# GEOPHYSICAL CONSTRAINTS ON THE STRUCTURE OF KILAUEA AND MAUNA LOA VOLCANOES AND SOME IMPLICATIONS FOR SEISMOMAGMATIC PROCESSES

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## ABSTRACT

**Seismic-refraction and gravity data on and adjacent to Kilauea and Mauna Loa Volcanoes show that the Cretaceous oceanic crust bends downward beneath the load of the volcanic edifice. Dip on the base of the crust (*M*-discontinuity) increases from 3° under the submarine flanks of the volcanoes to 6°–10° under the subaerial flanks; depth to the *M*-discontinuity increases from about 10 km beneath the open ocean to roughly 13 km and 18 km beneath the high, subaerial flanks of Kilauea and Mauna Loa, respectively. This profile of the oceanic crust is consistent with flexural-lithosphere models for flexural response to the load of the volcanoes. *P*-wave velocity structure of the volcano flanks supports the Moore-Fiske model for the volcanic pile, in which an inner wedge of submarine pillow basalt is covered by a mantle of halyoclastics and subaerial basalt flows. Fast *P*-wave traveltimes and positive gravity anomalies over rift zones and summit vent indicate that dense, intrusive cores to these structures form a significant mass-fraction of the volcanic edifice. The triangular shape of these intrusive cores supports the view that the volcanoes grow in part by the spreading of unbuttressed flanks along subhorizontal decoupling surfaces near the base of the volcanic pile in respones to forceful magma intrusion into the rift zones.**


## INTRODUCTION

The internal structure of the Hawaiian volcanoes and the underlying oceanic lithosphere hold important clues to processes governing the growth and contemporary seismomagmatic activity of these massive, midoceanic, basaltic shield volcanoes. Kilauea and Mauna Loa on the Island of Hawaii are of special interest in this regard because, together with Loihi Seamount 55 km south of Kilauea's summit, they are the youngest and most active volcanoes in the Hawaiian-Emperor volcanic chain, and they mark the present position of the Hawaiian hot spot beneath the Pacific plate (see Clague and Dalrymple, chapter 1).

Eaton (1962) first developed a model for the crustal structure of the Island of Hawaii using the traveltimes of *P*-waves recorded on seven seismic stations located at distances of 18–390 km from two magnitude 5 earthquakes that occurred beneath the southeast flank of Kilauea in 1955. His resulting one-dimensional model (velocity varying with depth only) for the average crustal structure of the Island of Hawaii indicated that the volcanoes forming the island rest on a slightly thickened oceanic crust, with the Mohorovicic (*M*) discontinuity at a depth of 14–15 km and with the upper most mantle having a *P*-wave velocity of 8.25 km/s. More recent efforts by Ward and Gregersen (1973), Crosson and Koyanagi (1979), and Klein (1981) to define an average velocity structure for Hawaii in terms of one-dimensional models are based on *P*-waves from a large number of local earthquakes recorded on an increasingly dense seismograph network. These models all place the *M*-discontinuity at depths between 13 and 16 km and indicate an upper mantle velocity of 8.2–8.3 km/s, in general agreement with Eaton's initial model. In these more recent models, however, the average depth to intermediate *P*-wave velocities (6.5–6.8 km/s) is significantly shallower (5–6 km) than the 11 km in Eaton's (1962) model.

An image of the three-dimensional structure of the Hawaiian volcanoes began to come into focus with the gravity survey of Hawaii by Kinoshita and others (1963), the series of seismic-refraction measurements around the island by Ryall and Bennett (1968), Hill (1969), Zucca and Hill (1980), and Zucca and others (1982), and the study of teleseismic *P*-wave delays by Ellsworth and Koyanagi (1977). Results from these experiments indicate that the central vents and rift zones of Kilauea and Mauna Loa are underlain by dense cores of intrusive rocks having high *P*-wave velocities (6.5–7.4 km/s) and reaching to within a few km of the surface; they further indicate that these intrusive cores represent a significant fraction of the volume of the volcanic pile forming the Island of Hawaii.

In this paper we reexamime some of the evidence from seismic-refraction experiments and gravity measurements on the structure of Kilauea and Mauna Loa Volcanoes. We will focus in particular on evidence concerning (1) the relation of the volcanic pile formed by the volcanoes to the underlying Cretaceous oceanic crust (upper lithosphere) and (2) the geometry of the dense intrusive masses forming cores beneath the summit vents and rift zones. We conclude

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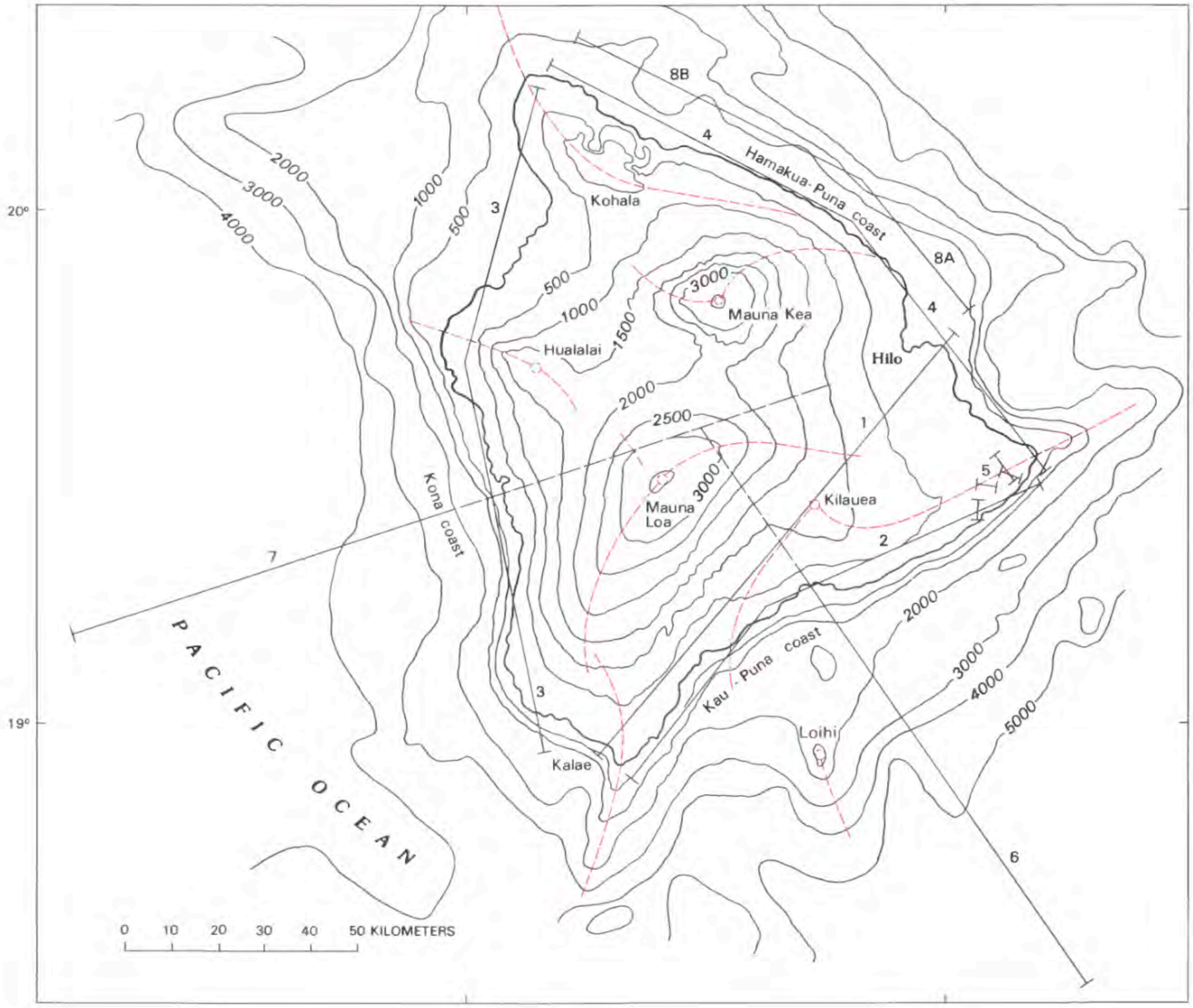


FIGURE 37.1.—The Island of Hawaii showing location of seismic-refraction profiles identified by number: 1, Ryall and Bennett (1968); 2–4, Hill (1969); 5, Broyles and others (1979); 6, Zucca and Hill (1980); 7, Zucca and others (1982); 8A–8B, Pollard and Eaton (1964). Dashed red lines, major rift zones. Elevation and depth contours in meters; contour interval 500 m on land, 1,000 m offshore.

with some thoughts on the significance of these structures for the processes governing the growth of Kilauea and Mauna Loa.

### ACKNOWLEDGMENT

We are grateful to Alan Ryall for providing the traveltime data from the Hilo and Kaloe shot points.

## SEISMIC-REFRACTION PROFILES

The locations of seismic-refraction profiles on and adjacent to the Island of Hawaii are shown in figure 37.1. With the exception of some of the short profiles across Kilauea's east rift zone done by the University of Hawaii (Broyles and others, 1979), all of the profiles were established using offshore shot points and onshore recording arrays. The permanent stations of the Hawaiian Volcano Observatory (HVO) seismic network form a common recording array for all of the profiles, although the number of stations in this array increased for successively later experiments as the HVO network grew with time (see Klein and Koyanagi, 1980). Additional portable stations were used to record the shots at either end of the profile across the summit of Kilauea (Ryall and Bennett, 1968), the series of shots forming the three coastal profiles (Hill, 1969), and the

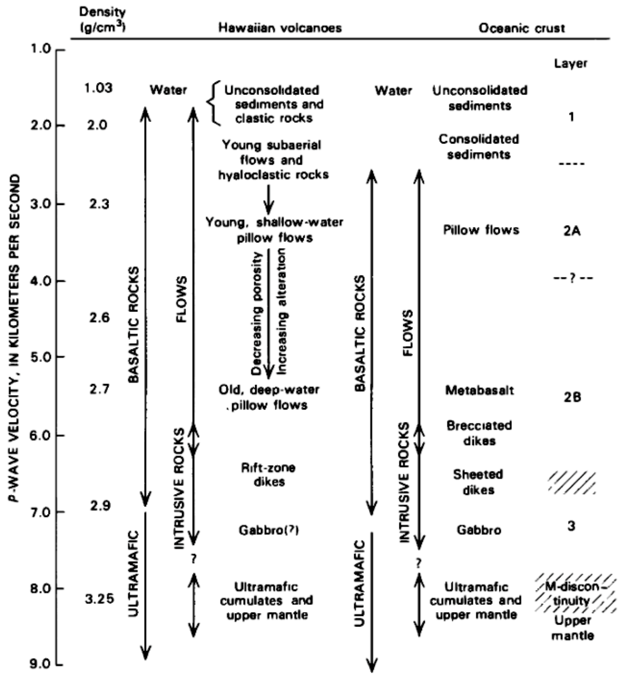


FIGURE 37.2.—Relation between $P$-wave velocity, rock density, and rock type for Hawaiian volcanoes and for oceanic crust. Density is used to compute gravity profiles in figures 37.7 and 37.8 (see Zucca and others, 1982). Moho is the M-discontinuity.

Mauna Loa offshore profile (Zucca and others, 1982). The series of shots perpendicular to the Kau-Puna coast (fig. 37.1, profile 6) were also recorded by two ocean-bottom seismographs deployed along the profile by the University of Hawaii (see Zucca and others, 1979).

These approximately linear seismic-refraction profiles are intrinsically two-dimensional. The internal structure of the volcanoes, however, is dominated by a roughly conical symmetry about the central vents. Profiles that cut obliquely across this dominant structure will be complicated by lateral (out-of-plane) reflections and refractions. Distortion in the interpretation from such out-of-plane ray paths will be more of a problem for coastal profiles than for profiles oriented radial to the central vents.

$P$-wave velocity structure serves as an important guide to subsurface rock types and geologic structure. The association of $P$-wave velocities with the rock types that we used in this paper for Kilauea and Mauna Loa Volcanoes and for the adjacent Cretaceous oceanic crust is illustrated in figure 37.2. This association is based on (1) published values of seismic velocities for rocks compiled by Christensen (1982), (2) petrographic models of the oceanic crust as described by Salisbury and Christensen (1978), and (3) recognition that the Hawaiian volcanoes consist almost entirely of basaltic rocks. Note in particular that $P$-wave velocities ranging from less than 2.0 km/s to more than 6.5 km/s on the Hawaiian volcanoes primarily reflect the wide variation in bulk porosity (bulk density) in rocks having essentially the same basaltic composition. The high bulk porosity in young, subaerial lava flows is produced by clinkery flow tops and bottoms, extensive sets of open fractures, large gas bubbles, and lava tubes; in older, deep-water pillow flows the porosity is reduced by alteration products that largely fill vesicles and cracks, and in basaltic dikes it is limited to small vesicles and intercrystalline cracks with dimensions of micrometers or less.

## VOLCANIC FLANKS AND THE OCEANIC CRUST

The two offshore profiles that extend radially away from the volcanic centers (fig. 37.1, profiles 6, 7) define the base of the oceanic crust (the $M$-discontinuity) from the open ocean to the subaerial flanks of Kilauea and Mauna Loa. Interpretations of these profiles by Zucca and Hill (1980) and Zucca and others (1982), illustrated in figures 37.3 and 37.4, indicate that, beginning 50–60 km from shore, the oceanic crust dips at about 3° beneath the submarine flanks of the volcanoes. The depth to the $M$-discontinuity increases from 10–11 km beneath the unperturbed oceanic crust at distances greater than 50 km from the island to about 13 km under the southeast coast of Kilauea on the Island of Hawaii and 14–15 km under the west coast of Mauna Loa. Zucca and others (1982) inferred that the dip on the base of the crust increases toward the center of the island and is nearly 9° under the west (Kona) coast of Mauna Loa.

Structural details of the unperturbed oceanic crust beyond the submarine flank of Hawaii remain poorly resolved. The Mauna Loa radial profile does not have reversed seismic-refraction coverage, nor does the seaward half of the Kilauea radial profile. The structure for the oceanic crust shown in figures 37.3 and 37.4 is simply a seaward horizontal extension of the structure found for the oceanic crust beneath the distal end of the volcanic wedge in each case. For the Kilauea profile, this structure is based on reversed and overlapping

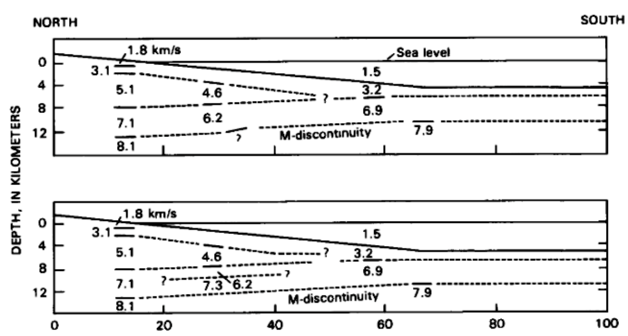


FIGURE 37.3.—$P$-wave velocity structure for the offshore profile perpendicular to the Kau-Puna coast of Kilauea (profile 6 in fig. 37.1). Numbers on profiles indicate $P$-wave velocity in kilometers per second. Upper and lower profiles show alternate interpretations of the traveltime data. Heavy lines and velocities at left (beneath the shoreline) indicate structure along the Kau-Puna coast from Hill (1969). Adopted from Zucca and Hill (1980).

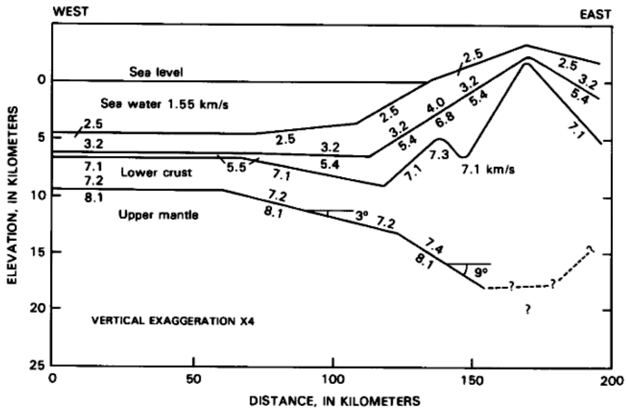


FIGURE 37.4.—*P*-wave velocity structure for the offshore profile perpendicular to the Kona coast of Mauna Loa (profile 7 in fig. 37.1). Inferred dip angle for base of crust shown for two parts of profile; other numbers on profile indicate *P*-wave velocity in kilometers per second. Adopted from Zucca and others (1982).

refraction coverage over the landward half of the profile (Zucca and Hill, 1982); for the Mauna Loa profile, however, it is based only on the overlapping (but unreversed) coverage provided by the array of onshore stations (Zucca and others, 1982). On the basis of the data from these profiles, we can say little more than the following: (1) The oceanic crust south and west of Hawaii is 5–6 km thick, with that west of Mauna Loa being slightly thinner than that southeast of Kilauea; (2) the *P*-wave velocity at the top of layer 2 is approximately 3.0 km/s, and layer 2 appears to be somewhat thicker west of Mauna Loa than it is southeast of Kilauea; and (3) the *P*-wave velocity in layer 3 (the deepest and thickest layer) is approximately 7.0 km/s. These data provide no direct information on the relatively thin sedimentary layer (layer 1) that covers the oceanic crust in the vicinity of the Hawaiian Islands (see Watts and others, 1985).

Our estimate of total thickness for the oceanic crust illustrated in figures 37.3 and 37.4 may be slightly low because the first arrival *P*-wave data did not sample the lower parts of layer 2. In most sections of the oceanic crust, the *P*-wave velocity in layer 2 increases rapidly with depth and averages about 5.6 km/s (see figure 37.2). If the average *P*-wave velocity in layer 2 in the vicinity of Hawaii is closer to 5.6 km/s than the 3.0–3.2 km/s value indicated by our first arrival data, we have underestimated the total crustal thickness by 1.0–1.4 km. In this case, the thickness of the oceanic crust off Hawaii is comparable to the 6.5-km thickness off of Oahu reported by Watts and others (1985).

The 7.9-km/s *P*-wave velocity for the uppermost mantle ($P_n$ velocity) derived from the offshore Kilauea profile (fig. 37.3) is slightly lower than the 8.0- to 8.4-km/s values determined elsewhere beneath the Island of Hawaii and along the Hawaii-Emperor chain (Furumoto and others, 1968; Watts and others, 1985). Because the 7.9-km/s value is based both on reversed traveltime branches over the submarine flank of Kilauea and on clear $P_n$ arrivals over the presumably flat oceanic crust beyond, we feel it is probably reliable. It is also consistent with evidence for anomalously low (by 1–2 percent) *P*-wave velocity in the top 35 km of the upper mantle off the southeast coast of Hawaii found by Ellsworth and Koyanagi (1977) in their three-dimensional inversion of teleseismic *P*-wave residuals recorded on Hawaii.

The wedge of volcanic products that forms the submarine flank of the volcanoes consists of an inner wedge having a *P*-wave velocity ranging from 4.6–5.4 km/s, which rests directly on the Cretaceous oceanic crust. Overlying this inner wedge is a blanket of material 1–3 km thick having a *P*-wave velocity of 3.0–3.2 km/s (see figs. 37.3, 37.4). This *P*-wave velocity structure is consistent with the model proposed by Moore and Fiske (1969) for the substructure of the volcano flanks from their study of dredge samples and ocean-bottom photographs. From the relation between *P*-wave velocities and rock types suggested in figure 37.2 and from the Moore-Fiske model, we infer that the inner wedge consists dominantly of deep-water pillow basalt and the overlying blanket consists dominantly of hyaloclastic rocks (vitric explosion debris, littoral-cone ash, and flow breccia). The shallower part of the 2.8–3.2-km/s layer probably also include young pillow flows and subaerial flows that have been submerged as the crust subsided in response to the growing load of the volcanoes.

Because we cannot distinguish the pillow flows and metabasalt forming layer 2 of the Cretaceous oceanic crust from the pillow flows and clastic rock that form the submarine flanks of the volcanoes on the basis of *P*-wave velocity (fig. 37.2), the position of the distal edge of volcanic debris from Kilauea and Mauna Loa remains ambiguous. The same is true for the position of the boundary between the bottom of the inner wedge of pillow lava and the top of layer 2. Moore (1964) has demonstrated that downslope movement in the form of massive submarine landslides is an important process on the steep submarine flanks of the Hawaiian volcanoes. This raises the possibility that the exceptional thickness of the 3.0-km/s layer in the oceanic crust off the coast of Mauna Loa (fig. 37.4) is due to accumulations of landslide debris extending well beyond the obvious break in slope at the distal end of the volcanic wedge.

The sections of the coastal seismic-refraction profiles interpreted by Hill (1969) that are not adjacent to major rift zones generally support the structural model for the flanks of the volcanoes based on the radial profiles just described (see fig. 37.5). Differences in structure between the coastal and radial profiles near their junction along the central section of the Kau-Puna coast are minor (figs. 37.3, 37.5).

Agreement between layer depths and *P*-wave velocities is not as close, however, for the coastal and radial profiles near their junction along the Kona coast on the west flank of Mauna Loa (see figs. 37.4, 37.5). The problem here probably lies with out-of-plane propagation paths complicating the two-dimensional interpretation of the coastal profile. This profile runs sub-parallel to a pronounced elongate gravity high located just inland of the Kona coast (see fig. 37.5), and it is likely that high *P*-wave velocity in the shallow crust associated with this gravity high provides a lateral, least-time path for seismic waves propagating between sources and receivers along the coast. Accordingly, we are inclined to place more confidence in the structure beneath the coast based on data from the radial profile

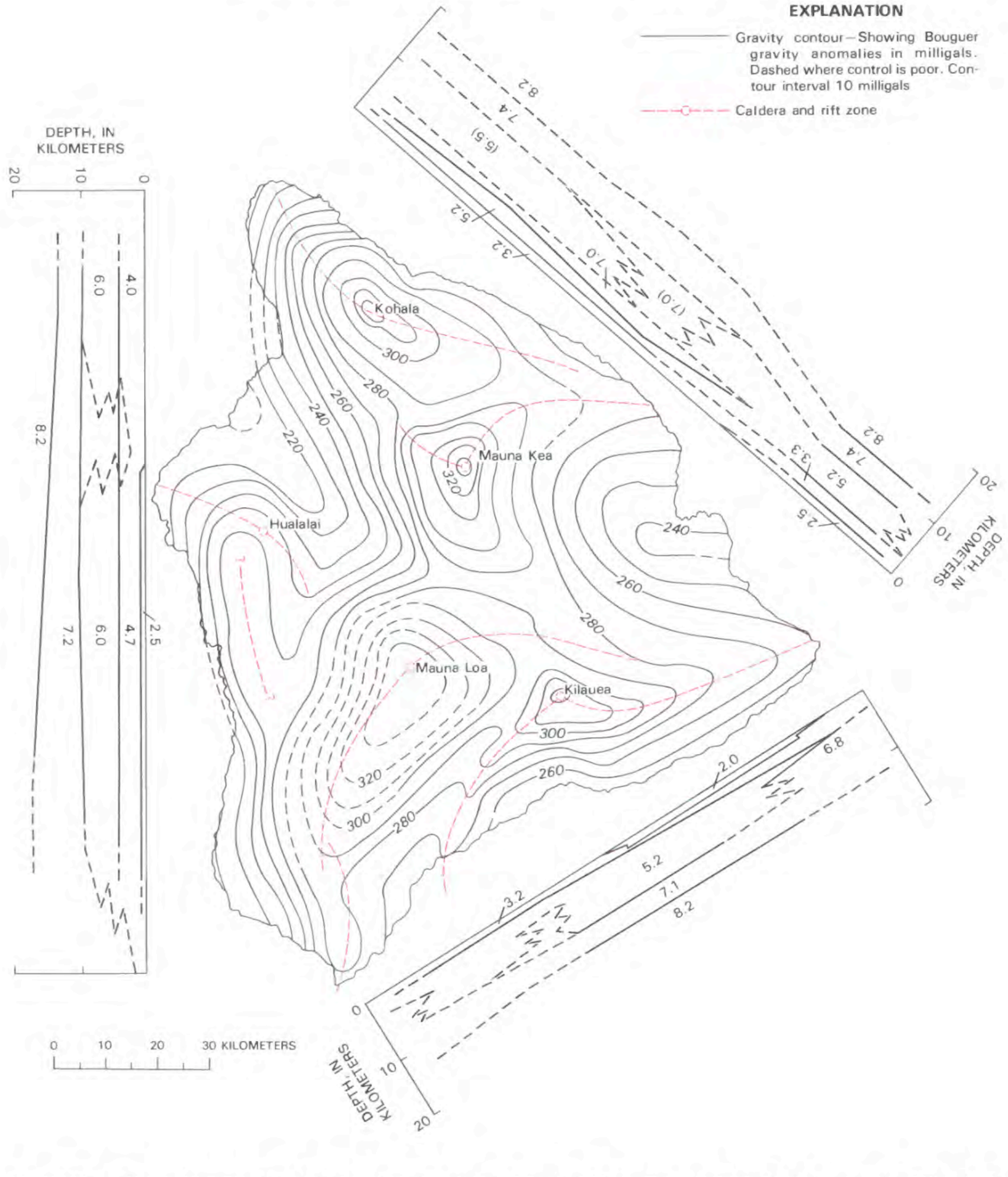


FIGURE 37.5.—Bouguer gravity map of Island of Hawaii (from Kinoshita and others, 1965) and *P*-wave velocity structure for the three coastal profiles (profiles 2–4 in fig. 37.1) around the island (adopted from Hill, 1969). Gravity contours dashed where control is poor; contour interval 10 milligals. Dashed red lines, major rift zones of the volcanoes.

(fig. 37.4) than in that based on the profile along the southern section of the Kona coast (fig. 37.5). In particular, we feel that the crustal thickness directly beneath the Kona coast west of Mauna Loa is closer to the 14 km indicated in figure 37.4 than to the 18 km indicated in figure 37.5 and that the layer depths and $P$-wave velocities for the southern section of the Kona profile in figure 37.5 give a poor image of the structure beneath the coast.

Results from profiles along the Hamakua-Puna coast (Hill, 1969) and from the Hilo-Kalae profile (Ryall and Bennett, 1968) indicate that the base of the crust deepens from about 13 km beneath Hilo to more than 15 km midway between Hilo and Kilauea and then appears to shallow again toward the summit region of Kilauea (figs. 37.1, 37.6). More than 10 km of effusive rocks with an average $P$-wave velocity of 5.3 km/s fill the resulting basin-like depression of the oceanic crust. This thick accumulation of rather porous flows is associated with a distinct gravity low centered just south of Hilo (fig. 37.5). Some of the deeper lava flows in this basin may be from Mauna Kea, but the bulk of the flows were probably fed by vents on Mauna Loa and, to a lesser degree, vents along the east rift zone of Kilauea. The increase in $P$-wave velocity from 3.3 to 5.2 km/s at about 2 km beneath the subaerial surface probably reflects reduced porosity due to alteration and compression; it is not necessarily the boundary between subaerial and pillow flows. Near intrusive cores, in particular, low-grade metamorphism between widely spaced dikes may markedly reduce the bulk porosity (and increase the $P$-wave velocity) in a mass dominated by hyaloclastic rocks and subaerial flows.

## RIFT ZONES

The major rift zones of Kilauea and Mauna Loa are marked by elongate ridges sloping away from the respective summits and by strong, elongate gravity highs closely coincident with the topographic ridges (see figs. 37.1, 37.5). Topographic expression of the rift zones on the older volcanoes of the Island of Hawaii (Mauna Kea, Hualalai, and Kohala) tends to be masked by lava flows from scattered vents that developed in the later evolutionary stages of the volcanoes, but the expression of these older rift zones in the gravity field remains clear. The strong, elongate gravity high extending southward from Hualalai subparallel to the Kona coast, for example, probably defines the location of an ancient rift zone (from Hualalai?) now buried by lava flows from Mauna Loa's actively growing west flank.

As is evident from figures 37.1 and 37.5, seismic-refraction data on the substructure of the rift zones are limited to profiles that either terminate in the vicinity of a rift zone or run subparallel to the axial trend of a rift zone. Taken together, these data provide a patchwork image for the rift-zone substructure. That the rift zones are underlain by rocks with high $P$-wave velocities is indicated by the uniformly early arrival times for $P$-waves propagating along ray paths that penetrate the rift zones. Apparent velocities for first-arrival traveltime curves on profiles that closely parallel rift zones, such as those along the eastern section of the Kau-Puna coast or the northern half of the Hamakua coast (fig. 37.1), are typically 6.5–7.3 km/s (see Hill, 1969, figs. 3, 5, 10). The waves propagating with these high apparent velocities have almost certainly followed laterally refracted ray paths. Nevertheless, reversed traveltime branches limit depths to the high-velocity rocks to no more than 2–5 km. Because of lateral refraction, the apparent depth to the high-velocity rift-zone rocks plotted in the cross sections in figure 37.5 should be regarded as a perpendicular distance (slant depth) between the average surface trace of the seismic-refraction profile and the subsurface flank of the adjacent rift zone. The series of short seismic-refraction profiles across Kilauea's lower east rift zone established by Broyles and others (1979) provides clear evidence that rocks with high $P$-wave velocities beneath the rift zones are at shallow depths. Their results indicate that, along the axis of the rift zone, rocks with $P$-wave velocites of 7.0 km/s lie at depths of only 2.1–2.3 km beneath recent lava flows with $P$-wave velocities averaging about 3.0 km/s.

Available seismic-refraction data provide only loose constraints on the deeper structure of rift zones. In general, it appears that the mass of intrusive rocks forming the high-velocity core of a rift zone broadens with both depth and proximity to the central vent of the volcano. Zucca and Hill (1980), for example, have used the pattern of $P_n$ delays recorded on selected stations of the HVO seismic network from the line of offshore shots southeast of Kilauea, together with average teleseismic residuals at the same stations reported by Ellsworth (1977), in an effort to put constraints on the deeper structure of the rift zones beneath Kilauea. The pattern of $P_n$ and teleseismic delays along a line of stations crossing the central section of the east rift zone indicates that the high-velocity core increases in breadth from about 2 km near the surface to perhaps 13–14 km at depths of 8–9 km, where it merges with the oceanic crust. Based on these sparse data, the sides of the high-velocity rift-zone core have an effective slope of 45°–50° (see Zucca and Hill, 1980, fig. 8). Results from a similar profile crossing the southwest rift zone and extending to the summit of Mauna Loa indicate that there the breadth of the high-velocity core increases more rapidly with depth (effective slope 30°–40°) and the rift-zone core merges with the massive high-velocity core associated with the central vent of Mauna Loa at a depth of 5 km or less (see Zucca and Hill, 1980, fig. 9).

The Bouguer gravity data for the Island of Hawaii (fig. 37.5) also suggest that the dense cores beneath the rift zones broaden with depth (Kinoshita and others, 1963; Broyles and others, 1979). Dense, closely spaced gravity measurements, however, currently stop at the shoreline, where both the topographic and gravity gradients are still quite steep. Until the gravity field is better defined offshore, beyond the steep gradients and the submarine flank of the volcanoes, it is hardly worth attempting to model the gravity field in terms of detailed three-dimensional density structure of the intrusive cores of the volcanoes.

## SUMMIT REGIONS AND CENTRAL CORES

The summits of both Mauna Loa and Kilauea contain large calderas, within which most summit eruptions occur. An elongate conduit system connects the summit vents and subjacent magma chamber located within the edifice of each volcano to magma sources in the upper mantle at depths of 40–60 km (Eaton and Murata,

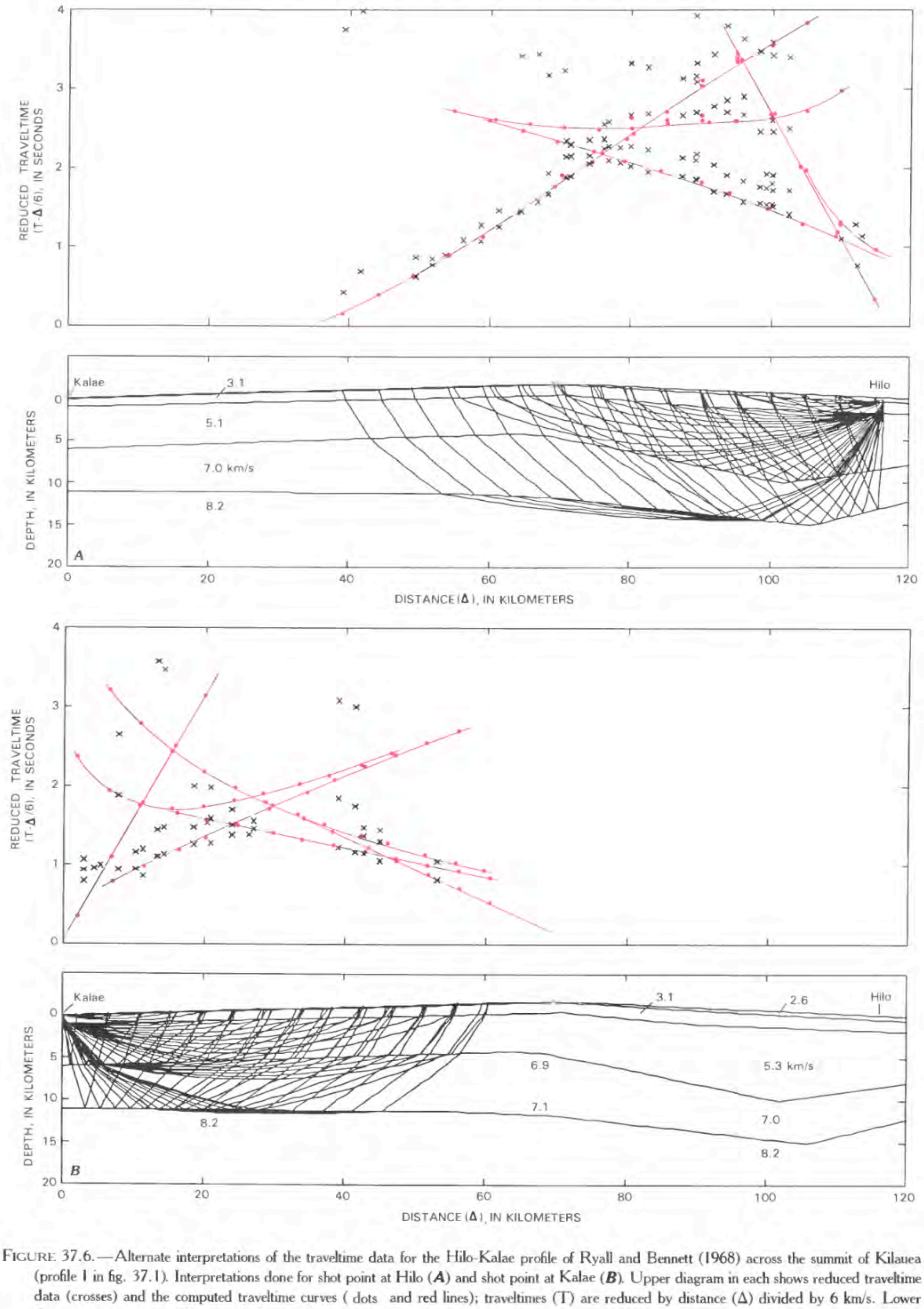


FIGURE 37.6.—Alternate interpretations of the traveltime data for the Hilo-Kalae profile of Ryall and Bennett (1968) across the summit of Kilauea (profile 1 in fig. 37.1). Interpretations done for shot point at Hilo (*A*) and shot point at Kalae (*B*). Upper diagram in each shows reduced traveltime data (crosses) and the computed traveltime curves ( dots and red lines); traveltimes (T) are reduced by distance (Δ) divided by 6 km/s. Lower diagram in each shows *P*-wave velocity model and ray paths.

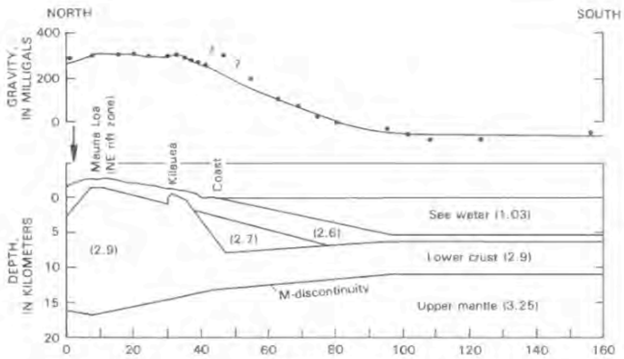


FIGURE 37.7.—Gravity profile and density structure along the offshore Kilauea profile (profile 6 in fig. 37.1) Dots, measured gravity values; gravity curve is that computed from the density model. Figures indicate density of layers in model, in grams per cubic centimeter. Arrow indicates intersection with the profile in figure 37.8. Adapted from Zucca and others (1982).

1960; Ryan and others, 1983). This conduit system forms a vertical axis about which the principal mass of each volcano has accumulated. A core of dense intrusive rock surrounds the conduit system and magma chamber within the summit edifice, producing the positive gravity anomalies over the summit area of each of the volcanoes (compare the topography and gravity fields in figs. 37.1 and 37.5).

The profile established by Ryall and Bennett (1968) between Kalae and Hilo passes immediately northwest of the summit of Kilauea Volcano (fig. 37.1). Although the opposing traveltime curves from the two shotpoints are not reversed (see fig. 37.6), they provide important clues on the deep crustal structure beneath the summit region of Kilauea.

Ryall and Bennett (1968) interpreted the Hilo-Kalae profile in terms of a major northwest-striking, northeast-dipping normal fault beneath the summit of Kilauea that offsets the oceanic crust by nearly 7.5 km and brings the base of the crust to within 10.3 km of the surface beneath the summit (Ryall and Bennett, 1968, fig. 7). We find that the traveltime data can also be fit rather well by a more continuous velocity structure that does not involve offset across a major crustal fault (fig. 37.6). In this structure, we interpret the thickening of a layer of velocity 6.9–7.1 km/s toward the summit region as due to intrusive cores of the southwest rift zone and the central conduit of Kilauea that lie just southeast of the profile where it crosses the summit. An intrusive core associated with the buried end of the Mauna Loa east rift zone may also contribute to the increased thickness of this layer northeast of Kilauea (see figs. 37.1, 37.5).

The profile extending northwest from Kalae runs subparallel to and about 5 km inland from the southeastern half of the Kau-Puna coastal profile (fig. 37.1). Both profiles show a similar structure for the 11 to 13-km-thick crust beneath the coastal flank of Mauna Loa between Kalae and the point where Kilauea's southwest rift zone intersects the coast (figs. 37.5, 37.6). The unreversed traveltime data from the Kalae profile, however, give no indication that the crust thickens significantly toward the summit of Kilauea, a result seemingly at odds with the evidence from the perpendicular profile (fig. 37.3) that the base of the crust dips at about 3° beneath the flank of Kilauea toward the summit region. Whether this apparent discrepancy means that the dip on the base of the crust flattens under the subaerial flank of Kilauea, or, as seems more likely, it reflects the level of uncertainty associated with the interpretation of these unreversed traveltime curves remains to be seen. In any case, the opposing traveltime data along the Hilo-Kalae profile indicate upper mantle $P$-wave velocities at fairly shallow depths (10–11 km below sea level) beneath the summit region of Kilauea. Thurber (chapter 38) finds independent evidence for unusually shallow depths of rocks with 8.0-km/s velocity beneath the summit of Kilauea in his inversion of local earthquake arrival-time data.

Constraints on the structure of the intrusive core beneath Mauna Loa derive from the onshore extensions of the two offshore radial seismic-refraction profiles (figs. 37.1, 37.3, 37.4). The $P$-wave velocity structure in figure 37.4 summarizes results obtained by Zucca and others (1982) from their inversion of the overlapping record sections from each of the nine stations forming the onshore section of the Mauna Loa profile passing just north of the summit. These results indicate that the mass of rocks with $P$-wave velocities of 7.0–7.3 km/s thickens dramatically toward the summit of Mauna Loa, beginning about 10 km west of the present coastline. The thickening occurs as a combination of an increased dip of the $M$-discontinuity and a decreasing depth to rocks with 7.0-km/s velocity toward the summit.

The fairly narrow zone of elevated $P$-wave velocity beneath the coast is consistent with but not required by the seismic-refraction data. A comparable zone of elevated density is required, however, by the elongate gravity high that extends southward along the coast from Hualalai (see fig. 37.5; Zucca and others, 1982). As mentioned in the previous section, this elongate zone of density and velocity probably corresponds to an ancient rift zone buried by the actively growing flank of Mauna Loa.

Unfortunately, none of the record sections on the Mauna Loa profile include waves that bottomed in the lower crust under the summit of Mauna Loa, and we are left with no information on the structure of the lower crust beneath the summit from these seismic-refraction data.

Further constraints on structural models for the Mauna Loa and Kilauea seismic-refraction profiles come from coincident gravity profiles, which Zucca and others (1982) constructed by joining the onshore (complete) Bouguer gravity data of Kinoshita and others (1963) with offshore free-air gravity data compiled by Watts and Talwani (1975). These profiles and their interpretation in terms of the density structure are reproduced in figures 37.7 and 37.8. Zucca and others (1982) modeled the density structure by assuming the relations between velocity and density summarized in figure 37.2 and modifying the layer configurations in the $P$-wave velocity models (figs. 37.3, 37.4) to obtain an acceptable fit between the computed and observed gravity profiles. They used a two-dimensional algorithm for computing the gravity field modified to approximate

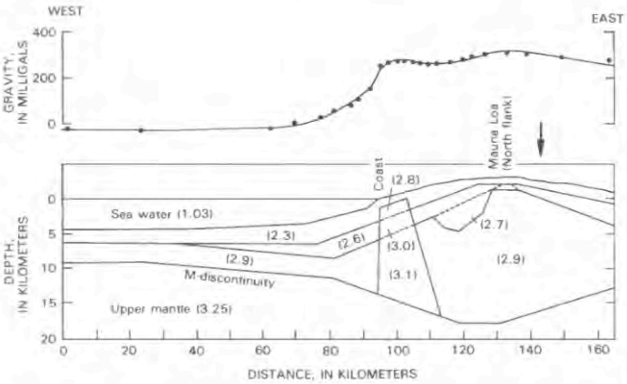


FIGURE 37.8.—Gravity profile and density structure along the offshore Mauna Loa profile (profile 7 in fig. 37.1). Dots, measured gravity values; gravity curve is that computed from the density model. Figures indicate density of layers in model, in grams per cubic centimeter. Arrow indicates intersection with profile in figure 37.7. Adapted from Zucca and others (1982).

the three-dimensional structure of the volcanoes by incorporating an approximate end correction developed by Cady (1977).

The density models in both profiles (figs. 37.7, 37.8) fit the gravity reasonably well, and the boundaries in the density models correspond closely to those in the $P$-wave velocity models of figures 37.3 and 37.4. Taken together, these velocity and density models indicate that the crust beneath Mauna Loa thickens to about 18 km toward the summit region and that the core of dense, high-velocity (intrusive) rocks is the dominant mass (and volume) forming Mauna Loa's central edifice.

## DISCUSSION

### GEOLOGIC SECTIONS

The cross sections in figures 37.9, 37.10, and 37.11 summarize our interpretation of the $P$-wave velocity structure and density structure of Mauna Loa and Kilauea in terms of geologic structures. Two possible geologic interpretations of the velocity structure along the profile crossing the summit of Kilauea from Kalae to Hilo are illustrated in figure 37.9$A$, $B$; a third possibility was shown in Ryall and Bennett (1968, fig. 7). These cross sections are drawn without vertical exaggeration, and the rock types illustrated are keyed to the $P$-wave velocity structure along corresponding seismic-refraction profiles by means of the relations indicated in figure 37.2.

The volcanoes are draped with a surficial blanket as much as 3 km thick that consists of young, subaerial basalt flows, shallow submarine flows, and clastic debris. As described by Moore and Fiske (1969), the submarine portions of this blanket (away from the submarine extensions of rift zones along which pillow flows predominate) consist mostly of clastic debris produced near sea level by the shattering of rapidly quenched lava flows and eruption of ash at littoral cones. Downslope movement in the form of massive submarine landslides carries this clastic debris to the distal ends of the submarine flank and onto the adjacent sections of the Cretaceous oceanic crust (Moore, 1964).

Older submarine pillow flows constitute the dominant effusive product of the volcanoes. These flows were erupted at depths of 100–500 m or more (Moore and Fiske, 1969) and have further submerged with the deflection of the lithosphere in response to the growing load of the volcanoes. Pillow flows probably make up the dominant fraction of the exceptionally thick sequence of effusive rocks in the basin-like depression between the east rift zones of Mauna Loa and Kilauea to the south and the east rift zone of Mauna Kea to the north (fig. 37.9).

The cores of the rift zones and summits of the volcanoes consist of a sequence of densely packed dikes analogous to the sheeted-dike complex forming the uppermost part of layer 3 in the oceanic crust (Salisbury and Christensen, 1978). These intrusive cores, which reach to within 2 km of the surface beneath the summit calderas and rift-zone axes, broaden with depth in a shape that is an exaggerated mimic of the overlying surface topography. The outer boundaries of the intrusive cores are no doubt transitional between altered effusive rocks with widely spaced dikes to a nearly solid mass of intrusive rocks. Walker (chapter 41) finds that, where the dense Koolau dike swarm is exposed on Oahu, most of this transition occurs within a distance of about 1 km. The wave lengths of the $P$-waves refracted by the intrusive rift-zone cores are 0.5–1.0 km for most of our data, and a 1-km-thick transition zone would appear as a relatively sharp boundary to these waves. At the other extreme, a poorly developed intrusive core such as that beneath Kilauea's southwest rift zone may be transitional across its entire width. In any case, though our sampling of the subsurface structure of these intrusive cores is still incomplete, it seems clear that they form an important fraction of the volume of Mauna Loa and Kilauea Volcanoes.

The orientation of the dikes within the intrusive cores reflects the direction of greatest horizontal principal stress at the time of their intrusion (Nakamura, 1977). Within the rift-zone cores, this direction is parallel or subparallel to the rift-zone axis (Fiske and Jackson, 1972). One might expect a more complex configuration of dikes within the summit cores. In the case of Kilauea, it appears as though the complexity in the orientation of more recently emplaced dikes is limited to a relatively small volume surrounding the current position of the central magma chamber beneath the southern margin of the summit caldera. Data from earthquake focal mechanisms (Endo, 1977; Thurber, chapter 38) and the orientation of fissures from recent eruptions (Holcomb, chapter 12) indicate that the stress field of the east rift zone (characterized by an east- to east-northeast-trending greatest horizontal principal stress axis) dominates both the northwest-trending upper east rift zone (see Hill, 1977) and much of the summit caldera. These same data indicate that the stress field of the southwest rift zone (characterized by a northwest trending greatest horizontal principal stress axis) extends only a short distance into the southern part of the summit caldera. The rather abrupt 30°–40° change in the orientation of the greatest horizontal principal stress where the stress fields from the two rift zones intersect coincides closely with the center of inflation for recent eruptive cycles and the current position of the central magma chamber (see Johnson,

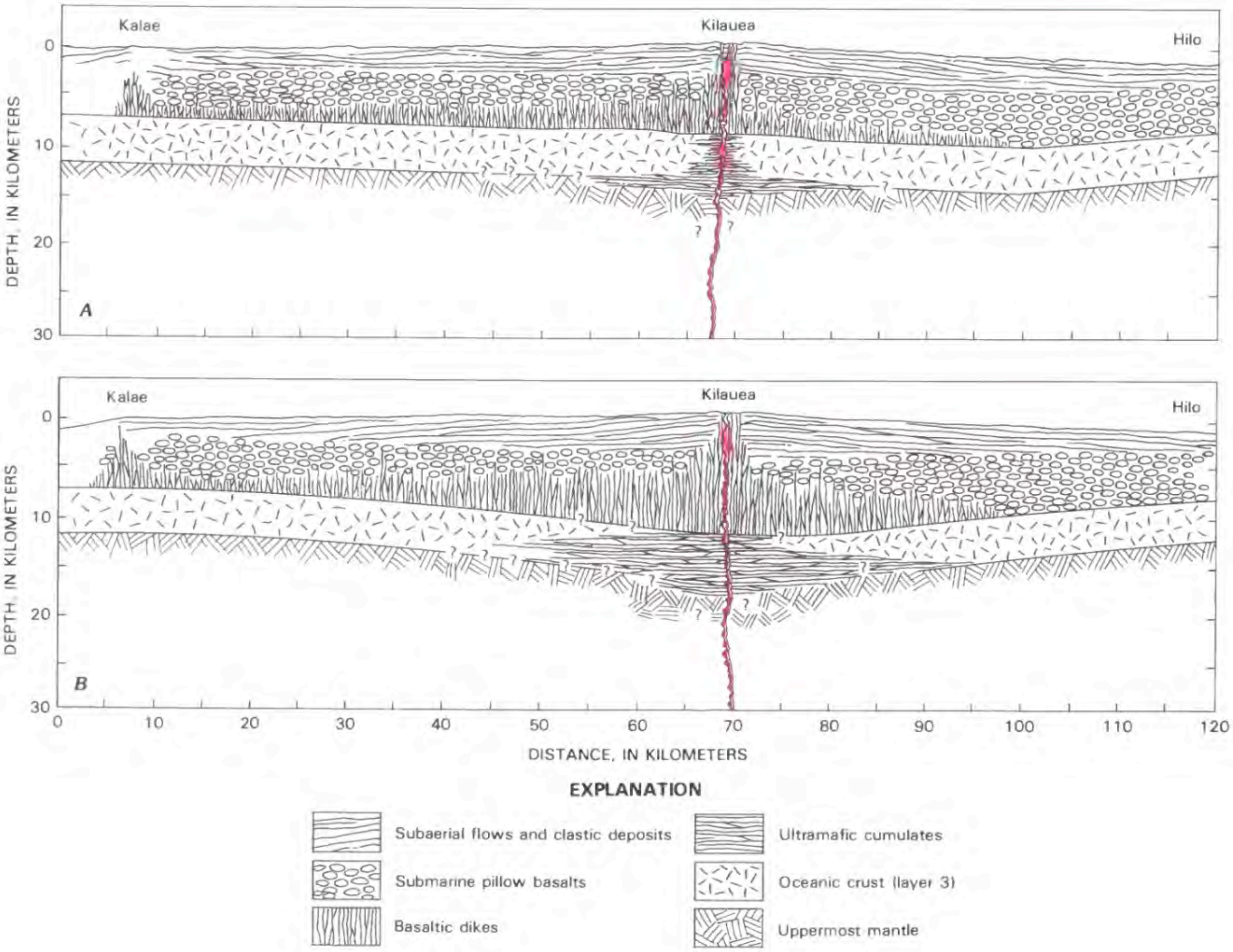


FIGURE 37.9.—Schematic cross sections showing two interpretations of the geologic structure along the Hilo-Kalae profile (profile I in fig. 37.1) based on *P*-wave velocity structure (see figs. 37.3, 37.6, 37.7). Main magma conduit and summit magma chamber indicated in red. The seismic profiles do not pass directly over the summit, and they provide no evidence for the small intrusive cupola included beneath the summit in these sections. ***A***, Interpretation involving little crustal depression and taking high-velocity material at depth beneath Kilauea to be the upper mantle. ***B***, Interpretation involving greater crustal depression and taking high-velocity material at depth beneath Kilauea to be ultramafic cumulate.

chapter 47). The stress field in the summit region may, of course, have had different orientations during earlier stages in the growth of the volcano, and the orientation of older (and deeper) dikes may therefore differ from that of dikes intruded under the influence of the present stress field.

Though dikes appear to be the dominant intrusive form in at least the outer portions of the intrusive cores, sills and somewhat larger, tabular bodies of gabbro are probably also locally important (see, for example, Ryan and others, 1983, p. 1452). Indeed, *P*-wave velocities near 7 km/s strongly suggest the presence of somewhat more mafic rocks such as pyroxene gabbro or olivine gabbro in the intrusive cores; *P*-wave velocities in basaltic (tholeiitic) dikes or sills seldom exceed 6.5 km/s under confining pressures lower than 200–300 MPa (Christensen, 1982). That isolated pockets of magma would be left in the rift zones to form gabbroic bodies seems plausible. The active intrusion of sills is limited to shallow depths in the volcanic edifice, where the overburden pressure forms a vertical least principal stress. It seems likely, for example, that the base of the intrusive cores may consist of a sequence of sills that were intruded below a fairly thin veneer of early pillow basalt on the oceanic crust.

The ultramafic cumulate forming a high-density, high-velocity inner core to the intrusive core of the volcanoes is the most speculative aspect of the structure suggested in figures 37.9 and 37.11. Its existence is consistent with but not required by the seismic-refraction data. Evidence from ultramafic inclusions in erupted basalt, however, indicates that early magma chambers must

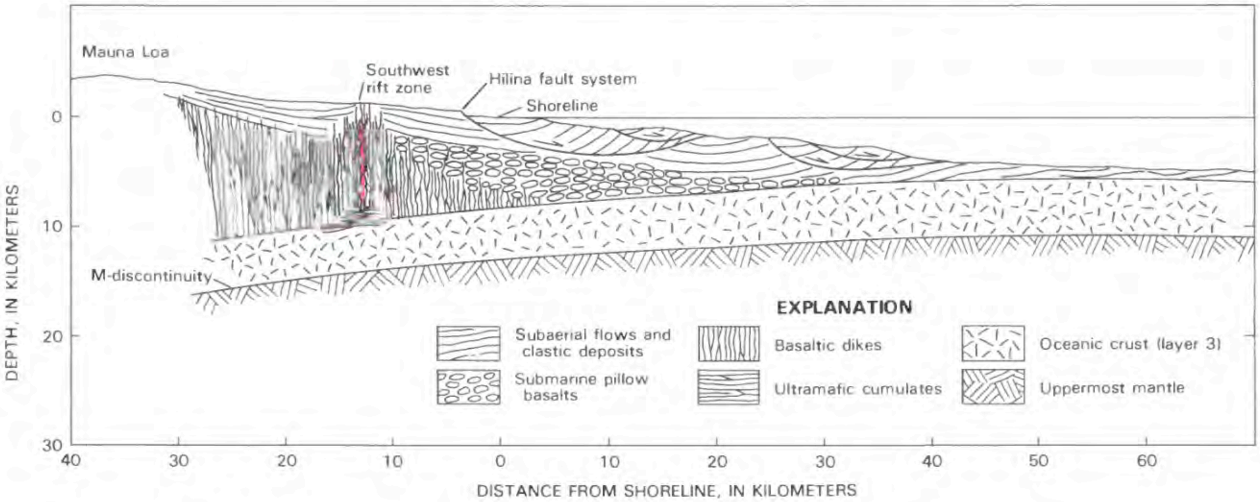


FIGURE 37.10.—Schematic cross section showing inferred geologic structure along the offshore Kilauea profile (profile 6 in fig. 37.1) based on *P*-wave velocity and density models in figures 37.3 and 37.7, respectively. Main magma conduit indicated in red.

have left a trail of residual ultramafic deposits as they migrated upward through the oceanic crust and volcanic pile during the construction of the submarine volcanic edifice (Jackson, 1968). The ultramafic cumulates shown in figure 37.9*A* are limited to a small volume around the central conduit. In this case, the base of the oceanic crust is only about 13 km below sea level under the summit of Kilauea, implying minimal deflection of the lithosphere under the load of Kilauea. In figure 37.9*B*, the ultramafic cumulate forms an extensive mass of sill-like bodies that have disrupted the depressed part of the oceanic crust beneath the summit region of Kilauea. The ultramafic cumulate has mantle-like *P*-wave velocity and its upper surface masquerades as the *M*-discontinuity beneath the summit region. Although we prefer this interpretation (fig. 37.9*B*) because it fits current ideas on the response of the oceanic lithosphere to the loads of the volcanoes, we cannot ignore the simpler interpretation (fig. 37.9*A*) of the Hilo-Kalae velocity structure (fig. 37.6*B*).

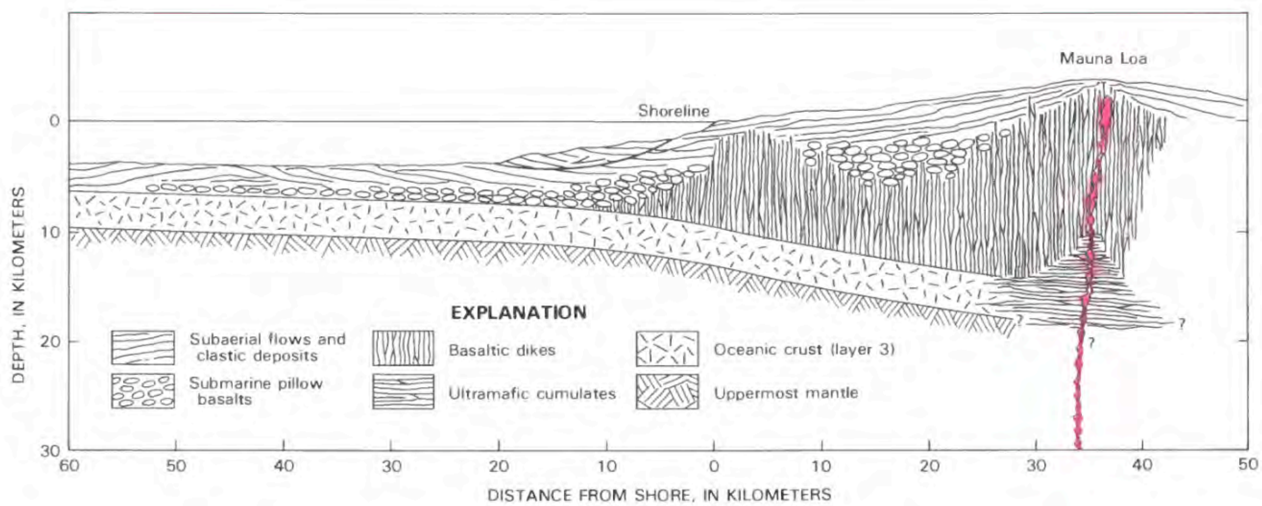


FIGURE 37.11.—Schematic cross section showing inferred geologic structure along the offshore Mauna Loa profile (profile 7 in fig. 37.1) based on *P*-wave velocity and density models in figures 37.4 and 37.8, respectively. Central magma conduit of Mauna Loa Volcano and its summit magma chamber shown in red.

It remains uncertain whether an inner core of ultramafic cumulate extending up to shallow depths under the central vents is an integral part of the Hawaiian volcanoes. The evidence described by Furumoto and others (1968) indicating protrusions of upper mantle $P$-wave velocity (7.7 km/s and above) to shallow depths beneath the center of the eroded core of Koolau Volcano on Oahu and beneath Penguin Banks (a submarine volcanic structure east of Molokai) is consistent with this possibility. Watts and others (1985) describe a body with $P$-wave velocity of 7.4–7.8 km/s as thick as 4 km that underlies the oceanic crust where their multichannel transect crosses the axis of the Hawaii-Emperor chain between Oahu and Molokai. They see no evidence in their data that this subcrustal body extends upward into the axis of the volcanic chain, but their transect does not cross the central vent of a major volcanic edifice.

The evidence for concave-downward curvature on the $M$-discontinuity and the oceanic crust under the subaerial flanks of Mauna Loa supports Walcott's (1970) model for the flexural response of the lithosphere to the load of the volcanoes. Watts and others (1985) find a similar result along the transect between Oahu and Molokai. The evidence for this model at Kilauea, however, is equivocal.

The 8.2-km/s $P$-wave velocity found from the $P_n$ waves propagating in the uppermost mantle under the Island of Hawaii is typical of uppermost mantle elsewhere within the Pacific plate, implying that at least the upper section of the lithosphere beneath Hawaii has undergone little heating from either the Hawaiian hot spot or the narrow conduits that supply magma to the summit reservoirs of Mauna Loa or Kilauea. Immediately southeast of Kilauea, however, and directly over the current position of the hot spot, the $P_n$ velocity is 7.9 km/s (Zucca and Hill, 1980) and the $P$-wave velocity is 8.0–8.1 km/s (1–2 percent low compared to 8.2 km/s) to depths of at least 35 km (Ellsworth and Koyanagi, 1977). Possible explanations for the anomalously low $P_n$ velocity measured on the profile southeast of Hawaii include the following: (1) Anisotropic $P$-wave propagation in uppermost mantle (Zucca and Hill, 1980) similar to that reported by Shor and others (1970) for an area northeast of Oahu; and (2) the presence of a small velocity gradient at the top of the mantle such that $P_n$ apparent velocity recorded at short distances (less than 100 km) would be somewhat lower than that recorded at greater distances (Zucca and others, 1982).

A third possiblility is that the low, lithospheric mantle $P$-wave velocity is related to the hot spot within the underlying asthenosphere. For this to be true, however, the effect must involve mechanical disruption of the upper lithosphere. Thermal conduction alone cannot supply measurable heat to the shallow parts of the lithosphere directly above the hot spot when the relative motion between the lithosphere and the hot spot is on the order of 10 cm/yr as is the case for Hawaii. Mechanical disruption of the lithosphere might be related to the southeastward progression of the flexural front associated with the growing loads of Kilauea and Loihi on the oceanic crust (see Moore, chapter 2). In principle, mechanical disruption of the lithosphere would also allow the rapid transfer of heat into the upper lithosphere by the upward migration of hot fluids from the asthenosphere. Conceivably, the low $P$-wave velocity off the southeast coast of Kilauea represents an early stage in the development of the subcrustal body with velocity of 7.4–7.8 km/s described by Watts and others (1985). Their subcrustal body, however, extends at least 100 km on either side of the ridge axis, and we see no evidence for reduced subcrustal velocity under the flanks of Mauna Loa. At this stage, we do not have adequate data to clearly discriminate between these several possibilities.

## GROWTH OF THE VOLCANIC PILE AND INTRUSIVE CORES

Moore and Fiske (1969) have outlined clearly the events in the growth of the effusive portion of a Hawaiian volcanic edifice: (1) progressive accumulation of pillow basalt flows fed by submarine vents; (2) mantling of the pile of pillow flows by hyaloclastic rocks produced when the vents reach within a few hundred meters of sea level; (3) progressive accumulation of subaerial flows as the central vents emerge above sea level, accompanied by continued production of hyaloclastic rocks where subaerial flows reach the sea; and (4) large-scale submarine slumping that transports the hyaloclastic rocks to the base of the volcanic pile and beyond.

The triangular cross section of the intrusive cores is most readily explained if the rift zones and central vents behave as spreading centers. Indeed, accumulating evidence indicates that spreading plays a central role in the growth of Hawaiian volcanoes. Swanson and others (1976) present a convincing case that the entire south flank of Kilauea Volcano has moved progressively upward and southward over the underlying oceanic crust in response to repeated episodes of forceful magma intrusion into both the summit region and east rift zone. The displacements associated with the magnitude 7.1 Kalapana earthquake of November 29, 1975, which indicate that the south flank of Kilauea moved 2–3 m southward on a low-angle plane at a depth of 8–10 km, add considerable support to this concept (Ando, 1979; Furumoto and Kovach, 1979; Nakamura, 1980).

A series of idealized structural cross sections (fig. 37.12) incorporate the concept suggested by Swanson and others (1976) together with the internal structure of rift zones summarized in figures 37.9–37.11 to illustrate sucessive stages in the growth of a rift zone. In this case the active core of the rift zone (the spreading center) moves away from the buttressed flank and pushes the unbuttressed (mobile) flank ahead of it over the oceanic crust. This process produces an intrusive core that broadens with depth as the progressively older dikes, which were injected into a progressively older and lower edifice, become further displaced from the zone of active intrusion (spreading).

If spreading within the rift-zone core were symmetric (that is, new dikes were nearly always injected midway between the next most recent series of intrusions). the mobile flank would move away from the buttressed flank at twice the rate of the active rift zone (twice the one-sided spreading rate). In the case of Kilauea, at least, it appears that active intrusion preferentially occurs on the side of the mobile flank, producing an asymmetric profile for both the rift-zone core (reflected in the gravity field) and the rift-zone topography (Swanson and others, 1976). In the absence of buttressing, the rift-zone axis would presumably remain stationary with respect to the under-

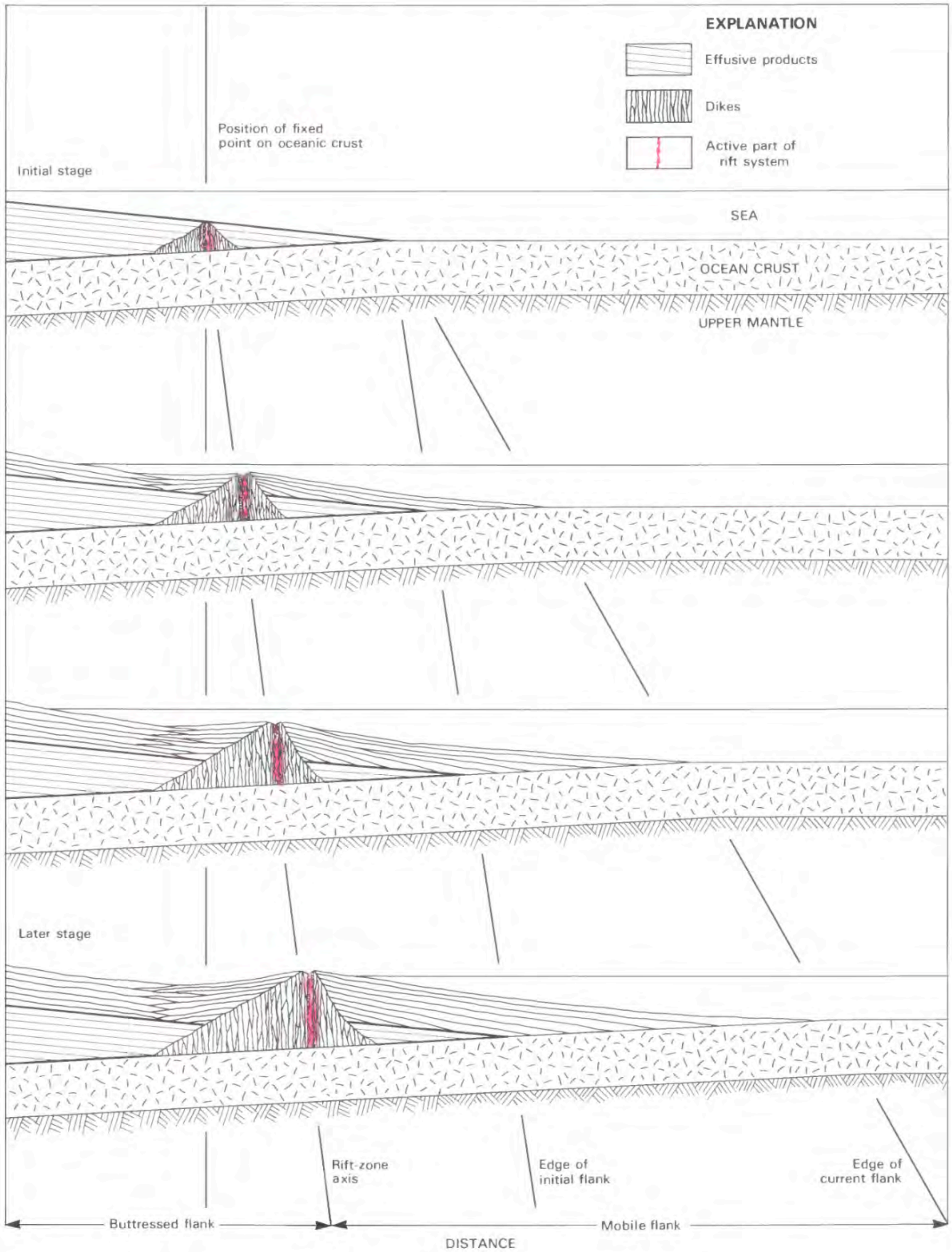


FIGURE 37.12.—Space-time diagram illustrating successive stages in the growth of a rift zone on a volcano's flank. Corresponds approximately to the east rift zone of Kilauea, buttressed on the north (here left) side by the still-growing flank of Mauna Loa. Heavy lines show successive configurations of the initial flank. Decoupling surface assumed to be at top of oceanic crust.

lying oceanic crust and both flanks would move away from the rift-zone spreading center at equal rates, resulting in symmetric topographic and core profiles.

One consequence of this process is that the axis of a rift zone with one buttressed flank will migrate with respect to the more stable position of a summit vent or the axis of a rift zone with two unbuttressed flanks. Indeed, the northward bend in Kilauea's upper east rift zone (see fig. 37.1) probably reflects the southward migration of the east rift zone with respect to the summit vent (Swanson and others, 1976). Similarly, the left-stepping offset in Mauna Loa's southwest rift zone probably reflects the westward migration of the upper part of the rift zone, which is buttressed on the east by Kilauea, with respect to the lower part, which has two unbuttressed flanks (Lipman, 1980).

The location and nature of the decoupling surface (or surfaces) along which the mobile flanks of the rift zones move under this process remains a matter of discussion. We feel that the most likely candidate for this surface is the sedimentary layer of the oceanic crust, as was suggested by Ando (1979), Furomoto and Kovach (1979), Nakamura (1980), and Zucca and Hill (1980). If this layer of abyssal ooze and clay were left more or less intact as the initial flows of pillow basalt were erupted onto the ocean floor, it would form a preexisting weak surface for later displacement between the oceanic crust and the overlying volcanic pile. Swanson and others (1976), however, suggest that decoupling occurs over a zone within the lower section of pillow basalt somewhere near the base of the volcanic pile. Fault plane solutions for the 1975 Kalapana earthquake show slip on a low-angle plane at about the proper depth for the sedimentary layer at the top of the oceanic crust beneath the southeast flank of Kilauea. Furumoto and Kovach (1979) conclude that the fault plane dips about 4° toward the island consistent with the local dip on the top of the oceanic crust. Ando (1979), however, concludes that the fault plane dips about 20° away from the island. Ando's solution is constant with a decoupling surface within the pillow basalt along the lines proposed by Swanson and others (1976). The differences between these two solutions to a large degree reflects the uncertainties involved in determining the dip on the fault plane of a large, shallow earthquake in a region with strong lateral variations in both velocity structure and topography.

We envision that the intrusive core beneath a central vent grows in much the same manner as suggested for the intrusive core within a rift zone. The principal differences are presumably that growth of the summit core would tend to develop with conical symmetry about the central conduit (in contrast to planar symmetry across the rift-zone axis) and the position of the central vent would be anchored by the deep-seated central conduit system. The angles with which rift zones intersect the summit region would no doubt strongly influence the growth pattern within the central core.

Many of the processes occurring at an actively spreading rift zone must be analogous with those occurring at the spreading center of a midocean ridge. In particular, it seems likely that the deeper parts of a rift zone might resemble a small-scale ophiolite sequence complete with pillow basalt, sheeted dikes, gabbroic bodies, and perhaps even layered ultramafic cumulate near the base of the complex. Spreading associated with the Hawaiian volcanoes, however, is a superficial process limited to the volcanic pile on top of the oceanic crust and underlying lithosphere.

## CONCLUSIONS

The seismic-refraction and gravity data reviewed in this paper provide some important clues to the structure of Kilauea and Mauna Loa Volcanoes, but critical details remain blurred. These data make clear that the intrusive cores to the summit regions and rift zones form a significant mass fraction of the volcanic edifice. Reliable estimates of this fraction, however, will require better definition of intrusive core morphology by seismic techniques and by densely spaced gravity measurements over the submarine flanks of the volcanoes. The wedge shape of the intrusive cores supports the idea that spreading plays an important role in the growth of the volcanic edifice. This spreading involves the lateral displacement of unbuttressed flanks over the oceanic crust in response to forceful magma intrusion in rift zones. Understanding the details of this process, however, demands a sharper image of the intrusive-core structure and better constraints on the nature of the decoupling zone between the mobile flanks and the oceanic crust. The profile of the oceanic crust beneath the volcanic pile appears to be consistent with the bending of a fractured lithosphere in response to the load of the pile. Clear delineation of the strongly three-dimensional structure of the deep crust beneath the volcano summits, however, presents a special challenge to high-resolution seismic and gravity techniques.

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