiated stored magma shortly before the eruption began (e.g., Anderson and Brown, 1993; Helz, 1987). This zone of shallow magma storage and mixing was most likely located beneath the caldera along the fissures that stretch from the eastern rim of Halemaʻumaʻu towards Kīlauea Iki (Wright and Klein, 2014), possibly coincident with the site of the present-day Halemaʻumaʻu magma body (Poland et al., 2014). The more primitive magma is thought to have bypassed the deeper portions of volcano's summit reservoir on its way to the surface (Anderson and Brown, 1993; Helz, 1987; Wright and Klein, 2014). This interpretation was based partly on the unique features of the eruption (noted above), but mostly on early observations (Wright, 1973) of the distinctive chemistry of the most primitive 1959 lava (especially its high CaO at a given MgO content), represented by sample S-1 (Helz, 1987). However, high MgO-normalized CaO contents along with elevated ratios of incompatible elements, such as Nb/Y (Pietruszka and Garcia, 1999a), are now known to be characteristics of summit lavas erupted from 1929 to 1959 (Garcia et al., 2003). Furthermore, the Pb isotope ratios of sample S-1 are similar to those of (1) sample S-2 (Table 1), which is thought to represent the cooler, more differentiated magma that was stored for several years prior to the start of the 1959 eruption (Helz, 1987) and (2) summit lavas erupted during the previous 30 years (Pietruszka and Garcia, 1999a). These compositional similarities indicate that the more primitive end-member magma of the 1959 eruption was likely supplied directly from Kīlauea's summit reservoir (rather than bypassing it). Thus, the $^{206}Pb/^{204}Pb$ ratio of sample S-1 (and sample S-22, which is nearly identical) probably records the composition magma from the summit reservoir in 1959. In this case, it is appropriate to model the temporal trend in Pb isotope ratios from 1959 to 1968 by the progressive mixing of two compositionally distinct magmas within a single body (Fig. 2c).

The location of magma mixing for the 1959–1968 period is poorly constrained, but there are three possibilities: (1) the Halemaʻumaʻu magma body alone (if it existed in 1959), (2) the SC magma body alone, or (3) both the Halemaʻumaʻu and SC magma bodies. Wright and Klein (2014) argue that a shallow magma body beneath the east rim of Halemaʻumaʻu during the period of sustained lava lake activity at Kīlauea's summit from the 19th to early 20th centuries was likely destroyed in the 1924 collapse of Halemaʻumaʻu lava lake and subsequent phreatic explosions. Modern geophysical studies definitively trace the presence of the Halemaʻumaʻu magma body back only to 1975 (Johnson et al., 2010). However, Kīlauea's summit reservoir had probably fully recovered from the events of 1924 by the time of the 1959 Kīlauea Iki eruption due to a post-1924 increase in the rate of magma supply from the mantle (Wright and Klein, 2014). This magmatic recovery, along with the hint that a deformation center may have existed beneath the east rim of Halemaʻumaʻu since at least the mid-1960s (Fiske and Kinoshita, 1969; Poland et al., 2014), suggests that the Halemaʻumaʻu magma body was present in 1959. Indeed, the 1961 and 1967–1968 eruptions probably tapped the shallow Halemaʻumaʻu magma body because their eruptive vents were located within Halemaʻumaʻu (which may exclude case #2 above). Thus, we conclude that the magma mixing recorded by the temporal trend in $^{206}Pb/^{204}Pb$ from 1959 to 1968 occurred either (1) within the shallow Halemaʻumaʻu magma body alone or (2) more likely, given the MgO-rich, high-temperature nature of the more primitive end-member melt erupted in 1959, within both the Halemaʻumaʻu magma body and the deeper (and presumably hotter) SC magma body. In the latter scenario, the two magma bodies must have been well connected and thoroughly mixed during or soon after the 1959 eruption, and later, on a time scale of the repose period between the 1959, 1961, and 1967–1968 eruptions.

The Pb isotopic composition of the parental magma delivered to Kīlauea's summit reservoir must have changed between 1959 and 1961 in order to account for the lower $^{206}Pb/^{204}Pb$ ratio of the 1961 lava and the temporal trend of decreasing $^{206}Pb/^{204}Pb$ ratios from 1959 to 1968. This melt delivery from the mantle may have been associated with a series of unusually deep (∼40–60 km) earthquake swarms (centered north of Kīlauea Caldera) from 1953 to 1960 (Wright and Klein, 2014), and a period of summit inflation that lasted for several months (Eaton and Murata, 1960) and culminated with the start of the 1959 eruption (Helz, 1987). Based on the compositional similarities between the 1959 Kīlauea Iki lavas and the previous 30 years of summit lavas (discussed above), it is unlikely that this new mantle-derived magma actually erupted in 1959. Instead, it may have forced resident magma out of the summit reservoir (accumulating olivine antecrysts along the rarely used path to the vent near Kīlauea Iki; Helz, 1987) and charged the magmatic plumbing system with the CO_2 that led to the spectacular lava fountains. A narrow maximum $^{206}Pb/^{204}Pb$ range of 18.494–18.502 for this mantle-derived input magma (#1 in Fig. 2a) is permitted by the mixing model, which indicates that 1982-type end-member magma with a relatively low $^{206}Pb/^{204}Pb$ ratio (∼18.456, based on the lower-trend Sep. 1982 lavas) was not present within the summit reservoir until after 1968. The mixing model for the 1959–1968 period suggests a best-fit magma residence time of 2.4–3.1 yr (with a maximum range of 2.3–3.2 yr), which corresponds to a magma volume of only ∼0.1–0.3 km^3 (0.1–0.4 km^3 maximum range) assuming a range of magma supply rates to the volcano's summit reservoir from an average of 0.06 km^3/yr for the 1959–1990 period to a maximum of 0.11 km^3/yr for the early years of the Puʻu ʻŌʻō eruption (Dvorak and Dzurisin, 1993). As discussed above, these estimates of magma residence time and volume most likely refer to the combination of the Halemaʻumaʻu and SC magma bodies.

4.3. The 1971–1982 period and inferences from the ongoing eruptions

The two separate decreasing temporal trends in Pb isotope ratios from 1971 to 1982 require that the upper- and lower-trend lavas were supplied from two distinct, poorly connected magma bodies (Fig. 5). For each eruptive period with Pb isotopic diversity (1971, 1974, and 1982), the lavas with the lowest $^{206}Pb/^{204}Pb$ ratios have the highest MgO contents (Fig. 4). These differences in MgO probably reflect a higher pre-eruptive magma temperature (up to ∼40 °C) for the lower-trend lavas (Appendix A). The association between the vent locations of the 1970s-era lower-trend lavas near the southeast rim of the caldera (Fig. 1) and their higher MgO contents suggest that these lavas tapped the relatively deep (∼2–4 km), and presumably, hotter SC magma body. In contrast, the 1970s-era upper-trend lavas (which erupted from vents within or near Halemaʻumaʻu) were likely derived from the magma body inferred to exist beneath the eastern rim of Halemaʻumaʻu (Anderson et al., 2015; Baker and Amelung, 2012; Cervelli and Miklius, 2003), confirming that this deformation center is associated with a site of magma storage that must have existed prior to 1975. The lower MgO contents of the upper-trend lavas can be explained if the Halemaʻumaʻu magma body is relatively shallow (<2 km) and cooled by an active hydrothermal system (Fig. 5).

Lavas from Sep. 1982 (erupted from vents on the floor of the southern portion of the caldera; Fig. 1) display a relatively large range in $^{206}Pb/^{204}Pb$ (for a <1-day eruption at Kīlauea's summit) that correlates with MgO content (Fig. 4) and the time of eruption. These rapid changes in lava chemistry probably reflect syn-eruptive mixing of two compositionally distinct magmas:

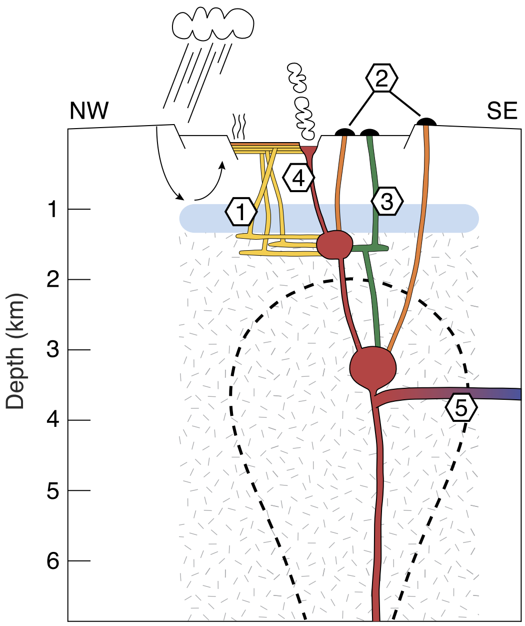

Fig. 5. Schematic cross section of Kīlauea's summit region, showing the internal geometry of the volcano's summit magma storage reservoir. Kīlauea's summit reservoir comprised two magma bodies (red) since 1971 (and probably since 1959): the shallower Halemaʻumaʻu magma body and the deeper "South Caldera" (SC) magma body. The dashed line marks the volcano's aseismic region (Klein et al., 1987), which represents the maximum possible volume of the summit reservoir (~40 km^3). However, residence time modeling based on the Pb isotope ratios of Kīlauea summit lavas suggests that the total volume of magma within the summit reservoir during the late 20th century was only ~0.1–0.5 km^3. This discrepancy can be resolved if the aseismic region includes an olivine-rich crystal-mush zone and hot, ductile rocks (stippled pattern) that likely surround the summit reservoir (e.g., Pietruszka and Garcia, 1999b). The blue area overlying the Halemaʻumaʻu magma body represents the volcano's active summit hydrothermal system. Eruptions within Halemaʻumaʻu (fed by transient eruption-related dikes and sills) that likely tapped the shallower magma body in 1961 and 1967–1968 led to ponded lava flows (#1, yellow) that are currently overlain by a ponded lava flow from the Sep. 1974 eruption (orange, at surface). The Aug. 1971 and Jul. 1974 eruptions (both represented by #2, orange) were supplied (separately) from both the shallower and deeper magma bodies. In Apr. and Sep. 1982, magma from both bodies was variably mixed, possibly in a small transient eruption-related sill or dike (#3, green). Today, an active lava lake within Halemaʻumaʻu is thought to be supplied directly from the shallower Halemaʻumaʻu magma body, which is likely connected to the deeper SC magma body (both #4, red). The location of the dike that transports magma from the summit reservoir to the Puʻu ʻŌʻō vent on the east rift zone is controversial. Cervelli and Miklius (2003) suggest that the Puʻu ʻŌʻō eruption is fed directly from the Halemaʻumaʻu magma body, whereas Poland et al. (2014) suggest that it is fed from the SC magma body (bypassing the Halemaʻumaʻu magma body). We favor the latter interpretation, and schematically represent a dike just below the base of the SC magma body to suggest that magma feeding the Puʻu ʻŌʻō rift eruption (#5) may partially bypass the summit reservoir (e.g., Garcia et al., 1996). (For interpretation of the references to color in this figure caption, the reader is referred to the online version of this article.)

(1) a high MgO, lower-trend end member derived from the SC magma body and (2) a low MgO, upper-trend end member derived from the Halemaʻumaʻu magma body (Fig. 5). Like one of the Sep. 1982 lavas, the Apr. 1982 lava that we analyzed has a $^{206}Pb/^{204}Pb$ ratio between the upper- and lower-trend (as defined by the end-member Sep. 1982 lavas). Thus, the Apr. 1982 lava is likely also a mixture of magmas from the two magma bodies. However, lavas from the Apr. 1982 eruption (which had vents within and near Halemaʻumaʻu) were compositionally homogeneous (Garcia et al., 2003), so any magma mixing must have been completed prior to the start of this <1-day eruption. The location of magma mixing for the Apr. and Sep. 1982 eruptions is unknown, but it could have occurred in one or more small sills or dikes (<0.003 km^3, based on the volume of erupted lava; Macdonald et al., 1983) that solidified soon after each eruption ended (Fig. 5).

The mixing model for the lower-trend lavas—erupted from the deeper SC magma body—suggests a best-fit magma residence time of 1.8–2.4 yr (with a maximum range of 1.5–2.9 yr), which corresponds to a magma volume of ~0.1–0.3 km^3 (0.09–0.3 km^3 maximum range) assuming a 0.06–0.11 km^3/yr range in the rate of magma supply to the volcano's summit reservoir. In contrast, the mixing model for the upper-trend lavas—erupted from the shallower Halemaʻumaʻu magma body—suggests a longer best-fit magma residence time of 5.9–8.3 yr (with a maximum range of 3.9–20.6 yr). A maximum $^{206}Pb/^{204}Pb$ range of 18.417–18.475 for the input magma related to the upper trend (not shown in Fig. 2a) is permitted by the mixing model. This value (like the maximum range of magma residence times) is poorly constrained, probably because the shallow Halemaʻumaʻu magma body was supplied from the deeper SC magma body, which was evolving in its Pb isotopic composition on a similar time scale.

The initiation of the five-year long Mauna Ulu rift eruption in 1969 coincided with an increasing diversion of magma from Kīlauea's summit reservoir to the east rift zone (Dvorak and Dzurisin, 1993), which (as it accelerated in the mid-1970s) may have starved the summit reservoir of magma and led to the partial disconnection of the Halemaʻumaʻu magma body from the SC magma body. Thus, the rate of magma supply to the Halemaʻumaʻu magma body from 1971 to 1982 is likely to have been much lower than the overall magma supply rate to the summit reservoir, particularly if the volcano's east rift zone primarily taps the deeper SC magma body (Poland et al., 2014). In this case, the partitioning of magma within Kīlauea's shallow plumbing system (Dvorak and Dzurisin, 1993) suggests that only ~0.01–0.03 km^3/yr of magma (Appendix A) was delivered to the Halemaʻumaʻu magma body from 1971 to 1982 (the remainder bypassed it on the way to the east rift zone). This corresponds to a magma volume of ~0.06–0.2 km^3 for the Halemaʻumaʻu magma body (0.04–0.6 km^3 maximum range). The sum of the best-fit magma volumes for the 1971–1982 period (~0.2–0.5 km^3) is probably not significantly different from the volume of magma within the summit reservoir for the 1959–1968 period (~0.1–0.3 km^3) given (1) the uncertainty associated with the estimated magma supply rate to the poorly connected Halemaʻumaʻu magma body from 1971 to 1982, and (2) the likelihood that the inferred magma volume for the Halemaʻumaʻu magma body from 1971 to 1982 represents a maximum (Appendix A) due to the continuous temporal decrease in the $^{206}Pb/^{204}Pb$ ratio of the input magma from the deeper SC magma body (Fig. 2d). Thus, we tentatively conclude that the average volume of magma within the summit reservoir remained essentially constant during the late 20th century (despite short-term changes; Dvorak and Dzurisin, 1993).

The $^{206}Pb/^{204}Pb$ ratios of the lower-trend Aug. 1971 lavas are significantly lower than both the 1968 lavas and the inferred Pb isotopic composition of the input magma for the 1959–1968 period (#1 in Fig. 2a). This suggests that a second compositionally distinct, mantle-derived magma intruded Kīlauea's summit reservoir between 1968 and Aug. 1971 (#2 in Fig. 2a), following the 8-month long 1967–1968 eruption in Halemaʻumaʻu (itself preceded by a 10-month period of continuous summit inflation; Wright and Klein, 2014) and preceding the initiation of the sustained Mauna Ulu eruption in 1969. A narrow maximum $^{206}Pb/^{204}Pb$ range of 18.451–18.458 for this input magma is permitted by the mixing model. This relatively high $^{206}Pb/^{204}Pb$ ratio, compared to Puʻu ʻŌʻō lavas that erupted after the end (in 1985) of the early period of magma mixing with an evolved rift-stored magma (Garcia et al., 1992), indicates that Puʻu ʻŌʻō-type magma was not present within the summit reservoir until after Sep. 1982. Instead, a compositionally distinct, Puʻu ʻŌʻō-type magma was probably delivered to the volcano from the mantle shortly before, during, or after the initiation of this eruption in 1983 (#3 in Fig. 2a). It is unknown if (or when) this early Puʻu ʻŌʻō-type parental magma entered the summit reservoir (due to the lack of summit eruptions between 1982 and 2008). The observation that most of this period was dominated by summit deflation (Johnson et al., 2010; Poland et al., 2012) suggests that the rate of magma accumulation in the summit reservoir decreased significantly after 1982 as the mantle-derived input magma was preferentially transferred to the eruptive vent on the volcano's east rift zone. Thus, the short-term fluctuations in the $^{206}Pb/^{204}Pb$ ratios of Puʻu ʻŌʻō lavas (Fig. 2a)—if they are derived from the melting of small-scale heterogeneities within the Hawaiian plume (Greene et al., 2013; Marske et al., 2008)—may be preserved because the parental magmas for this eruption partially bypassed the summit reservoir (Garcia et al., 1996), and avoided its buffering effects (Fig. 5).

Tephra that erupted from Halemaʻumaʻu in 2009 (Fig. 2b) has a $^{206}Pb/^{204}Pb$ ratio similar to contemporaneous Puʻu ʻŌʻō lavas, confirming the observation of Thornber et al. (2015) that a Puʻu ʻŌʻō-type parental magma eventually infiltrated Kīlauea's summit reservoir. However, subsequent tephra from the Halemaʻumaʻu eruption in 2010 trend towards lower $^{206}Pb/^{204}Pb$ ratios (Fig. 2b) and lower $^{208}Pb/^{204}Pb$ at a given $^{206}Pb/^{204}Pb$ than Puʻu ʻŌʻō lavas from 2008–2012 (Fig. 3a). This difference suggests that a compositionally distinct, mantle-derived magma was recently delivered to the summit reservoir (#4 in Fig. 2a) that has not yet (as of 2012) erupted at Puʻu ʻŌʻō. One possibility for the timing of this input is the 2003–2007 surge (peaking in 2007) of magma (Poland et al., 2012), which led to a reversal of the long-term summit deflation and exceptionally high rates of magma supply from the mantle ($\sim$0.2 km^3/yr).

4.4. Comparison with estimates based on geophysics

Estimates for the volume of Kīlauea's summit reservoir (based mostly on geophysics) vary widely from $\sim$0.1–6 km^3 for the Halemaʻumaʻu magma body to $\sim$13–27 km^3 for the SC magma body (Anderson et al., 2015; Dawson et al., 1999; Johnson, 1992; Poland et al., 2009; Segall et al., 2001), as reinterpreted by Anderson et al. (2015) and Poland et al. (2014). Most estimates for the volume of the Halemaʻumaʻu magma body have been obtained using the so-called "manometer" approach (Anderson et al., 2015; Johnson, 1992; Segall et al., 2001), in which measurements of the change in the volume of the volcanic edifice (from geodetic data) are compared with measurements related to a change in the pressure of the underlying magma (e.g., the level of an active lava lake). The results of this approach (0.1–6 km^3) range to much higher values than our best estimate for the volume of magma within the Halemaʻumaʻu magma body from 1971 to 1982 (0.06–0.2 km^3). An even larger discrepancy is observed in the geophysics-based estimates for the volume of the SC magma body. Johnson (1992), as reinterpreted by Poland et al. (2014), used a "manometer" approach to estimate a volume of 13 km^3 for the SC magma body, whereas a volume of 27 km^3 was found using anomalies in seismic velocity beneath Kīlauea's summit (Dawson et al., 1999). These estimates are much larger than our best estimate for the volume of magma within the SC magma body from 1971 to 1982 (0.1–0.3 km^3). Sensitivity tests of the model (Appendix A) show that the short residence times (and thus, small magma volumes) based on the Pb isotope ratios of Kīlauea summit lavas cannot be reconciled with the much larger volumes based on geophysics. However, the discrepancy can potentially be resolved (e.g., Pietruszka and Garcia, 1999b) if (1) the geophysics-based estimates include an olivine-rich crystal-mush zone and hot, ductile rocks (Fig. 5) that likely surround the summit reservoir and (2) the estimates based on the Pb isotope ratios of the lavas represent only the hotter, molten core of each body in which magma mixing occurs.

5. Identification of magma batches at Kīlauea Volcano

The identification of magma batches, and their specific compositions free of the effects of shallow crustal processing (i.e., magma mixing and differentiation), can potentially be used to pinpoint the chemical and isotopic signatures of mantle heterogeneity and melt generation within the Hawaiian plume (e.g., Pietruszka et al., 2013). Early studies (e.g., Wright and Fiske, 1971; Wright and Tilling, 1980) considered the relatively uniform and unique chemistry of lavas from many individual Kīlauea eruptions to represent discrete magma batches from the mantle. Indeed, the classic hypothesis for Kīlauea states that individual batches of magma are delivered from the mantle for each summit eruption, and summit eruptions end when a particular batch of magma has been exhausted (e.g., Wright and Klein, 2014). However, residence time modeling based on the Pb isotope ratios of Kīlauea summit lavas (Fig. 2) suggests that most of the compositional differences for summit lavas erupted during the late 20th century (beyond the effects of crystal fractionation or accumulation) are related to magma mixing within the volcano's summit reservoir. Indeed, the model results indicate that only two mantle-derived magma batches were delivered to the summit reservoir from 1959 to 1982, each with a narrow maximum range in $^{206}Pb/^{204}Pb$ of only $\sim$0.008 for the 1959–1968 period (#1 in Fig. 2a) and $\sim$0.007 for the 1971–1982 period (#2 in Fig. 2a). Thus, melt extracted from Kīlauea's source region within the Hawaiian plume can, at times, remain essentially constant in composition on a time scale of up to $\sim$10 years. Each of these two relatively large, homogeneous magma batches were delivered to the volcano nearly continuously for much longer than the duration of individual summit eruptions during the late 20th century ($<$1 day to $\sim$5 weeks, except for the 8-month long 1967–1968 Halemaʻumaʻu eruption).

Following the input of a third mantle-derived magma batch to the volcano shortly before, during, or after the initiation of the Puʻu ʻŌʻō eruption in 1983 (#3 in Fig. 2a), Puʻu ʻŌʻō lavas display short-term fluctuations in their Pb isotope ratios on a time scale of $<$10 years (Fig. 2a) that are potentially controlled by mantle processes (Greene et al., 2013; Marske et al., 2008). These observations for Puʻu ʻŌʻō lavas suggest that, unlike the 1959–1982 period, the composition of melt extracted from Kīlauea's source region within the Hawaiian plume can also vary almost continuously (in this case, with a larger total range in $^{206}Pb/^{204}Pb$ of $\sim$0.05; Fig. 2a). This behavior implies that more frequent magma batches delivered to the volcano since the start of the Puʻu ʻŌʻō eruption have been relatively small in volume (and poorly homogenized during transport from the source to the surface). The apparent differences in the volume of the magma batches delivered to Kīlauea since 1959 may be related to the size of the small-scale heterogeneities that are melting within the volcano's mantle source region, from thin filaments $\sim$1–3 km in diameter (Greene et al., 2013) to larger $\sim$5-10 km thick blobs (Marske et al., 2007). A fourth mantle-derived magma batch (#4 in Fig. 2a) must have been recently delivered to Kīlauea's summit reservoir, probably during the peak of the magma surge in 2007 (Poland et al., 2012). Continued monitoring is required to see if the temporal variations in the Pb isotope ratios of Halemaʻumaʻu tephra since 2009 (Fig. 2b) are part of a new trend of (1) magma mixing within the summit reservoir similar to the 1959–1968 and 1971–1982 periods or (2) rapid fluctuations analogous to Puʻu ʻŌʻō lavas erupted since 1985. In any case, the input of four of these melts from the mantle (three of which are known to have entered the summit reservoir) correlate with the initiation of energetic (1959 Kīlauea Iki) and/or sustained (1969–1974 Mauna Ulu, 1983-present Puʻu ʻŌʻō, and 2008-present Halemaʻumaʻu) eruptions (Fig. 2a). This association suggests that Kīlauea's eruptive behavior is partly tied to the delivery of new magma batches from the Hawaiian plume.

6. Summary and broader implications

Deliveries of isotopically distinct, mantle-derived magma batches to Kīlauea reveal unambiguously—for the first time based on variations in lava chemistry—that the volcano's summit reservoir comprised two magma bodies since 1971 (and probably since at least 1959). The locations of these magma bodies—a deeper one ∼2–4 km beneath the southern rim of the caldera and a shallower one <2 km beneath the east rim of Halema'uma'u—are inferred to coincide with two major deformation centers identified by geodetic monitoring of the volcano's summit region (Anderson et al., 2015; Baker and Amelung, 2012; Cervelli and Miklius, 2003). The total volume of magma within the summit reservoir from 1959 to 1982 was only ∼0.1–0.5 km^3. This is, at most, a ∼1- to 8-year supply of magma given a 0.06–0.11 km^3/yr range of magma supply rates (Dvorak and Dzurisin, 1993), and may be a short as six months in recent years due to even higher rates of magma supply (up to ∼0.2 km^3/yr from 2003 to 2007; Poland et al., 2012). The volume of magma within the summit reservoir probably decreased during the first 20 years of the Pu'u 'Ō'ō eruption (Thornber, 2003), until the period of summit deflation ended in 2003 (Johnson et al., 2010; Poland et al., 2012). Thus, the ongoing, 32-year old Pu'u 'Ō'ō eruption (>4 km^3 of lava erupted; Orr et al., 2015) must be sustained by an essentially continuous supply of new melt from the mantle.

The geometry of Kīlauea's summit magma storage reservoir is, in many ways, similar to the magmatic plumbing system beneath the summit of Piton de la Fournaise, a frequently active basaltic shield-building volcano on the Island of Réunion (Indian Ocean). Petrologic, geochemical, and geophysical observations (e.g., Di Muro et al., 2014; Famin et al., 2009; Peltier et al., 2009; Prôno et al., 2009) suggest that summit eruptions at Piton de la Fournaise are supplied from a deep magma body near the base of the oceanic crust (∼7.5 km deep), a shallow magma body near sea level (∼2.5 km deep), and a near-surface complex of partially solidified dikes and sills (<1 km deep). Estimates for the volume of magma within the shallow body (based on lava chemistry) range from ∼0.1 to 0.4 km^3 (Albarède, 1993; Sigmarsson et al., 2005; Vlastélic et al., 2009), which is remarkably similar to our estimate for the summit magma storage reservoir of Kīlauea Volcano during the late 20th century. This comparison suggests that (1) frequently active basaltic volcanoes may commonly have two or more magma bodies at different depths beneath their summits, and (2) the volume of magma within these bodies at any one time may be exceedingly small. These inferences, if correct, have implications for the hazards associated with these volcanoes (e.g., Michon et al., 2013; Swanson et al., 2012, 2014), such as the maximum potential volume of caldera collapse and the amount of juvenile material that may be emitted during explosive eruptions.

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Appendix A. Supplementary material

Supplementary material related to this article can be found online at http://dx.doi.org/10.1016/j.epsl.2014.12.040.

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