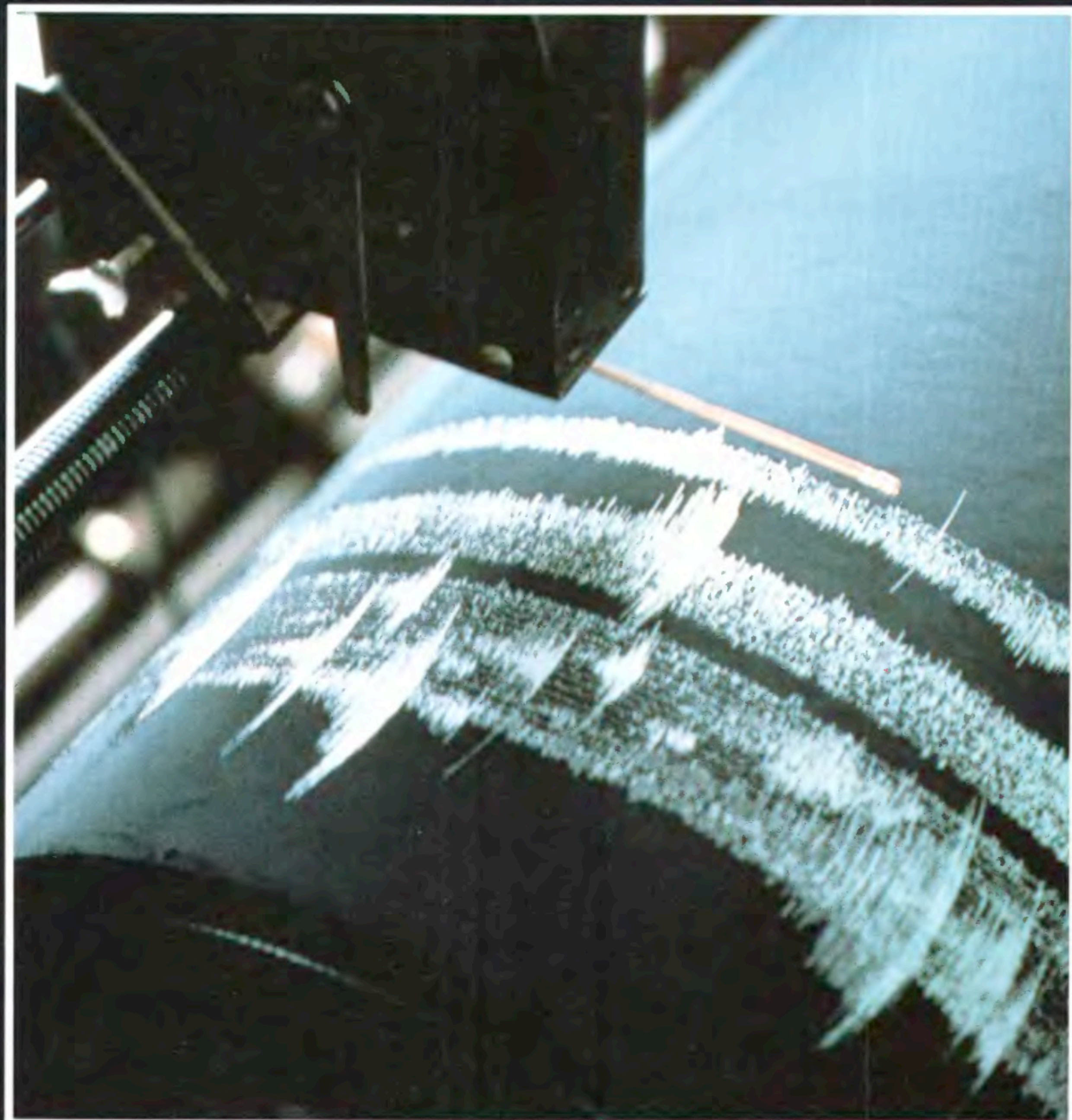


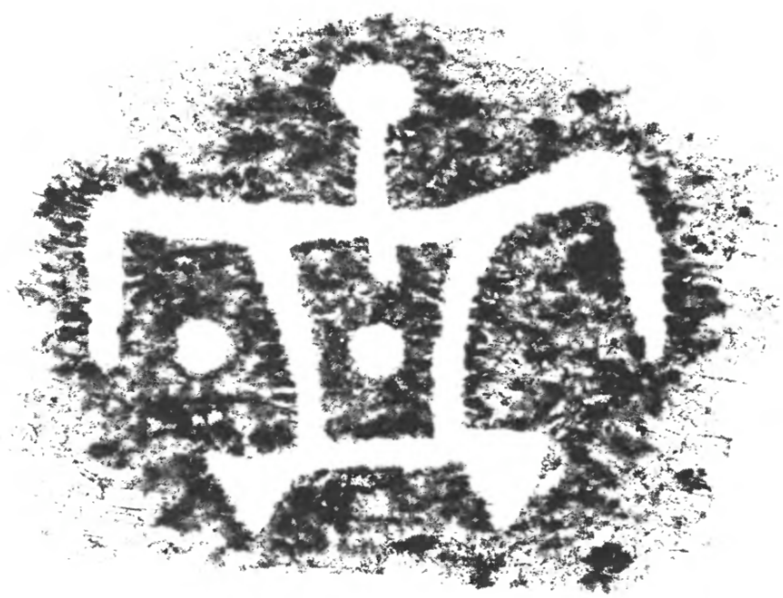
Earthquake swarm and volcanic tremor being recorded on seismograph at Hawaiian Volcano Observatory, November 16, 1979. Photograph by R.W. Decker.

DYNAMICS

Much of the present research on Hawaiian volcanoes is directed towards understanding how they work. It is appropriate, therefore, that this section on dynamics is the longest in the two volumes on Volcanism in Hawaii.

The eighteen chapters in this section address questions about the physical and chemical dynamics of magma formation, accumulation, ascent, storage, intrusion, and eruption. These questions cannot be investigated in isolation, and many of the papers in the section on petrogenesis and volcanic gases and in that on structure are closely related.

An overview chapter is followed by three chapters on earthquakes and volcanic tremor, four chapters on ground-surface deformation related to magma movements, and one on the geoelectric contrast between the crust and molten core of a lava lake. Chapter 51 points out the resemblance of the cyclical behavior of Hawaiian volcanoes to patterns predicted by attractor dynamics, a new conceptual approach to understanding the interactions in complex systems. The next two chapters delve into the rock mechanics that constrain the physical dynamics of Hawaiian volcanoes, and chapters 54 through 56 consider the thermodynamics and geochemistry of the magmatic-geothermal systems formed on active Hawaiian volcanoes. The last three chapters in this section discuss the flow dynamics of Hawaiian lava.



Turtle
Hawaiian Petroglyph
Puuloa, Hawaii

DYNAMICS OF HAWAIIAN VOLCANOES: AN OVERVIEW

By Robert W. Decker

ABSTRACT

Eaton and Murata's cross section of Kilauea Volcano, published in 1960 (Science, v. 132), remains the basic model of how Hawaiian volcanoes work. Magma forms by partial melting of mantle rock at depths between 60 and 170 km in a specific area of mantle known as the Hawaiian hot spot. The melt accumulates by migrating through small fractures, or by shear coalescense, into volumes large enough to ascend. The ascensive pressure is caused by the greater density of the rocks surrounding and overlying the magma. Ascent through the lithosphere occurs in discontinuous conduits that are generated by magmafracting. Apparently there are enough of these ascensive conduits so that magma is supplied to a shallow reservoir system 3–7 km beneath the summit caldera at rates that vary by less than an order of magnitude over time scales of months to decades. When the shallow magma reservoir fills to about lithostatic pressure, magmafracting emplaces dikes upward or laterally into the rift zones; those dikes that reach the surface erupt. Eruptions at high-volume rates rapidly reduce the pressure in the shallow magma reservoir and are of brief duration. Long-lived eruptions occur at low-volume rates and are sustained by the resupply of magma from depth. Lava fountains are produced by the rapid expansion of magmatic gases—H_2O, CO_2, and SO_2—at near-surface pressures. Major caldera collapses, sometimes associated with explosive eruptions, occur repeatedly at intervals of a few thousand years. One probable cause of these collapses is intrusion or eruption on the submarine rifts which produce major draindown of the shallow reservoir system. The newly formed volcanic islands subside at exponentially decreasing rates as the Pacific plate moves slowly northwestward past the Hawaiian hot spot.

INTRODUCTION

The literature on Hawaiian volcanism contains many descriptions of eruptions, but fewer interpretations of the dynamic processes that form magma and that cause it to ascend and erupt at the surface. Present studies, however, are concentrating on obtaining the necessary surface and subsurface data to compare with inferences from dynamic models of how Hawaiian volcanoes work. Those models, constrained by geological, geophysical, and geochemical data, are beginning to give a glimpse of the complex, hidden processes that occur beneath Kilauea and Mauna Loa Volcanoes. The purpose of this chapter is to synthesize these studies into a comprehensive model of the dynamics of Hawaiian volcanoes, particularly during their stage of vigorous growth.

The earliest recorded speculations on the subsurface structure and dynamics of Hawaiian volcanoes appear in the journal of William Ellis, a Christian missionary from England, who walked around the Island of Hawaii in 1823. While visiting Kilauea caldera, Ellis (1827, p. 171) noted that his Hawaiian guides told him about many eruptions on the flanks of Kilauea "on which occasions they supposed Pele went by a road under ground from her house in the crater to the shore." Present-day interpretations of rift-zone eruptions on the flanks of Kilauea are considerably more detailed, but in essence are the same as this "road under ground" of the native Hawaiians.

Ellis goes on to speculate about the dynamic processes beneath the active lava lakes that he saw in Kilauea caldera (p. 181): "Perforated with innumerable apertures in the shape of craters, the island forms a hollow cone over one vast furnace." This concept has not survived as well as Pele's road.

J.D. Dana (1890, p. 174–175), geologist with the U.S. Exploring Expedition to Hawaii in 1840, was the first to distinguish clearly the more or less steady ascensive force of magma at depth from the rapid fountaining of erupting lava that is caused by the sudden boiling off of gases at near-surface pressures. Dana (1895, p. 303) attributed the ascensive force in the volcano to "1) the expansive action of moisture from the deep-seated source of the lavas; and 2) the gravitational pressure of the contracting crust of the globe, forcing up the lavas* * *." He noted that the top of the Kilauea magma column had risen 360 feet in 6 years. Both of Dana's interpretations are close to, but not quite the same as, present interpretations. These two causes of ascensive force could now be reworded as follows: (1) The expansive action of fluids, particularly CO_2, from the deep-seated source of the lava, forming separate phases; and (2) the gravitational pressure of the overlying, heavier crust and upper mantle of the globe, forcing up the lighter lava.

Eaton and Murata (1960) were the first to draw a dynamic cross section of a typical Hawaiian volcano (fig. 42.1). In their model, magma moves upward by density contrast, more or less continuously, through conduits from the top of a source region at 60 km depth to a storage reservoir at a depth of 3–7 km. Intermittent eruptions from the shallow magma reservoir to the surface occur through newly formed fractures breaking upward to the caldera or laterally into the rift zones. The eruption rates from the shallow magma reservoir to the surface are often much higher than the rate of resupply from depth, and thus the eruptions tend to be self-limiting and intermittent. Sometimes low-volume-rate eruptions form active lava lakes in the caldera, presumably sustained by the resupply rate. Such more or less steady-state eruptions can last for months or years. This overall model fits well with the seismic data and observed ground-surface deformation on Kilauea Volcano, and it is still the

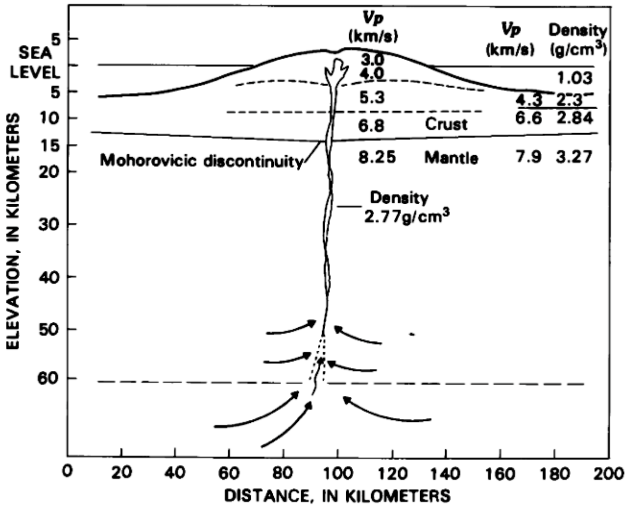


FIGURE 42.1.—Schematic cross section through idealized Hawaiian volcano. Magma from a source about 60 km deep streams up through conduits and collects in shallow reservoir beneath caldera. Occasional discharge of lava from shallow reservoir through dikes that split to surface constitute eruptions. Note slight depression of Mohorovicic discontinuity beneath volcano. V_p is the velocity of seismic *P*-waves. Modified from Eaton and Murata (1960).

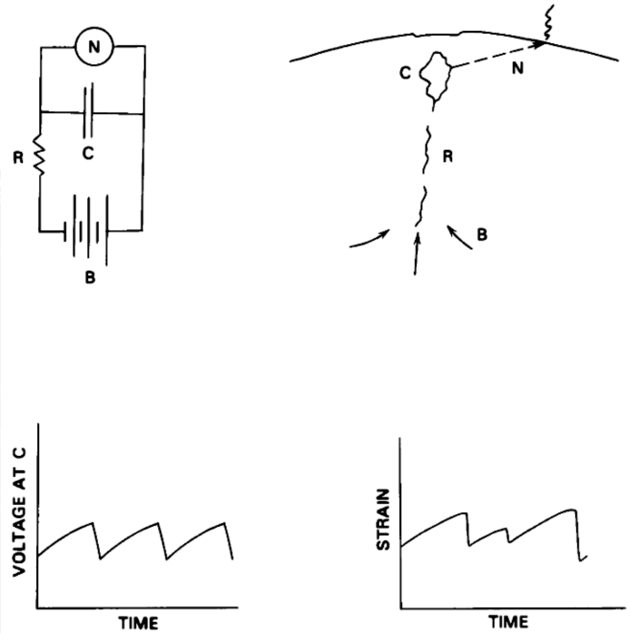


FIGURE 42.2.—Electrical analog of dynamics of Kilauea Volcano. Electrical circuit is a battery (B) charging a condenser (C) through a resistor (R). When voltage across condenser reaches firing voltage of neon bulb (N), bulb winks on until voltage drops below sustaining level. In Kilauea deep ascensive force of magma represents battery, conduits from deep source to shallow magma reservoir form resistor, and shallow magma reservoir is condenser. Dike breaking to surface or intruding a rift zone reduces magma pressure (voltage) and is equivalent to neon bulb. Variation of strain in volcanic edifice is similar to variation of voltage in circuit, but less regular in its cycles.

basic concept of how Hawaiian volcanoes work.

Building on Eaton and Murata's model, Decker (1968) and Shimozuru (1981) have noted that the fluid and fracture dynamics of this model are analogous to the storage and discharge of electricity in a resistance-capacitance (RC) circuit (fig. 42.2). The ascensive pressure is analogous to a battery, the resistance to fluid flow through the system is analogous to a network of electrical resistors—within the battery, between the battery and the condenser, and between the condenser and the surface—and the shallow magma reservoir is analogous to a condenser. If the resistance between the condenser and the surface operates in two modes—almost infinite resistance or very low resistance, as in a neon bulb—then intermittent discharges of the condenser in the electrical model will be analogous to eruptions.

Although this electrical analog model is greatly oversimplified, it has the advantage of allowing thought experiments about what happens when changes occur in the ascensive pressure, the resistance to flow, or the volume of the shallow magma reservoir, either separately or in combination. For example, if the ascensive pressure were doubled, the repose time between eruptions would be halved and long-term lava output would double, but the individual eruptions would have the same volume as before. Increasing the capacity of the shallow magma reservoir would increase the repose time between eruptions, but the eruptions would be of proportionately greater volume. Changing the breakdown strength of the rocks between the shallow magma chamber and the surface—analogous to changing neon bulbs—could change both the repose time between eruptions and the volume of the eruptions. If the values of the various components are computed or assumed, then their quantitative interrelation can be calculated using well-established electrical or hydraulic equations.

Realistic models of the structure and dynamics of Hawaiian volcanoes are strongly dependent on the physical properties of magma and its host rocks. Some of these properties vary greatly, and the numerical values of others are only poorly known. Refined models are therefore dependent on improved knowledge of these physical properties.

Attempting to understand the hidden dynamics of an active volcano calls upon the most eclectic abilities of volcanologists. A valid dynamic model must explain the geologic and eruptive history of the volcano, as well as the seismic, deformational, and electromagnetic changes that occur before, during, and after eruptions. In addition it must explain the composition and changes in composition of the solid, liquid, and gaseous products emitted from the volcano. Besides all this, the model must be consistent with the laws of physics and chemistry and, if possible, satisfy the philosophy of Ockham's Razor—that is, not to be unnecessarily complicated. Such a task clearly requires a team effort. This article is therefore a selected synthesis of the data and interpretations of many students of Hawaiian volcanism.

ACKNOWLEDGMENTS

Jerry Eaton, Howard Powers, and Gordon Eaton were my kind hosts and patient teachers while I was a guest investigator at the Hawaiian Volcano Observatory (HVO) in the 1960's and 1970's. Robert Tilling offered me the opportunity of returning to HVO as Scientist in Charge, and my fine colleagues at HVO continued my education during my tenure there from 1979 to 1984. Jerry Eaton, Dan Dzurisin, and Barbara Decker gently but thoroughly reviewed earlier versions of this chapter. My warmest thanks to all these fellow students of volcanoes.

MAGMA FORMATION

When, where, and how does magma form? There are no clear answers to these questions, but there is no lack of hypotheses. Since volcanism has been a persistent geologic process for billions of years, it seems reasonable to assume that magma has continued to form during most of the Earth's history. A counter argument could be that magma only formed in the early history of the Earth, has always been present within the Earth, and has been released slowly enough to stretch out volcanism for billions of years. This alternative has not been proven wrong, but since most geologic processes work at levels close to equilibrium, it is difficult to accept a process that could be restrained for such enormous periods of time. For this reason it is assumed that magma is forming within the Earth at the present time.

There is reasonably good agreement about where basaltic magma is formed. Yoder (1976) carefully reviewed the geophysical and geochemical evidence and concluded that basaltic magma may be generated in broad regions generally at depths of 50–170 km; in more restricted environments the source of basaltic magma may be as deep as 300–400 km. Eaton and Murata (1960) concluded on the basis of depths of earthquake sources that the magma source region beneath Hawaii has its top at a depth of 60 km, but they were not able to infer how far the source region may extend below this level. Wright's model (1984) for the origin of Hawaiian tholeiite proposes a depth of formation ranging from 60 to 90 km. Recent studies by Anderson and Dziewonski (1984) using seismic tomography indicate an anomalously slow (hot) zone in the Earth's mantle at a depth of 350 km over a broad area south of Hawaii. Mantle flow in this area is apparently upward. At shallower depths in the upper mantle, the anisotropy of Rayleigh wave velocity suggests horizontal flow in a southeastward direction beneath Hawaii. As a working hypothesis, it is assumed that magma is currently being formed beneath Hawaii at some depth interval between 60 and 170 km.

The question of how magma forms is even more complex than when or where it forms. From what rock material is it melted, and what is the source of heat? Yoder (1976) thinks that the best candidate for the parent rock of basaltic liquids is garnet peridotite, a crystalline aggregate of olivine, orthopyroxene, clinopyroxene, and garnet. He accepts this source—tentatively—and lists the following features of garnet peridotite as supportive evidence (Yoder, 1976, p. 42):

(1) Its occurrences are appropriate to deep-seated environ ments of origin.

(2) Its close compositional relation to meteorites supports a derivation from material accumulated in the primordial Earth.

(3) Partial melts of natural samples at high pressure have basaltic compositions.

(4) The mineral asmblage was found experimentally to be stable at high pressure and temperature.

(5) It has appropriate density and seismic velocity.

Wright (1984) prefers a source rock for Hawaiian tholeiite magma composed of lherzolite, a crystalline aggregate of olivine, orthopyroxene, and clinopyroxene, with some other constituents (amphibole, nephelinite, ilmenite, apatite, and ferrous iron oxide) added from depth.

The source of heat to melt a basaltic liquid from a crystalline rock aggregate must be either internal or external to the rock being melted. If it is external, some thermal process such as conduction, convection, or radiation must transport heat into the source region. If it is internal, the heat must be generated by some process such as radioactive decay or mechanical friction, unless it is heat that is already present in the rock. In the latter case, some other process must be called upon to lower the melting temperature of the basaltic liquids, such as lowering the confining pressure or introducing fluxing elements or compounds into the source region.

The problem is not a lack of possibilities, but too many permutations and combinations to choose among. Yoder (1976, p. 60–61) lists 21 proposed mechanisms to melt rock into magma. Several of these are not tenable for Hawaii; for example, those mechanisms that require subsidence of continental crustal layers of the Earth by 5 km or more (Hess, 1960). Among the more reasonable mechanisms for generating Hawaiian magmas are the following (in alphabetical order by author):

(1) Melting caused by the friction of shear strain released from stored elastic strain energy along a propagating crack (Griggs and Baker, 1969).

(2) Internal radioactive heat production (Joly, 1909).

(3) Conductive heat trapping caused by changes in thermal conductivity with temperature and pressure (McBirney, 1963).

(4) Rise of the source material as a buoyant diapir, thus reducing confining pressure and melting temperature (Ramberg, 1972).

(5) Melting caused by the friction of viscous strain. In this thermal feedback or runaway model, constant shear stress on a crystalline rock at its melting temperature causes small viscous strains that increase the temperature. This temperature increase lowers the viscosity, increases the strain rate, and thus further increases the temperature (Shaw, 1973).

(6) Melting caused by the friction of shear strains associated with tidal energy dissipated in the solid Earth (Shaw, 1970).

(7) Rise of the source region by solid-state convection, thus reducing the confining pressure and melting temperature (Verhoogen, 1954).

(8) Reduction in melting temperature of the basaltic liquids by addition of volatiles (H_2O, CO_2) to the parent-rock source region (Yoder and Tilley, 1962).

Radioactive uranium, potassium, and thorium isotopes are apparently ubiquitous in upper-mantle rocks but low in amount. Hawaiian lava exhibits no abnormal content of these radioactive isotopes, so radioactive heating (mechanism 2) does not seem to be a principal cause of the high rate of Hawaiian magma production. However, on a global basis and over tens to hundreds of millions of years, radioactive heat production probably contributes significantly to heating the rocks of the upper mantle to near melting temperatures.

Choosing among the 7 remaining mechanisms to explain Hawaiian magma generation depends in part on tectonic prejudices. Those who favor the hot-spot hypothesis (chapter 1, part I) will probably favor the diapiric rise (4) and (or) volatile transfer (8) mechanisms, while those who favor a propagating fracture may prefer the stored elastic energy release (1) and (or) the conductive heat trap (3) mechanisms. Wright's model (1984) combines a thermal plume to drive up the added chemical elements from depth and thermal feedback to provide the principal heat of melting. It is probably best to say that the melting problem is not resolved at this time. Detailed seismic tomography studies may soon provide more definite evidence on this important but difficult subject of the heat source of Hawaiian magma.

In all the proposed mechanisms for melting crystalline rock to form magma, the enthalpy of melting is about $4.2–5.2 \times 10^5$ J/kg (100–125 cal/g). This heat of fusion has a major buffering effect on the temperature and rate of melting or freezing in the source region, implying that basaltic magmas of similar composition at similar depths of formation are nearly isothermal. At 60-km depth, Hawaiian tholeiite magma has a calculated formation temperature of 1,350–1,400 °C (T.L. Wright, oral commun., 1985). Measured temperatures of erupting Hawaiian lava are in the range of 1,100–1,200 °C.

MAGMA ACCUMULATION AND ASCENT

Melting a garnet peridotite or lherzolite mantle involves a 10–15 percent volume increase from the crystal aggregate to the basaltic fluid. A change from solid to 30 percent partial melt would thus bring about a 3–5 percent increase in the bulk volume of the rock-liquid mixture.

In experimental studies on crystalline silicate aggregates, the first melt forms films between mineral grains that then unite into a liquid-bonded aggregate—a network of melt coating the mineral grains with a thin film of fluid (Yoder, 1976, p. 164-165). The volume of these molten films can reach 10 percent of the mineral-liquid mixture. As the volume of melt increases, deformation from internal volume expansion and (or) interaction with external stresses takes on an increasingly important role.

Extension fractures perpendicular to the direction of the least compressive stress can form even at mantle depths if the pore-fluid pressure exceeds the tensile strength of the rock. Shaw (1980) found that these conditions are probable in the thermal and stress domains estimated to occur in the asthenosphere beneath Hawaii. His model of melt accumulation and ascent is shown in figure 42.3; in it the accumulation fractures are subhorizontal (sills) if the least compressive stress is vertical, or they are nearly vertical fractures (dikes) if the intermediate compressive stress is vertical. Because of small transient changes in the shear stresses, pore pressures, or gravitational load on this deep region where the three major compressive stresses are nearly but not quite equal, Shaw thinks both of the above conditions will occur and will generate a set of planar melt lenses oriented at 50°–90° to one another, with intersections parallel to the general direction of maximum compressive stress. These intersections are shown schematically in figure 42.3 by the broken horizontal line segments within the asthenosphere. The inferred lenses of melt accumulation would tend to lie both (vertically) in the plane of the cross section and (horizontally) perpendicular to it. Since there is no seismic evidence for major brittle fracturing in the asthenosphere beneath Hawaii, these proposed cracks would have to be small or form slowly.

An alternative way of expressing the fracture-accumulation hypothesis is shearing of the films of melt into progressively larger pockets of molten rock. Shaw (1969, p. 533) refers to the melt fraction as being "kneaded from the crystalline source"; Weertman (1972), on the basis of observed bubble coalescense in deforming glacial ice, believes a similar process may occur during shear creep in partially melted rock. Spera (1980) points out that the rate of melt segregation is governed by the rheology of the rock-liquid mush and by the balance among buoyancy, pressure, and viscous forces. He also notes that he can estimate buoyancy reasonably well and stress gradients to within an order of magnitude, but viscous forces and permeabilities only poorly. Nevertheless, Spera, on the basis of correlation to rigid porous-media systems, estimates that the segregation rate is highly dependent on the melt fraction; in dimensionless terms it would be 6,500 times larger for a 30-percent partial melt than for a 2-percent partial melt.

Whatever the mechanism of melt segregation is beneath Hawaii, it must be rapid enough throughout the source volume to provide about 0.1 km^3/yr of magma to Kilauea Volcano for time periods longer than a few decades (Swanson, 1972; Dzurisin and others, 1984) and a roughly similar but less well known amount to Mauna Loa Volcano. Assuming that the source region is approximately cylindrical, as wide as the island (100 km), and 30 km thick, the source volume would be approximately 2.5×10^5 km^3. Using a partial melt of 30–40 percent (Wright and others, 1979; Wright, 1984), the calculated average fraction of the total melt delivered to shallow levels in Hawaii on a yearly basis is only $2–3 \times 10^{-6}$. About 2.5×10^{17} J (6×10^{16} cal) of heat of fusion are needed by this slowly evolving source region on a yearly basis to maintain a steady-state supply rate of 0.2 km^3/yr. Under the above assumptions, this translates into about 3.2×10^{-5} J/m^3/s (7.7×10^{-12} cal/cm^3/s), an amount of heat nearly 10^3 times greater than the heat production from radioactivity in an average garnet peridotite. These numbers assume that the melting and segregation processes are spread evenly throughout the source region; it is likely that these processes in fact become more concentrated near the apex of the source region.

The mechanism of the ascent of magma through the more rigid rocks of the lithosphere must be both qualitatively and quantitatively

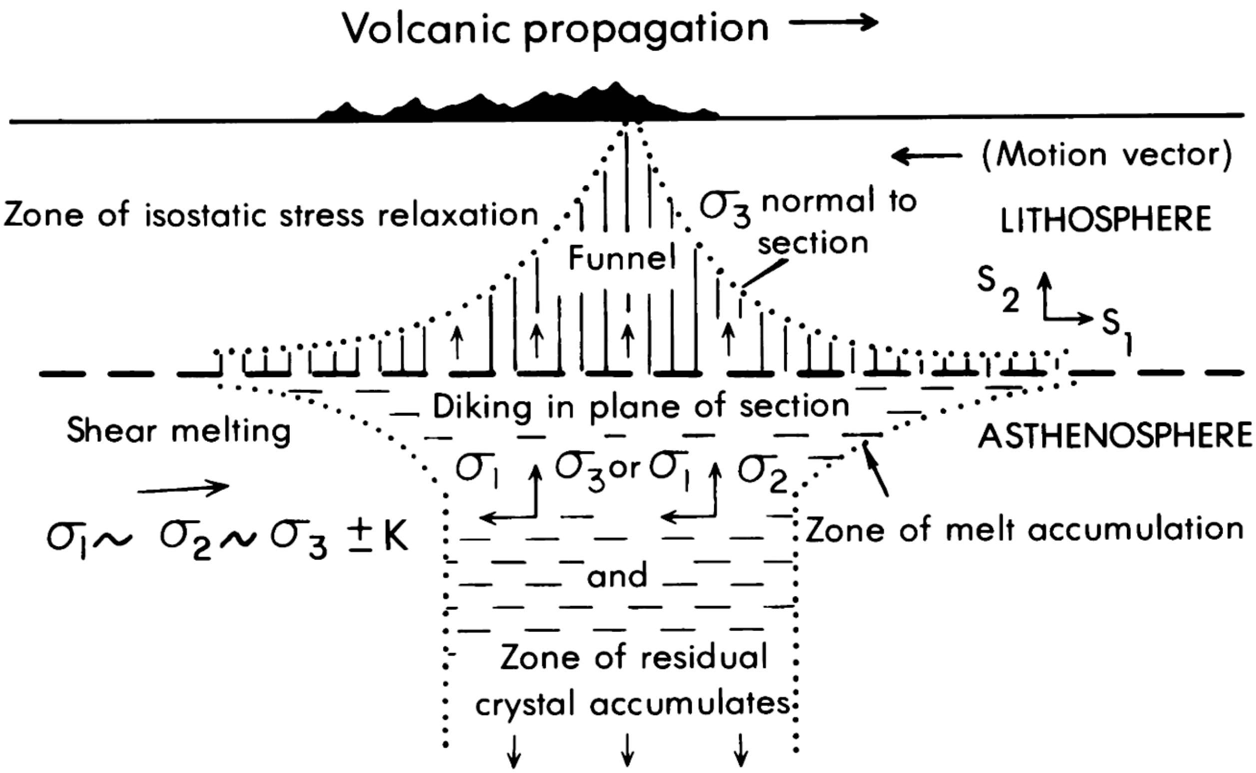


FIGURE 42.3.—Schematic longitudinal section showing stress domains beneath active Hawaiian volcanoes. S, total normal stress acting at a point in rock; σ, effective normal stress acting in rock containing an interconnected fluid phase. Subscripts 1, 2, and 3 refer to maximum, intermediate, and minimum principal stresses, respectively. K, tensile strength. In zone of melt accumulation, fractures would be either nearly vertical in plane of section, or nearly horizontal in plane perpendicular to section. In ascent funnel, fractures would be nearly vertical in plane of section. Modified from Shaw (1980).

different from that of its segregation and ascent in the source region. The tensile strength of the lithospheric host rocks is greater than that in the asthenosphere, although it may become weakened by heating from ascending magma. Deviatoric stresses in the lithosphere may be much greater than in the asthenosphere, and beneath Hawaii the lithosphere is seismically active whereas the asthenosphere is apparently not.

Magma is pushed upward into and through the lithosphere by the pressure gradient caused by gravity acting on the density difference between lighter magma and heavier host rocks. This explanation of the ascensive force was first clearly proposed by Daly (1914, p. 183) who suggested "that magmatic eruption is, in the first instance, a hydrostatic phenomenon." Daly later became even more convinced. His second edition (1933, p. 247) states "The leading cause for the ascent of primary magma, assumed to move up abyssal fissures, is naturally found in the weight of the adjacent crust." Almost no one would now disagree with this explanation of the ascensive force; the origin of the abyssal fissures, however, is much more debatable.

Does Hawaiian magma passively rise through conduits in the lithosphere generated by external stresses, or does it actively generate or help to generate these conduits? Turcotte and Oxburgh (1978) propose that deformation of the Pacific plate in adjustment to thermal and Earth-curvature changes may cause a major propagating intraplate fracture. They suggest that magma leaking up this propagating fracture has progressively formed the Hawaiian volcanic chain. Since Wilson (1963) proposed the hot-spot origin for the Hawaiian Islands, most other researchers believe the melting anomaly is fixed in position and that plate motion over the hot spot causes the age progression of the Hawaiian chain (Dalrymple and others, 1973; McDougall, 1979). Implicit in the hot-spot hypothesis is the idea that magma in sufficient quantities can somehow force itself upwards through the lithosphere. Possibly the stresses proposed by Turcotte and Oxburgh (1978) may assist in this magmafracting process.

The concept that magma can actively fracture its way upward is strengthened by the distribution of earthquake hypocenters beneath Kilauea Volcano (see chapter 43). Over periods of years this earthquake distribution outlines a zone of active fracturing in the lithosphere that is shaped like a steeply dipping inverted funnel reaching generally to a depth of 45 km and occasionally to nearly 60 km (Eaton, 1962). Close association of the earthquakes with the inferred conduit path of the rising magma, rather than with the site of

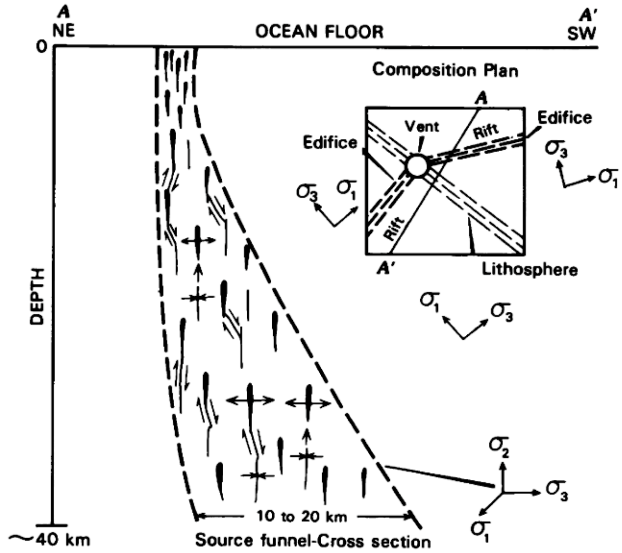


FIGURE 42.4.—Schematic cross section of lithosphere beneath Kilauea Volcano showing magma ascent funnel as region of extensional fractures and magma batches migrating upward from asthenosphere. Inset shows simplified plan of rift zones. Stress orientations and dike orientations have different directions in lithosphere and in edifice of volcano. Symbols have same meaning as in figure 42.3. Modified from Shaw (1980).

the postulated intraplate fracture, indicates that the magma plays an active role in the formation of the conduits.

Shaw (1980) proposes that magma is injected into the lithosphere by a process analogous to hydraulic fracturing of the wall rocks in a borehole by increasing fluid pressure. The normal compressive stresses in the buried host rock can be balanced or exceeded by the magma pressure to produce zero or even negative effective stresses. When these effective stresses equal or exceed the tensile strength of the rock, extensional fractures form perpendicular to the direction of least compressive stress (figure 42.4). Yoder (1976, p. 179) has called this process magmafracting. In this view most of the earthquakes beneath Hawaii are the effect, not the cause, of the transport of magma.

As the vertical dikes of magma shown in figure 42.4 intermittently force their way upward by brittle fracture, they must be resupplied with more magma from the source region, or the trailing edge of the rising dike will close off because of the recovering compressive stresses (Pollard, 1976). Hill (1977) shows that shear stresses will cause the active dike system to become a complicated network of dike-like lenses connected by faults. Movement of magma within one of the dikes will change the local pattern of deviatoric and effective stresses, and this may trigger new fractures or movements on the existing fracture network. Hill uses this model to explain earthquake swarms in Hawaii, and Shaw calls such resulting swarms failure cascades.

Shaw (1980) then inverts the seismic data from Hawaii in order to estimate the volume and magma fraction of the active intrusion zone in the lithosphere located between the deep source region in the asthenosphere and the shallow magma-reservoir system beneath Kilauea Volcano. His calculations suggest that this zone has a volume of about 10^3 km^3 and a melt fraction between 10^{-5} and 10^{-4} (that is, 0.01–0.1 km^3 of magma). For continuity, the 0.1-km^3 yearly magma supply to Kilauea would thereby have an average residence time of only a month to a year while rising through this zone. Magma ascent may appear relatively continuous over time spans of months to years, but can be quite variable at time scales of minutes to days. Several batches of lava may be moving upward in independent spurts. Some batches may rise 30 km in a few weeks, while other batches may never reach shallow levels. The varying rate of inflation of the summit region of Kilauea, as measured by continuously recording tiltmeters, supports this model of semicontinuous rise of magma into the shallow reservoir system.

MAGMA STORAGE

Batches of Hawaiian magma rising through the lithosphere apparently take various paths and various amounts of time in their ascent, but most of the magma supplying Kilauea Volcano during the past several decades has paused in a shallow magma storage system before erupting to the surface.

Mogi (1958), using data from Wilson (1935), first identified the presence and depth of this shallow magma reservoir by analysis of surface uplift and subsidence of the summit and upper flanks of Kilauea. Using elastic-deformation theory he identified two major sites of pressure and (or) volume change in the region beneath Kilauea caldera: one well-defined site at 4 km depth and another less well defined site at 25 km depth. Eaton (1962) rejected the deeper site as an apparent artifact of leveling-rod error in the long survey traverse from Hilo to the summit of Kilauea. Using additional level and tilt data, Eaton confirmed the shallow reservoir system and showed it on his schematic cross section (fig. 42.1) as an irregular vertical cylinder about 4 km wide and at 3–7 km depth beneath the summit of Kilauea.

Patterns of continuously recorded summit tilt over the past two decades (fig. 42.5) indicate that this shallow magma-reservoir system slowly inflates over periods of months to years during the intervals between eruptions and then rapidly deflates over periods of days to weeks during flank eruptions or lateral intrusions into the rift zones. Expansions and contractions of the caldera diameter by as much as a few meters accompany these inflations and deflations.

On many of the inflation cycles shown in figure 42.5, the rise of the summit shows a high rate of inflation immediately after a deflation event. This is followed by an exponentially decreasing rate of inflation, probably caused by the decreasing pressure gradient between the shallow magma-reservoir system and the sources of magma at greater depths and (or) by the increased leakage of magma into parts of the shallow magma reservoir away from the summit as the pressure in the shallow reservior system increases.

Considerable research has been done to attempt to define the depth, shape, and dynamic nature of the shallow magma-reservoir

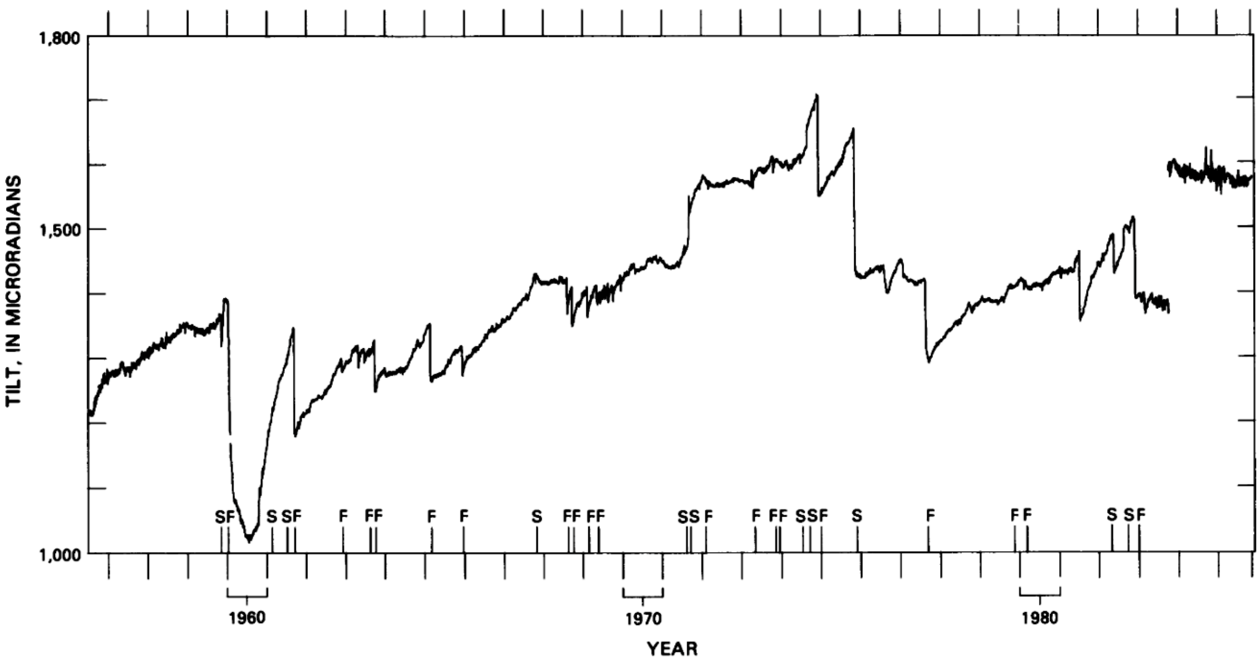


FIGURE 42.5.—Radial tilt at Hawaiian Volcano Observatory (HVO) on northwest rim of Kilauea caldera. Increasing tilt (radially outward) indicates inflation of summit region; decreasing tilt indicates deflation. One hundred microradians (μrad) of tilt at HVO indicates uplift of apex of summit bulge by about 50 cm. Rapid deflation marked by sharp drops in tilt represent volume change of bulge of about 3.3×10^5/μrad and pressure drop in shallow magma reservoir system of about 90 kPa/μrad. Magma is added slowly to shallow reservoir system from depth between eruptions and is rapidly removed during flank eruptions. Vertical lines indicate eruptions; S, summit eruptions; F, flank eruptions. Offset in tilt record in November 1983 was caused by *M*-6.6 earthquake (see Buchanan-Banks, chapter 44).

system. Geologic evidence from the location and form of Kilauea caldera and the pit craters along the upper east rift zone gives some indication of the subsurface location and lateral extent of the reservoir system. The caldera and pit craters have formed by collapse and thus identify the subsurface regions from which magma has been removed in such large quantities that the deflation cannot be accommodated by elastic or plastic surface deformation. Presumably these large removals occur during major eruptions from the distant submarine portions of the east rift zone. Interpretation of this morphological evidence indicates that the main magma reservoir beneath Kilauea caldera is (or was) roughly elliptical in plan—4 km NE.-SW. by 3 km NW.-SE.—and that about 3 km^3 of magma had been removed from the shallow subsurface reservoir to accommodate the volume of the caldera as it appeared in 1823.

The location of the pit craters is evidence that the magma storage system extends (or extended) southeast and then northeast beneath the upper east rift zone at least as far as Napau Crater, 15 km east of the summit caldera. The pit craters range in diameter and volume from 1 km and 77×10^6 m^3 (Makaopuhi before 1965) to 25 m and 90×10^3 m^3 (Devil's Throat); these figures put some limits on the width and volume of previous magma-storage regions beneath the east rift zone. There are only two small but deep pit craters along the southwest rift zone, indicating its subordinate role as a magma-storage region. Duffield and others (1982) conclude that the southwest rift zone breaks primarily because of accumulated movements of the east rift zone and to a lesser degree because of magmatic pressure from the summit reservoir system.

Geodetic data on the changing elevations and displacements of the deforming ground surface above the shallow magma-reservoir system have provided the best evidence on the dynamic nature of these magma storage chambers (Ryan and others, 1983). Detailed leveling of the summit area of Kilauea during an inflation period by Fiske and Kinoshita (1969) shows that the apex of inflation migrates over distances of 1–3 km within or near the caldera (figure 42.6). This shift of the center of inflation indicates that the summit reservoir is a plexus of interconnected storage zones. It can be compared to a Mickey Mouse balloon whose head inflates separately from the ears. Recording tiltmeters near HVO show that this migration of the apex of inflation or deflation during the larger cycles of magma influx and discharge has a repetitive pattern. After a major summit deflation, reinflation begins beneath the northern part of the caldera and then moves southward; during the next major deflation, the northern part of the caldera also subsides first and is quickly followed by the south end subsiding. This implies that the main input and withdrawal conduits for the summit reservoir plexus are on the north side of the general storage zone, which lies mainly beneath the south end of the caldera.

Dvorak and others (1983) analyzed the surface deformation

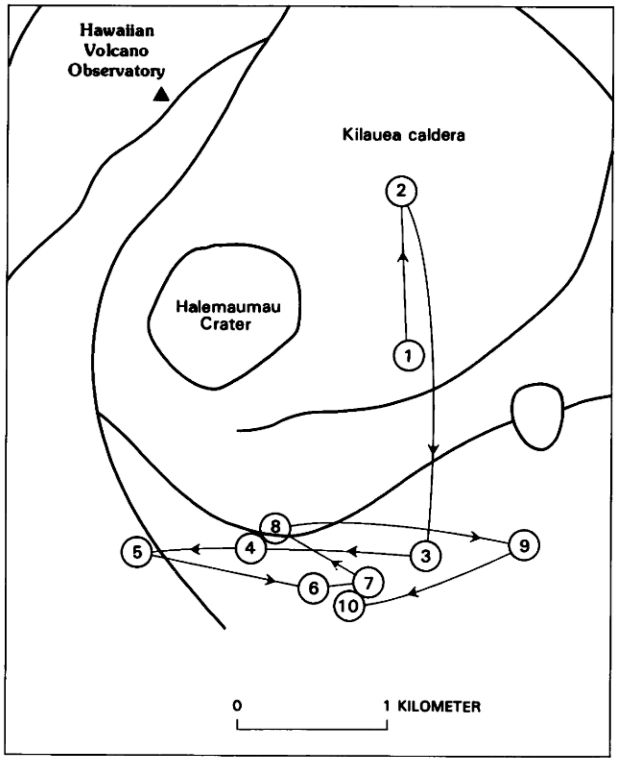


FIGURE 42.6.—Lateral shift of apex of uplift of inflating bulge at summit of Kilauea Volcano, modified from Fiske and Kinoshita (1969). Level surveys demonstrate shifting center of uplift: (1) January–July 1966; (2) July–October 1966; (3) October 1966–January 1967; (4) January–February 1967; (5) during February 1967; (6) February–May 1967; (7) May–June 1967; (8) June–July 1967; (9) July–September 1967; (10) September–October 1967. Position of Hawaiian Volcano Observatory and caldera and crater walls (heavy lines) shown for reference. Total uplift of the 10-km-wide bulge at its apex from January 1966 to October 1967 was approximately 70 cm; summit eruption of Kilauea began in November 1967.

data on Kilauea for the period 1966–1970 by simultaneous inversion of the level, tilt, and horizontal-distance measurements. They determined 32 centers of inflation and deflation, which clustered into two groups, one at the south end of Kilauea caldera at depths from 2–7 km and the other near Makaopuhi Crater on the east rift zone at depths of 3–8 km. Their results show that although the general deformation pattern is apparently elastic, there are departures from ideal elastic behavior. The differences between measured and calculated horizontal displacements indicates that a gradual north-south extension across the summit caldera occurred during 1966–1970.

Changes in gravity as much as a few tenths of a milligal accompany the inflation and deflation of the summit area of Kilauea (Johnson, chapter 47; Dzurisin and others, 1980; Jachens and Eaton, 1980). These changes are more complex than those expected from simple changes in elevation, and they indicate that mass that is not proportional to the volume changes can be both added to and removed from the summit region.

The volume of the shallow magma-reservoir system beneath Kilauea can be estimated by assuming that it is roughly spherical with a diameter similar to the 3.5-km diameter of the caldera. Because the magma-reservoir system is probably a network of magma bodies separated by screens of solid rock, only part of this 22-km^3 volume is estimated to be molten rock. A net volume of 11 km^3 would be the same as the estimate of magma-reservoir volume made by Wright (1984, p. 3238) on the basis of his summary analysis of the geodetic data.

Earthquake hypocenters also provide important and independent evidence on the shallow magma-reservoir system. Koyanagi and others (1976), on the basis of an aseismic zone largely surrounded by an envelope of seismic sources beneath Kilauea caldera, estimated that the zone of primary summit magma storage has a mean diameter of 3 km and extends from a depth of 3 km to 6 km beneath the surface. Their interpretation is based on the concept that the magma reservoir is too ductile to produce earthquake-generating fractures, whereas the rocks above and around the magma storage zone respond to inflation and deflation of the magma reservoir by brittle fracturing.

Ryan and others (1981) expanded this approach and produced a detailed model of Kilauea's magma-reservoir system from the surface to a depth of 15 km (fig. 42.7). Their model has a primary conduit extending vertically beneath the south end of the caldera, an upper-east-rift-zone pipe extending vertically through depths of 2–6 km beneath the Koae fault zone 1 km southwest of Kokoolau Crater, two horizontal ducts connecting the primary conduit and the pipe at depths of about 2 km and 6 km, and an extension of the upper horizontal duct beneath the east rift zone to the Mauna Ulu eruption site.

The main shallow magma-storage reservoir beneath Kilauea is apparently not completely aseismic. During times of rapid summit collapse and initial reinflation, swarms of peculiar microearthquakes have their apparent source within or very near the margins of the summit magma reservoir. These earthquakes generally have magnitudes less than 1 and have a low-frequency, emergent signal. They are difficult to locate because they are only recorded on a few seismographs near the summit of Kilauea and their onset signals are not clear enough to establish accurate arrival times. These swarms of long-period earthquakes are thought to be related to rapid readjustments among magma pockets within the plexus of the main magma reservoir.

The question of a deeper storage reservoir about 25 km beneath Kilauea remains unresolved. At present, deformation data are not accurate enough to detect small inflations and deflations of a zone that deep. The region at that depth is relatively aseismic, but no envelope of earthquake hypocenters surrounds it. Wright (1971) proposed a region of temporary accumulation of magma at a depth of 20–30 km beneath Kilauea to allow for removal of olivine and pyroxene (fig. 42.8) He called this region the intermediate staging area.

The dynamic nature of the well-established shallow magma-reservoir system beneath Kilauea is currently being investigated from several directions. Epp and others (1983) have shown that there is a

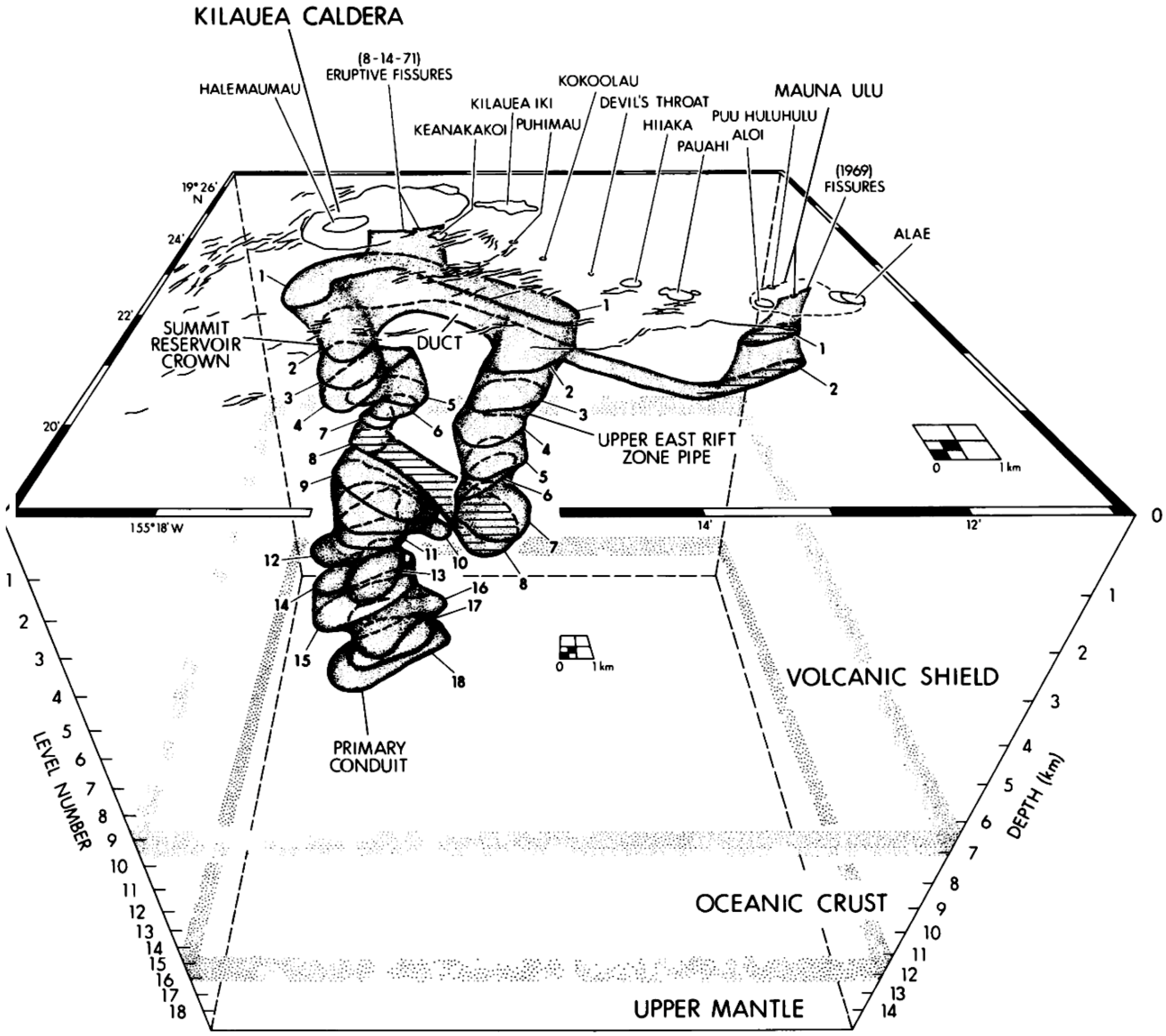


FIGURE 42.7.—Perspective view northward and downward into interior of Kilauea Volcano, from Ryan and others (1981). Magma conduits are defined by zones of earthquake hypocenters that surround them over periods of several years. Numbers on magma conduits represent horizontal sections at depths corresponding to level-number scale on left side.

strong linear correlation between the amount of summit deflation and the elevation of a flank eruption of short duration (fig. 42.9). Summit eruptions involve little or no deflation, whereas low-elevation flank eruptions are accompanied by large deflations. A simple explanation of this relation is that the magma pressure in the reservoir drops until it is balanced by the back pressure from the column of magma in the conduit from the reservoir to the surface. If this is the case, the magma pressure in the reservoir at the end of an eruption is simply the product of the average density of the magma column multiplied by the elevation difference and the gravitational constant. For example, assuming the average density of the magma column is 2.7 g/cm^3, the pressure in the magma reservoir 3 km below sea level beneath Kilauea following the eruption at Kapoho (30-m elevation) in 1960 would have been 83 MPa (830 bars); the lowest pressure in the magma reservoir during the past few decades. Since there are no major deflations associated with summit eruptions, there must be little or no pressure drop in the magma reservoir during these eruptions. The pressure of a magma column from the present floor of Kilauea caldera to a depth of 3 km below sea level would be 109 MPa. This difference of 26 MPa (109 minus 83) is the approxi-

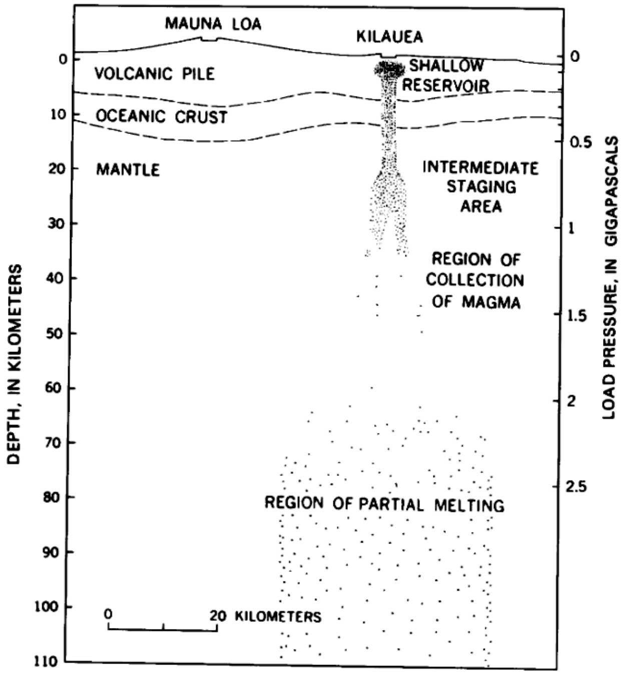


FIGURE 42.8.—Schematic cross section of Kilauea Volcano by Wright (1971) outlining regions of melting, transport, and storage of magma. Fractionation of magma occurs mainly in intermediate staging area and shallow reservoir, but also during its ascent from 60 to 30 km.

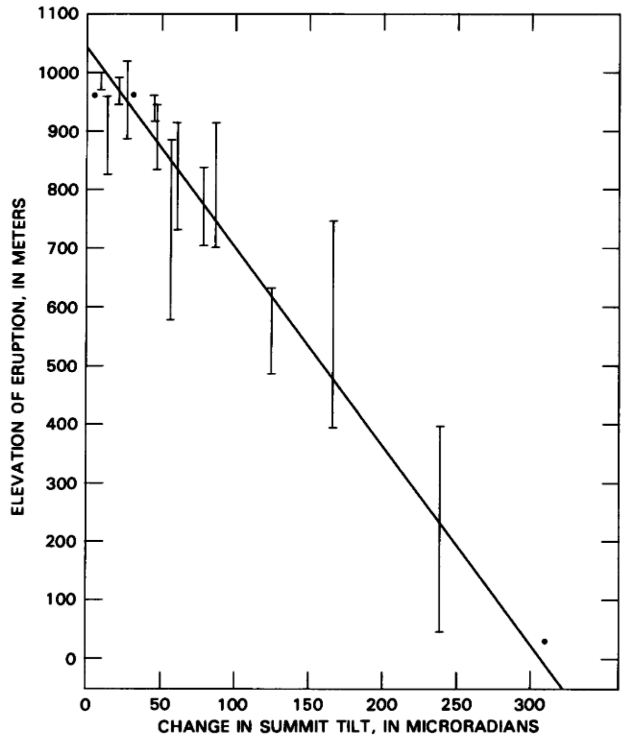


FIGURE 42.9.—Deflation measured by change in radial summit tilt at Kilauea Volcano versus elevation of associated vents during eruptions from 1955–1979. Bars indicate that eruptions occurred from fissures with considerable range in elevation. Linear relation of deflation and vent elevation indicates that pressure drop in summit magma reservoir is closely related to hydraulic head of magma at erupting vent. Modified from Epp and others (1983).

mate maximum range of pressure change in the shallow magma-reservoir system beneath Kilauea during the past few decades. In addition, the pressure change related to the amount of rapid deflation can be estimated from the data of Epp and others as 0.17 MPa (1.7 bars) per centimeter of summit deflation. Davis and others (1973, 1974, 1979) concluded from the absence of any marked piezomagnetic effects related to tilt changes at the summit of Kilauea that the volcano can support no large-scale pattern of shear stresses. They suggested that the rocks surrounding the shallow magma reservoir are of low shear strength and may behave in a plastic manner. Because the summit region of Kilauea probably does not behave in a perfectly elastic manner, particularly over time periods of months to years, slow inflations may represent volume changes more closely than pressure changes in the magma reservoir.

Magmafracting of the host rocks surrounding the magma reservoir appears to take place when the magma pressure reaches or slightly exceeds lithostatic pressure. This indicates low bulk tensile strength in the surrounding rocks.

The residence time for magma storage in the shallow reservoir system beneath Kilauea is easily estimated from the annual supply rate (0.11 km^3/yr, Swanson 1972; 0.086 km^3/yr, Dzurisin and others, 1984) and the volume of the reservoir system (11 km^3, Wright, 1984). The result, 100–120 yr, should be regarded as an average residence time. Some magma may crystallize in remote niches in the reservoir system and never erupt, while other magma batches, such as some of that which supplied the Kilauea Iki eruption in 1959, may spend only a small fraction of the average residence time in the magma reservoir system.

Another dynamic quality of shallow Hawaiian magma-reservoir systems on a vastly different time scale is their apparent ability to migrate upward as the volcano they supply grows in height. The submarine volcano Loihi is assumed to be in a youthful stage of growth, yet already a caldera and rift zones have apparently developed upon it (Klein, 1982a; Malahoff and others, 1982). This implies that it has probably developed a shallow magma-reservoir system. The depth to Loihi's magma reservoir is not known, but the summit area of Loihi is about 1 km below sea level. Mauna Loa Volcano, at the other extreme, has reached a height of 4,169 m, and recent studies by Decker and others (1983) indicate that it has a shallow magma-reservoir system similar to that beneath Kilauea. The source of inflation shown by deformation measurements before the 1984 eruption of Mauna Loa was at a depth of only 3 km beneath the summit. Thus, the top of the magma-reservoir system beneath Mauna Loa is at nearly the same elevation as the ground surface at the summit of Kilauea. The long-term evolution of Hawaiian volcanoes apparently involves progressive upward remelting, stoping, and (or) shoving aside of the rocks overlying the shallow magma-reservoir systems.

SHALLOW INTRUSIONS

As the shallow magma-reservoir system slowly inflates with magma added from depth, the increasing pressure and volume puts new stresses on the surrounding rocks. Sometimes the summit deflates slowly or remains level as magma moves almost aseismically into the open magma conduit of the upper and middle east rift zone of Kilauea. These slow intrusions or readjustments of the existing shallow magma-reservoir system involve volume flow rates of 1–10 m^3/s. When stresses reach the breaking strength of the host rocks surrounding the magma- reservoir system, fractures occur and a dike or sill is intruded upward or laterally into the shallow, brittle rocks of the volcanic edifice. These rapid intrusions are common at the summit and along the rift zones of Hawaiian volcanoes, and they occur at volume flow rates of 10^2–10^3 m^3/s. During 1959–1980, 21 rapid intrusions without eruptions occurred at Kilauea Volcano and 25 other rapid intrusions led to eruptions (Klein, 1982b).

The orientation of rapid, shallow intrusions is controlled by the strength of the rocks and by ambient stresses generated by both internal magma pressure and external gravitational forces. Fiske and Jackson (1972) have shown that the topography of the earlier volcanoes in Hawaii affects the orientation of the rift zones of later forming, adjacent volcanoes (see Peterson and Moore, chapter 7, fig. 7.7). Rift zones form perpendicular to the direction of least compressive stress; once a topographic ridge has formed from numerous dike-fed eruptions along a rift zone, the gravitational slumping of this ridge reinforces the least compressive stress in a direction perpendicular to the ridge. The name Mauna Loa means long mountain, and this great shield volcano is markedly elongated by the gently-sloping ridges along its northeast and southwest rift zones. Kilauea Volcano is younger than Mauna Loa Volcano and inherited the gravitational stress system on the southeast flank of Mauna Loa. Kilauea thus developed a principal east rift zone and subsidiary southwest rift zone.

A typical rapid intrusion into a rift zone of Kilauea Volcano is characterized by a swarm of small earthquakes occurring at rates of 2–4 events per minute that lasts for a few hours to a few days. The sources of these earthquakes are concentrated at 2- to 4-km depths beneath a segment of the rift zone several kilometers long. The hypocenters of the swarm outline a tabular shape a few kilometers high and about 1 km wide. The hypocenters also migrate progressively in time and space, both vertically and along the rift zone, indicating that the intruding fracture is propagating at a velocity of a few hundred to a few thousand meters per hour. The linear trace of intrusions that reach the surface as eruptions, and the eroded rift zones of older Hawaiian volcanoes, indicate that the rapid intrusions are steeply dipping dikes that parallel the rift zones. However, the shapes of the rapidly intruded bodies and the more open magma conduit along the upper and middle east rift zone of Kilauea below a few kilometers depth have not been established. The concentration of earthquake hypocenters at depths of 2–4 km in the seismic swarms associated with rapid intrusions suggests that the intrusion has some sort of axis at that depth.

In addition to their associated earthquakes, rapid intrusions are generally characterized by rapid deformation of the summit area and ground deformation above the intruded region. Volcanic tremor, new gas emissions and electromagnetic field changes also accompany some intrusions.

Rapid intrusions into the summit region of Kilauea Volcano are accompanied by rapid summit inflations, whereas rapid flank intrusions into the rift zones remove magma from the summit region and are accompanied by rapid subsidence of the summit area. Rapid intrusions into the rift zones of Kilauea have typical magma volumes of 10^6–10^7 m^3 and they occur at volume rates of 10^2–10^3 m^3/s. These typical volumes are associated with subsidence of the summit by 1–15 cm and pressure drops of 0.3–3 MPa in the shallow magma reservoir. An approximate rate of magma supply to a dike intruding the rift zones can be estimated from the tilt record. The radial tilt at HVO on the northwest side of Kilauea caldera has been calibrated as representing about 3.3×10^5 m^3 of subsided volume for each microradian (μrad) of inward tilt (Dzurisin and others, 1984). This is a minimum value for the amount of magma removed because of the unknown values of strength in the rocks surrounding the summit magma reservior and of the bulk compressibility of the reservoir. In addition, one microradian of inward radial tilt at HVO also indicates an approximate vertical subsidence of 5 mm at the apex of the deformation bulge and an approximate pressure drop in the summit magma reservoir of 90 kPa (0.9 bars).

Ground deformation in the area above a rapid intrusion causes rapid excursions and sometimes reversals of tilt vectors on nearby recording tiltmeters. Ground cracking is also common with the larger and shallower intrusions, and new or renewed volcanic-gas fumaroles are sometimes formed along these surface fractures.

When the volume rate of magma intrusion exceeds about 0.5 μrad of summit subsidence per hour (about 50 m^3/s), shallow volcanic tremor begins to appear as a background vibration between individual events in the earthquake swarm. Shallow volcanic tremor is a continuous ground vibration with a predominant frequency of 2–5 Hz, and it is associated with shallow subsurface movement and eruption of magma.

Increases in the electrical self-potential anomalies associated with active rift zones and changes in the electrical conductivity of the summit region of Kilauea also occur in association with the larger intrusions. The absence of observable magnetic anomalies associated with rapid subsidence of the summit area of Kilauea has led Davis and others (1974) to postulate a plastic envelope of low-rigidity rock material surrounding the summit magma reservoir.

During 1980 there were five rapid intrusions without eruptions into the east rift zone of Kilauea Volcano (table 42.1). During 1981 the intrusions shifted into Kilauea's southwest rift zone and on August 10–12 a large intrusion affected an 18-km-long segment of that rift. Excellent observations were made on this rapid intrusion, and they are summarized as follows:

TABLE 42.1.—*Rapid intrusions without eruptions into the east rift zone of Kilauea Volcano during 1980*

[*, possible eruption of a few cubic meters of lava; EM, electromagnetic; n.d., not determined]

Date	Location	Length (km)	Height (km)	Minimum depth (km)	Volume ($m^3 \times 10^6$)	Propagation rate (m/h)	Gases	EM anomalies
March 2	Hiiaka	2	2	1	.6	n.d.	no	no
March 10–12*	Mauna Ulu	5	3	.5	5.3	200–500	yes	no
August 27–28	Puhimau	4	3	.5	2.3	1,000	yes	yes
October 22	Mauna Ulu	2	2	2	.6	n.d.	yes	yes
November 2	Kokoolau	2	2	1	2.0	700	no	no

The August 10–12, 1981, event involved at least 4×10^7 m^3 of magma and was the largest rapid intrusion without an eruption of Kilauea Volcano since increased seismic and deformation monitoring began in the late 1950's.

Thousands of small earthquakes occurred in a swarm at rates as high as 3–4 per minute and although most were microearthquakes, a few reached magnitudes of 3.5–3.9. The hypocenters of the earthquakes in the region of the intrusion formed a zone 18 km long, 1–2 km wide, and extending from the surface to 5-km depth beneath the southwest rift zone (see Klein, chapter 43, fig. 43.90). Downrift migration of the earthquake locations indicates an intrusion propagation rate of 2 km/h slowing to 0.5 km/h (fig. 43.90).

Shallow volcanic tremor and rapid summit subsidence at about 10 μrad/h indicate a high-volume rate of intrusion, as much as 10^3 m^3/s. The tilt at HVO totaled 107 μrad of deflation, and an elastic model of the summit level changes indicates 4×10^7 m^3 of subsidence volume. The ratio of the volume rate to the downrift migration rate gives an estimate of the cross sectional area of the intrusion. This area is 2×10^3 m^2 , and if the intrusion is a dike 3 km high, then its width is about 0.7 m.

Large horizontal and vertical displacements occurred along the southwest rift zone, indicating major extension across the rift on the order of 1 meter; and a horst-and-graben pattern developed that indicates a shallow dike intrusion parallel to the rift and dipping steeply southeast. New surface cracks parallel to the rift follow the trend of the 1974 eruption vents and continue southwest for several kilometers (fig. 42.10). Much of the new surface cracking involves reactivation of 1974 or older fractures.

Renewed gas emissions and increased temperatures were measured at a fumarole on a newly activated surface crack. Self-potential anomaly changes were measured in the Koae fault system 5 km east of the trend of the seismic swarm; except near the renewed gas vent, no significant self-potential changes occurred over the seismic zone.

During 1982, rapid intrusions causing eruptions occurred twice in the summit area of Kilauea, in April and September. In June 1982 there was another large rapid intrusion without eruption into the southwest rift zone. Activity then switched back to the east rift zone and a major rapid intrusion and eruption occurred there during January 2–3, 1983. That event formed the magma conduit that has erupted through 47 episodes so far (as of June 1986) during the long-lasting 1983–1986 (and continuing) eruption of Kilauea Volcano.

The dike swarms formed by many rapid intrusions over the years tend to wedge apart the volcanic edifice and lead to long-term dynamic evolution of the rift zones.

RIFT ZONES

The rift zones of Hawaiian volcanoes are "Pele's under ground roads" referred to by Ellis' Hawaiian guides in 1823 (Ellis, 1827). They are persistent zones of weakness in the volcanoes' flanks, resulting from thousands of rapid lateral intrusions. Many of these rapid intrusions also cause eruptions.

The rift zones are not primary conduits that penetrate to magmatic source depths; rather they are secondary features that form from radial pressures generated in the summit reservoirs at depths of

FIGURE 42.10.—Aerial photograph of ground cracks formed during August 10, 1981, intrusion beneath southwest rift zone of Kilauea Volcano. Area here covered by barren lava flows and by reworked volcanic ash from 1790 eruption. Cracks are as much as 1 m wide, and a shallow graben about 20 m wide has formed between principal fractures. Some of these cracks existed before the August 10, 1981 intrusion and were reactivated by the event. U.S. Geological Survey photograph by Norman Banks.

a few kilometers. Earthquake locations along the rifts of Kilauea Volcano indicate that the main dynamic axes of the rifts are about 3 km beneath the surface and that the overall rift zones extend from the surface to a depth of about 8 km below sea level. Because of isostatic subsidence of the Island of Hawaii, 8 km is about the depth of the old Cretaceous sea floor on which the Hawaiian volcanoes formed. It appears that the rift zones penetrate virtually the entire volcanic pile but do not extend down into the underlying oceanic crust.

Fiske and Jackson's (1972) analysis of the evolution and orientation of Hawaiian rift zones is also consistent with the concept that the rift zones are fairly shallow and secondary features compared to the deep magma conduits that supply the summit magma reservoirs. In their view, the rifts are fed by lateral intrusions from the summit magma reservoirs, and their orientations are controlled by the gravitational stresses of the surface topography.

The north-south elongation of the submarine topography and seismicity of Loihi Seamount (Klein, 1982a) indicates that rift zones are generated in the early stages of life of Hawaiian volcanoes, and the major rift systems still active on Hualalai and Haleakala Volcanoes indicate that rift zones persist into the waning stages. Thus, although the rifts are secondary features, they play a major and long-lived role in the structure and dynamics of Hawaiian volcanism.

The rift zones of Kilauea Volcano have been very active during the past few decades, and their dynamic behavior is therefore fairly well known. From 1959 through 1984 there were 20 rapid intrusions into the east rift zone without eruptions and 19 rapid intrusions that resulted in eruptions. During the same period there were 8 rapid intrusions into the southwest rift zone without eruptions and 2 rapid intrusions that resulted in eruptions. Summit activity over this same period consisted of 4 rapid intrusions without eruptions and 12 with eruptions. Eleven of these intrusions beneath the summit area were emplaced between 1966 and 1984.

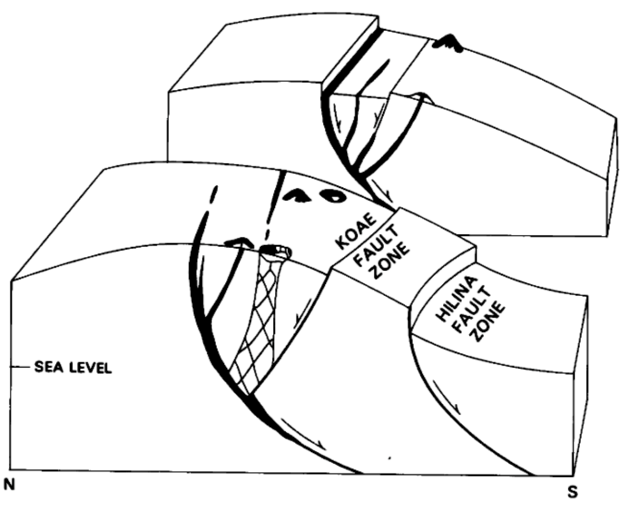


FIGURE 42.11.—Schematic block diagrams of the east rift zone of Kilauea Volcano looking downrift, from Moore and Krivoy (1964). In this interpretation stresses causing rift are largely gravitational, and rift zone dips seaward at depth. Pit craters (cross-hatched example) form by collapse into voids within rift if magma is drained far downrift.

Since each rapid intrusion and eruption involves a dike being injected into the carapace or flank of the volcano, the volcano must widen to accommodate these multiple intrusions. Using an estimate of 0.5 m for an average dike width, we can infer that the summit and upper rift zones of Kilauea Volcano should have widened by 5.5 m over this 19-year period. The actual measured widening of Kilauea caldera, between HVO on the northwest rim and Ahua on the southeast is 4.2 m from 1966 to 1984. There are various possible reasons why the estimate and the corrected measurement differ: the average dike width may be less than 0.5 m; each dike probably extends only part way across the summit area; and there probably is some elastic compression stored in the rocks adjacent to the intruded dikes. Even allowing for these considerations, it is clear that major widening of the summit of Kilauea Volcano perpendicular to its rift zones has occurred in the past few decades.

The widening of the rift zones becomes less downrift because fewer dikes reach long distances from the summit. The fact that some summit eruptions are followed by progressive downrift migration of rapid intrusions and flank eruptions can be explained by this progressive summit wedging effect extending down the rift zones.

Eaton and Murata (1960) suggested that a core of magma in the east rift zone of Kilauea remains molten between some eruptions, and Swanson and others (1976b) discussed this concept more thoroughly on the basis of seismic evidence from 1968 and 1969 eruptions on the east rift zone. Hardee (1982) has shown by theoretical heat-flow calculations that if a sufficient number and volume of intrusions are injected into a region, part of these intrusions will remain molten. He calculated (written commun., 1983) that the recorded intrusions into the east rift zone of Kilauea Volcano should result in a molten conduit from the summit to about 30–40 km downrift. The number and volume of intrusions into the southwest rift zone, however, have not been sufficient to maintain a molten conduit. The evidence for slow intrusions into the east rift zone provided by slow aseismic deflations of the summit and measurable inflation along the east rift zone as far as 32 km from the summit (Dieterich and Decker, 1975; Swanson and others, 1976a) give strong support to Hardee's conclusion.

Moore and Krivoy (1964) were among the first to consider the dynamics of the east rift zone of Kilauea Volcano. They suggested that gravitational slumping was the major factor causing the rift to widen and that magma moved rather passively into the fractures formed in the head of the slumping block (fig. 42.11). Swanson and others (1976a) took the opposite position that magmatic pressure pushed the rift apart and that slumping was the result rather than the cause of the rifting process (fig. 42.12). In a sense they are all correct; the gravitational and magmatic stresses are additive, and together they control the evolution of the rift zone.

Swanson and others (1976a) measured the progressive shortening of survey lines across the Hilina fault system on the south flank of Kilauea Volcano south of the east rift zone. They recognized that

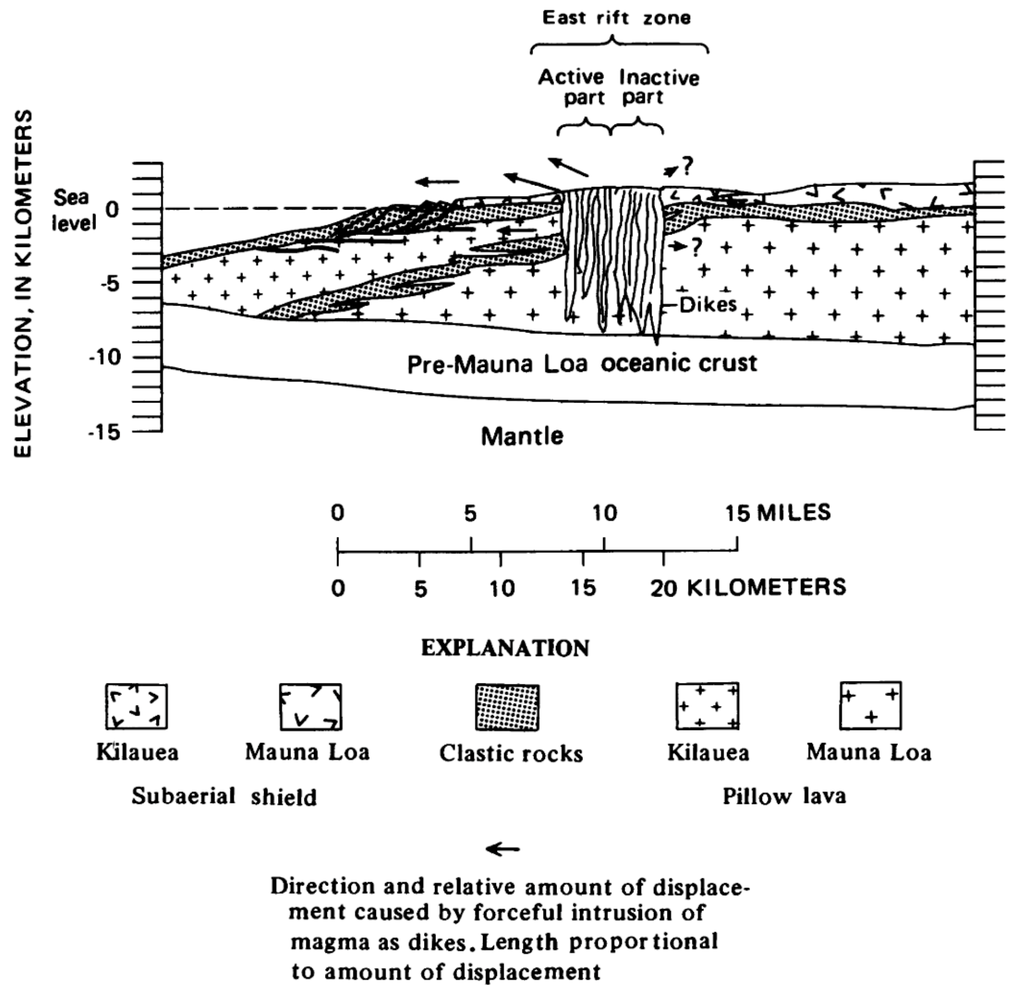


FIGURE 42.12.—Schematic cross section at true scale of east rift zone of Kilauea Volcano looking uprift, from Swanson and others (1976). In this interpretation stresses causing rift arise from forceful injection of magma, and dikes are nearly vertical.

the intrusions and eruptions occurring on the east rift zone in the late 1960's and early 1970's were wedging the rift apart but that much of the widening was being accommodated by compression and uplift of the flanks of the rift zone. They realized that this accumulating strain might be released by sudden failure of the Hilina fault system, and they wrote in 1974 (Swanson and others, 1976a, p. 35): "Some of the data suggest that the Hilina fault system is poised for another episode of subsidence, with only minor additional ground displacement needed to trigger it. Most of the strain acquired since 1965 remains, however, and we anticipate a subsidence (strain-release) event of unknown magnitude in the not too distant future."

Fifteen months later, but before their work was formally published, a magnitude-7.2 earthquake released the pent-up compressive strain in the south flank of Kilauea. Portions of the south coast moved seaward several meters and subsided as much as 3.5 m (Tilling and others, 1976; Lipman and others, 1985). Swanson and others (1976a) had clearly forecast the nature and imminence of that major earthquake.

Analysis of the source mechanism and aftershock pattern for this earthquake (Ando, 1979; Furumoto and Kovach, 1979) indicates that the strain was released by rupture on a nearly horizontal fault about 8–10 km beneath the south flank of Kilauea along 50 km of the coastline and across a region 10–20 km wide, bounded on the north by the east and the southwest rift zones. The slip on the fault was several meters in a seaward direction away from the rift zones (fig. 42.13). The net long-term result of dikes wedging into a rift zone therefore appears to be widening of the rift zone down to a depth of about 8 km, and this widening is accommodated by intermittent movement of the seaward flank of the volcano along a nearly horizontal slip surface close to the old sea floor on which the volcano grew.

More than 3 m of extension occurred on survey lines on the

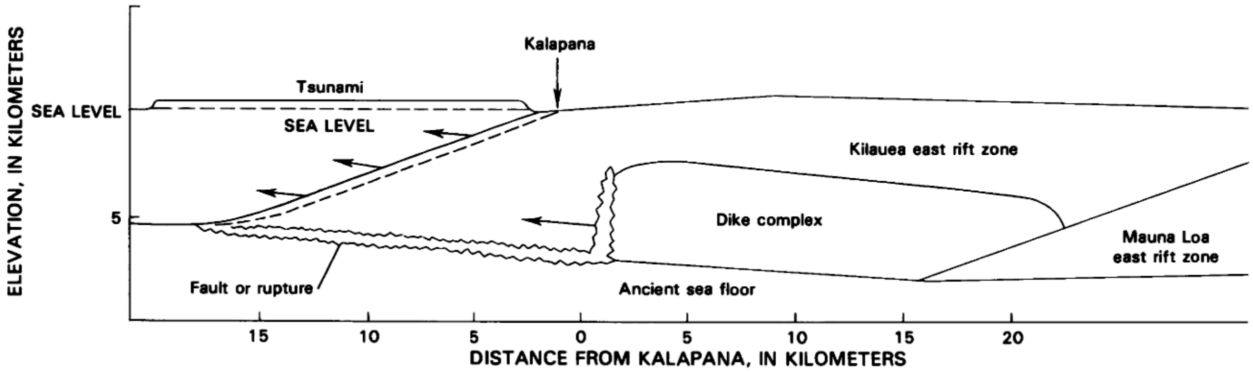


FIGURE 42.13.—Schematic cross section of east rift zone of Kilauea Volcano looking uprift and showing inferred mechanism of 1975 earthquake (M=7.2) and tsunami. Dike intrusions (at left side of dike complex) compress adjacent rocks over time. When stress reaches breaking point, unbuttressed flank of rift zone suddenly fails. Modified from Furumoto and Kovach (1979).

south flank of Kilauea Volcano during the period of the major earthquake; figure 42.14 shows changes from 1970 to 1984 in the 8.6-km-long line that extends from the north side of the Hilina fault system to the south coast.

It appears that Hawaiian volcanoes do not only grow upward from successive lava flows piled one upon another, but they also grow sideways by successive intrusions into the summits and rift zones and by intermittent failure of the seaward flanks of the rift zones to accommodate these intrusions. The east rift zone of Kilauea Volcano slopes down gently for 50 km from the summit at 1,200 m to sea level. It continues for another 70 km eastward as a slightly steeper submarine ridge to depths of nearly 5,000 m below sea level (Moore, 1971). Since Epp and others (1983) have shown that eruptions at lower elevations on the rift tend to be more voluminous than single short-lived eruptions high on the rift, it is clear that eruptions must become progressively much rarer at large distances from the summit. This decrease is probably caused by at least two factors: the widening of the rift zone and its wedging effect apparently decrease with distance from the summit; and the molten core in the rift zone also apparently decreases in size and finally disappears with distance from the summit. The very even slope of the ridge along the axis of the east rift zone indicates that the frequency and volumes of eruptions maintain a steady equilibrium over long time spans.

ERUPTIONS

Most volcanic eruptions in Hawaii are outpourings of incandescent lava from linear or point-source vents. Lava fountains from linear vents are called curtains of fire and reach heights up to about 200 m. Lava fountains from point-source vents reach heights up to about 600 m.

A reduction in the volume rate of eruption or in the gas content reduces the vigor of lava fountaining until incandescent lava simply pours out from the linear or point-source vent. Vents in the bottoms or sides of craters feed active lava lakes that, at low-volume rates of eruption, can remain in eruption for years.

Lava flows fed by the erupting vents move down the slopes of the volcano or pond in topographic depressions. The formation of various lava textures and structures such as aa, pahoehoe, and lava tubes depends upon the complex interaction of the rheological properties of the lava, the volume rate and continuity of the eruption, and the environment and topography over which the flows are moving (Swanson, 1973; Holcomb and others, 1974; Peterson and Swanson, 1974; Peterson and Tilling, 1980). Lava commonly drains back down old surface cracks or even down the active fissures as the eruption wanes.

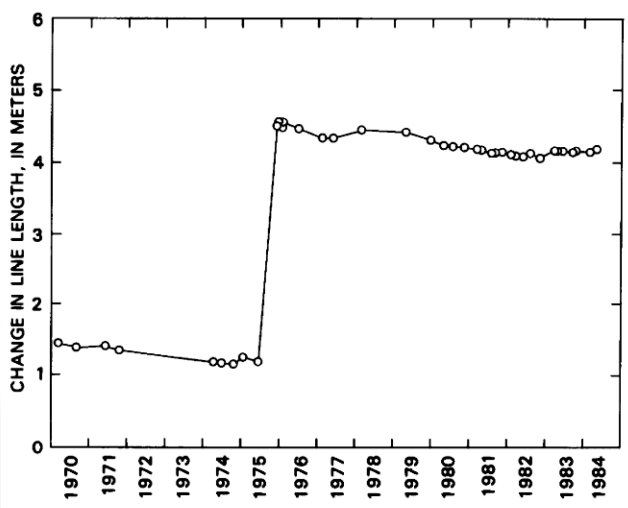


FIGURE 42.14.—Multiple surveys (small circles) of 8.6-km-long line on south flank of upper east rift zone of Kilauea Volcano indicate gradual contraction from 1970 to 1975, sudden extension related to the M=7.2 earthquake of November 1975, and renewed gradual contraction since 1975. Surveyed line runs from Goat to Apua Point.

Data on the eruptions of Kilauea and Mauna Loa Volcanoes since 1823 are compiled in Peterson and Moore (chapter 7, tables 7.3, 7.4). The long-lived, low-volume-rate eruptions of active lava lakes in the caldera of Kilauea Volcano that continued with a few interruptions from their first-recorded observation in 1823 until 1924 are not included in table 7.4.

The quality of these data on eruptions becomes, in general, better for more recent times. Continuous scientific observation began in 1912, and aerial observation and helicopter access to remote eruption sites has increased progressively during the past few decades.

Several major changes in the character of eruptions of Kilauea are evident in table 7.3. Between 1823 and 1919, during the century of active lava lakes and slow filling of Kilauea caldera, there was only one confirmed east-rift-zone eruption (in 1840) and one southwest-rift-zone eruption (1868). Beginning in 1919 and culminating with a probable east-rift submarine eruption in 1924, three rift eruptions marked the end of a long period of predominance of summit activity. The 1924 summit eruption, associated with subsidence of more than 400 m in the magma column feeding the active lava lake in Halemaumau Crater, has been the only explosive eruption in Hawaii since written records began in 1823.

From 1924 to 1934, seven small eruptions filled the bottom of the deep Halemaumau Crater formed by subsidence in 1924; from 1934 until 1952 there were no eruptions of Kilauea. This hiatus was ended by two eruptions in Halemaumau Crater in 1952 and 1954, and then a major eruption on the lower east rift zone in 1955 began a still-continuing period (1955–1986) of major eruptive activity on the east rift zone (18 eruptions), the summit (12 eruptions), and the southwest rift zone (2 eruptions).

What constitutes a separate eruption of Kilauea is not always clear. The breakout of an eruption in an entirely new location—not just an extension of a fracture associated with an ongoing rapid intrusion—is one criterion. Another criterion is the occurrence of a seismic swarm at the beginning of an eruption in a familiar location, like Halemaumau Crater. This indicates that a new subsurface fracture has been opened to feed the surface eruption. In long-lived eruptions such as the Mauna Ulu eruption on the upper east rift zone from 1969 to 1974 and the current Puu Oo eruption on the middle east rift zone from 1983 to 1986 (and still continuing), there have been periods of repose between episodes of high lava fountaining or other changes in the character of the eruption phases. However, the same general conduit system appears to be involved throughout these continuing eruptions, and on this basis these distinct episodes or phases are not considered separate eruptions.

Under these limits of definition, the eruption statistics for Kilauea from 1918 through 1984 are as follows: The average time from the beginning of one eruption to the next is 494 days (range is 2 to 6,504 days). The average duration of an eruption is 86 days (range is less than 1 to 1,884 days). The average volume of an eruption is 2.7×10^7 m^3 (range is less than 10^5 to over 4×10^8 m^3). The average area coverered by an eruption is 4.4 km^2 (range is 0.1 to 55.7 km^2).

Dzurisin (1980) has compared tidal stresses to the onset times of eruptions of Kilauea and concludes that more eruptions begin near maximum fortnightly tidal stress (new moon and full moon) than would be expected by random chance at the 90-percent confidence level. He ascribes this to the triggering effect of tidal stresses on a system that has almost reached the point of eruption.

Klein (1982b) has investigated the statistical pattern of the eruptions of Kilauea since 1918 and concludes that the overall pattern is largely random, but that some non-random features appear. He finds that summit eruptions tend to cluster, that large-volume eruptions are usually followed by repose times that are longer than average, and that long periods of increased activity of Kilauea—for example during 1960–1979, when the average repose period was only 261 days—are associated with periods of less vigorous activity of Mauna Loa Volcano.

A complete list of known historical Mauna Loa eruptions is presented in table 7.4. Mauna Loa has not had any sustained lava-lake activity during this time period. There is less apparent variation in the pattern of Mauna Loa eruptions than in that of Kilauea eruptions, but the eight summit eruptions of Mauna Loa that occurred between 1870 and 1876 show an unusual clustering of locations and shorter-than-average repose times. In addition, the complete lack of eruptions between 1950 and 1975 was an unusually long repose period for Mauna Loa.

Statistics for Mauna Loa eruptions are as follows: The average time period from the beginning of one eruption to the next is 1,459 days (range is 104 to 9,165 days). The average duration of an eruption is 69 days (range is less than 1 to 547 days). The average volume of an eruption is 1.88×10^8 m^3 (range is 2.7×10^7 to 6.88×10^8 m^3). The average area covered above sea level by an eruption is 34 km^2 (range is 5 to 91 km^2).

All historical eruptions of Kilauea and Mauna Loa Volcanoes appear to have started with a rapid intrusion injecting a dike upwards or sideways from some part of the shallow magma reservoir into the summit or flank of the volcano. These rapid intrusions precede the eruptions by minutes to days, and provide important warnings of both the probable imminent occurrence and the probable location of an eruption.

The character of an eruption depends on several factors and is best explained by examples:

The long-lived eruptions of active lava lakes in Kilauea caldera begin with a dike injection into the summit area. Pressure in the shallow magma reservoir is high enough to push magma to the surface, and the resupply rate from depth is sufficient to match the low-volume rate of eruption, which is typically about 3 m^3/s. The rift zones are strong enough to withstand magmafracting. These eruptions have become less common as Kilauea caldera has refilled because the higher the magma column, the more pressure there is to fracture the rift zones and thereby lower the magma column. The latest long-lived eruption of an active lava lake occurred in Halemaumau Crater in 1967 and lasted 251 days (Kinoshita and others, 1969).

Short-lived eruptions at the summits and along the rift zones of Kilauea and Mauna Loa Volcanoes are the most numerous. The volume rates of these eruptions are typically 10–200 m^3/s; since this exceeds the resupply rate from depth, pressure rapidly drops in the shallow magma reservoirs. The eruptions stop when the reservoir pressure can no longer push magma to the surface. The latest eruption of this type on Kilauea Volcano occurred in the caldera in September 1982 and lasted 16 hours (Scientific Event Alert Network, 1982). The latest one of this type on Mauna Loa Volcano occurred along the northeast rift zone in March–April 1984 and lasted 21 days (Lockwood and others, chapter 19).

Although long-lived flank eruptions are more unusual, two have occurred on Kilauea Volcano since 1969: the Mauna Ulu eruption on the upper east rift zone from 1969 to 1974 (Tilling and others, chapter 16; Peterson and others, 1976; Swanson and others, 1979), and the Puu Oo eruption on the middle east rift zone from 1983 to 1986 and still continuing (Wolfe and others, chapter 17). These eruptions consist of many episodes from essentially the same vent area. The subsurface magma conduits apparently remain open between eruptive events because the onsets of new episodes are not preceded or accompanied by earthquake swarms. Individual episodes of lava fountaining involve volume flow rates of 50–400 m^3/s, but they generally last only about 10–20 hours and produce lava volumes of about 5×10^6–15×10^6 m^3. The longer term volume flow rates through several fountaining episodes or during sustained low-volume-rate phases average about 3 m^3/s, which is approximately the resupply rate of Kilauea Volcano for the past two decades.

The spectacular lava fountains and curtains of fire that characterize many Hawaiian eruptions are caused almost entirely by the rapid effervescense of gases boiling out of the magma at near-surface pressures and by the geometry of the vents. High-volume-rate eruptions of gas-rich magma in constricted vents produce the highest fountains. Gerlach and Graeber (1985), and Gerlach (written commun., 1985) have determined that magmatic gases of the volume and composition normally occurring in flank eruptions of Kilauea Volcano begin to boil rapidly out of magma at pressures equivalent to depths of only 40–100 m. In addition, as Paul Greenland has noted (written commun., 1984), near-surface pressures will move downward into an open magma column as rapid lava fountaining progressively unloads the upper part of the column, and the actual depth at which sustained effervescense is formed will be deeper than 40–100 m below the surface.

Most Hawaiian eruptions begin with linear vents (curtains of fire). After a few hours or days the fountaining becomes restricted to one or a few principal point-source vents. Apparently zones along the erupting fissure that were wider than average at the start of the eruption become enlarged by melting and erosion, while the narrower zones along the original fissure tend to freeze shut (Delaney and Pollard, 1982).

Once the magma flow supplying a fountaining vent is no longer replenished from depth, the fountaining will cease, and degassed surface lavas ponded near the vent will drain back into the conduit. Drainback in an active lava lake may also occur if the feeder dike is extended or some newly formed dike is generated in the subsurface. Magma in hydraulic communication will move toward the lowest pressure region in the system.

CALDERA COLLAPSE AND EXPLOSIVE ERUPTIONS

Hawaiian volcanoes have a reputation for producing numerous but relatively harmless eruptions of fluid lava. Explosive eruptions, similar to those at Surtsey Volcano in Iceland in 1963 (Thorarinsson, 1965), were probably common during the transition from submarine to subaerial eruptions in the Hawaiian Islands, but these usually cease after the summit of a volcano grows above sea level. Less than 1 percent of Hawaiian volcanic products above sea level are pyroclastic (Macdonald, 1972, p. 355), and most of these are tephra from high lava fountains. Nevertheless, major explosive eruptions of hydromagmatic and hydrothermal character have occurred in Hawaii. Their occurrence appears to be closely related to episodes of major caldera collapse.

The last major collapse and explosive eruption of Kilauea caldera occurred in 1790 (Swanson and Christiansen, 1973; Christiansen, 1979; Decker and Christiansen, 1984). The caldera was more than twice its present depth when visited by Ellis in 1823 (Ellis, 1827), but its depth and configuration prior to 1790 are not known.

Holcomb's paleomagnetic data (1980) indicate that lava overflows were occurring from the summit region of Kilauea during the period of about 1–0.3 ka, and Ellis' Hawaiian guides told him that "in earlier ages it used to boil up, overflow its banks, and inundate the adjacent country" (Ellis, 1827, p. 171).

A much smaller collapse accompanied by small explosive eruptions occurred at Kilauea in 1924 (Jaggar and Finch, 1924). A nearly full lava lake in Halemaumau Crater drained down more than 400 m as a major intrusion was injected into the lower east rift zone of Kilauea near Kapoho. This intrusion possibly fed an eruption on the submarine extension of the east rift zone, but there was no direct evidence of this.

As the magma column in Halemaumau Crater drained below zones of subsurface water beneath the summit of Kilauea Volcano, the equilibrium between the magma column and subsurface water was destroyed and steam explosions of hydromagmatic and hydrothermal origin issued from the crater (fig. 42.15).

The subsidence in 1924 occurred during February through April, and the multiple steam explosions occurred for 18 days in May. Since magma moves out of the summit reservoir through conduits in the rift zones, the volume rate of magma removal is limited and the process of caldera collapse occurs over periods of weeks to months rather than minutes to hours. In theory there should be time for people to evacuate the summit region of Kilauea in case of future caldera collapses.

Overflows of lava that were erupted from the summit region of Mauna Loa Volcano as recently as 590 years ago as determined by radiocarbon dating (J.P. Lockwood, oral commun., 1985) are beheaded by Mauna Loa caldera. When first mapped in 1841 (Wilkes, 1845), the caldera was 50–100 m deeper than the present surface of ponded lavas. Thus it appears that the present caldera on Mauna Loa is similar in many ways to Kilauea caldera; it formed only a few hundred years ago and has been rapidly refilled with

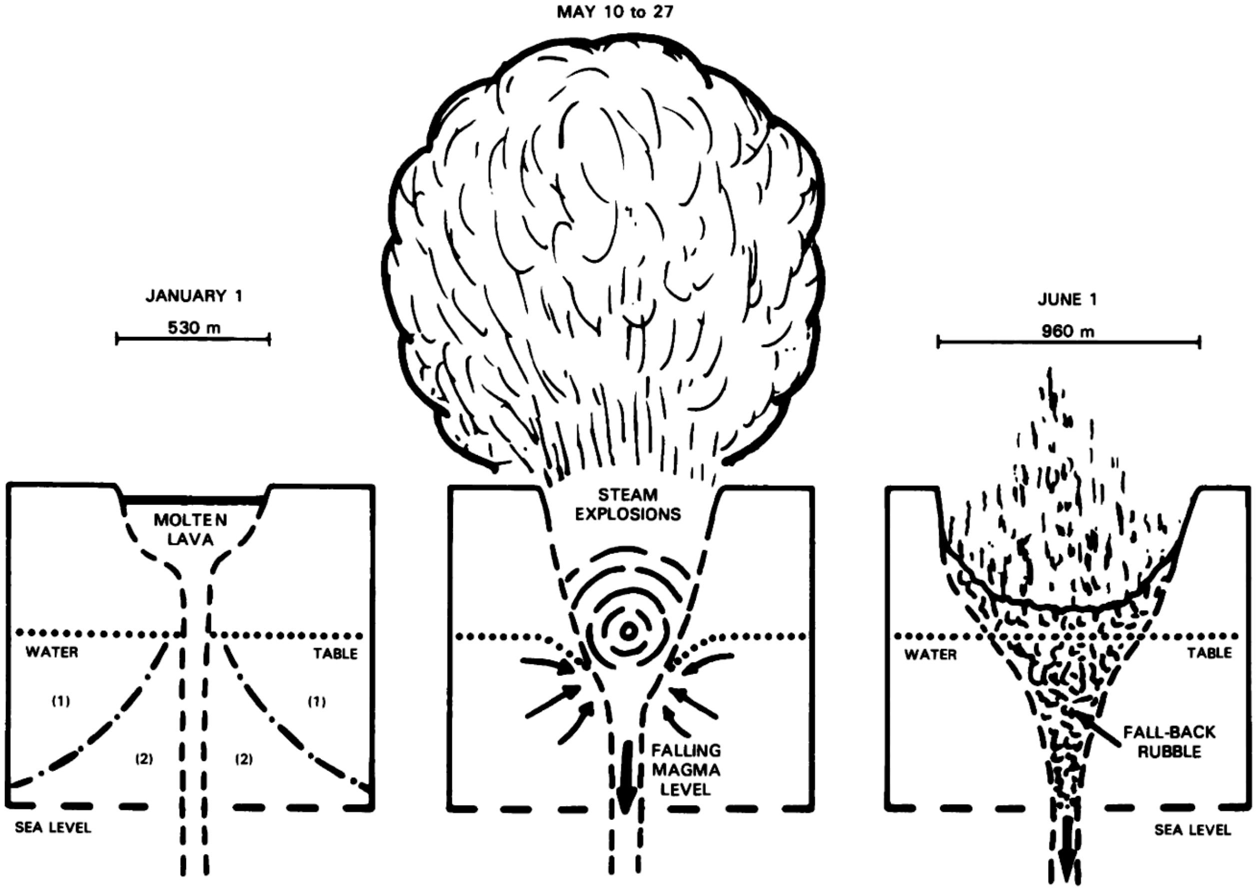


FIGURE 42.15.—Schematic cross sections showing conditions at Halemaumau Crater before, during, and after 1924 explosive eruption. On January 1, one zone of subsurface water (1) is cooler than boiling temperature, and other zone (2) is at boiling temperature for that depth; dash-dot line marks approximate boundary. During May 10–27, as magma column subsides below water table, equilibrium between subsurface water and magma is destroyed, and water from zone 1, rushes into broken hot rocks surrounding subsiding magma column and flashes into steam. Water from zone 2 also flashes into steam as hydrostatic pressures are suddenly reduced. Explosions are multiple because of progressive lowering of the magma column; they stop in late May with exhaustion of thermal energy above the subsided magma column, leaving a collapsed crater. From Decker and Christiansen (1984).

ponded lavas. There are scattered blocks of explosive ejecta on the rim of Mauna Loa caldera, but any explosive eruptions that may have accompanied the caldera collapse were minor compared to the 1790 eruption of Kilauea. This difference in explosive habits of Kilauea and Mauna Loa Volcanoes is probably related to the greater depth from the surface to zones of subsurface water beneath Mauna Loa.

Major caldera collapses of Kilauea Volcano have occurred repeatedly, with an average recurrence time of about 2,000 years (Decker and Christiansen, 1984; Dzurisin and Casadevall, 1986). Refilling the caldera with ponded lavas apparently takes several hundred years. Since both the young submarine volcano Loihi and Kilauea's older neighbor volcano Mauna Loa have calderas, this repeated process of caldera collapse and refilling apparently occurs during most of a Hawaiian volcano's active life span—roughly 0.5–1 m.y. If the caldera recurrence interval of 2,000 years at Kilauea is typical, this indicates that 200–500 generations of major caldera collapse and refilling occur during the evolution of a Hawaiian volcano.

This concept of repeated caldera collapse and refilling, combined with the evidence of progressively upward migration of the shallow magma reservoir systems of Hawaiian volcanoes (fig. 42.16), suggests that some lavas ponded in early calderas may be partly reassimilated into the summit magma-reservoir systems.

SUBSIDENCE OF THE ISLANDS

Hawaiian volcanoes form a massive load on the lithosphere of the Pacific plate. The great positive gravity anomalies of young Hawaiian volcanoes—as much as 330 mGal Bouguer (Kinoshita, 1965)—indicate that they are not compensated by local roots of less dense rock. As Hawaiian volcanoes form, therefore, they also begin

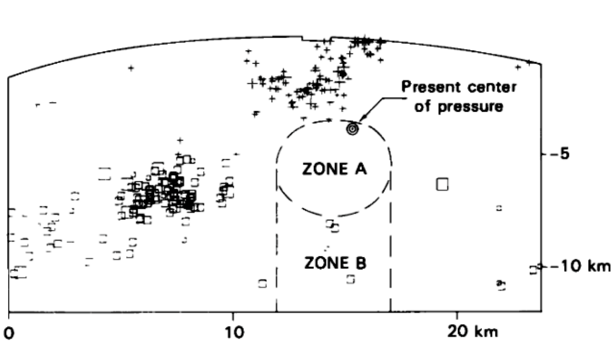


FIGURE 42.16.—Schematic cross section at true scale of inferred magma-reservoir system beneath summit of Mauna Loa Volcano. Zone A is more active region of magma addition and removal. Zone B is more passive; it contains magma and allows magma to move through it, but does not increase or decrease in volume as does zone A. Zone A migrates upward as the volcano grows in height. Pluses and squares represent well-located hypocenters of earthquakes (M=2 or greater) that occurred from January 1962 through May 1983 within 2.5 km of the cross-section plane. From Decker and others (1983).

to sink, and this process has been largely completed when the volcano becomes extinct.

Moore (chapter 2) and Moore and Fornari (1984) have determined from tide-gauge and drowned-reef data that the average rate of subsidence of the west shore of the Island of Hawaii is about 2 mm/yr over the past 150,000 yr and that the rate of subsidence has been generally increasing during this time period. The subsidence is apparently more rapid near the center of the island than at the shoreline. Eustatic sea-level changes are superimposed on this general subsidence and complicate the actual change of sea level. For example, the worldwide rise in sea level over the past century added to the subsidence of the Island of Hawaii has caused a rise in sea level at Hilo of about 4 mm/yr during the past few decades. Tide-gauge data at Honolulu reflect a eustatic rise in sea level but do not indicate any measurable subsidence of the Island of Oahu.

Very high sea-level stands reported by Stearns (1978), which imply periods of major uplift following the subsidence of some Hawaiian islands, have recently been interpreted by Moore and Moore (1984) as deposits from giant waves. Present opinion favors general subsidence of the Hawaiian islands rather than some complex oscillating tectonic movements.

Tectonic subsidence probably results from at least four factors: (1) removal of magma from beneath the active volcanoes; (2) elastic compression of the lithosphere nearly as fast as load is added; (3) plastic flow in the lower lithosphere and upper mantle as a result of volcanic load over periods of 10^5–10^6 yr; and (4) the slow thickening of the oceanic lithosphere and its subsidence as it loses heat over periods of 10^7 to 10^8 yr (Yoshii and others, 1976). Liu and Kosloff (1978) calculated factors (2) and (3) using an elastic/plastic plate-bending model with laboratory data on rock deformation.

This concept of general subsidence of the Hawaiian Islands has some interesting geologic implications. If any mineral deposits were ever formed by interaction of magma and subsurface water, these deposits would probably not be exposed above sea level by later erosion. In addition, subaerial erosion would in general not expose any Hawaiian volcanic rocks or structures formed below sea level.

FUTURE RESEARCH DIRECTIONS

As in most geologic problems, our understanding of how Hawaiian volcanoes work decreases with depth below the surface. Nevertheless, there are some research paths that look promising over the next decade or two for providing new insights into the structure and dynamics of Hawaiian volcanoes.

More detailed information on the variations in temperature and kinetics of the Earth's mantle should be forthcoming from worldwide seismic tomography. The deep anomalies beneath the central Pacific, emerging in the present work of Anderson and Dziewonski (1984), should become clearer if a worldwide net of digital seismometers is deployed, particularly at sites on the ocean floor.

Deployment of a few ocean-bottom seismometers around the Island of Hawaii would also greatly improve the capability of the present seismometer network of the Hawaiian Volcano Observatory. Potential specific benefits from this type of research effort include better location of earthquake sources beneath Loihi Seamount, the ability to determine the source functions of deeper earthquakes beneath Hawaii, and the ability to gain more depth and detail for the seismic velocity structure beneath Hawaii.

Volcanic tremor speaks in a language we do not yet clearly understand, but the technology probably exists to translate this babble into meaningful words and sentences. Arrays of portable digital broad-band seismometers are needed to record volcanic tremor in sufficient detail to interpret its source functions. With such data in hand, we may be able to resolve some of the ambiguity in present models of how volcanic tremor is generated.

Recent developments in satellite geodesy suggest that a revolution in the technology of measuring the surface deformation on active and potentially active volcanoes is only a few years away. The ability to measure distance changes between points that are not intervisible, and the potentially great consequent improvements in the quality of far-field data, would make observations possible that cannot now be made with conventional surveying devices. These new observations are essential for testing more complex theoretical models of the size, shape, and depth of subsurface magma bodies. Add to this the great potential savings in surveying time and manpower, and the attractiveness of satellite global positioning systems becomes clear.

The recent successes in observing detailed gravity changes related to activity of Kilauea and Mauna Loa Volcanoes (Johnson, chapter 47; Dzurisin and others, 1980; Jachens and Eaton, 1980), and the fact that these new data indicate important mass changes in the shallow magma-storage systems not quantitatively shown by surface deformation measurements, clearly demonstrate the importance of long-term measurement of gravity changes on the summits and rift zones of Kilauea and Mauna Loa Volcanoes. Recording gravity meters in conjunction with recording tiltmeters would provide the most definitive data on the mass and volume changes in the shallow magma reservoir systems.

Significant changes in the electrical, magnetic, and electromagnetic fields associated with the activity of Kilauea Volcano (Davis and others, 1973; Zablocki, 1976; Dzurisin and others, 1980; Jackson, 1983) indicate that these methods of geophysical research should continue to be pursued vigorously at HVO. Needed most in this research are a comprehensive analysis of the available data and the generation of theoretical models to try to explain the data. This type of effort would define more clearly what types of field observations are most important.

Great progress has been made in the past few years in determining the composition of gases in Hawaiian magma, and in understanding the complex manner in which some of these gases exsolve from the melt and escape to the atmosphere (chapters 27–36; Gerlach and Graeber, 1985) These studies have set the stage for determining the role of volcanic gases in the dynamics of the ascent, the shallow storage and intrusion, and the eruption of magma. Carbon dioxide apparently forms a separate but dense fluid phase at depths as great as 40 km beneath the summits of Hawaiian volcanoes. Studies of the thermodynamic and mechanical behavior of oxygen, carbon, hydrogen, and sulfur in the generation, ascent, shallow storage and intrusion, and eruption of Hawaiian magmas is a research opportunity of great importance.

The destruction of homes and property in the 1983–1985 eruption of Kilauea Volcano, and the threat to Hilo from flows of the 1984 Mauna Loa eruption, make it clear that we need to know more about the mechanical processes of Hawaiian lava flows in order to predict their behavior better. The complex interaction of the volume rate of eruption, the rheological properties of lava, and the topography over which the flows are moving combine to make a challenging research area that needs more study. Careful plans must be made for future experiments. When an eruption begins it is usually too late to start planning.

As a parting note, one area of the dynamics of Hawaiian volcanism that has been almost entirely overlooked is the study of the interaction of magma with subsurface water beneath the summits and rift zones of the active volcanoes. Few direct data are available: only one 1,260-m drill hole into the summit of Kilauea (Keller and others, 1979), and a few geothermal wells on the east rift zone of Kilauea whose data are still largely proprietary. Some resistivity data on the depth to the water table at various places are available. The potential for important new insights into the interaction of magma and ground water awaits a creative hydrogeologist who wants to venture into volcanology.

REFERENCES CITED

Anderson, D.L., and Dziewonski, A.M., 1984, Seismic tomography: Scientific American, v. 251, no. 4, p. 60–68.

Ando, M., 1979, The Hawaii earthquake of November 29, 1975: Low angle faulting due to forceful injection of magma: Journal of Geophysical Research, v. 84, p. 7616–7626.

Christiansen, R.L., 1979, Explosive eruption of Kilauea Volcano in 1790 [abs.]: Hawaii Symposium on Intraplate Volcanism and Submarine Volcanism, Hilo, Hawaii, Abstract Volume, p. 158.

Dalrymple, G.B., Silver, E.A., and Jackson, E.D., 1973, Origin of the Hawaiian Islands: American Scientist, v. 61, p. 294–308.

Daly, R.A., 1914, Igneous rocks and their origin: New York, McGraw-Hill, 563 p.

——— 1933, Igneous rocks and the depths of the Earth (2nd ed.): New York, McGraw-Hill, 598 p.

Dana, J.D., 1890, Characteristics of volcanoes: New York, Dodd Mead and Co., 399 p.

——— 1895, Manual of geology (4th ed.): New York, American Book Company, 1088 p.

Davis, P.M., Hastie, L.M., and Stacey, F.D., 1974, Stresses within an active volcano—with particular reference to Kilauea: Tectonophysics, v. 22, p. 355–362.

Davis, P.M., Jackson, D.B., Field, J., and Stacey, F.D., 1973, Kilauea Volcano, Hawaii: A search for the volcanomagnetic effect: Science, v. 180, p. 73–74.

Davis, P.M., Stacey, F.D., Zablocki, C.J., and Olson, J.V., 1979, Improved signal discrimination in tectonomagnetism: Discovery of a volcanomagnetic effect at Kilauea, Hawaii: Physics of the Earth and Planetary Interiors, v. 19, p. 331–336.

Decker, R.W., 1968, Kilauea volcanic activity: an electrical analog model [abs.]: Eos, Transactions American Geophysical Union, v. 49, p. 352–353.

Decker, R.W., and Christiansen, R.L., 1984, Explosive eruptions of Kilauea Volcano, Hawaii, *in* Explosive volcanism: Inception, evolution, and hazards: Washington, D.C., National Academy of Science, p. 122–132.

Decker, R.W., Koyanagi, R.Y., Dvorak, J.J., Lockwood, J.P., Okamura, A.T., Yamashita, K.M., and Tanigawa, W.R., 1983, Seismicity and surface deformation of Mauna Loa Volcano, Hawaii: Eos, Transactions of the American Geophysical Union, v. 64, p. 545–547.

Delaney, P.T., and Pollard, D.D., 1982, Solidification of basaltic magma during flow in a dike: American Journal of Science, v. 282, p. 856–885.

Dieterich, J.H., and Decker, R.W., 1975, Finite element modeling of surface deformation associated with volcanism: Journal of Geophysical Research, v. 80, p. 4094–4102.

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

Dvorak, J.J., Okamura, A.T., and Dieterich, J.H., 1983, Analysis of surface deformation data, Kilauea Volcano, Hawaii: October 1966 to September 1970: Journal of Geophysical Research, v. 88, p. 9295–9304.

Dzurisin, D., 1980, Influence of fortnightly earth tides at Kilauea Volcano, Hawaii: Geophysical Research Letters, v. 7, p. 925–928.

Dzurisin, D., Anderson, L.A., Eaton, G.P., Koyanagi, R.Y., Lipman, P.W., Lockwood, J.P., Okamura, R.T., Puniwai, G.S., Sako, M.K., and Yamashita, K.M., 1980, Geophysical observations of Kilauea Volcano, Hawaii, 2. Constraints on the magma supply during November 1975–September 1977: Journal of Volcanology and Geothermal Research, v. 7, p. 241–269.

Dzurisin, D., and Casadevall, T.J., 1986, Stratigraphy and chemistry of Uwekahuna Ash: Product of prehistoric phreatomagmatic eruption at Kilauea Volcano, Hawaii [abs.]: Auckland, N.Z., Abstracts volume, International Volcanological Congress, p. 100.

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

Eaton, J.P., 1962, Crustal structure and volcanism in Hawaii: American Geophysical Union, Geophysical Monograph 6, p. 13–29.

Eaton, J.P., and Murata, K.J., 1960, How volcanoes grow: Science, v. 132, p. 925–938.

Ellis, W., 1827, Journal of William Ellis: London, reprinted 1963 by Advertiser Publishing Co., Honolulu, Hawaii, 342 p.

Epp, D., Decker, R.W., and Okamura, R.T. 1983, Relation of summit deformation to east rift zone eruptions on Kilauea Volcano, Hawaii: Geophysical Research Letters, v. 10, p. 493–496.

Fiske, R.S., and Jackson, E.D., 1972, Orientation and growth of Hawaiian volcanic rifts: The effect of regional structure and gravitational stresses: Proceedings of the Royal Society of London, Ser. A, v. 329, p. 299–326.

Fiske, R.S., and Kinoshita, W.T., 1969, Inflation of Kilauea Volcano prior to its 1967–1968 eruption: Science, v. 165, p. 341–349.

Furumoto, A.S., and Kovach, R.L., 1979, The Kalapana earthquake of November 29, 1975: An intra-plate earthquake and its relation to geothermal processes: Physics of the Earth and Planetary Interiors, v. 18, p. 197–208.

Gerlach, T.M., and Graeber, E.J., 1985, Volatile budget of Kilauea Volcano: Nature, v. 313, p. 273–277.

Griggs, D.T., and Baker, D.W., 1969, The origin of deep-focus earthquakes, *in* Mark, H., and Fernback, S., eds., Properties of matter under unusual conditions: New York, Interscience Publishers, p. 23–42.

Hardee, H.C., 1982, Incipient magma chamber formation as a result of repetitive intrusions: Bulletin Volcanologique v. 45, p. 41–49.

Hess, H.H., 1960, Stillwater igneous complex, Montana: Geological Society of America Memoir 80, 230 p.

Hill, D.P., 1977, A model for earthquake swarms: Journal of Geophysical Research, v. 82, p. 1347–1352.

Holcomb, R.T., 1980, Preliminary geologic map of Kilauea Volcano, Hawaii: U.S. Geological Survey Open-File Report 80–796, 2 sheets, scale 1:50,000.

Holcomb, R.T., Peterson, D.W., and Tilling, R.I., 1974, Recent landforms at Kilauea Volcano, a selected photographic compilation, *in* Greeley, R., ed., Hawaiian Planetology Conference, NASA TMX 62362, p. 49–86.

Jachens R.C., and Eaton, G.P., 1980, Geophysical observations of Kilauea Volcano, Hawaii, 1. Temporal gravity variations related to the 29 November, 1975, M=7.2 earthquake and associated summit collapse: Journal of Volcanology and Geothermal Research, v. 7, p. 225–240.

Jackson, D.B., Kauahikaua, J., and Zablocki, C.J., 1983, Resistivity monitoring of an active volcano: Application of the controlled-source electromagnetic technique at Kilauea Volcano, Hawaii [abs.]: Eos, Transactions of the American Geophysical Union, v. 64, p. 1072.

Jaggar, T.A., and Finch, R.H., 1924, The explosive eruption of Kilauea in Hawaii: American Journal of Science, Ser. 5, v. 8, p. 353–374.

Joly, J., 1909, Radioactivity and geology: London, Archibald Constable and Co., 287 p.

Keller, G.V., Grose, L.T., Murray, J.C., and Skokan, C.K., 1979, Results of an experimental drill hole at the summit of Kilauea Volcano, Hawaii: Journal of Volcanology and Geothermal Research, v. 5, p. 345–385.

Kinoshita, W.T., 1965, A gravity survey of the Island of Hawaii: Pacific Science, v. 19, p. 339–340.

Kinoshita, W.T., Koyanagi, R.Y., Wright, T.L., and Fiske, R.S., 1969, Kilauea Volcano: The 1967–68 summit eruption: Science, v. 166, p. 459–468.

Klein, F.W., 1982a, Earthquakes at Loihi submarine volcano and the Hawaiian hot spot: Journal of Geophysical Research, v. 87, p. 7719–7726.

——— 1982b, Patterns of historical eruptions at Hawaiian volcanoes: Journal of Volcanology and Geothermal Research, v. 12, p. 1–35.

Koyanagi, R.Y., Unger, J.D., Endo, E.T., and Okamura, A.T., 1976, Shallow earthquakes associated with inflation episodes at the summit of Kilauea Volcano, Hawaii: Bulletin Volcanologique, v. 39, p. 621–631.

Lipman, P.W., Lockwood, J.P., Okamura, R.T., Swanson, D.A., and Yamashita, K. M., 1985, Ground deformation associated with the 1975 magnitude-7.2 earthquake and resulting changes in activity of Kilauea Volcano, Hawaii: U.S. Geological Survey Professional Paper 1276, 45 p.

Liu, H., and Kosloff, D. 1978, Elastic-plastic bending of the lithosphere incorporating rock deformation data, with application to the structure of the Hawaiian Archipelago: Tectonophysics, v. 50, p. 249–274.

Macdonald, G.A., 1972, Volcanoes: Englewood Cliffs, N. J., Prentice-Hall, 510 p.

Malahoff, A., McMurtry, G.M., Wiltshire, J.M., and Yeh, H., 1982, Geology and chemistry of hydrothermal deposits from Loihi, the latest active submarine volcano of the Hawaiian hot spot: Nature, v. 298, p. 234–239.

McBirney, A.R., 1963, Conductivity variations and terrestrial heat-flow distribution: Journal of Geophysical Research, v. 68, p. 6323–6329.

McDougall, I., 1979, Age of shield-building volcanism of Kauai and linear migration of volcanism in the Hawaiian Island chain: Earth and Planetary Science Letters, v. 46, p. 31–42.

Mogi, K., 1958, Relation between the eruptions of various volcanoes and the deformation of the ground surface around them: Tokyo, Bulletin of the Earthquake Research Institute, v. 36, p. 99–134.

Moore, J.G., 1970, Relationship between subsidence and volcanic load, Hawaii: Bulletin Volcanologique, v. 34, p. 562–576.

——— 1971, Bathymetry and geology—East Cape of the Island of Hawaii: U.S. Geological Survey Miscellaneous Geological Investigations Map I–677, Scale 1:62,500.

Moore, J.G., and Fornari, D.J., 1984, Drowned reefs as indicators of the rate of subsidence of the Island of Hawaii: Journal of Geology, v. 92, p. 752–759.

Moore, J.G., and Krivoy, H.L., 1964, The 1962 flank eruption of Kilauea Volcano and structure of the east rift zone: Journal of Geophysical Research, v. 69, p. 2033–2045.

Moore, J.G., and Moore, G.W., 1984, Deposit from a giant wave on the Island of Lanai, Hawaii: Science, v. 226, p. 1312–1315.

Peterson, D.W., Christiansen, R.L., Duffield, W.A., Holcomb, R.T., and Tilling, R.I., 1976, Recent activity of Kilauea Volcano, Hawaii, *in* International Association for Volcanology and Chemistry of Earth's Interior, Symposium on Andean and Antarctic Volcanology Problems, Santiago, Chile, Proceedings, p. 646–656.

Peterson, D.W., and Swanson, D.A., 1974, Observed formation of lava tubes during 1970–71 at Kilauea Volcano, Hawaii: Studies in Speleology, v. 2, p. 209–233.

Peterson, D.W., and Tilling, R.I., 1980, Transition of basaltic lava from pahoehoe to aa, Kilauea Volcano, Hawaii: Field observations and key factors: Journal of Volcanology and Geothermal Research, v. 7, p. 271–293.

Pollard, D.D., 1976, On the form and stability of open fractures in the Earth's crust: Geophysical Research Letters, v. 3, p. 513–516.

Ramberg, H., 1972, Mantle diapirism and its tectonic and magmagenetic consequences: Physics of Earth and Planetary Interiors, v. 5, p. 45–60.

Ryan, M.P., Blevins, J.Y.K., Okamura, A.T., and Koyanagi, R.Y., 1983, Magma reservoir subsidence mechanics: Theoretical summary and application to Kilauea Volcano, Hawaii: Journal of Geophysical Research, v. 88, p. 4147–4181.

Ryan, M.P., Koyanagi, R.Y., and Fiske, R.S., 1981, Modeling the three-dimensional structure of macroscopic magma transport systems: Application to Kilauea Volcano, Hawaii: Journal of Geophysical Research, v. 86, p. 7111–7129.

Scientific Event Alert Network, 1982–1985, Monthly accounts of the activity of Kilauea and Mauna Loa Volcanoes, Hawaii: SEAN Bulletin v. 8-10, no. 1-12, Washington, D.C., Smithsonian Institution.

Shaw, H.R., 1969, Rheology of basalt in the melting range: Journal of Petrology, v. 10, p. 510–535.

——— 1970, Earth tides, global heat flow, and tectonics: Science, v. 168, p. 1084–1087.

——— 1973, Mantle convection and volcanic periodicity in the Pacific; evidence for Hawaii: Geological Society of America Bulletin, v. 84, p. 1505–1526.

——— 1980, The fracture mechanisms of magma transport from the mantle to the surface, *in* Hargraves, R.B., ed., Physics of Magmatic Processes: Princeton, New Jersey, Princeton University Press, p. 201–264.

Shimozuru, D., 1981, Magma reservoir systems inferred from tilt patterns: Bulletin Volcanologique, v. 44-3, p. 499–504.

Spera, F.J., 1980, Aspects of magma transport, *in* Hargraves, R.B., ed., Physics of Magmatic Processes: Princeton, New Jersey, Princeton University Press, p. 265–323.

Stearns, H.T., 1978, Quaternary shorelines in the Hawaiian Islands: Bernice P. Bishop Museum, Bulletin 237, Honolulu, 57 p.

Swanson, D.A., 1972, Magma supply rate at Kilauea Volcano, 1951–1972: Science, v. 175, p. 169–170.

——— 1973, Pahoehoe flows from the 1969–1971 Mauna Ulu eruption, Kilauea Volcano, Hawaii: Geological Society of America Bulletin, v. 84, p. 615–626.

Swanson, D.A., and Christiansen, R.L., 1973, Tragic base surge in 1790 at

Kilauea Volcano: Geology, v. 1, p. 83–86.

Swanson, D.A., Duffield, W.A., and Fiske, R.S., 1976a, Displacement of the south flank of Kilauea Volcano: The result of forceful intrusion of magma into rift zones: U.S. Geological Survey Professional Paper 963, 39 p.

Swanson, D.A., Duffield, W.A., Jackson, D.B., and Peterson, D.W., 1979, Chronological narrative of the 1969–71 Mauna Ulu eruption of Kilauea Volcano, Hawaii: U.S. Geological Survey Professional Paper 1056, 55 p.

Swanson, D.A., Jackson, D.B., Koyanagi, R.Y., and Wright, T.L., 1976b, The February 1969 east rift eruption of Kilauea Volcano, Hawaii: U.S. Geological Survey Professional Paper 891, 30 p.

Thorarinsson, S., 1965, Surtsey: Island born of fire: National Geographic Magazine, v. 127, p. 713–726.

Tilling, R.I., Koyanagi, R.Y., Lipman, P.W., Lockwood, J.P., Moore, J.G., and Swanson, D.A., 1976, Earthquake and related catastrophic events, Island of Hawaii, November 29, 1975: A preliminary report: U.S. Geological Survey Circular 740, 33 p.

Turcotte, D.L., and Oxburgh, E.R., 1978, Intraplate volcanism: Philosophical Transactions Royal Society of London, Ser. A., v. 288, p. 561–579.

Verhoogen, J., 1954, Petrological evidence on temperature distribution in the mantle of the earth: Eos, Transactions of the American Geophysical Union, v. 35, p. 85–92.

Weertman, J., 1972, Coalescence of magma pockets into large pools in the upper mantle: Geological Society of America Bulletin, v. 83, p. 3531–3632.

Wilkes, C., 1845, Narrative of the United States exploring expedition during the years 1838–1842: Philadelphia, v. 4, p. 87–231.

Wilson, J.T., 1963, Continental drift: Scientific American, v. 208, p. 86–100.

Wilson, R.M., 1935, Ground surface movement at Kilauea Volcano, Hawaii: University of Hawaii, Honolulu, Research Publication 10, 56 p.

Wright, T.L., 1971, Chemistry of Kilauea and Mauna Loa lava in space and time: U.S. Geological Survey Professional Paper 735, 40 p.

——— 1984, Origin of Hawaiian tholeiite: a metasomatic model: Journal of Geophysical Research, v. 89, p. 3233–3252.

Wright, T.L., Shaw, H.R., Tilling, R.I., and Fiske, R.S., 1979, Origin of Hawaiian tholeiitic basalt: a quantitative model [abs.]: International Association for Volcanology and Chemistry of the Earth's Interior, Symposium on intraplate volcanism and submarine volcanism, Hilo, Hawaii, July 16–22, 1979, Abstracts Volume, p.

Yoder, H.S., 1976, Generation of basaltic magma: Washington, D.C., National Academy of Sciences, 265 p.

Yoder, H.S., and Tilley, C.E., 1962, Origin of basalt magmas: an experimental study of natural and synthetic rock systems: Journal of Petrology, v. 3, p. 342–532.

Yoshii, T., Kono, T., and Ito, K., 1976, Thickening of the oceanic lithosphere: American Geophysical Union, Monograph 19, p. 423–430.

Zablocki, C.J., 1976, Mapping thermal anomalies on an active volcano by the self-potential method, Kilauea, Hawaii: United Nations Symposium on the development and use of geothermal resources, 2nd San Francisco, Proceedings, v. 2, p. 1299–1309.