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REVIEW ARTICLE

# Geology and Volcanology of the Hawaiian Islands[1]

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ABSTRACT: This article summarizes the present state of knowledge of the geology of Hawaii. It tends to concentrate on aspects not usually covered by review articles. Current ideas on hotspots and mantle plumes are applied to the specific example of Hawaii, the eight volcanic systems currently nourished by the hotspot are identified, and gross differences in magma-supply rate are related to position of these systems on the hotspot. The important role played by level of neutral buoyancy in distributing incoming magma between magma chambers, rift zones, intrusions, and surface flows is discussed. This is important because volcanic edifices may expand nearly as much by growth of subsurface intrusions as by surface lava outpourings. Recent discoveries, however, show how strongly volcano growth is countered by subsidence and major collapses. A brief description is given of styles of volcanism in Hawaii, and recent ideas on how formation of aa and pahoehoe depends on eruption discharge rate are discussed. A brief summary description pointing to highlights of each volcano is then presented. Finally, I indulge in speculations regarding geographical distribution of the volcanoes and show how, by postulating that a considerable strike-slip motion has occurred on two faults, a much more orderly arrangement of volcano and rift-zone alignments appears, leading to a dynamic model of island-chain growth that is simpler then current models. Proceeding from Kaua'i toward the southeast, an alternating sequence of southeast and west-southwest alignments is revealed. These alignments may be related, respectively, to fractures propagated against the plate motion direction (because of extensional stresses resulting from diverging flow in the mantle plume) and along faults of the Moloka'i fracture zone.

THIS IS IN PART A REVIEW article and synthesis of the present state of knowledge of the geology of the Hawaiian Islands. It is in part also a personal view and includes some speculative new ideas on the distribution and orientation of the Hawaiian volcanoes; inclusion of these ideas is intended to emphasize the fact that concepts must continually change, new ideas must be tested, and even some of the most time-hallowed concepts are capable of alternative interpretations. This article is not concerned with the history of development of ideas, and most of the references accordingly are to the most recently published works.

The past decade has seen a massive increase in the research effort that is directed at the geology of the Hawaiian Islands. Currently, some 30 research papers are being published each year, new editions have appeared of two popular texts (Macdonald et al. 1983, Stearns 1985), a comprehensive reference work on source materials has been compiled (Wright and Takahashi 1989), the entire collection of Volcano Letter has been reprinted (Fiske et al. 1987), two massive volumes have been published on volcanism in Hawaii (Decker et al. 1987), a new map of Mauna Loa has been compiled (Lockwood et al. 1988), and a new geologic map of the island of Hawai'i is nearing completion.

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*The Hawaiian Hotspot and Mantle Plume*

The Hawaiian Islands are the tops of great volcanoes, built from the accumulations of countless volcanic eruptions at the Hawaiian hotspot. A hotspot is a place where voluminous mid-plate volcanism occurs or where anomalously voluminous volcanism occurs on a spreading ridge (Burke and Wilson 1976). The Hawaiian hotspot is of the first type and occurs near the middle of the Pacific Plate. It vies with the Icelandic hotspot as being the most productive on earth.

Volcanism at a hotspot is typically sustained for tens of millions of years, and as the lithosphere moves over the hotspot a chain of volcanoes (a hotspot trace) results (Wilson 1963). The Hawaiian-Emperor Chain extends for 6000 km across the North Pacific to the edge of the Asian continent and is the finest hotspot trace known. It includes over 100 major volcanoes (most of them seamounts) and shows a general progression in age from the 70-million-yr-old Meiji Seamount off Kamchatka to the still actively growing island of Hawai‘i (Clague and Dalrymple 1987, 1989). It is supposed that still older parts of the chain were subducted under the Asian continent. The Pacific Plate is moving over the hotspot at 9 cm/yr. A change in plate motion about 43 million yr ago (Ma) possibly related to the separation of Australia from Antarctica produced the dogleg west of Midway.

It is widely thought that a hotspot is a manifestation of a mantle plume (Morgan 1972, White and McKenzie 1989). A mantle plume is a pipelike current of mantle material that ascends because it is less dense than the surrounding mantle (because it is hotter, different in composition, or both). Opinions differ on whether mantle plumes originate at the core/mantle boundary or within the mantle. Where a plume reaches the base of the lithosphere it spreads laterally and forms a mushroom-shaped head that may be from several hundred to as much as 2000 km wide (Figure 1).

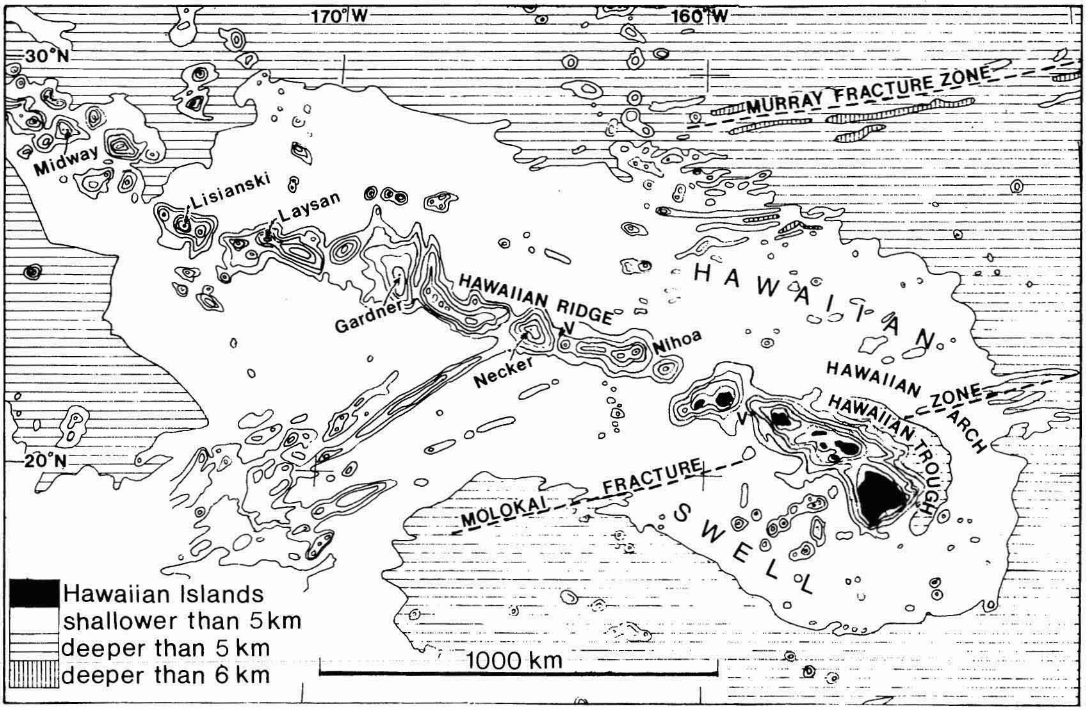
The arrival of a plume at the base of the lithosphere may split a continent and result in a voluminous outpouring of flood basalts; the subsequent ascent of the plume tail sustains volcanism at a lesser intensity and so constructs the hotspot trace. Subsequent plate motion may carry the flood basalts far away from the hotspot, although they remain connected to it by the hotspot trace. Thus, the Deccan is connected with the Réunion hotspot (Duncan 1981). No flood basalts are known associated with the Hawaiian plume: possibly any such basalts were subducted below Asia. Plate reconstructions (Rea and Duncan 1986) suggest that about 100 Ma the Hawaiian hotspot was situated at a triple junction (at the join of three spreading ridges). Spreading was possibly initiated by the arrival of the plume from below.

The size and position of the Hawaiian plume may be indicated by the Hawaiian Swell, a broad elongate positive bathymetric feature 1200 km wide on the axis of which the Hawaiian Ridge stands (Crough 1983). The oceanic crust below Hawaii was formed in Cretaceous times, at 80–90 Ma. The depth of the ocean is closely related to the age of the underlying crust (because the lithosphere cools and subsides as it moves away from a spreading ridge). The Hawaiian Swell is an area in which the ocean is up to 1.2 km shallower than is expected for its age. The Swell can be attributed to the upbowing of oceanic crust over the anomalously low-density material of the mantle plume.

*Generation of Magma in the Plume*

Beneath Hawaii the boundary between the lithosphere (the outer, rigid shell of the earth) and the underlying asthenosphere (a yielding layer) is about 60 km deep. The deepest Hawaiian earthquakes have their focus at about this depth (Klein et al. 1987, Thurber 1987), and this is where Hawaiian magmas are generated. The Mohorovicic Discontinuity, or Moho, which is the boundary between the basaltic crust and the peridotite mantle, is about 12 to 18 km deep beneath the islands and is shallower offshore (Hill and Zucca 1987).

Practically all of the earth's mantle is solid and sufficiently rigid to transmit earthquake waves. Its temperature increases downward, but the melting temperature also increases



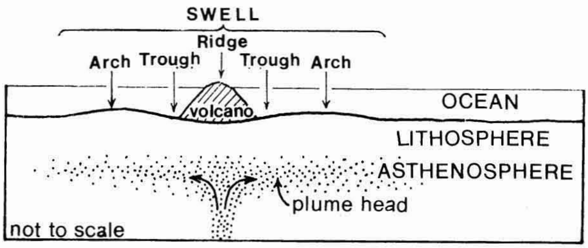




FIGURE 1. *Above*: Bathymetric map of the Hawaiian Swell, marked approximately by water shallower than 5 km. This swell may coincide roughly with the extent of the hotspot and mantle plume-head. V, sites of possible submarine eruptions in 1955 and 1956. Bathymetric contours at 1-km intervals. *Below*: Southwest-northeast profile across the Hawaiian Swell, showing (stippled) the inferred anomalously hot mantle plume-head. Vertical exaggeration × 100.

downward as the pressure increases (Oxburgh 1980). The upper mantle consists of crystalline ultramafic rocks exemplified by the lherzolite, pyroxenite, and garnet pyroxenite inclusions that are brought to the surface by many volcanoes in Hawaii and elsewhere (Jackson and Wright 1970, Sen 1988). If the plume material is sufficiently hot—perhaps 100° to 300° C hotter than the surrounding mantle—then partial melting may occur (Wyllie 1988). If the immediately overlying lithosphere is stretched and becomes thinner, then the plume material rises, the pressure on it decreases, and its melting temperature is lowered, so favoring partial melting; no evidence for this thinning is known below Hawaii.

Consider now what happens when part of the mantle melts. Laboratory experiments show that silicate rocks in general do not have a fixed melting point but melt over a range of temperatures. When peridotite is heated, melting begins at crystal boundaries as the "solidus" temperature is passed. The proportion of rock that melts increases as the temperature rises, and the peridotite becomes totally molten only when the "liquidus" temperature is reached. For many silicate rocks the liquidus temperature is about 200° C hotter than the solidus temperature. Total melting of peridotite will generate peridotite magma, but partial melting of peridotite occurs at a much lower temperature and generates basaltic magma.

The basaltic magma that is generated just above the solidus temperature with a very small degree (a few percent) of partial melting is undersaturated in silica and highly alkalic, and in it are concentrated all the least refractory constituents of the peridotite (the "incompatible" elements) such as alkalies and volatiles (Clague and Frey 1982). At a still higher temperature with perhaps about 30% of partial melting a tholeiitic basalt magma results. At high degrees of partial melting the magma is picritic (magnesium-rich).

Partial melts are generated in a thin film at crystal boundaries in the melting peridotite, and calculations show that this film is capable of segregating when it constitutes even as little as a few percent of the mantle rock (Yoder 1988). Then, when enough magma has segregated, it may ascend to a high crustal level or erupt at the surface. Ascent occurs because the basaltic melt has a lower density than rocks in the mantle and much of the crust. Evidently magma is able to penetrate the lithosphere when about 0.1 to 1.0 $km^3$ of it has segregated out.

The magma that actually erupts from a Hawaiian volcano may be considerably different from that which leaves the asthenosphere. This is because some form of fractionation occurs in transit. The first-formed crystals that grow in a cooling magma are enriched in the more refractory elements (e.g., olivine crystals are enriched in magnesium) compared with the magma. If the crystals are then removed (e.g., by settling out), the magma that is left has a changed composition because it is then impoverished in these more refractory elements. Fractionation is favored if a magma resides for a long time in a magma chamber. Geochemical tests can discriminate between magma that fractionated deep in the crust and magma that fractionated at a shallow level.

Some tholeiitic lava flows, called picrite basalts, are conspicuously rich in olivine crystals. Such magnesium-rich melts have a very low viscosity, and olivine that readily crystallizes in them is much denser than the melt; the crystals therefore rapidly settle out and are often found concentrated in the lower parts even of thin lava flows (Rowland and Walker 1988). Given the ease with which this settling takes place in surface flows, considerable separation can be expected in a magma chamber.

This tendency of settle can make it difficult to determine the origin of picrite basalts. Picrite basalts may represent the direct crystallization of high-Mg melts or may be the result of accumulation of olivine crystals into less Mg-rich melts. Similarly, some of the associated Mg-poor basalts may be primary melts, and others may result from the loss of olivine from a more picritic melt. Geochemical tests of these alternative mechanisms exist and are discussed by Wilkinson and Hensel (1988) and Nicholls and Stout (1988).

Picrite basalts appear to be more abundant at lower elevations on Hawaiian volcanoes, good examples being the 1840 flow that erupted low on Kīlauea's east rift zone and the 1868 flow that erupted low on Mauna Loa's southwest rift zone. This greater abundance may result from discharge rates being generally higher at lower elevations, giving the dense crystals less time to separate out. The erupting lava at Kīlauea Iki in 1959 was much more magnesium-rich when the discharge rate was high (Murata and Richter 1966).

### *Volcanic Systems Sustained by the Hawaiian Hotspot*

When considering the volcanism of the Hawaiian Islands, it is convenient to adopt the concept of volcanic systems (Saemundsson 1986). A system consists of a volcano together with its complex of co-magmatic intrusions, magma chambers, rift zones, and magma-supply conduits. A system may comprise a single shield volcano such as Mauna Loa together with its well-developed roots and rift zones, or it may comprise a field of monogenetic volcanoes (volcanoes that erupt once, and once only) such as the cluster of relatively young "rejuvenation-stage" Honolulu Volcanics (exemplified by Diamond Head) on O'ahu.

The Hawaiian hotspot currently sustains at least eight volcanic systems (Figure 2). Three, Kīlauea, Mauna Loa, and Lō'ihi, are in their tholeiitic shield-building stage (Figure 3) in which the volcanoes are most productive, and it is inferred that these systems derive their magmas from the central and hottest part of the plume (above the ascending limb) where the degree of partial melting is highest. Two, Hualālai and Mauna Kea, are in their alkalic-cap stage and have entered a period of decline that may last as long as a million years as they gradually move off the center of the hotspot. At least two, Haleakalā and the Honolulu Volcanics, are in the rejuvenation stage of volcanism, which may lag by as much as 5 million yr behind the main shield-building stage (Figure 4).

Hawaiian volcanoes at the peak of their activity attain an output unsurpassed by any other volcano on earth and probably approached or matched only by volcanoes in Iceland, Réunion, and the Galápagos. The amount of heat carried by the magma is ample to maintain hot pathways from the magma source to the subsummit chambers of the volcanoes (Figure 3) and thence into parts of the rift zones.

At the very peak of activity magma may flow uninterruptedly for long periods from the mantle source to the surface through hot and stable conduits, residing for only a short time, if at all, in the subsummit magma chamber. Such continuous activity generates extensive pahoehoe flows. Kīlauea had such activity from 1919 to 1924, 1970 to 1972, 1973 to 1974, and 1986 to the present and probably for much longer periods in prehistoric time.

The subsequent decline in magma generation rate and delivery rate to the volcano changes the thermal condition of the volcano roots; the amount of heat carried in becomes insufficient to maintain the hot pathway from the magma source through the subsummit chamber and thence into rift zones. The chamber solidifies, caldera subsidence no longer occurs, and eventually each magma batch must create its own independent path all the way to the surface with little guidance either from the original central conduit or the rift zones. The vents then become widely scattered monogenetic structures such as the cones on Mauna Kea (Figure 3).

The eruption frequency decreases to one per several centuries or millennia, giving time for soils to develop or erosion to occur between eruptions. The eruptions tend to increase in magnitude and power, in part because of the higher content of magmatic gas. They build large cinder cones and erupt large-volume (mostly aa) lava flows. Many magma batches bring to the surface inclusions from the mantle or from intrusions in the volcano core because the magma does not reside in a magma chamber where these very dense objects can settle out (Clague 1987b). Such inclusions are locally abundant in the alkalic lavas of Hualālai, Mauna Kea, and the rejuvenation-stage Honolulu Volcanics of O'ahu.

It is inferred that the rejuvenation-stage magmas come from near the edge of the plume head where the degree of partial melting is low. The eruption rate in this stage has decreased to one per several tens of millennia, and the lavas typically rest on strongly eroded and weathered rocks; hence the common designation, "posterosional" volcanics.

### *Styles of Volcanic Activity*

HAWAIIAN STYLE. In this, the most common style of volcanism in Hawaii, lava spurts out of a fissure in so-called fire fountains in which the foamy lava is torn apart into tatters. The fountain height varies from under 5 m to more then 500 m and is directly related to the volumetric discharge rate and magmatic gas content (Head and Wilson 1989). If the discharge rate is high, most of the ejected lava fragments remain inside the optically thick central part of the fountain where they undergo minimal cooling. They are hence still hot when they land, and most of them coalesce and flow away.

Generally only an insignificant proportion of the total erupted volume fails to flow away, forming pyroclastic deposits instead. The lava tatters that are still plastic when they land flatten against the ground to form spatter and

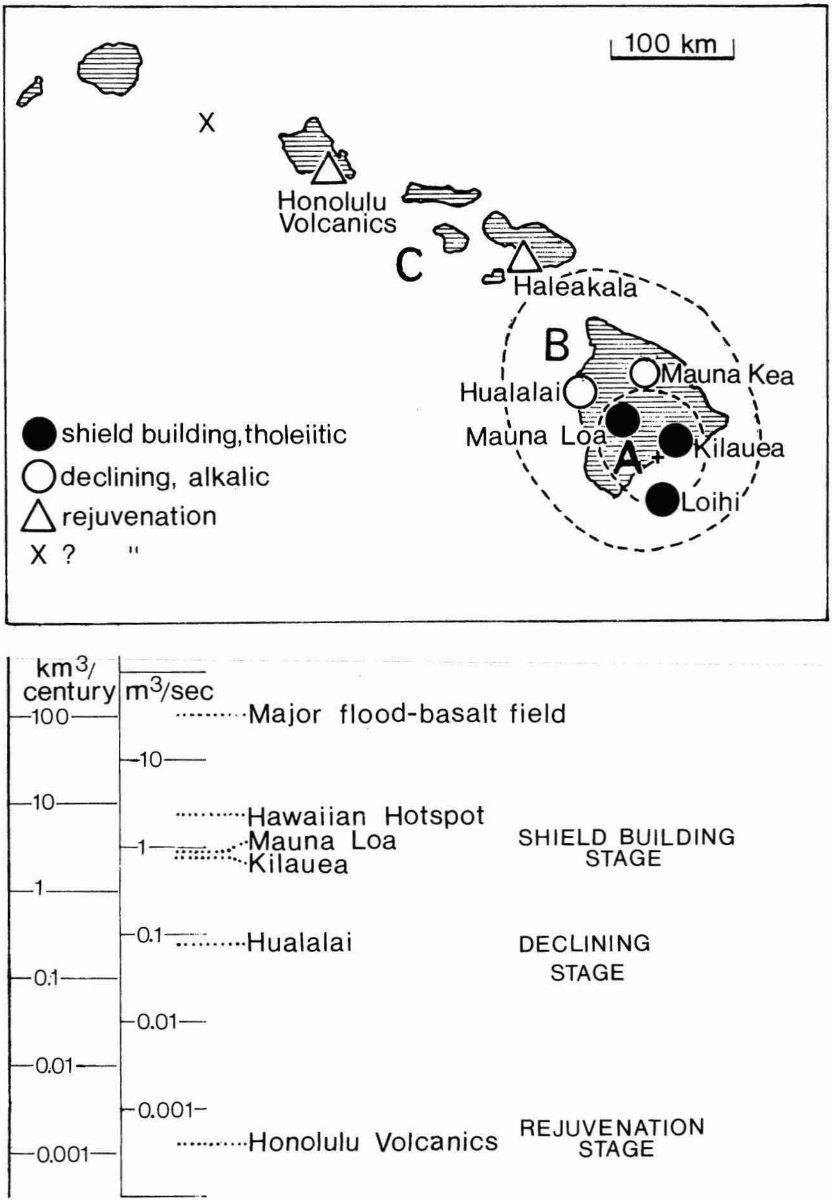



FIGURE 2. *Above*: Volcanic systems currently sustained by the Hawaiian hotspot. Zone A, tholeiitic shield-building stage, inferred to overlie the ascending limb of the mantle plume; +, a narrow zone of earthquakes >30 km deep interpreted to mark the magma-supply conduit from the mantle. Zone B, declining (alkalic cap) stage: the magma-supply rate and eruption frequency have drastically declined. Zone C, rejuvenation stage, representing a further decline; X, site of possible submarine eruption in 1956 (Macdonald 1959); a better documented eruption occurred 700 km farther west-northwest off Necker Island (Figure 1) in 1955. Some of the lava flows recorded by Holcomb et al. (1988) from the ocean floor around Hawaii may relate to other volcanic systems. *Below*: Estimated time-averaged output rate of Hawaiian volcanoes in different stages: Mauna Loa and Kīlauea for historical period (Lockwood and Lipman 1987); Hualālai for past 3000 yr (Moore et al. 1987).

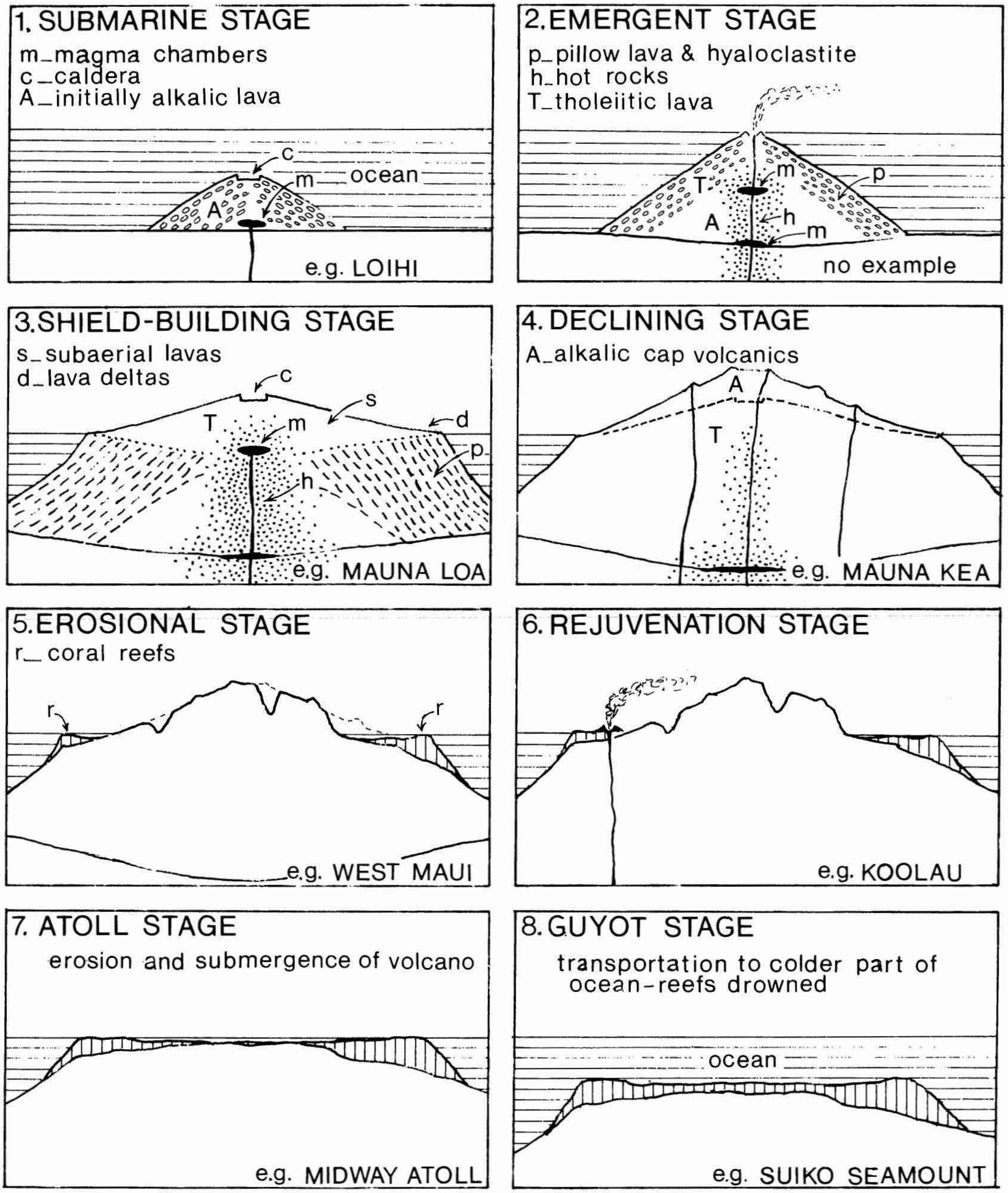



FIGURE 3. Stages in the life of a typical major volcano in the Hawaiian-Emperor Chain, modified after Stearns (1946), Macdonald et al. (1983), and Peterson and Moore (1987). Stage 4 is variously referred to as the alkalic-cap or postshield stage.

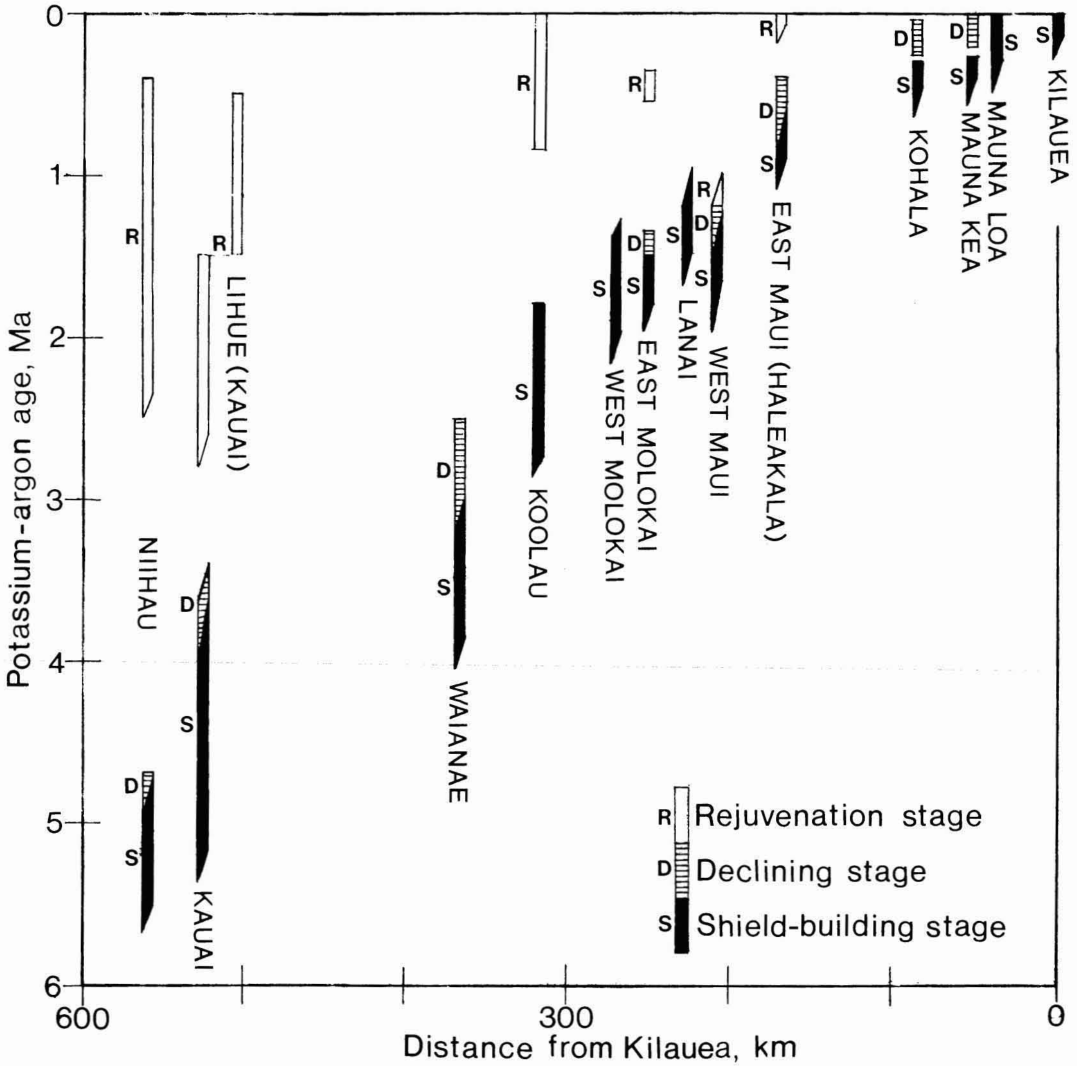



FIGURE 4. Potassium-argon ages of rocks from the Hawaiian Islands in millions of years against distance of the volcano centers from Kīlauea, from data referenced and collated by Clague and Dalrymple (1987, 1989). Note that the dated rocks were collected from the island tips of largely submerged volcanic edifices and do not give the time of inception of each edifice.

weld to one another to form agglutinate. They construct spatter ramparts usually less than 10 m high along the lengths of eruptive fissures and spatter rings or cones where eruptions are concentrated at a point. Good examples are seen on the edge of Keanakāko'i and Puhimau Craters on Kīlauea. The lava tatters that are solid on landing form cinders and build cinder cones. These cones are usually only tens of meters high. Pu'u O'o, a cone of pyroclastic material interstratified with thin lava flows that built up to a height of 280 m during 47 fire-fountaining episodes from January 1983 to July 1986, is exceptional.

Lava fragments in the margins of the fountains cool more rapidly, and those that are blown off the top of the fountain by the wind may travel several kilometers before falling.

These fragments consist mostly of lightweight foamy pumice. Reticulite, an incredibly fragile and lightweight lava foam in which millimeter-size bubbles occupy about 98% of the total space, is also produced in minor quantity in the fire fountains, and so are Pele's tears (small lava droplets) and Pele's hair (lava stretched into fibers: natural fiberglass).

STROMBOLIAN-STYLE activity results when the lava has a higher gas content and a somewhat higher viscosity; the resulting eruptive column tends to be higher (although the height depends partly on the discharge rate), and the lava is more highly fragmented. Most lava fragments cool significantly before they land and form loose accumulations of scoria or cinders. Large cinder cones accumulate, together with areally extensive cinder deposits around or downwind of them. Cinder cones in Hawaii are commonly 50 to 200 m high and have a crater that is commonly 100 to 400 m wide. They particularly characterize the declining and rejuvenation-stage eruptions of Hawaiian volcanoes. Fine examples form the summit area of Mauna Kea (Porter 1972b) and occur in the saddle between Mauna Kea and Mauna Loa, and numerous cones occur in Haleakalā Crater on Maui.

The highly distinctive spindle-shaped volcanic bombs are common in the strombolian-type cinder cones (e.g., the alkalic-cap cones on Mauna Kea), but are absent from the Hawaiian-style tholeiitic vents. The spindle bombs typically have a higher density than the associated cinders and apparently represent lava fractions that resided for longer than average in the vent (giving time for gas loss and viscosity increase) before being thrown out. The elongate shape results from stretching when one end of an ejected lava "rope" travels through the air faster than the other end.

SURTSEYAN-STYLE ERUPTIONS occur at the coast or in the shallow ocean, where copious amounts of water can enter the vent and mingle with the ascending lava. The eruption style is dominated by a great ascending steam cloud, and the lava is fragmented to a sandy and glassy ash that accumulates at the vent and builds an ash ring. Eruption-column collapses that generate outward-moving base surges occur repeatedly and give rise to dune-bedded deposits such as those conspicuously displayed in the Koko fissure (Fisher 1977) and Salt Lake tuff rings on O'ahu.

Chemical changes then hydrate the glassy ash to palagonite and cause the crystallization of such minerals as calcite and zeolites in the pore spaces; within a few years they convert the ash into hard tuff and the ash ring into a tuff ring (Jakobsson 1978). Examples of tuff rings are Molokini (between Maui and Kaho'olawe), Diamond Head, Hanauma Bay, and Punchbowl (O'ahu), Kīlauea Head (Kaua'i), Lehua (Ni'ihau), and Kapoho Crater (Kīlauea). Explosions sometimes occur where a lava flow enters the ocean from land and build small littoral cones (Moore and Ault 1965, Fisher 1968). The black sand beach of Kalapana and the new black sand beach generated since 1986 west of Kalapana consist of glassy ash formed this way.

PHREATIC EXPLOSIONS of great violence occasionally occur in the summit area of Kīlauea volcano and are attributed to steam explosions when drainback of lava occurs from the upper part of the magma conduit system (Decker and Christiansen 1984). This enables hot water that was trapped alongside the conduits to escape, flashing into steam as it does so. It also enables groundwater to enter the vacated conduit system and be rapidly converted to steam in contact with hot rocks. Collapse of the debris-laden eruptive column may generate dangerous base surges.

### *Pahoehoe and Aa Lava Flows*

Lava flows are formed in most eruptions of Hawaiian volcanoes. Two common structural types occur, named pahoehoe and aa. Pahoehoe has a smooth and commonly wrinkled ("ropy") surface, and the abundant gas bubbles (vesicles) in it tend to be spherical in shape. Aa has a rough surface that consists of an untidy assemblage of more or less loose clinkery and rubbly fragments, and this surface layer is underlain by massive lava in which the vesicles are commonly scarce and have strongly deformed shapes. Pahoehoe lava flows are compound, composed of a great

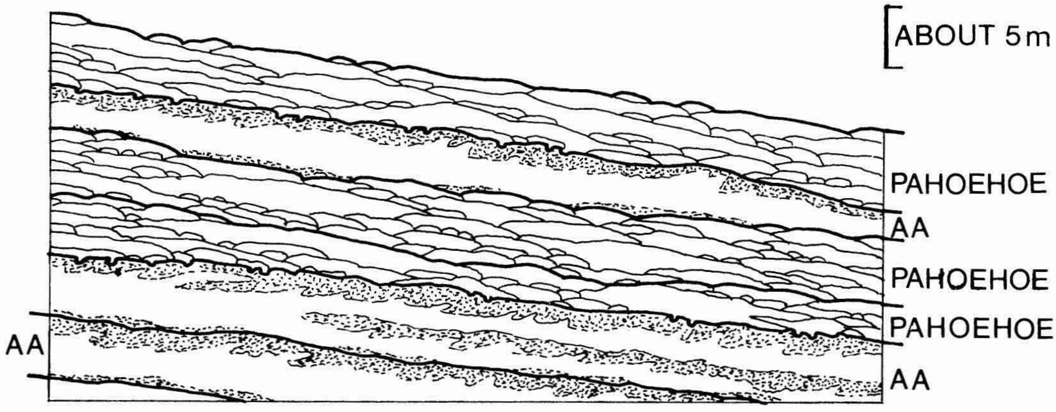



FIGURE 5. Typical lava structure on the flank of a Hawaiian shield. Six lava flows are shown; three are pahoehoe, subdivided into numerous flow-units, and three are aa in which massive lava is sandwiched between rubble layers (stippled).

number of separate flow units, each of which is enclosed in a chilled crust having very small vesicles. The smaller flow units measure under 1 m in size and are called toes. Aa flows are less consistently compound in character than pahoehoe (Figure 5).

Recent studies (Peterson and Tilling 1980, Rowland and Walker 1990) relate the difference between aa and pahoehoe to the different flow dynamics of the actively moving lava. Aa in Hawaii typically forms when lava erupts at a high volumetric discharge rate (exceeding 5 or 10 $m^3$/sec) and flows away from the vent as one or several powerful lava rivers in open channels. Pahoehoe typically forms when lava erupts at a low volumetric rate and the discharge is dispersed into a multitude of small and mostly slow-flowing lava lobes; the dispersal takes place either around the vent or after lava has traveled through and emerged from tubes. Aa continues to flow until, with cooling, the lava viscosity is too high to repair torn crust. In contrast, pahoehoe lobes are so small that they become static while the viscosity is still low enough to repair torn crust.

*Submarine Volcanics*

When lava erupts under water or flows from land into water, the resulting products are somewhat different from those formed on land. The observed subdivision of pahoehoe into flow units on land is represented by a more extreme subdivision under water, the flow units then being called pillows. Pillows form in much the same way as pahoehoe flow units (Moore 1975). The observed partial fragmentation of aa on land is represented by a more thorough fragmentation under water, the resulting glassy debris being called hyaloclastite. Hyaloclastite is particularly prone to form in the wave-swept coastal surf zone and on the steep submarine slopes of lava deltas (Moore et al. 1973). The morphology of the submerged part of a rift zone is graphically described by Lonsdale (1989).

*Constructional Form and Collapses of Hawaiian Volcanoes*

The Hawaiian Islands are the tops of great basaltic volcanoes, the major parts (about 85%) of which are hidden below the surface of the ocean. Each volcano has the subaerial form of a shield, shown to perfection by Mauna Loa. Recognizable remnants of the original shield surface survive on the lower slopes of even deeply eroded shields such as Koʻolau on Oʻahu. It is often stated that Hawaiian shields have slopes of only a few degrees, and this is indeed true of some. Slopes of 10° to 25° (measured from the horizontal) are not, however, unusual (Figure 6). The

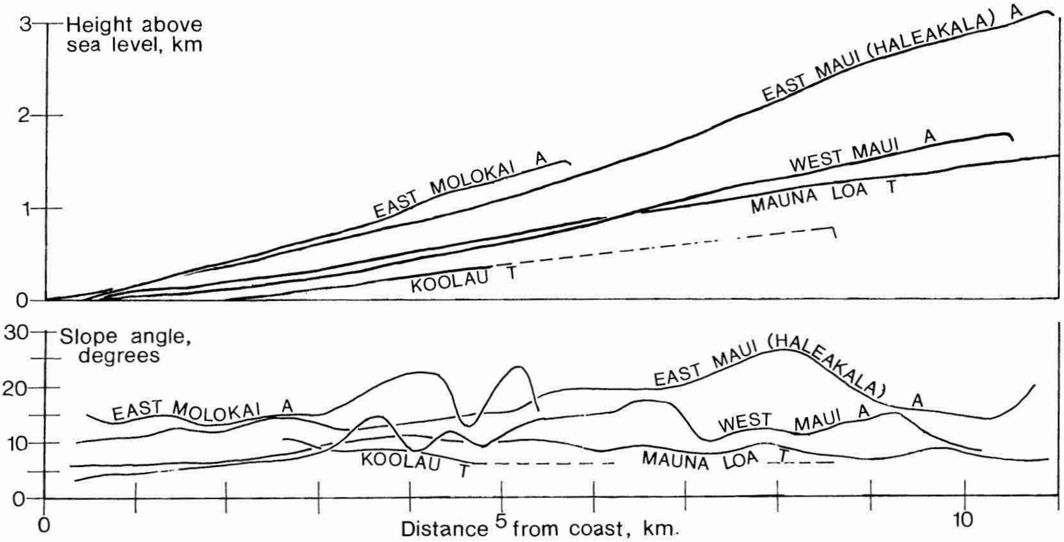



FIGURE 6. *Above*: Representative profiles across parts of unbuttressed flanks of some typical Hawaiian volcanoes transverse to their rift zones (T, tholeiitic shields: A, shields with alkalic cap). No vertical exaggeration. East Moloka'i, south flank along 156° 52′ meridian; Haleakalā, south flank N 10° W to Magnetic Peak; West Maui, west flank due east from coast to 3000 ft (915 m) and thence S 80° E to Pu'u Kukui; Mauna Loa, west flank along parallel of 19° 25′ N; Ko'olau, N 45° E along St. Louis Heights: dashed line extrapolated to summit of Mt. Olympus. *Below*: Generalized slope angles based on contour spacing, in degrees measured from the horizontal, along the same profiles as above. Slopes of 5° to 20° occur in all, with little tendency to vary systematically with height; shields with an alkalic cap are generally steeper.

steeper slopes characterize volcanoes that have a thick cap of alkalic lava.

The low-angle shieldlike form of Hawaiian tholeiitic volcanoes is often attributed to the low viscosity of the lava, but there are at least five other factors that may be equally important: (1) Lava erupted at a high discharge rate (aa) is channeled in powerful lava rivers that rapidly convey it far from the vent. (2) Lava erupted at a low discharge rate (pahoehoe) advances slowly, but discharge is often sustained sufficiently long that stable tube systems develop, through which the lava travels far. (3) Many eruptions occur in rift zones on the volcano flanks and do not contribute to the building of a central cone. (4) Persistent subsidence of the summit area of each volcano is sufficiently fast almost to compensate for upbuilding of lava there and generates a central caldera that acts as a "sink" and engulfs most of the summit-erupted lava. (5) The volcanoes are significantly widened by the injection of dikes in rift zones.

A coastal terrace having slopes commonly of $<5°$ occurs on the active volcanoes and is bounded by steeper underwater slopes on the seaward side. This terrace consists of lava deltas formed where lava flowed from land into water. The older volcanoes possess a similar feature underwater regarded as a submerged lava-delta terrace.

Hawaiian volcanoes grow to a great height —Mauna Loa rises nearly 10 km above the deep ocean floor. Partly because of their large size and partly because of fissuring and dike injections that bodily shoulder aside the rifted edifice, they tend to form mechanically unstable structures. Massive collapses therefore occasionally occur. One collapse took away nearly half of Ko'olau volcano on O'ahu, exposing the Pali cliff, and another took away the northern half of Moloka'i, exposing the great cliffline of Moloka'i's north coast (Moore 1964). Part of the west side of Mauna Loa, from Kealakekua Bay southward, also collapsed, and although subsequently erupted

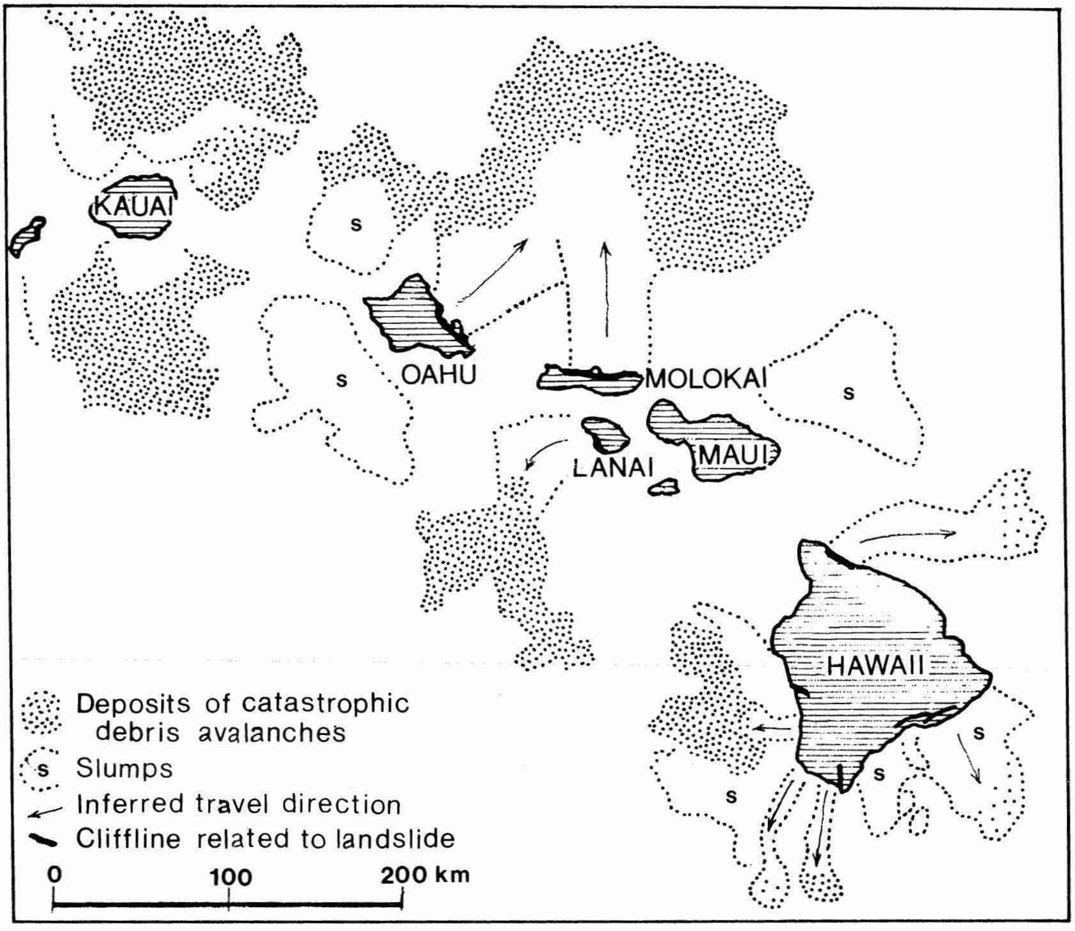



FIGURE 7. The southeastern Hawaiian Ridge showing slides (delineated by dashed lines) and areas of hummocky bottom topography (stippled) identified as debris accumulations (after Moore et al. 1989).

lava has largely healed the scar the lower slopes of the volcano are still anomalously steep in this sector.

The wide extent of the debris from these volcano-collapse events (Figure 7), clearly visible on sonar images of the ocean floor, suggests that they were catastrophic: the debris from the west side of Mauna Loa (the ʻĀlika Slide), for example, is spread over an area of ocean floor almost as large as the island of Hawaiʻi (Lipman et al. 1988).

One consequence of catastrophic volcano collapses is the generation of great tsunamis. Limestone debris has long been known to occur up to 326 m above sea level on the surface of Lānaʻi, up to 73 m on West Maui, and up to 65 m on Molokaʻi and was originally thought to have been deposited at a time of exceptionally high stand of the ocean (Stearns et al. 1940). It was recently reinterpreted as the deposits of a great tsunami (Moore and Moore 1984).

Volcano collapse is not necessarily catastrophic; slow collapse of the mobile south flank of Kīlauea volcano is currently occurring at an average rate of tens of centimeters per year (Swanson et al. 1976). Movement is accommodated partly by a widening of Kīlauea's rift zones as dikes are injected and partly by slip on the great (presumably listric) faults of the Hilina system. Some of the movement is episodic and accompanies major earthquakes. Much of the south coast of Kīlauea subsided from 1 to 3.5 m and moved southward by as much as 7 m in the magnitude 7.2 earthquake of November 1975, and tsunamis devastated coastal villages in the earthquake of 1868 (Tilling et al. 1976, Lipman et al. 1985).

The south flank as well as much of the area between Kīlauea and Mauna Loa is seismically very active, and the depth of many earthquake foci is consistent with lateral slippage of these parts of Kīlauea over a detachment surface at about the level where the base of Kīlauea rests on old ocean crust. Detachment may occur in the layers of marine sediments anticipated to occur there (Nakamura 1980). The exact relationship between this inferred detachment structure and the faults of the Hilina and Koa'e systems (Duffield 1975) is yet uncertain.

*Subsidence of the Hawaiian Islands*

All the Hawaiian Islands are slowly subsiding and will eventually become seamounts. There is abundant evidence for subsidence; for example, the terrace of reef limestones on which most of Honolulu is built is more than 250 m deep, and because reef corals grow only in the shallow water of the photic zone, this implies that about 250 m of subsidence has occurred since the deeper part of the terrace was constructed. A change of bathymetric slope that appears to mark the submerged coastal terrace of lava deltas occurs 600 m deep around O'ahu, implying that 600 m of subsidence has occurred since O'ahu's volcanoes became inactive about 2 Ma (average subsidence rate 0.3 mm/yr).

The volcanoes in the Hawaiian-Emperor Chain become progressively deeper toward the older (northwestern) end, and the upper part of Suiko Seamount, now at a water depth of 2400 m, consists of lava flows formed on land (Jackson et al. 1980), implying that at least 2400 m of subsidence has occurred in 64 Ma (average 0.04 mm/yr).

This slow subsidence is attributable to a general cooling and thickening of the Hawaiian lithosphere as it moves away from the hotspot (Crough 1983). The subsidence rate is too small to cause concern to man, unlike the rapid fluctuations in sea level that may accompany future changes in the Antarctic and Greenland ice budgets.

Hawaii is subsiding faster than the other islands. Rapid subsidence is evident from cursory observations: for example, at Pu'uhonuao Hōnaunau on the west coast of Mauna Loa palm-tree molds occur in young lava at sea level, where trees do not grow today. Subsidence at Hilo is documented by tide gauge measurements, and from 1946 to 1983 averaged 2.4 mm/yr (relative to Honolulu, assumed to be stable; Moore 1987). Geothermal drill holes in the Puna district entered pillow lava (formed under water) at about 250 m below sea level (Moore and Thomas 1988), indicating that 250 m of subsidence had occurred (neglecting possible sea-level changes).

The most striking evidence for rapid subsidence is supplied by the remarkable series of drowned reefs that occurs at depths varying from 100 m to 1300 m west of Lāna'i, Kaho'olawe, and Kohala (Moore and Fornari 1984, Moore and Campbell 1987). Reef-forming organisms in the latitude of Hawaii can evidently keep pace with a 2.4 mm/yr subsidence rate, but at times during the Ice Age sea level would have risen at several millimeters per year because of rapid melting of continental ice sheets. Reef-growing organisms could not keep pace with land subsidence combined with sea-level rise, and the reefs drowned.

The tilting of these drowned reefs toward the island of Hawai'i at 5 m per kilometer shows that in the past half-million years since the reefs formed Hawai'i subsided more rapidly than the islands farther northwest. This rapid subsidence is linked with the active volcanism. Volcanism involves a massive transfer of molten rock from below the lithosphere to a high crustal level (where intrusions form) and to the surface (where lava erupts).

The subsidence rate increases inland, and seismic refraction profiles (Zucca et al. 1982, Hill and Zucca 1987) indicate that the base of the crust has been depressed about 8 km beneath Mauna Loa. A narrow annular "moat" of deeper ocean, the Hawaiian Trough or Deep, occurs around the islands, and shallower ocean floor of the Hawaiian Arch occurs outside this moat. The geometry of the Hawaiian Trough and Arch is such as would be expected by a localized excess load (the Hawaiian Ridge) causing flexuring of a rigid and elastic lithosphere (Watts and Ten Brink 1989). These flexural structures are superimposed on the broader feature of the Hawaiian Swell.

The summit caldera of each active volcano is caused by localized excessive subsidence, commonly explained as accompanying removal of underlying magma during voluminous flank eruptions (Macdonald 1965). Alternatively it could result from an excessive loading of the lithosphere by dikes and very dense crystalline intrusive rocks underlying the subsummit magma chamber (Walker 1988).

One can speculate on whether the extinct volcanoes of the Hawaiian Chain, such as Koʻolau on Oʻahu, were ever as high as Mauna Loa or Haleakalā are today and whether they have subsided to their present lower heights.

*Level of Neutral Buoyancy*

Thought to be of great importance in understanding Hawaiian volcanoes and their roots is the possession by the volcanoes of a density layering such that the uppermost layers are less dense than common basaltic magmas and underlying layers are more dense. Ascending basaltic magma tends to pond at the level of neutral buoyancy, where rocks less dense than magma rest on rocks more dense than magma (Walker 1986, Ryan 1987*b*). Magma at this level is in a gravitationally stable position. It can reside stably in a magma chamber, or it can travel along the level of neutral buoyancy and form intrusions there. Injection and residence of magma at the level of neutral buoyancy is a viable alternative to eruption at the surface.

Consider the reasons for the density layering. Subaerial lavas of the upper layers have a low density because they contain many vesicles and other voids. Below the local water table all the voids are filled with water. Underlying pillow lavas are less vesicular but have numerous spaces between pillows, and associated hyaloclastites consist largely of glass, which is less dense than crystalline basalt. Intrusion complexes consist of rocks that may lack voids and include cumulates rich in olivine crystals that have a particularly high density. Pervasive hydrothermal alteration of the rocks at moderate depths replaces high-density minerals such as olivine with low-density hydrous minerals and infills the voids with similar minerals. In addition, contraction of the rocks (Ryan 1987*a*) because of pressure causes a general density increase with depth as voids (including microfractures) are progressively eliminated and at deeper levels as the compressibility of the minerals becomes significant.

To some extent the density structure can be investigated by seismic study, because the seismic-wave velocities through rocks are related to the rock density, and refraction of seismic waves occurs at density steps. Measurements of the gravitational force at the earth's surface also provides information on the integrated density in the underlying rock prism.

More than one level of neutral buoyancy may occur for a given magma, and magmas having different densities may be neutrally buoyant at different levels. For tholeiitic magma the level is likely to be shallow, whereas for more picritic magmas it is deeper. A seismic layer that locally occurs at the Moho (Ten Brink and Brocher 1987) is interpreted to consist of mafic intrusions underplating the crust and may have formed at a deep level of neutral buoyancy there, and Delaney et al. (1990) present evidence for the existence of a deep magma body beneath Kīlauea.

*Intrusions*

Magma that solidifies underground forms intrusions. The most abundant intrusion type in Hawaiian volcanoes is the dike, a wall-like body of rock formed where magma infills a crack. Only a few dikes are seen on the actively growing volcanoes, but great complexes consisting of hundreds to many thousands are revealed by erosion in the older volcanoes. They average about 75 cm wide. In most, the plane of the dike is inclined at 70–85° to the horizontal. The most flat-lying ones are inclined at $<45°$ and are called intrusive sheets or sills.

A dike complex is the subsurface expression of a rift zone and yields important insights into the subsurface processes of volcanoes. Some dikes were channelways by which lava traveled to the surface and erupted. Other dikes represent noneruptive magma excursions into the rift zones such as have often been documented below Kīlauea by their accompanying seismicity (Klein et al. 1987). Narrow dikes in particular are likely to solidify so quickly that they fail to reach the land surface (Bruce and Huppert 1989).

Dikes are injected so that the plane of the dike is at right angles to the direction of least compressive stress. In one of the most outstanding contributions to Hawaiian geology, Fiske and Jackson (1972) noted the superficial character of Hawaiian rift zones and explained their orientation as being determined by the force of gravity acting on the volcanic edifices. They conducted experiments in which they injected a fluid into the centers of gelatin models simulating volcanic edifices and observed the outward propagation of bladed "dikes" along the elongation axis of these models. They also pointed out that the rift zone orientation in a younger edifice is strongly influenced by the buttressing effect of an earlier contiguous edifice.

There are weaknesses in the Fiske and Jackson model; for example, the injection of only a limited number of dikes would be enough to relieve stress in a real edifice (Rubin and Pollard 1987). Also although activity in a Hawaiian rift zone may indeed be shallow, it is shallow relative to today's land surface; earlier in the history of a volcano the rift activity was similarly shallow, but it was deep relative to today's surface. Hawaiian rift zones should therefore have been initiated in underlying oceanic crust.

Opinions differ on whether dikes are forcibly injected into rift zones (Swanson et al. 1976) or passively enter positions in the volcano that have a particularly favorable stress field (Dieterich 1988). Whichever view, dike injection causes lateral extension, and one or both flanks must be capable of moving so as to accommodate the extension. If the flanks are not capable of moving, then injections of flat-lying intrusive sheets or sills are favored instead of dikes.

### *Dike Concentration and Parallelism in Rift Zones*

The rift zones of Hawaiian volcanoes in their tholeiitic shield-bearing stage are narrow (typically 1 to 4 km wide), and the visible eruptive and noneruptive fissures have a parallel strike. Modeling of the Bouguer gravity anomalies (Furumoto 1978) indicates that the underlying dike complex is wider. In some examples the surface rift zone lies to one side of the gravity high (Lipman 1980). The dike complex of the eroded Ko'olau volcano exceeds 7 km wide and shows a very high concentration and high degree of parallelism of dikes.

Later in this article reasons are presented for thinking that rift zones are initiated in the underlying oceanic crust at the time of inception of Hawaiian volcanoes. Whether this is so or not, there are three reasons why a rift zone, once established, tends to be perpetuated and why the dike intensity in it is high: (1) An intense rift zone consisting of $\geqslant 50\%$ largely nonvesicular dike rocks has a higher bulk-rock density than dike-free volcanic rocks; the edge of the dike complex is therefore a position of neutral buoyancy to typical basaltic magmas and strongly influences where subsequent dikes are injected (Figure 8) (Walker 1986). (2) If magma excursions occur frequently into a rift zone, then earlier dikes that are still hot and may indeed still contain magma pockets provide thermally favorable pathways for subsequent dike injection. Wilson and Head (1988) considered the question of frequency, and Garcia et al. (1989) presented petrological evidence for the persistence of magma pockets in Kīlauea's east rift zone. (3) Each wall of a dike is a plane of weakness that may guide injection of a younger dike.

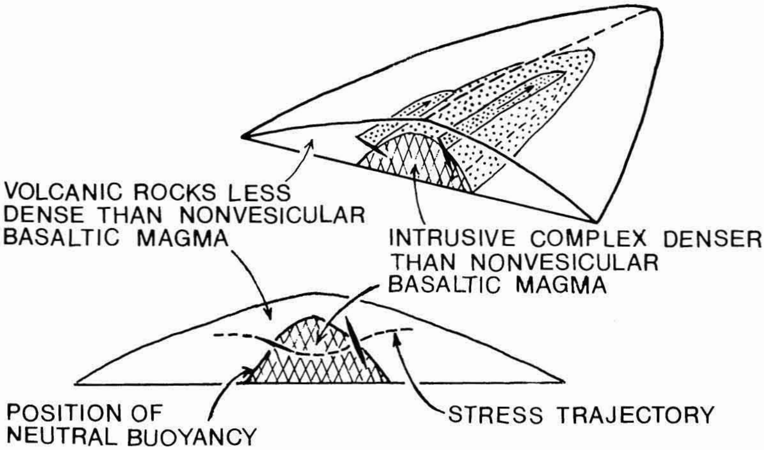



FIGURE 8. Schematic view and profile across a rift zone showing bulk-rock density zones. A high-density intrusion complex projects up into relatively low-density volcanics, and the position of neutral buoyancy (LNB) for typical basaltic magmas occurs between them. Intrusions strongly favor this position and (according to the stress field) form either nonvertical dikes or inclined intrusive sheets. Once established, an intrusion complex tends to be self-perpetuating.

*Geology and Volcanism of the Island of Hawai'i*

The island of Hawai'i consists of the tops of five great shield volcanoes. Two, Kīlauea and Mauna Loa, have had many eruptions in historic time, and Hualālai has had one historic eruption. Their growth rate exceeds their erosion rate. The other two, Mauna Kea and Kohala, are respectively dormant and extinct and are being strongly eroded.

KĪLAUEA volcano is highly active and is regarded as the youngest volcano on Hawai'i. About 90% of the surface area of 2500 km$^2$ consists of lava flows (mainly pahoehoe) younger than 1100 yr (Holcomb 1987), and the oldest rocks (the Hilina Basalt) seen at the base of fault escarpments on the mobile south flank are thought to be between 100,000 and 30,000 yr old (Easton 1987). Kīlauea had significant flank eruptions low on its east rift zone (ERZ) in 1790 and 1840 and low on its southwest rift zone in 1823, but activity between 1830 and 1934 was concentrated in its summit caldera, which acted as a sink for most of the volcano's output. The central vent in the caldera is the pit crater of Halema'uma'u. A lava lake was active there for many decades. The many pit craters in the ERZ are subsidence pits up to 325 m deep and are ephemeral features destined to be infilled with lava. One formed in 1921, but two were buried by the Mauna Ulu eruptions of 1969–1974.

Halema'uma'u varied greatly in depth as periods of infilling alternated with subsidences. The greatest depth of 900 m was reached after the violent phreatic explosions in 1924. Similar but larger explosions, accompanied by base surges that killed a party of warriors, occurred in or around 1790 (Swanson and Christiansen 1973).

Kīlauea was in repose from 1934 until 1952, but since then has been the most productive volcano in Hawaii and probably in the world (Swanson 1972, Dzurisin et al. 1984). An eruption low on the ERZ occurred in 1955. Kīlauea Iki pit crater was partly infilled to a depth of 130 m by lava in 1959, and another eruption low on the ERZ followed in 1960. Other ERZ eruptions built the 140-m-high lava shield of Mauna Ulu in 1969–1974 (Swanson et al. 1979, Tilling et al. 1987), the 280-m-high pyroclastic cone of Pu'u O'o that fed many aa flows in 1983–1986 (Wolfe et al. 1987), and the extensive pahoehoe flow-field from Kupaianaha lava lake since mid-1986.

Many dike injection events, with or without eruptions, have also occurred since 1952 (Klein et al. 1987).

MAUNA LOA rises to 4167 m above sea level and about 10 km above the isostatically depressed ocean floor. It is the largest volcano on earth, having a volume estimated to exceed 40,000 km$^3$ (Bargar and Jackson 1974). Moku'āweoweo caldera occurs at the summit (Macdonald 1977), and rift zones extend southwest and northeast from it; in addition, isolated radial fissures occur on the northwestern flanks. Moku'aweoweo is a relatively young structure that truncates lava flows only 590 yr old. The period since 1843 has been one of net infilling, the caldera acting as a sink that has absorbed about 30% of the volcano's output.

Since 1843 Mauna Loa has had 32 eruptions. Many began with firefountaining from fissures in or near Moku'āweoweo, and in more than half of them this was followed within a few days by extension of the fissures into a rift zone and voluminous flank eruptions. Two flank lava flows, those of 1859 and 1881, are more than 45 km long and the latter reached the edge of Hilo. The latest lava flow, in 1984 (Lockwood et al. 1987), traveled toward and caused concern in Hilo, but stopped short of the city.

It is thought that Mauna Loa took about 0.5 million yr to form. It is a highly productive volcano, and about 40% of the surface consists of lava younger than 1000 yr (Lockwood and Lipman 1987). Declines in its output began in 1868 and again since 1950, but it is not known if this declining trend will continue.

There has been much speculation over possible relationships between Mauna Loa and Kīlauea. Seismicity thought to be related to ascent of magma indicates that the plumbing systems converge at depth (Klein et al. 1987). The rarity of concurrent eruptions suggests that the two volcanoes compete for the same magma supply. The decline in Mauna Loa's output since 1950 is matched by a corresponding increase in Kīlauea's since 1955 (King 1989). Distinct geochemical differences between the lavas point to separate plumbing systems and geochemically different magma sources. Recent studies, however, suggest that Kīlauea's magma system has been invaded in comparatively recent times by Mauna Loa magma (Rhodes et al. 1989).

A major anomaly on Mauna Loa is the Nīnole Hills, a group of relatively old (about 0.2 million yr old) lava outliers that are evidently erosional remnants of an older portion of Mauna Loa. The lavas of these hills have dips parallel to the younger Mauna Loa flows that surround them and presumably originated from much the same part of Mauna Loa (the southwest rift zone) as these younger flows.

Possibly the Nīnole Hills are a surviving remnant of an older Mauna Loa cone, much of which collapsed to generate the 'Ālika and other slides (Lipman et al. 1988).

HUALĀLAI volcano has a well defined northwest-striking rift zone along which many cinder cones and several small pit craters occur, and from the Bouguer gravity anomaly a south rift zone extends under Mauna Loa lavas. The entire visible part of Hualālai consists of alkalic lava (Moore et al. 1987), but tholeiitic basalts occur offshore and in drill holes. The surface lavas are mostly alkalic olivine basalt, but include trachyte, which forms the prominent cone of Pu'u Wa'awa'a and the 100- to 200-m-thick lava that extends downslope from it. Some obsidian occurs on this, one of the most silicic lavas known in Hawaii.

Hualālai erupted seven times in the past 2100 yr from vents mostly in the northwestern part of the rift zone. The only historic eruption produced two flows, one largely aa lava in 1800 and the other (on which much of Keāhole airport is built) mainly pahoehoe in 1801. The 1800 lava is noteworthy for the extraordinary volume of dunite and gabbroic blocks it brought to the surface, thought to be fragments of intrusions in the volcano core (Kirby and Green 1980, Jackson et al. 1981). Lag accumulations of these blocks are admirably seen below the microwave station 2.5 km east of Hualālai Ranch (Jackson and Clague 1982).

The latest prehistoric eruptions occurred 300, 710, 800, 900, 1180, and 2030 yr ago (Moore et al. 1987) and were larger than average Hawaiian eruptions. That 710 years ago included a powerfully explosive phase that dispersed debris over a wide area. Future eruptions are anticipated, and Hualālai may be considered as potentially the most dangerous Hawaiian volcano. A considerable and rapidly growing population lives on its flanks and at its foot, quite close to its rift zone and on or below steep slopes. The 1800 lava evidently flowed fast, and it can be anticipated that future flows also will travel fast.

MAUNA KEA, at 4205 m the highest volcano in Hawaii, last erupted about 3600 yr ago and will likely erupt again. The whole visible edifice is alkalic but tholeiitic basalts are known offshore and in drill holes. Alkalic cap eruptions produced over 300 widely scattered cinder cones and lava flows (mostly hawaiite, with minor mugearite; West et al. 1988, Frey et al. 1990) that conceal earlier structures such as rift zones and a probable caldera (Porter 1972*a*). The volcano is elongated on an east-west axis that probably marks the strike of its main rift zones (the eastern continues as a prominent underwater ridge off Hilo), and an ill-defined south rift zone also occurs. Some alkalic cap eruptions brought up abundant dunite and gabbroic blocks thought to be derived from intrusions in the volcano core. Lava flows traveled down and locally ponded in several valleys; for example, impressively prismatic-jointed lava locally with pillows at its base flowed down the Wailuku River at Boiling Pots, and a lava delta formed at the seaward end of a valley at Laupāhoehoe.

Mauna Kea is the only summit in the tropical mid-Pacific that is known to have experienced glaciation. Glaciers extended over more than 64 km$^2$ and reached down to 2800 m elevation. Four glacial drifts are found interstratified with volcanic rocks of the past 280,000 yr (Porter 1979).

KOHALA is the oldest volcano on the island and last erupted about 60,000 yr ago; it is regarded as extinct. It has well-defined collinear rift zones striking northwest and southeast from the summit area. Many cinder cones occur on these rifts, and the highest reaches 1644 m above sea level. Erosion has carved deep canyons on the windward side, including Waipi'o, which is as much as 750 m deep and has an alluvial infill apparently extending to well below sea level. Most of the surface lavas are alkalic, but the deep canyons penetrate down into underlying tholeiitic flows, about 100 m of which are exposed; for example, in the Waipi'o valley (Spengler and Garcia 1988). Alkalic cap lavas range in composition from hawaiite to trachyte and total a maximum of about 150 m thick. Their eruption frequency averaged about one per 2000 yr. The straight line of coastal cliffs 600 m high extending 17 km northwest from Waipi'o appears to be fault-controlled. This prominent feature has been identified as the headwall of a large and mainly submarine collapse feature called the Pololū debris avalanche by Moore et al. (1989).

LŌ'IHI is a submarine seamount volcano 28 km south of the island of Hawai'i and is thought to be the youngest edifice constructed on the Hawaiian hotspot (Malahoff 1987, Fornari et al. 1988). The base is about 4000 m deep, and the top reaches to 969 m below sea level. A shallow caldera or plateau 2 to 3 km wide that has pit craters on its floor occurs at the summit. Well defined south-southeast and north ridges extend from the summit area and are interpreted to be rift zones. Low-temperature geothermal vents and very fresh lavas occur in the summit area, and Lō'ihi is seismically very active (Klein 1982), although eruptive activity has not yet been observed in progress. Alkalic basalts occur among the products of Lō'ihi (Frey and Clague 1983), suggesting that alkalic magmas erupt in an early as well as a late stage in the history of Hawaiian volcanoes. Lō'ihi is not expected to become a new island for 10,000 yr or more.

### *Volcanoes of Maui Volcanic Complex*

The cluster of islands comprising Maui, Kaho'olawe, Moloka'i, and Lāna'i is sometimes referred to as the Maui Volcanic Complex and was once a single island consisting of probably seven major volcanoes.

EAST MAUI OR HALEAKALĀ is the larger of two volcanoes that constitute Maui (Stearns and Macdonald 1947). It rises to 3055 m above sea level and is elongated on east and southwest rift zones, which are marked by many prominent cinder cones. A scatter of cones that occurs over the north flank may mark a diffuse north rift zone. Most of the visible part of Haleakalā consists of alkalic lavas, but lavas of the tholeiitic shield are exposed around parts of the north coast and in several deep valleys. Scenically spectacular Haleakalā Crater (Macdonald 1978) is a cliff-girt depression 11 km long by 3 km wide by 800 m deep. It is essentially an erosional feature, consisting of the expanded and merged heads of two major valleys (Kaupō on the south and Ke'anae on the north). A third valley, Kīpahulu, to the southeast is separated from the main crater by a narrow ridge and will eventually merge. Subsequent volcanic activity built cinder cones across the crater, and voluminous lava flows traveled down the three valleys and built lava deltas that extended the shoreline outward. Some of the lavas are ankaramite and rich in crystals of pyroxene and olivine. Other lavas contain large crystals of plagioclase or hornblende. Extensive rejuvenation-stage volcanism occurred in the southwest and east rift zones (West and Leeman 1987). The latest eruption occurred in about 1790 from two vents low on the southwest rift zone, near the 465-m and 170-m levels, and produced aa lava flows that form a conspicuous lava delta at Cape Kīna'u.

WEST MAUI is a tholeiitic lava shield capped by a discontinuous layer 20 to 225 m thick of alkalic rocks. It has a maximum elevation of 1764 m at Pu'u Kukui. The original shield form is clearly recognizable despite dissection by valleys as much as 1 km deep. A small infilled caldera is identified in the 'Īao Valley southeast of the volcano center, from which several rift zones (trending south-southeast, north-northeast, north-northwest, and southwest) appear to radiate. According to Diller (1982), the principal rift zones that contain the most dikes trend north-northwest and south-southeast; and the southwest rift zone is a minor one that was increasingly favored in the declining stage of activity. Several trachytic domes occur in the alkalic cap, including Pu'u Māhanalua 5 km southeast of Lahaina and Puu Ka'ae on the northeast coast. Two minor rejuvenation-stage vents occur on the western slope and include the cinder cones of Keka'a Point and Pu'u Lahaina (upslope from Lahaina); the latter erupted picritic lava. Another cone (now largely quarried away) is Pu'u Mele near the southeastern tip of West Maui.

KAHO'OLAWE. This almost-uninhabited island is a low-angle lava shield only 443 m high with an infilled caldera 5 km wide at Kanapou Bay on the eastern side. Precaldera lavas are tholeiitic, and at least part of the caldera infill is alkalic (Fodor et al. 1987). A well defined west-striking rift zone occurs, with many dikes exposed at Kanapou Bay, and this zone is in line with the small tuff ring of Molokini and the main rift zone of Haleakalā. Kaho'olawe is elongated on the same line. Samples of cap lavas from the west side of the island gave K-Ar ages of slightly over 1 million yr. Some vents that erupted spatter and cinder on the cliff at Kanapou are reported to have a very youthful appearance (Stearns et al. 1940). Because of the low rainfall and overgrazing by sheep in the last century, the upper layers of soil were stripped off Kaho'olawe by wind and water to expose a barren red subsoil over much of the island.

LĀNA'I is a shield volcano that is elongated in a general northwest direction along its main northwest and southeast rift zones. Numerous dikes of the southeast rift zone and also a number of parallel faults are exposed in the seacliffs between Mānele Bay and Kapoho Gulch on the south coast. The flat-floored Pālāwai and Miki basins near the south end appear to be partially infilled subsidence structures on the floor of a caldera that is about 7 km wide. The highest elevation, 1027 m, is on the northwestern caldera rim. A minor rift zone appears to strike southwest from the caldera. So far as is known, all the lavas of Lāna'i are tholeiitic. A particularly noteworthy feature is the occurrence of fossiliferous marine surface deposits up to 326 m in elevation, now interpreted to be the deposits of a giant tsunami (Moore and Moore 1984). The seacliff up to 300 m high on the southwest coast may mark where a major (possibly tsunamigenic) collapse occurred.

EAST MOLOKA'I. This, the larger of the two volcanoes that form the island of Moloka'i, rises to 1813 m above sea level and has depositional slopes of 6° to 22°. It has two rift zones represented by impressive dike complexes, one extending east and the other west-northwest from the summit area. A caldera has been identified in the upper Pelekunu and Wailau valleys from the thicker and more massive character of the lava flows, the widespread hydrothermal alteration of the rocks, and the presence of gabbro. Much of the shield consists of tholeiitic basalts, but alkalic lavas including hawaiite and trachyte form a cap. A great cliffline up to nearly 1200 m high bounds the north side of Moloka'i and marks where the entire northern half of the volcano collapsed into the ocean (Moore 1964). Rejuvenation-stage volcanism generated the Kalaupapa peninsula at the foot of this cliff and the tuff ring of the two tiny islands of Mokuho'oniki and Kanahā off the southeast coast. Kalaupapa is a pahoehoe shield, and its crater (Kauhakō) has a rim that rises 123 m above sea level and contains a small lake that has been plumbed to a depth of 255 m (Clague et al. 1982). Kalaupapa forms a very prominent bathymetric feature and has a total estimated volume of 3 km$^3$.

WEST MOLOKA'I volcano is a shield that has a low profile and is elongated east-northeast on what appears to be its main rift zone. Another rift zone strikes northwest, and a number of fault scarps having the same strike occur in the saddle between West and East Moloka'i volcanoes. West Moloka'i appears to consist largely of tholeiitic lavas, with a thin discontinuous alkalic cap (Clague 1987*a*). A deep red soil resulting from prolonged weathering caps the volcano except for the prominent strip of sand dunes consisting of calcareous sand that extends for 7 km southeastward from Mo'omomi Beach. Parts of the dunes are lithified and no longer active and were blown by wind from beaches exposed when sea level was lower.

PENGUIN BANK is a broad submarine feature, a nearly flat shoal 54 m below sea level, that was probably a low-profile volcano like West Moloka'i but was truncated by erosion and then capped by coral. It is in line with the main rift zone of West Moloka'i volcano.

### *Volcanoes of O'ahu*

O'ahu consists of the remnants of two eroded lava-shield volcanoes, Wai'anae and Ko'olau.

WAI'ANAE rises to 1227 m above sea level at Mt. Ka'ala. It consists of a tholeiitic shield with a thick cap of transitional to alkalic rocks (Sinton 1987). Rejuvenation-stage volcanics of undetermined age occur in Kolekole Pass and also form a line of well-preserved cinder cones (including Pu'u Makakilo and Pu'u Kapolei) on the south flank of Wai'anae. Wai'anae is very deeply eroded, and Lualualei and Wai'anae valleys present a landscape that is unusual for Hawaii, in which narrow and nearly vertical-sided ridges, some of them isolated, rise abruptly from the flat alluvial valley floors. A large caldera has been identified in Lualualei and Wai'anae valleys. Rhyodacite and icelandite occur among the lavas filling the caldera. Despite the deep valleys, erosion does not expose hydrothermally altered rocks of a former high-temperature geothermal system such as occur in the core of Ko'olau. Dike orientations define northwest and south-southeast rift zones (Zbinden and Sinton 1988), but dike intensities are lower than in the Ko'olau dike complex.

KO'OLAU is a tholeiitic shield that lacks an alkalic cap and is the only major Hawaiian island edifice that is predominantly composed of reversely magnetized rocks (Moberly and Campbell 1984). It rises to 960 m above sea level at Mt. Kōnāhuanui. The windward half of the Ko'olau shield is missing because of collapse (Moore 1964), and the pali (cliffline) marks the eroded-back collapse scar. An extraordinarily intense dike complex contains an estimated 7400 dikes totaling 4 km wide in one transect (Walker 1986, 1987). It is particularly well exposed in the Kapa'a Quarry and on the tiny Mokulua Islands.

There is clear evidence for a caldera centered on Kailua, where the lavas show a centripetal dip and are hydrothermally altered (Fujishima and Fan 1977). The caldera is centered on a great positive Bouguer gravity anomaly paralleling the rift zone (Strange et al. 1965) where seismic refraction demonstrated the presence at shallow depth of rocks having a high seismic velocity (Adams and Furumoto 1965). The dike intensity drops to a very low value in the center of the caldera where subsidence evidently kept pace with dike injection (Walker 1988).

Ko'olau has a wide scatter of rejuvenation-stage vents termed the Honolulu Volcanics (Figure 9), including the landmarks of Diamond Head, Punchbowl, Hanauma Bay, and Koko Crater (on the Koko fissure) and the Salt Lake/Āliamanu crater cluster. Peridotite inclusions occur in some deposits: dunite at Ulupa'u Head, spinel lherzolite at a vent near the Pali Lookout, and spinel lherzolite, pyroxenite, and garnet pyroxenite (also sparse rocks containing kaersutite or phlogopite) at Salt Lake and Āliamanu. Alignments of vents is on northeast-striking fissures such as the Koko fissure and the fissure on which Diamond Head, the Kaimukī pahoehoe lava shield, and Mu'umae cinder cone occur.

The highly porous lavas of Ko'olau and Wai'anae are great water-storage systems that yield 500 million gallons (1.9 billion liters) of groundwater per day to support O'ahu's population of nearly 1 million. Small dike-impounded reservoirs exist in the Ko'olau range; otherwise the water table is very flat (gradient typically 0.3 m/km (Figure 10); Hunt et al. 1988). Much of Honolulu is built on a prominent raised-beach platform consisting of limestone with interbedded sediments as well as rejuvenation-stage ashes and lavas. Lithified calcareous sand dunes occur at Lā'ie Point and elsewhere.

## *Volcanoes of the Kaua'i Group*

KAUA'I consists apparently of a single lava shield, reaching its maximum elevation of 1598 m in the Wai'ale'ale plateau near the center of the island. Dikes are numerous, but less concentrated into swarms than in other Hawaiian volcanoes; a swarm striking northeast and west-southwest can, however, be distinguished, in line with Ni'ihau and Ka'ula. Outstanding scenic features are the great cliffline of the Nā Pali coast on the northwest, and the 600-m-deep Waimea Canyon on the west. The cliffs, the country on the west side of the canyon, and much of the middle of the island including Wai'ale'ale are composed of tholeiitic basalt lavas of the Kaua'i shield. A plateau interpreted to be the infilling by alkalic basalt of a very large caldera 20 km across occupies a large area northeast of Waimea Canyon, and lavas occupying a graben extend south of the caldera.

Rejuvenation-stage volcanism was particularly widespread on Kaua'i, and about 40 scattered vents and associated lavas (the Koloa Volcanics) have been identified (Clague and Dalrymple 1988). Their eruption began about 2.65 Ma in the west-northwestern half of Kaua'i, with apparently no appreciable time hiatus between it and the declining-stage volcanics. All the Koloa Volcanics younger than 1.5 million yr occur in the east-southeastern half of Kaua'i centered on the Līhu'e basin, and they compose most of this half of the island. The youngest dated lavas are about 0.5 million yr old, but a submarine eruption is thought to have happened south of Kaua'i in 1956 (Macdonald 1959). An extensive lava flow of undetermined age occurs on the ocean floor of the Hawaiian Arch east of Kaua'i (Holcomb et al. 1988).

NI'IHAU is the eroded remnant of a single shield volcano made mostly of tholeiitic basalts rising to an altitude of 390 m and elongated west-southwest on the Kaua'i-Ni'ihau-Ka'ula line. Numerous dikes having the same trend have been mapped along the southeast side of the shield and mark the rift zone of the Ni'ihau shield (Stearns 1947). The shield is abruptly truncated on the east by a seacliff up to 360 m high, and apparently a large part of the volcano, including any caldera it might have possessed, has been lost to erosion. The shield remnant is rimmed on the north, west, and south sides by a low

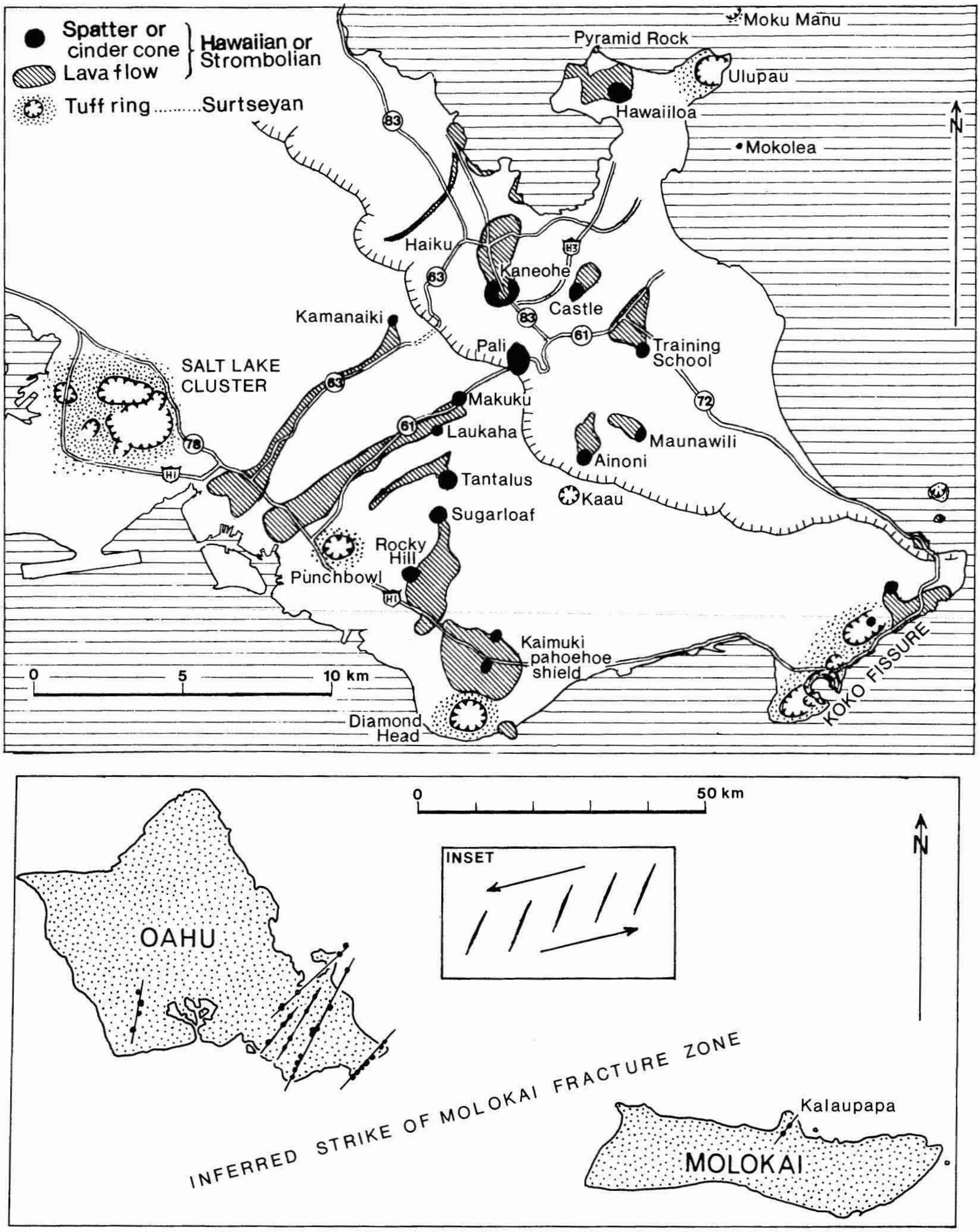



FIGURE 9. *Above*: Distribution of the Honolulu Volcanics, the products of rejuvenation-stage volcanism from monogenetic vents in the eastern half of O'ahu. *Below*: Northeast alignments of rejuvenation-stage vents of Wai'anae, the Honolulu Volcanics, and Kalaupapa (G. P. L. Walker and C. R. Coombs, unpubl. data) are consistent with tension gashes developed by a left-lateral shear motion (inset) over a locked Moloka'i fracture zone.

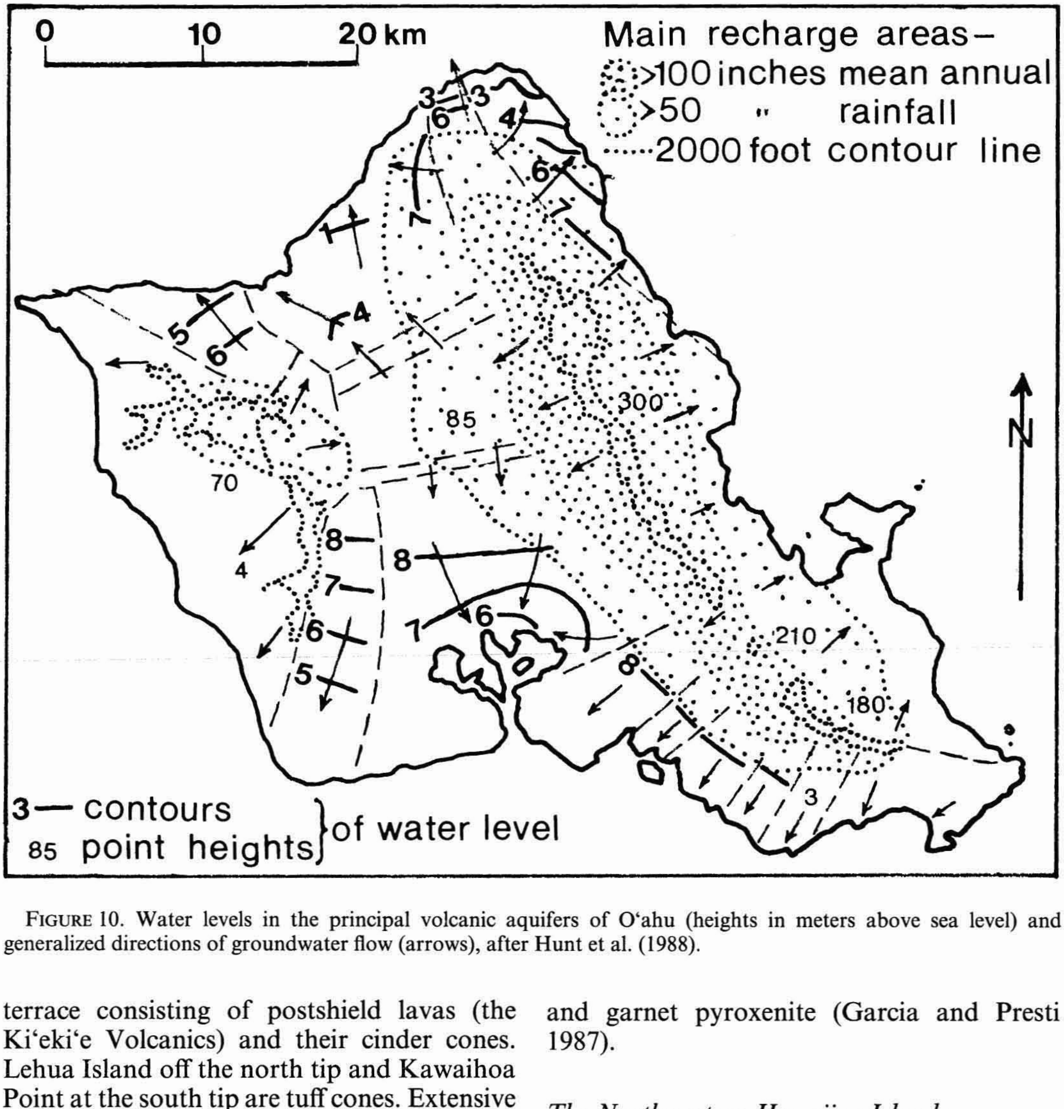



FIGURE 10. Water levels in the principal volcanic aquifers of O'ahu (heights in meters above sea level) and generalized directions of groundwater flow (arrows), after Hunt et al. (1988).

terrace consisting of postshield lavas (the Ki'eki'e Volcanics) and their cinder cones. Lehua Island off the north tip and Kawaihoa Point at the south tip are tuff cones. Extensive active and lithified dunes of calcareous sand also occur on the terrace, particularly near its southern end.

KA'ULA Island is a youthful, isolated tuff cone 160 m high. It is the only land on an extensive submarine bank that averages a little more than 60 m below sea level and appears to be the eroded top to a large shield volcano. Ka'ula is composed of nephelinite (Garcia et al. 1986) and is regarded as a rejuventation-stage vent. The cone contains mantle-derived inclusions of dunite, lherzolite, and garnet pyroxenite (Garcia and Presti 1987).

*The Northwestern Hawaiian Islands*

The Northwestern Hawaiian Islands consist of about 11 tiny islands, totaling 5.2 km$^2$ in area, spread over nearly 2200 km of ocean northwest of Kaua'i, and are all that is seen of the continuation of the Hawaiian Chain in that direction (Macdonald et al. 1983). Nihoa Island is 1.3 km long and is an eroded remnant of a tholeiitic shield cut by dikes. Necker Island is about the same size and consists of alkalic lava and dikes. Gardner Pinnacles are two sea stacks of basalt cut by dikes. The other islands are limestone reefs or atolls. One drill hole at Midway entered basalt, however, at a depth of only 55m. A volcanic eruption apparently occurred 90 km east of Necker Island in 1955, in an area with a water depth of 3600 m.

*Speculation on Position and Rift-zone Orientation of Hawaiian Volcanoes*

This final section speculates on the controls that determine the positions of volcanoes and the orientation of their rift zones. Fiske and Jackson (1972) pointed out the superficial character of the rift zones and the lack of any clear correlation between strikes of rift zones and the strike of either the Moloka'i fracture zone or the axis of the Hawaiian Archipelago (Figure 11). They concluded that "the influence of regional structure on the orientation of the rifts is at best, obscure" and explained the orientation as being determined by stress fields generated by the force of gravity acting on the volcanic edifices.

An alternative view of the interrelationships between Hawaiian volcanoes is presented in Figure 12. A critical element in this model is that compensation is made for a postulated substantial strike-slip movement along faults striking parallel with the Moloka'i fracture zone north and south of O'ahu. Seismic refraction study (Lindwall 1988) confirmed the existence in both places of major faults that bring contrasting crustal profiles into juxtaposition, but did not resolve the question of the amount or sense of motion that has occurred.

The postulated movements of up to 105 km are in the same sense as, and much less than, the known displacement along the Moloka'i fracture zone. According to current views, however, this displacement occurred earlier,

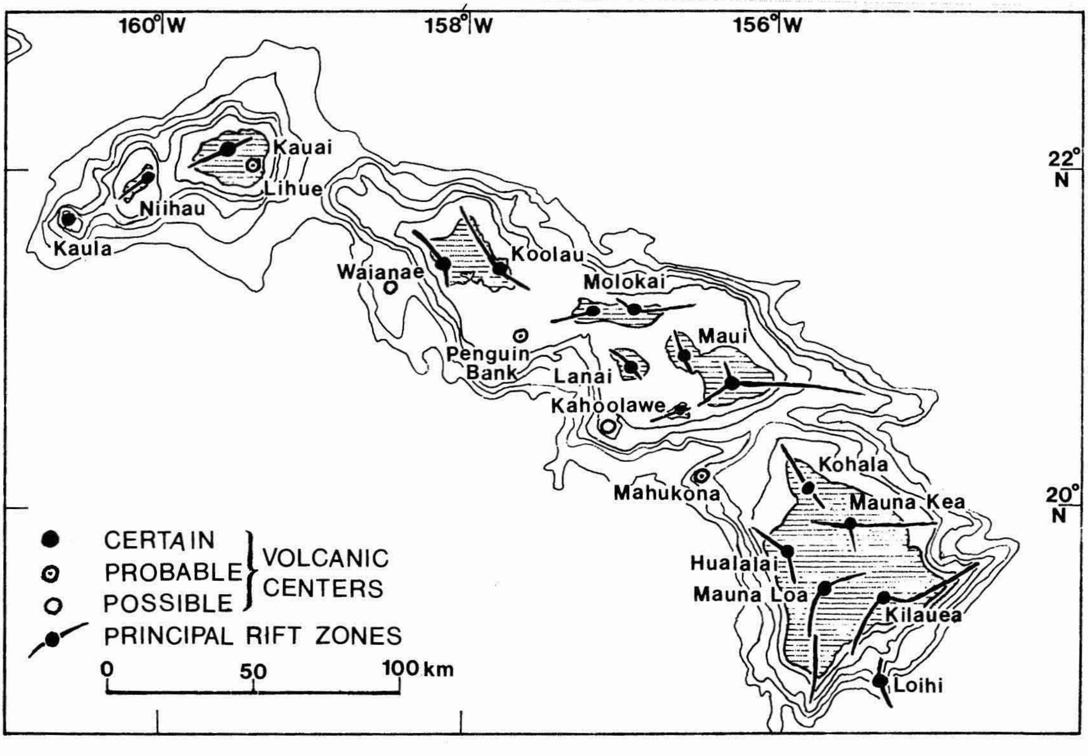



FIGURE 11. Distribution of volcanic centers and rift zones in the Hawaiian Islands. Two dominant alignments occur. One, approximately east-northeast, is parallel with faults in the Moloka'i fracture zone. The other is southeast; it is not readily explained. Bathymetric contours at 400-fathom (730-m) intervals down to 2000 fathoms (3660 m).

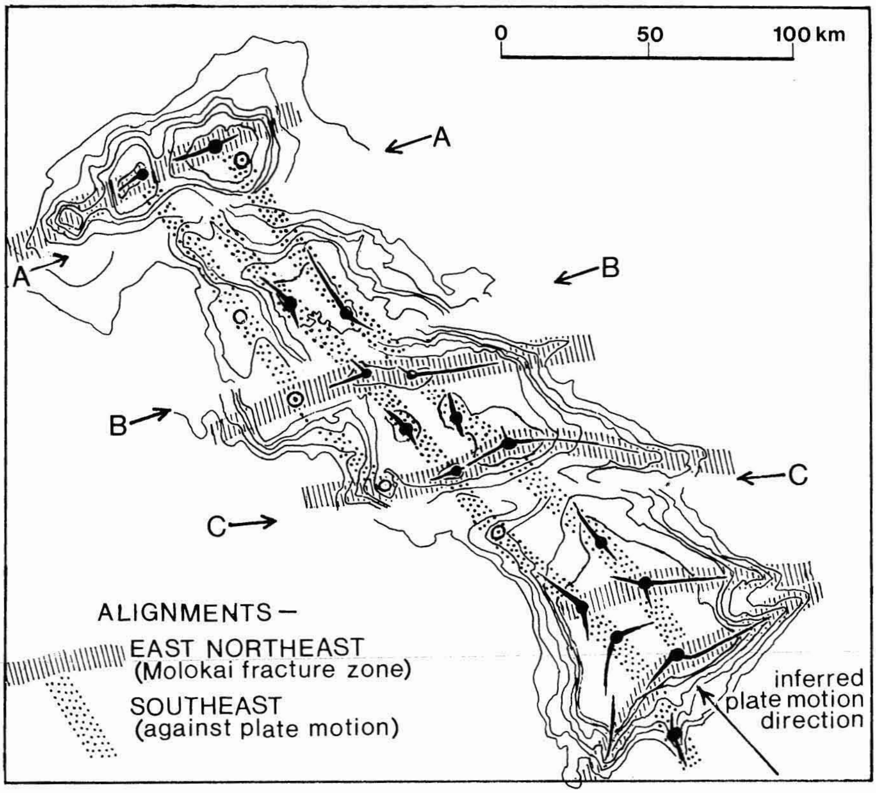



FIGURE 12. As Figure 11, after removing postulated left-lateral transcurrent motions of 105 km on a fault (A) between Kaua'i and O'ahu, 50 km on a fault (B) between O'ahu and Moloka'i, and 15 km on a fault (C) between Maui and Hawai'i. Proceeding southeast, volcanic centers and rift zones are alternately distributed along east-northeast lines (parallel with the Moloka'i fracture zone) and parallel southeast lines (against the plate-motion direction).

at the time when the zone was a transform fault. The following is therefore highly speculative.

When this compensation for fault displacements is made, a much-simplified structural pattern emerges in which the volcanic centers and their principal rift zones are seen to lie on (1) two or locally three chain-parallel lines having a southeast strike (against the inferred plate-motion direction), and (2) near-parallel transverse lines having an east-northeast strike (parallel with faults of the Moloka'i fracture zone).

The southeast alignments could be interpreted to mark fractures propagated in the oceanic lithosphere as it moved over the hotspot; radial outward flow in the mantle plume-head would create extensional stresses favoring such fracturing. The east-northeast alignments could be interpreted to overlie fractures in the underlying oceanic lithosphere parallel with the Moloka'i fracture zone.

Proceeding southeastward along the Hawaiian Chain from Kaua'i, the volcanic alignments show a clear alternation of transverse (predominantly east-northeast) and chain-parallel (predominantly southeast) trends.

Kaua'i, Ni'ihau, and Ka'ula lie on an east-northeast line, as do the long axis of Ni'ihau, the strike of the Ni'ihau dike swarm, the long axis of the tholeiitic shield of Kaua'i, and the strike of the dike swarm exposed in the Waimea Canyon on Kaua'i. The younger southwest part of Kaua'i appears to mark a change to the southeast tend, apparently with a late shift in the locus of volcanism to the

Līhu'e Basin; the axis of the Bouguer gravity anomaly of Kaua'i also follows this trend.

The next volcanoes are Wai'anae and Ko'olau both with southeast-trending principal rift zones. Ko'olau shows evidence for a southeastward progression of volcanic centers (Knight and Walker 1988). Two submarine ridges occur northwest of Ka'ena Point, O'ahu. One is likely a continuation of Wai'anae's northwest rift zone, and the other could be a parallel rift zone related to a possible unnamed volcano west of Wai'anae that is now part of the Wai'anae slump of Moore et al. (1989).

The next two volcanoes are the two on Moloka'i and the probable submerged volcano of Penguin Bank, aligned on an east-northeast-trending submarine ridge over 100 km long. Some dikes on East Moloka'i volcano and a group of young faults on West Moloka'i volcano, however, have a southeast trend. Lāna'i and West Maui volcanoes revert to the southeast trend.

Then, proceeding further down the Hawaiian Chain, Kaho'olawe and East Maui volcanoes occur on another transverse ridge over 100 km long. Both volcanoes are elongated on an east-northeast axis, their principal rift zones have the same strike, and the young tuff ring of Molokini occurs in the same line. Another volcano may possibly form the shallow bank 20 km west-southwest of Kaho'olawe. The submarine ridge is strongly curved east of Maui.

The northern volcanoes on Hawai'i, Kohala and Hualālai and presumably also the submerged volcano Māhukona, revert to the southeast strike. The alternation becomes less clear, however, farther southeast. Kīlauea has well-defined rift zones with an east-northeast strike, and Lō'ihi reverts to the other trend.

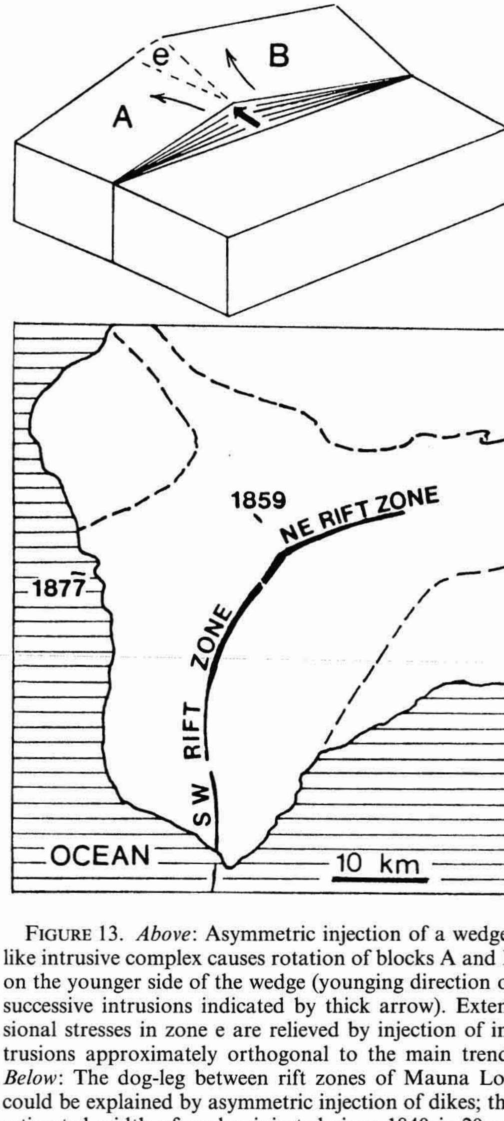



FIGURE 13. *Above*: Asymmetric injection of a wedge-like intrusive complex causes rotation of blocks A and B on the younger side of the wedge (younging direction of successive intrusions indicated by thick arrow). Extensional stresses in zone e are relieved by injection of intrusions approximately orthogonal to the main trend. *Below*: The dog-leg between rift zones of Mauna Loa could be explained by asymmetric injection of dikes; the estimated width of wedge injected since 1840 is 20 m. The 1859 and 1877 flank eruptions were from fissures approximately orthogonal to the rift zones.

## *Noncollinearity of Rift Zones*

Any study of the parallelism of rift zones in the Hawaiian Chain is complicated by the fact that the two principal rift zones of a Hawaiian volcano are generally not exactly collinear, but subtend an angle commonly of about 150°. A third lesser rift zone or scatter of radial fissures also generally occurs in the obtuse angle between the principal zones. In the eroded volcanoes dikes that occur approximately orthogonal to the principal dike strike may be members of the third rift zone.

A plausible explanation for these relationships depends on the fact that, in a given time interval, the lateral widening of a volcanic edifice by dike injection is greatest at the summit and decreases downrift. If two rift zones are initially collinear, asymmetric widening (by dike injection concentrated at one side of the rift zones) will cause a bend to develop. If widening as at Ko'olau is 4.5 km at the summit, and if it decreases to zero 25 km downrift, then the rift zones will come to subtend an angle of 160°. Extensional stresses are concentrated in the obtuse angle of the bend, which is where the third rift zone develops (Figure 13).

It is proposed, therefore, that rift zones of Hawaiian volcanoes were originally collinear and were aligned in either the southeast or east-northeast fracture direction. Asymmetric growth of the dike complexes then caused the rift zones to become noncollinear and resulting stresses were relieved by radial fissuring or formation of a third rift zone.

### *Trend of Rejuvenation-stage Fissures*

The ability of rejuvenation-stage magmas to attain the surface and erupt is generally attributed to tensional stresses set up in the lithosphere as it is flexed by passage across it of the Hawaiian Arch (Jackson and Wright 1970). There is, however, another possibility because of the fact that many of the rejuvenation vents lie on lines having a northeasterly to north-northeasterly trend. These alignments are clearly apparent among the Kōloa volcanics of Kaua'i, the Honolulu Volcanics and the young vents on southern Wai'anae in O'ahu, and on Kalaupapa on Moloka'i (Figure 9).

The trend of these alignments is consistent with their origin as tensional fractures relieving a strike slip stress pattern in the vicinity of "locked" faults of the Moloka'i fracture system.

### *Why Discrete Volcanoes?*

A basic question is why the Hawaiian Ridge consists of discrete volcanoes and is not a continuous uniform ridge. Several explanations have been proposed (Shaw et al. 1980). One is that conduits through the lithosphere are bent by plate motion, and when a conduit is bent to beyond some critical angle a new conduit develops (Whitehead 1982).

A very different approach to the problem was the observation by Moberly and Campbell (1984) that most of the volcanoes of the Hawaiian Chain are normally magnetized. From this they inferred that igneous activity along the chain occurred predominantly during the intervals of normal polarity of the earth's magnetic field, even though the earth's field polarity was reversed during about half of that time. Ko'olau is the only volcano in the Hawaiian Islands that is magnetically reversed. They concluded that magma generation in the mantle is modulated by processes in the core.

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