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CHAPTER 3:
MAGMA GENESIS IN MID-OCEAN RIDGE AND CONTINENTAL INTRA-PLATE

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Natura of the Magma Conduit Under the East Rift Zone of Kilauea Volcano, Hawaii

A. S. FURUMOTO *Hawaii Institute of Geophysics, University of Hawaii, Honolulu, Hawaii 96822*

ABSTRACT

From a combination of results of gravity, magnetic and seismic refraction surveys, the dike complex under the east rift zone of Kilauea Volcano in Hawaii was found to extend for 110 km from the summit area of the volcano to a point 60 km at sea beyond the eastern tip of the island. Near the summit the complex is 20 km wide, and at about 40 km distance from the summit, the complex narrows to 12 km wide. The main body of the dike complex is 2.3 km deep, but some parts are as shallow as 1 km.

2.3 km 7546'
1 km 3281'
" 0,62 mi

From extrapolation of temperature data of a deep well and from analysis of magnetic data, it was inferred that temperature of the dike complex is above the Curie point of 540°C. The internal part of the complex can approach the melting point of 1060°C.

The dike complex was formed by numerous excursions of magma from the holding reservoir under the volcano summit. The theory of forceful intrusion of magma into rift zones accounts for the magma excursions and migration of the passageways.

Gravity and seismic velocity data indicate that density of the material left in the dike complex is 3.1 g/cm^3. In the light of recent density determinations of Hawaiian rocks under high pressure and temperature, it is concluded that during Hawaiian volcanic activity, less dense components of the parent magma erupt through surface vents while the more dense components remain trapped below. Samples of the dense material from the dike complex are required before we can have a complete picture of the parent magma of Hawaiian volcanoes.

The dike complex is the source of thermal energy for a commercial quality geothermal reservoir that was found by drilling.

INTRODUCTION

As part of the geothermal development project on the island of Hawaii, a series of exploratory geophysical surveys were carried out over the island, especially over the east rift zone of Kilauea Volcano (FURUMOTO, 1976). In addition an exploratory well, HGP-A, drilled to a depth of 1960 m, found a commercial quality geothermal reservoir. One of the results of combining geophysical data and log data from the well was the delineation of the subsurface geological structure of the east rift land mass (BROYLES *et al.*, in prep). Once the geological structure became known, the thermal process of the rift zone and the nature of the magma conduit under the rift zone became clear. In this paper I present the derivation of the geological structure and discuss the nature of the magma conduit under the rift zone.

37 mi

GEOLOGICAL STRUCTURE OF THE EAST RIFT ZONE OF KILAUEA

Kilauea is the youngest and most active of the five volcanoes that comprise the island of Hawaii. Because of the cumulative accomplishments of many investigators since 1900, good qualitative knowledge of the geological structure of Kilauea Volcano and its east rift was available at the time the geothermal project began. The geothermal program determined quantitatively the geological structure of the area east of 155°W longitude (*cf.*, Fig. 1) which comprises about one-third of the length of the land portion of the east rift.

The source of magma for Kilauea Volcano is located at about 60-km depth, according to earthquake data (EATON and MURATA, 1960). From this source, magma rises through nearly vertical conduits to enter a holding reservoir under the summit caldera. Earthquake data indicate that the holding reservoir is located 3 to 4 km below the ground surface (MINAKAMI, 1974; KOYANAGI *et al.*, 1977). From the holding reservoir, magma can erupt through craters and vents in the summit area or move laterally through subterranean passageways into the rift zones.

Of the two rift zones associated with Kilauea Volcano, the east and southwest rift zones, I shall concentrate on the

USGS [illegible]
Sept 78 identifies
[illegible] median zone

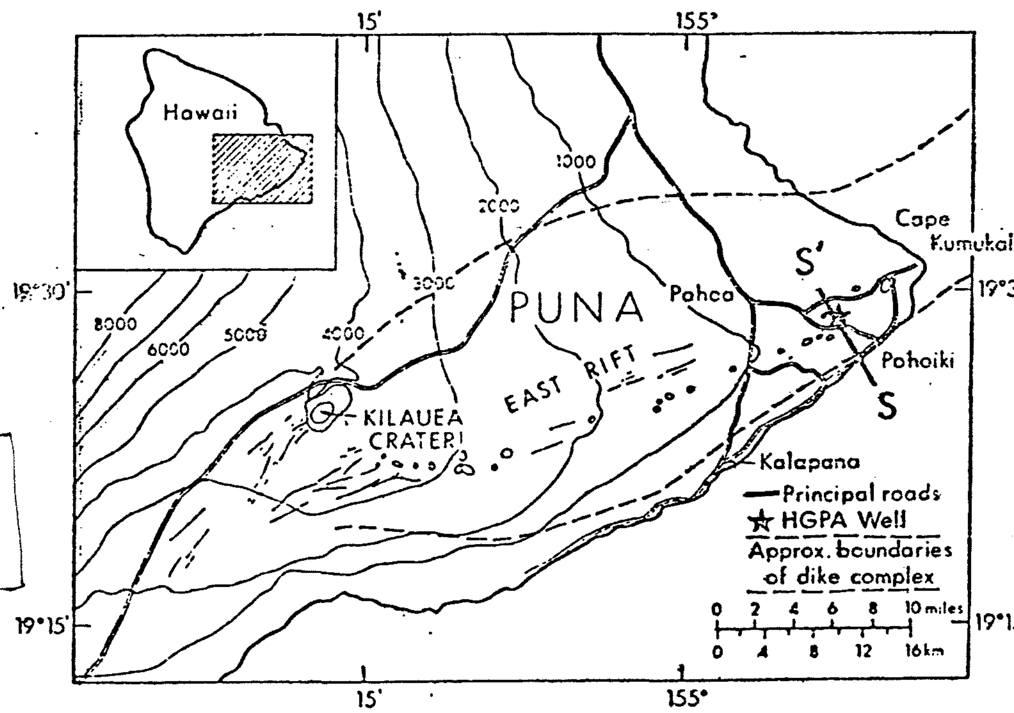

FIG. 1 - Map of the east rift of Kilauea with the approximate boundaries of the dike complex underlying the rift zone. The star marks the location of well HGP-A; line SS' shows the location of the transverse section structure of Fig. 2. The earthquake of November 29, 1975 occurred under the village of Kalapana. (Base map adapted from DEPT. OF GEOGRAPHY, 1973, «Atlas of Hawaii»).

east rift zone. The east rift zone is manifested on the surface by a line of vents and craters that begins in the southeast direction from the summit caldera but at a point roughly 8 km from the caldera the line bends nearly 90° and heads east-northeast. The line

Submarine pillow lavas ; low perm.

lo electric resist

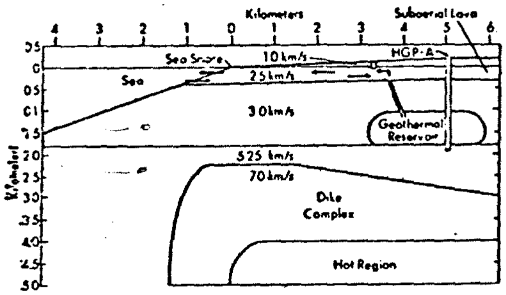

FIG. 2 - Velocity-depth profile of transverse section of the east rift. The profile corresponds to the section under line SS' of Fig. 1.

then passes through what is geographically known as the Puna District; beyond Cape Kumukahi on the eastern tip of the island, it enters the sea and continues for another 60 km (Fig. 1).

The mode of magma excursion down the rift zone through subterranean passageways can be accounted for by the theory of forceful intrusion (FISKE and JACKSON, 1972; SWANSON *et al.*, 1976). According to this theory, gravitational slumping of the mountain mass produces vertical cracks parallel to the ridge crest of the east rift zone, and into these cracks magma forcefully injects itself. Under pressure of the magma, the cracks then widen and extend. Because the east rift zone is anchored against the older flanks of Mauna Loa, the flanks of Mauna Loa exert a control on the gravitational field of the east rift mass, and the stress from the gravitational field in turn directs the propagation of the cracks. Over geological ages, with numerous magma excursions, many passageways were formed and these

1 Tension opens cracks

2 magma then enlarges the fracture net.

remained as dikes filled with magma residues. Because the flanks of Mauna Loa act as barriers to the north, the passageways tend to migrate to the south. The accretion of dikes eventually built up a dike complex (SWANSON *et al.*, 1976).

The above was the general picture of the east rift structure available at the initiation of the geothermal project. In the project, seismic refraction surveys derived a velocity-depth profile as shown in Fig. 2 (BROYLES *et al.*, in prep) which represents a transverse section across the rift zone along line S of Fig. 1. Line S cuts through the geothermal well HGP-A. The structure under the line consists of four nearly horizontal layers with P-wave velocities of 1.0, 2.5, 3.0 and 5.25 km/sec. The upper two layers are composed of subaerially extruded lava characterized by high porosity and

GRAVITY

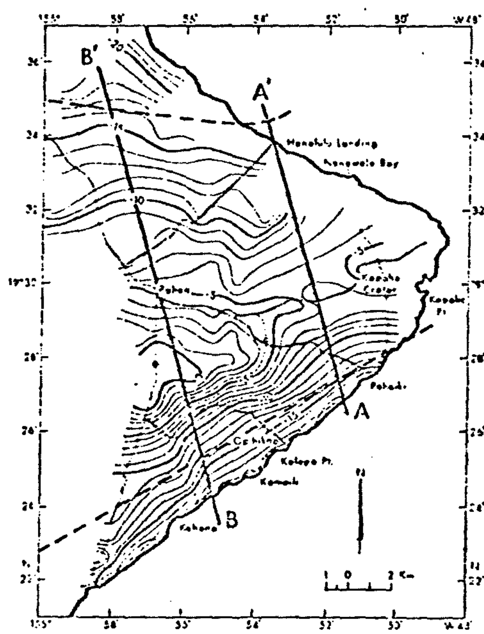

FIG. 3 - Bouguer gravity map of the eastern section of Puna District (FURUMOTO *et al.*, 1976). The dashed lines mark the boundaries of the dike complex underlying the rift zone. Lines AA' and BB' the sections analyzed by model fitting.

permeability. The top of the 2.5 km/sec layer coincides with the water table, which is almost at sea level. The 3.0-km/sec layer is made up of pillow lavas with low permeability. From core samples of HGP-A it was found that there are numerous fractures and fissures in the 3.0-km/sec layer, and because of hydrothermal alteration it was apparent that heated groundwater circulated through these cracks. The 5.25-km/sec layer, coincides with very high electrical resistivity as determined by KELLER *et al.* (1977). This means that geothermal fluids are not found in the 5.25-km/sec layer.

water in ? gas phase

At a depth of 2.3 km the 5.25-km/sec layer is intruded by a body with a velocity of 7.0 km/sec. For outlining the dimensions of this intruding body, gravity data obtained by FURUMOTO *et al.* (1976) (Fig. 3) were analyzed in the following way. The 7.0-km/sec velocity was considered to correspond to a density of 3.1 g/cm³, and 5.25 km/sec to 2.5 g/cm³ from the empirical density-velocity relationship determined by MANGHNANI and WOOLLARD (1968) for Hawaiian rocks. This in turn means a density contrast of 0.6 g/cm³ for the intruding body. The Bouguer gravity map of Fig. 3 was then analyzed for anomalous mass, with the constraint that the body started from a depth of 2.3 km. Two profiles AA′ and BB′ in Fig. 3 were selected for analysis by means of the two-dimensional model-fitting program developed by TALWANI *et al.* (1959). The results obtained by BROYLES *et al.* (*in prep*) are shown in Fig. 4. Given the shape, the anomalous body that can account for the gravity data is 12 km wide under AA′ and 17 km wide under BB′, irrespective of the choice of density

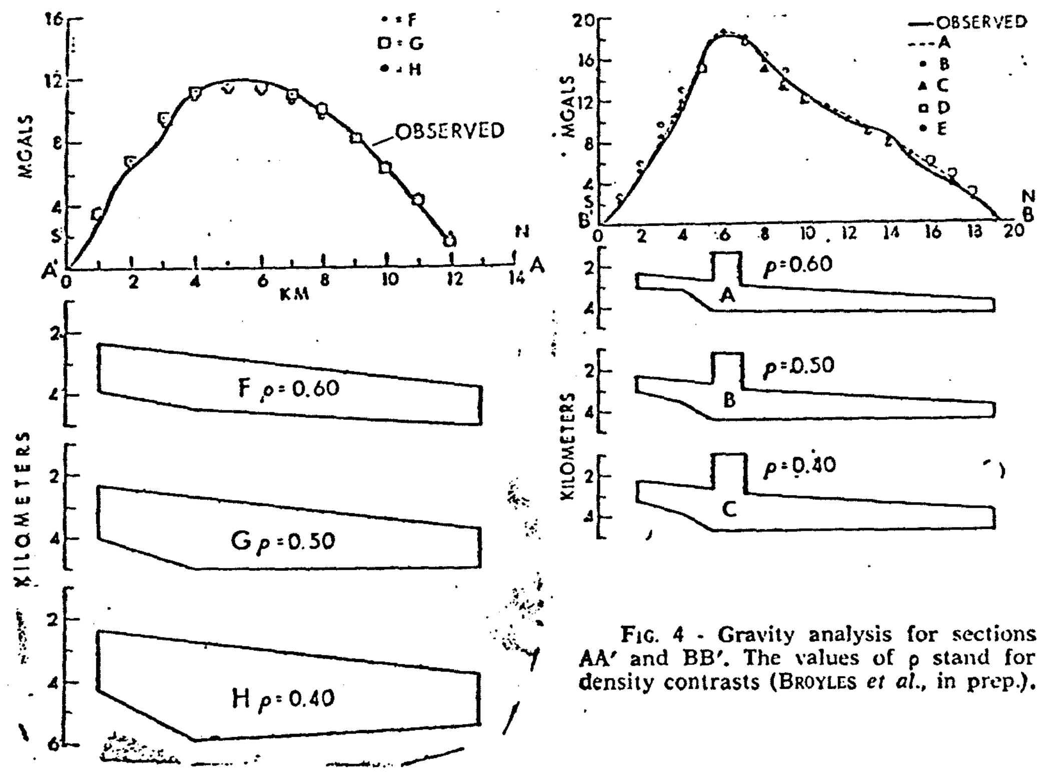

FIG. 4 - Gravity analysis for sections AA′ and BB′. The values of ρ stand for density contrasts (BROYLES *et al.*, in prep.).

contrast. The anomalous body was then outlined between AA′ and BB′ by following the isogal lines, as illustrated by the dashed lines in Fig. 3.

This dense body was interpreted to be the dike complex formed by numerous magma excursions from the summit holding reservoir down the rift zone. The successive excursions took new passageways, and in time the set of closely packed dikes had expanded into a complex over 17 km wide at BB′. The narrowing of the complex from 17 km to 12 km probably represents the expanding front of the growing dike complex. In time the 12-km width under AA′ will become much wider with successive magma excursions from the summit. The relationship of the dike complex to other structures is shown in Fig. 5. Atop the dike complex sits a geothermal reservoir which was found by drilling, and to the west sprouts a chimney-like superstructure which was outlined by gravity and magnetic data. The source of thermal energy for the geothermal reservoir is the dike complex, which is still very hot.

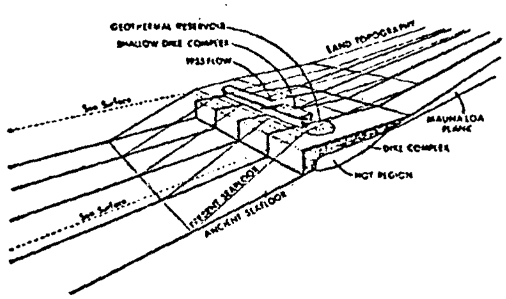

FIG. 5 - A perspective view of the dike complex underlying the east rift zone and the relation of the dike complex to the geotermal reservoir and chimney-like structure. The view is from east to west.

The success of the gravity interpretation for the Puna area east of 155°W prompted an analysis of the regional gravity data compiled by KINOSHITA (1965) (Fig. 6). Two profiles CC′ and FF′ farther up the rift zone were selected for similar analysis by the two-dimensional model-fitting program. As different density contrasts affect thickness of the anomalous rather than width, only a contrast of 0.6 g/cm³ was used. The results are shown in Fig. 7. Under profile CC′ the body is 40 km wide, while under FF′ it is 33 km wide. These broad widths are due to the ancient rift zones of Mauna Kea and Mauna Loa that underlie the Kilauea east rift zone. Because of this interference, the northern boundary of the Kilauea east rift could not be determined by gravity data. The southern boundary of the rift zone, however, could be defined easily, as shown in Fig. 6.

Four outlining the northern boundary of the rift zone, magnetic data proved very handy. As discussed in a later section, the extrapolation of the temperature data from well HGP-A led to the conclusion that the dike complex was hot enough to be above the Curie temperature. This means that the dike complex should show up as a nonmagnetic anomaly in an otherwise strongly magnetized matrix. In fact, the aeromagnetic map published by MALAHOFF and WOOLLARD (1966) shows a trough over the Puna District (Fig. 8), an indication of subsurface nonmagnetic material. For the analysis of aeromagnetic data, two sections NN′ and RR′ were selected from the aeromagnetic map to be subjected to the two-dimensional magnetic model-fitting program of TALWANI and HEIRTZLER (1964). For the datum plane the height of the flight paths at 3.6 km (12,000 ft) above sea level was used. For the analysis of section NN′, first the magnetic profile that would be produced by a cold basaltic island mass with the same shape and size as the island topography under section NN′ was calculated. In Fig. 9, the structure labeled « Topography » approximates the island mass under NN′, and the magnetic profile that would result from such structure is plotted as a broken line in the upper graph. The heavy line represents the data profile. Next, parts of the bottom of the island mass were removed so as to obtain a

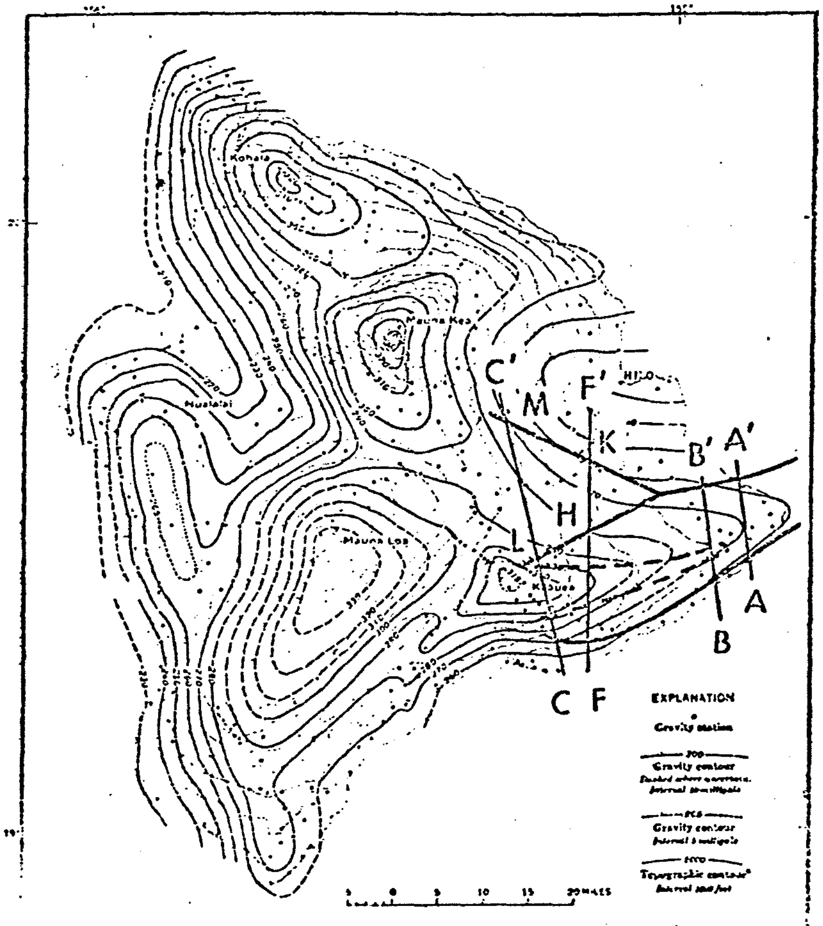

FIG. 6 - Bouguer gravity map of the island of Hawaii (KINOSHITA, 1965), showing sections CC' and FF' that were selected for analysis. Sections AA' and BB' are the same as those of Fig. 3. The heavy lines represent the dike complexes and the dashed lines represent the chimney-like superstructure on the main complex.

magnetic profile that approximates the data profile. After many attempts a best fit model designated as « Model » in Fig. 9 yielded a magnetic profile represented by a. dotted line in the graph. Although there are some discrepancies between the best fit model and the data, the agreement is good, considering that any small change in the island mass results in large fluctuations of the magnetic profile. The best fit model required a nonmagnetic section from 2.4 km below sea level, agreeing with the dike complex, which starts from a depth of 2.3 km. An interesting new discovery was that the undersea portion of the island mass, which presumably should be cool, also showed up as nonmagnetic. The nonmagnetic property of the undersea portion is puzzling, with no plausible explanation in sight. A similar study was done for section RR′, but the model fitting process started out with setting the undersea portion of the island mass nonmagnetic. Figure 10 shows the data profile in heavy line, the island mass profile in broken line and the best fit model profile in dotted line. The best fit model again comes up with nonmagnetic material at a depth of 2.4 km.

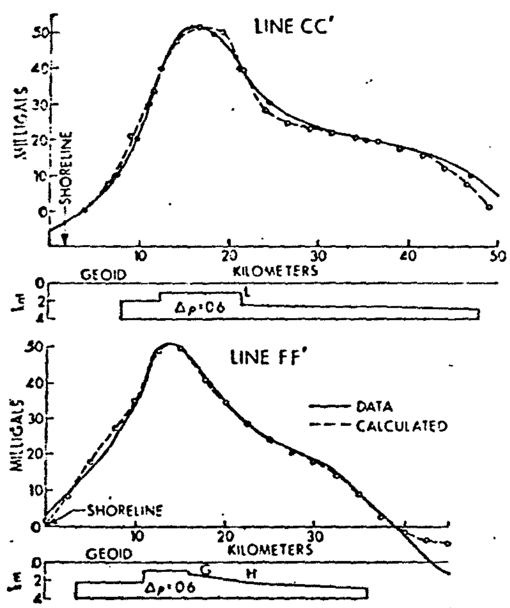

FIG. 7 - Anomalous mass models for sections CC′ and FF′.

Analysis of magnetic data, then, helped in outlining the dike complex under the east rift, especially, the northern boundary. The northern boundary thus outlined is very near point L of the gravity model CC′ and point H of model FF′ in Fig. 7. Because the southern boundary of the dike complex was outlined by gravity methods, a combination of gravity and magnetic methods now provides sufficient information so that a perspective view of the dike complex from the summit area to Cape Kumukahi can be drawn, as shown in Fig. 11, in which sections CC′, FF′, BB′ and AA′ and points G, L, H, K, and M correspond to so mentioned features in Figs. 6 and 7. The dike complex of the east rift is outlined as follows: the southern boundary is marked by the edge along points C, F, B and A; the northern boundary is delineated by points L, H, D, B′ and A′. The triangular sector enclosed by L, M, K, D, H and L includes dense material from the east rift of Mauna Loa and the southeast rift of Mauna Kea. Thus, the dike complex of Kilauea east rift is roughly 20 km wide from the summit to the section BB′, where it starts narrowing to 12 km at AA′.

The submarine portion of the dike complex east of Cape Kumukahi had been outlined previously by MALAHOFF and MCCOY (1976), who used shipboard magnetic data for their analysis. According to them the complex is 12 km wide and extends 60 km east-northeast beyond Cape Kumukahi. The boundaries determined by them matched the boundaries determined by BROYLES *et al.* (in prep) at the seashore. When the land part and submarine part of the dike complex are joined, the lateral dimensions of the dike complex from the summit area to the far end under the sea can be illustrated, as shown by heavy lines in Fig. 8. The total length of the dike complex is 110 km. It is 20 km wide for the first 40 km and then narrows to 12 km for the rest of the length.

MALAHOFF and MCCOY (1976) found that the submarine portion of the dike complex was polarized normally with respect to the present geomagnetic field, while our analysis found the dike complex nonmagnetic under the island. This discrepancy can be readily explained. The submarine portion is much narrower than the land portion and much less frequented by magma excursions from the summit. Because of the smaller size and smaller supply of thermal energy, the submarine portion has cooled below the Curie temperature.

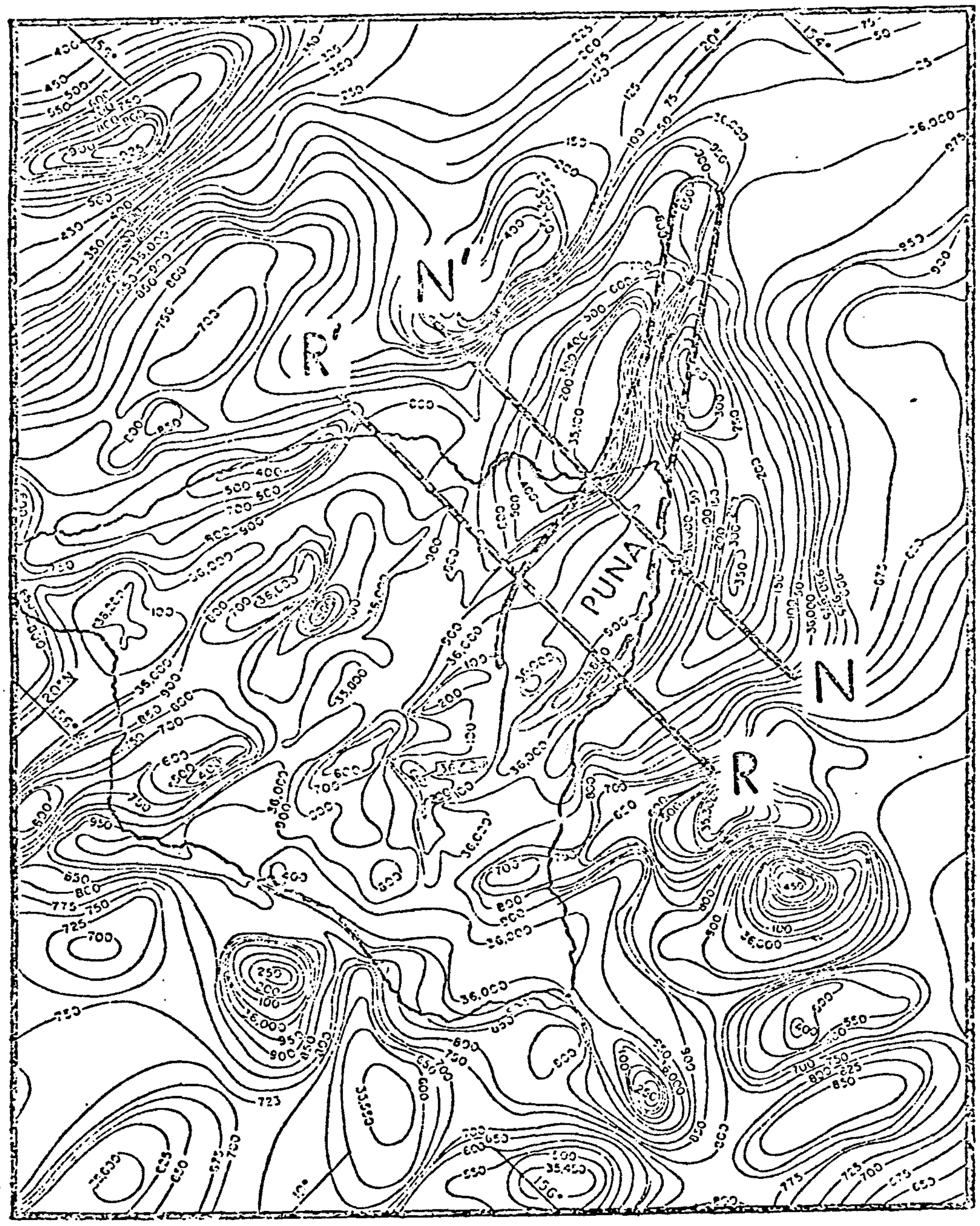

Fig. 8 - Aeromagnetic map of the island of Hawaii (Malahoff and Woollard, 1968), showing sections FF' and RR' which were selected for analysis. The dike complex is outlined in heavy lines.

BOTTOM OF THE DIKE COMPLEX

Current hypotheses on the internal structure of Kilauea Volcano advocate or at least imply that the magma for the rift zones comes from the central holding reservoir through subterranean passageways above the ancient seafloor or oceanic crust (*e.g.*, FISKE and JACKSON, 1972; MACDONALD and ABBOTT, 1970; SWANSON *et al.*, 1976). The rift zones are not primary features, *i.e.*, the rift zones do not have conduits directly connected to magma sources in the deep mantle. The various theories may differ in details but they all agree on the concept of a central reservoir and lateral movement through shallow subterranean passageways. Evidence for shallow passageways comes mainly from earthquake monitoring data.

During the east rift eruptions of 1959-1960, the movement of magma through

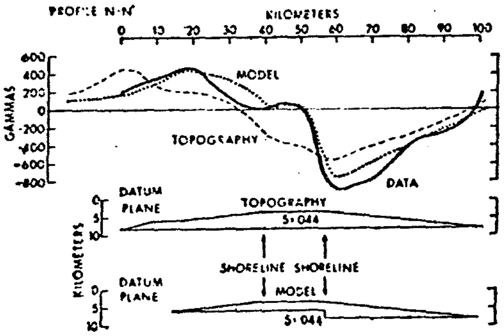

FIG. 9 - Magnetic models that gave rise to magnetic profiles for section NN'. The model designated as « TOPOGRAPHY » corresponds to a cool island mass of basalt sitting on an oceanic crust. The model designated « MODEL » is the best-fit model for magnetic data. Magnetic susceptibility is in m.c.u.

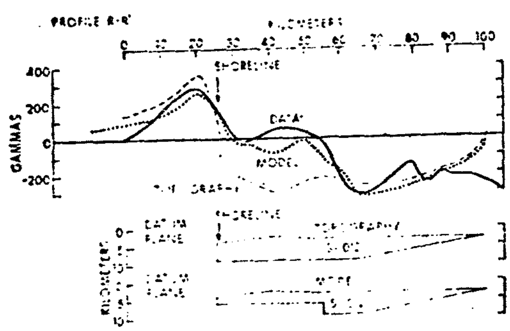

FIG. 10 - Models that gave rise to magnetic profiles for section RR'. The « TOPOGRAPHY » model has the undersea portion removed. « MODEL » is the best-fit model for the magnetic data.

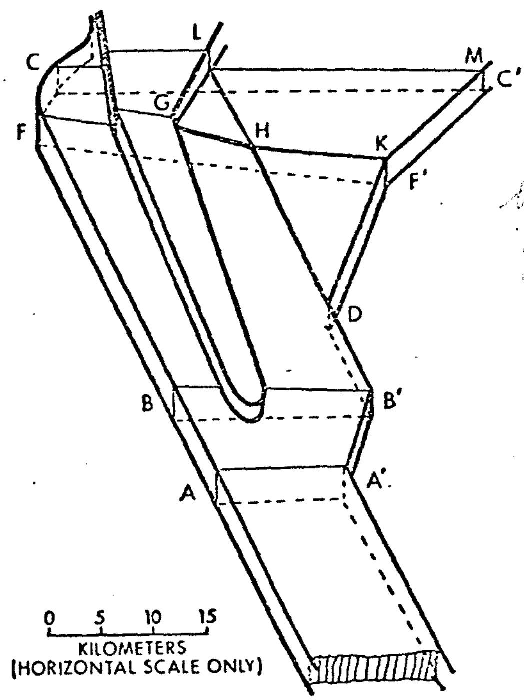

FIG. 11 - A perspective view of the structure of the dike complex from the summit area to Cape Kumukahi. The northern boundary of the Kilauea east rift runs through points L, H, D, B' and A'. The triangular sector of LMKDH is made up of rift material from Mauna Loa and Mauna Kea.

the passageways was tracked by earthquake monitoring and by measurements of ground deformation (RICHTER *et al.*, 1970). Lava eruptions began in the summit area, and then later eruptions appeared at the ends of the rift zone. The shift in activity was accompanied by deflation on the mountain near the summit and the appearance of earthquakes down the rift zone. By earthquake monitoring, volcanologists were able to forecast the general area where new vents were liable to appear.

In 1975, as part of the geothermal exploration program, a network of seismographs was set up to monitor the area near HGP-A well site, about a year prior to the beginning of drilling (SUYENAGA, in prep). A plot of the epicenters of more than 30 earthquakes recorded during a month of monitoring is shown in Fig. 12, in which the black dots represent seismographs and the epicenters are shown by numbers that represent depth of foci in the kilometers. The deepest earthquake occurred at 14-km depth; all but one earthquake occurred at depths shallower than the Mohorovicic discontinuity, which in this area is at 13.3 km (HILL, 1969). When the foci of these earthquakes were projected onto a vertical plane AA', three distinct clusters of earthquakes were seen (Fig. 13). The first cluster outlined an elliptical region around what later became well HGP-A. The second cluster was grouped along a plane dipping 60° to the south, and the third cluster outlined a horizontal plane 5 km deep.

The first cluster of earthquakes are associated with the geothermal reservoir. The earthquakes were probably due to fracturing resulting from groundwater coming into contact with hot rock.

1978

? Months in 1975

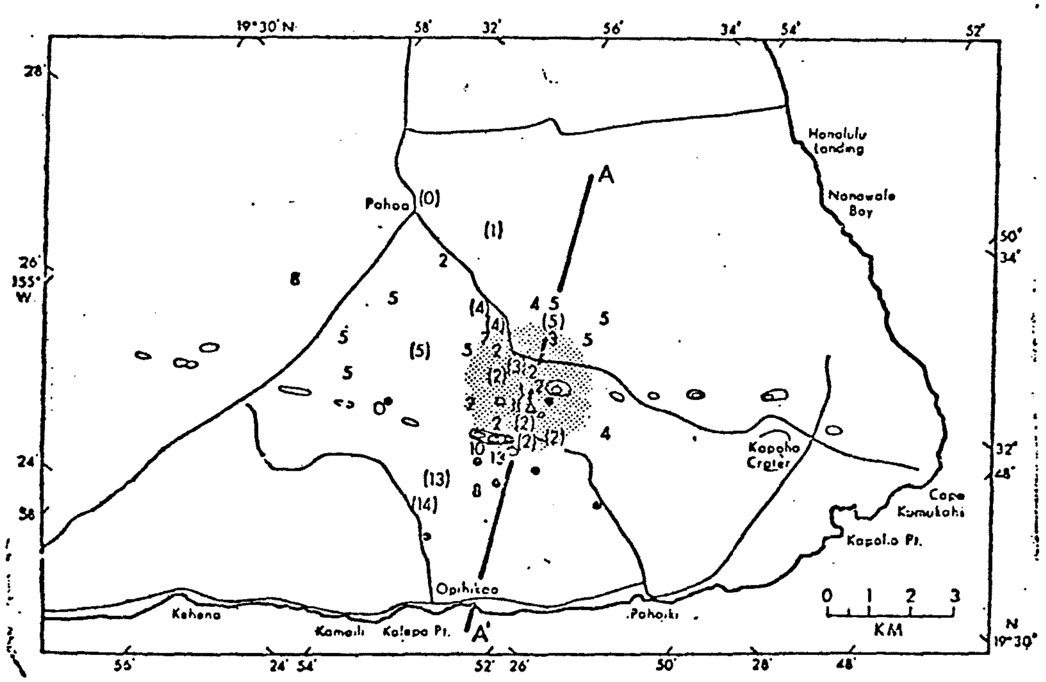

FIG. 12 - Epicenter map of the microearthquakes along the east rift in eastern Puna (SUYENAGA, in prep.). The black dots represent seismometers; the epicenters are marked by numbers that indicate depth in kilometers. Section AA' is the vertical plane corresponding to Fig. 13.

Because the base temperature of the reservoir is 300°C, such fracturing can well be expected. Further discussion of the geothermal reservoir is reserved for another paper.

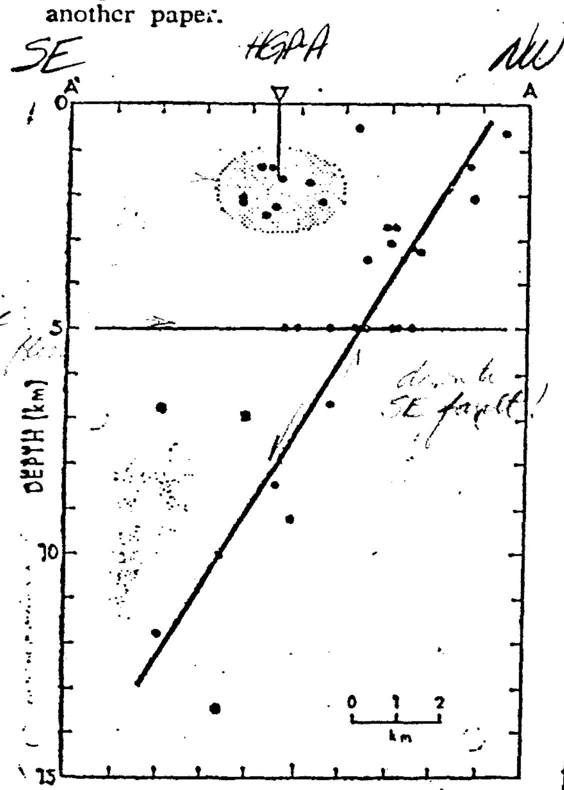

FIG. 13 - Projection of the foci of earthquakes onto the plane AA' (SUYENAGA, (in prep.). The triangle marks the position of HGP-A well.

The second cluster of earthquakes, aligned roughly along a plane dipping 60° to the south-southeast, disclosed an active local tectonic process. Further data on this cluster were provided by by KELLER (1975), who encountered the same 60° dipping plane during a similar microearthquake survey in the area. From first motion data, he was able to obtain a composite source mechanism solution, which is shown as upper hemisphere projection in Fig. 14 (KELLER, 1975). Keller interpreted the mechanism in terms of a series of normal faults with 45° dip arranged en echelon along a plane dipping 60°. A tectonic process to account for such faulting was not proposed.

The source mechanism of Fig. 14 can also be interpreted in terms of a system of tensional and compressional stresses, whereby the tensional stresses are horizontal in the north-south direction, the compressional stresses are vertical, and the neutral axis is horizontal in the east-west direction. Such a pattern of stresses can be produced by forceful intrusion of magma into fissures, which causes dilatation and expansion of the dike complex. These stresses would induce earthquakes in the vicinity of the dike complex, but at distances away from the dike complex the diminishing effect of the stresses would not be able to induce earthquakes. This explains the shallow depths of the earthquakes. In a different

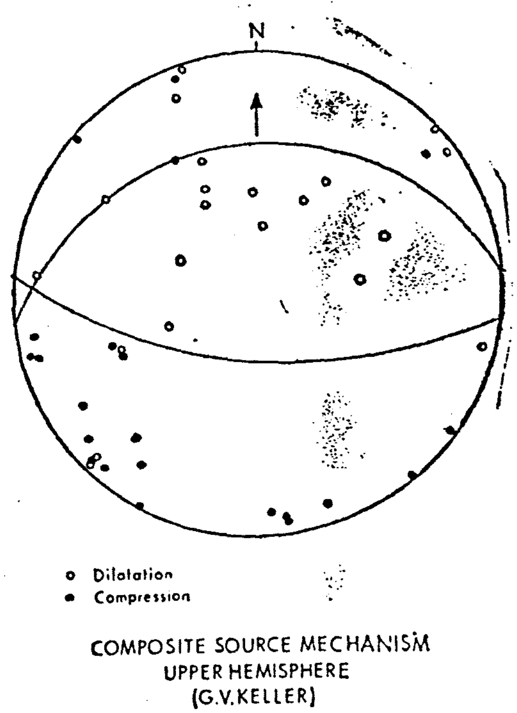

FIG. 14 - Composite focal plane solution of microearthquakes along the east rift by KELLER (1975).

study, KOYANAGI *et al.* (1972) in so many words also proposed the expanding dike complex as the cause of earthquakes that they observed on the south flank of the east rift of Kilauea.

The third group of earthquakes were neatly aligned along a horizontal plane 5 km deep (¹). This leads to the inference that there is some sort of discontinuity at 5-km depth. The most likely discontinuity is the ancient seafloor, which preexisted the birth of the volcano and which is the unconformity between the island mass and the oceanic crust. Even the depth of 5 km agrees well with the expected depth of the ancient seafloor, since the average depth of the Pacific Basin is 5 km.

Further information on the bottom of the dike complex was provided by the Kalapana earthquake of November 29, 1975 (Fig. 1). The parameters of the earthquake were: epicenter 19°21′06″N, 155°01′45″W; depth 7 km; origin time November 29, 1975, 14 h 47 m 42 s (GMT); magnitude 7.2. The analysis of its source mechanism by use of teleseismic and local P wave arrivals and Love wave distribution pattern showed that the slip motion was a low-angle overthrust of the southern flank of the east rift across a fault plane with a 4° slope to the south (FURUMOTO and KOVACH, 1978). The interpretation of the source mechanism in terms of geological structure is shown in Fig. 15. The fault plane with the 4° slope to the south was equated to the unconformity between the island mass and the oceanic crust.

(¹) A suspicion may arise wheter such an alignment was an artifact of the epicenter determination program, since the program used began with an assumed depth. In the analysis reported here, the data were processed twice, first with an assumed initial depth of 5 km and the second run with an initial depth of 10 km. Figure 13 represents the results of the first run. In the second run the earthquakes in question had depths ranging from 4 to 6 km. Hence these earthquakes do cluster around a plane with a median depth of 5 km.

The foregoing analysis results in a model where the ancient seafloor is dipping under the island at 4°, a downwarping of the oceanic crust caused by the load fo the island mass. The depth

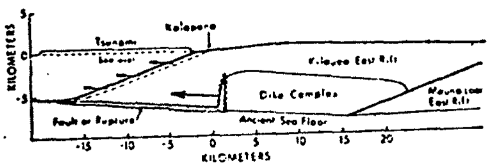

FIG. 15 - The slip vector of the Kalapana earthquake of November 29, 1975, as determined by FURUMOTO and KOVACH (1978).

of the earthquake was given as 7 km, a value 2 km deeper than the bottom of the dike complex determined by microearthquake data. As discrepancies of 2 km can be expected in depth determinations of earthquakes, there is no inconsistency in affirming that the Kalapana earthquake occurred at the ancient seafloor.

TEMPERATURE PROFILE OF DIKE COMPLEX

Temperature profiles logged from well HGP-A showed a high temperature gradient at the bottom of the well. Attempts at extrapolating temperature to greater depths yielded results consistent with the structural picture obtained by geophysical methods, namely, that the dike complex outlined above was hot enough to be above the Curie temperature. In the extrapolation process the heat flux of the area was determined by use of the well temperature profile, and the values of density, thermal diffusivity and conductivity were obtained from core samples from the well. Once the heat flux was determined, that value was held constant; and the extrapolation was done in stages, with use of varying temperature gradients depending upon changes in the other parameters.

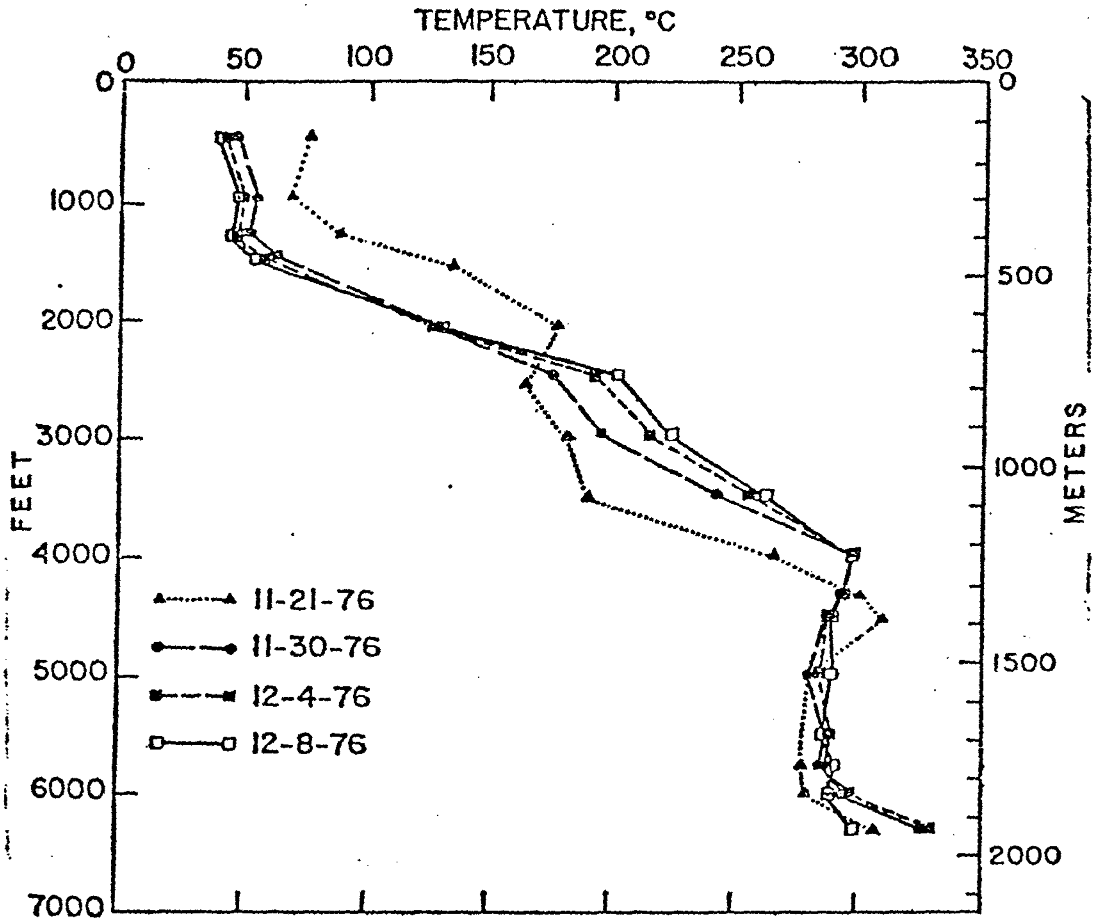

FIG. 16 - Temperature profile of HGP-A after the flow test of November 1976 (YUEN *et al.*, 1977).

Of the many well temperature profiles available, the one logged in December 1976 was selected, because it was farthest removed in time from disturbing operations such as flow tests (Fig. 16). In the figure, the temperature gradients are high between depths of 500 m to 1300 m; below that is an isothermal section; then another gradient section near the bottom. The lithological log of the well mentions that the section from 680-m depth to 1100-m depth is characterized by sealing of fractures in the rock by secondary mineral deposits (PALMITER, 1976; Fig. 13). In this section, conduction dominated convection as the mode of heat transport.

The associated parameters for calculating heat flux are density 2.3 g/cm^3 and heat capacity 0.18 cal/g (MANGHNANI, personal communication). Thermal diffusivity of core samples was found to vary with temperature and depth at sample origin (MANGHNANI *et al.*, 1977).

TABLE 1 - Conduction Heat Flux at Different Levels in Well HGP-A.

Depth (from ground surface) m	Depth (from ground surface) ft	Depth (from sea level) m	Temperature °C	Gradient, °C/m	Diffusivity, cm²/sec	Conduction heat flux, µcal/sec-cm²
1800	549	349	95	4.0×10^{-3}	6.4×10^{-3}	10.5
2300	701	501	157	6.1	6.14	15.4
2750	838	638	212	1.5	6.14	3.8
3250	991	791	242	2.4	6.14	6.1
3750	1143	943	277	2.3	6.22	5.9
6150	1875	1675	307	3.1	7.5	9.6

The conduction heat flux values calculated with the above-mentioned parameters are listed in Table 1. The maximum value of 15.4 µcal/sec-cm² was found at a depth of 701 m, where the rock fissures and fractures were sealed by secondary minerals (PALMITER, 1976). For calculation purposes, we can take this heat flux value as total heat flux.

In the temperature extrapolation process, in the first step the temperature was straightforwardly extended to a depth of 2.0 km (1.8 km below sea level) using the gradient of 3.1×10^{-3} °C/cm observed in the bottom of the well. At 1.8 km below sea level, the temperature would be 364°C. At this depth, seismic refraction survey encountered a rock layer with a seismic P-wave velocity of 5.25 km/sec, which corresponds to a density of 2.6 g/cm³, according to the velocity-density relationship by MANGHNANI and WOOLLARD (1968). Electrical resistivity values became very high, so that « infinity » was assumed for this solutions. Also, the temperature of 364°C is not far from the critical temperature of water, which is 374°C. Hence, in the extrapolation process it was decided to consider that below this depth heat transport was due almost entirely to conduction. With heat flux values of 15.4 µcal/cm² sec, density 2.6, and specific heat capacity of 0.18 cal/g, a temperature gradient of 4.4×10^{-3} °C/cm was derived. When temperature was further extrapolated with that gradient value, at 2.2 km below sea level, the temperature became 540°C, and at 2.3 km it became 583°C (Table 2).

As the value of 540°C was found to be the apparent Curie temperature for Hawaiian rock, as measured over a cooling lava lake (ZABLOCKI and TILLING, 1977), it was concluded that magnetic properties vanished at 2.2-km depth. As 2.3 km was found to be the upper boundary of the dike complex from refraction data, it was concluded that the dike complex was hot enough as to be nonmagnetic.

Seismic velocity changes to a value of 7.0 km/sec at a depth of 2.3 km, according to refraction data, and the density also changes to 3.1 g/cm³. This entailed a recalculation of the temperature gradient to 3.7×10^{-3} °C/cm. When that gradient was extrapolated, the temperature of 1060°C, which is considered to be the melting point of Hawaiian rocks, was reached at a depth of 3.6 km below sea level. The extrapolated temperature

705°F

3 temps 374 705
540 1000
1060

profile and its relation to structure is shown in Fig. 17.

The special feature about the dike complex which was found to be a large structural unit, is the high seismic P_i-wave velocity, 7.0 km/sec, measured by seismic refraction methods. High seismic velocity has been observed previously: 7.7 km/sec has been measured for a layer at 5-km depth under the northeast rift of Koolau Volcano on the island of Oahu (FURUMOTO *et al.*, 1965). From the empirical velocity-density relation of Hawaiian rocks (MANGHNANI and WOOLLARD, 1968), a velocity of 7.0 km/sec corresponds to a density of 3.1 g/cm³.

TABLE 2 - Thermal Parameter Variation with Depth.

Depth (from sea level), m	*Temperature,* °C	*Density and Temperature Gradient*
1715	325	$\rho = 2.3$ g/cm³
		$\frac{\Delta T}{\Delta z} = 3.1 \times 10^{-3}$ °C/cm
1800	364	$\rho = 2.6$ g/cm³
2201	540	$\frac{\Delta T}{\Delta z} = 4.4 \times 10^{-3}$ °C/cm
2300	584	
3600	1060	$\rho = 3.1$ g/cm³
		$\frac{\Delta T}{\Delta z} = 3.7 \times 10^{-3}$ °C/cm

Gravity data also demand a dense anomaly under the east rift. To account for the gravity data, a density contrast of 0.4 g/cm³ to 0.6 g/cm³ is necessary for the anomaly. As the rock samples from HGP-A well in the range of 1.5-km depth already weighed 2.56 g/cm³ on the average for bulk density (MANGHNANI *et al.*, 1977), the density of the material in the dike complex would be 2.96 to 3.16 g/cm³. Therefore, from seismic refraction data and from gravity analysis, we have evidence for very dense material at a shallow depth of 2.3 km. It is difficult to conceive of such heavy mineral in bulk at such shallow depths. This poses an interesting puzzle to petrologists.

The puzzle remains unsolved when the following is considered. Up to the present time, physical properties of Hawaiian rocks under high pressure and temperature have been obtained by taking lava samples that were extruded on land or on the ocean floor and then subjecting them to tests in the laboratory. These samples were extruded lava, not magma samples from depth obtained by drilling. It is proposed here that extruded lava comprises the lighter components of the original magma, and that the heavier residues have remained trapped underground, even in the case of the east rift eruptions.

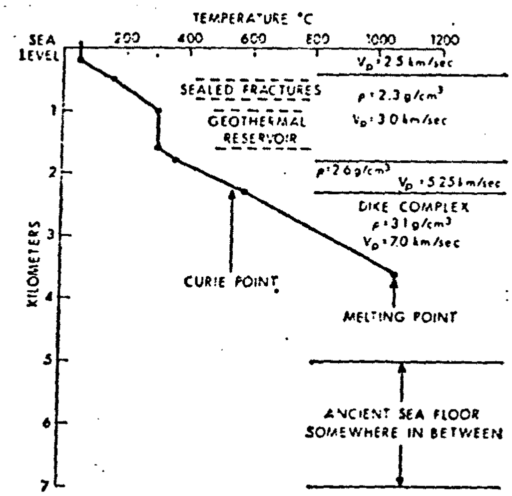

FIG. 17 - Thermal profile of the east rift and associated structures.

FUJII and KUSHIRO (in press) have taken lava samples and subjected them to various pressures and temperatures to measure the parameters of viscosity and density. At 1060°C and 1 kilobar pressure, which corresponds to 3-km depth, the lava sample had a density of 2.63 g/cm³. This is a great difference from 2.96 to 3.16 g/cm³, which geophysical data demand. At this point the following conceptual equation is proposed:

$$\text{Magma (? g/cm)}^3 \rightarrow \text{lava (2.63 g/cm}^3\text{)} + \text{residue (3.1 g/cm}^3\text{)},$$

to account for the various phenomena found in field and laboratory data. The magma from the summit area differentiated at shallow depths in the rift conduits, so that the less dense components erupt through surface vents, while the dense components remain trapped as residue. The residue can be a very small percentage of the parent magma. At the present time, petrologists have observed and measured the parameters of only the less dense components; they have not been able to acquire a sample of the residue. We volcanologists do not know what the residue is, nor do we know completely the nature of the parent magma.

The above argument is limited to oceanic intraplate volcanism of the Hawaiian type. Although we have seen exposed parts of ancient volcanic rift zones, such as those on the rocky coast of the island of Niihau (DALRYMPLE *et al.*, 1973, fig. 5), these exposed rocks were originally emplaced at depths above sea level. So far, no example has been found in the Hawaiian Islands of an outcrop or exposure of rocks emplaced by intrusion at 2.3 km below sea level or deeper. Comparative petrological study with Skaergaard intrusion of East Greenland may prove enlightening, but it is beyond the scope of this paper.

Density data also cast doubt on the existence of a magma body in the order of kilometers in diameter under the east rift. At a depth of 3 km, the fluid lava would have a density of 2.63 g/cm³ (FUJII and KUSHIRO, in press). With a density contrast of roughly 0.5 g/cm³ with respect to the surrounding rocks and with low viscosity in the range of 100 poises (KUSHIRO *et al.*, 1976), the fluid would tend to rise by upward melting or forceful intrusion into existing fissures or zones of weakness. Analytical studies and model experiments of HARDEE and LARSON (1977) showed that the upward migration would be rapid in either process. The fluid would reach the ground surface in a matter of days to a few years.

Even if a pocket of fluid material were trapped below by an effective roof, the existence of such a pocket is easily detectable by gravity methods. A sphere of 300-m diameter with a density contrast of 0.5 g/cm³ would produce a negative gravity anomaly of 3 milligals. In the Bouguer gravity map of Figure 3, the undulations in the otherwise smooth gravity contour lines are, at most, variations of 0.5 milligals. Gravity data do not show evidence of fluid pockets as large as 300 m in diameter.

It is unikely that the dense residue in the dike complex is in the fluid state. As it is difficult to even conceive of a solid mineral in bulk as having a density of 3.0 to 3.1 g/cm³, it is all the more difficult to conceive of a fluid as having such density.

If any form of magma remains underground, it would be a fluid filling the cracks and fissures but not large pockets. Except in periods of eruptive activity, fluid material would amount to, at most, only a few percent of the mass of the dike complex.

In summary, the dike complex has a high temperature, well above the Curie temperature, and its high density causes speculation about the nature of its material. Geophysical data and laboratory tests on lava samples support the thesis that material extruded in lava eruptions is the less dense components of parent magma and that dense residue remained underground. Only by drilling down to 3 or 4 km depths will we be able to retrieve samples of the dense residue. Then we will be able to determine the composition of the parent magma.

CONCLUSION

Geophysical surveys of the geothermal exploration program provided the necessary details to yield a coherent, clear picture of the structure and thermal processes of the east rift of Kilauea Volcano.

The edifice of the east rift was built up by successive lateral magma excursion from the summit holding reservoir trough subterranean passageways. These passaggeways, first formed by gravitational slumping, were thin vertical cracks into which magma forcefully injected itself and by its hydrostatic pressure forced the cracks to propagate. Magma made its own passaggeway down the rift zone, although it was controlled by the gravitational stress field. The less dense components of the magma worked upward to erupt as lava through the numerous fissures and vents. The more dense components remained as residue, filled the cracks and solidified to form dikes, so subsequent magma excursions wedged into new cracks to form new passaggeways.

During the early stage of construction of the east rift the passageways were to the north against the flanks of the older volcano of Mauna Loa. Northward migration of the passaggeways was blocked by Mauna Loa, so migration shifted southward. Eventually, after geological ages, the numerous dikes formed and left by successive magma activity built up a dike complex 20 km wide near the summit. At the present time the 20-km width extends to a distance of 40 km from the summit. There the width narrows to 12 km.

The dike complex extends for about 110 km. Half of the length is on land and half of it is under the sea. On the whole, the top of the dike complex is about 2.3 km below sea level. Closer to the summit area is a chimney-like superstructure which is only 1 km below sea level. The bottom of the dike complex rests on what was the ancient seafloor before the birth of the volcano. The volcano edifice, including the dike complex and its flanks, sits on the oceanic crust. Because of the weight of the volcanic edifice, the oceanic crust has subsided. At Kalapana, the subsidence has produced a slope of 4° on the oceanic crust.

The temperature of the land portion of the dike complex is above the Curie temperature of 540°C. This condition produces a low magnetic profile over the east rift of Kilauea. The temperature of the inner part of the dike complex is well over 1000°C. In spite of the high temperature, the material is probably not fluid, but rather in a glassy state. As we have not obtained a rock sample from the dike complex we can only guess what the material could be. Density considerations point to a rock with density of 3.0 to 3.1 g/cm³. Minerals of such high density are not usually encountered at such shallow depths in a basaltic province.

Repeated recent eruptions along the east rift reveal that the dike complex is still growing. The continued growth produces tensional stresses that produce earthquacks in the rock layers above and below the dike complex. In the overlying layers, tensional earthquakes produce fissures. Groundwater, percolating into the fissures, is heated so that geothermal fields are produced.

Layey
sedimentary

The geothermal exploration program gave us not only knowledge about geothermal fields, a topic to be treated in another paper, but also valuable information on the structure and thermal process of rift zones associated with oceanic basaltic volcanoes.

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