Characteristics of Hawaiian Volcanoes
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Chapter 5

Magma Supply, Storage, and Transport at Shield-Stage Hawaiian Volcanoes

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Abstract

The characteristics of magma supply, storage, and transport are among the most critical parameters governing volcanic activity, yet they remain largely unconstrained because all three processes are hidden beneath the surface. Hawaiian volcanoes, particularly Kīlauea and Mauna Loa, offer excellent prospects for studying subsurface magmatic processes, owing to their accessibility and frequent eruptive and intrusive activity. In addition, the Hawaiian Volcano Observatory, founded in 1912, maintains long records of geological, geophysical, and geochemical data. As a result, Hawaiian volcanoes have served as both a model for basaltic volcanism in general and a starting point for many studies of volcanic processes.

Magma supply to Hawaiian volcanoes has varied over millions of years but is presently at a high level. Supply to Kīlauea's shallow magmatic system averages about 0.1 km^3/yr and fluctuates on timescales of months to years due to changes in pressure within the summit reservoir system, as well as in the volume of melt supplied by the source hot spot. Magma plumbing systems beneath Kīlauea and Mauna Loa are complex and are best constrained at Kīlauea. Multiple regions of magma storage characterize Kīlauea's summit, and two pairs of rift zones, one providing a shallow magma pathway and the other forming a structural boundary within the volcano, radiate from the summit to carry magma to intrusion/eruption sites located nearby or tens of kilometers from the caldera. Whether or not magma is present within the deep rift zone, which extends beneath the structural rift zones at ~3-km depth to the base of the volcano at ~9-km depth, remains an open question, but we suggest that most magma entering Kīlauea must pass through the summit reservoir system before entering the rift zones. Mauna Loa's summit magma storage system includes at least two interconnected reservoirs, with one centered beneath the south margin of the caldera and the other elongated along the axis of the caldera. Transport of magma within shield-stage Hawaiian volcanoes occurs through dikes that can evolve into long-lived pipe-like pathways. The ratio of eruptive to noneruptive dikes is large in Hawai‘i, compared to other basaltic volcanoes (in Iceland, for example), because Hawaiian dikes tend to be intruded with high driving pressures. Passive dike intrusions also occur, motivated at Kīlauea by rift opening in response to seaward slip of the volcano's south flank.

Introduction

Perhaps the most fundamental, yet least understood, aspects of volcanic activity involve the supply, storage, and transport of magma. Is magma supplied to a volcano episodically or continuously (and if the latter, what is the rate of supply)? What determines the location and geometry of subsurface magma plumbing? How does magma move between source reservoirs and eruptive vents? Despite significant advances in volcanology, especially since the 1950s, these questions continue to motivate much of the research into volcanic and magmatic processes.

Magma supply to subsurface reservoirs is possibly the most important control on eruptive and intrusive activity at a volcano (Wadge, 1982; Dvorak and Dzurisin, 1993). In many cases, supply is episodic, and the arrival of magma batches is frequently invoked as a trigger for eruption. For instance, an intrusion of fresh basalt into a dacite magma body is interpreted as the trigger for the 1991 eruption of Mount Pinatubo, Philippines (Pallister and others, 1992). Similarly, repeated episodes of magma intrusion and accumulation beneath Eyjafjallajökull, Iceland, over at least 18 years culminated in an eruption in 2010 (Sigmundsson and others, 2010). Episodes of magma accumulation are not always associated with eruption, however, as exemplified by the sudden onset and gradual cessation of inflation, inferred to be the result of magma accumulation, near South Sister, Oregon (Dzurisin and others, 2009). At other volcanoes, like Kīlauea, magma supply is inferred

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to be nearly continuous over decadal time scales, with the supply rate roughly proportional to the eruption rate (for example, Swanson, 1972; Dvorak and Dzurisin, 1993; Cayol and others, 2000).

Magma supplied to a volcano is typically stored in reservoirs below the surface prior to intrusion or eruption. Such reservoirs occupy a wide depth range. For example, seismic and geodetic data place New Mexico's Socorro magma body at a depth of ~20 km (Reilinger and others, 1980; Fialko and Simons, 2001), but a zone of magma accumulation at Kīlauea may be only 1–2 km beneath the surface (Poland and others, 2009; Anderson and others, in press). Reservoir depth is controlled by (1) density, as magma ascent stalls where magma and host rock density are equal (for example, Ryan, 1987; Pinel and Jaupart, 2000; Burov and others, 2003), and (2) mechanical resistance, where stress barriers due to the thermal or mechanical properties of the host rock, relative to the magma, inhibit magma flow (for example, Gudmundsson, 1987; Martel, 2000; Burov and others, 2003). Magma reservoir volumes also vary widely, from hundreds of cubic kilometers, such as those that fed silicic ash-flow eruptions, like that from the Yellowstone Caldera about 0.64 Ma (Christiansen, 2001), to about 1 cubic kilometer for the shallow magma reservoir at Kīlauea (Poland and others, 2009).

From storage areas, magma is transported toward the surface via porous flow through partially molten rock, diapiric ascent, or dike and sill intrusion (Rubin, 1995). Dikes and sills are perhaps the most common means of magma transport, as suggested by their abundance in eroded volcanoes (for example, Gudmundsson, 1983; Walker, 1987), as well as monitoring data from active volcanoes (for example, Pollard and others, 1983). Dike propagation is triggered by both source overpressure (for example, Rubin, 1990) and tectonic extension (Bursik and Sieh, 1989) and can be either vertical or lateral, possibly resulting in eruptions far from the source magma reservoir (for example, Sigurdsson and Sparks, 1978). Despite research into the mechanics of magma transport, important problems remain—for instance, distinguishing between intrusions that will erupt versus those that will not, which is especially important during volcanic crises (for example, Geshi and others, 2010).

Kīlauea and Mauna Loa volcanoes, in Hawai‘i, offer an outstanding opportunity to explore magma supply, plumbing, and transport, owing to their accessibility, frequent eruptive activity, and excellent long-term monitoring record made possible by over 100 years of study by the Hawaiian Volcano Observatory (HVO) and collaborators. This chapter synthesizes past research and new insights from recent volcanic activity at Kīlauea and Mauna Loa to develop refined models for magma supply, magma storage, and magma transport at shield-stage Hawaiian volcanoes. We discuss those three topics, in turn, beginning each discussion with a review of current understanding, based on studies of volcanoes around the world, before focusing on insights provided by Hawaiian volcanoes. Our goal is to derive new knowledge of magmatic processes in Hawai‘i that may serve as an example for other volcanoes, following the well-established tradition that models based on Hawaiian volcanism provide a basis for understanding volcanic activity on the Earth and on other planets.

A comprehensive review of even just one aspect of Hawaiian volcanism (surface deformation or seismology, for example) is impossible, owing to the vast literature on such topics. Instead, we seek to mark HVO's 2012 centennial and beyond by (1) building on previous studies (the examination of supply, storage, and eruption by Decker [1987] from HVO's 75th-anniversary volume, for example) and (2) by incorporating insights from new datasets (including Interferometric Synthetic Aperture Radar [InSAR] and the Global Positioning System [GPS]) and from recent volcanic activity to develop our conceptual models. A more historically comprehensive analysis of the eruptive and intrusive history of Kīlauea, with special emphasis on magma supply, storage, and transport, is provided by Wright and Klein (2014).

Magma Supply

Magma supply from a source region to the base of a volcano can be estimated only indirectly. Over long time periods, magma supply can be approximated by modeling the volume of melt needed to sustain a measured heat flow at the surface. For example, Bacon (1982) calculated a basaltic magma supply rate of 570 km^3/m.y. for the Coso volcanic field, California, on the basis of heat flow measurements. Similarly, Guffanti and others (1996) used a petrologic model of basalt-driven volcanism to determine the basalt influx per unit area necessary to account for erupted volumes and heat flow at two volcanic centers in northern California: the Caribou volcanic field (minimum basalt influx calculated to be 0.3 km^3 per km^2 per m.y.) and the Lassen volcanic center (minimum basalt influx of 1.6 km^3 per km^2 per m.y.). The same principle can also be applied over shorter time scales. For instance, Francis and others (1993) inferred heat loss from the Halema‘uma‘u lava lake that occupied Kīlauea's summit caldera before 1924 and calculated that the magma supply was much higher than the eruption rate, implying that the volcano grew endogenously through subsurface intrusion and cumulate formation.

Calculations of contemporary rates of magma supply have been made at only a few volcanoes on Earth and are most often accomplished using eruption rate, gas emission, and deformation data. For example, Wadge (1982) calculated the supply rates to several volcanoes on the basis of eruption volumes over time. Burt and others (1994) and Wadge and Burt (2011) inferred a 2.5-fold increase in magma supply to Nyamulagira, Democratic Republic of the Congo, starting in 1977–80, based on an increase in erupted volumes over time. Lowenstern and Hurwitz (2008) used CO_2 degassing to infer an influx of 0.3 km^3/yr of mantle-derived basalt beneath Yellowstone Caldera, which is consistent with measured heat flow. Allard (1997) and Allard and others (2006) determined that about 3–4 times more magma degassed than erupted at Mount Etna during 1975–2005, which suggests a supply rate to the shallow volcanic system of less than 0.1 km^3/yr.

Owing to infrequent eruptions at many volcanoes and the challenge in collecting comprehensive gas emission datasets in general, magma supply rates are commonly estimated from deformation data. At Okmok volcano, Alaska, Lu and others (2005) and Fournier and others (2009) inferred variable supply to the volcano's shallow magma reservoir on the basis of fluctuations in inflation rate prior to eruptions in 1997 and 2008. Similarly, deformation measurements at Sierra Negra volcano, in the Galápagos Islands, indicated variable rates of subsurface magma accumulation, and supply by inference, prior to an eruption in 2005 (Chadwick and others, 2006). After eruptions at many volcanoes (including Okmok and Sierra Negra), reinflation is rapid but gradually wanes, suggesting that an initially large pressure difference between a deep magma source and the shallow magma storage reservoir drives magma ascent. The pressure difference decreases as the shallow reservoir fills, causing the supply rate to decrease. This concept explains the exponential decay of deformation rates at Kīlauea (Dvorak and Okamura, 1987) and several other basaltic volcanoes (for example, Lu and others, 2005, 2010; Fournier and others, 2009; Lu and Dzurisin, 2010) and has been used to argue for a "top-down" influence on magma supply—in other words, the pressure state of the shallow reservoir influences the rate of magma ascent from depth (Dvorak and Dzurisin, 1993).

Unfortunately, studies based on deformation data alone do not account for magma compressibility, which, in the case of strong host rock and the presence of exsolved volatiles within the magma, may cause 80 percent or more of the magma supplied or withdrawn from a reservoir to be accommodated without measurable surface inflation or deflation (Johnson, 1992; Johnson and others, 2000). Better constraints on magma supply are achieved by combining gas emissions, deformation, and effusion rate (in the case of ongoing eruptions). Using such a combination of data, the magma supply rate to Soufrière Hills volcano, Montserrat, was estimated to be about 0.06 km^3/yr and steady over several years, with much of that volume accommodated by compression of magma stored in a crustal reservoir (Voight and others, 2010). Physics-based models that utilize multidisciplinary datasets also hold promise for elucidating magma supply characteristics, as demonstrated by Anderson and Segall's (2011) determination that little or no recharge was associated with the 2004–08 eruption of Mount St. Helens, Washington.

Magma Supply to Hawaiian Volcanoes

The Hawaiian-Emperor chain of seamounts and islands extends over 6,000 km across the Pacific Ocean (fig. 1), tracing a record of volcanic activity over at least the past 70 m.y., with the age of volcanism progressively younger to the southeast (Clague and Dalrymple, 1987; Tilling and Dvorak, 1993). Volcanism is attributed to a mantle melting anomaly, termed a "hot spot" (Wilson, 1963). The origin of the hot spot is generally assumed to be a plume of high-temperature material upwelling from the deep mantle, as originally proposed by Morgan (1971), and supported by geochemical evidence of a primitive source region (for example, Kurz and others, 1983). Paleomagnetic evidence suggests that the plume is not fixed in the deep mantle but drifts over time, like during the formation of the Emperor Seamounts (Clague and Dalrymple, 1987; Tarduno and others, 2003), but plate motion models argue that Hawaiian hot spot motion may be relatively small (Gripp and Gordon, 2002; Wessel and others, 2006; Wessel and Kroenke, 2008). The plume hypothesis has been challenged as dogma that fails to explain numerous geochemical and geophysical datasets (see Hamilton, 2011, and references therein, for alternatives to a plume origin), but seismic tomography studies indicate a deep mantle source for the melting anomaly, providing evidence in favor of the presence of a mantle plume (Wolfe and others, 2009, 2011). Regardless of its origin, the Hawaiian hot spot is the most productive on Earth, based on buoyancy fluxes (for example, Sleep, 1990) and eruption rates (Wadge, 1982). Tracking magma supply from the hot spot to the crust thus provides important constraints on the production of melt over time.

Estimates of the Hawaiian hot spot's magma supply rate—the flux of magma from the source region to the surface—over millions of years are based on the volumes and ages of the islands and seamounts of the Hawaiian-Emperor chain. Early calculations used bathymetric and topographic maps to determine the volumes of individual volcanoes (Vogt, 1972; Bargar and Jackson, 1974). Later, more sophisticated studies incorporated seismic, gravity, and crustal loading data to quantify mass flux (White, 1993; Van Ark and Lin, 2004; Vidal and Bonneville, 2004; Robinson and Eakins, 2006). Regardless of the method used, all long-term magma supply calculations have found an increase in the activity of the Hawaiian hot spot over the past 30 m.y. after relatively constant melt production during the formation of the Emperor chain, with the highest rates occurring at present (fig. 2). The current high supply may be reflected in the geochemistry of erupted lavas. Weis and others (2011) suggested that the strong appearance of the "Loa" geochemical composition (see Clague and Sherrod, this volume, chap. 3; Helz and others, this volume, chap. 6) may coincide with increased magma supply over the past 5 m.y. Variations in the supply rate on the scale of millions of years have been attributed to changes in lithospheric thickness (especially where the hot spot crossed fracture zones), temperature, and age, as well as pulsation of the hot spot itself (Vogt, 1972; White, 1993; Van Ark and Lin, 2004; Vidal and Bonneville, 2004).

Although the magma supply rates calculated by various studies differ by several times (fig. 2), they are similar in two important respects. First, activity of the Hawaiian hot spot (manifested primarily by erupted volume) varies over millions of years, with the highest values occurring in recent times (Clague and Dalrymple, 1987). Second, a dominant influence on hot-spot activity is the rate of melt production. The volume

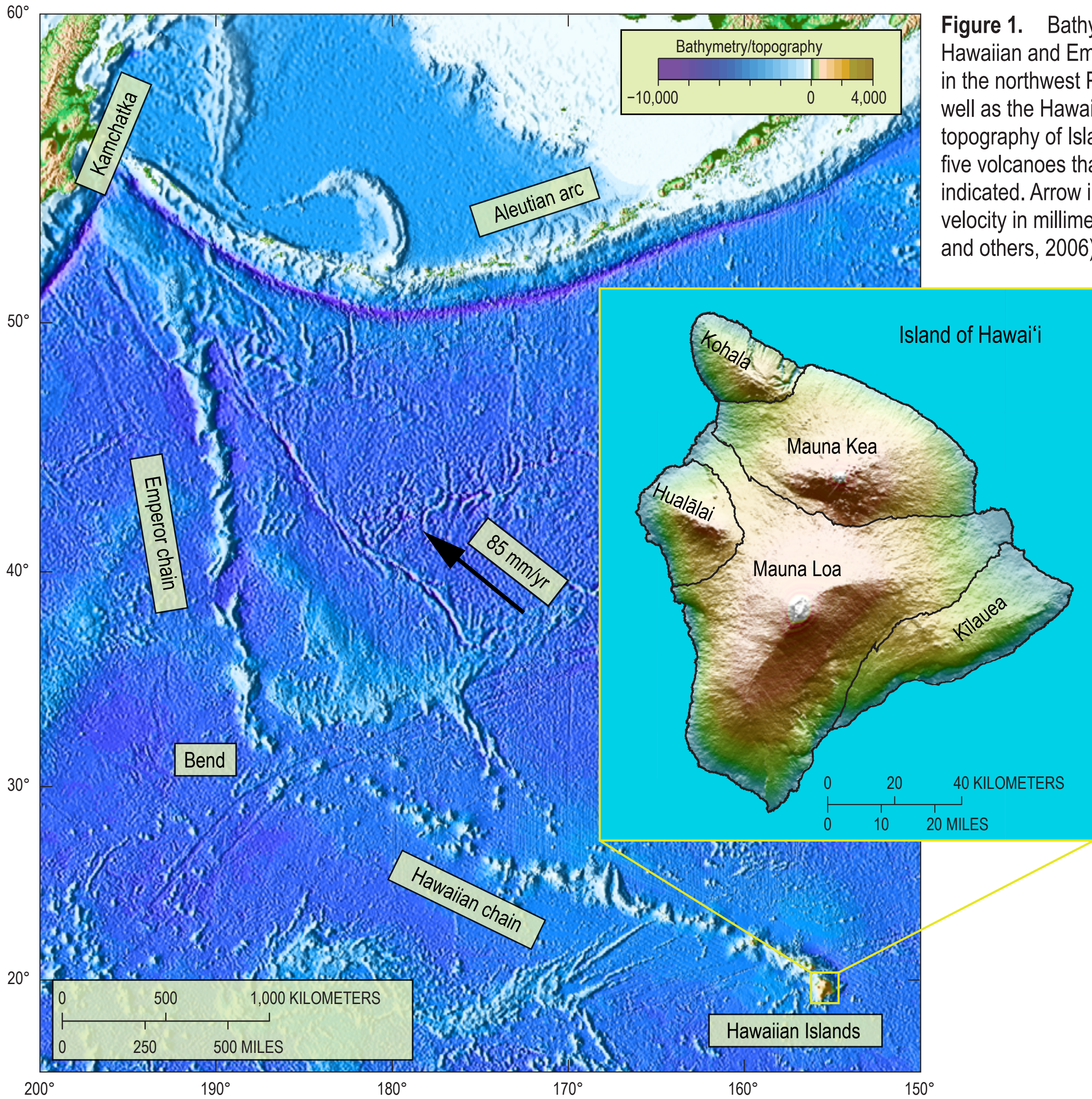


Figure 1. Bathymetric map showing the Hawaiian and Emperor seamount chains in the northwest Pacific Ocean basin, as well as the Hawaiian Islands. Inset shows topography of Island of Hawaiʻi, with the five volcanoes that make up the island indicated. Arrow indicates plate motion velocity in millimeters per year (Simkin and others, 2006).

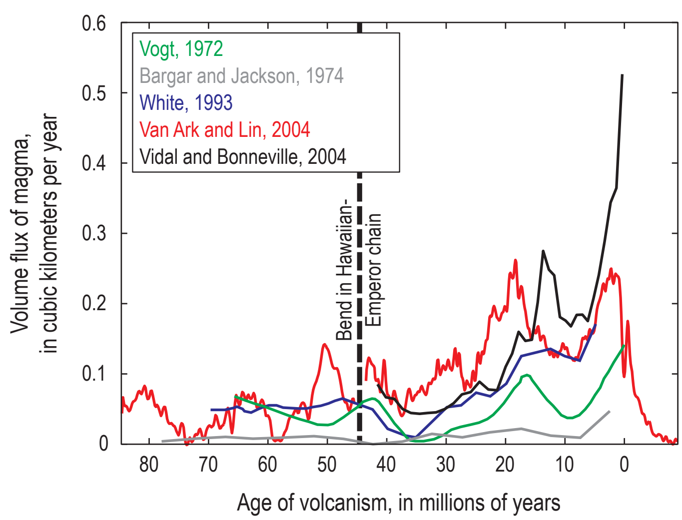


Figure 2. Plot of the volume flux of the Hawaiian hot spot over time, as calculated by various authors. The estimate of Bargar and Jackson (1974) is based on the volume of seamounts and islands in the Hawaiian-Emperor chain, with the timing established by Robinson and Eakins (2006). Vogt (1972) used a similar volume-age relation. White (1993) based his volume flux curve on crustal thickening determined seismically, coupled with the volume of seamounts and volcanoes and their underplated roots, assuming Airy isostasy. Vidal and Bonneville (2004) distinguished topography/bathymetry created by the Hawaiian Swell (caused by uplift driven by the mantle plume) from that caused by volcanism, which is a better measure of the magma production rate. Finally, Van Ark and Lin (2004) calculated crustal thicknesses from gravity data to compute hot-spot volume flux over time. Dashed vertical line is the location of the bend in the Hawaiian-Emperor chain of seamounts and islands (see fig. 1).

of melt provided to a volcano by the hot spot apparently varies over millions of years, and perhaps on even shorter timescales. The composition of Mauna Kea basalt accessed through drilling, for example, suggests waxing and waning supply within the shield-building stage of that volcano (less than 1 m.y. in duration), instead of a smoother increase and then decrease in supply as the volcano grew and was then carried away from the hot spot (Stolper and others, 2004; Rhodes and others, 2012). The possibility thus exists that shorter-term variations in melt production occur in addition to longer-term changes documented by studies of the Hawaiian hot-spot track.

Contemporary Estimates

Determinations of the volume of erupted material over time at Mauna Loa and Kīlauea provide estimates of the minimum contemporary magma supply. Stearns and Macdonald (1946, p. 113–114), for example, noted that the total volume of erupted products at Kīlauea during 1823–1945 was approximately 3.2 km^3, corresponding to an annual rate of about 0.03 km^3/yr. Moore (1970) calculated a combined average annual eruption rate of 0.05 km^3/yr for Kīlauea and Mauna Loa during 1823–1969 and noted that the volume was less by a factor of 5 than the annual volume of subsidence caused by loading. He attributed the discrepancy to an unknown volume of magma added by intrusion and submarine eruption. Both of these estimates are consistent with that of Quane and others (2000), who determined an eruption rate of 0.05 km^3/yr over the past 350,000 years at Kīlauea, based on lava ages from a drill core from the lower East Rift Zone (ERZ).

Swanson (1972) made the first estimate of magma supply rate to Kīlauea by accounting for volumes of both stored magma and erupted lava. His calculation of 0.11 km^3/yr (dense rock equivalent) during 1952–71 was based on the volumes of three effusive eruptions (each of which lasted at least 4.5 months), during which there was little change in the amount of magma stored beneath the summit. In addition, he speculated that this supply rate was constant even between eruptions, when Kīlauea inflated to accommodate magma influx. He interpreted the supply rate to reflect the total amount of magma produced by the hot spot, given the lack of Mauna Loa eruptions during 1952–71. Swanson's (1972) supply rate eliminates the discrepancy between the volumes of subsidence and eruption rate cited by Moore (1970) when simple isostatic compensation is assumed.

Subsequently, numerous authors built on Swanson's (1972) work by combining models of surface deformation with extrusion volumes to calculate supply rates (table 1). For example, Dzurisin and others (1984) calculated an average supply rate of 0.09 km^3/yr during 1956–1983, although the period included short-term fluctuations associated with changing eruptive activity. Dvorak and Dzurisin (1993) found that magma supply between the 1960s and 1990s varied over days to years between 0.02 km^3/yr and 0.18 km^3/yr, with changes in rate controlled by the pressure in the summit magma reservoir. Estimates of the supply rate during the sustained Puʻu ʻŌʻō ERZ eruption, which started in 1983, fall within this range. Wolfe and others (1987) determined a supply rate of 0.12 km^3/yr from the first 20 episodes of the eruption in 1983–84, while Denlinger (1997) inferred a supply rate of 0.08 km^3/yr based on the extrusion rate during a period of zero summit elevation change. Heliker and others (2003) calculated an average effusion rate of 0.12 km^3/yr over 19 years that were dominated by summit deflation (implying that nearly all of the magma entering the volcano's summit reservoir system was erupted), although Garcia and others (1996) suggested that minor magma supply variations sometimes influenced the composition of erupted lava during the Puʻu ʻŌʻō activity.

The average magma supply rate in published reports covering the period 1952–2002 (table 1) is 0.1 km^3/yr. Wright and Klein (2014) presented a comprehensive analysis of magma supply since 1790, independent of the studies summarized in table 1, and also found that the supply rate since the 1960s was relatively steady. Short-term increases or decreases in rate are probably caused by fluctuations in pressure within Kīlauea's shallow summit magma reservoir system that promote or suppress supply from depth (for instance, Cervelli and Miklius, 2003). There is some evidence that the supply rate has been steady, at least since the 1800s. Francis and others (1993) determined that a magma supply rate of about 0.12 km^3/yr was needed to account for heat loss from the lava lake occupying Kīlauea's summit prior to the 1924 collapse, even though the eruption rate—which was up to an order of magnitude lower (Francis and others, 1993; Wright and Klein, 2014)—indicated that most of the supplied magma was stored, not erupted.

Models of geodetic data from Kīlauea suggest deep (~4–10 km) dilation beneath the East and Southwest Rift Zones (for example, Dieterich, 1988; Delaney and others, 1990; Owen and others, 1995, 2000a; Cayol and others, 2000). If this opening volume is filled by magma, as is interpreted from seismic data (Ryan, 1988; Wyss and others, 2001), the magma supply rate to Kīlauea could be almost twice that estimated from eruption and shallow magma storage alone. Cayol and others (2000) suggested that the magma supply rate to Kīlauea during 1961–91 was relatively steady, at 0.18 km^3/yr, based on deformation modeling, consistent with the analysis of Wright and Klein (2014). Such a high value implies that the CO_2 content of the primary magma supplied to Kīlauea is ~0.70 percent, which agrees with previous estimates, while a supply rate of 0.1 km^3/yr would require 1.21 percent, which may be unrealistic (Gerlach and others, 2002). The higher supply rate calculated by Cayol and others (2000) and Wright and Klein (2014) does not invalidate the previous estimates (summarized in table 1) but, instead, adds a new dimension to the discussion—is magma supplied to, and stored within, Kīlauea's deep rift zones? We address this question below in the sections entitled "Magma Supply Dynamics Between Kīlauea and Mauna Loa" and "Deep Rift Zones."

Table 1. Estimates of the magma supply rate during various times in Kīlauea's recent history. Shaded estimates include the volume created by deep rift opening. Table does not include estimates by Wright and Klein (2014), which span all time periods.

Time period	Supply (km^3/yr)	Method	Reference
1918–1979	0.08	Ratio between repose times and erupted volumes	Klein, 1982
1919–1990	0.09	Effusion rate of several sustained eruptions	Dvorak and Dzurisin, 1993
1952–1971	0.11	Effusion rate of three sustained eruptions	Swanson, 1972
1956–1983	0.09	Average summit and rift deformation and eruption volumes	Dzurisin and others, 1984
1959–1990	0.06	Average based on deformation and eruption volumes	Dvorak and Dzurisin, 1993
1960–1967	0.02–0.18	Deformation-inferred refilling of summit reservoir	Dvorak and Dzurisin, 1993
1961–1970	0.18	Eruption rate and deformation volumes	Cayol and others, 2000
1966–1970	0.07	Deformation and eruption volumes	Dvorak and others, 1983
1967–1975	0.05–0.18	Deformation and eruption volumes	Wright and Klein, 2008
1971–1972	0.08	Deformation and erupted volumes	Duffield and others, 1982
1975–1977	0.07–0.16	Microgravity and deformation	Dzurisin and others, 1980
1976–1982	0.19	Eruption rate and deformation volumes	Cayol and others, 2000
1983–1984	0.12	First 20 episodes of Puʻu ʻŌʻō eruption	Wolfe and others, 1987
1983–1991	0.14–0.18	Eruption rate and deformation volumes	Cayol and others, 2000
1983–2002	0.12	Puʻu ʻŌʻō eruption volumes	Heliker and Mattox, 2003
1983–2002	0.13	SO_2 emissions from ERZ	Sutton and others, 2003
1991	0.08	Deformation and effusion rates	Denlinger, 1997

Only sparse attempts have been made to document magma supply to Mauna Loa. This lack of attention is unsurprising, given the paucity of eruptive activity since 1950 (compared to Kīlauea) and the dearth of deformation data (in time and space). Lipman (1995) calculated a magma supply rate for Mauna Loa of 0.028 km^3/yr over the past 4 k.y. by summing the eruption rate over that time period (0.02 km^3/yr)—an estimate of the volume added by shallow dike emplacement in the rift zones (0.004 km^3/yr) and an estimate of magma accumulation at depth (0.004 km^3/yr)—and by assuming that cumulative summit inflation over the time period was negligible (based on analogy with Kīlauea). Such a supply rate, too small to account for the size attained by Mauna Loa (80,000 km^3) over its lifetime, led Lipman (1995) to speculate that magma supply rates must have been higher in the past—perhaps similar to the 0.1 km^3/yr rate that has characterized Kīlauea over the last several decades—and decayed as Mauna Loa was carried away from the hot spot locus by Pacific Plate motion.

2003–07 Increase in Magma Supply to Kīlauea

Deformation, seismic, gas, and geologic data indicate an increase in the rate of magma supply from the mantle to Kīlauea during 2003–07. In late 2003, deformation at Kīlauea's summit (fig. 3) transitioned from long-term deflation to inflation, as indicated by the cessation of subsidence at GPS station AHUP and increasing distance between GPS stations AHUP and UWEV (fig. 4*A*). SO_2 emission rates from Kīlauea's ERZ remained nearly constant at about 1,370 metric tons per day (t/d) during this period (fig. 4*B*), however, indicating no change in the amount of magma being transported from the summit to the ERZ—about 0.13 km^3/yr (Sutton and others, 2003). The inflation was therefore not caused by a decrease in effusion rate and subsequent backup in Kīlauea's magma plumbing system, as has occurred at other times during the Puʻu ʻŌʻō eruption (Kauahikaua and others, 1996; Miklius and Cervelli, 2003).

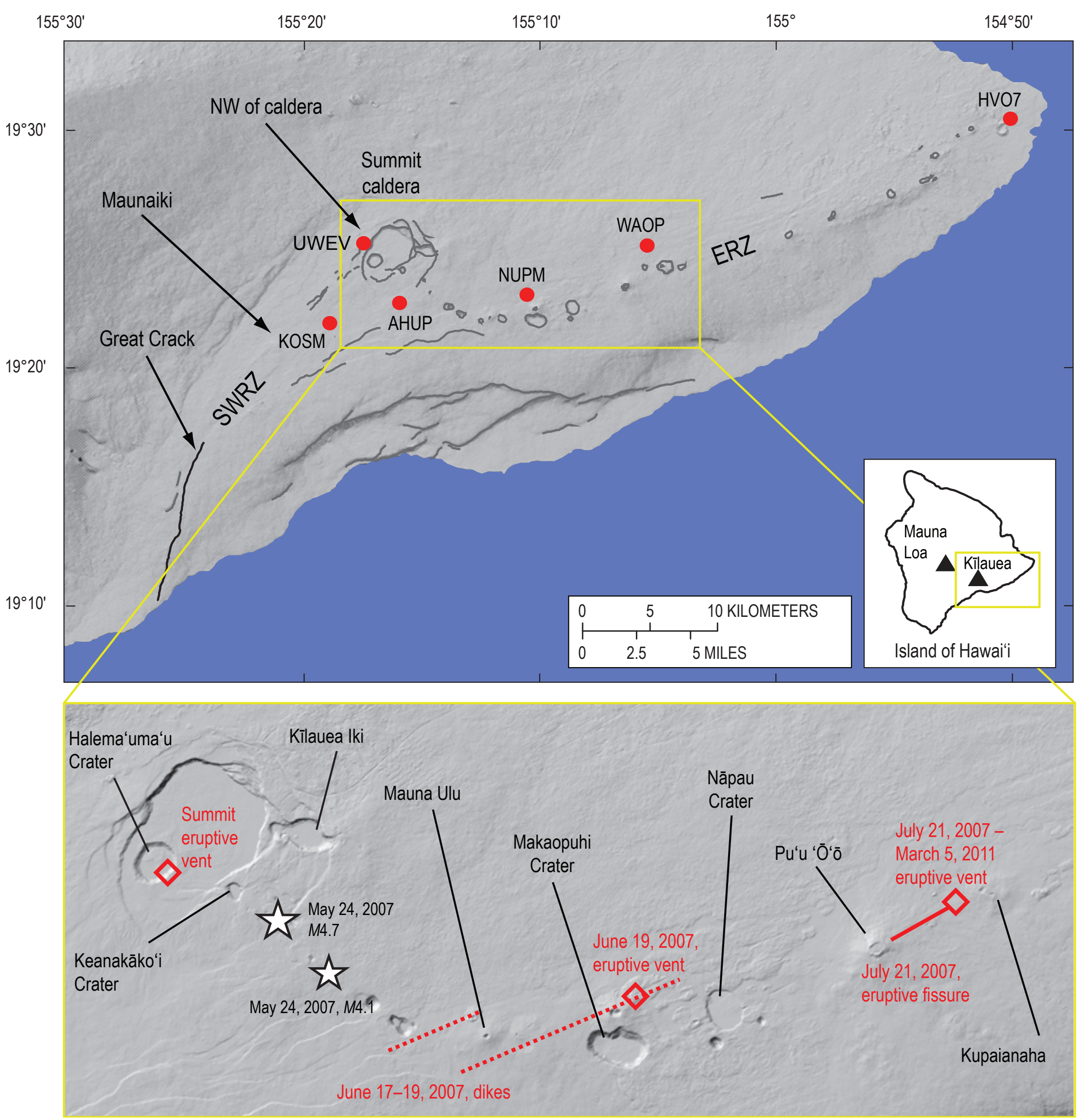


Figure 3. Shaded topographic map of Kīlauea Volcano. Labeled red dots give the locations and site names of continuous Global Positioning System stations mentioned in the text and other figures. ERZ, East Rift Zone; SWRZ, Southwest Rift Zone. Bottom image shows detail of Kīlauea's summit and ERZ, with major features discussed in the text labeled. Red diamonds give the locations of eruptive vents discussed in the text. White stars are earthquakes that occurred on May 24, 2007 (Wauthier and others, 2013). Dashed red lines give the geometry of a pair of en echelon dike segments that were emplaced during June 17–19, 2007 (from Montgomery-Brown and others, 2010). Solid red line shows location of the July 21, 2007, eruptive fissure.

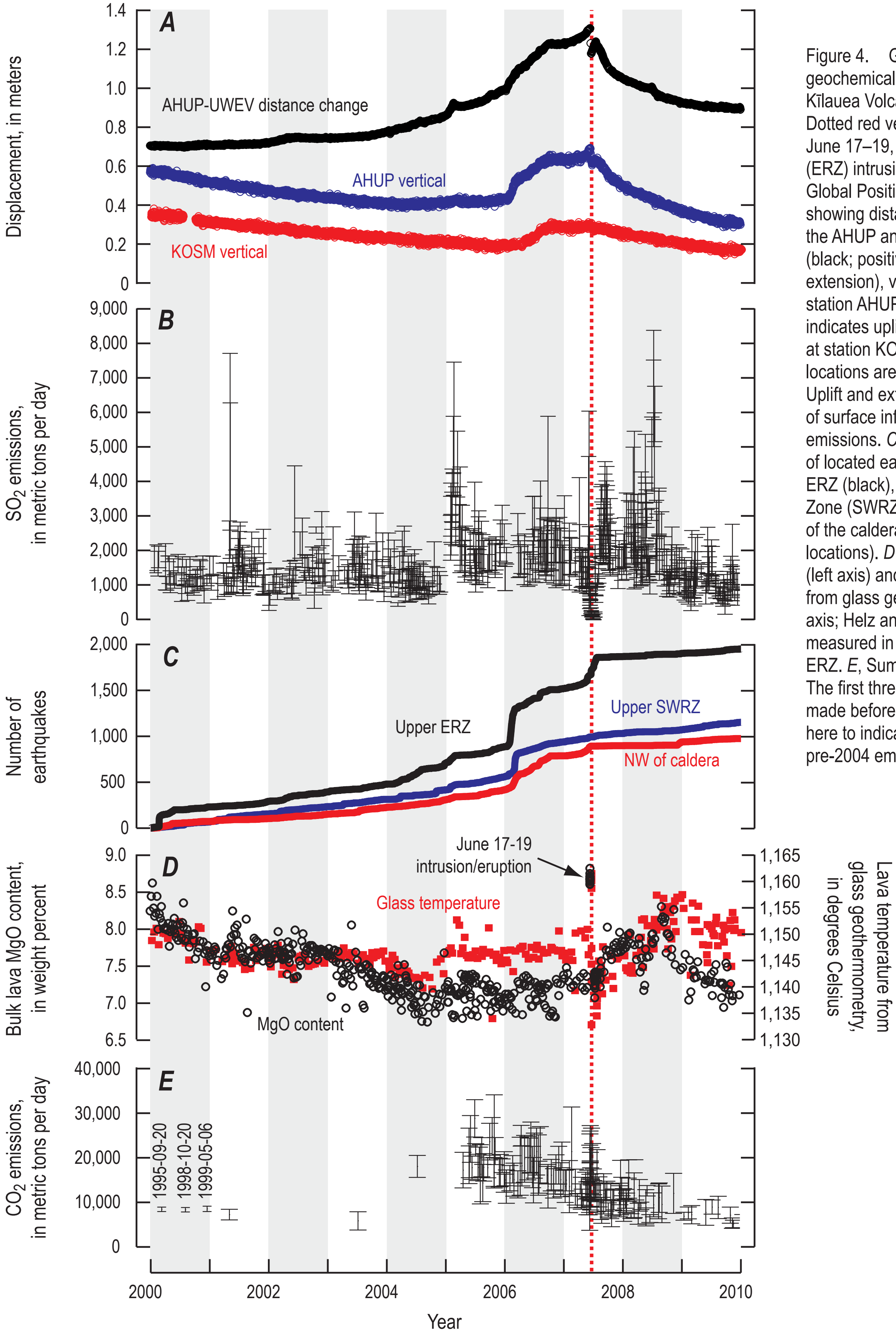


Figure 4. Geophysical and geochemical time-series data from Kīlauea Volcano during 2000–10. Dotted red vertical line indicates June 17–19, 2007, East Rift Zone (ERZ) intrusion and eruption. *A*, Global Positioning System data showing distance change between the AHUP and UWEV stations (black; positive change indicates extension), vertical change at station AHUP (blue; positive change indicates uplift), and vertical change at station KOSM (red). Station locations are given in figure 3. Uplift and extension are indicative of surface inflation. *B*, ERZ SO_2 emissions. *C*, Cumulative numbers of located earthquakes in the upper ERZ (black), upper Southwest Rift Zone (SWRZ; blue), and northwest of the caldera (red) (see fig. 3 for locations). *D*, MgO weight percent (left axis) and eruption temperatures from glass geothermometry (right axis; Helz and Thornber, 1987), measured in lavas erupted from the ERZ. *E*, Summit CO_2 emissions. The first three measurements were made before 2000 but are shown here to indicate the consistency of pre-2004 emission rates.

Since there was no precursory deflation event that would have created a pressure imbalance and a “top-down” change in magma supply, the most likely explanation for the inflation is increased magma flux from the hot spot. In early 2005, a pulse of magma from the summit to the ERZ resulted in both temporarily heightened rift SO_2 emissions (fig. 4*B*), increased lava effusion from the Puʻu ʻŌʻō eruptive vent, and a several-week period of summit deflation as magma drained from the summit to feed the East Rift Zone effusive surge (fig. 4*A*). By mid-2005, summit inflation had resumed (fig. 4*A*), and rift SO_2 emissions had increased to a new average rate of ~1,900 t/d (fig. 4*B*), implying that about 0.18 km^3/yr of magma was then being delivered from the summit to the ERZ.

The summit continued to inflate, following the 2005 surge in ERZ effusion, indicating that the ERZ conduit could not accommodate the elevated amount of magma being supplied to Kīlauea and that magma was accumulating beneath the summit. The rate of summit inflation increased rapidly in early 2006 (fig. 4*A*), and by the middle of that year, uplift was also occurring beneath the Southwest Rift Zone (which had been subsiding since 1982), suggesting magma accumulation beneath that part of the volcano, as well (fig. 4*A*, station KOSM). Heightened stress in the summit region caused by the inflation was indicated by swarms of short-period earthquakes in the upper parts of the East and Southwest Rift Zones, as well as northwest of the caldera (fig. 4*C*).

Modeled volume changes suggest that at least 0.01 km^3 of magma accumulated beneath the summit and Southwest Rift Zone in 2006 (Poland and others, 2012) which, when combined with ERZ effusion, implies a minimum supply rate of 0.19 km^3/yr to Kīlauea’s shallow magma system during that year. In addition, GPS data reveal that the normal pattern of ERZ subsidence stopped or switched to uplift at several places during 2003–07 (fig. 5), indicating magma storage in the ERZ and that the magma supply rate to Kīlauea’s shallow magmatic system must have been even higher. No change in the rate of deep opening of the ERZ was detected during 2003–07, as suggested by steady motion of Kīlauea’s south flank during this time period (fig. 6), so the increased magma supply apparently affected only Kīlauea’s shallow magma plumbing system.

Petrologic changes to lava erupted from the ERZ also reflect the magma supply increase. Lava erupted from Puʻu ʻŌʻō is a hybrid between a high-MgO, high-temperature magma supplied from the hot spot and a partially degassed, lower-MgO, lower-temperature resident magma that has partially crystallized due to storage at shallow levels within the volcano (Thornber, 2003; Thornber and others, 2003). MgO values and lava temperature had been declining since 2000 (fig. 4*D*), reflecting an increase in the proportion of lower-temperature magma being erupted. In addition, the mineralogical composition of the lava included a nonequilibrium assemblage of high-temperature olivine plus low-temperature clinopyroxene, olivine, and plagioclase (Thornber and others, 2010). MgO content stabilized at

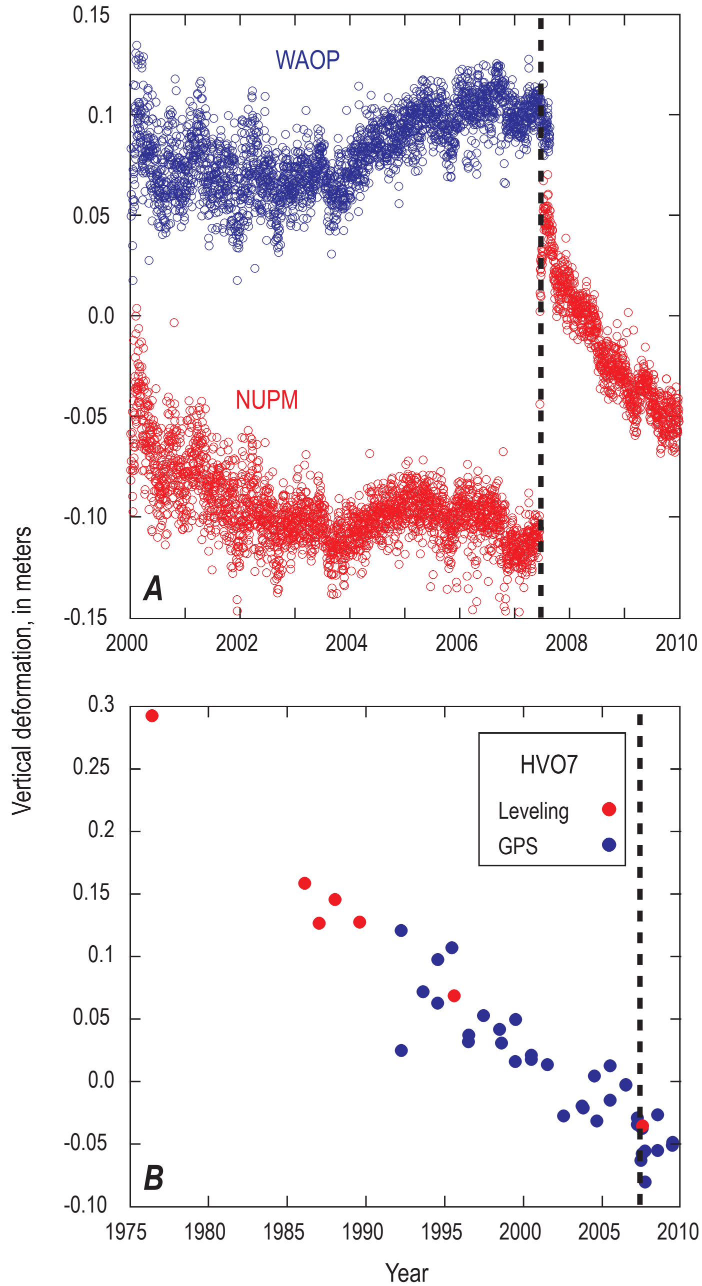


Figure 5. Vertical deformation at sites along Kīlauea’s East Rift Zone. Station locations are given in figure 3. Dashed lines indicate June 17–19, 2007, intrusion and eruption. *A*, Plot of elevation change at Global Positioning System (GPS) stations NUPM and WAOP. Both sites show a transition from subsidence to uplift in late 2003 at the same time that summit inflation commenced. No data are available from WAOP after August 24, 2007, because the station was overrun by advancing lava. *B*, Time series of vertical elevation change at site HVO7 from leveling (red circles) and campaign GPS (blue circles). The trend of long-term subsidence at the site was interrupted during 2003–07, coincident with the period of increased magma supply. The station is located 6.5 km from a geothermal power plant, which is distant enough to be free from deformation associated with geothermal production.

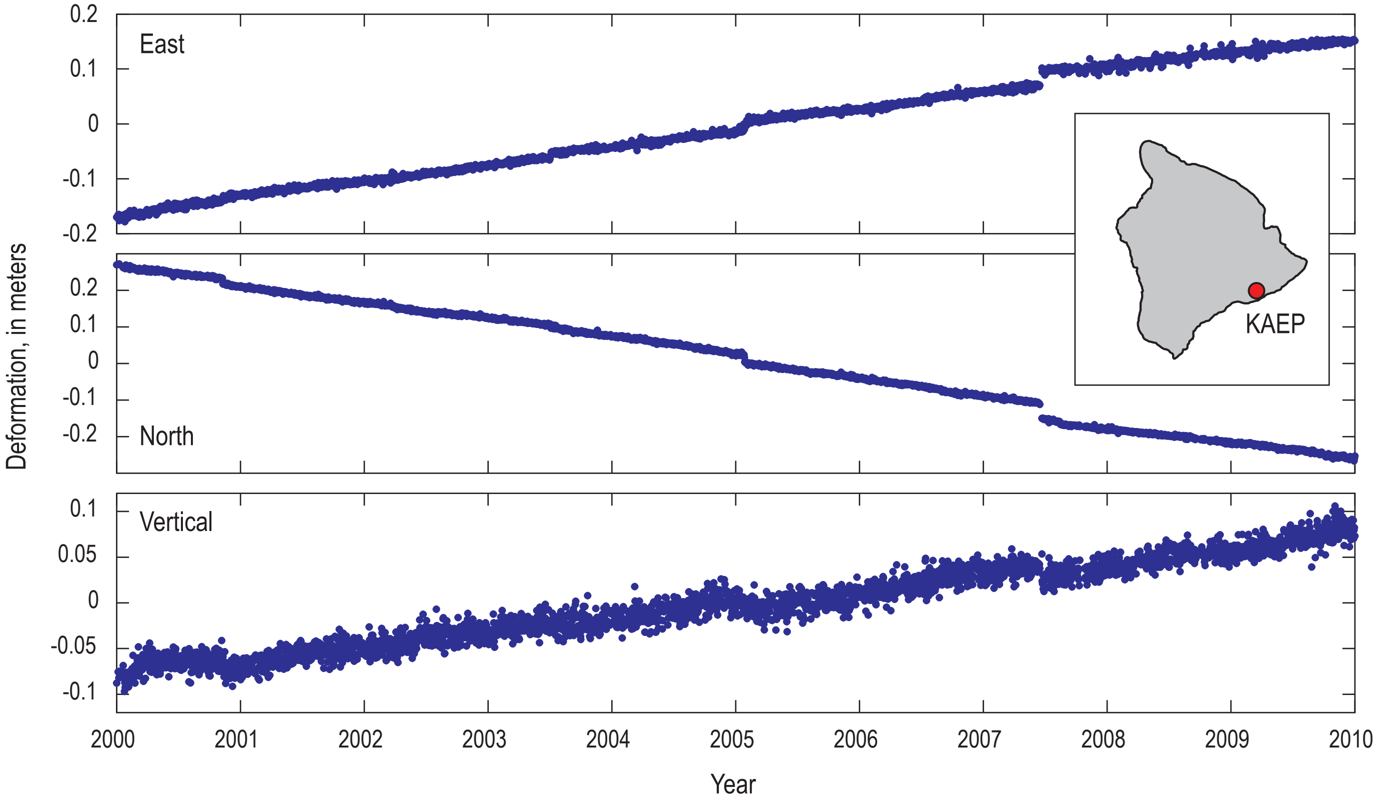


Figure 6. Plots showing east (top), north (middle), and vertical (bottom) components of deformation at Global Positioning System station KAEP, located on the south flank of Kīlauea Volcano (location given in inset). Linear deformation suggests no change in the opening rate of the deep rift zone during 2000–10. Small offsets in the time series are related to aseismic slip events on the south flank (Brooks and others, 2006; Montgomery-Brown and others, 2009).

~7.0 percent in early 2004 and, along with lava temperature, remained steady until mid-2007 (fig. 4*D*)—a sign of shallow mixing between hotter recharge magma and shallow resident magma, suggesting that cooler parts of the magma storage and transport pathways in Kīlauea's summit and ERZ were being stirred by an influx of new magma. Lava erupted on June 19, 2007, near Makaopuhi Crater (see section entitled "Consequences of Changes in Supply"), was the hottest and most primitive since 1998, representing the high-MgO, high-temperature magma component. After that eruption, MgO content and temperature increased steadily in lava erupted from Puʻu ʻŌʻō until mid-2008 (fig. 4*D*), suggesting that heightened magma supply introduced more of the high-temperature magmatic component to Kīlauea.

CO_2 emissions from Kīlauea are an independent indication of deep magma supply, because gas begins to exsolve from ascending magma at a depth of ~30 km and is therefore mostly insensitive to shallow processes (Gerlach, 1986; Gerlach and others, 2002). Before 2005, sporadic CO_2 emission measurements yielded constant values of about 8,000 t/d, but a measurement in July 2004 was 18,200±2,500 t/d (fig. 4*E*), and more frequent measurements, starting in 2005, averaged about 20,000 t/d for several months. The increase in CO_2 emissions began sometime between mid-2003 and mid-2004—the same period of time that summit inflation commenced. Interestingly, the CO_2 emission rate did not track summit inflation but, instead, reached a maximum more than a year before the 2005 increase in lava effusion and three years prior to the highest rates of summit inflation, emphasizing the importance of CO_2 monitoring for forecasting changes in volcano behavior.

By the end of 2008, geological, geophysical, and geochemical data indicated a return to pre-2003 rates of magma supply. CO_2 emission rates, which began to decline in 2006, reached pre-2003 levels (fig. 4*E*), deflation dominated the summit after formation of a new ERZ vent on July 21, 2007 (fig. 4*A*), and SO_2 emissions from the ERZ had decayed towards pre-2003 rates by late 2008 (fig. 4*B*), indicating a decrease in the quantity of magma being transported from the summit to the ERZ eruption site. Summit seismicity returned to background levels following the intrusion and eruption of June 2007 (fig. 4*C*), although tremor levels increased in the months prior to the start of the 2008 summit eruption (Poland and others, 2009). Finally, the MgO content of lavas erupted from the ERZ began to decline in mid-2008 (fig. 4*D*), indicating a decrease in the proportion of newly supplied magma.

The long-term Puʻu ʻŌʻō eruption on Kīlauea's ERZ, which started in 1983 and continued unabated through 2003–07, suggests that the increased supply was not caused by pressure fluctuations in the shallow summit magma system, as has been proposed for previous changes in supply rate

(Dvorak and Dzurisin, 1993). Instead, the increase must have been driven by a greater flux of magma from the mantle, which is also supported by heightened CO_2 emissions. This documentation of a short-term change in hot-spot activity demonstrates that supply of magma from the mantle to Hawaiian volcanoes can vary on time scales of years.

Consequences of Changes in Supply

Variations in magma supply to Kīlauea and Mauna Loa have been associated with major changes in the style of activity at the two volcanoes, as well as compositional variations in erupted lavas (Garcia and others, 1996, 2003). For example, Dzurisin and others (1984) calculated high supply rates to Kīlauea in the years before and during the large East Rift Zone eruptions that occurred in 1960 (Kapoho) and 1969–74 (Mauna Ulu). Dvorak and Dzurisin (1993) proposed that magma supply rates to Kīlauea were low during the 1800s and early 1900s but higher starting in the 1950s, implying that summit eruptive activity was favored during periods of low supply, and that ERZ activity and sustained rift eruptions were a consequence of high supply—similar to the model of Wright and Klein (2014) for magma supply during the 20th and early 21st centuries. Geochemical variations in lava flows erupted during the early 19th to middle 20th centuries suggest decreased partial melting of the source region, also indicative of a lower magma supply rate (Pietruszka and Garcia, 1999a). In contrast, modeled heat loss from the summit lava lake prior to 1924 argues for higher supply (Francis and others, 1993), exemplifying the ambiguity in magma supply estimates.

Changes in volcanic activity as a consequence of increased magma supply were best documented following the 2003–07 magma supply surge to Kīlauea, described above and by Poland and others (2012). Rapid summit inflation due to the supply surge caused an increase in stress on the caldera-bounding normal faults and resulted in a pair of *M*4+ earthquakes on the southeast margin of the caldera on May 24, 2007 (fig. 3; Wauthier and others, 2013). Continued accumulation of pressure at the summit drove an intrusion from the summit into the ERZ during June 17–19, 2007, which resulted in a small eruption of high-MgO lava (fig. 4*D*) just north of Makaopuhi Crater (fig. 3) on June 19 (Poland and others, 2008; Fee and others, 2011). Increased compressive stress on Kīlauea's south flank due to ERZ opening during June 17–19 apparently triggered an aseismic slip event on the basal detachment fault that underlies Kīlauea shortly after the onset of the intrusion (Brooks and others, 2008; Montgomery-Brown and others, 2010, 2011). Eruptive activity resumed in Puʻu ʻŌʻō crater on July 1, but pressure buildup beneath the vent caused a fissure eruption on the east flank of the cone on July 21, leading to the formation of a new long-term eruptive vent about 2 km downrift of Puʻu ʻŌʻō (fig. 3; Poland and others, 2008). The formation of an eruptive vent at Kīlauea's summit (fig. 3) on March 19, 2008 (Wilson and others, 2008; Houghton and others, 2011; Patrick and others, 2011), has also been attributed indirectly to the magma supply increase. Poland and others (2009) suggested that decompression of the inflated summit due to formation of the July 21, 2007, East Rift Zone eruptive vent caused exsolution of volatiles, which ascended through existing fractures and reached the surface along the southeast margin of Halemaʻumaʻu Crater. Increasing volatile pressure dilated the pathways and caused the March 19 explosion that formed the new eruptive vent, allowing magma to passively rise toward the surface without breaking rock and without causing earthquakes or deformation.

Variable magma supply has also been cited as a cause for changes in eruptive activity at Mauna Loa and even the structural evolution of that volcano. The eruption rate between 1843 (the first recorded eruption of Mauna Loa) and 1877 was more than twice that of the post-1877 rate (Lockwood and Lipman, 1987; Lipman, 1995). Decreasing incompatible-element abundances in erupted lavas during 1843–77, coupled with the heightened eruption rate, might be a sign of increased magma supply (Rhodes and Hart, 1995). Riker and others (2009) proposed that an eruption of MgO-rich lava in 1859 from a radial vent on the northwest flank of the volcano (outside a rift zone) was a consequence of high magma supply based on the primitive composition and vent location.

Deformation and gravity data spanning Mauna Loa's 1984 eruption indicate that the volume of extruded lava was much greater than the volume removed from the summit magma reservoir. This discrepancy, coupled with the observation that CO_2 emissions were at their highest during the later part of the eruption, prompted Johnson (1995a) to suggest that magma supply to Mauna Loa is episodic, with most influx occurring concurrently with eruptions—akin to the "top-down" model of Dvorak and Dzurisin (1993) for Kīlauea. Unfortunately, testing these hypotheses will be difficult without more frequent eruptive activity at Mauna Loa than characterized the latter half of the 20th century and the start of the 21st century.

Magma Supply Dynamics Between Kīlauea and Mauna Loa

The possibility of a connection between the magmatic systems of Kīlauea and Mauna Loa has been a source of debate since the volcanoes were first described scientifically in the mid-1800s (see discussion in Stearns and Macdonald, 1946, p. 132–135). In general, petrologic differences in lava erupted from the two volcanoes provide convincing evidence that their magma sources are geochemically distinct and their magma plumbing systems are independent (for example, Wright, 1971; Rhodes and others, 1989; Frey and Rhodes, 1993). Mauna Loa-like magmas have erupted from Kīlauea (Rhodes and others, 1989), however, and lavas erupted at Kīlauea during 1998–2003 showed signs of an increasing proportion of a Mauna Loa component (Marske and others, 2008). In addition, small-scale compositional heterogeneities have been seen to affect both volcanoes at roughly the same time, suggesting that the source regions for the two volcanoes are not widely separated (Marske and others, 2007). In fact, seismic evidence supports the possibility of an interconnected melt zone between the two volcanoes below about 30-km depth (Ellsworth and Koyanagi, 1977).

Observations of eruptive activity at Kīlauea and Mauna Loa highlight correlations that argue for some sort of shared magma supply, with several authors noting an inverse relation between their eruptions (for example, Moore, 1970; Klein, 1982). During 1934–52, for instance, Kīlauea was dormant but Mauna Loa erupted six times, while during 1952–2014, Mauna Loa erupted only twice and Kīlauea was frequently (and often continuously) active. Klein (1982) found this anti-correlation to be statistically significant and suggested that the volcanoes were competing for a common supply of magma. Kīlauea and Mauna Loa have also behaved sympathetically. In May 2002, an effusive surge from Kīlauea's ERZ occurred at the same time as the onset of inflation at Mauna Loa. Miklius and Cervelli (2003) suggested that input of magma into Mauna Loa increased the pressure in Kīlauea's magma system, triggering the ERZ effusive episode.

A more direct case of sympathetic behavior between the two volcanoes is suggested by activity spanning 2002–07. Mauna Loa began to inflate in 2002 after nearly a decade of deflation (Miklius and Cervelli, 2003; Amelung and others, 2007). The inflation rate increased in late 2004 and was accompanied by a swarm of thousands of long-period earthquakes at >30-km depth (Okubo and Wolfe, 2008), but inflation waned and ceased by the end of 2009 (fig. 7). The fact that Mauna Loa's inflation occurred at approximately the same time as the 2003–07 surge in magma supply to Kīlauea is an unlikely coincidence and suggests the possibility that an increase in magma supplied from the mantle affected both volcanoes. Dzurisin and others (1984) proposed that a similar relation existed in the late 1970s, when their model suggested a period of increasing magma supply to Kīlauea during an episode of inflation at Mauna Loa. Similarly, a simultaneous change in the composition of erupted lavas at both volcanoes occurred during 250–1400 C.E. (Marske and others, 2007), indicating that Kīlauea and Mauna Loa have previously been affected by mantle source processes at the same time. Gonnermann and others (2012) argued that the two volcanoes may be dynamically linked through an asthenospheric porous melt zone located tens of kilometers deep. In their model, shallow magma storage at each volcano is distinct, and the magma feeding systems tap different parts of the mantle source (thereby explaining the overall petrologic differences in erupted lavas). Changes in magma pressure, however, can be transmitted between crustal storage reservoirs through the asthenospheric melt zone on time scales of less than a year without requiring direct melt transport between volcanoes. The model therefore provides a mechanism by which Kīlauea and Mauna Loa display complementary modes of behavior without requiring a shallow connection.

An asthenospheric melt zone that links Mauna Loa and Kīlauea might also explain observations of CO_2 emissions at the two volcanoes. As discussed above, Gerlach and others (2002) favored the 0.18 km^3/yr supply rate of Cayol and others (2000), because lower rates imply that the CO_2 content of the primary magma supplied to Kīlauea is unrealistically high (1.21 percent versus 0.70 percent). This higher supply rate is based on the assumption that nearly half of the magma fed to Kīlauea is stored in the deep ERZ, which is shallow enough that CO_2 would degas, but deeper than the exsolution pressures of other volatile species. If no magma is stored in this region (see "Deep Rift Zones" section below), however, how can the lower supply rate be reconciled with the favored CO_2 content of the magma?

A possible explanation is that CO_2 from most magma supplied by the hot spot degasses through Kīlauea's summit. Deep seismicity (primarily tremor and long-period earthquakes) at ~40-km depth offshore of Kīlauea's south flank has been interpreted as the magma source that feeds the active volcanoes of the Island of Hawai‘i (Aki and Koyanagi, 1981; Wright and Klein, 2006). Wright and Klein (2006) further proposed that the deep feeder is linked to Kīlauea via a subhorizontal zone of magma transport at about 30-km depth, similar to the asthenospheric melt zone envisioned by Gonnermann and others (2012). CO_2 starts to exsolve at about this depth (Gerlach and others, 2002), so CO_2 bubbles might ascend along the path closest to the deep conduit from the source—the nearby conduit to Kīlauea's summit—regardless of the ultimate destination of the magma, be it Kīlauea, Mauna Loa, or possibly Lō‘ihi. Such a model (fig. 8) has the potential to explain not only the CO_2 content discrepancy raised by Gerlach and others (2002) at Kīlauea, but also why so little CO_2 has been emitted from Mauna Loa despite periods of magma accumulation and eruption (Ryan, 1995). The recognition that Mauna Loa magmas have infiltrated Kīlauea's magma plumbing system (Rhodes and others, 1989) lends additional support to this hypothesis. This proposal is speculative but provides an interesting alternative to requiring higher average rates of magma supply to Kīlauea (~0.18 km^3/yr, based on CO_2 emissions, versus ~0.1 km^3/yr from deformation and effusion-rate data) and storing that excess magma in Kīlauea's deep rift zones.

Summary

Volcanism from the Hawaiian hot spot is driven by upwelling of high-temperature material, probably originating deep within Earth's mantle. The rate of magma supplied by the plume over millions of years has varied by an order of magnitude but has been increasing since about 30 Ma and is currently high, with the contemporary rate of magma supply to Kīlauea's shallow magmatic system best approximated by the eruption rate of long-term effusions—at least 0.1 km^3/yr. Historical changes in the rate of magma supply have been driven both by pressure fluctuations in shallow crustal reservoirs ("top-down") and variations in the volume of magma supplied from the mantle ("bottom-up"). Bottom-up control of magma supply affects both volcanoes, as demonstrated by 2002–07 activity at Kīlauea and Mauna Loa. Supply also varies between volcanoes, even when

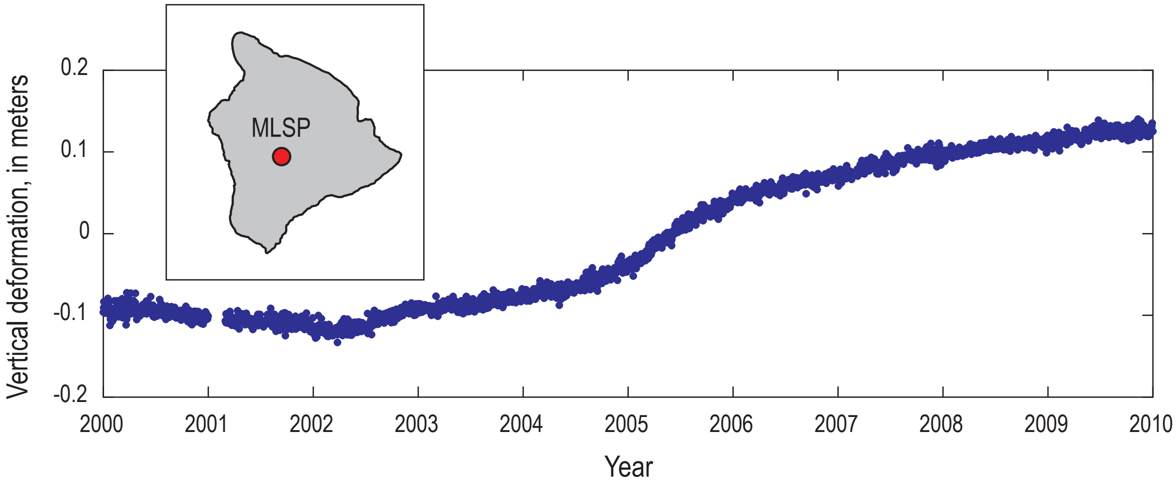


Figure 7. Plot of vertical elevation change at Global Positioning Station station MLSP, located on the south side of Mauna Loa's summit caldera (location given in inset). Uplift and inflation began in 2002, accelerated in 2004–05, and gradually waned to no deformation by the end of 2009 (Miklius and Cervelli, 2003; Amelung and others, 2007).

magma ascends from the hot spot at a steady rate, causing periods of few eruptions at one volcano and vigorous activity at the other. For example, since 1952 and through 2014, Mauna Loa erupted only twice (in 1975 and 1984), while Kīlauea erupted quasi-continuously. Short-term (days to weeks) fluctuations in magma supply are superimposed on the overall rate and are "top down" processes that are driven, for instance, by pressure in a shallow magma reservoir that is low enough to promote magma ascent from depth. Kīlauea and Mauna Loa share magma supply from the hot spot, which explains observations of both inverse patterns of eruptive activity and sympathetic patterns of magma accumulation. Differences in mantle source composition for Kīlauea and Mauna Loa reflect compositional heterogeneity in the asthenospheric melt region that links the two volcanoes and through which pressure is transmitted without requiring direct melt transport. A model of shared magma supply may also explain why CO_2 emissions from Kīlauea imply a higher rate of supply than is suggested by long-term eruptive activity. CO_2 from all hot-spot magma may degas through Kīlauea's summit, regardless of whether that magma ultimately feeds the shallow systems of Kīlauea, Mauna Loa, or even Lō‘ihi.

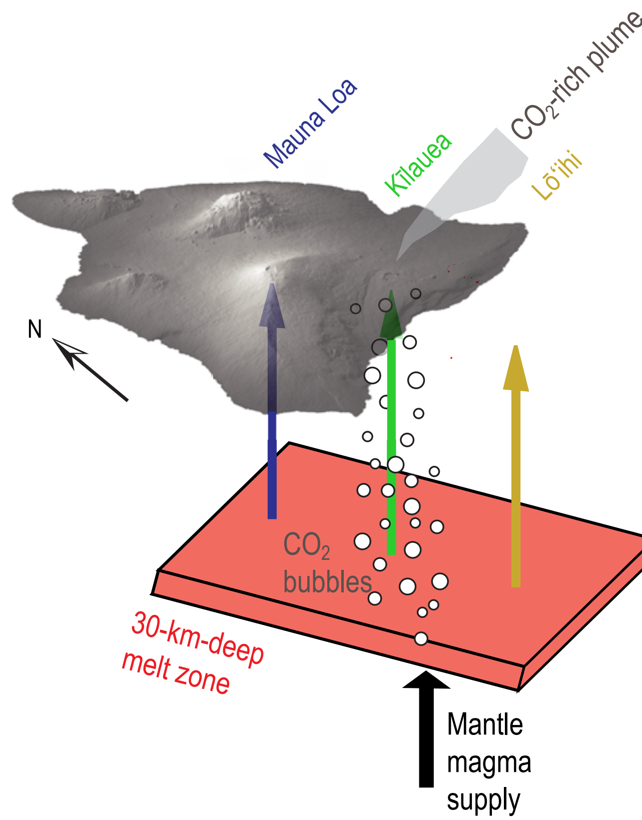


Figure 8. Schematic diagram of possible magma supply pathways beneath the Island of Hawai‘i, based in part on figure 2 of Gonnermann and others (2012). Black vertical arrow represents magma supplied from the mantle hot spot to a subhorizontal melt zone at about 30-km depth (red plane), located beneath Kīlauea's Southwest Rift Zone, based on long-period seismicity and tremor. Colored arrows depict magma transport paths to the most active volcanoes of Hawai‘i. Most exsolved CO_2 might ascend with magma that is fed to Kīlauea, since that volcano is closest to the source of mantle supply, resulting in a CO_2-rich plume from Kīlauea's summit.

Magma Storage

Before erupting at the surface, magma is often (although not always) stored in subsurface reservoirs, where chemical differentiation produces a variety of compositional products. Petrologists and structural geologists have spent considerable effort attempting to understand the formation of such reservoirs (for example, Glazner and Bartley, 2006) and their physical and chemical evolution (for example, Marsh, 1989).

The identification of active magma storage areas is most often accomplished by geophysical studies, especially seismology and geodesy (Dvorak and Dzurisin, 1997). For instance, earthquake hypocenters in the years following the May 18, 1980, eruption of Mount St. Helens, Washington, defined a vertically elongate aseismic zone between depths of about 7 and 13 km, interpreted to delineate a magma body (Scandone and Malone, 1985). Models based on 2004–08 co-eruptive GPS displacements favored magma storage in the same area (Lisowski and others, 2008). Surface inflation near South Sister volcano, Oregon, detected by Interferometric Synthetic Aperture Radar (InSAR) and corroborated by GPS and leveling, suggested magma accumulation at a depth of approximately 5 km located about 5 km west of the volcano's summit (Dzurisin and others, 2009). Continuous GPS data from Mount Etna, Italy, have been used to map a multilevel magma plumbing system, with magma storage areas at 6.5, 2.0, and 0.0 km below sea level (Aloisi and others, 2011).

Magma reservoir location and geometry are best determined when a variety of data are available. Deformation, seismicity, and geology indicate the presence of at least three regions of magma storage beneath Piton de la Fournaise volcano, Réunion Island, at about 2.3 km, 7.5 km, and 15 km beneath the surface (Peltier and others, 2009). Physics-based models incorporating diverse datasets offer hope for constraining not just reservoir depths and locations but also such elusive parameters as magma reservoir volume, overpressure, and volatile content (Anderson and Segall, 2013).

Mapped regions of magma accumulation and storage span a range of geometries, depths, and sizes. Reservoir geometry is generally depicted as spherical or ellipsoidal, primarily because such shapes are simple to visualize and model (for example, Davis, 1986; Delaney and McTigue, 1994; Ohminato and others, 1998; Yang and others, 1992) and are favored by thermal and mechanical considerations (Gudmundsson, 2012). High-spatial-resolution deformation data sometimes suggest more complex geometries (for example, Masterlark and Lu, 2004), and magma reservoirs have been described as amalgamations of small bodies separated by screens of solid rock or semisolid magma mush (for example, Fiske and Kinoshita, 1969; Gudmundsson, 2012). Kühn and Dahm (2008) pointed out that the stress field induced by a plexus of dikes and sills would resemble that of a simpler ellipsoidal source. Indeed, the very definition of a magma chamber is vague, as magma storage areas are not homogenous but are, instead, mush zones of melt, crystals, and exsolved volatiles that will have different behaviors, effective sizes, and shapes over different time scales and strain rates (for example, Johnson, 1992; Gudmundsson, 2012).

Depths of magma accumulation vary widely among volcanoes. Magma accumulation can occur tens of kilometers beneath the surface, as is apparently the case at Hekla, Iceland (Ofeigsson and others, 2011), or only 1–2-km deep, as modeled at Kīlauea (Poland and other, 2009; Anderson and others, in press). At many volcanoes, like Shishaldin, Alaska, no magma chamber has been detected in spite of frequent eruptive activity (Moran and others, 2006), while other volcanoes have a series of vertically stacked reservoirs, like Soufrière Hills volcano in Montserrat (for example, Voight and others, 2010), Piton de la Fournaise (Peltier and others, 2009), and Etna (Aloisi and others, 2011). The depth of magma accumulation depends on the density contrast between magma and country rock, as well as on the presence of structural discontinuities that may inhibit upward magma flow (Ryan, 1987; Burov and others, 2003; Peltier and others, 2009). Magma chamber volumes can exceed 100 km^3, as indicated by the volume of ignimbrite sheets that represent single eruptive events (for example, Christiansen, 2001). The largest eruptions aside, magma chambers are more generally in the range of tens of cubic kilometers (for example, Mastin and others, 2009; Paulatto and others, 2012) to about 1 km^3 (for example, Poland and others, 2009).

The magma plumbing systems of Kīlauea and Mauna Loa are well known, compared with most other volcanoes on Earth. Magma pathways in Hawaiian volcanoes are established early in the shield stage (Clague and Sherrod, this volume, chap. 3) but evolve over time, along with the volcano, for example, as calderas form and fill (Swanson and others, 2012a) and rift zones migrate (Swanson and others 1976a). Some datasets, like magnetics and gravity, reflect the cumulative development of a volcano's magmatic system. We use constraints provided by such data in combination with monitoring results (especially deformation and seismic), geologic studies, and observations of eruptions collected during the 100+ years of HVO's existence to investigate the current magma plumbing configurations at Kīlauea and Mauna Loa. Below, we present refined models of the magmatic systems at both volcanoes.

Kīlauea

The general model for Kīlauea's magma plumbing system, first proposed by Eaton and Murata (1960) and refined by Tilling and Dvorak (1993), is simple: magma generated in the mantle ascends and is stored in reservoirs that are one to a few kilometers beneath the summit, from which it may eventually erupt within the caldera or be transported laterally into the East or Southwest Rift Zones as intrusions that may feed eruptions far from the summit (fig. 9*A*). While this overall depiction remains largely unchanged, the characteristics of specific parts of the magma plumbing system have been the focus of numerous studies. Ryan (1988) used seismic data to define areas of subsurface magma transport and storage, including magma storage at 2–4-km depth beneath the summit and almost wholly molten rift zones at 3–10 km beneath the surface (fig. 9*B*), although the conduit he proposed to 30-km depth beneath Kīlauea's summit was later interpreted to be a mantle fault zone (Wolfe and others, 2003). Other studies have employed deformation measurements to map the locations of magma reservoirs beneath the summit (for example, Cervelli and Miklius, 2003; Baker and Amelung, 2012; Anderson and others, in press), constrain the relation between summit and rift zone magma storage (for example, Dzurisin and others, 1980,

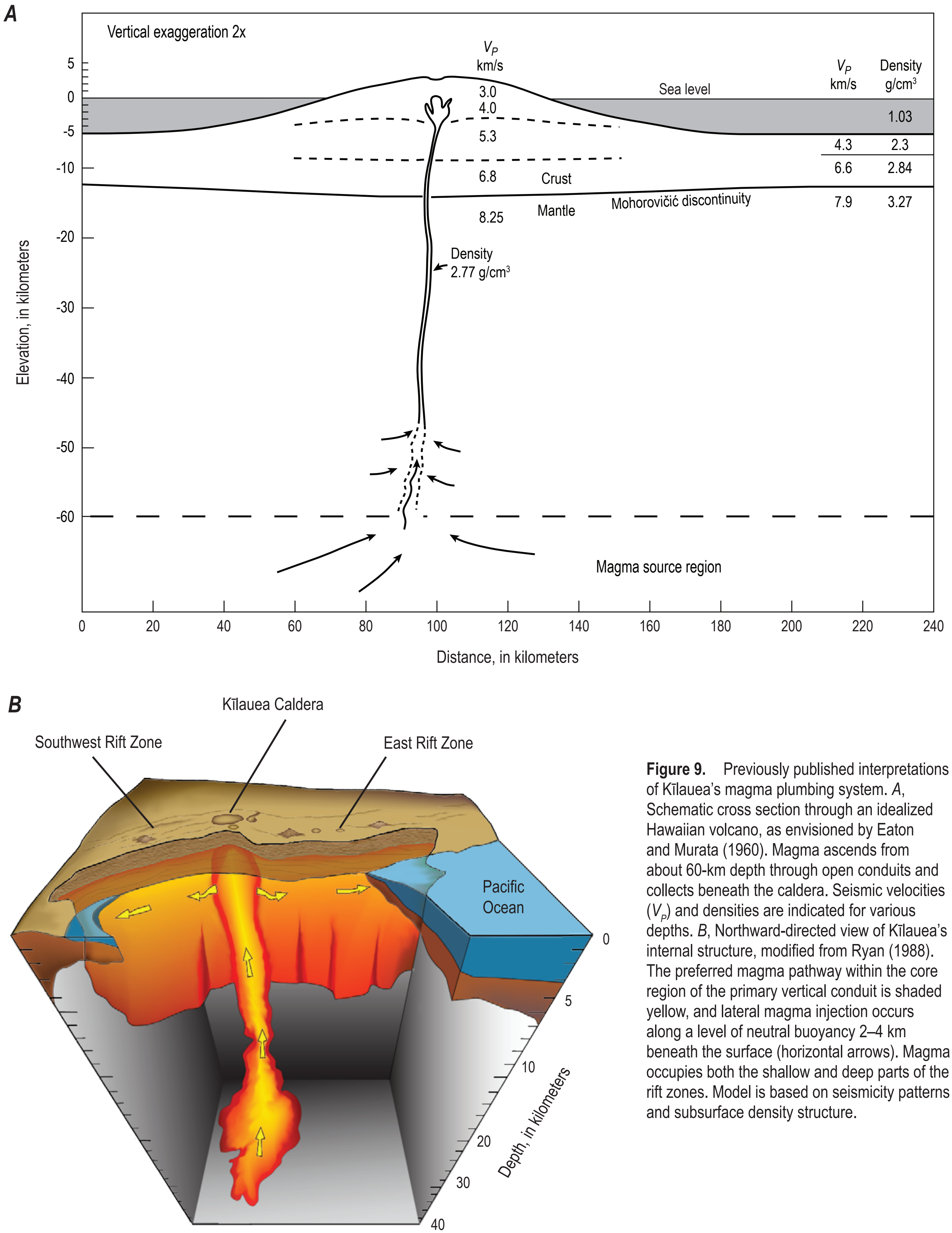


Figure 9. Previously published interpretations of Kīlauea's magma plumbing system. *A*, Schematic cross section through an idealized Hawaiian volcano, as envisioned by Eaton and Murata (1960). Magma ascends from about 60-km depth through open conduits and collects beneath the caldera. Seismic velocities (V_P) and densities are indicated for various depths. *B*, Northward-directed view of Kīlauea's internal structure, modified from Ryan (1988). The preferred magma pathway within the core region of the primary vertical conduit is shaded yellow, and lateral magma injection occurs along a level of neutral buoyancy 2–4 km beneath the surface (horizontal arrows). Magma occupies both the shallow and deep parts of the rift zones. Model is based on seismicity patterns and subsurface density structure.

1984; Johnson, 1995b), and propose the presence of magma in the deep rift zones (Delaney and others, 1990; Owen and others, 1995, 2000a; Cayol and others, 2000). Petrologic arguments have also been utilized to develop models of Kīlauea's magma plumbing system. Geochemical data from historical summit eruptions favor a simple magma storage zone of a few cubic kilometers at 2–4 km beneath the summit (Pietruszka and Garcia, 1999b; Garcia and others, 2003). Primitive compositions and deformed olivine phenocrysts derived from cumulate rocks have been interpreted as evidence that magma may bypass summit storage on its way to erupting on the ERZ (for example, Trusdell, 1991; Wright and Helz, 1996; Vinet and Higgins, 2010).

Geodetic data collected in the 1990s and 2000s, combined with seismic and petrologic data, provided improved resolution of the geometry of Kīlauea's magma storage zones and transport pathways. This improvement is largely due to the excellent temporal resolution of continuous ground-based sensors, including GPS stations and borehole tiltmeters, and outstanding spatial resolution from InSAR. We synthesize these data, especially those acquired during the 2003–07 magma supply increase to Kīlauea (see "2003–07 Increase in Magma Supply to Kīlauea" section and fig. 4), with previously published studies to propose a refined model for the geometry of magma storage and transport at Kīlauea that is consistent with petrologic, geophysical, and geologic data (fig. 10).

Our model contains several elements: (1) a summit magma system consisting of two long-term reservoirs, one at ~3 km beneath the south caldera (SC in fig. 10) and a second at 1–2 km beneath the caldera center (H in fig. 10), as well as occasional magma storage beneath Keanakāko‘i Crater (K in fig. 10); (2) a seismic Southwest Rift Zone (so called because it is easily recognized from earthquake hypocenters) at ~3-km depth that is connected to the south caldera magma reservoir and that stores magma during times of heightened summit magma pressure; (3) an East Rift Zone with a molten core (from the summit to at least its distal subaerial extent) at ~3-km depth (with compositionally isolated pods of stored magma at that depth and shallower), which is also connected to the south caldera magma reservoir and has been active continuously since 1983; (4) a volcanic Southwest Rift Zone (so named because it hosts more eruptive vents and fissures than the seismic Southwest Rift Zone) within ~1 km of the surface that is connected to the shallower summit magma reservoir and extends southwest from Halema‘uma‘u Crater; and (5) a "Halema‘uma‘u-Kīlauea Iki trend" (abbreviated HKIT in fig. 10) within ~1 km of the surface that is also connected to the shallower summit magma reservoir but extends east from Halema‘uma‘u Crater towards Kīlauea Iki. The model does not include a deep rift zone that allows magma to bypass the summit magma storage system and intrude into, or erupt from, the rift zones. Instead, we favor the model of Johnson (1995b), who proposed that the East and Southwest Rift Zones host molten cores at 3–5-km depth and supply magma vertically both towards the surface and to greater depths. Below, we describe these zones in detail.

Summit Magma Storage

Storage of magma beneath Kīlauea's summit was suspected by observers in the 1800s, including the first non-Hawaiian visitor to the volcano, William Ellis (Ellis, 1825), and no doubt by Hawaiians before then, on the basis of summit eruptive activity and the presence of the caldera, which was thought to be related to removal of subsurface magma (see discussion in Peterson and Moore, 1987). Early geophysical evidence for magma storage was provided mostly by deformation measurements. For example, deflation of the summit associated with magma withdrawal in 1924 (fig. 11) resulted in subsidence of several meters near Halema‘uma‘u Crater (which collapsed as a result of the activity) and horizontal displacements of over 1 m toward the center of the caldera (Wilson, 1935; Dvorak, 1992). These data were used by Mogi (1958) in a now-classic paper that modeled the deformation as due to pressure decrease caused by withdrawal of magma from point sources at 3.5-km and 25-km depth (although Eaton (1962) rejected the deeper source as an artifact of scale error in the wooden leveling rods, and Dvorak (1992) found that a single source at 4.5-km depth was sufficient to fit the data).

Tilt measurements that started in 1913 with horizontal pendulum seismographs (Finch, 1925) and continued with the development of water tube tilt measurements in the 1950s (Eaton, 1959) revealed that the summit was almost always in some state of inflation or deflation, often associated with eruptive and earthquake activity (fig. 12). Seismic evidence for magma storage was provided by data from the modern seismic network installed at Kīlauea in the 1950s (Eaton and Murata, 1960; Eaton, 1962; Okubo and others, this volume, chap. 2), which made high-resolution earthquake locations possible.

The geometry of magma storage is complex. Fiske and Kinoshita (1969) tracked inflationary deformation during the 22.5 months between Kīlauea's 1965 ERZ eruption and 1967–68 summit eruption, noting that the locus of maximum uplift changed over time (fig. 13*A*), with deformation during several time periods best approximated by a source at about 2–3-km depth beneath the southern part of the caldera. Their conceptual model of the summit magma system was a plexus of dikes and sills, with different zones of the system activated at different times—similar to the model of Dieterich and Decker (1975) for the same time period, and a proposal adopted by numerous authors to explain changes in surface deformation over time (for example, Schimozuru, 1981; Yang and others, 1992). Ryan and others (1981) and Ryan (1988) proposed a series of interconnected magma storage and transport zones beneath Kīlauea's summit and rift zones from seismic data and numerical models.

Deformation during the 2003–07 magma supply increase (see section entitled "2003–07 Increase in Magma Supply to Kīlauea" and Poland and others, 2012) followed a progression that highlights several discrete areas of magma accumulation (fig. 14). InSAR provides an especially clear view of deformation that is indicative of distinct magma storage areas

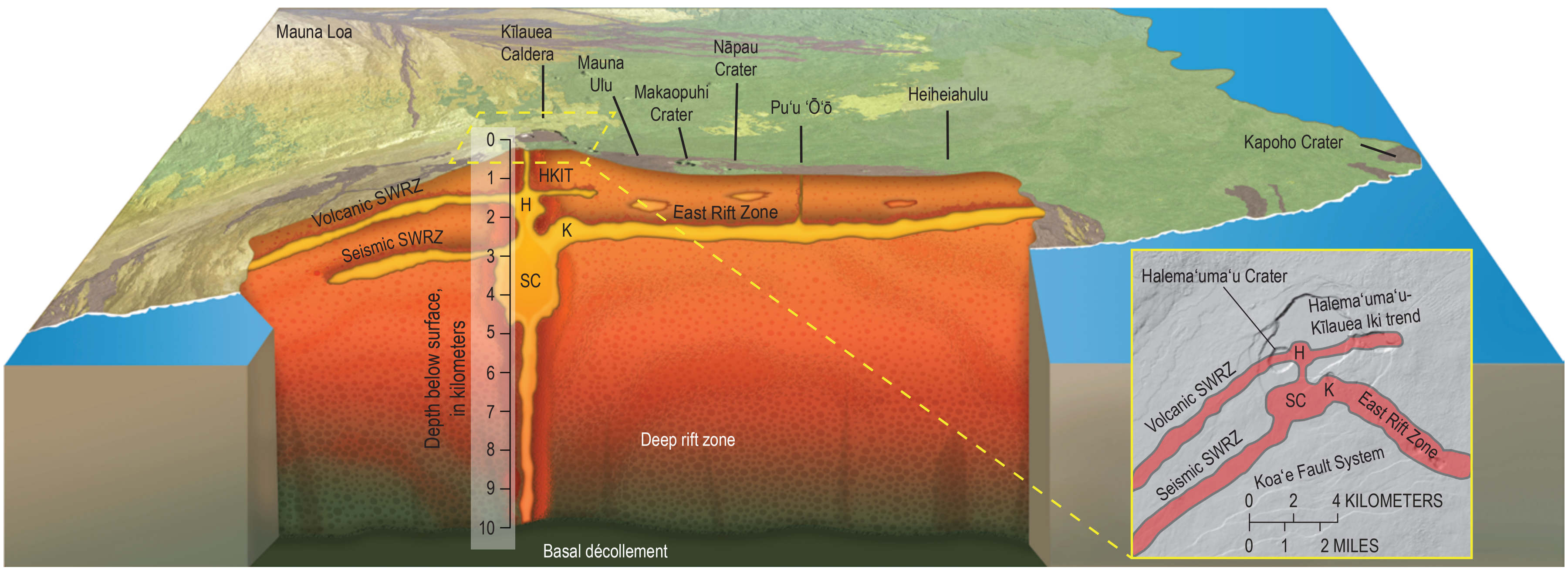


Figure 10. Illustration of proposed structure of Kīlauea’s subsurface magma plumbing system. Schematic cut-away shows a cross section through Kīlauea’s summit and rift zones. Magma pathways and storage areas are exaggerated in size for clarity. H, Halema‘uma‘u reservoir; K, Keanakāko‘i reservoir; SC, south caldera reservoir; SWRZ, Southwest Rift Zone. Plan view gives the relations of magma pathways to surface features and topography in the vicinity of Kīlauea Caldera.

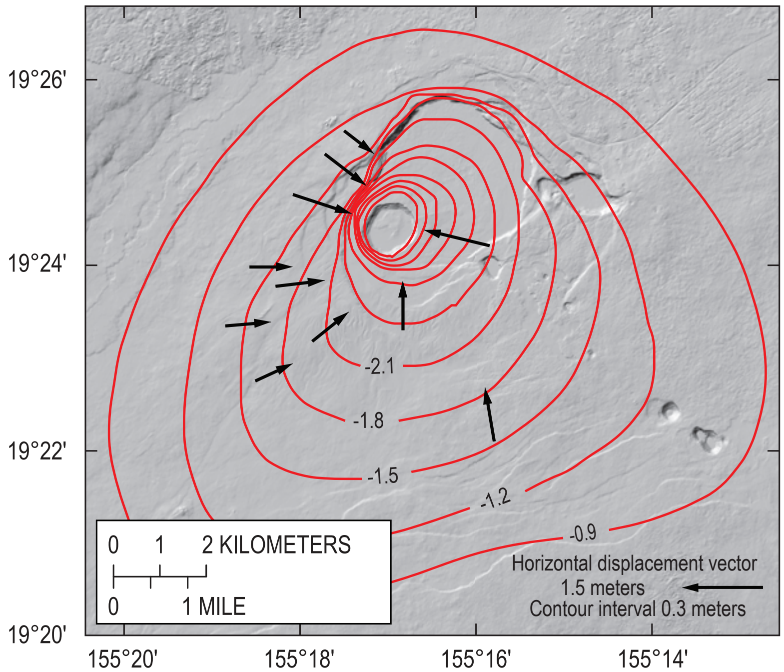


Figure 11. Map of deformation at Kīlauea's summit during the 1920s. Arrows show horizontal displacements from triangulation surveys spanning 1922–26, and contours give vertical displacements from leveling surveys spanning 1921–27. Maximum measured subsidence is about 4 m. Most of this deformation was associated with downdrop of the Halemaʻumaʻu lava column and enlargement of Halemaʻumaʻu Crater in 1924. Adapted from figures 2 and 8 of Wilson (1935).

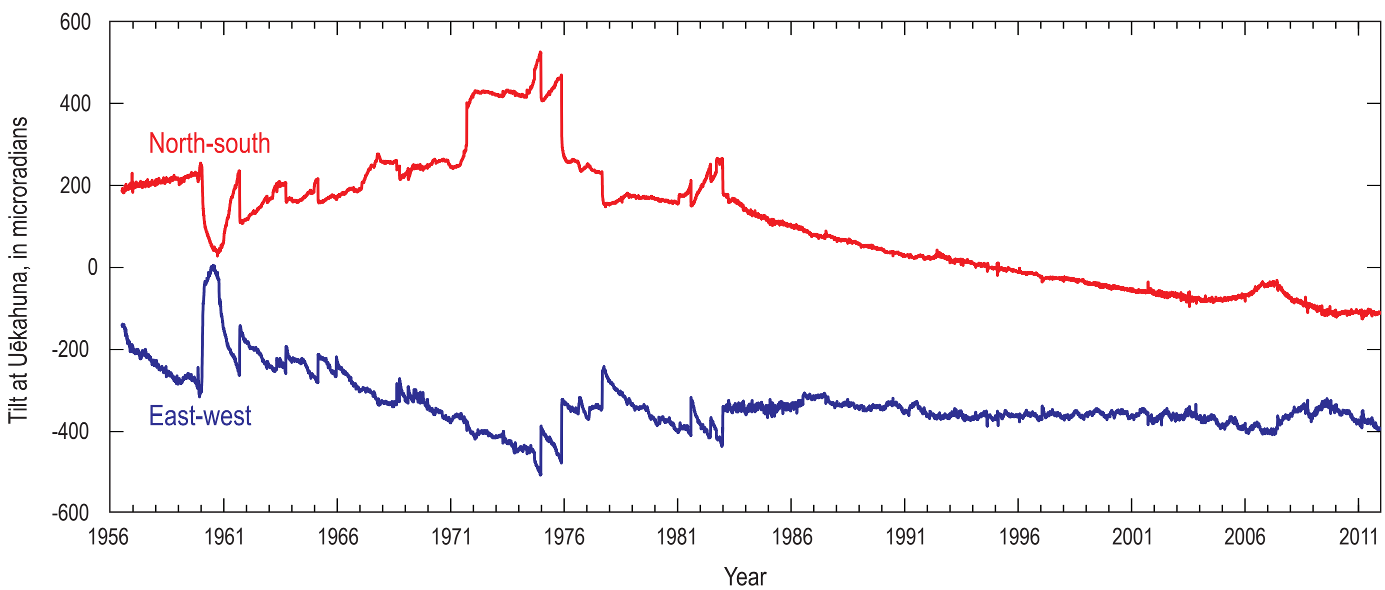


Figure 12. Plot of north-south (red) and east-west (blue) tilt from 1956 through 2011 at the Uēkahuna water-tube tiltmeter, located about 300 m west-northwest of the Hawaiian Volcano Observatory on the northwest rim of Kīlauea Caldera. Positive tilt is to the north or east, and negative tilt to the south or west. Northwest tilt indicates inflation and southeast tilt, deflation. Some offsets are associated with large earthquakes (for example, the 1975 earthquake on Kīlauea's south flank). A large offset associated with the November 1983 Kaʻōiki earthquake, beneath Mauna Loa's southeast flank, has been removed from the time series because the vault that houses the tiltmeter was damaged by the shaking.

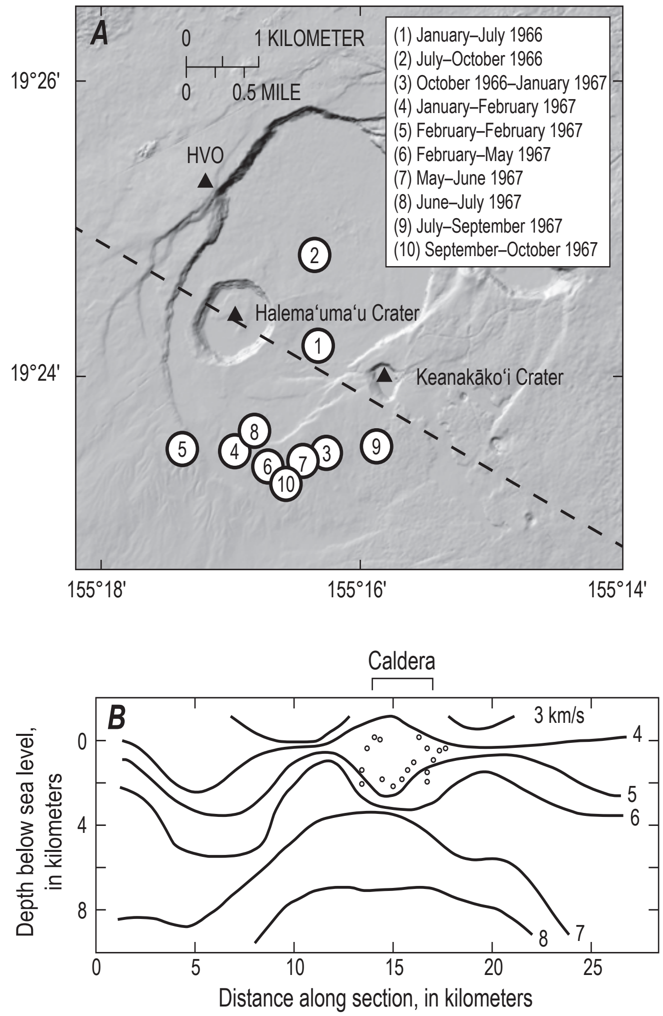


Figure 13. Deformation and seismic data indicating magma storage beneath Kīlauea Caldera. *A*, Map of centers of uplift determined by leveling during 1966–67. Dashed line indicates trend of section in part *B*. Adapted from figure 5 of Fiske and Kinoshita (1969). *B*, Cross section through Kīlauea Caldera showing seismic velocities (in kilometers per second) and shallow earthquakes in the vicinity of the caldera during 1980–81 (circles). Zone of low seismic velocity surrounded by earthquakes beneath the caldera suggests the presence of magma. Cross section extends both northwest and southeast of the dashed line in part *A*, and caldera bounds are indicated at the top of the section. From figure 3 of Thurber (1984).

beneath the summit (Baker and Amelung, 2012). Inflation associated with subsurface magma accumulation was centered just east of Halemaʻumaʻu Crater in 2003 (fig. 3, 14*A*), near Keanakākoʻi Crater in 2004–05 (fig. 3, 14*B*), and in the south caldera in 2006 (fig. 14*C*). Synthesizing these results with past studies of deformation, gravity, and seismicity suggests two long-term, interconnected magma reservoirs: one at ~3 km beneath the south caldera (which we term the "south caldera reservoir") and another at ~1–2 km beneath the caldera center (our "Halemaʻumaʻu reservoir"). In addition, at least one region of intermittent storage exists beneath the Keanakākoʻi Crater area near where the ERZ and summit intersect. While satisfying geophysical data, the presence of multiple discrete magma reservoirs beneath Kīlauea's summit is also consistent with petrologic observations of the preservation of distinct magma batches over time (Helz and others, this volume, chap. 6).

South Caldera Reservoir

The southern part of Kīlauea Caldera has long been recognized as the main locus of persistent deformation and, therefore, the main region of magma storage beneath the summit. Starting in the late 1950s and early 1960s, models of deformation consistently located a source of volume change beneath the south caldera at about 3–4-km depth (Eaton, 1959, 1962). The map-view location of the magma reservoir is determined geodetically and, thus, is independent of models for depth and shape. Dvorak and others (1983) modeled numerous episodes of inflation and deflation during 1966–70 using point sources, which showed a clustering at 2–4 km beneath the south caldera. Davis (1986) introduced an ellipsoidal source located in the same place to fit deformation of Kīlauea's summit during the 1970s. An ellipsoidal source with a center about 2 km below the south caldera was also modeled by Dieterich and Decker (1975) from vertical and horizontal deformation data collected during inflation in January–February 1967, although the top of their reservoir extended to less than 1 km beneath the surface. Yang and others (1992) suggested that all deflations and inflations were caused by volume changes in a single spherical body at 2.6 km beneath the south caldera, and that migrating deformation maxima during periods of uplift reflected dike intrusions beneath the caldera center and upper parts of the rift zones. The geodetically determined location of the south caldera magma reservoir correlates with both an aseismic zone at 3–6-km depth (Koyanagi and others, 1976; Ryan and others, 1981) and low P-wave velocities (Thurber, 1984; Rowan and Clayton, 1993)—characteristics consistent with magma storage (fig. 13*B*).

Summit deformation during the first 20 years of Kīlauea's 1983–present (as of 2014) ERZ eruption was dominated by deflation centered on the south caldera, which has been modeled by several authors as a small amount of pressure (or volume) loss at ~3-km depth (Delaney and others, 1990, 1993; Owen and others, 1995, 2000a; Cervelli and Miklius, 2003; Baker and Amelung, 2012). Similarly, gravity data collected during the eruption found mass changes centered on the south caldera with mass loss of only a few percent of the erupted volume, which supports interpretations based on deformation data that the south caldera reservoir served as a waypoint for magma that entered the volcano en route to the eruption site (Johnson, 1992; Kauahikaua and Miklius, 2003; Johnson and others, 2010). Inflation of the south caldera and the upper part of the Southwest Rift Zone (SWRZ) in 2006 was imaged with excellent spatial resolution by InSAR (fig. 14*C*; Myer and others, 2008; Baker and Amelung, 2012). Modeling the inflation observed by InSAR and GPS—and assuming a region

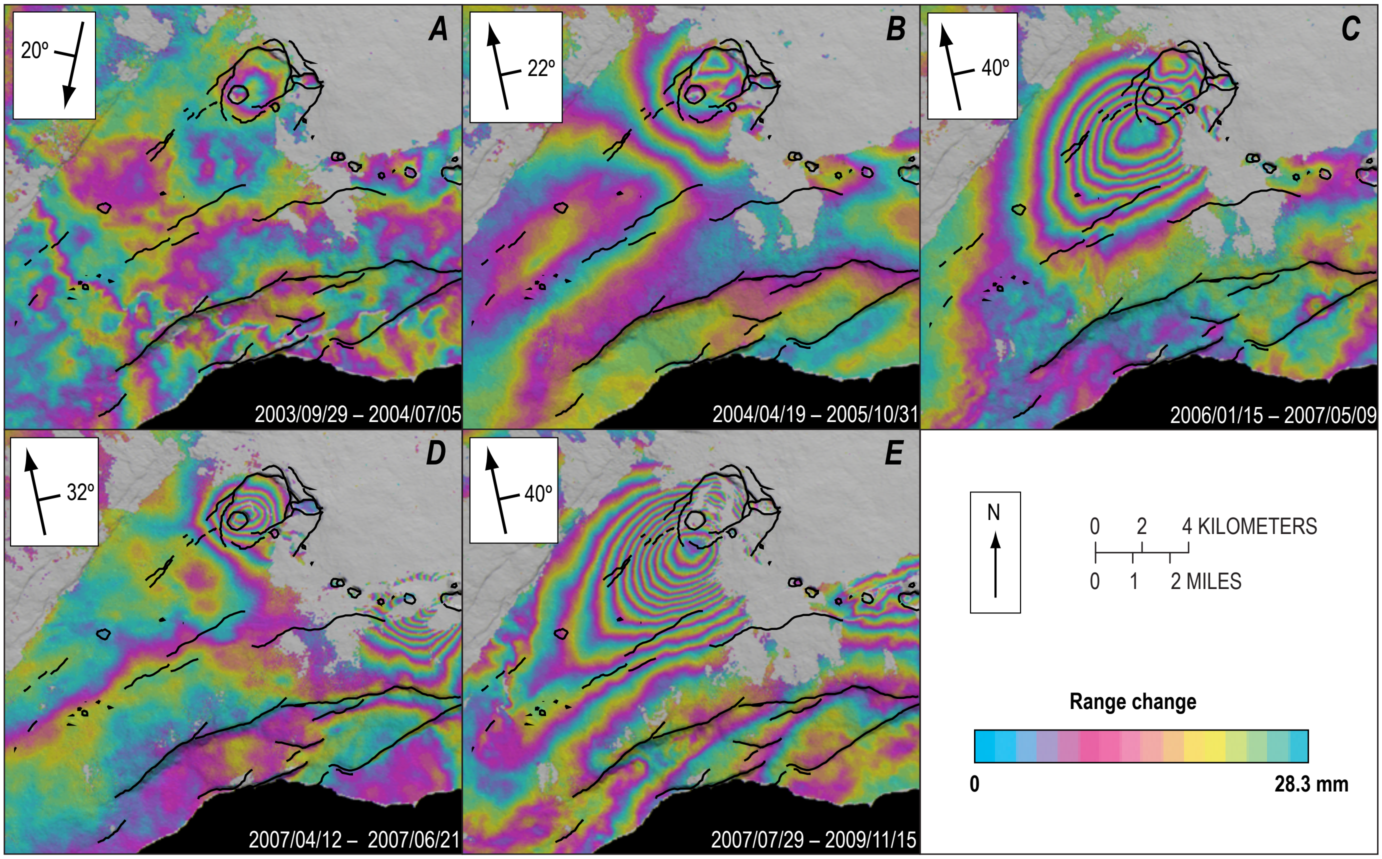


Figure 14. Interferograms detailing surface deformation of Kīlauea's summit area during the 2003–07 magma supply increase. Data are from the Advanced Synthetic Aperture Radar (ASAR) instrument on the Envisat satellite. For all images, dates spanned are in the lower right, and upper left inset gives flight direction (arrow) and look direction (orthogonal line with angle in degrees from vertical). Color scale in the lower right panel applies to all five interferograms. One fringe is 28.3 mm of range change along the radar line of sight, where positive (magenta to yellow) indicates increasing range, that is, ground motion away from the satellite: subsidence. *A*, Uplift is focused within the caldera, near Halema'uma'u Crater. *B*, Uplift has shifted to near Keanakāko'i Crater. *C*, Uplift rate has increased and is centered on the south caldera and upper Southwest Rift Zone. *D*, Subsidence near Halema'uma'u Crater and uplift of the East Rift Zone caused by magma withdrawal from the summit and intrusion into the East Rift Zone during June 17–19, 2007. *E*, Subsidence centered in the south caldera and upper Southwest Rift Zone, as well as in the East Rift Zone.

of distributed opening beneath the south caldera and upper SWRZ (see appendix for details on modeling procedure)—suggest a source depth of 3 km (95-percent confidence bounds are 2.1–5.1 km) beneath the south caldera and upper SWRZ (fig. 15), essentially the same location as that modeled by numerous other workers.

Volume estimates of the south caldera magma reservoir are diverse but generally cluster between 3 and 20 km^3. Dawson and others (1999) identified a low P-wave velocity anomaly in the south caldera with a volume of 27 km^3 (on the basis of a 5 percent reduction in velocity contrast from an initial model). Decker (1987) pointed out that magma storage must have been, at some point, at least the volume of the caldera collapse, estimated at 3 km^3 and possibly as large as 22 km^3, assuming a spherical shape with a diameter of 3.5 km (roughly equivalent to the caldera dimensions). Only part of this region may currently be molten, however, which would bring the volume closer to the 11 km^3 estimate that Wright (1984) derived from geodetic data. Johnson (1992) calculated an effective volume (that is, the volume that behaves like a fluid) of about 13 km^3. At the low end of the spectrum is a volume of 0.08 km^3, based on the magma supply rate and average repose period between eruptions (Klein, 1982). Residence time analysis of rapid geochemical fluctuations led Pietruszka and Garcia (1999b) and Garcia and others (2003) to infer a simple source geometry with a volume of 2–3 km^3, which may be smaller than most geophysical estimates because it represents the hotter core of the reservoir in which magma mixing occurs. As an upper bound, Denlinger (1997) calculated a volume of 240 km^3 from the ratio of pressure change to volume, but this estimate is probably more reflective of the entire volume of magma within and beneath Kīlauea and not solely that of the south caldera reservoir (Ryan, 1988).

The wide range of these volume estimates demonstrates the difficulty in defining just what constitutes a magma reservoir. Zones of magma storage probably grade from host rock to molten liquid and will contain regions of crystal mush and exsolved volatiles that will respond differently to applied stress (for instance, changing pressurization due to episodes of magma accumulation and withdrawal), depending also on the

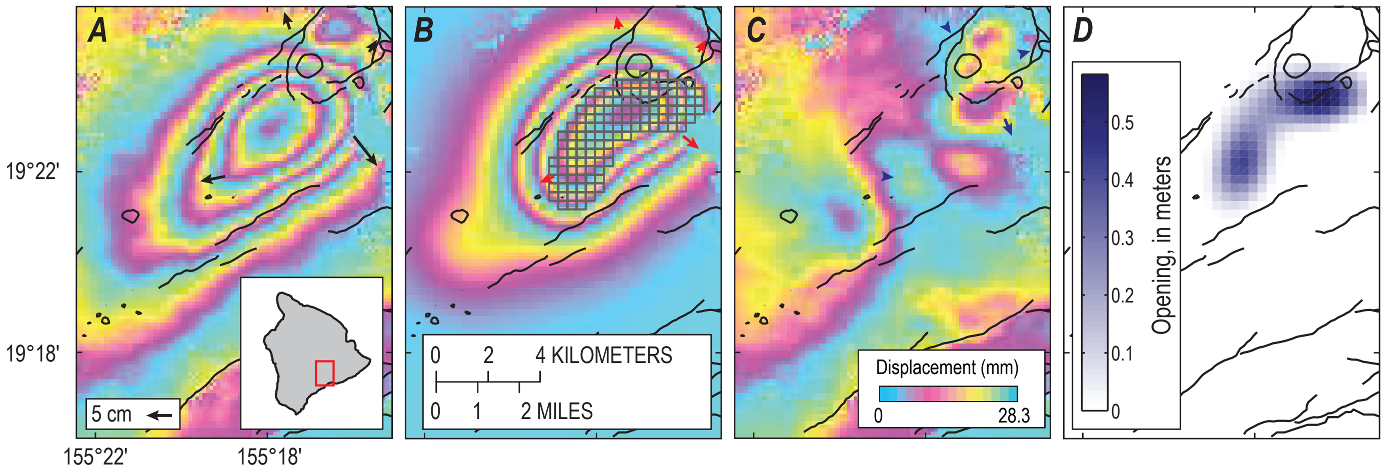


Figure 15. Source model of summit inflation due to volume increases in the south caldera and upper Southwest Rift Zone. Interoferometric Synthetic Aperature Radar (InSAR) and Global Positioning Ssystem (GPS) data span May 13–September 30, 2006. InSAR data are from the Advanced Synthetic Aperture Radar (ASAR) instrument on the Envisat satellite, beam mode 3, track 365. Arrows are horizontal displacement vectors scaled according to magnitude of deformation; where small, only arrowhead may show. In contrast to figure 14, one fringe represents 28.3 mm of displacement, where positive (magenta to yellow) indicates line-of-sight uplift. *A*, Observed deformation from InSAR and GPS. *B*, Modeled GPS and InSAR displacements assuming distributed opening of a sill-like source at 3-km depth. Boxes indicate sill elements that opened by 1 cm or more. *C*, Residual (observed minus modeled) GPS and InSAR displacements based on the model in part *B*. *D*, Map showing distribution of model sill opening.

magnitude and time scale of the stress (Gudmundsson, 2012). Such a question is inherent to any study of magma storage. We therefore caution against strict interpretations of reservoir volumes without a thorough understanding of the underlying assumptions that went into volume calculations.

Halemaʻumaʻu Reservoir

A second long-term zone of magma storage beneath Kīlauea's summit caldera is suggested by deformation results that indicate volume/pressure change beneath, or east of, the east margin of Halemaʻumaʻu Crater (see, for example, location 2 in fig. 13*A*). Leveling and tilt data from the summit over various time periods show deformation centered just east-northeast of Halemaʻumaʻu Crater, distinct from the south caldera reservoir (Fiske and Kinoshita, 1969; Dvorak and others, 1983), although the presence of an independent magma storage zone in this area (as opposed to occasional dike intrusions) was not proposed until later (Cervelli and Miklius, 2003). The reservoir probably feeds the eruptive vent within Halemaʻumaʻu Crater that formed in March 2008, but the connection between the vent and storage area is complex (Chouet and Dawson, 2011).

The installation of electronic borehole tiltmeters at the summit of Kīlauea, starting in 1999, provided further evidence of a magma storage reservoir beneath the east margin of Halemaʻumaʻu Crater. The tiltmeters record repeated transient tilt events that last hours to days and are characterized by sudden deflation, followed by equally sudden inflation ("DI" events; see "Characteristics of Magma Transport at Hawaiian Volcanoes" section and Anderson and others, in press). Cervelli and Miklius (2003) modeled four of these events as due to a point source of pressure change at 500–700-m depth about 0.5 km east of Halemaʻumaʻu Crater. With the start of summit eruptive activity in 2008 (ongoing as of 2014), DI events became much more common (~5–10 per year during 1999–2007, before the eruption, compared to more than 50 per year in 2009–13, during the eruption). Modeling the tilt events (see appendix for details on modeling procedure) as due to a point source of pressure change (a more complex source geometry is not resolvable with the limited number of tilt stations at the summit) indicates a location about 1–2 km beneath the east margin of Halemaʻumaʻu Crater (fig. 16), although the depth is poorly constrained. Anderson and others (in press) inverted tilt data recorded during more than 450 DI events that occurred between 2000 and 2013 and found the same location and depth range, indicating that the source position does not vary over time.

Rapid deflation of the Halemaʻumaʻu reservoir has been documented repeatedly since the start of the 1983–present (as of 2014) ERZ eruption and is a result of magma drainage to feed ERZ intrusions and eruptions, as exemplified by activity in 1997 (Owen and others, 2000b), 2007 (Montgomery-Brown and others, 2010), and 2011 (Lundgren and others, 2013). Models of these subsidence events suggest a source depth of 1–2 km beneath the east margin of Halemaʻumaʻu Crater (Poland and others, 2009; Montgomery-Brown and others, 2010; Lundgren and others, 2013). As an example, inversion of InSAR and GPS data for summit subsidence associated with the 2011 ERZ fissure eruption (see appendix for details on modeling procedure) gives a depth of 1.4 km (95 percent confidence range is 1.0–1.9 km; fig. 17).

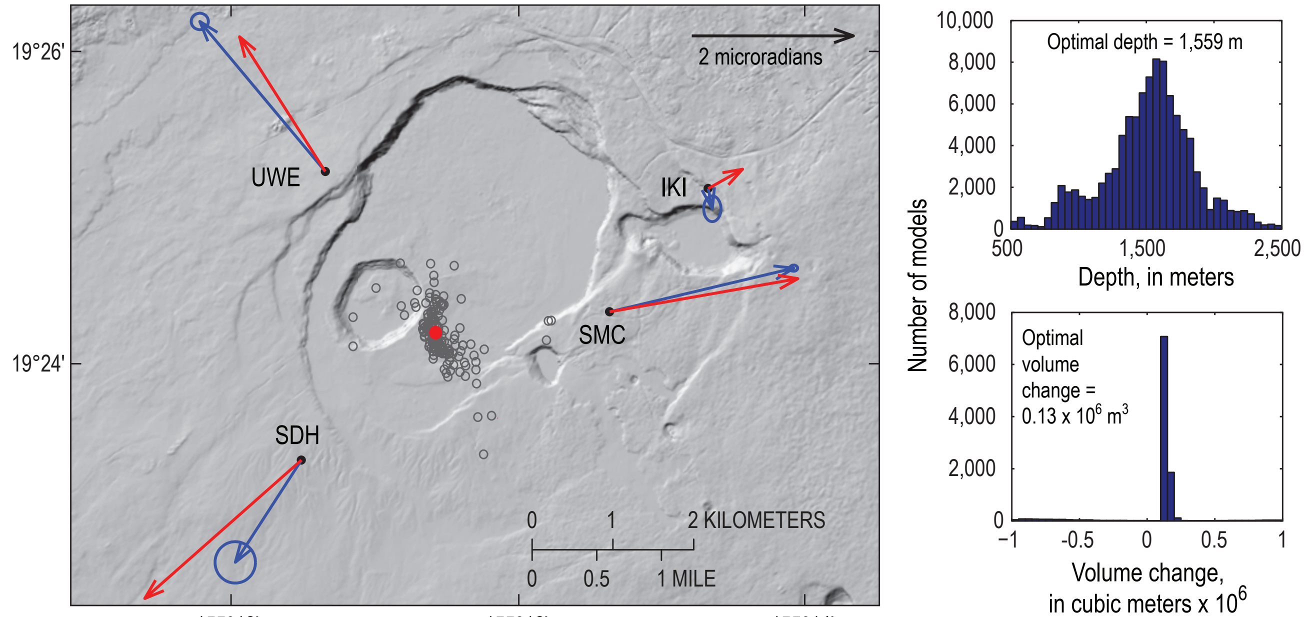


Figure 16. Source modeling of transient tilt events at Kīlauea's summit. Modeling assumes a point-source geometry. Arrows show magnitude and direction of observed (blue) and predicted (red) tilts for a sample model at the sites of four tiltmeters (IKI, SDH, SMC, and UWE) positioned around Kīlauea Caldera. Best-fitting source location marked by red dot. Distributions of depth and volume change for the sample model are given at right. Source locations for models of 151 additional tilt events that occurred during 2000–09 are given by gray circles. Depth ranges for these additional models are widespread but generally fall between 0.2 and 2 km beneath the surface.

Seismic and gravity evidence for the Halema'uma'u reservoir is also convincing. Broadband seismic stations in Kīlauea Caldera detected very-long-period (VLP) tremor (Ohminato and others, 1998) and VLP seismic events (Almendros and others, 2002; Chouet and others, 2010) originating about 1 km beneath the east margin of Halema'uma'u Crater, possibly an indication of flowing magma. High-resolution (0.5 km) velocity models imaged a low-velocity P-wave anomaly in about the same location (Dawson and others, 1999), which corresponds to a cluster of long-period (LP) seismicity (Battaglia and others, 2003); tremor and LP events also occur just above this region (Almendros and others, 2001). Finally, gravity data spanning 1975–2008 detected mass increase beneath the east rim of Halema'uma'u Crater, although no long-term deformation was measured in this location. Models of the gravity data suggest a source depth of 1 km, and the lack of deformation accompanying the gravity increase implies that magma was accumulating in void space that may have been created when several tens of millions of cubic meters of magma drained from the summit as a result of the 1975 earthquake (Dzurisin and others, 1980; Johnson and others, 2010).

The Halema'uma'u magma reservoir is an order of magnitude smaller than the south caldera reservoir. Johnson (1992) suggested that the volume was at least 1.6 km^3, based on summit deflation associated with draining of the Mauna Ulu lava lake in 1973. A model of deformation and gas emission data during rapid deflation of the source associated with the June 2007 ERZ intrusion and eruption led Poland and others (2009) to determine a volume of 0.2–1.2 km^3. Anderson and others (in press) related lava-level changes within the summit eruptive vent to ground tilt recorded during two especially large-magnitude DI events in February 2011 and found a volume of 0.15 to 2.7 km^3. Segall and others (2001) used deformation associated with the brief 1997 ERZ eruption to model a volume for the Halema'uma'u magma reservoir—shown by Owen and others (2000b) to be the deflation source at Kīlauea's summit during that event—of ~20 km^3. They assumed a value for the elastic modulus of 20 GPa; reducing this by an order of magnitude, which is reasonable, given the fractured nature of Kīlauea's shallow subsurface (for example, Rubin and Pollard, 1987), would correspondingly decrease the modeled volume to ~2 km^3 (Anderson and others, in press).

Keanakāko'i Reservoir

Multiple authors have noted that modeled locations of long-term summit subsidence (in other words, subsidence not associated with rapid drainage of magma to feed rift zone intrusions and eruptions) cluster beneath the south caldera, whereas inflation sources are distributed over a broader area (for example, Dvorak and others, 1983; Yang and others, 1992; fig. 13*A*). This observation led Yang and others (1992) to suggest that inflation of the south caldera reservoir was accompanied by dike intrusion, but the intruded dikes did not subsequently close during reservoir deflation. As an alternative, we suggest that the

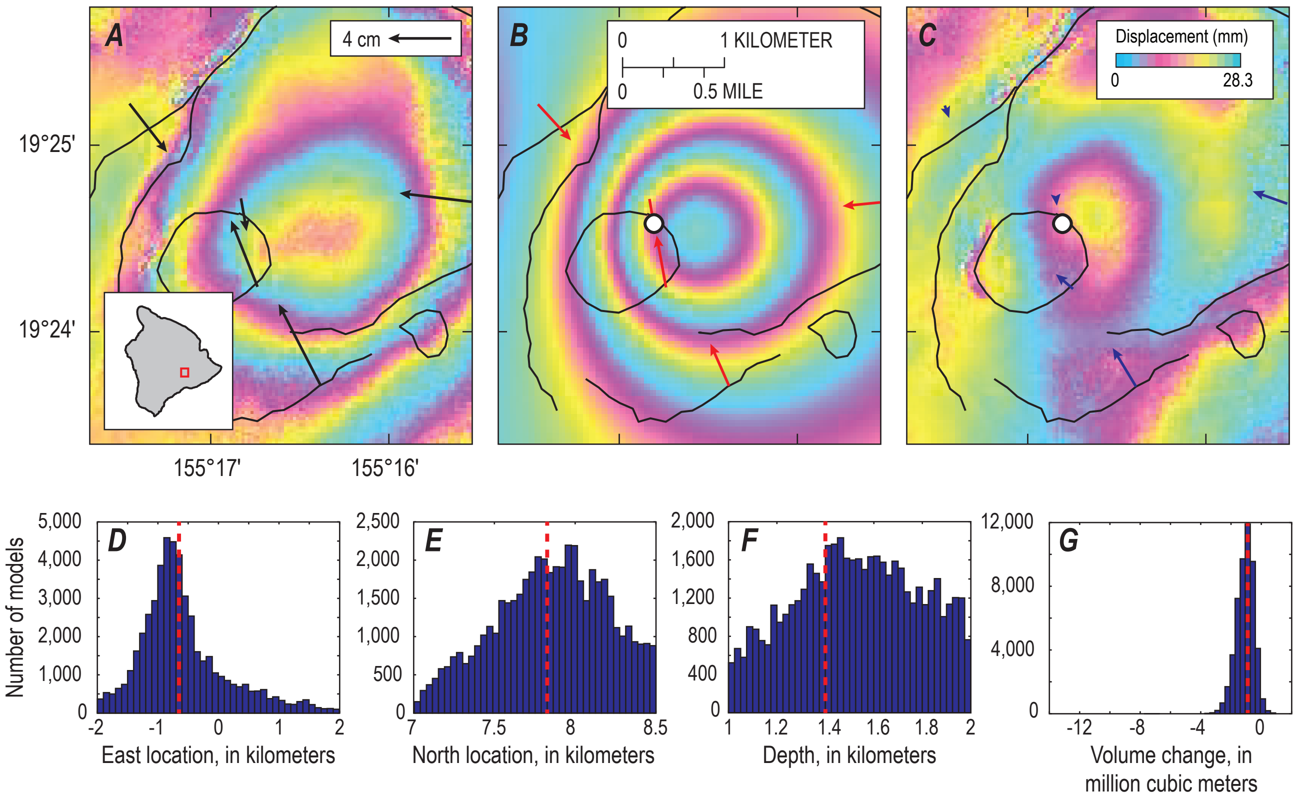


Figure 17. Source model of summit subsidence associated with the 2011 Kamoamoa fissure eruption. Interoferometric Synthetic Aperature Radar (InSAR) and Global Positioning System (GPS) data span January 19–March 6, 2011. InSAR data are from the PALSAR instrument on the ALOS satellite, orbital path 598. Color scale and arrows as in figure 15. *A*, Observed deformation from InSAR and GPS. *B*, Modeled GPS and InSAR displacements assuming a point source of volume change (white circle) at 1.4 km beneath the northeast margin of Halema‘uma‘u Crater. *C*, Residual (observed minus modeled) GPS and InSAR displacements, based on the point source model in part *B*. *D–G*, Histograms showing the distributions of model parameters, including east location (*D*, with arbitrarily sourced x-axis), north location (*E*, with arbitrarily sourced x-axis), depth (*F*), and volume change (*G*). Dashed red lines indicate best-fitting solutions, which do not necessarily align with the peaks of the distributions.

distribution of inflationary centers reflects the combined effects of several persistent zones of magma storage. In addition to the Halema‘uma‘u and south caldera reservoirs, magma is also stored beneath the area near Keanakāko‘i Crater (fig. 3).

Relocated LP earthquakes cluster at about 4-km depth beneath Keanakāko‘i Crater (Battaglia and others, 2003), which is also a region of low P-wave velocity (Dawson and others, 1999). In addition, uplift has frequently been localized in that area during periods of summit inflation; for example, between the 1959 Kīlauea Iki and 1960 Kapoho eruptions (Wright and Klein, 2014), before the 1967–68 summit eruption (location 1 in fig. 13*A*; Fiske and Kinoshita, 1969), and before the 1974 summit eruption (Lockwood and others, 1999). Many dikes modeled for episodes of uplift cluster near Keanakāko‘i Crater, and seismicity indicates that ERZ intrusions are frequently initiated near Keanakāko‘i (Yang and others, 1992; Klein and others, 1987), both suggesting magma storage in the area. Historical eruptions in this region occurred within Keanakāko‘i Crater in 1877 (Peterson and Moore, 1987) and nearby in 1971 (Duffield and others, 1982) and 1974 (Lockwood and others, 1999).

InSAR data that span 2004–05 indicate uplift immediately north and east of Keanakāko‘i Crater as part of the sequence of deformation associated with the 2003–07 magma supply increase (Baker and Amelung, 2012; fig. 14*B*). A model of InSAR and GPS displacements from 2004–05 (see appendix for details on modeling procedure) suggests a deformation source depth of about 2.6 km (95 percent confidence range is 2.0–4.9 km) near Keanakāko‘i Crater (fig. 18).

We propose that the localized uplift near Keanakāko‘i Crater observed occasionally during periods of caldera inflation represents a transient accumulation of magma. The accumulation may occur due to a backup of magma that cannot be accommodated by the ERZ conduit. Eruptive activity in 2004–05 included a sudden increase in the effusion rate from the ERZ eruptive vent in February 2005 (Poland and others, 2012), suggesting that the rift conduit began to carry more magma from the summit. Continued summit inflation after this date, including uplift near Keanakāko‘i Crater (figs. 14*B* and 18), implies that the rift conduit was full and could not transport larger volumes of magma towards the ERZ eruption site. Pressure increase where the ERZ and summit magma systems intersect near Keanakāko‘i Crater, due to conduit back-ups, such as the one that occurred in 2005, may eventually lead to an eruption from this accumulation zone, as exemplified by activity in 1971 and 1974. Each of those eruptions was preceded by waning activity at Mauna Ulu, presumably due to a blockage between the summit and that vent, and the lava erupted at the summit was similar in composition to that which had been erupted during the preceding activity at the then-active Mauna Ulu (Duffield and others, 1982; Lockwood and others, 1999). Indeed, the very presence of Keanakāko‘i Crater, and the existence of pit craters in general, argues for magma storage, since the crater may have formed during sudden drainage of magma from a subsurface storage area (for example, Swanson and others, 1976b).

Rift Zones

Kīlauea is generally described as having two rift zones that radiate to the east and southwest from the summit. Geologic data suggest that each rift zone is comprised of two distinct magma pathways with different trends and surface expressions (for example, Holcomb, 1987; Fiske and others, 1993)—a model that is also supported by geophysical results. The shallower pathways, described below as the volcanic Southwest Rift Zone and Halema‘uma‘u-Kīlauea Iki Trend, are within 1 km of the surface and are fed from high-level parts of the summit magma system (specifically, the Halema‘uma‘u reservoir). Slightly deeper pathways at about 3-km depth—the East Rift Zone and seismic Southwest Rift Zone, as described below—are fed by the south caldera reservoir (fig. 10). The volcanic Southwest Rift Zone and Halema‘uma‘u-Kīlauea Iki Trend are the sites of numerous eruptions but are structurally superficial to the volcano (Fiske and others, 1993), at least during recent times, while the slightly deeper seismic Southwest Rift Zone and East Rift Zone represent the current structural boundaries of the mobile south flank (for example, Cayol and others, 2000). The East and Southwest Rift Zones may therefore be viewed as parallel structures, each with both shallow and deeper magma pathways that are connected to the summit magma system at different depths, and with the shallower pathways located north of the deeper pathways. Such geometry may be a consequence of seaward migration of both rift zones over time, as suggested by Swanson and others (1976a). In addition, both rift zones are underlain by a zone of deep extension (see the “Deep Rift Zones” section below) that has been proposed as a region of magma accumulation and transport (for example, Delaney and others, 1990), but for which evidence for magma storage is contradictory.

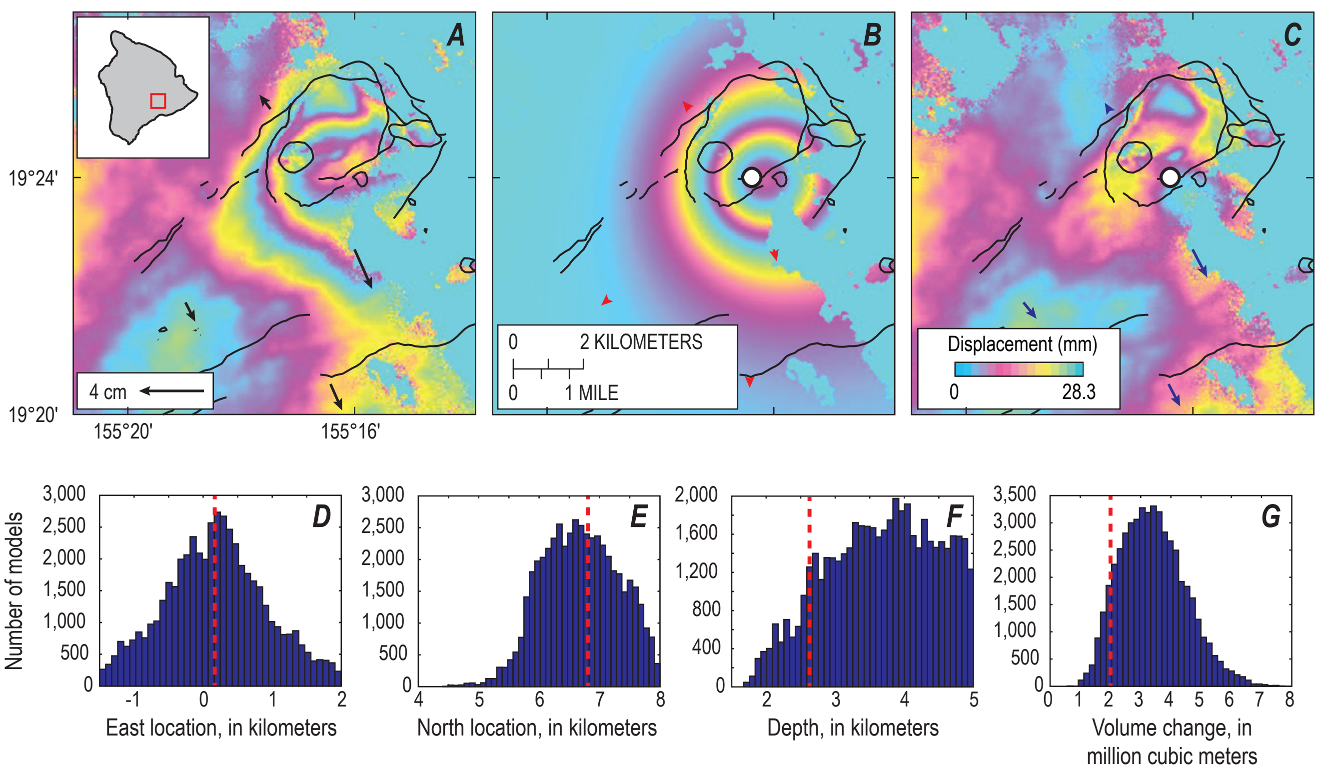


Figure 18. Source modeling of summit inflation recorded by Interoferometric Synthetic Aperature Radar (InSAR) and Global Positioning System (GPS) data spanning November 3, 2004–October 19, 2005. InSAR data are from the ASAR instrument on the Envisat satellite, beam mode 2, track 429. Color scale and arrows as in figure 15. *A*, Observed deformation from InSAR and GPS. *B*, Modeled GPS and InSAR displacements assuming a point source of volume change (white circle) at 2.6-km depth beneath the south part of the caldera, near Keanakāko‘i Crater. *C*, Residual (observed minus modeled) GPS and InSAR displacements based on the point source model in part *B*. *D*–*G*, Histograms showing the distributions of model parameters, including east location (*D*, with arbitrarily sourced x-axis), north location (*E*, with arbitrarily sourced x-axis), depth (*F*), and volume change (*G*). Dashed red lines indicate best-fitting solutions, which do not necessarily align with the peaks of the distributions.

Seismic Southwest Rift Zone

Southwest of Kīlauea Caldera is a broad alignment of eruptive vents, fractures, and seismicity that extends toward the coast. This zone, the Southwest Rift Zone (SWRZ), is composed of distinct strands (Holcomb, 1987). We follow previous authors in recognizing two primary strands, one defined largely by seismicity (the "seismic SWRZ") and the other marked mostly by alignments of fissures and eruptive vents (the "volcanic SWRZ," described in a section of the same name below). The seismic SWRZ, following the naming convention of Wright and Klein (2006, 2014), defines the boundary between the stable north sector of the volcano and the mobile south flank, based on modeling of rift zone opening (Cayol and others, 2000), and corresponds to the "middle rift strand" of Holcomb (1987).

The seismic SWRZ is marked by earthquakes that trend south and then southwest from the south caldera magma reservoir at a depth of ~3 km, southwest of the alignment of eruptive vents and fissures that defines the volcanic SWRZ (Klein and others, 1987; Fiske and Swanson, 1992). The seismic swarms mark magmatic intrusions, as indicated by deflation of the summit during propagation of earthquakes to the southwest, such as in 1974–75 (Lockwood and others, 1999). That episode resulted in the only post-18th century eruption from the seismic SWRZ on December 31, 1974. The eruption lasted a single day and was characterized by gas-rich lava that formed shelly pāhoehoe and fountain-fed ʻaʻā flows. In the days following the eruption, deformation and seismic evidence indicated intrusion of magma to lower parts of the seismic SWRZ (fig. 19*A*), but no lava erupted (Lockwood and others, 1999).

Since 1974, at least three additional magmatic intrusions have occurred in the seismic SWRZ, as judged by deformation measurements and seismicity (many additional intrusions occurred before 1974, as well; Klein and others, 1987). The first half of 1981 was characterized by small earthquake swarms in the seismic SWRZ that led Klein and others (1987) to hypothesize a continuous intrusion that culminated in August (fig. 19*B*). In June 1982, an earthquake swarm in the seismic SWRZ (fig. 19*C*) was accompanied by summit deflation and rift extension. Wallace and Delaney (1995) modeled this event as a dike intrusion coupled with seaward slip of the volcano's south flank. In April 2006, uplift was detected by InSAR (fig. 14*C*) and GPS (fig. 4*A*) in the portion of the seismic SWRZ near the caldera (Myer and others, 2008; Baker and Amelung, 2012; Poland and others, 2012). This deformation is best modeled as inflation of a sill-like body beneath the south caldera and upper seismic SWRZ at a depth of 3 km (fig. 15) and is correlated in time with shallow, high-frequency seismicity in the same area (fig. 19*D*).

Seismicity related to intrusions into the seismic SWRZ extends as far as the Kamakaiʻa Hills, where the trace of epicenters bends abruptly southward and the seismicity deepens as if the intrusions activate south-flank faults (fig. 19; Klein and others, 1987). Deep extension modeled from deformation data terminates in about the same place (Cayol and others, 2000). Geophysical data therefore indicate that the seismic SWRZ does not extend all the way to the coast but, instead, ends somewhere in the vicinity of the Kamakaiʻa Hills.

When not inflating, the normal deformation mode of the seismic SWRZ is opening and subsidence. Leveling results during 1996–2002, although dominated by south caldera deflation, include a component of subsidence along the seismic SWRZ (fig. 5 in Cervelli and Miklius, 2003). GPS data from station KOSM, located about 5 km southwest of Halemaʻumaʻu Crater (fig. 3), also show persistent subsidence of several centimeters per year before and after the 2003–07 magma supply surge (fig. 4*A*). Most of this subsidence is probably a result of rift-zone opening due to south-flank motion, as indicated by models of deformation data (Johnson, 1987; Owen and others, 2000a; Cayol and others, 2000), although a component of the subsidence may be caused by magma withdrawal (Johnson, 1995b) and (or) cooling and contraction of stored magma.

East Rift Zone

Post-18th century eruptions from the ERZ were sparse before the 1950s, occurring in 1840, 1922, and 1923 (Holcomb, 1987). Starting in the 1950s, the ERZ became the most active part of Kīlauea's magmatic system, with weeks-long eruptions from the lower part of the rift zone in 1955 and 1960. At least eight eruptions occurred in the 1960s from the middle ERZ, culminating in the 1969–74 eruption of Mauna Ulu—the longest-lasting post-18th century eruption along the rift zone to that time. Following the 1975 south flank earthquake, numerous ERZ intrusions and a few eruptions characterized the latter part of the 1970s. A series of intrusions occurred in early 1980, after which the ERZ was quiet until the start of eruptive activity in 1983 at Puʻu ʻŌʻō (Holcomb, 1987), which continues as of 2014.

Like the SWRZ, the ERZ is connected to the south caldera magma reservoir at a depth of ~3 km. Such a depth is indicated by the location of seismicity associated with ERZ dike intrusions (Klein and others, 1987; Wolfe and others, 1987). For instance, dikes that intruded into, and ultimately erupted from, the middle ERZ in June 2007 and March 2011 both ascended from depths of about 3 km, according to earthquake locations and deformation modeling (Syracuse and others, 2010; Montgomery-Brown and others, 2010; Lundgren and others, 2013).

The ERZ contains a molten core that connects the summit to at least the distal subaerial end of the rift zone, over 50 km from the summit, and possibly into the submarine part of the rift zone beyond (Fiske and others, 1993). Such a continuous magma system was suggested by Dana (1849, description starting p. 188), based on second-hand reports of summit lava lake drawdown during the 1840 flank eruption. Johnson (1995b) proposed that the ERZ molten core, at 3–5-km depth, could feed magma vertically downward to deeper levels, towards the surface and eruption, and laterally along the rift zone. A hydraulically connected magma system along the rift zone is

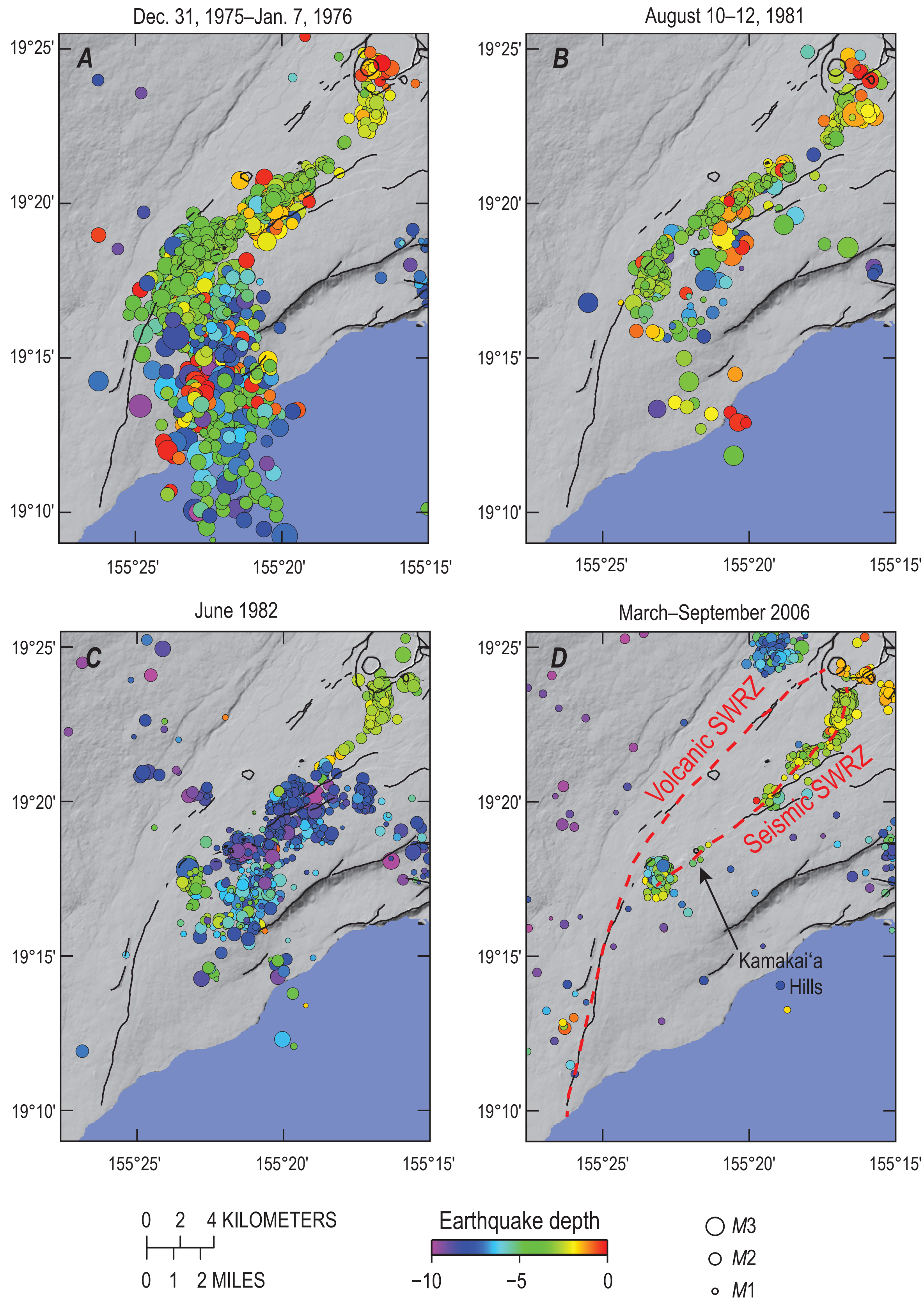


Figure 19. Maps of seismicity along Kīlauea's seismic Southwest Rift Zone (SWRZ). *A*, December 31, 1974–January 7, 1975. *B*, August 10–12, 1981. *C*, June 1982. *D*, March–September 2006. Color and size of circles gives depth and earthquake magnitude, respectively. Dashed red lines in part *D* indicate seismic SWRZ and volcanic SWRZ, and the Kamakai'a Hills are labeled.

required to explain summit subsidence during ERZ eruptions, which indicates magma withdrawal from beneath the summit (Dvorak and Okamura, 1987). This is especially apparent when summit subsidence begins after the start of an ERZ eruption, as was the case on the lower ERZ during 1955 and 1960 (Eaton and Murata, 1960; Helz and Wright, 1992; Fiske and others, 1993; Wright and Helz, 1996). Episodes of summit inflation also demonstrate hydraulic connectivity between the summit and lower ERZ. For example, in late 2003, as the summit began to inflate due to a surge in magma supply (see “2003–07 Increase in Magma Supply to Kīlauea” section and Poland and others, 2012), uplift began in both the middle (fig. 5*A*) and possibly lower (fig. 5*B*) ERZ, indicating magma accumulation within the rift zone. During prolonged ERZ eruptions (including the Mauna Ulu and Puʻu ʻŌʻō eruptions), magma might enter the volcano’s summit magma system and flow through the south caldera reservoir directly to the ERZ without spending time in storage (or displace a similar amount of stored magma into the ERZ), as suggested by gravity and deformation measurements from the 1980s through 2000s (Johnson, 1987; Kauahikaua and Miklius, 2003; Johnson and others, 2010).

Geophysical, petrologic, seismic, and physical evidence demonstrate that the ERZ contains multiple areas of shallow magma storage along its length. A reservoir beneath the Makaopuhi Crater area was first suggested by Jackson and others (1975) on the basis of seismicity and deformation associated with ERZ eruptions in 1968, and again by Swanson and others (1976b), using similar data from the 1969 ERZ eruption. Local gravity highs in the vicinity of Makaopuhi and Nāpau Craters suggest the presence of dense bodies—probably solidified magma storage zones—in these areas (Kauahikaua and others, 2000). High b-values (a seismic quantity associated with magma storage in volcanic environments) at multiple locations along the ERZ, including near Makaopuhi, also suggest magma storage (Wyss and others, 2001). Wolfe and others (1987) argued for a reservoir near Makaopuhi, because seismicity associated with the first episode of the Puʻu ʻŌʻō eruption started in that location (a pattern that was repeated in March 2011; Lundgren and others, 2013), and episodes of lava fountaining during 1983–86 suggested that a secondary reservoir between the summit and Puʻu ʻŌʻō acted as a valve that controlled eruptive activity. Owen and others (2000b) and Segall and others (2001) similarly found that including magma storage near Makaopuhi improved models of deformation from data recorded during a 22-hour eruption in Nāpau Crater in 1997, and Cervelli and others (2002) argued for a source of magma, possibly beneath Makaopuhi Crater, contributing to the 1999 intrusion.

The somewhat evolved compositions of many lava flows erupted from the ERZ require storage at shallow levels within the rift zone (Wright and Fiske, 1971; Thornber and others, 2003). Likewise, compositions of lava from the early stages of the Puʻu ʻŌʻō eruption resulted from mixing between isolated stored magmas and a more mafic end member (Garcia and others, 1989, 1992, 2000; Thornber, 2003). Lava flows from the lower ERZ in 1955 and 1960 had a common source and show signs of mixing between evolved magmas stored within the rift zone (which crystallized pyroxenes and plagioclase as well as olivine) and more primitive magmas supplied from the summit (which crystallized only olivine; Helz and Wright, 1992; Wright and Helz, 1996). Drilling at a geothermal power plant in the lower ERZ in 2005 intersected a magma body with a dacitic composition at a depth of 2.5 km (Teplow and others, 2009). That such magma bodies exist is not surprising, given the frequent and sometimes long-lived activity along the ERZ (Parfitt, 1991); they were inferred from vent distributions and petrology prior to being drilled (Moore, 1983). The ongoing eruption at Puʻu ʻŌʻō, for instance, has formed a small (~1×10^7 m^3) magma storage area beneath the vent, as interpreted from geophysical, geochemical, and fluid dynamic evidence (Wilson and Head, 1988; Hoffmann and others, 1990; Garcia and others, 1992; Owen and others, 2000b; Segall and others, 2001; Heliker and others, 2003; Thornber and others, 2003; Shamberger and Garcia, 2007; Mittelstaedt and Garcia, 2007).

Volcanic Southwest Rift Zone

Holcomb (1987) suggested that the SWRZ included a northern strand that extended from Halemaʻumaʻu Crater to 5 km southwest of Maunaiki (fig. 3) and that was connected to the deeper magma plumbing system by way of Halemaʻumaʻu. Fiske and others (1993) termed this the “classic” SWRZ and suggested that it was a superficial feature fed by shallow dikes intruded laterally from high levels of the summit magma system. We term this strand the “volcanic SWRZ” to avoid confusion with the “seismic SWRZ” (described in a section of the same name above). The volcanic SWRZ follows an alignment of fissures and eruptive vents, including the Great Crack (fig. 3), as far as the coast and was the source of post-18th century eruptions in 1823, 1868, 1919–20, and 1971.

Instead of tapping the summit magma system at the level of the south caldera reservoir at ~3-km depth, as do the seismic SWRZ and ERZ, the volcanic SWRZ appears to be within 1 km of the surface and is fed directly from the Halemaʻumaʻu reservoir and, sometimes, from Halemaʻumaʻu Crater itself. This connection was demonstrated by the 1919–20 Maunaiki (fig. 3) eruption, when the active lava lake at Halemaʻumaʻu drained into a fissure that propagated southwest from the caldera. Jaggar (1919) observed lava flowing in cracks just beneath the surface, and occasionally reaching the surface, at multiple locations between Halemaʻumaʻu Crater and Maunaiki, attesting to its shallow nature. When the summit lava lake drained in 1922 and 1924, a dike-like structure interpreted to be the 1919–20 conduit was exposed in the walls of Halemaʻumaʻu Crater (fig. 20)—further evidence of its shallow connection to the summit magma system. This interpretation is at odds with the existence of higher gravity along the volcanic SWRZ, compared to the seismic SWRZ (Kauahikaua and others, 2000), implying a deeper and more important magma pathway. Gravity, however, reflects cumulative magma storage throughout Kīlauea's evolution. The current gravity field may indicate that the volcanic SWRZ was a more extensive magma pathway earlier in Kīlauea's history than it currently is, much like gravity data that also suggest southward migration of the ERZ over time (Swanson and others, 1976a).

Duffield and others (1982) provided an overview of historical eruptive activity southwest of the summit caldera, noting that the eruptions of 1868, 1919–20, and 1971 were fed from lava lakes at Halemaʻumaʻu Crater or magma stored at shallow depth beneath the crater (on the basis of drops in lava level or surface sagging coincident with the eruptions). The 1823 Keaīwa eruption issued from the Great Crack about 20 km from the summit and included exceptionally thin pāhoehoe (Stearns, 1926; Duffield and others, 1982; Soule and others, 2004). The eruption was interpreted by Ellis (1825) to be contemporaneous with a drop in lava level at the summit, forming the "black ledge" that existed at the time of his visit to Kīlauea Caldera in 1823, just a few months after the Keaīwa outbreak.

Volcanic SWRZ eruptions differ considerably in terms of eruption rate, eruption style, vent location, and composition from the 1974 eruption, which had a source in the seismic SWRZ. These differences led some authors to propose that the 1971 and 1974 eruptions southwest of the summit originated from different structural domains of the volcano (Duffield and others, 1982; Lockwood and others, 1999). Evidence from deformation suggests that dikes from both the seismic SWRZ and volcanic SWRZ are more or less vertical (Pollard and others, 1983; Dvorak, 1990; Lockwood and others, 1999); thus, the volcanic SWRZ eruptions are probably not a result of south-dipping dikes that originate from the seismic SWRZ and intersect the surface along the volcanic SWRZ trend.

Halemaʻumaʻu–Kīlauea Iki Trend (HKIT)

A shallow magma pathway extends east from Halemaʻumaʻu Crater towards Kīlauea Iki, defining what Hazlett (2002) termed the Halemaʻumaʻu-Kīlauea Iki rift zone (his figure 31) and what we refer to as the Halemaʻumaʻu-Kīlauea Iki Trend (HKIT). Like the volcanic SWRZ, the HKIT is connected to the shallow Halemaʻumaʻu magma reservoir, transports magma within about 1 km of the surface, and is defined by an alignment of eruptive vents and fissures (fig. 21), although no accompanying gravity high is apparent (Kauahikaua and others, 2000). The western end of the HKIT is defined by frequent historical activity adjacent to Halemaʻumaʻu Crater, with eruptions in 1954, 1971, 1975, and 1982 (Holcomb, 1987; Neal and Lockwood, 2003). Farther east, the HKIT is marked by eruptions in 1832 on Byron Ledge (which separates Kīlauea Iki Crater from the deeper part of the caldera), in 1868 within Kīlauea Iki Crater, in 1877 along the east side of the caldera, and in 1959 at Kīlauea Iki. The site of the ʻAilāʻau eruption, which occurred in the 15th century and was ongoing for about 50–60 years (Clague and others, 1999; Swanson and others, 2012a), also lies along that trend immediately east of Kīlauea Iki, and lava flows from ʻAilāʻau would obscure any vents that might have formed prior to the 15th century outside the current caldera. The HKIT is therefore currently expressed only within the caldera, and it is not clear to what extent caldera formation (which occurred after the ʻAilāʻau eruption; Swanson, 2008; Swanson and others, 2012a) might have influenced the geometry of existing or subsequent magma pathways or whether caldera-bounding faults exert control on eruptive vent locations.

Magma transported by way of the HKIT passes through, and may be stored within, the shallow Halemaʻumaʻu magma reservoir before eruption, as demonstrated in 1959. The Kīlauea Iki eruption of that year was fed, at least in part, by a batch of rapidly ascending, primitive magma, as indicated

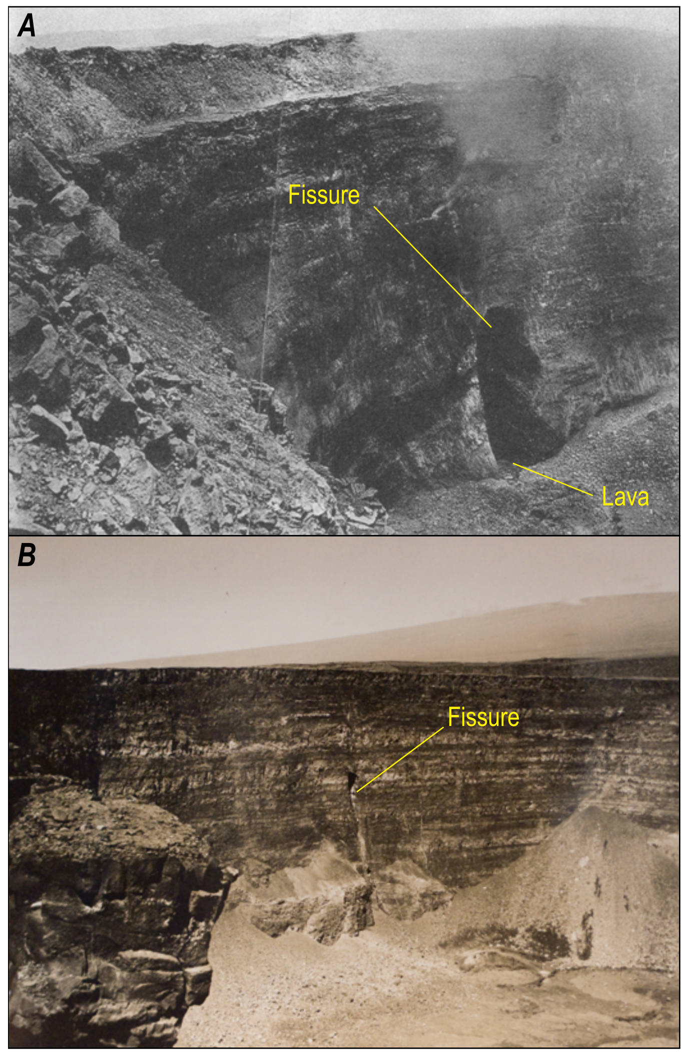


Figure 20. Photos of the southwest wall of Halemaʻumaʻu Crater showing the fissure through which the 1919–20 Maunaiki eruption is thought to have been fed. A, USGS photograph by Thomas A. Jaggar, Jr., taken on May 25, 1922, following a collapse of Halemaʻumaʻu Crater. A small amount of lava is present at the base of the fissure. B, USGS photograph by Howard A. Powers, taken on August 18, 1947, with the south flank of Mauna Loa in the background.

by a deep (~55 km) earthquake swarm three months prior to eruption and summit inflation that began two months before the eruption (Eaton and Murata, 1960). At least two lines of evidence suggest that some of this magma was stored beneath the caldera (and probably in the shallow Halemaʻumaʻu magma reservoir) before erupting. First, deformation and seismicity indicate a source near Halemaʻumaʻu Crater (Eaton and others, 1987; Wright and Klein, 2014), consistent with the location of the shallow reservoir. Second, petrographic and petrologic evidence from 1959 lava and scoria confirm that at least two components—a juvenile magma and a shallowly stored magma that is closely related to lava erupted in 1954—mixed before and during the eruption (Wright, 1973; Helz, 1987; Wright and Klein, 2014; Helz and others, this volume, chap. 6). Melt inclusions from Kīlauea Iki scoria also indicate mixing between a primitive, parental component and a component that had been stored at shallow depths (<3 km); the mixing probably occurred during summit inflation in the two months before the eruption (Anderson and Brown, 1993). The period of storage of the juvenile magma was sufficiently brief and deep that little volatile exsolution (except for CO_2) occurred before the eruption; gas was therefore available to drive high fountaining (Stovall and others, 2012). Lacking concurrent continuous deformation data, it is not possible to ascertain the relation between the 1959 eruption and the HKIT, but we speculate that magma erupted in 1959 utilized this pathway to travel from the storage reservoir to the surface.

Deep Rift Zones

Many models of Kīlauea's magmatic system incorporate a deep (~3–9 km) rift zone below the ERZ and seismic SWRZ that is thought to contain a mix of magma and solidified intrusions. This deep rift zone provides a pathway by which magma may enter the ERZ and seismic SWRZ without passing through summit reservoirs (for example, Ryan, 1988), as well as internal pressure to force the volcano's south flank seaward (for example, Dieterich, 1988; Delaney and others, 1990; Fiske and others, 1993; Cayol and others, 2000; Wright and Klein, 2014). Petrologic evidence has been used to argue that the deep rift zone provides a path for primitive magmas to circumvent the summit reservoir system prior to eruption

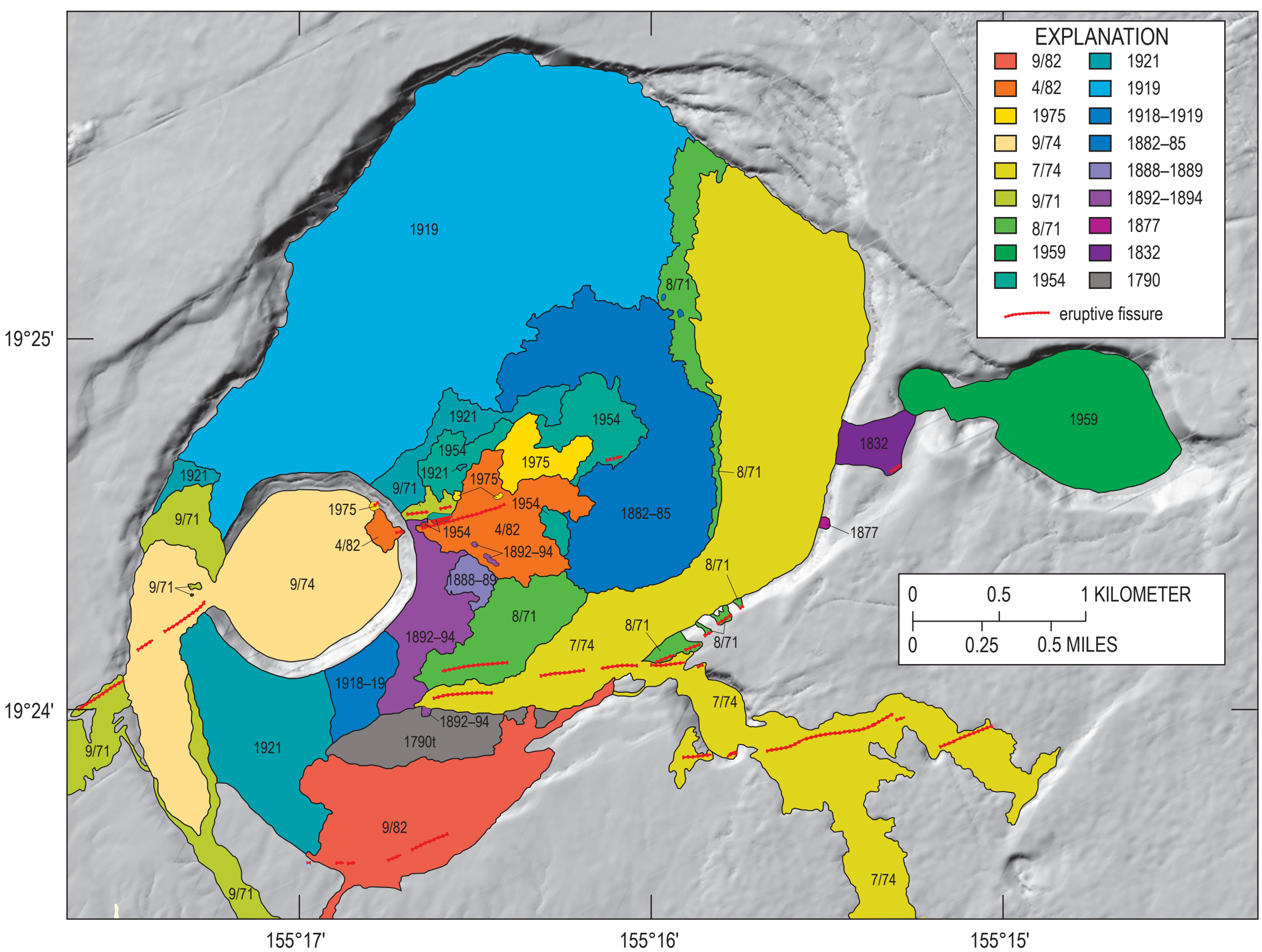


Figure 21. Geologic map showing post-18th century lava flows of the summit region (warm colors represent more recent flows than cool colors) and eruptive fissures (hatched red lines). Tephra deposits, like the 1959 Kīlauea Iki deposit, are not shown. From Neal and Lockwood (2003).

along the ERZ. For example, Trusdell (1991) suggested that picritic (19 percent MgO) magma, which erupted from fissures along the lower ERZ in 1840 (a mixed magma was extruded as part of the same eruption in the middle ERZ), bypassed the summit magma reservoir by way of the deep rift zone. Vinet and Higgins (2010) cited the presence of deformed olivine in some Mauna Ulu lava flows as evidence that magma had moved along the basal décollement to erupt on the ERZ without passing through the summit reservoir. Wright and Helz (1996) proposed that compositional variations in lavas erupted during the 1960 Kapoho eruption reflected magma flow along two different paths from the summit region to the lower ERZ, with various amounts of mixing between magmatic components.

Fundamentally, the petrologic arguments above are based on the inference that the eruption of olivine-rich lavas on the ERZ would not be possible if the magma had passed through the summit reservoir system, or even the shallow part of the rift zone, where it would have mixed with resident magma before erupting. This inference is not necessarily accurate. Most primitive magmas erupted from the ERZ have complex evolutions, having fractionated and (or) mixed with stored magmas at the base of Kīlauea's summit magma system (for example, Helz and others, this volume, chap. 6). Olivine cumulates grow in that area, are available to be picked up by magma passing through the summit and out to the ERZ, and are deformed by flow through the magma plumbing system (Clague and others, 1995). Involvement of the deep rift zone is therefore not necessary for production of an olivine-rich flow with deformed crystals. Indeed, Macdonald (1944) proposed such a model for the 1840 eruption of Kīlauea, with picritic lava extruded from the lower ERZ indicating an origin in the deeper part of the summit reservoir system, while olivine-poor lava that erupted from the middle ERZ came from higher levels of the summit magma complex. Even high-MgO Mauna Ulu lava, thought by Vinet and Higgins (2010) to indicate a source deep beneath Kīlauea's ERZ, has olivine compositions that are consistent with a source no deeper than the base of the summit reservoir (Helz and others, this volume, chap. 6). Melts with high (>14 percent by weight) MgO have been recovered on the Puna Ridge (the submarine extension of the ERZ) and provide evidence that high-MgO magma can occasionally traverse long distances through Kīlauea's plumbing system, perhaps related to periods of heightened magma supply (Clague and others, 1991). Most high-MgO lava on the Puna Ridge, however, show evidence of mixing between volatile-rich and shallowly degassed end members and subsequent fractional crystallization (for example, Dixon and others, 1991), indicative of complex evolution; similar picritic lava is also found on the submarine southward extension of Mauna Loa's Southwest Rift Zone (Helz and others, this volume, chap. 6).

Geophysical evidence for melt in the deep rift zones of Kīlauea is ambiguous. The rift zones are largely aseismic below 4 km, which supports the presence of melt and (or) hot cumulates (Ryan, 1988; Clague and Denlinger, 1994; Denlinger and Okubo, 1995). High b-values in the deep ERZ suggest melt, although the high values extend several kilometers south of the rift zone into areas unlikely to host magma (Wyss and others, 2001). In addition, models of deformation data require opening of the rift zones below about 3 km and above the basal décollement (which defines the base of the volcano's mobile south flank) at ~9-km depth (Dieterich, 1988; Delaney and others, 1990; Owen and others, 1995, 2000a; Cayol and others, 2000; Denlinger and Morgan, this volume, chap. 4). This opening is consistent with magma accumulation and deep rift pressurization—a potential mechanism for seaward motion of Kīlauea's south flank (for example, Dieterich, 1988; Delaney and others, 1990; Cayol and others, 2000; Wright and Klein, 2014).

Gravity and seismic-velocity data contradict evidence for magma in the deep rift zones. Kīlauea's rift zones, particularly the ERZ, are characterized by positive residual gravity anomalies that imply dense intrusions extending from the base of the volcanic pile to high within the edifice (Kauahikaua and others, 2000). Similarly, seismic velocities within the deep ERZ are elevated, relative to adjacent areas, suggesting gabbro-ultramafic cumulates (Okubo and others, 1997; Park and others, 2007, 2009). Low V_p/V_s ratios are also inconsistent with the presence of melt (Hansen and others, 2004). This dense, seismically fast region has been proposed to be a hot cumulate core that is capable of plastic flow and might drive south flank motion by gravitational forces (for example, Clague and Denlinger, 1994; Park and others, 2007, 2009; Plattner and others, 2013; Denlinger and Morgan, this volume, chap. 4). Magnetic data support such an interpretation, as low magnetization at depth below the ERZ could be a result of hot, and possibly altered, rock, as opposed to solidified, more magnetic intrusions in the shallower part of the rift zone (Hildenbrand and others, 1993).

Given the conflicting petrologic and geophysical interpretations, the pressing questions become, is molten magma present in Kīlauea's deep rift zones, and does the deep rift zone provide a pathway for magma to bypass the summit reservoir? These questions can be addressed by three possible models for deep rift zone structure: (1) the deep rift zones are solid and do not allow for magma transport; (2) the deep rift zones are a mix of cumulates and melt and provide a means for magma to bypass the summit reservoir; and (3) the deep rift zones are a mix of cumulates and melt but are fed by downward draining of magma from a molten core that originates from Kīlauea's ~3-km-depth summit magma storage area. Of these models, only the third satisfies geophysical and geological constraints.

While model 1 explains the dense, seismically fast structure of the deep rift zone, it is not consistent with the model for ERZ development. The asymmetric ERZ positive gravity anomaly (with a steeper gradient on the north than on the south) suggests southward migration of the ERZ over time (Swanson and others, 1976a; Kauahikaua and others, 2000). The maximum residual gravity is near the current

ERZ axis, indicating a complex of solidified magma that extends well beneath the transport region of the present rift zones (Kauahikaua and others, 2000). If the ERZ migrated southward over time but no magma were present below 3-km depth, how can solidified intrusions extend to the base of the volcano under the present ERZ? At some point, magma must have been present in the deep rift zones, and the solidification of that magma provides the dense core imaged by gravity and seismic measurements.

Model 2 is compatible with a dense and seismically fast deep rift zone (implying solidified magma and cumulates) that is also aseismic, opening, and has low magnetization (implying the presence of melt, or at least hot and altered rock), but is unsupported by observations from recent eruptions and intrusions. If a semimolten deep rift zone provides a pathway for magma to bypass Kīlauea's summit magma storage areas, why was there no change in the rate of deep rift zone opening during the 2003–07 surge in magma supply (fig. 6)? Similarly, why was there no contraction of the deep rift zone during the 1997 fissure eruption in Nāpau Crater, which was caused by extensional stress from deep rift opening and would presumably have received magma from the deep rift zone (Owen and others, 2000b)?

Model 3 is essentially the same as the molten core model of Johnson (1995b) and Fiske and others (1993). In model 3, the ERZ has a semicontinuous molten core at 3–4-km depth from which magma moves vertically, either upward towards the surface or downward, into the deeper part of the rift zone, and all magma that enters the ERZ passes through the summit reservoir system first. The ERZ is characterized by an interconnected melt zone for at least its entire subaerial length (figs. 5, 10). Olivine-rich magma that erupts on the ERZ does not bypass the summit reservoir but, instead, traverses tens of kilometers, mixing with older batches of rift-stored magma along the way (for example, Wright and Fiske, 1971; Thornber and others, 2003). The Father's Day dike intrusion of June 2007 provided a glimpse of the MgO-rich magma fed into Kīlauea's summit reservoir system. The dike was driven by summit overpressure, and the lava that erupted when the dike reached the surface between the summit and Pu'u 'Ō'ō (after propagating at a depth no greater than 3 km; Montgomery-Brown and others, 2010) was much hotter and richer in MgO than lava that had been erupting from Pu'u 'Ō'ō a few days earlier (fig. 4*D*; Poland and others, 2012). Similarly, the composition of lava extruded three weeks after the start of the 1960 Kapoho eruption was relatively olivine-rich and contained up to 30 percent of the juvenile component of the 1959 Kīlauea Iki eruption, suggesting that 1959 summit magma was able to reach an eruption site about 50-km distant in a few weeks (Wright and Helz, 1996). Finally, if all magma passes through the summit reservoir system, it ascends through cumulates at the base of the summit reservoir complex (Clague and Denlinger, 1994; Clague and others, 1995; Kauahikaua and others, 2000) and may pick up deformed olivine crystals that might later erupt at the summit (for example, Helz, 1987) along the subaerial ERZ (for example, Wright and Helz, 1996; Vinet and Higgins, 2010) or even the submarine extension of the ERZ (Clague and others, 1995), although submarine olivine compositions are more magnesian than those of the summit and subaerial ERZ (Helz and others, this volume, chap. 6). Observing the seismicity and deformation associated with any future submarine eruptions on the Puna Ridge will be invaluable in interpreting the relation between the summit magma system and submarine part of the rift zone.

Downward transport of magma from the molten core explains geophysical measurements that imply the presence of both melt and cumulates extending to the base of the volcano beneath the rift zones. Magma that passed though the summit and intruded along the ERZ but did not erupt would gradually crystallize, with dense minerals retained at depth to create a massive, high-velocity body. Degassed magma, some of which may have drained back into eruptive fissures following extrusion (which is commonly observed in Hawai'i—see, for example, Richter and others, 1970), might also sink to lower levels. The deep rift zone would therefore contain a mix of degassed melt and dense crystals that could explain deep opening modeled from deformation data, the lack of seismicity, low magnetization, high seismic velocities, and positive gravity.

The deep rift zone is ambiguous in character, yet critical to overall models of Kīlauea's magma supply and storage system. For example, whether all magma passes through the summit storage network or intrudes into the deep rift zone and bypasses the summit has important implications for magma supply calculations and CO_2 degassing (see "Magma Supply" section above). Further work is obviously necessary to investigate the characteristics and evolution of the deep rift zones on Hawaiian volcanoes, but we are hopeful that this model will serve as a starting point for studies of Kīlauea's past, present, and future eruptive activity.

Model Summary and Complications

We propose a refined model of Kīlauea's magma plumbing system that includes multiple areas of magma storage beneath the summit, with the primary storage reservoir at ~3–5 km beneath the south caldera and a smaller reservoir ~1–2 km beneath the east margin of Halema'uma'u Crater. Magma is also at least occasionally stored beneath Keanakāko'i Crater. Kīlauea's major rift zones radiate east and southwest from the deeper south caldera reservoir, while shallower pathways extend from the Halema'uma'u reservoir east toward Kīlauea Iki and southwest toward, and beyond, Maunaiki.

Individual elements of this model are long established, having been proposed previously by other authors (for example, Holcomb, 1987; Klein and others, 1987; Fiske and others 1993), but our overall depiction represents an attempt to combine geologic, geochemical, and geophysical observations, especially those collected by seismic and geodetic techniques since the 1990s, into a comprehensive model for Kīlauea.

The model nevertheless remains an oversimplification. For example, we overlook the geometrical complexity of the ERZ, particularly the bend that occurs southeast of the summit that may have formed as a result of southward rift zone migration over time (Swanson and others, 1976a). We chose, instead, to focus on the general configuration of Kīlauea's magma plumbing system as defined geophysically, leaving these and other complications for future study.

We have also neglected important structural elements of the volcano, most notably the Koaʻe Fault System (fig. 10), which is itself poorly understood (Duffield, 1975; Swanson and others, 1976a; Fiske and Swanson 1992; Lockwood and others, 1999) but was recognized by Holcomb (1987) as being associated with a "southern strand" of the SWRZ. Magma was inferred to have intruded the Koaʻe Fault System in 1965 (Fiske and Koyanagi, 1968) and 1999 (Cervelli and others, 2002), and intrusions were confirmed in 1969 (Swanson and others, 1976b) and 1973 (Zablocki, 1978; Tilling and others, 1987), with the 1973 dike extending several kilometers into the central Koaʻe Fault System from the ERZ, based on leveling data (Swanson and others, 2012b). Dikes underlying the Koaʻe Fault System have also been inferred from a variety of geophysical measurements (for example, Flanigan and Long, 1987). Small eruption sites of unknown age, but covering ash deposited in 1790, were discovered in the central part of the Koaʻe by Swanson and others (2012b), demonstrating that these dikes sometimes erupt. The Koaʻe Fault System may therefore structurally and magmatically link the ERZ and seismic SWRZ in a single "breakaway" rift (Fiske and Swanson, 1992) and may one day become an important site for the injection and storage of magma (Fiske and Swanson, 1992; Lockwood and others, 1999). Geophysical monitoring of future intrusions and eruptions in the Koaʻe Fault System will be critical in elucidating the role and character of the region, as well as incorporating its structure into the magmatic model of Kīlauea.

Mauna Loa

Mauna Loa's magmatic system is less well known than Kīlauea's, owing to a comparative lack of geophysically monitored eruptive activity. The modern seismic network on Mauna Loa was established in the 1950s (Okubo, 1995; Okubo and others, this volume, chap. 2), and deformation monitoring began in the 1960s (Decker and others, 1983). These measurements, however, missed the period of frequent Mauna Loa activity—31 eruptions occurred during 1843–1950 (Lockwood and Lipman, 1987), but only two eruptions took place from 1951 to 2014 (in 1975 and 1984). In addition, while geodetic monitoring characterized important changes over time (fig. 22), the network was too sparse and the deformation insufficient in magnitude to facilitate anything except generalized modeling. Despite these shortcomings (compared to Kīlauea), enough measurements exist to provide a rough picture of Mauna Loa's magma plumbing system.

Both deformation and seismicity suggest the existence of a zone of magma storage at 3–4 km beneath the southeast margin of Mokuʻāweoweo Caldera. A magma accumulation zone is hypothesized to correspond to an aseismic region that is capped by earthquakes at about 4-km depth beneath the south-southeastern part of the caldera (Decker and others, 1983; Okubo, 1995). Vertical deformation from leveling during 1977–81 was best modeled by a source of inflation at 3.1-km depth below the southern part of the caldera. Combining vertical, horizontal, and tilt over the same time period suggests a depth of almost 4 km (Decker and others, 1983; Lockwood and others, 1987). The 1984 eruption was accompanied by summit deflation as a dike propagated into, and erupted from, the Northeast Rift Zone. Tilt and leveling data spanning the eruption were modeled by a source of volume loss at about 3.5-km depth beneath the southeast margin of the caldera, coupled with dike opening from 0–5-km depth in the Northeast Rift Zone (Johnson, 1995a). Following the eruption, Electronic Distance Mesurement (EDM) and leveling data indicated inflation with a model location just east of the caldera at 3.7-km depth (Miklius and others, 1995).

An episode of inflation began in May 2002 and was well monitored by both continuous GPS (Miklius and Cervelli, 2003) and InSAR (Amelung and others, 2007). InSAR data spanning 2002–05 are best fit by magma accumulation in two sources: a spherical body located 4.7 km beneath the southeast margin of the caldera and a dike-like structure extending the length of the caldera and into both rift zones, with most opening occurring at 4–8-km depth (Amelung and others, 2007). GPS displacements from 2004 to 2005, the period of most rapid inflation, can be modeled by similar sources and demonstrate the importance of the dike to the overall inflation pattern and rate (fig. 23). It is unknown whether the dike source is a new feature associated with the 2002 onset of inflation or was previously present but unresolvable. Given the limited spatial resolution of the campaign-style measurements (including tilt, leveling, EDM, and GPS) used before the availability of InSAR and continuous GPS, we suspect that the dike source has always been present. The fact that both sources accumulated magma implies that they form a single interconnected magma storage system. During the latter half of the inflation episode (which lasted until 2009), GPS data indicated that the source beneath the southeast caldera was the dominant region of magma storage, suggesting that the dike source may be most active during periods of rapid supply (which was apparently the case during 2004–05).

In addition to regions of magma storage, deformation of Mauna Loa also indicates motion of the volcano's southeast flank. With the establishment of campaign GPS stations around the volcano in the 1990s, surface displacements on Mauna Loa could be measured on a broader scale than was previously possible. These measurements revealed that the southeast flank of Mauna Loa was moving southeast at a rate of about 4 cm/yr (Miklius and others, 1995). Similar to Kīlauea's, such motion may reflect slip along a deep subhorizontal fault underlying the volcano's flank. The role of this flank motion in magmatic and tectonic activity at Mauna

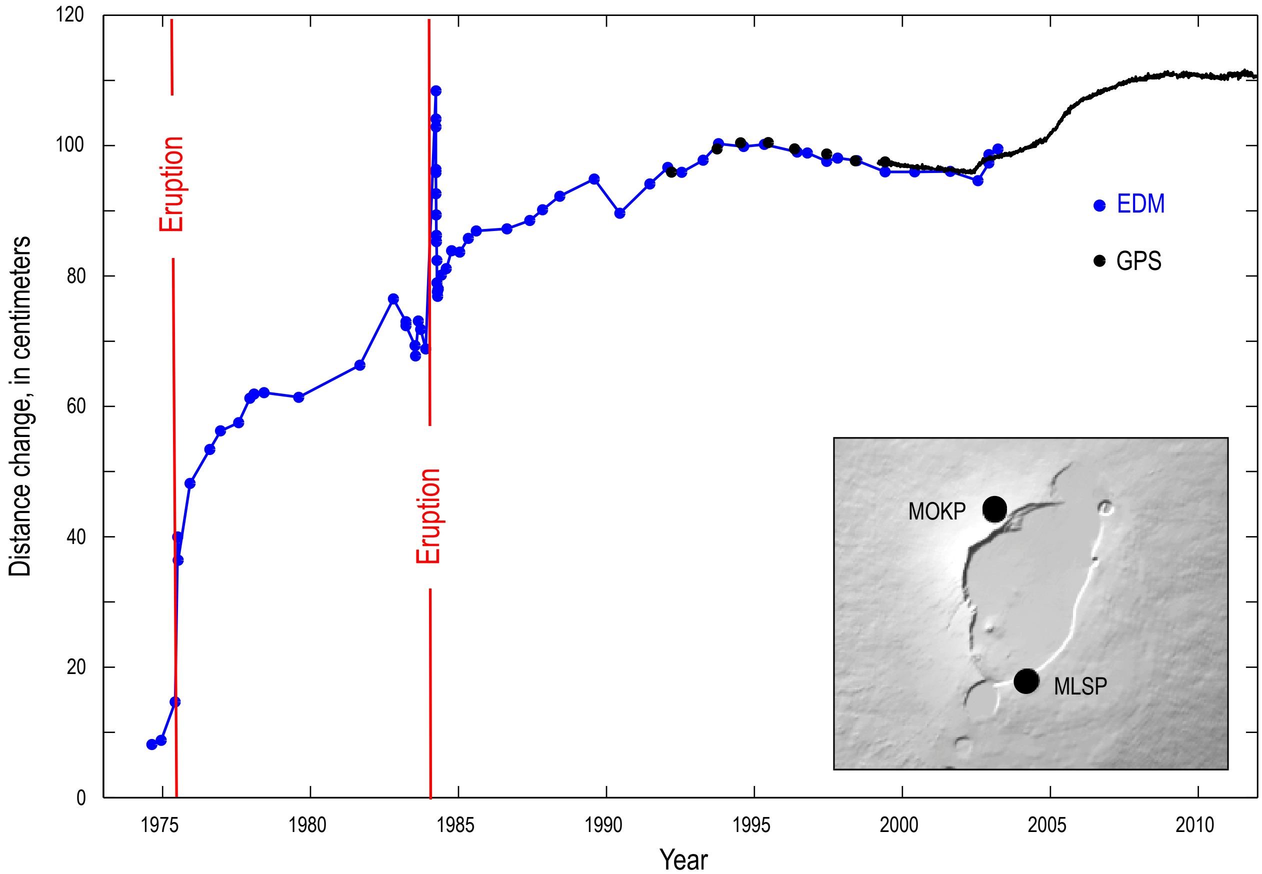


Figure 22. Plot of distance change across Mauna Loa's summit caldera during 1974–2011. Blue dots give Electronic Distance Measurement (EDM) results, black dots are campaign Global Positioning System (GPS) data, and black line is from continuous GPS. Positive change is interpreted as inflation and negative, deflation. Inset map shows station locations and names relative to Mauna Loa's caldera.

Loa is unclear, although stress modeling suggests a feedback where flank earthquakes encourage dike intrusion, and vice versa (Walter and Amelung, 2004, 2006).

Mauna Loa eruptions occur not only from the summit and rift zones, but also from submarine and subaerial vents oriented radially from the summit on the northwest flank of the volcano (Lockwood and Lipman, 1987). The radial vents are located in the area of greatest horizontal tension, where the volcano's stress field is controlled by intrusions into the two rift zones, suggesting that radial dike eruptions occur during periods of heightened magma pressure within the volcano (Rubin, 1990). Petrologic studies of radial vent eruptions have found both primitive high-Mg lava (Riker and others, 2009), as well as evolved alkalic lava (Wanless and others, 2006). If the source magmas for these eruptions ascended through the main Mauna Loa magmatic system, their compositional signatures would probably have been overprinted by the more common tholeiitic magma that undoubtedly dominates the magma plumbing system. Radial vent eruptions may therefore be evidence of secondary magma pathways that bypass the summit plumbing system, allowing evolved magmas to reach the surface (Wanless and others 2006), and (or) evidence of episodic increases in magma supply that allow primitive magmas to erupt (Riker and others, 2009).

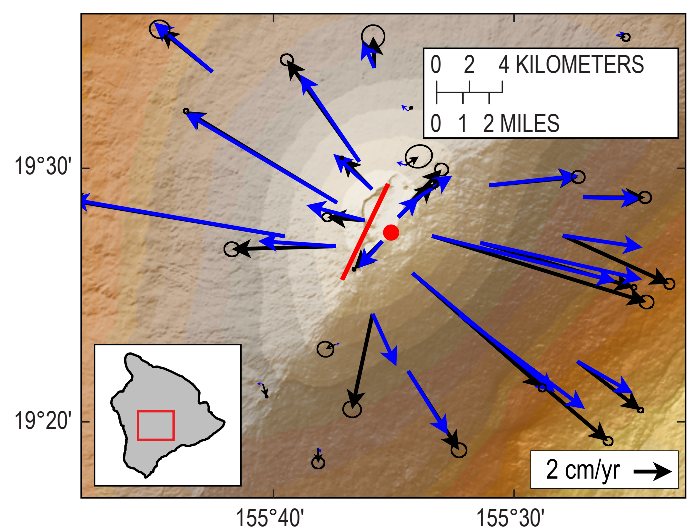


Figure 23. Map of displacements from Global Positioning System data from Mauna Loa during 2004–05. Observed displacements are black, with 95-percent confidence ellipses, and modeled displacements are blue. Model includes a point source at 3.5-km depth beneath the southeast caldera rim (red dot), with a volume increase of 4×10^6 m^3/yr, and a vertical dike along the length of the caldera (red line), with a top at 3.8-km depth (the bottom is difficult to resolve but is probably no more than a few km below the top) and a volume increase of 30×10^6 m^3/yr.

Summary

Frequent eruptive activity and an excellent record of geophysical and geological monitoring during the period 1950–2014 have provided a high-resolution view of Kīlauea's magma plumbing system. Our model of Kīlauea's magma storage areas and transport pathways builds upon previous models (for instance, Eaton and Murata, 1960; Decker and others, 1987; Holcomb, 1987; Klein and others, 1987; Ryan, 1988; Fiske and others, 1993; Tilling and Dvorak, 1993) and incorporates data collected by new techniques (for example, continuous GPS and InSAR). The improved resolution of the different components of Kīlauea's plumbing system provided by our conceptual model offers a new framework for interpreting volcanic and tectonic activity at the volcano. Although less active during the same time period, the general outline of Mauna Loa's magmatic system should prove valuable for understanding future volcanic and seismic activity at that volcano.

Curiously, both Kīlauea and Mauna Loa are characterized by primary magma storage areas located about 3–4 km beneath their summit calderas. This correspondence suggests that the magmatic systems of Hawaiian volcanoes migrate upward as the volcanoes grow (Decker and others, 1983; Lockwood and others, 1987; Dvorak and Dzurisin, 1997), and that the 3–4-km depth represents a favorable level of magma accumulation, possibly because it is a level of neutral buoyancy (Ryan, 1987; Burov and others, 2003). Magma storage zones exist at similar depths at other basaltic shields—for example, Piton de la Fournaise (Peltier and others, 2009), Etna (Aloisi and others, 2011), and Axial seamount (Chadwick and others, 1999; Nooner and Chadwick, 2009)—implying that levels of neutral buoyancy and magma trapping may be comparable at large basaltic volcanoes worldwide. Such volcanoes are also often characterized by multiple, vertically stacked magma reservoirs and rift zones that radiate from the summit, resembling the model of Kīlauea's magma system.

Magma Transport

At basaltic volcanoes, transport of magma away from a source reservoir is most often accomplished by means of dikes and sills. Intrusion initiation, propagation, and eruption have been the subjects of a large body of literature, a complete (or even partial) review of which is impractical. We therefore refer the reader to articles cited within this section for additional information, although these few articles only scratch the surface of research into the process of intrusion. We also focus specifically on dike intrusion, which is common at Hawaiian volcanoes, although we note that other basaltic systems seem to be dominated by sill intrusion over dikes (for example, Bagnardi and others, 2013).

Dike propagation occurs both vertically and laterally and is controlled by numerous factors, the most important of which include magma pressure, the preexisting stress field, magma viscosity, and host rock properties (for example, Rubin, 1995; Taisne and Jaupart, 2009; Traversa and others, 2010). When dike driving pressure (defined as the difference between the magma pressure and the least compressive stress in the crust) is sufficiently high, the dike will reach the surface and erupt; however, geologic evidence suggests that most dikes are arrested before they reach the surface, probably because mechanical and stress barriers inhibit propagation (for example, Gudmundsson and others, 1999; Gudmundsson, 2002; Rivalta and others, 2005; Taisne and Tait, 2009; Taisne and others, 2011a; Geshi and others, 2012). Understanding the conditions that favor dike eruption versus arrest is therefore critical for volcano monitoring and hazards mitigation (for example, Geshi and others, 2010).

Dike emplacement is accompanied by both deformation and seismicity. For example, earthquakes indicated lateral emplacement of dikes at Krafla, Iceland (Brandsdóttir and Einarsson, 1979; Einarsson and Brandsdóttir, 1980), and at Miyakejima, Japan (Toda and others, 2002), as well as vertical ascent of magma at Piton de la Fournaise (Battaglia and others, 2005). Seismicity and tilt were used to follow the transition from vertical to lateral propagation of dikes at Etna (Aloisi and others, 2006) and Piton de la Fournaise (Toutain and others, 1992; Peltier and others, 2005). Rates of dike propagation in these and similar cases are on the order of 0.1–1 km/hr, occasionally reaching several kilometers per hour (for example, Dvorak and Dzurisin, 1997; Grandin and others, 2011; Wright and others, 2012) and sometimes displaying a complex propagation history in both rate and direction (for example, Taisne and others, 2011b). Traversa and Grasso (2009) and Traversa and others (2010) noted that earthquakes may not necessarily track the propagation of the dike tip. In some cases, however, earthquakes must indicate the location of a dike's leading edge—in Iceland in 1977, a small eruption occurred through a drill hole when seismicity associated with a propagating dike reached the hole (Brandsdóttir and Einarsson, 1979).

Exposures within the cores of eroded Hawaiian islands, such as O‘ahu, illustrate the importance of dike intrusions to volcano construction and evolution (Walker, 1986, 1987), and many of the primary characteristics of active dike emplacement were deduced from Hawaiian examples (for instance, Pollard and others, 1983). Exceptional monitoring data and frequent episodes of dike emplacement, especially at Kīlauea, provide views into the process of dike formation and intrusion (for example, Segall and others, 2001; Rivalta, 2010). Flow of magma through conduits can also be studied, thanks to long-lived eruptive activity (for example, Cervelli and Miklius, 2003). Here, we summarize observations of active magma transport in Hawai‘i and explore the mechanisms for the formation of dikes within Hawaiian volcanoes.

Characteristics of Magma Transport at Hawaiian Volcanoes

Dike emplacement is the most important means of transporting magma from source reservoirs to intrusion and eruption sites at shield-stage Hawaiian volcanoes. The eroded Koʻolau volcano, on the Island of Oʻahu, displays thousands of dikes that, at outcrop or map scale, make up 50–65 percent of the host rock. These dikes form highly concentrated swarms, similar to oceanic sheeted dike complexes, that delineate rift zones in which intrusion is concentrated (Walker, 1986, 1987), with Hawaiian rift zones apparently formed and maintained by gravitational forces (for example, Fiske and Jackson, 1972).

Compared to dikes exposed by erosion in Iceland—another region of frequent basaltic intrusions—those of Hawaiʻi are thin. This observation led Rubin (1990) to suggest that Hawaiian dikes are driven by higher magma pressures than their Icelandic counterparts (which are more heavily influenced by tectonic stress), and therefore are more likely to reach the surface—a model later supported by Segall and others (2001). During 1960–75, dikes reached the surface at Kīlauea about twice as often as they stalled (Dzurisin and others, 1984)—a proportion much greater than that implied by geologic studies in Iceland (for example, Gudmundsson and others, 1999). Changes in the stress state of the volcano directly influence the proportion of magma that is stored versus erupted. For example, in the months following the *M*7.7 earthquake beneath Kīlauea's south flank in 1975 (Nettles and Ekström, 2004; Owen and Bürgmann, 2006), the ratio of eruptions to noneruptive intrusions changed from 4:1 to 1:4, presumably because dilation of the volcano's magma system caused by the earthquake promoted magma storage over eruption (Dzurisin and others, 1980, 1984; Cayol and others, 2000). A similar relation was found at Piton de la Fournaise, which experienced a period dominated by noneruptive intrusions following a caldera collapse in 2007 (Peltier and others, 2010).

Rates of dike propagation at Kīlauea and Mauna Loa are comparable to those determined at other basaltic volcanoes. At Mauna Loa, lateral dike propagation associated with the 1984 Northeast Rift Zone eruption occurred at a rate of 1.2 km/hr, based on the appearance of eruptive fissures (Lockwood and others, 1987). The relative lack of seismicity during the dike propagation was attributed by Lockwood and others (1987) to magma flow into a mostly open conduit created by an intrusion in 1975 (which immediately followed the summit eruption of that year) or perhaps to a stress drop caused by a nearby *M*6.6 earthquake in 1983. In contrast, Kīlauea has experienced tens of intrusive episodes since the start of dense geophysical monitoring in the 1950 and 1960s. Klein and others (1987) provided a comprehensive overview of Kīlauea seismicity during 1963–83, subdividing earthquake swarms into those that were associated with magma intrusion and eruption, inflation of the magma system, and constant flow of magma from the summit into the rift zones. Seismic swarms due to rapid magma intrusion or eruption have migration speeds ranging from 0.1 to 6 km/hr. Propagation directions are generally away from the summit, although uprift migrations of seismicity have been observed within the ERZ between the summit and Mauna Ulu, possibly reflecting a complex interplay between varying stresses, pressure gradients, and magma conduits (Klein and others, 1987).

Numerous magmatic intrusions have occurred in Kīlauea's ERZ since 1983, despite nearly continuous eruptive activity (ongoing as of 2014) and the existence of a magma conduit between the summit and ERZ eruption site (fig. 24; see "Mechanisms of Dike Emplacement at Hawaiian Volcanoes" section for a discussion of intrusion driving forces). Five intrusions into the upper ERZ took place during 1990–93, some of which were associated with pauses as long as 8 days in the ongoing eruption (Heliker and Mattox, 2003). A fissure eruption in Nāpau Crater in 1997 caused collapse of Puʻu ʻŌʻō and a pause of 24 days (Owen and others, 2000b; Heliker and Mattox, 2003; Thornber and others, 2003), and dikes in 1999 and 2000 intruded the ERZ but failed to erupt (Cervelli and others, 2002; Heliker and Mattox, 2003), although they caused the ongoing eruption to pause for 11 days and 7 hours, respectively. A dike consisting of two discrete segments intruded between Mauna Ulu and Nāpau Crater and resulted in a small eruption on the northeast flank of Kanenuiohamo in June 2007 (Poland and others, 2008; Montgomery-Brown and others, 2010; Fee and others, 2011). The intrusion propagated downrift discontinuously in a series of pulses, as indicated by bursts of seismicity coincident with increases in the rate of summit deflation, with periods of propagation characterized by rates of about 3 km/hr (fig. 25). During March 2011, the ongoing ERZ eruption was again interrupted, this time by a 5-day fissure eruption between Nāpau Crater and Puʻu ʻŌʻō (Lundgren and others, 2013).

These dikes followed the trends established by previous intrusions at Kīlauea. The 1997 and 2011 fissure eruptions were preceded by only a few hours of intense and localized seismic activity, suggesting that a dike propagated from the existing ERZ magma conduit at ~3-km depth to the surface at an average rate of about 1 km/hr (Owen and others, 2000b; Thornber and others, 2003). Tilt records over the 1999 intrusion suggest a similar velocity in a downrift direction (Cervelli and others, 2002).

In addition to hosting repeated dike intrusions, Kīlauea provides an opportunity to study magma flow through established conduits, thanks to the occurrence of long-term eruptive activity. Such a conduit has existed since the start of the 1983–present (as of 2014) ERZ eruption and has evolved over time from an episodically active to a continuously active pathway, possibly due to the relation between cooling of magma within the conduit and pressure in the summit magma reservoir system (Parfitt and Wilson, 1994). The primary vent for the eruption, Puʻu ʻŌʻō, is located about 20 km from Halemaʻumaʻu Crater along the curving trend of the ERZ. Given that the average

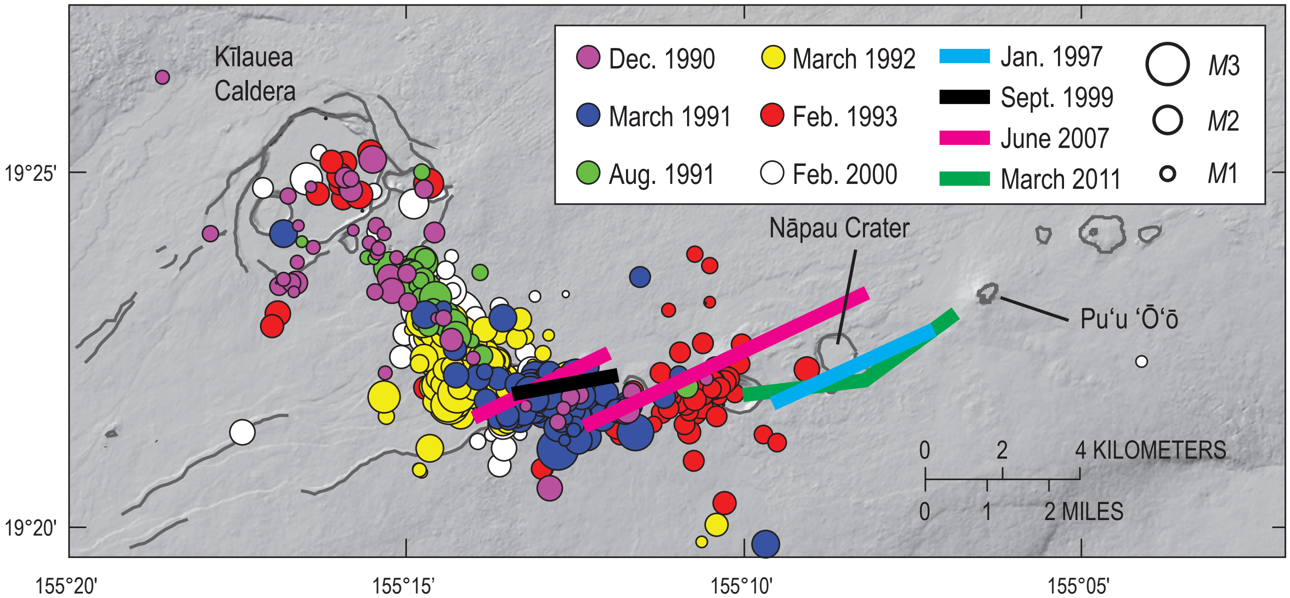
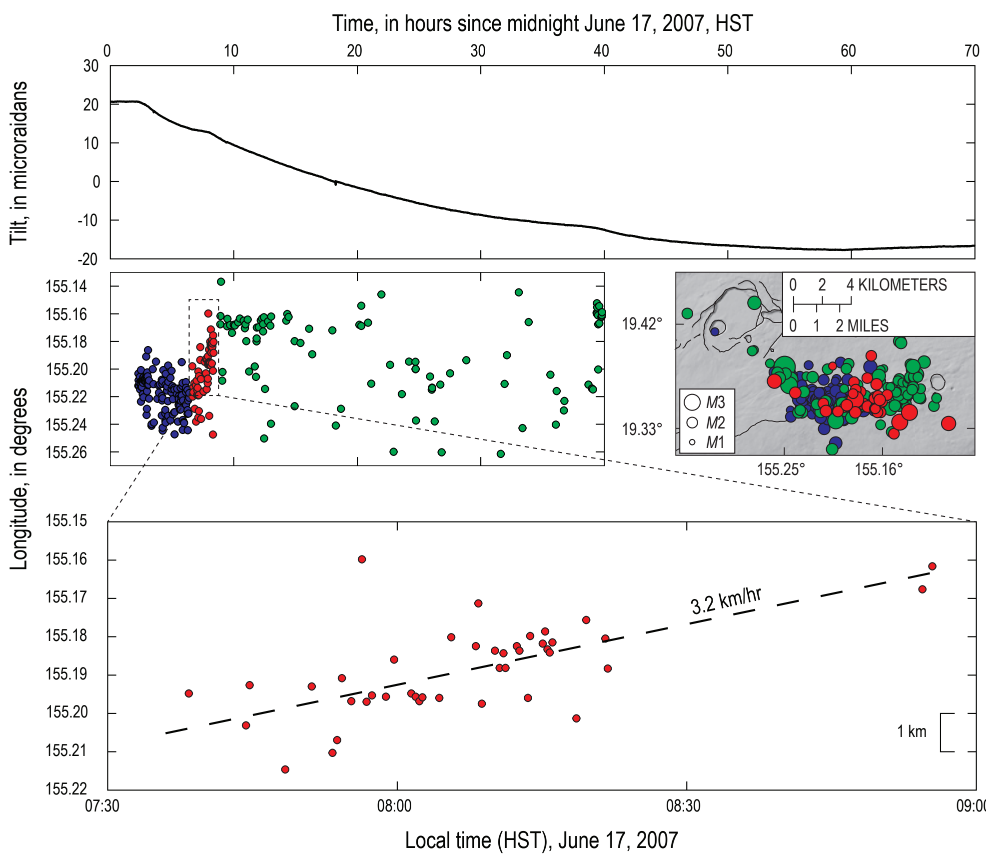


Figure 24. Map of Kīlauea's East Rift Zone showing intrusive activity during 1990–2011. Seismicity (colored circles) is shown for intrusions for which little or no geodetic data are available. Model geometries (colored lines) are given for those dikes that were detected by deformation measurements (tilt, Global Positioning System data, and (or) Interferometric Synthetic Aperature Radar), including the 1997 Nāpau fissure eruption (Owen and others, 2000b), 1999 noneruptive dike (Cervelli and others, 2002), 2007 Father's Day intrusion/eruption (Montgomery-Brown and others, 2010), and 2011 Kamoamoa fissure eruption (Lundgren and others, 2013).

Figure 25. Plots of tilt and seismicity associated with Kīlauea's June 17–19, 2007, East Rift Zone intrusion and eruption. Tilt time series (top) is from a station located about 300 m west-northwest of the Hawaiian Volcano Observatory. Tilt azimuth is 327°, which is approximately radial to the source location, implying that positive tilt is inflation and negative tilt is deflation (assuming a spherical source). Seismicity (middle) occurred along an approximately east-west trend, so longitude is roughly perpendicular to the strike of the dike. Map shows distribution of earthquakes. Earthquake colors on map and time series indicate timing (blue, 00:00–07:30 HST on June 17; red, 07:30–09:00 on June 17; green, 09:30 on June 17 to the end of June 19). The most rapid propagation, during 07:30–09:00 HST on June 17 (zoomed plot at bottom), occurred at a rate of about 3.2 km/hr (the two earthquakes at the end of the zoomed plot do not significantly influence the trend line).



propagation rate of a dike at Kīlauea is about 1 km/hr (Klein and others, 1987), one can reasonably infer that the transport rate of magma along an open conduit is no slower than that, and that magma would take no more than 20 hours to travel from the summit to Puʻu ʻŌʻō. In fact, the rate of transport is possibly much higher, as suggested by ground tilt at the summit and at Puʻu ʻŌʻō (Cervelli and Miklius, 2003).

Transient tilt events consisting of deflation-inflation-deflation (DID) or deflation-inflation (DI) cycles (fig. 26) have been recorded at Kīlauea since the installation of borehole tiltmeters in 1999 (Cervelli and Miklius, 2003; Anderson and others, in press). The general pattern includes summit deflation, followed within minutes to hours by deflation at Puʻu ʻŌʻō and a waning of eruptive activity. After a period of hours to a few days, inflation begins at the summit and, with a lag of minutes to hours, at Puʻu ʻŌʻō (sometimes accompanied by a surge in lava effusion). DID and DI events were probably occurring before 1999, based on summit tilt cycles that correlated with pauses and surges in lava effusion from Puʻu ʻŌʻō, but tilt measurements were not available from Puʻu ʻŌʻō o for confirmation (Heliker and Mattox, 2003). The coincidence between changes in tilt at the summit and at Puʻu ʻŌʻō implies a hydraulic connection between the vents (for example, Dvorak and Okamura, 1987), and the time lag may represent the flow rate of magma through the conduit for at least three reasons: (1) summit and Puʻu ʻŌʻō deflation indicate that pressure within the conduit was low before the onset of inflation; (2) if the inflation were caused by a pressure wave, the pressure pulse would approximate the speed of a P-wave through magma, arriving at Puʻu ʻŌʻō a few seconds after occurring at the summit (Cervelli and Miklius, 2003); and (3) a poroelastic effect (for example, Lu and others, 2002) is unlikely, given the presence of an open magma conduit between the two locations. Alternatively, the delay time may be a function of mechanical factors, like dike-wall-rock elasticity and magma viscosity, as modeled by Montagna and Gonnermann (2013).

The delay between the onset of inflation at the summit and Puʻu ʻŌʻō during DID and DI events generally falls in the range of 1–2 hours (fig. 26). If related to magma transport, the delay time suggests a magma flow rate of 10–20 km/hr in the ERZ conduit—similar to the gravity-driven velocity of lava flowing through tubes (for example, Kauahikaua and others, 1998), and a few times higher than flow rates modeled for early lava fountaining episodes of the Puʻu ʻŌʻō eruption (Hardee, 1987). Such a simple model, however, does not account for magma that is already in the conduit and that would be pushed at the front of a surge from the summit (for example, Parfitt and Wilson, 1994), implying a shorter transport time than indicated by the tilt patterns. The model also ignores the mechanism for varying delay times (which can reach up to many hours) and for DI events at the summit that were not reflected at Puʻu ʻŌʻō during 2005–07 despite continued ERZ eruptive activity. The time lag in tilt between the summit and Puʻu ʻŌʻō therefore represents a fast end member for the transport time along the ERZ.

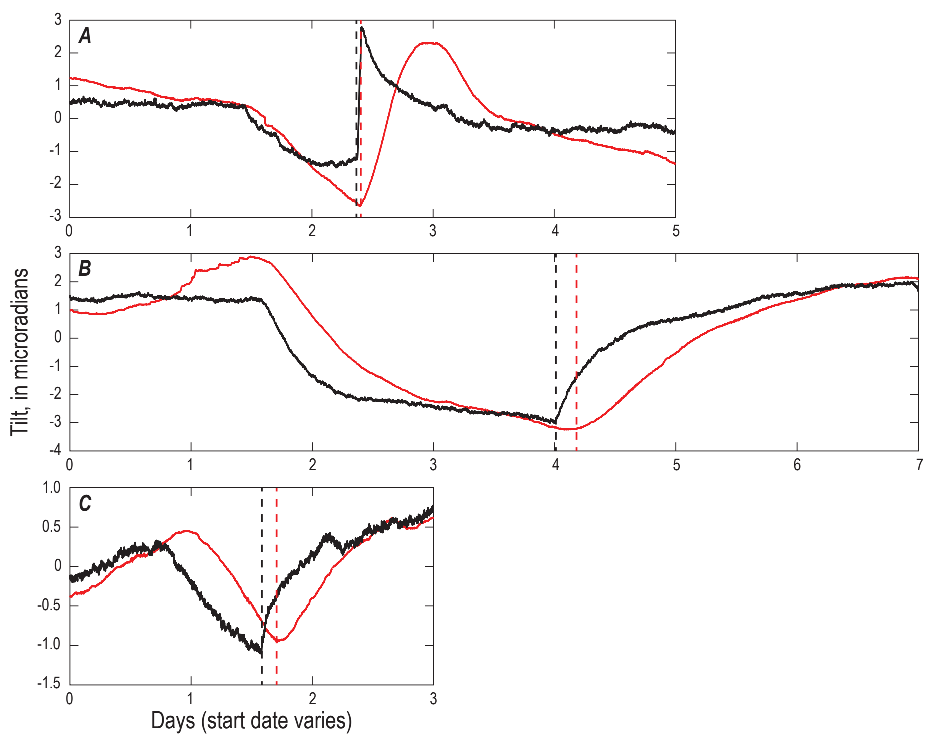


Figure 26. Plots of some example transient tilt events at Kīlauea's summit and Puʻu ʻŌʻō. Summit tiltmeter (black line) is located about 300 m west-northwest of the Hawaiian Volcano Observatory, and tilt azimuth is 327° (northwest). Puʻu ʻŌʻō tiltmeter (red line) is located on the north flank of the eruptive cone, and tilt azimuth is 308°. Both tilt azimuths are radial to deformation sources (assumed spherical), so that positive tilt changes are interpreted to be inflation and negative changes are deflation. The time axes for all three plots are at the same scale for ease of comparison. *A*, Tilt record spanning December 8–12, 2001, showing a deflation-inflation-deflation (DID) event. The time between the onset of inflation at Puʻu ʻŌʻō (dashed red line), relative to the summit (dashed black line), is approximately 36 minutes. *B*, Tilt record spanning March 21–27, 2010, showing a deflation-inflation (DI) event with a U-like shape. The time between the onset of inflation at Puʻu ʻŌʻō relative to the summit is approximately 271 minutes. *C*, Tilt record spanning September 27–29, 2010, showing a DI event with a V-like shape. The time between the onset of inflation at Puʻu ʻŌʻō, relative to the summit, is approximately 177 minutes.

Mechanisms of Dike Emplacement at Hawaiian Volcanoes

The driving pressure of a dike can be enhanced by either increasing the magma pressure in the dike (active intrusion) or by decreasing the least compressive stress (σ_3) that opposes dike opening (passive intrusion). Both conditions facilitate intrusion, but eruption is more likely when magma pressure is high (Rubin, 1990). At Hawai‘i's active volcanoes, dikes have been observed to have characteristics suggesting emplacement due to both increased magma pressure and decreased σ_3.

Dikes that are initiated due to high magma pressure commonly occur after periods of repose at Kīlauea and Mauna Loa and are usually preceded by summit inflation and (or) earthquake activity. For example, both the 1975 and 1984 eruptions of Mauna Loa were heralded by about a year of elevated seismicity, although cross-caldera extension consistent with inflation was observed only before the 1975 eruption (Koyanagi and others, 1975; Lockwood and others, 1987; fig. 22). At Kīlauea, summit inflation is a common precursor to dike intrusions, as demonstrated by summit tiltmeter records (fig. 12), and is usually accompanied by seismicity (Klein and others, 1987). The 1959 Kīlauea Iki eruption provides an excellent example. Following a swarm of deep earthquakes, Kīlauea's summit began to inflate in August 1959. The inflation accelerated in October and November, and a swarm of small, shallow earthquakes began and increased in number as the inflation rate accelerated. This activity culminated in an eruption on November 14 at Kīlauea Iki (Eaton and Murata, 1960).

Passive intrusions were documented in 1997 and 1999 during the 1983–present (as of 2014) ERZ eruption of Kīlauea. The 1997 intrusion resulted in a 22-hour-long eruption in Nāpau Crater (Owen and others, 2000b; Thornber and others, 2003), but the 1999 dike did not breach the surface (Cervelli and others, 2002). The intrusions are thought to be a consequence of shallow rift zone opening in response to a buildup of extensional stress above the opening deep rift zone. The extension is ultimately caused by long-term seaward motion of Kīlauea's south flank (Owen and others, 2000b; Denlinger and Morgan, this volume, chap. 4), possibly enhanced by the presence of preexisting melt pockets within a few kilometers of the surface that create zones of weakness (Thornber and others, 2003). Such a model is supported by the fact that neither the 1997 nor the 1999 intrusion was preceded by summit inflation; instead, the summit was deflating prior to both events, suggesting that increasing magma pressure did not trigger dike emplacement (Owen and others, 2000b; Cervelli and others, 2002). Seismicity during these two intrusions was localized along the ERZ and did not propagate downrift from the summit—additional evidence that the intrusion process was initiated from the ERZ conduit.

The presence of an active magma conduit between the summit and Pu‘u ‘Ō‘ō is probably a prerequisite for these passive intrusions, since much of the magma supplied to the 1997 and 1999 dikes came from the conduit system itself (Owen and others, 2000b; Cervelli and others, 2002). In addition, stress modeling of the 1999 intrusion based on a model of long-term GPS velocities from Owen and others (2000a) indicates extensional stress favoring dike opening in the area of the ERZ where the dike actually intruded (Cervelli and others, 2002). Passive dike initiation and ascent is not unique to the time period of the Pu‘u ‘Ō‘ō eruption but probably also occurred at other times in Kīlauea's history. For example, negligible summit deformation was recorded before the 1955 (Helz and Wright, 1992; Wright and Klein, 2014) and September 1977 (Dzurisin and others, 1980, Wright and Klein, 2014) ERZ eruptions, both of which involved fractionated magmas thought to have been stored within the rift zone for years to decades (Wright and Klein, 2014).

The June 2007 Father's Day intrusion and eruption provides an example of a hybrid type of dike intrusion in which both magma pressure and extensional stress contributed to dike growth. Summarized here is the chronology and interpretation of activity (Poland and others, 2008; Montgomery-Brown and others 2010, 2011). At 02:16 HST on June 17, 2007, summit tiltmeters began recording rapid deflation as an earthquake swarm began on the ERZ, indicating magma withdrawal from the summit and intrusion into the ERZ near Pauahi Crater. After ~6 hours, seismicity jumped downrift, and the rate of summit deflation increased as the intrusion migrated to the east (fig. 25). During the intrusion, GPS stations on the south flank of Kīlauea suggested the occurrence of an aseismic slip event along the basal décollement (Brooks and others, 2008; Montgomery-Brown and others, 2010, 2011). Just after midnight HST on June 19, a small-volume (~1,500 m^3) fissure eruption of high-MgO basalt (compared to what had been erupting from Pu‘u ‘Ō‘ō) began on the northeast flank of Kanenuiohamo (Poland and others, 2008, 2012; Fee and others, 2011). Summit deformation reversed to inflation at about 10:30 HST on June 19, indicating an end to the Father's Day intrusive and eruptive activity. Models of deformation data suggest that the intrusion comprised two en echelon dike segments (Montgomery-Brown and others, 2010).

The Father's Day event was preceded by an increase in magma pressure at the summit of Kīlauea, as suggested by summit inflation that had been ongoing since 2003, due to an increase in magma supply to the volcano (see "2003–07 Increase in Magma Supply to Kīlauea" section above and Poland and others, 2012). The magma pressure finally ruptured the already-full ERZ conduit early on June 17, 2007, and magma was pushed from the summit into the ERZ, forming the western dike segment (using the terminology of Montgomery-Brown and others, 2010). The fact that summit deflation occurred contemporaneously with the onset of the intrusion, instead of lagging by minutes or hours, also argues for summit magma pressure (and not extension due to south flank motion) as a trigger for dike initiation.

The decreasing rate of summit deflation after the first 6 hours indicates that growth of the intrusion was waning when the aseismic décollement slip event began at about 08:00 HST on June 17. Stress models suggest that the décollement slip may have been triggered by the intrusion (Brooks and others, 2008). Once begun, the décollement slip would have promoted additional dike opening, possibly prompting the jump of dike emplacement to the eastern segment (Montgomery-Brown and others, 2010), which ultimately became larger in volume and had a longer period of growth than the western segment. Feedback between dike opening and décollement slip therefore combined to drive much of the Father's Day dike emplacement activity, demonstrating that the event was a combination of active and passive processes.

Summary

Dikes are the dominant means of initial magma transport within shield-stage Hawaiian volcanoes away from the summit reservoir system. Most dikes in Hawai‘i appear to be driven by high magma pressure, as opposed to low levels of stress in the crust that oppose dike opening, which means that many Hawaiian dikes erupt. This condition can be reversed (intrusion favored over eruption) during times of enhanced extensional stress, as occurred following the 1975 earthquake beneath Kīlauea's south flank, but such periods apparently last only months to years. Rates of dike propagation in Hawai‘i are similar to those measured at other basaltic volcanoes worldwide (generally about 1 km/hr). During long-term eruptions, the initial dikes evolve into open pipe-like conduits, and magma flow rates may reach 10–20 km/hr based on tilt measurements during transient deformation events.

Dike emplacement in Hawai‘i is a consequence of both active and passive processes. Active intrusions, common following periods of intrusive/eruptive repose, are driven by magma pressure in summit reservoirs at Kīlauea and Mauna Loa. Passive intrusions, in contrast, are driven by extensional stress (especially prevalent at Kīlauea due to seaward motion of the volcano's south flank), the best examples of which include the 1997 and 1999 ERZ dikes at Kīlauea. Feedback between dike emplacement and flank slip creates intrusions that are a consequence of both magma pressure and extensional stress, like the 2007 Father's Day intrusion and eruption along Kīlauea's ERZ.

Data collected on dike emplacement in Hawai‘i provide important insights into magma transport at basaltic volcanoes. Perhaps most critical from a hazards perspective is that active intrusions are preceded by summit inflation, earthquake swarms, and other activity, making them straightforward to forecast (although predicting their precise timing remains elusive). Previous work has demonstrated as much—for example, Koyanagi and others (1975) correctly interpreted the buildup in seismicity at Mauna Loa during 1974–75 as a precursor to eruption (the eruption occurred shortly after their manuscript had been submitted for publication). Passive intrusions, on the other hand, are a consequence of extensional stress buildups and have no obvious precursors. The likely location of such activity can be forecast from stress models (Cervelli and others, 2002), but their general timing cannot, as yet, be anticipated.

Conclusions

In the more than 100 years since the Hawaiian Volcano Observatory was founded by Thomas A. Jaggar, Jr., in 1912, a wide array of volcanic and tectonic activity has been tracked in Hawai‘i. Monitoring of Kīlauea has resulted in one of the most comprehensive sets of geological, geophysical, and geochemical data available for any volcano in the world. Mauna Loa is nearly as well observed, especially since the establishment of seismic monitoring in the 1950s and continuous deformation networks in the 1990s. These data provide a context for research into magma supply to, and storage and transport within, shield stage basaltic volcanoes.

Examinations of the topography, bathymetry, and gravity of the Hawaiian-Emperor chain of islands and seamounts have allowed for the reconstruction of the magma supply from the hot spot over time. The rate of supply has fluctuated over millions of years, with the present representing a peak in supply rate. Measurements of deformation and erupted volumes suggest a relatively constant supply rate to Kīlauea over decades, although supply appears to fluctuate between Mauna Loa and Kīlauea. For example, between 1952 and 2014, Kīlauea has been almost continuously active while Mauna Loa has erupted only twice (individual eruptions may be large in volume, but Mauna Loa's time-averaged effusion rate has dropped; Lockwood and Lipman, 1987). The supply rate to Kīlauea from the mantle more than doubled during 2003–07, indicating that short-term changes in supply do occur. The 2003–07 increase in supply caused important changes in volcanic and tectonic activity at the surface and was preceded by seismic activity, inflation, and increased gas emissions. It should be possible to forecast future changes in supply by tracking these indicators, especially CO_2 emissions. During periods of heightened magma supply, both Mauna Loa and Kīlauea may receive magma from the hot spot, as indicated by inflation at Mauna Loa during 2002–09, suggesting a deep connection between the hot spot source of magma, Kīlauea, and Mauna Loa. The lack of significant CO_2 emissions from Mauna Loa raises the intriguing possibility that most of the CO_2 degassed by magma supplied from the hot spot may ascend through, and be emitted from, Kīlauea.

Magma storage within Hawaiian volcanoes is accommodated by a complex series of interconnected reservoirs and pathways. We introduce a refined model for Kīlauea's magma plumbing system based largely on past studies, combined with more recent seismic imaging and deformation data collected by GPS and InSAR. The model is consistent with geochemical data and previous geophysical results. There are at least two long-term (that is, decades or longer) magma storage areas beneath Kīlauea's summit, each of which is connected to a rift zone system. The deeper summit magma reservoir, located beneath the south caldera, is the primary magma storage area for Kīlauea and feeds magma into the East and seismic Southwest Rift Zones at about 3-km depth; both rift zones themselves contain isolated pockets of stored, crystallizing magma emplaced at earlier times. A shallower reservoir, beneath the east margin of Halema‘uma‘u

Crater, transports magma to summit eruptive vents, as well as to the east (toward Kīlauea Iki) and southwest (toward Maunaiki) at depths of 1 km or less. Magma also accumulates episodically beneath Keanakāko‘i Crater. The question of magma storage in Kīlauea’s deep rift zones is still a matter of uncertainty and is an important topic for future research.

Mauna Loa’s magma system appears less complicated, but a lack of eruptive activity there since the advent of high spatial- and temporal-resolution deformation measurements limits our ability to detect all elements of the volcano’s plumbing system. The main magma reservoirs beneath Mauna Loa are located beneath the southeast margin of the caldera and along the length of the caldera, probably reflecting an interconnected storage area with complex geometry. These conceptual models of the current magma plumbing systems of Kīlauea and Mauna Loa will no doubt be refined as new geophysical and geochemical techniques are developed and applied, and as future eruptive activity offers additional opportunities to study magma storage. For now, they provide a framework for interpreting magmatic and volcanic activity in Hawai‘i and, by extension, other basaltic volcanoes.

Given their frequent eruptions, Hawaiian volcanoes afford an unprecedented opportunity to observe magma transport. Studies of magma intrusion in Hawai‘i have provided much of the foundation for understanding dike emplacement in general, from the mechanics of propagation to the seismic and geodetic changes that result. During long-lasting eruptions, dikes evolve into open conduits, and magma flows freely between storage areas and the eruption site. Dike intrusion is triggered by both an increase in magma pressure within a storage reservoir (which generally occurs following periods of repose or during surges in supply) and buildups in extensional stress (generally due to flank motion), even during ongoing eruptions, when an open magma conduit already exists. These two processes reinforce each other, resulting in complex triggering relations between magmatic and tectonic activity, as demonstrated by the June 2007 Father’s Day intrusion and eruption at Kīlauea.

During the first century of the Hawaiian Volcano Observatory’s operation, understanding of magma supply, storage, and transport at Hawaiian volcanoes has evolved from a rudimentary set of observations and generalizations to quantitative models and detailed characterizations. These insights have been applied to numerous volcanoes on the Earth and on other planets and have aided in the development of volcano monitoring techniques and hazard mitigation efforts. We expect that Hawaiian volcanoes will continue to serve in this role for the next 100 years.

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Appendix. Inverse Modeling Methods

All inverse models in this study (shown in figs. 15–18) are kinematic and assume a homogeneous, isotropic, linearly elastic half-space, with data weighted according to their respective variances. We assume a Poisson's ratio of 0.25 and a shear modulus of 3×10^{10} Pa. Modeled displacements are computed from expressions that relate deformation of point (Mogi, 1958) or planar (Okada, 1985) sources to surface displacements. Parameters estimated for point sources include east position, north position, depth, and volume change. The parameters estimated for the planar sources are length, width, depth, dip, strike, east position, north position, and opening.

Model parameters (east position, north position, depth, and volume change) for point sources in the Halema‘uma‘u (figs. 16 and 17) and Keanakāko‘i (fig. 18) models were determined using a Markov Chain Monte Carlo (MCMC) optimization algorithm (Metropolis and others, 1953). We assumed that the a priori distributions of the parameters are uniform between broadly chosen bounds. The posterior distributions show the range of parameters that fit the data acceptably. The preferred model is the one with the smallest misfit.

Sill opening for the south caldera plus upper Southwest Rift Zone model (fig. 15) was determined by first constraining the parameters describing a single, uniformly opening planar dislocation (length, width, depth, dip, strike, east position, north position and opening) with an MCMC optimization (fig. 27). The preferred dislocation was then expanded horizontally and subdivided into 375-m by 375-m squares for the distributed opening model. We used a non-negative least squares algorithm that minimizes the L2-norm of the weighted residuals. Spatial smoothing was applied using a Laplacian operator with the optimal weight chosen by the L-curve criterion (fig. 28; Hansen, 1992).

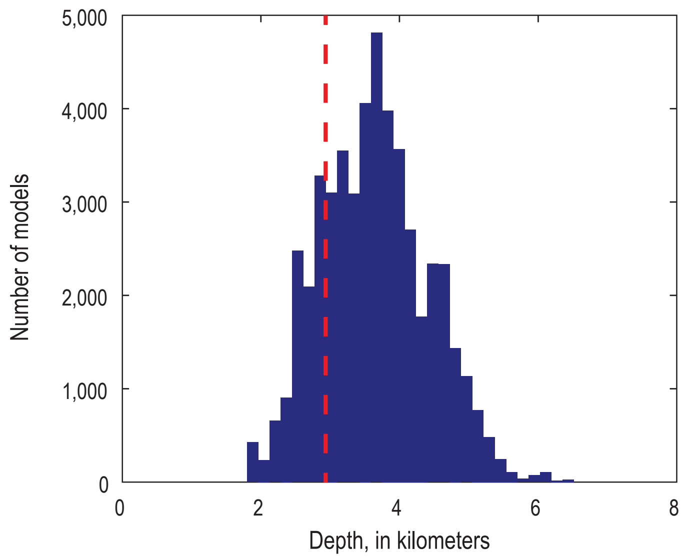

Figure 27. Histogram showing the distribution of sill depth that fits deformation of the south caldera and upper Southwest Rift Zone (fig. 15*A*). The best-fitting sill, which is used to constrain the distributed opening model (fig. 15*B*), is at 2.9-km depth (red dashed line), with 95-percent confidence values spanning 2.1–5.1 km. The best-fitting depth does not align with the peak of the distribution.

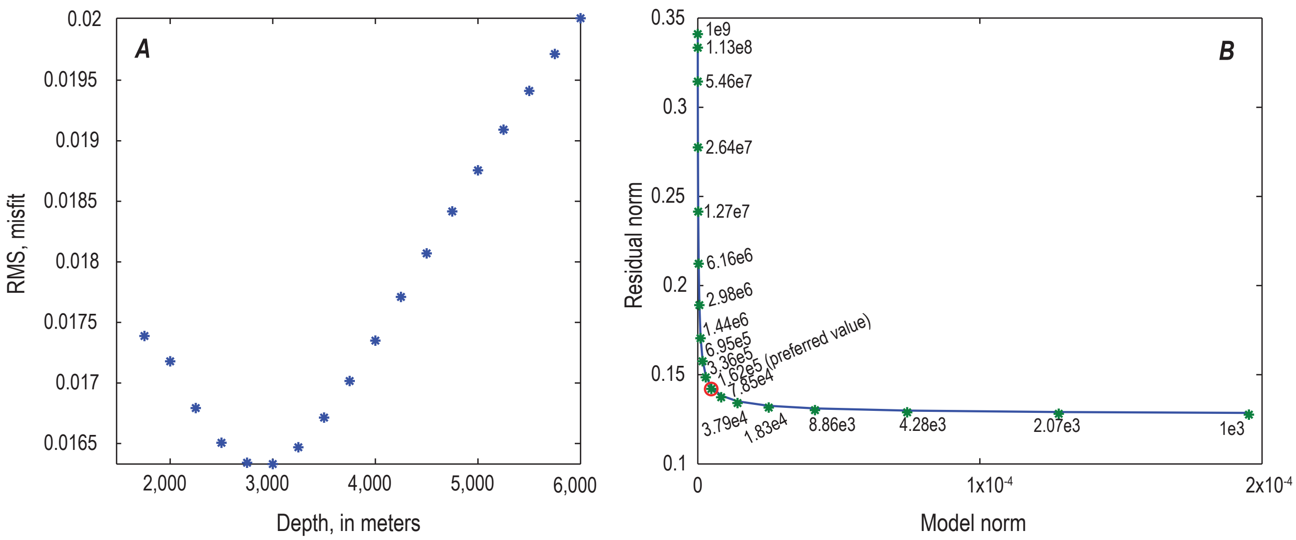

Figure 28. Plots of model misfit and smoothing parameters for distributed-opening sill model shown in fig. 15. *A*, Root-mean square (RMS) misfits for sill models of varying depths. The misfit is minimized at ~3-km depth. *B*, L-curve (Hansen, 1992) for models of varying values of the smoothing weight. The curve compares the weighted residual norm (a measure of data misfit) versus model norm (a measure of model roughness). The optimal smoothing weight is chosen as the corner value that represents the smoothest model with minimal increase in misfit.