Detailed Gravity of E. Rift Zone

GL03310 2 of 6

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STRUCTURE OF THE LOWER EAST RIFT ZONE OF KILAUEA VOLCANO, HAWAII, FROM SEISMIC AND GRAVITY DATA

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ABSTRACT

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Two seismic refraction surveys were carried out in 1976 and 1977 on the east rift zone of Kilauea volcano as part of an exploratory program for geothermal resources. The short traverse seismic refraction survey of January 1976 delineated the upper surface structure of the east rift, revealing velocities of 2.5 km/s under the Kalapana line and 3.1 km/s under the Leilani line beneath a surface layer of low, but variable velocity. This survey was not successful in determining the depth of the high-velocity material. The longer traverse seismic refraction survey of 1977 outlined a high-velocity body at a depth of 2.1—2.3 km. Crustal layer velocities of 3.0, 5.25 and 7.0 km/s were observed with "kinks" in the travel-time curves suggesting that the 7.0 km/s material is intruding into the 5.25 km/s zone. The high-velocity body was interpreted to be a complex of dikes containing solidified magma from past volcanic eruptions that provide energy for the hydrothermal processes associated with the rift zone. Analysis of gravity data from a survey of 1975 provided constraints on the horizontal extent of the dike complex indicating a width varying from 12 to 17 km. Density contrasts of 0.4—0.6 were obtained by using velocity-density systematics demonstrating that the extruded surface lava is quite different from that which remains at depth. One possible conclusion is that magma differentiation has taken place.

INTRODUCTION

Since 1973 various types of geophysical and geochemical surveys have been conducted over the east rift zone of Kilauea volcano, Hawaii, as part of a geothermal exploration program to locate resources and to understand the thermal processes underlying the rift zone (Furumoto, 1976a). During the course of the geothermal project a test hole, well HGP-A, drilled to a depth of 1960 m, revealed a commercial quality geothermal reservoir with a base temperature of 300°C. From analytical and interpretive studies of the survey data and well logs, gradually a picture of the structure of the east rift emerged and with it came an insight into the thermal processes of the rift zone. Of the many surveys carried out, seismic refraction and

HIG Contribution No. 934

gravity surveys provided the necessary information to derive the subsurface geological structure of that part of the rift zone east of 155°W longitude. Extension of the results to the rift zone west of 155°W has been treated in another paper (Furumoto, 1978).

The present paper relies upon geophysical data obtained during the geothermal project. Generally, our geophysical results are consistent with interpretations from surface geological evidence but there are some differences especially on size of subsurface structure. For example, our results show a dike complex with a width varying from 12 to 17 km, while Swanson et al. (1976) estimated a width of about 4 km. Discrepancies between geophysical and geological interpretations have occurred frequently in Hawaiian geological studies. The dense volcanic neck of Koolau volcano on the island of Oahu has been calculated to have a diameter of 16 km by gravity data (Strange et al., 1965) and by magnetic data (Furumoto, 1976b), while speculation from geological data inferred that the dense plug is about 4.6 km wide (Macdonald and Abbott, 1970, p.141). In the interest of volcanology, these discrepancies due to different techniques should be pointed out.

For the present study, two separate seismic refraction surveys were carried out: one in January 1976 and the second in January 1977. Details of the surveys are in Suyenaga et al. (1978). The first survey with short traverses did not have adequate depth penetration for determining structure; the second survey with traverses 25 km long derived a velocity-depth profile of 4 km. The velocity-depth profile then provided constraints for the interpretation of gravity data. The utilization of seismic results in gravity interpretation depends on the relationship between P-wave velocity and density determined empirically by Manghnani and Woollard (1968). Without the velocity-density relationship, interpretations in this paper would not have been possible.

PREVIOUSLY AVAILABLE DATA

Of the five volcanoes that make up the island of Hawaii, Kilauea volcano is the youngest and was formed by eruption through a pre-existing flank of Mauna Loa volcano, which is located to the north. The east rift of Kilauea volcano was selected for geothermal exploration because reconnaissance surveys showed it to have many promising features (Furumoto, 1976a). The exploration program included the following types of surveys: gravity, magnetic, electrical resistivity, electrical self potential, groundwater temperature, microearthquake monitoring, ground noise, infrared scanning and geochemical. Data in preliminary processed form from all of the above-mentioned surveys, including the short traverse refraction survey of January 1976 plus logs from the 1960-m-deep well, HGP-A, were available when we felt that longer traverse seismic refraction data were necessary. Long line seismic refraction surveys were not included in the original plans for exploration because refraction data for the area were already available in published form; subsequently, the published data proved inadequate for our purpose.

Previous seismic refraction studies were for the purpose of determining the crustal structure of the island. Ryall and Bennett (1968) carried out a seismic refraction survey across the island of Hawaii with a straight-line traverse extending from Hilo to Kalae, the southern tip of the island (Fig. 1). Their results, which showed variation in crustal thickness from 12 to 16 km, were not applicable to our study on the east rift zone since the traverse was far away from the rift zone.

Hill (1969) carried out offshore seismic traverses around the island of Hawaii. The shot and recorder locations of the Hill traverse near the east rift are shown on the map of Fig. 1; the shots were offshore from the Kilauea east rift. The resulting velocity depth profile by Hill is shown in Fig. 2. Hill assumed that the first layer had a P-wave velocity of 1.8 km/s and then derived the rest of the velocity depth structure as a 3.1 km/s layer overlying a 5.1 km/s layer. Because his data were somewhat ambiguous, Hill offered two solutions below the 5.1 km/s layer: a 7.0 km/s layer or a 6.3 km/s layer (Fig. 2). The 7.0 km/s interface dips toward the southwest, while the interface of the 6.3 km/s layer dips toward the northeast.

Ward and Gregersen (1973), after some microearthquake studies in the western section of the rift zone, concluded that there was evidence of a high-velocity body at shallow depths under the rift zone. This finding has

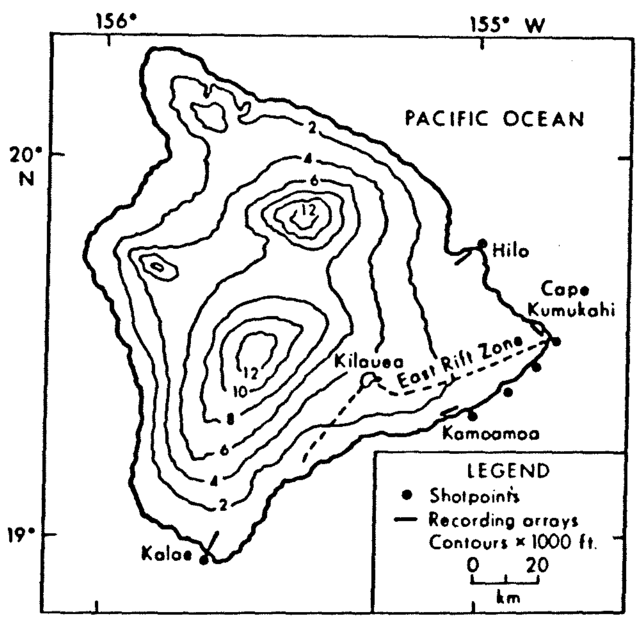


Fig. 1. Map of the Island of Hawaii with locations of seismic refraction traverses by Ryall and Bennett (1968) and Hill (1969). The rift zones of Kilauea volcano are shown for easy reference. Base map from Macdonald and Abbott (1970).

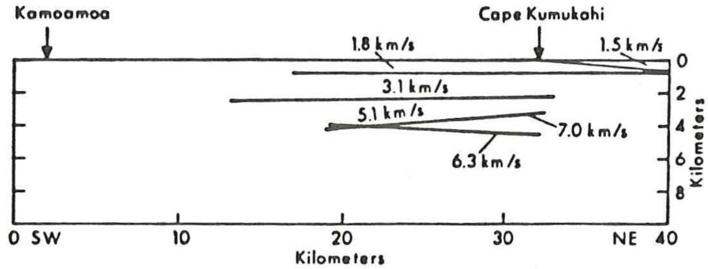


Fig. 2. Velocity depth profile along the shore of Puna District by Hill (1969). Hill gave two possible solutions for the fourth layer, both of which are shown.

been confirmed by Ellsworth and Koyanagi (1977). The high-velocity zone probably continues eastward as suggested by gravity data and microearthquake epicenters (Suyenaga and Furumoto, 1975).

Gravity data from two different surveys indicated that there is a dense body beneath the east rift zone. The gravity survey of the island by Kinoshita (1965) found a Bouguer anomaly of about 15 mgal over the east rift, and a more detailed survey by Furumoto (1976a) confirmed the anomaly. As dense material usually is associated with high velocity, the gravity data may be interpreted as indicating a high-velocity structure at relatively shallow depths under the east rift zone of Kilauea.

SHORT TRAVERSES OF 1976

Two short reversed refraction lines were completed in January 1976 (Suyenaga et al., 1978). The Leilani line, with shot points W and E, was located approximately over the crest of the rift zone; the Kalapana line, with shot points N and S, was on the southern flank of the rift zone (Fig. 3).

At each end of the two refraction lines, six shots were fired at a depth of 5 m. Shot sizes ranged from four to sixteen pounds of explosive. Field recording was done on magnetic tape systems and later filtered and transferred to paper records. The quality of first arrivals was good out to 2 km distance and poor beyond about 3 km. Arrival times were corrected for elevation but not for weathering because the thickness of the soil layer in all cases was negligible.

The travel time plots for both traverses are shown in Fig. 4. From shotpoint W on the Leilani line (Fig. 4A), a low-velocity surface layer of approximately 0.7—0.8 km/s is defined by two points on both sides of the shot. A second layer is defined by first arrivals at travel distances of 1.0 and 2.0 km, and a second arrival at travel distance of 2.4 km. This layer has an apparent velocity of 2.5 km/s. At about 2.4 km, the slope of the travel time plot first decreases, then gradually increases to a distance of about 3 km, beyond which first arrivals were not recorded because of distance attenuation.

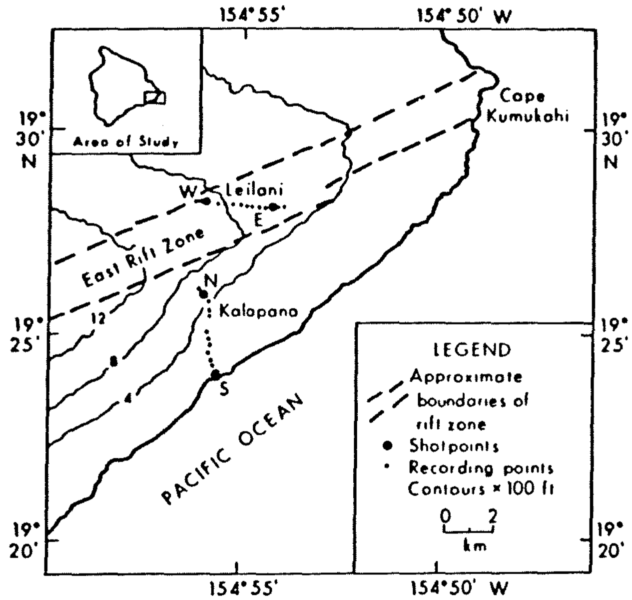


Fig. 3. Location of short refraction lines Kalapana and Leilani in the Puna District. The broken lines indicate the boundaries of the surface expression of the rift zone.

From shot point E, the surface layer is recorded by one point on each side of the shot. To the east the velocity was 1.1 km/s, and to the west it was 1.6 km/s. Large-amplitude secondary arrivals at distances of 1.0—1.2 km were interpreted as arrivals along the surface layer with a velocity of 0.9 km/s, indicating that the surface layer has a low but highly variable velocity. A shot recorded between 1.0 and 1.2 km showed a very poor pattern of arrivals; this plus the change in slope at 2.3 km from W, suggests that the structure is complex in that vicinity. At distances between 1.8 and 2.5 km from E, first arrivals indicate an apparent velocity of 3.9 km/s. The extension of this branch leads to an end travel time of 1.2 s at W, which agrees within 5% of the reciprocal travel time in the other direction, indicating that the apparent velocity of 3.9 km/s probably continues out that far.

The Leilani travel times were fitted by eye to the standard form:

$$t = t_i + \Delta/V$$

where t is travel time, t_i is the intercept time, Δ is distance, and V is velocity (Table 1). Neglecting the already noted complexity near 1 km from E, we constructed a simple model by assuming a surface velocity of 0.8 km/s with a dipping second layer (Fig. 5). The interface between the two layers dips 3° toward the east and velocity of the second layer is 3.1 km/s.

Fig. 4B is the Kalapana line travel time plot. From N, a surface velocity of 0.8 km/s was measured toward the north. This velocity was assumed to

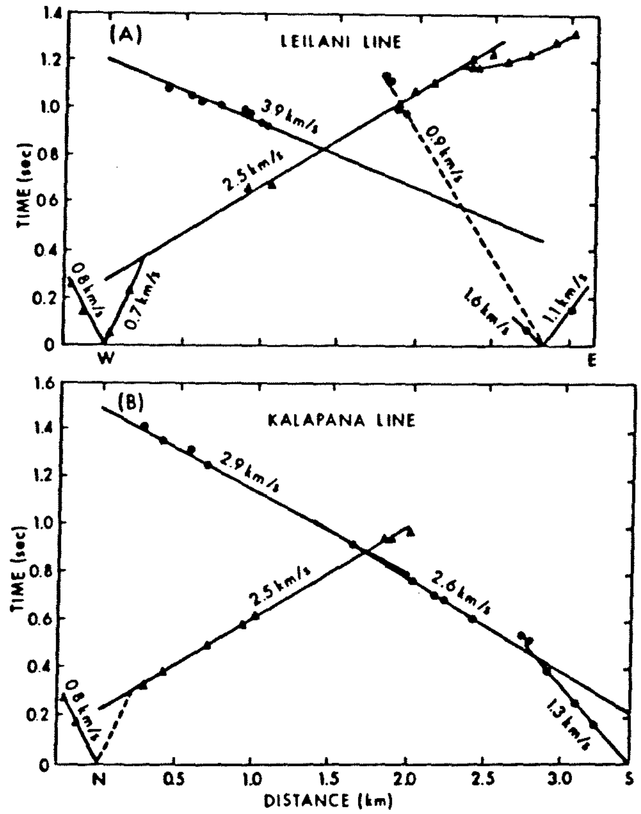


Fig. 4. Travel time plots of lines Leilani and Kalapani. Triangles indicate receptions from shotpoints W (top) and N (bottom); circles indicate receptions from shotpoints E (top) and S (bottom).

TABLE 1

Travel times for short line refraction

Refraction line	Shot point	Travel time curves* (seconds)
Leilani	W	$t_0 = 0.00 + \Delta/0.66$
		$t_1 = 0.27 + \Delta/2.51$
	E	$t_0 = 0.00 + \Delta/1.61$
		$t_1 = 0.44 + \Delta/3.86$
Kalapana	N	$t_0 = 0.00 + \Delta/0.82$
		$t_1 = 0.202 + \Delta/2.51$
	S	$t_0 = 0.00 + \Delta/1.35$
		$t_1 = 0.215 + \Delta/2.56$
		$t_2 = 0.30 + \Delta/2.92$

*Δ = distance (km).

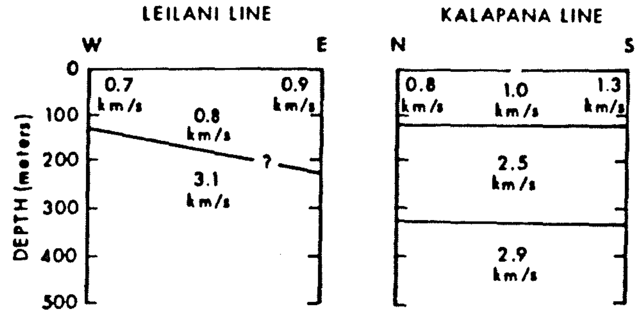


Fig. 5. Velocity depth profiles for Leilani (left) and Kalapana (right).

apply southward also and is indicated by the dashed line. An apparent velocity of 2.5 km/s was measured for a second layer by first arrivals from 0.3 to 2.1 km. Arrivals at greater distances were not detected since shot size was limited at N. From S, a surface velocity of 1.3 km/s was measured by use of three first arrivals, and a well-defined second layer of 2.6 km/s was recorded out to 1.7 km. Velocity increased at that point to 2.9 km/s.

The Kalapana line travel times were analyzed in a manner similar to that used for the Leilani data (Table 1). The data indicated a surface layer of variable velocity (0.8—1.3 km/s) overlying a layer of 2.5 km/s. The interface between the two layers was assumed to have no dip since the apparent velocities in either direction are comparable and the travel time plots intersect at the midpoint of the line. The assumption of no dip is reasonable even if the internal layers follow the topography since the latter dips only 2° from north to south over the refraction line. A third layer of velocity 2.9 km/s is indicated from one end on the line; data were not available from the other end. A three-layer model was derived (Fig. 5) by use of an average surface velocity of 1.0 km/s. The depths to the 2.5 - and 2.9 - km/s layers were calculated to be 110 m and 320 m, respectively.

The primary difference between this structure derived here and in previous studies (Ryall and Bennett, 1968; Hill, 1969) is the presence of the low-velocity surface layer. We believe the reason for this difference is that those studies did not have sufficiently close shot-receiver distances.

The low-velocity layer is comprised of interlayered aa and pahoehoe lava flow (Palmiter, 1976). Rinehart and Greeley (1971) reported near-surface velocities of 1.2—2.4 km/s on a pahoehoe flow. The lower velocities reported here are attributed to the large voids between flows and within the aa flows which decrease the bulk density of the entire structure.

Beneath the surface layer, a velocity of 2.5 km/s was observed under the Kalapana line and 3.1 km/s under the Leilani line. Such velocities are not thought to be associated with any drastic change in lithology (Palmiter, 1976). The change in velocity between the surface and second layer probably reflects saturation with water at the water table. At shallow depths, the compressional velocity in rocks is strongly affected by saturation (Christensen and Salisbury, 1975).

This survey was not successful in determining the depth to the high-velocity, high-density material. We measured second-layer velocities of 3.1 km/s at 100—200 m depth under the crest of the rift zone. This is higher than the 2.5 km/s under the flanks but is not indicative of a high-density zone which would most likely have a velocity near 6—7 km/s (Manghnani and Woollard, 1968). A refractor with this high velocity at depths shallower than about 1 km would probably have been observed. Since it was not, it was evident that a longer refraction line was needed. In the next section such an experiment is described.

LONGER TRAVERSES OF 1977

Two seismic refraction traverses exceeding 20 km were carried out in the Puna District in January 1977; the first in the east-west direction from Cape Kumukahi to the area around the geothermal well HGP-A and the second in the north-south direction nearly perpendicular to the first line (Fig. 6). Shots were fired at sea by the research vessel "Noi'i", and various types of recording units were deployed on land. The shot and ship positions were zeroed-in by the transit parties on land during the shooting sequence, which lasted an hour or more for each traverse. The traverses were carried out on two separate nights. On each traverse a sonobuoy was deployed before firing commenced and picked up after the shooting sequence. As indicated in Fig. 6, the sonobuoy drifted each time from position S1 to position S2. We assumed that the drift was linear, as the shooting run was of short duration.

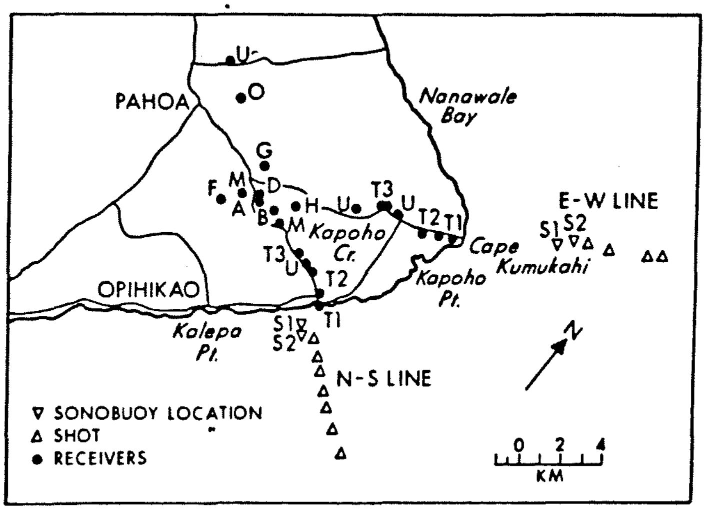


Fig. 6. Locations of shots and recorders for the two long traverses. S1 and S2 are positions of recording sonobuoys before and after each shooting run. Station A was located at site of well HGP-A.

The firing tone and seismic signals were recorded against WWVH signals that came in strong during the course of the shooting. On the vessel "Noi'i", the recording of WWVH was intermittent, limited to shots 4—7 of the north-south line.

The recording units deployed on land contained an assortment of instruments, and all of the recording was continuous on tape recorders or smoked paper drums. The geophones were of a 1-Hz and a 4.5-Hz type; the difference in natural frequencies was not significant since the signals were in the higher frequencies.

Samples of the recordings at station T1 on the north-south line are displayed in Fig. 7. The rather high signal-to-noise ratio is apparent. The recordings include P-wave and T-phase arrivals; the label *SB* in the recorder indicates the acoustic signal from the detonation as picked up by the sonobuoys and transmitted to station T1.

In the processing of data, all arrivals were reduced to the datum plane at sea level. A uniform value of 1 km/s for the layer above sea level was used for this reduction in accordance with results from the previous section.

The travel time plots for shots 2 to 5 on the east-west line are displayed in Fig. 8 and the travel times listed in Table 2. In these plots we may consider the shot points as fixed and the recording points as variable with distance. On the other hand, in Fig. 9, the stepout times of the shots are plotted with the recording stations T1 and T2 considered as fixed. In effect, with the arrangements of these plots, a split traverse for the east-west line is constructed. On the east-west line for these fixed shot point plots a high-velocity layer is observed across the entire array. On the north-south line, however, a surface layer of velocity 3.0 km/s is observed out to 6—8 km on the array (Fig. 10). Since the arrays on the east-west line are located no closer than about 8 km to the recorders, a 3.0-km/s layer would not be visible in this configuration.

An attempt was made to delineate the surface layer on the east-west line by plotting the shots for the two closest fixed stations. Corrections were made to the travel time for the depth of the water at each shot as well as for the apparent slope of the ocean bottom out to the most distant shot. The slope was estimated to be about 7° from a bathymetric chart by Malahoff and McCoy (1967). The shots were in effect brought to the ocean bottom and the correction was made assuming the surface was the sloping ocean bottom. The travel time data are given in Fig. 9 and indicate an apparent velocity of 3.0 km/s.

We feel confident in assigning a velocity of 3.0 km/s to the first layer as, in addition, an upper layer of apparent velocity 3.0 km/s was observed just below the sea level datum plane on the north-south travel time plots, and in previous studies (Hill, 1969) of this region.

CAUTION

For this survey, unfortunately, we do not have any reversed profiles, and to designate velocities rigidly is not justified. Our assigned velocity values are estimated to have an uncertainty of ± 0.25 km/s, however.

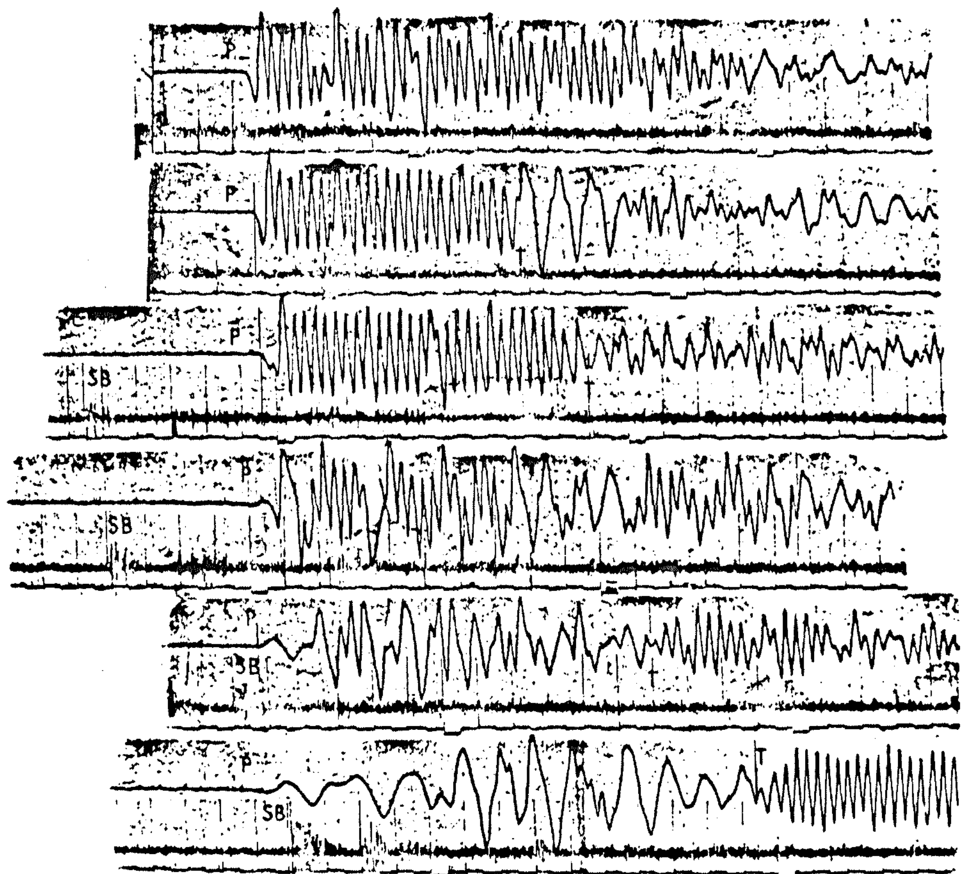


Fig. 7. Sample of seismograms from N-S line. The high-frequency traces are the acoustic arrivals detected by the sonobuoys and telemetered to manned recording stations.

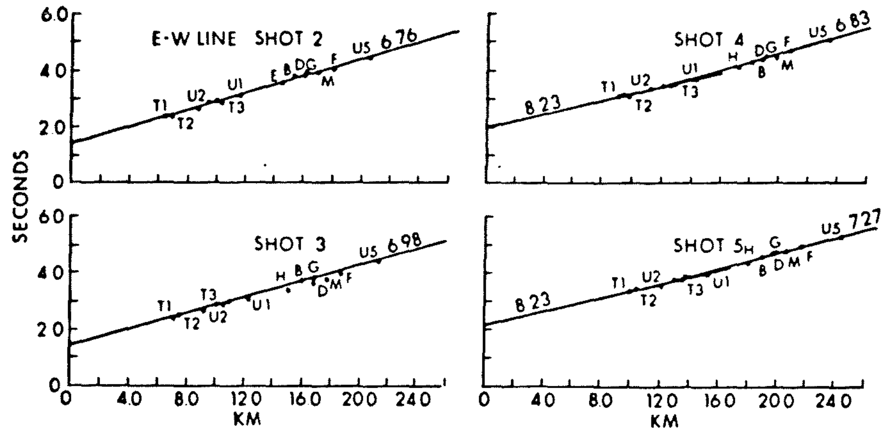


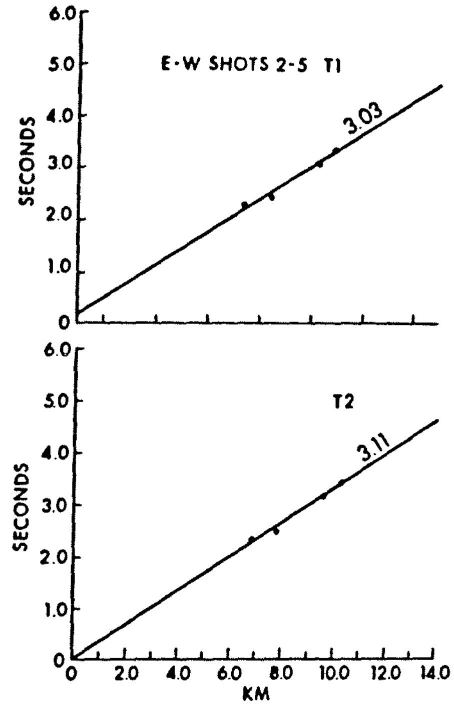
Fig. 8. Travel time plots of E-W line for shots 2—5 with recorders as stepout variables. The surface layer was undetected.

TABLE 2

Travel times for the 1977 east rift refraction

Refraction line	Shot point	Travel time* (seconds)
East-west	2	$t = 1.40 + \Delta/6.76$
	3	$t = 1.40 + \Delta/6.98$
	4	$t = 2.00 + \Delta/8.23$
		$t = 1.70 + \Delta/6.83$
	5	$t = 2.20 + \Delta/8.23$
		$t = 2.00 + \Delta/7.27$
North-south	1	$t = 0.00 + \Delta/3.00$
		$t = 1.00 + \Delta/5.25$
	2	$t = 0.15 + \Delta/3.00$
		$t = 1.10 + \Delta/5.64$
	3	$t = 0.45 + \Delta/3.43$
		$t = 1.35 + \Delta/5.71$
	4	$t = 0.9 + \Delta/5.25$
		$t = 0.9 + \Delta/5.80$
	5	$t = 1.0 + \Delta/5.00$
		$t = 0.9 + \Delta/5.70$
	6	$t = 0.9 + \Delta/5.25$
		$t = 1.2 + \Delta/6.80$
	7	$t = 1.45 + \Delta/7.50$

*Δ = distance (km)



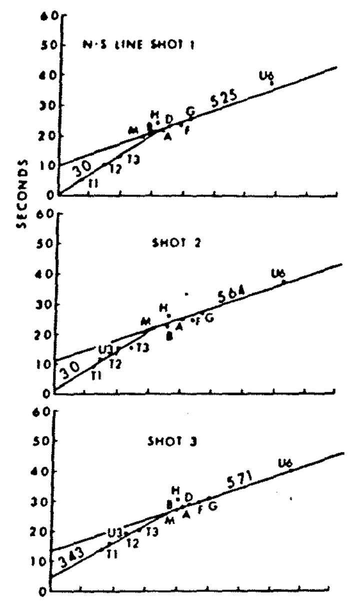


Fig. 9. Travel time plots of E-W line for stations T1 and T2 with shots as stepout variables.
Fig. 10. Travel time plots of N-S line for shots 1, 2, and 3.

The travel time plots of the north-south line are listed in Table 2 and are displayed in Figs. 10 and 11. Since the detonation time was not well recorded for shots 4—7 (Fig. 11), the intercept times at the zero distance are not reliable. The velocities derived from the stepout times, however, are accurate. Also the bends or "kinks" in the curves are reliable.

No 5.2 km/s layer on E-W line!

Recall 5.2 km/s layer has high R
Ferromito "No GT FLUIDS"

From the above travel time plots the following velocity structures were derived. On the east-west line, the first layer has a velocity of 3.0 km/s and at a depth of 2.1 km there is a high-velocity layer of 7.0 km/s. On the north-south line, the first layer has a velocity of 3.0 km/s and is underlain by a second layer of 5.25 km/s at a depth of 1.6—1.8 km. The 5.25-km/s value was selected as the average of velocities detected in the north-south profiles for the second layer. Velocities of about 7.0 km/s are observed on two of the north-south travel time plots and "kinks" in the curves of shots 4—6 in the north-south traverse suggest that a high-velocity material is intruding into the 5.25-km/s layer. The high-velocity intrusive material is interpreted as the extension of the 7.0-km/s layer detected in the east-west traverse.

Due to the presence of "kinks" in the north-south travel time graphs, we resorted to graphical solutions by ray tracing methods to come up with an

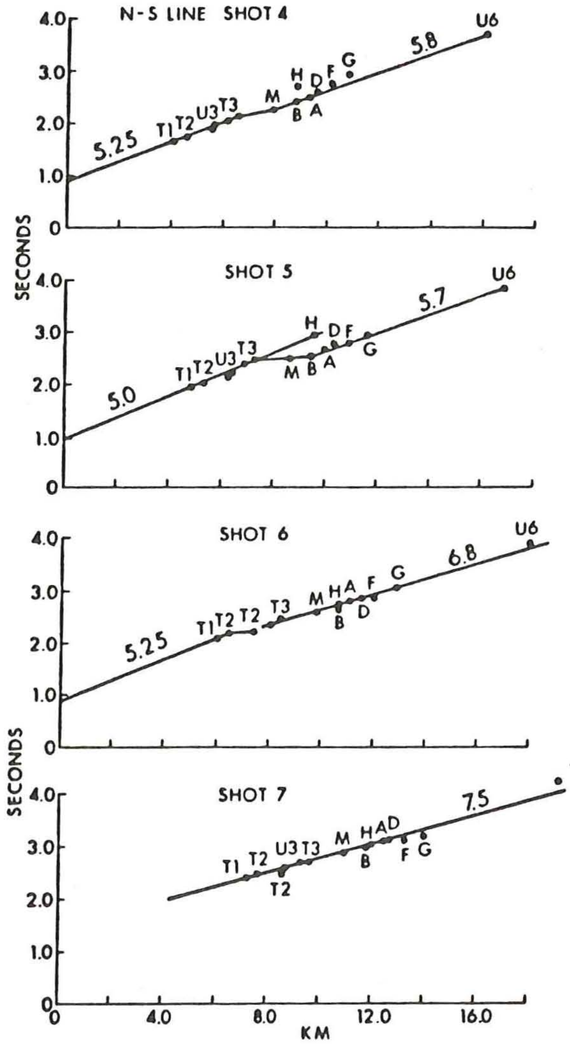


Fig. 11. Travel time plots of N-S line for shots 4—7. Notice the "kinks" in the lines.

appropriate model. A model that satisfies the travel time data is illustrated in Fig. 12.

An intrusive body with a velocity of 7.0 km/s was deposited in the 5.25-km/s layer. The sum of the travel times along the segments of the ray paths agree closely with the observed travel times for the various shots. A similar ray tracing exercise for the east-west line was also performed to check the derived model.

The depth to the intrusive high-velocity body was determined from refraction data. The width was constrained by gravity data as will be shown later. The intrusive complex was calculated to dip 6—9° toward the north-

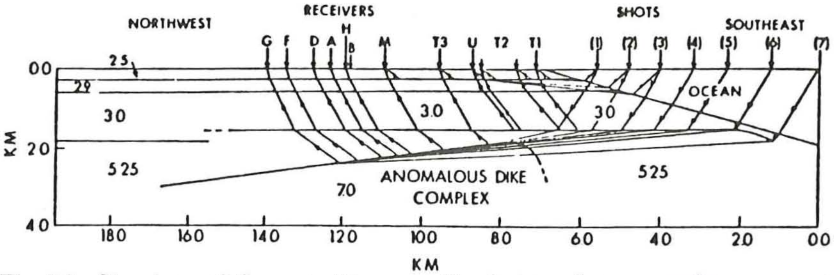


Fig. 12. Structure of the east rift zone with shot-receiver ray paths.

west to account for the 5.75-km/s branch observed beyond the kinks in shots 4 and 5 (Fig. 11).

The velocity-depth profiles for the east-west and north-south traverses are shown in Fig. 13. The east-west traverse is considered representative of

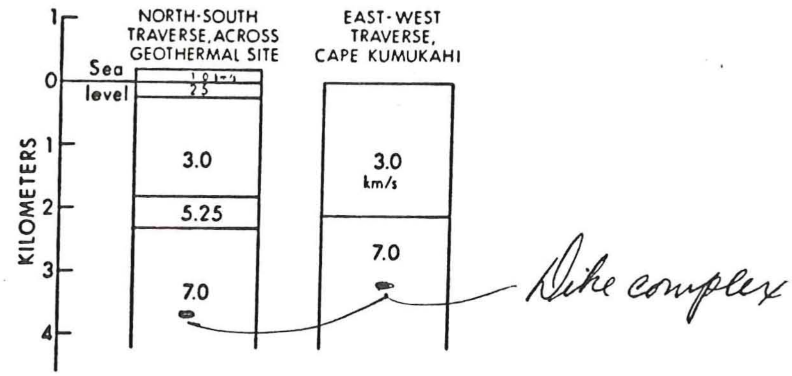


Fig. 13. Velocity depth profiles for N-S and E-W traverses. The N-S profile cuts across the geothermal site; the E-W profile is representative of the eastern end of the rift zone.

the subsurface structure under Cape Kumukahi and the north-south traverse representative of the structure in the vicinity of the geothermal well. The 7.0-km/s layer may be horizontally layered, but the difference in depths of 2.3 km and 2.1 km for the traverses are within observational error.

Our structure differs from that obtained by Hill (1969), which is indicated in Fig. 2. The difference may be explained in that Hill's profile shows the crustal structure offshore along a line drawn through the shot points of Fig. 1, while our profile shows the structure on land beneath the rift zone. An important conclusion of our refraction survey is that we have detected a high-velocity body of 7.0 km/s at shallow depths of 2.1—2.3 km.

ANALYSIS OF GRAVITY DATA

Using the results of the seismic refraction survey, which indicate a 7.0-km/s layer at depths of 2.1—2.3 km, available gravity data were analyzed to outline the lateral dimensions of the high-velocity layer. The gravity data used were from a survey carried out for the geothermal project in the summer of 1975 (Furumoto et al., 1976). The Bouguer map resulting from the survey, which consisted of 210 stations at average grid spacing of 0.5 km, is shown in Fig. 14. Cross sections of the gravity profile along lines AA' and BB' were taken to find density anomalies that would fit the data. The line AA' approximates the north-south seismic refraction traverse.

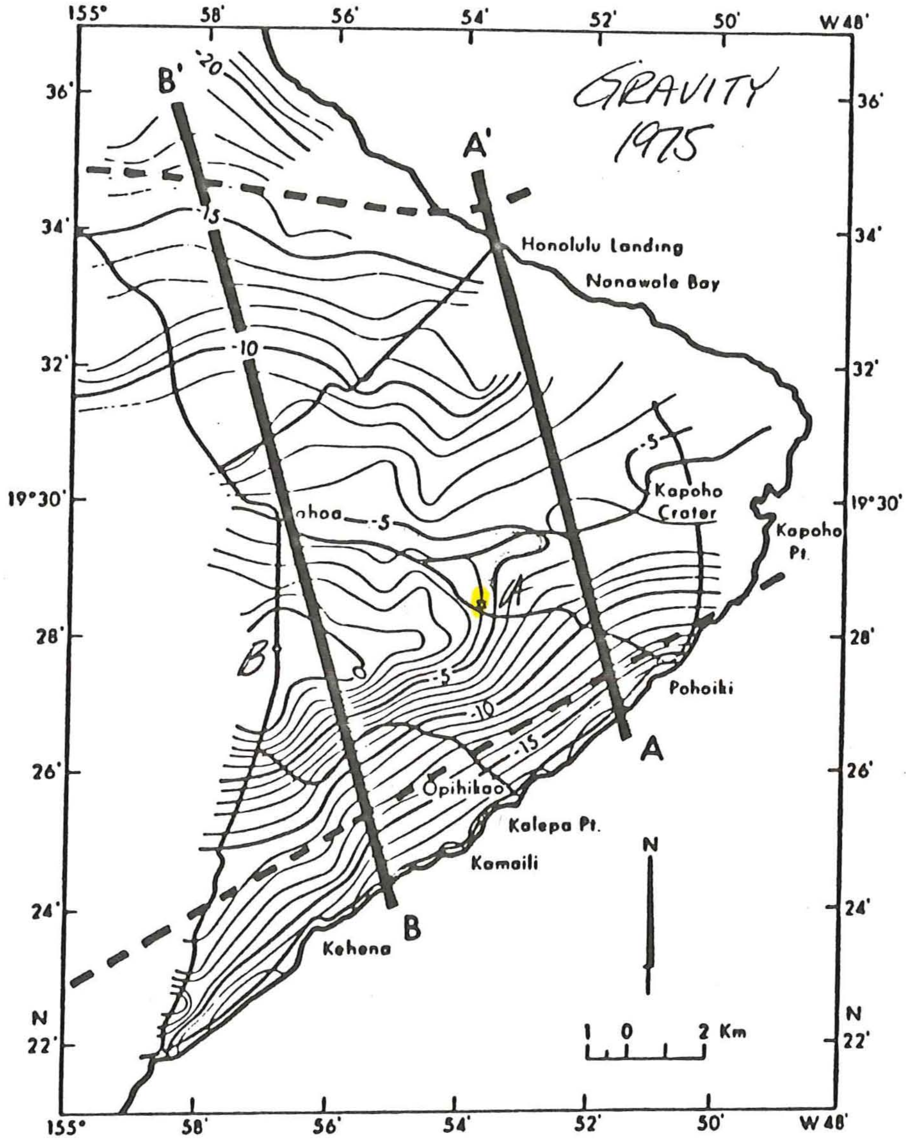
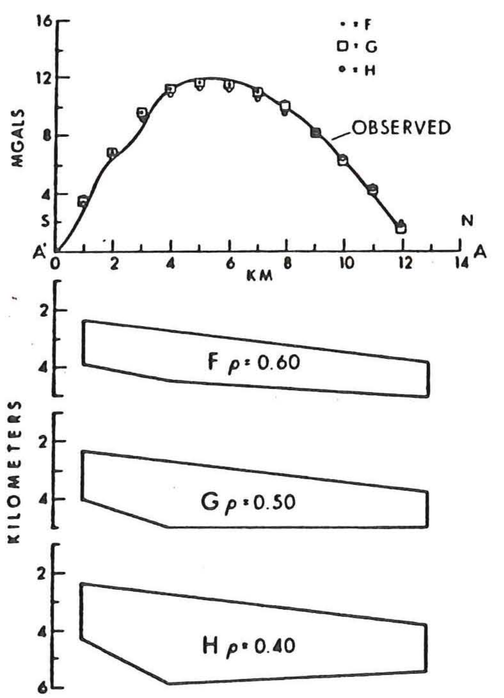


Fig. 14. Bouguer gravity map of Puna District with maximum value taken as zero. Lines AA' and BB' are the profiles where two-dimensional analyses were made. The dotted lines represent the boundaries of the dike complex under the east rift. The star indicates location of well HGP-A.

For gravity modeling, we used a two-dimensional type computer programming (Talwani et al., 1959). From the seismic refraction data, we assumed that the 7.0-km/s layer represented a high-density anomaly located at a depth of 2.3 km. For the density contrast, we considered values of 0.4, 0.5 and 0.6 g/cm^3 as approximating the contrast between 7 km/s and 5.25 km/s according to velocity-density relations obtained by Manghnani and Woollard (1968) for Hawaiian basalt. The various models that fit the data are shown in Figs. 15 and 16 for profiles AA' and BB', respectively. The significance of this exercise is that we have determined the width of the high-velocity, dense anomaly to be about 12 km at AA' and 17 km at BB'. If we had been able to carry out reversed traverses in the seismic refraction surveys, we could have obtained the width of the high-velocity body from seismic data. In lieu of data from reversed profiles, however, gravity data gave satisfactory results.

We have equated the dense 7.0-km/s body with a complex of dikes that previous investigators (e.g., Swanson et al., 1976) have said to underlie the east rift zone. The outline of the horizontal extent of the dike complex under profiles AA' and BB' is indicated in Fig. 14. An interesting behavior of the dike complex is its apparent narrowing in the region of the geothermal well HGP-A.



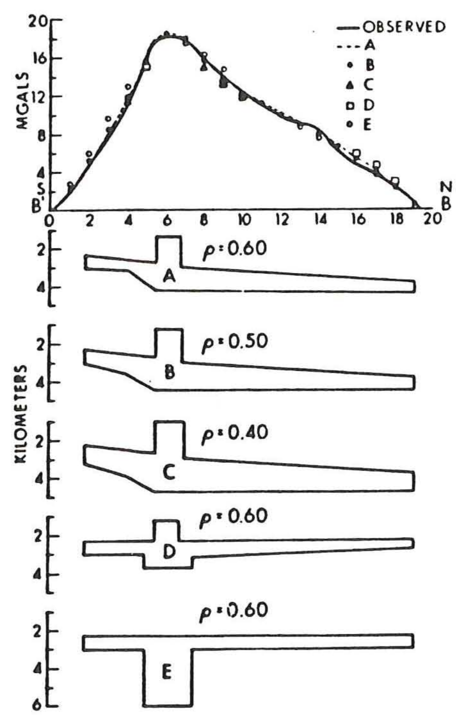


Fig. 15. Bouguer gravity for profile AA' and a few models that fit the data.
Fig. 16. Bouguer gravity for profile BB' and a few models that fit the data.

DISCUSSION

The hypothesis of magma moving down the rift zone from a central holding reservoir through subterranean passageways of thin vertical dikes was proposed sometime ago (Fiske and Jackson, 1972). From geological and geodetic data, Swanson et al. (1976) proposed that the active part of the rift zone tended to migrate southward and that the rift zone including the active and inactive parts was about 4 km wide. On the other hand, Malahoff and McCoy (1967) had calculated from marine magnetic data that the submarine portion of the rift zone was 12 km wide. What was found in this study was that the dike complex including active and inactive dikes was very wide, ranging from 12 to 17 km in the lower rift zone (Fig. 17). Our results corroborated the findings of Malahoff and McCoy.

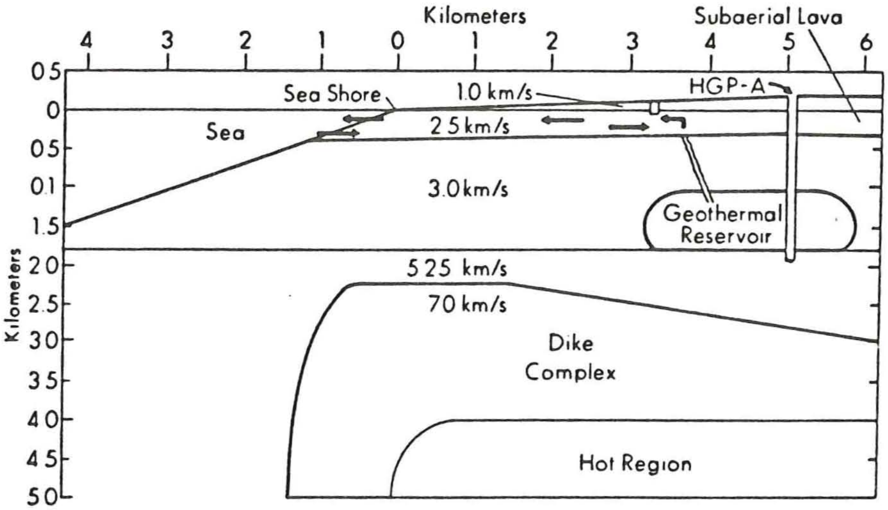


Fig. 17. Geological structure in the vicinity of the geothermal reservoir as derived from various data.

The discrepancy in estimated sizes of subsurface structure by geophysical and by geological methods has occurred often. The case of estimating the size of the plug under Koolau caldera has already been mentioned. Generally, geophysical methods come up with structure two to four times larger than geological methods do. A probable explanation to resolve the difference is that lava flows cover up and obscure previously existing structure, and what we observe on the ground surface in a given area are the results of the last of volcanic activities and their associated by-products. On the other hand, geophysical methods such as gravity, magnetic and seismic refraction surveys can uncover hidden structure. In the case of the east rift zone in the Puna area, the narrow band of craters, fissures, vents and cracks indicating the active rift shows the location of a part of the subterranean passageways in operation today. Gravity data revealed the existence of an inactive part of the dike complex that is as much as 10 km wide to the north of the present active rift.

To infer the size of even the active part of the dike complex by geological data alone may be risky. Swanson et al. (1976) estimated the active part to be 2—3 km wide, about the width of the band of craters and vents. But our gravity data show a substantial portion of subsurface dense material to the south of the present rift zone. This means that there could be dikes and passageways underground without such surface manifestations as vents, fissures and craters. Evidence for the active nature of the dense material south of the present rift zone was provided by the source mechanism of the Kalapana earthquake of November 29, 1975, the epicenter of which was located at 19°20′06″N, 155°01′45″W, the southern margin of the dike complex we have outlined. The source mechanism has been interpreted as a low angle overthrust from sudden rifting at shallow depth, perhaps due to magmatic pressure (Furumoto and Kovach, 1979). With these considerations, the active part of the dike complex is much wider than the 2—3 km estimated by Swanson et al. (1976).

The existence of a broad inactive part of the dike complex north of the rift zone supplies evidence for the theory of southward migration of the active zone as described by Swanson et al. (1976). It can be conjectured that the rift zone migrated 10 km since its inception to the present, but this would put the first position of the rift zone high up on the submarine portion of the slopes of the then existing Mauna Loa. This is rather unlikely considering the gravitational effects of a mountain slope. A more likely scenario is that the first rift zone position was somewhere to the north of the present position and the original rift split into two with one part migrating north and the other migrating south. The northward-migrating part ceased operation a long time ago and became covered with lava flows. The southward-migrating part is still active today.

A geothermal reservoir with a base temperature of 300°C was found by drilling. An isothermal region of 300°C extended from a depth of 1.0—1.6 km below sea level. The nature of the reservoir and its thermal processes will be discussed in detail in a future paper, but the salient properties will be mentioned here. The reservoir appears to be located within the 3.0-km/s layer (Fig. 17). It is unlikely that the reservoir extends into the 5.25-km/s layer because the higher velocity means higher density and hence less pores and cracks present. The lateral dimension of the reservoir has been inferred from microearthquake data. By extrapolation of the temperature gradient of the reservoir, it has been inferred that the dike complex is hot enough as to be above the Curie temperature; and analysis of magnetic data supports a hot dike complex (Furumoto, 1978). The geothermal reservoir derives its thermal energy by conduction from the hot dike complex.

km³ volume

The high velocity and rather high density of the material making up the dike complex are intriguing. By the velocity-density relation of Manghnani and Woollard (1968), a velocity of 7.0 km/s corresponds to a density of 3.0–3.1 g/cm^3, while 5.25 km/s corresponds to a density of 2.55 g/cm^3. Fujii and Kushiro (1977) melted samples of lava from Kilauea and subjected them to high pressures to measure density and viscosity. At 1 kbar pressure, which approximated 3 km depth, the density was only 2.63 g/cm^3. This means that the rock extruded on the ground surface as lava differs from the dike material at 3 km depth by about 0.4 g/cm^3, under the same conditions. It is clear that what we observe as extruded lava is quite different from what remains trapped below among the interstices of the dike complex. Unfortunately, we have not yet obtained any samples from 3 km depth to ascertain what sort of mineral assemblage is responsible for the high density. It may be inferred that even in the dikes of the rift zone some magma differentiation takes place.

SUMMARY

Seismic refraction surveys resulted in velocity depth profiles of the east rift of Kilauea that were consistent with previously published data. In addition, the results provided necessary vertical constraints for the proper interpretation of gravity data. The gravity data in turn outlined the lateral dimensions of the dike complex, which infer the location of the passageways through which magma from the central holding reservoir migrated down the rift zone. The commercial quality geothermal reservoir found by well HGP-A seems to be, for the most part, located in the layer with a P-wave velocity of 3.0 km/s.

The outlining of subsurface structure by seismic refraction and gravity methods makes possible the understanding of the thermal and hydrothermal processes associated with the east rift of Kilauea. One attempt at interpretation has been made (Furumoto and Broyles, 1977); other interpretive studies by interested volcanologists are welcomed.

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