# GEOPHYSICAL RECONNAISSANCE OF PROSPECTIVE GEOTHERMAL AREAS ON THE ISLAND OF HAWAII USING ELECTRICAL METHODS

by

Jim Kauahikaua
U.S. Geological Survey

and

Mark Mattice
Hawaii Institute of Geophysics

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## ABSTRACT

Resistivity data from several areas were compiled, analyzed, and interpreted in terms of possible geologic models. On the basis of this analysis alone, two areas have been ruled out for possible geothermal exploitation, two have been interpreted to have a moderate-temperature resource, and two have been interpreted to have a high-temperature resource.

The two areas which have been ruled out are the Keaau and South Point areas. The Kawaihae area and the lower northwest rift zone of Hualalai appear to have anomalous resistivity structures, which suggest a moderate-temperature resource in each of these areas. Finally, specific areas in the lower southwest and lower east rift zones of Kilauea have been outlined as locations where high-temperature fluids may exist at depth.

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## INTRODUCTION

Hawaii is the largest and youngest of the Hawaiian islands and is located at the southeastern end of the chain. It is composed of the products of five volcanoes (see Fig. 1). Kohala and Mauna Kea are the northernmost and oldest; Hualalai, Mauna Loa, and Kilauea dominate the southern half of the island and are the youngest. All three have been active in historic time, although only Mauna Loa and Kilauea have remained active to the present day.

Active volcanoes are abundant sources of heat and would be natural targets for geothermal exploration if surface manifestations of heat alone were an adequate heat resource; however, the real resource in volcanoes is the heat that may be trapped in their interiors. The source of this heat is the numerous bodies of initially molten magma which are intruded into the summit and rift zones of active volcanoes. Although large amounts of heat can be expected beneath active volcanoes, significant amounts of heat may still be trapped beneath older volcanoes. In addition, the lower risk of damage to resource development from volcanic activity also makes older volcanoes more desirable targets.

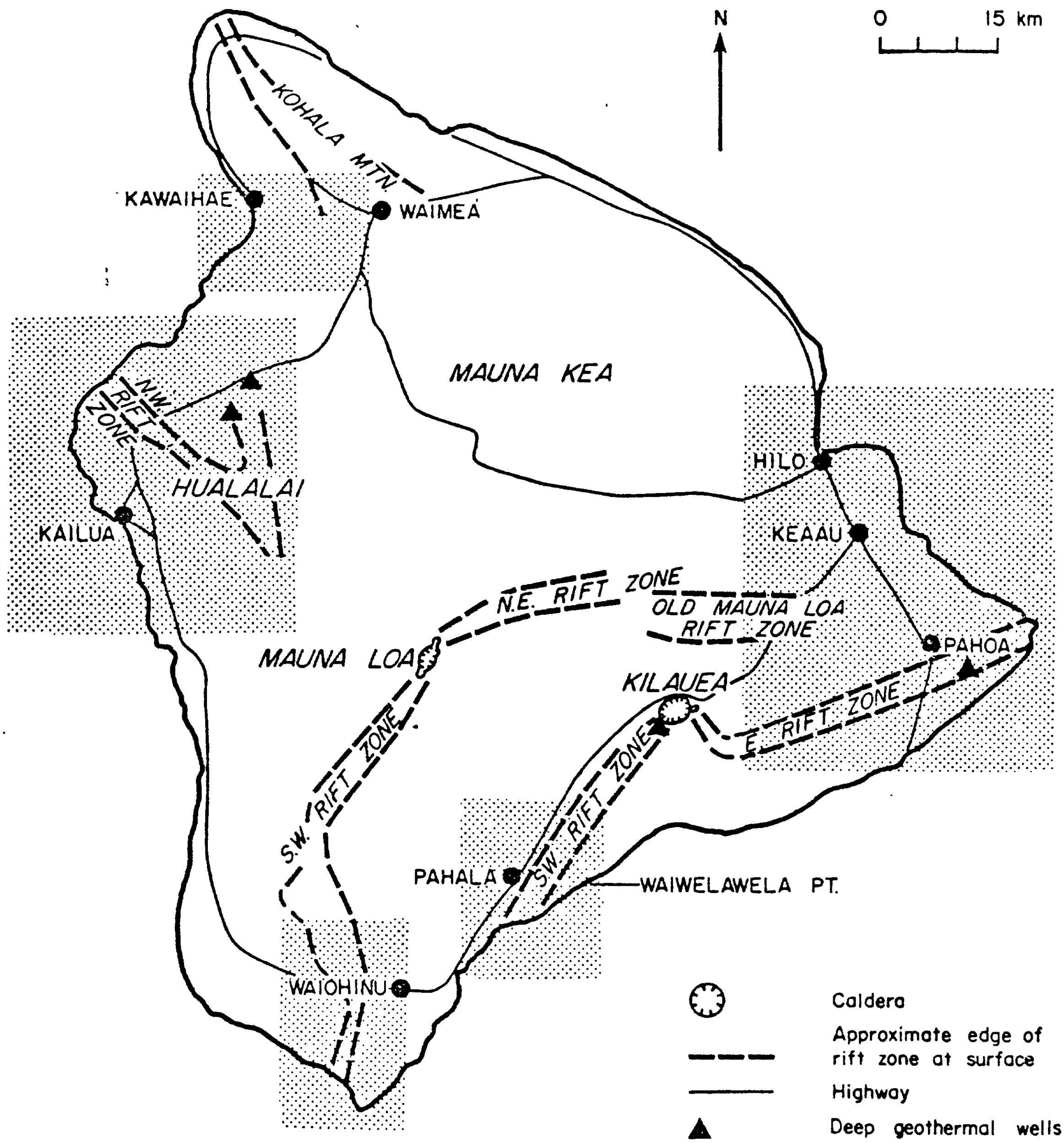



Figure 1: Map of the island of Hawaii showing prominent volcanic structures and the locations of the subareas (stippled) discussed in this report.

In practice, a geothermal resource requires not only heat, but also abundant water with which to extract the heat from the ground. Hawaii groundwater comes from two basic sources: rainfall which percolates down through porous volcanic rock to a basal water body, and seawater which infiltrates the porous island below sea level and is present beneath most of the island. The basal freshwater floats on the underlying seawater in an equilibrium which can be disrupted by excessive pumping of wells, low recharge from rainfall, or higher-than-normal temperatures.

Although most groundwater on the island is in a basal configuration, it is the water confined within dike complexes which is probably most important with regard to an area's geothermal potential. The normal basal lens is never more than 300 or 400 m thick because water is free to flow laterally and is eventually lost to the ocean; however, dike-confined water can achieve much larger dimensions owing to its confinement. In many areas, the water is impounded to such great elevations above sea level, that it is difficult to imagine that seawater still exists at the great depth required for support. Of the two deep geothermal wells drilled into Kilauea, the one on the summit encountered relatively freshwater at an elevation of 700 m above sea level, and the other on the lower east rift zone encountered relatively freshwater throughout its length even though it was drilled to 1750 m below sea level. Both wells must have been located within dike-confined water bodies; both were anomalously warm.

It is because water is such an important part of the resource that electrical geophysical methods are so well suited for geothermal exploration. The electrical properties of most rocks are ordinarily a function of the electrical properties of the water in its pores; anomalously low electrical resistivity can be a direct indicator of high pore-water temperature, or salinity, or high rock porosity.

This paper assesses several prospective geothermal areas on Hawaii primarily using electrical geophysical data. Emphasis is placed on the results of a recent survey, although data from previous investigations are reviewed. Six areas are discussed; they are the area near the town of Kawaihae, Hualalai volcano, the south rift of Mauna Loa, the southwest rift of Kilauea, the area around the town of Keaau, and the east rift of Kilauea (see Fig. 1).

# 2

## DESCRIPTION OF DATA AND ELECTRICAL EXPLORATION TECHNIQUES

The data used in this assessment comes from three sources. Primary emphasis is placed on the results of a survey completed during the summer of 1979 by a crew of USGS and Hawaii Institute of Geophysics (HIG) staff. Twenty-six vertical electric soundings (VES), thirteen time-domain electromagnetic (TDEM) soundings and one self-potential profile from this survey are presented for the first time. The findings are coupled with those of a similar survey carried out by HIG staff during the summer of 1974 which used VES, TDEM, equatorial DC soundings and dipole-bipole techniques (Klein and Kauahikaua, 1975). The third source of data is a series of groundwater exploration surveys done by Water Resources Research Center (Univ. of Hawaii) staff during the late 1960's. These surveys made use of local aeromagnetic and gravity coverage as well as various electrical techniques.

As already mentioned, most of the electrical data were obtained with two techniques: VES, using the Schlumberger electrode configuration, and TDEM sounding, using a horizontal grounded-wire source and a vertical magnetic-field sensor. Both methods yield information on resistivity variations with depth; however, in our present application, each method was used to sound different ranges of depths. Any one technique would require an unrealistically large range of either spacings (for a VES method) or frequencies or times (for an EM method) to adequately resolve the whole range of resistivities present within the first few kilometers beneath a typical area of Hawaii. The largest resistivity contrast is between rocks that are saturated and those that are unsaturated with groundwater; rocks above the water table may have typical resistivities in the thousands of ohm-m whereas those below may have resistivities lower than 5 ohm-m (Zohdy and Jackson, 1969; Kauahikaua and Klein, 1977a). Because of the inherent sensitivity of EM techniques to low resistivities, the TDEM technique was used to resolve resistivities within the saturated rocks; the VES technique was used to resolve resistivities within the unsaturated range of depths and to determine the depth to saturated rock.

## 2.1 VERTICAL ELECTRIC SOUNDING METHOD

In any direct-current (DC) technique, the depth of investigation is a function of the distances between the current electrodes and the voltage-measuring electrodes, and the resistivity structure of the ground. The basic relationship is that the greater the distance between current and voltage electrodes, the greater the depth of investigation. The field application of the Sclumberger VES method requires that a pair of current electrodes be progressively expanded in opposite directions about a point at which two more closely-spaced electrodes are being used to measure the resulting voltages. Data from larger current-electrode expansions provide information from greater depths.

Our field method consisted of expanding the current electrodes so that one-half the current electrode separation varied from 3 m (10 ft) to 914 m (3000 ft) at logarithmically-equal intervals. Ten spacing expansions per decade was the maximum data density for each sounding. The voltage-measuring electrodes were expanded twice per decade. All four electrodes were in line and the distance between the current electrodes was at least five times the distance between the voltage electrode spacing. By maintaining this minimum ratio between current and voltage electrodes, the maximum measurement error, due to a finite voltage-electrode separation, is less than 13% (Mundry,1980).

At each spacing an apparent resistivity was computed from the measured voltage, the known current, and the known distances between electrodes (Keller and Frischknecht, 1966, p. 95). The apparent resistivities at each electrode spacing were then used as input to a computer program which could find an earth model made of discrete, horizontal layers whose response fit the data in a least-squares sense (Anderson, 1979). The number of layers in the model was increased and the program was rerun until the match between data and model response was no longer improved significantly. At this point, estimates of the errors in the determined parameters (resistivities and thicknesses of each layer in the model), as well as estimates of the correlation between parameter errors, were computed (Inman, 1975). The final interpretation was based not only on the parameters of the final model, but on an analysis of the parameter error and correlation estimates as well.

The data for each sounding were plotted as apparent resistivity versus current electrode half-separation (denoted as AB/2) on log-log graph paper. If fences or pipe lines were crossed during a sounding, the point of crossing was noted on the plot. The interpreted model is shown at the bottom of the plots in bar form with resistivity, thickness and the respective errors specified for each layer. Those resistivities or thicknesses which do not have errors specified were poorly resolved (standard deviations greater than the parameter value) or were not allowed to vary during the fitting process (signified by an equal sign with three dashes). Several soundings did not require detailed interpretations in order to extract the information relevant to geothermal assessment; only the data are plotted for those soundings. Finally, because most soundings reflect the local hydrology, sea level is indicated by a dashed line in each sounding.

In practice, the VES method, as described above, works well for delineating the sequences of resistive, unsaturated rock units to depths of 600 to 800 m. However, when water-saturated rocks are encountered, usually at depths less than 600 m, the resistivity is generally poorly resolved due to the high contrast between unsaturated and water-saturated rock resistivities. On the island of Hawaii resolution of the saturated rock resistivity with the VES technique has been generally poor, but significant improvement in resolution would require electrode expansions of at least eight times the elevation (Roy and Apparao, 1971). Doing soundings at lower elevations means doing soundings near the ocean's edge, and the effect of a low-resistivity ocean would have to be considered. The ocean has the effect of decreasing resolution since electrical current would have a tendency to flow laterally into this extremely conductive medium rather than penetrating the rock layers below (Mundry and Worzyk, 1979). By contrast, EM soundings need source-sensor distances of only one or two times the elevation to resolve saturated rock resistivities. Thus, an EM method is used for exploring the restivities of the saturated rocks.

## 2.2 TIME-DOMAIN ELECTROMAGNETIC (TDEM) SOUNDING METHOD

EM methods exploit the time-varying nature of electromagnetic fields produced either by natural sources (geomagnetic field variations, lightning, etc.) or by artifical, controlled sources to penetrate the earth. The depth of investigation of these methods is primarily a function of the frequency of the EM field and the resistivity structure of the ground, and secondarily a function of the distance between the observation point and the EM source. For natural source methods, the source of the EM field is generally assumed to be hundreds of kilometers from the survey area so that the EM field variations take the approximate form of a plane wave impinging upon the earth. Penetration is independent of the distance to the source for this case (Keller and Frischknecht, 1966).

For controlled-source methods, observations of the field are made at distances of much less than 100 km from the EM source. The depth of penetration of the EM energy is still predominantly a function of its frequency; however, it is more difficult to accurately measure the changes in the transmitted signal brought about by the earth's resistivity structure at distances from the source of less than one wavelength. A wavelength of EM energy in the earth is a function of the resistivity structure and the frequency only. Therefore, application of the controlled-source EM technique would be optimum if the range of source-sensor distances and

EM frequencies were chosen such that a sensor was never significantly closer to the source than the shortest wavelength used.

For all EM methods, the depth of investigation is larger for EM energy at lower frequencies than it is for EM energy at higher frequencies. When using a controlled EM source, energy at several frequencies can be transmitted simultaneously and the resulting transient EM field can be recorded and analyzed as a function of time rather than frequency. For these so-called time-domain EM (TDEM) methods, the depth of investigation is larger at later times (relative to the start of the source signal).

The TDEM sounding method was used in both the 1979 and the 1974 surveys but in slightly different applications. In both surveys, the vertical magnetic field produced by a horizontal, grounded wire was recorded as a function of time. Several sources were used; each was 400 to 3000 m long and was pulsed with up to 15 amps of electrical current for 12 seconds out of every 24 seconds. Data were recorded less than 10 km from each source. The difference between the two surveys was in the method of magnetic-field detection. The 1974 survey used a horizontal wire loop, which is sensitive to the time-derivative of the vertical magnetic field; the 1979 survey used a cryogenic magnetometer, which is sensitive to the magnetic field directly.

The TDEM data were also analyzed in different ways for each survey. Data from the 1974 survey was manually digitized from paper records made in the field; several TDEM responses were stacked for each sounding; each sounding was digitally corrected for the imperfect response of the recording electronics; finally each sounding was interpreted using a computer program similar to the one described for VES sounding interpretation (Kauahikaua, 1980). Full details of the data acquisition, reduction, and analysis may be found in Klein and Kauahikaua (1975), Kauahikaua and Klein (1977b), and Kauahikaua (1981b). Data from the 1979 survey was recorded digitally in the field. Reduction consisted only of stacking several responses for each sounding. The reduced data were then transformed to apparent resistivities versus apparent depths of penetration; apparent depth of penetration is defined as the deepest possible vertical penetration for a horizontally-layered earth with a perfectly conductive basement (Kauahikaua, 1981a). Because of the reconnaissance nature of the survey, it was decided that detailed interpretation would not be necessary. An adequate picture of the resistivity structure could be obtained from a qualitative evaluation of the data in this form.

The TDEM soundings were effective in determining the saturated rock resistivities in several areas. An earlier application of the technique at the summit of Kilauea volcano (Jackson and Keller, 1972) delineated low resistivities between 900 and 2100 m below the surface which were thought to indicate increased temperatures at these depths (Keller and Rapolla, 1976). The 1974 survey confirms a similar model for the East Rift Geothermal Area (Kauahikaua, 1981b). The 1979 TDEM results were not as densely spaced in other areas and, as a consequence, structures were less clearly resolved. The 1979 data were obtained at greater distances from the source than earlier surveys and probably resolved deeper structures. For example, all 1979 soundings showed a tendency for increased resistivities below 3 km depth; earlier TDEM data only resolved resistivities less than 2 km deep and did not detect this generally-present strata. The details of each sounding will be discussed in the appropriate sections.

3

# RESISTIVITY STRUCTURE AND GEOTHERMAL IMPLICATIONS IN SIX AREAS

## 3.1 THE CHOICE OF EXPLORATION AREAS

The choice of the areas for exploration that we discuss in this report was based in general on previous geothermal assessments by Macdonald (1973) and Thomas and others (1979). Further specific criteria pertinent to the choices are as follow:

1. proximity to eruptive centers which have been active within the last 200 years,
2. existence of either anomalous well temperatures or infrared anomalies on the ground or in coastal waters, and
3. maximum depth of 2000 m to rock saturated with water.

A brief explanation of criterion 3 may be useful, as it is based on the current capabilty of drilling contractors and on the known depth of geothermal resources that have been drilled in Hawaii. The maximum drillable depth for rigs in Hawaii appears to be less than 3000 m. In both geothermal holes drilled on Kilauea so far, elevated temperatures were reached only after drilling at least 1000 m below the surface. Therefore, it seems likely that only prospects at less than 2000 m elevation will be recoverable in the near future and that this should be a criterion for reconnaissance-level exploration.

The six areas chosen are the following (in the order that they will discussed):

1. the Kawaihae area,
2. Hualalai,
3. the southwest rift zone of Mauna Loa (South Point),
4. the southwest rift zone of Kilauea,
5. the Keaau area, and
6. the east rift geothermal area of Kilauea (Puna).

The areas are outlined on the map in Fig. 1 .

## 3.2 KAWAIHAE

The Kawaihae exploration area is centered on highway 19 between the towns of Kawaihae and Waimea (see Fig. 2). The area along the highway and to the south has a number of water wells whose temperatures range from 26° to 28° C. One well has a temperature of 37° C. There are no known structures within this area which might be responsible for the heat, although the area is bordered by rift zones and cinder cones of both Kohala and Mauna Kea volcanoes. The youngest of these cones appears to be Puu Loa on Kohala with a K/Ar age of 80,000 years (Malinowski, 1977).

Few geophysical surveys had been done previously in the area. A DC Wenner resistivity profile run along the coast

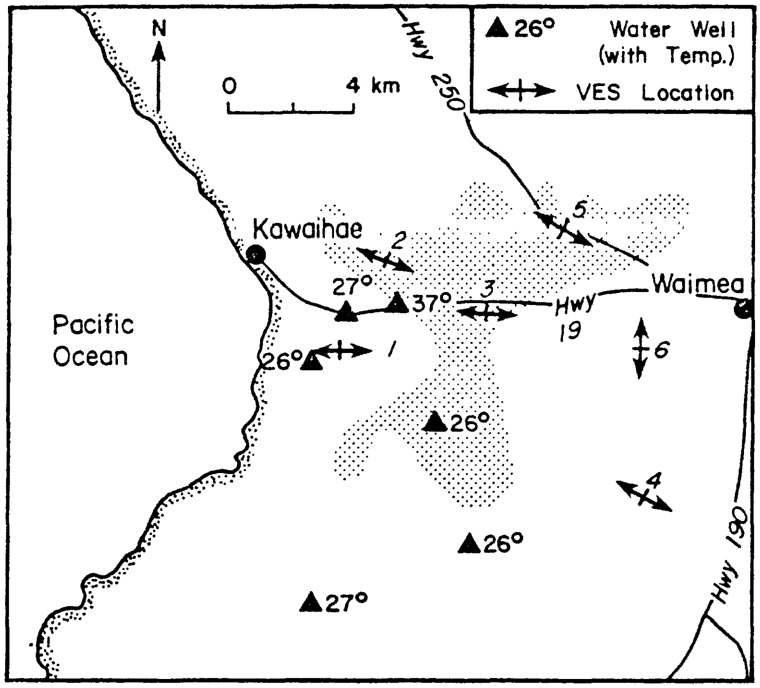
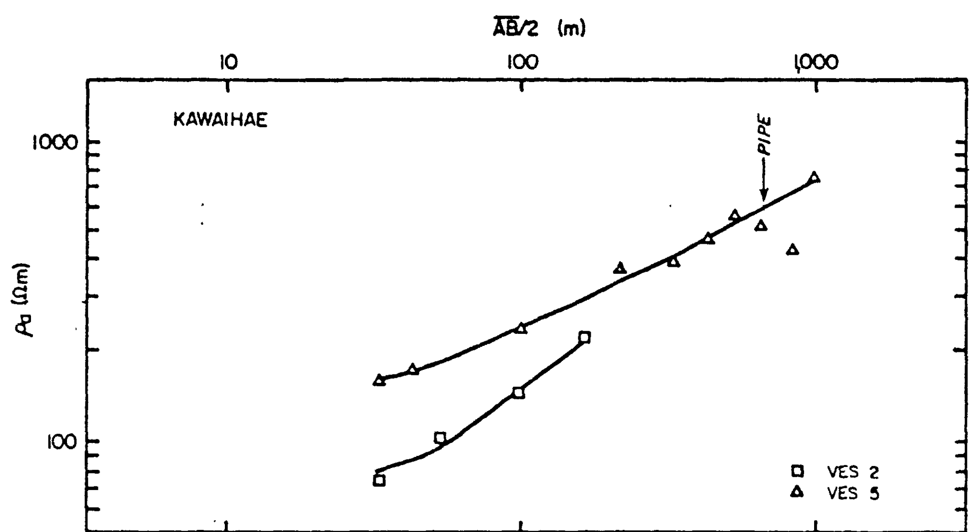



Figure 2: Map of Kawaihae area showing prominent volcanic features, and VES locations. The stippled area is a magnetic high (greater than 36,200 gammas) observed by Malahoff and Woollard (1965).

between Kawaihae and Mahukona to the north (Adams, 1968) suggests that water-saturated rocks near Kawaihae are more conductive (less than 10 ohm-m) than those to the north (greater than 10 ohm-m). This might be explained by warmer seawater or, alternatively, a larger proportion of seawater saturating the rocks nearest Kawaihae.

From an audiomagnetotelluric (AMT) profile along highway 190 and a total-field aeromagnetic survey (flown at 1200 meters), Adams and others (1969) concluded that a deep lateral change in resistivity takes place about 16 km south of Waimea along a roughly east-west boundary. The Waimea Plain to the north appeared to be underlain by more resistive strata than to the south; the higher resistivities were attributed to "the higher resistivity of Mauna Kea lavas or to a higher water table . . ."

The aeromagnetic data show a complex anomaly, which is part of the larger Kohala volcanic pipe anomaly (Malahoff and Woollard, 1965), crossing Hwy 19 west of the warmest well. Six Schlumberger soundings were situated in and around this magnetic high to see whether the body responsible for the magnetic anomaly might be detected by resistivity methods. The sounding locations are shown in Fig. 2 and the apparent resistivity data and interpreted resistivity sections are shown in Fig. 3 and 4.




| | | |
|---|---|---|
| $\rho_1$ = 71 Ωm ± 19%<br>$d_1$ = 37 m ± 36% | $\rho_2$ = 830 Ωm ± 86% | VES 2 |

| | | | |
|---|---|---|---|
| $\rho_1$ = 111 Ωm ± 64%<br>$d_1$ = 21 m ± 110% | $\rho_2$ = 400 Ωm ± 34%<br>$d_2$ = 190 m ± 52% | $\rho_3$ = 1450 Ωm ± 31% | VES 5 |

Figure 3: VES obtained on Kohala lavas. The approximate position of a pipe crossed during sounding is noted for VES 5.

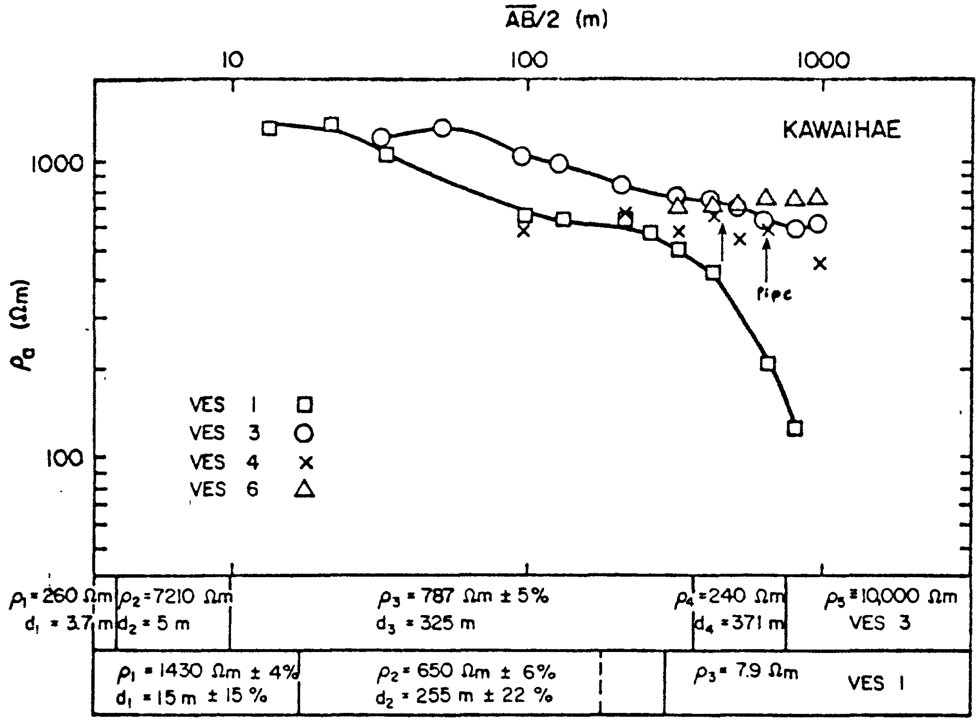



Figure 4: VES obtained on Mauna Kea lavas. The relatively small amount of data for VES 4 and 6 precluded quantitative interpretation.

The soundings appear to be of two basic types depending on whether they were on Kohala or Mauna Kea lavas. VES 2 and 5 were located on Kohala lavas; each is dominated by an increase in apparent resistivity with increasing AB/2. VES 1, 3, 4, and 6 were located on Mauna Kea lavas; in each of these soundings, the apparent resistivities either increase slightly or decrease with increasing AB/2.

The difference between the two groups appears to be confined to the surface material since rocks below 30 m are uniformly between 650 and 850 ohm-m in all soundings. VES 1 is the only sounding which has a descending branch at the largest spacings. This is interpreted as the top of a low-resistivity layer, thought to be the normally-found case of seawater-saturated rock beneath a freshwater lens. The other three soundings on Mauna Kea lavas - VES 3, 4, and 6, do not encounter a low resistivity basement; in fact, modeling VES 5 with a highly resistive basement about 160 m below sea level improves the fit by 30% over layered models with a conductive basement. This result is unusual for Hawaii unless we assume that the sounding was done at an elevation high enough so that the base of the freshwater lens was out of reach of our range of spacings. VES 4 and 6 might be explained by a freshwater lens whose surface was about 6 m above sea level depressing the base of the freshwater 240 m below sea level; however, the VES 3 location would require a water table at an unlikely 12 m above sea level. Of the six soundings, VES 3 is the only sounding which cannot be explained with normal, hydrologic models.

A bipole-dipole resistivity survey along the highway (Keller and others, 1977) gives results that support the interpretation of a high resistivity basement just west of the warm water well and a conductive basement near the coast. However, the data were primarily collected along roads with buried water pipes, which may have channelled current along the roads. There is a strong possibility that the data have been contaminated by this channelling.

Deep EM soundings would be able to answer many questions about the Kawaihae area and its geothermal potential.

Only limited conclusions can be drawn about the existence and location of a heat source from the data compiled so far. the best candidate is the body responsible for the magnetic anomaly crossing Hwy 19. TDEM soundings were tried during the 1979 survey but failed because of problems in obtaining large enough currents in the grounded wire source. The fine ash covering most of this area is very dry and electrical grounding is very difficult in it. Geothermal prospects are neither ruled out nor confirmed; further work should use EM soundings, preferably with an ungrounded source.

### 3.3 HUALALAI

Hualalai is a dormant volcano on the west coast of Hawaii. Most of its exposed lavas are alkalic olivine basalt which are thought to mantle a tholeiitic shield volcano beneath. There is no summit caldera on Hualalai nor is there evidence that one ever existed. Volcanic activity has occurred along two principal rift zones which trend northwest and southeast from the summit. A third minor rift zone extends north from the summit. Puu Waawaa, a large trachyte pumice cone from which an extensive trachyte lava flow issued, is located on the minor rift zone. The only recorded activity on Hualalai was two eruptions in 1800 and 1801 on the northwest rift and an intensive swarm of earthquakes in 1929 under the volcano's north flank (summarized from Macdonald and Abbott, 1970; see Fig. 5).

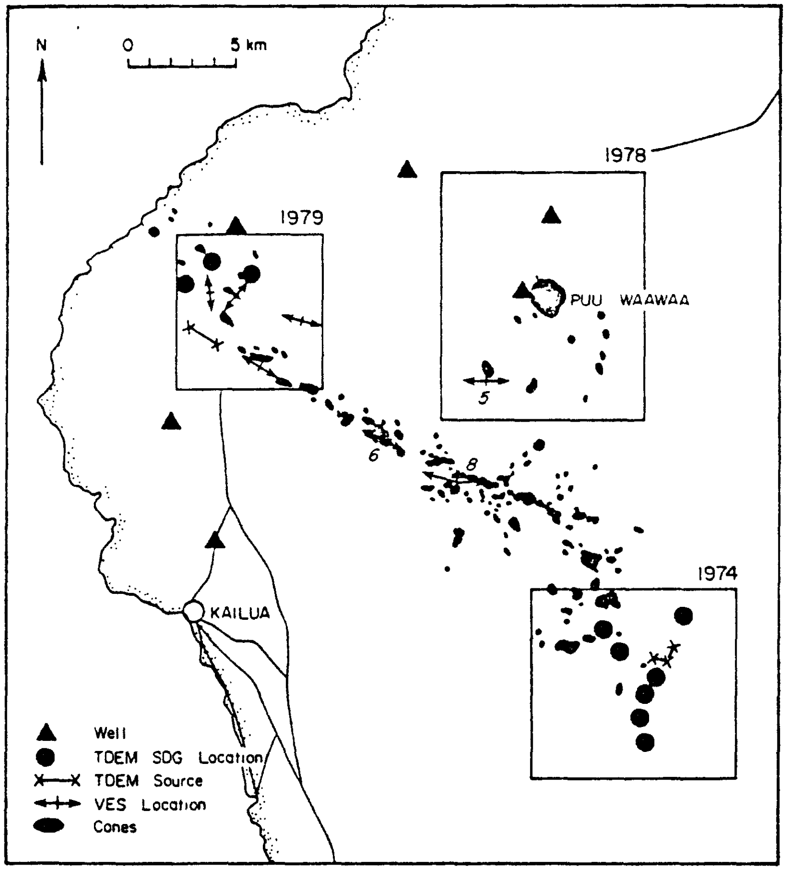



Figure 5: Schematic geologic map of Hualalai volcano (after Macdonald and Abbott, 1970). Superposed on the geology are outlines of the particular areas studied in 1974, 1978, and 1979.

The earliest geophysical work on Hualalai was a ground-water assessment survey (Adams and others, 1969), which covered the north and west slopes below 600 m elevation with low-level aeromagnetics, resistivity profiling and audiomagnetotelluric (AMT) profiling. Based on preliminary interpretations, the investigators concluded that Hualalai's north and northwest rift zones are marked by high horizontal magnetic gradients and lower than normal apparent resistivities; groundwater was expected to be channelled from higher elevations by the north rift structure. The significance of the rift's geophysical characteristics was not discussed.

The first exploration survey for geothermal resources on Hualalai concentrated on the southeast rift zone (see Fig. 5). Two separate grounded-wire sources were set up and electric field and time-domain electromagnetic field data were collected at sites up to 2400 meters from the source (Klein and Kauahikaua, 1975). A reexamination of that data showed uniformly high resistivities in excess of 50 ohm-m from the TDEM and 98 to 683 ohm-m from the electric field data. Taken together, the data preclude any anomalously conductive zones under the area studied to a depth of roughly 2 km; the results are not encouraging for a major geothermal resource but, as a sidelight, may suggest high-level groundwater less than 900 m from the surface. The electric field apparent resistivity values seem to cluster between 500 and 700 ohm-m, a value that is associated with freshwater-saturated basalts on Oahu (Zohdy and Jackson, 1969).

Most recently, a commercial geothermal exploration survey was conducted on the north slopes of Hualalai for the Puu Waawaa ranch (Charles Helsley, written communication, 1978). Reconnaissance magnetotellurics, seismicity, resistivity sounding, soil mercury, and electrical self-potential surveys were conducted; all results seemed to point to the areas immediately around Puu Waawaa cone as anomalous. The prospect was drilled at two locations in early 1979 to depths of 1.7 and 2 km, respectively; however, abundant but exceptionally cold groundwater was encountered and the effort was abandoned (A. Shadduck, oral communication, 1979).

The 1979 phase of exploration consisted of seven Schlumberger soundings and three TDEM soundings on the northwest rift and summit areas of Hualalai (Fig. 5). The most intensive effort was on the lower northwest rift in the vicinity of the 1801 eruptive vent; the trace of the rift becomes more diffuse in this area as is shown in Fig. 6 . Data and interpretations from the two lower Schlumberger soundings, VES 1 and 7, are shown in Fig. 7 . They do not differ greatly in character nor in interpretation; both show surface resistivities of about 8000 ohm-m, a second layer of a few thousand ohm-m, and a basement of very low but unresolvable resistivity. High surface resistivities are common in recent volcanic terrain on Hawaii and represent fresh, unaltered lava flows containing very little water. The intermediate resistivity layer may represent a slight increase in the moisture content of these lava flows. The very low resistivity of the basement or lowermost layer, suggests completely saturated rocks with the pore fluid being brackish water or seawater. The elevation of the top of this basement with respect to sea level should determine whether there may be freshwater floating on salt water at this location. VES 1 suggests no freshwater as depth to

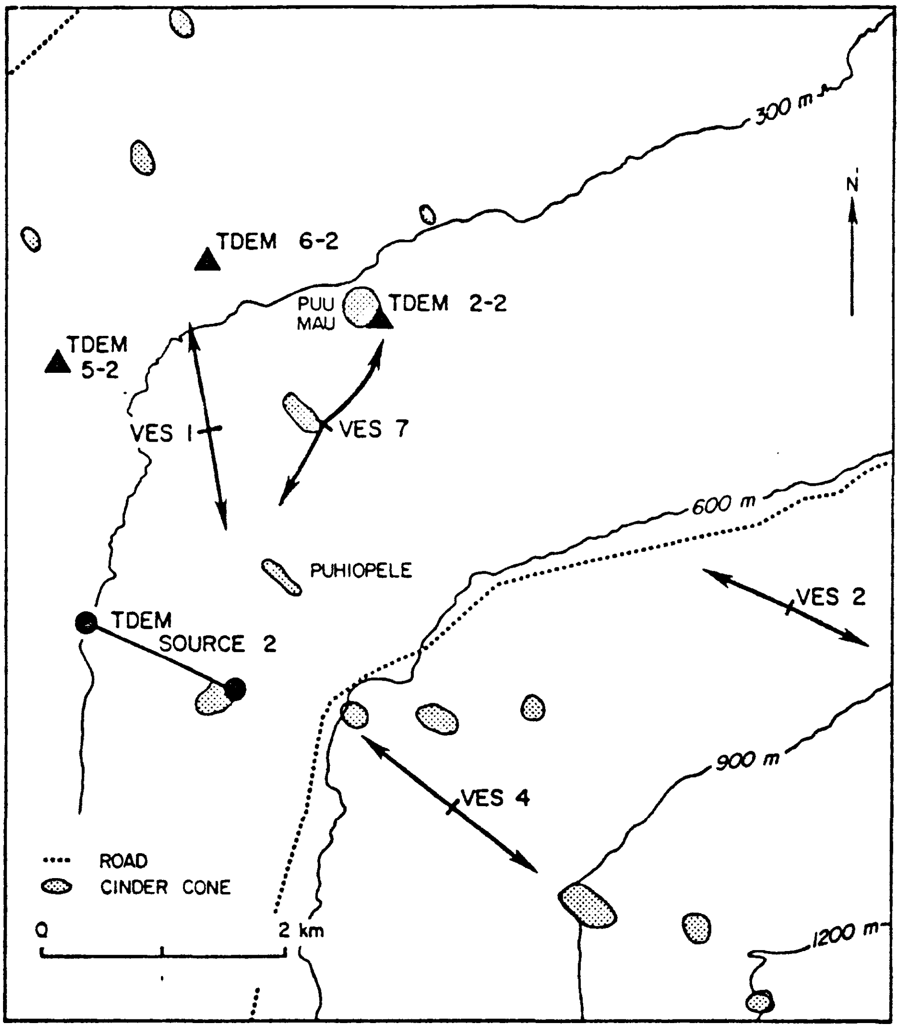



Figure 6: Detailed locations for TDEM and VES soundings within the area labeled 1979 in Fig. 4. Puhiopele is the vent for the 1801 Hualalai eruption.

basement equals the elevation, and VES 7 suggests a hydraulic head of no more than 1.6 m. In good agreement, water wells in the immediate vicinity yield slightly brackish water with a hydraulic head of .3 to 1.3 m (Davis and Yamanaga, 1973).

VES 2 and 4 were located at higher elevations and did not detect the low-resistivity basement; however, the section of undersaturated rock is detailed (see Fig. 8). VES 4 shows a well-developed thickness of low resistivity soil at

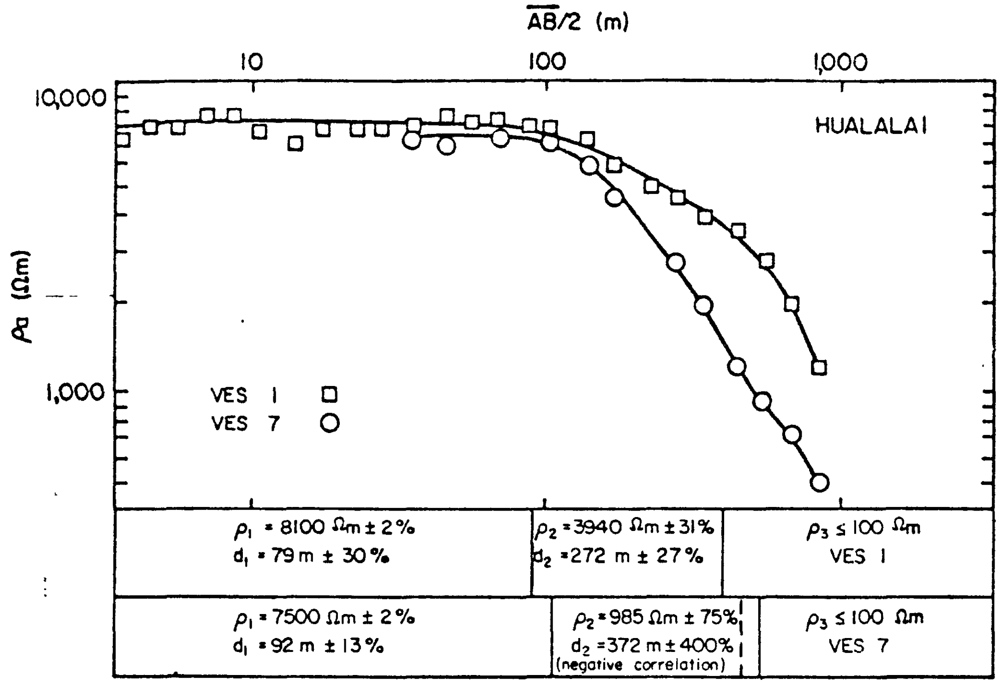



Figure 7: Data and interpretations for Hualalai VES 1 and 7.

the surface which overlies highly resistive lava flows. The soil layer is absent in VES 2 where resistive lava flows outcrop at the surface.

It is the low-resistivity basement, that is, the water-saturated rocks, which are of direct geothermal interest; therefore, three TDEM soundings were obtained near the 1801 vent to better determine its deep structure. The data and the derived apparent resistivity versus apparent depth curves are presented in Fig. 9 . Soundings 5-2 and 6-2 show almost identical structures of about 15 ohm-m decreasing to 9 to 12 ohm-m at a depth of 1500 to 1800 m. Sounding 2-2 shows quite a different picture: apparent resistivities as low as 9 ohm-m over a more resistive basement at 800 to 900 m depth. The lower resistivity at the shallower depths in

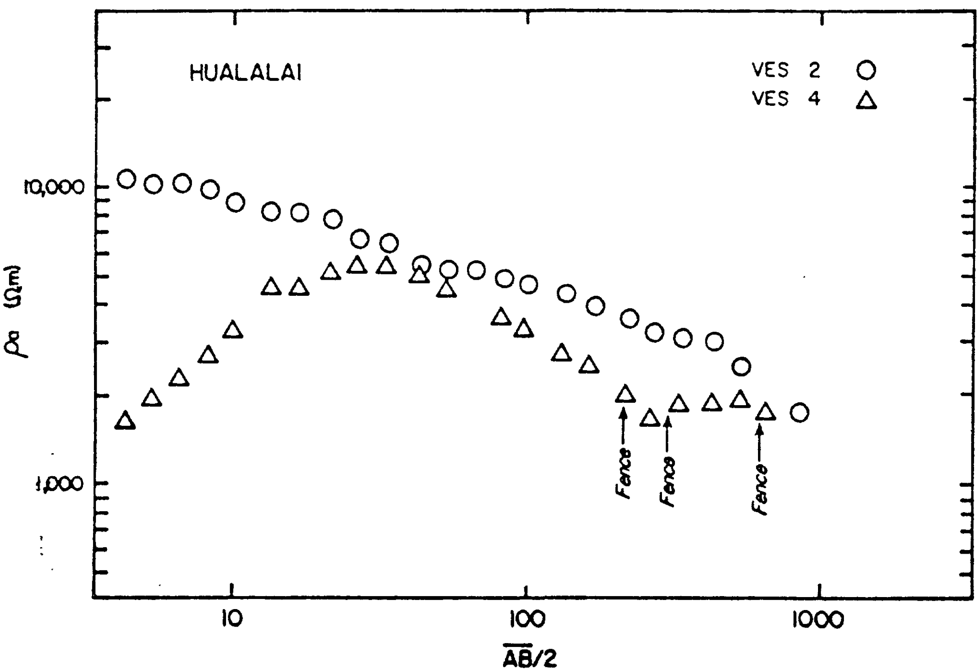



Figure 8: Data for VES 2 and 4. Approximate position of fences crossed during sounding are noted.

the water table could suggest local heat while the resistive basement may be the heat source, analogous to the Kawaihae model discussed previously. The resistivities interpreted for VES 7, which is very close to TDEM 2-2 are also significantly lower than for VES 1; in particular, the intermediate layer has a resistivity of about 1000 ohm-m in VES 7 and about 4000 ohm-m in VES 1. In other words, the Schlumberger results are consistent with the TDEM results for the Puu Mau area, even though the two types of soundings do not detail the same depth range. Resistivities are locally lower at least to depths of 800 to 900 m and may signify the effects of heat.

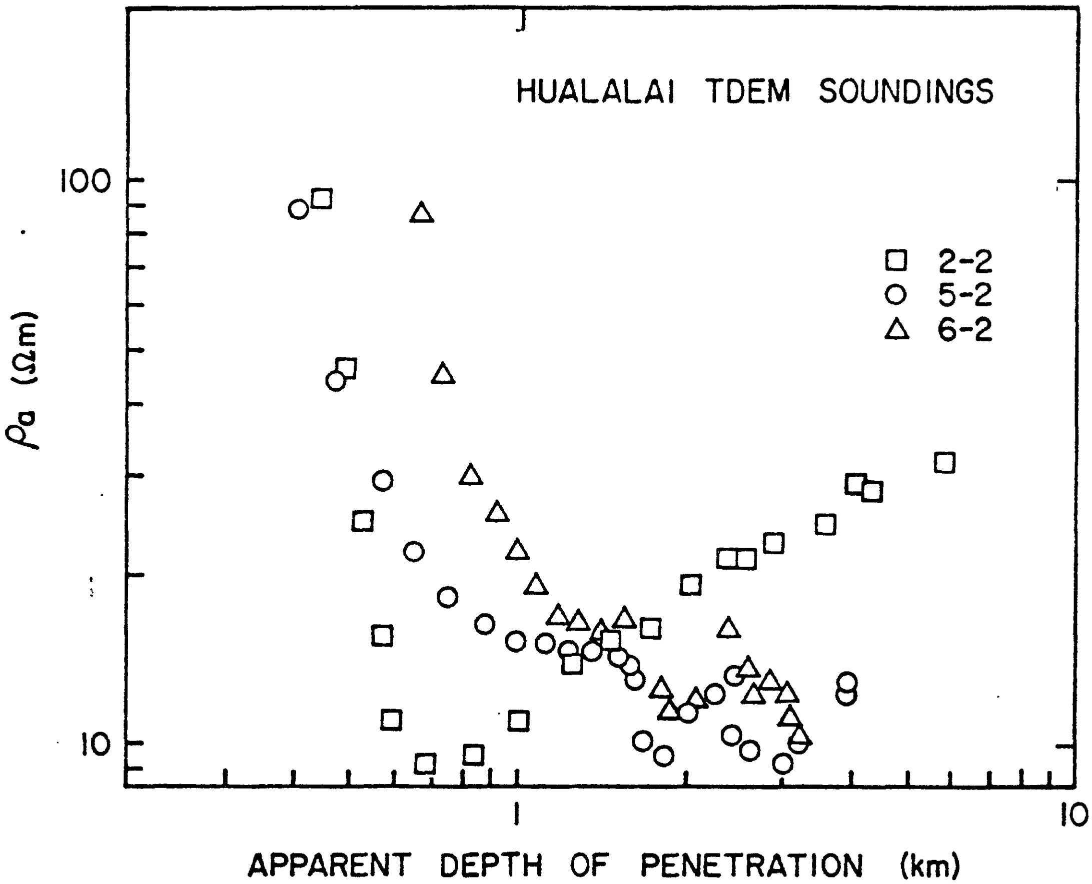



Figure 9: Converted TDEM data from the lower northwest rift zone of Hualalai.

Three more Schlumberger soundings were located at higher elevations along the northwest and north rifts and the summit. Their data and interpretations are shown in Fig. 10 . The shallowest 300 m of each of them consisted of one or two 3000 to 6000 ohm-m layers and a very high resistivity layer. These two layers totally describe the VES 6 sounding located slightly southwest from the 1800 Kaupulehu vent. By interpretation, no water-saturated rock was found to 900 m depth. An 800 ohm-m basement was detected only 300 m beneath the surface by VES 5 on the upper north rift which may represent high-level freshwater.

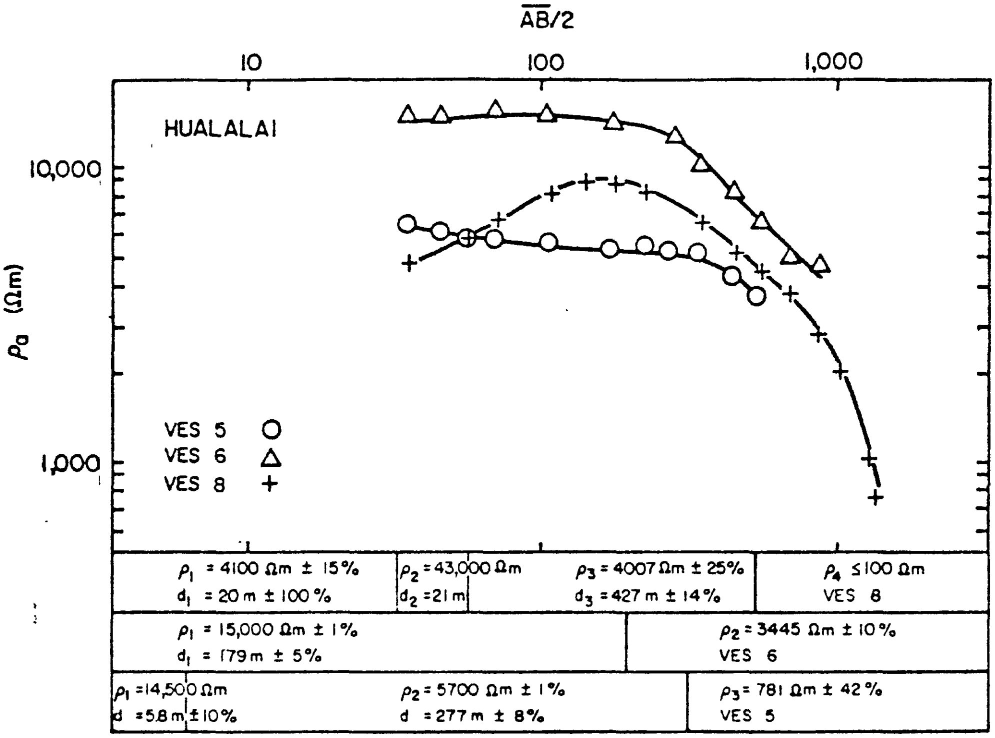



Figure 10: Data and interpretations for VES 5, 6, and 8.

Of more direct geothermal significance is the low-resistivity basement delineated by VES 8 at Hualalai's summit. It's resistivity is too low to be uniquely determined by this sounding but it must be less than 100 ohm-m; its surface is 480 m below the summit. A similar resistivity section has been interpreted beneath the summit region of Kilauea volcano (Dallas Jackson, oral communication, 1979) where the low resistivity basement coincides with water-saturated rocks discovered by drilling (Zablocki and others, 1974). The temperature of the shallow water at Kilauea is not particularly warm, but it does increase to about 130° C at the hole bottom. On the basis of the similarity between VES 8 and the more completely studied structure at Kilauea, we believe that water-saturated rocks occur 480 m beneath the summit of Hualalai; however, the water may not be anomalously warm.

In summary, there are no strong anomalies on Hualalai, but there are two areas that warrant future exploration: the lower northwest rift around Puu Mau, and the summit. The resistivity data in the Puu Mau area directly suggest a heat anomaly; however, the source may be small. Interpreted resistivity structures beneath the summit are ambiguous as far as their geothermal significance is concerned. Both prospects are marginal on the basis of the data compiled here, but may be worth exploiting because of their close proximity to major demand centers (Thomas and others, 1979).

## 3.4 THE SOUTHWEST RIFT ZONE OF MAUNA LOA (SOUTH POINT)

The southwest rift zone is the only one of Mauna Loa's two major rift zones that extends to the coast. It has had seven historic eruptions, all but one occurring above 2300 m altitude (Macdonald, 1977). The eruptions below 3000 m altitude have generally occurred at progressively higher altitudes with time, and consequently farther from the coast. Along the lowest 17 km of the rift, a west-facing, normal hinge fault marks the rift trace. This fault, known as the Kahuku fault, extends at least 40 km offshore (Fornari and others, 1979).

The primary indicators for geothermal potential in this area were the numerous, recent eruptions and anomalies delineated by aerial infrared photography. Fischer and others (1964) noted thermal sources along the rift and warm springs flowing into the ocean where the rift intersects the coast. Abbott (1975) photographed thermal anomalies along the cliff face of the Kahuku fault (possibly residual solar heat), as well as a large patch of warm water directly offshore.

Most of the initial electrical exploration was concentrated on the area east of Kahuku fault, principally to evaluate possible sources of the cliff-face anomalies. The 1974 survey crew completed 10 TDEM soundings and two equatorial DC soundings. A preliminary interpretation of the TDEM soundings found that subsurface resistivities might be as low as 10 ohm-m (Klein and Kauahikaua, 1975); however, the DC results showed a much more resistive structure. Even with the large amount of scatter in Fig. 11, the representative vertical section is obviously 800-1000 ohm-m to about 1 km depth, underlain by 10-100 ohm-m material. Although it would be difficult to prove, the TDEM soundings were probably distorted by the many kilometers of galvanized pipe laid over this area to water cattle. The pipe network may have an effect similar to that of a highly conductive surface layer, and might explain the discrepancy between TDEM and DC results.

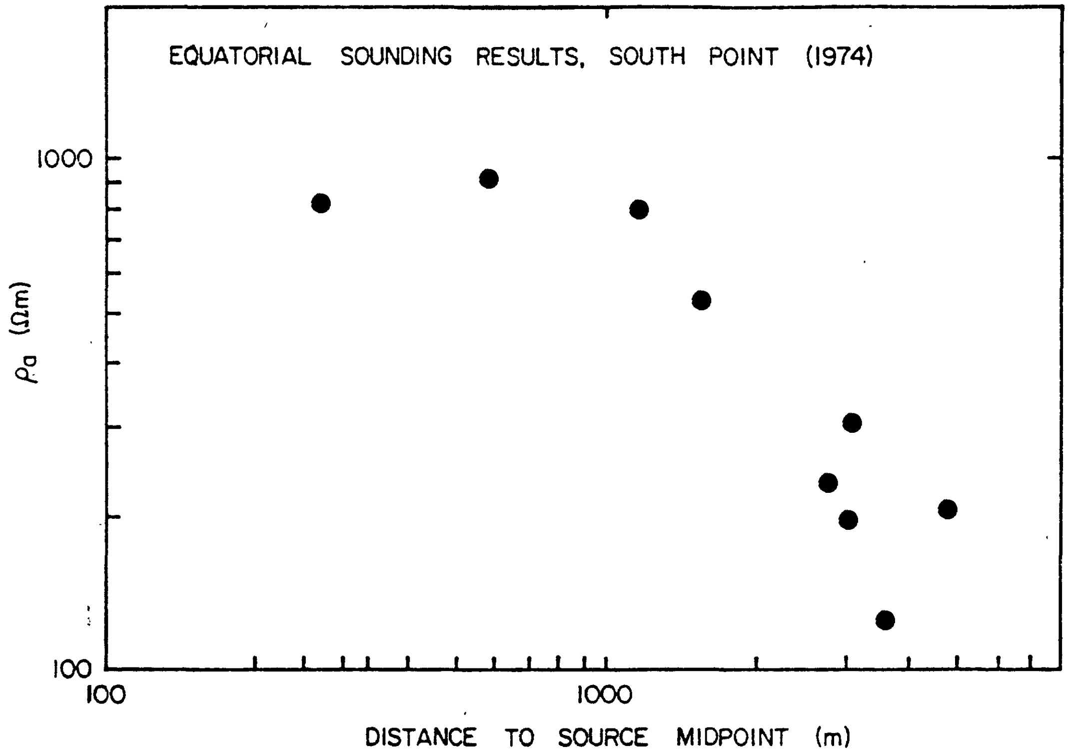



Figure 11: Equatorial sounding results from the 1974 survey of South Point.

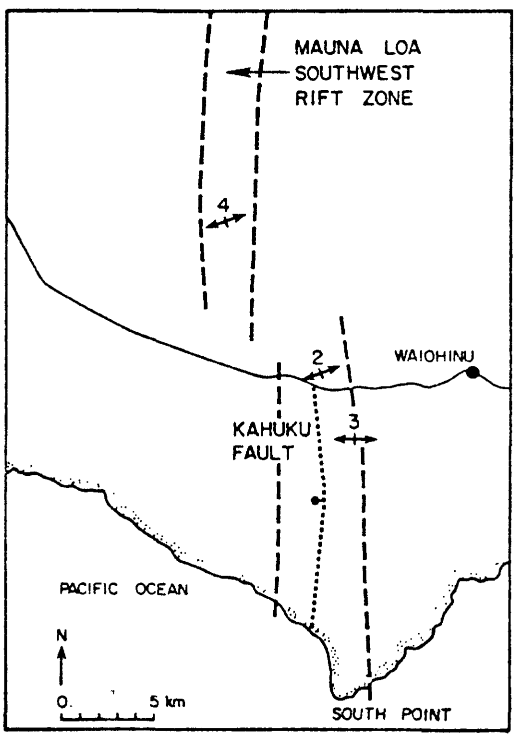



Figure 12: Map of the lower southwest rift zone of Mauna Loa showing VES locations.

The TDEM method was not used in the small amount of follow up work during the 1979 survey. Several VES were attempted in locations east and north of the Kahuku fault (Fig. 12), but only three were successful; pipes were found to be so close together that an undistorted VES was impossible in many places. The three Schlumberger soundings and their interpretations are shown in Figs. 13 and 14 . Anomalously low resistivities were not detected in any of the soundings.

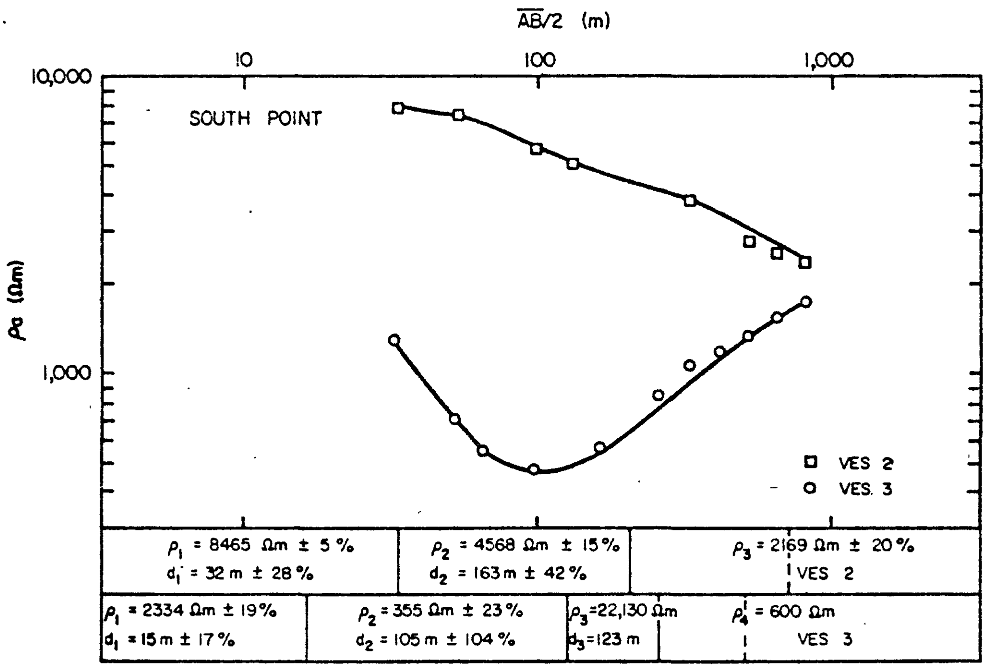



Figure 13: Data and interpretations for VES 2 and 3.

VES 4 was noticeably distorted between the 100- and 200-meter spacings. These spacings were precisely where the

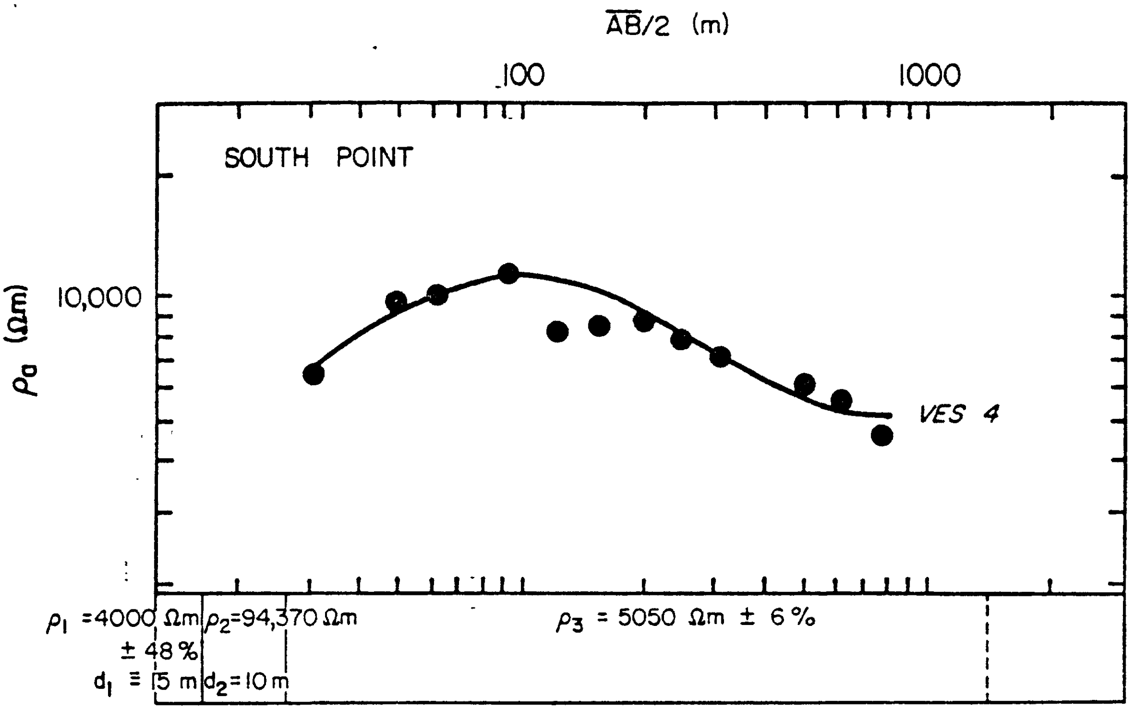



Figure 14: Data and interpretations for VES 4.

electrodes crossed a prehistoric eruptive fissure. In order to determine more about this fissure, a self-potential (SP) traverse was run along the VES line; the results are plotted in figure 15.

A narrow -140 mV anomaly was found centered over the fissure. Because SP anomalies caused by subsurface heat are generally positive (Zablocki, 1976), this fissure is possibly not of geothermal interest.

The admittedly sparse reconnaissance has uncovered no evidence to suggest that warm water exists beneath the South Point area, despite the infrared findings. Reconnaissance-level work still needs to be done west of Kahuku fault for complete coverage.

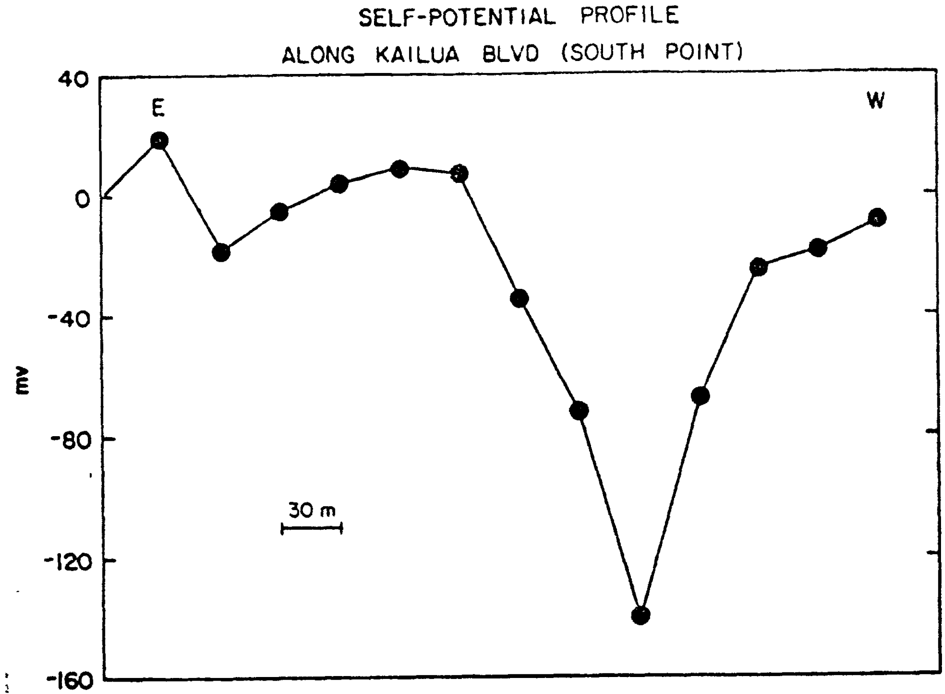



Figure 15: SP data measured along the line of VES 4. Negative peak is centered over a prehistoric fissure.

## 3.5 THE SOUTHWEST RIFT ZONE OF KILAUEA

The southwest rift is the least active of Kilauea's two rift zones having only erupted five times in history. This includes a fissure eruption in 1823 involving about 21 km of the 'Great Crack', the building of the Mauna Iki shield in 1919 (Macdonald and Abbott, 1970, p. 71,77), and the two most recent eruptions which occurred on subparallel sets of en echelon fissures near the summit in 1971 and 1974. Eruptions have progressed towards the summit in this period, and since 1975, activity in the form of seismicity, has been confined principally to the upper few kilometers of the rift zone. This portion of the rift zone is within the boundaries of Hawaii Volcanoes National Park and is off-limits for exploitation; the following discussion focuses on the lower rift zone.

Groundwater exploration was the first application of geophysical methods in this area. Hussong and Cox (1967) delineated the southwest and southeast boundaries of high-level water beneath the town of Pahala by means of several Schlumberger (VES) soundings. The authors hypothesize that the principal impounding structure is a series of northeast-striking dikes below a prehistoric eruptive fissure just southwest of Pahala (see Fig. 16). The water table elevation drops from 61 m beneath Pahala to about 20 m seaward of the fissure. Subsequent studies by Adams and others (1970) substantiated these conclusions by demonstrating that the boundaries they had inferred from DC soundings were detectable using shallow resistivity, gravity profiling and low-level aeromagnetics.

A recent examination of data taken during the 1974 survey (Klein and Kauahikaua, 1975) demonstrates that the fissure boundary is also a contact between high resistivity rocks to the west and low resistivity rocks to the east (Kilauea side). Fig. 16 shows the locations of three DC bipole sources and several dipole measurement locations, whereas Fig. 17 shows three plots of apparent resistivity versus distance along two lines normal to the rift trend. The abrupt change in apparent resistivities along these lines marks the contact quite precisely: near Pahala (line B-B') the contact coincides with the prehistoric fissure thought by Hussong and Cox (1967) to be a hydrologic barrier, and farther north (line A-A'), the contact is about 1 km west of the Great Crack. The same contact can be traced farther up the rift zone to the summit caldera using the dipole-bipole mapping data of Keller and others (1977, p. 42), and probably represents the easternmost extent of dikes or cracks associated with the southwest rift. The resistivity data shows that the contrast is deep enough (below sea level) to require water saturation; the Mauna Loa side has a resistivity of a few hundred ohm-m and probably represents cold, freshwater saturated rocks (Hussong and Cox, 1967), while the Kilauea side has a resistivity of 2-3 ohm-m, which may represent warm, water-saturated rocks.

Several TDEM soundings were obtained during the 1974 survey; however, consistent interpretations from sounding to sounding along each of the two lines were impossible. Nothing could be gained from the data set as a whole, therefore the soundings are not presented here.

The available data are encouraging and suggest that a geothermal resource is located in the southwest rift area. Surface data suggest that the subsurface resistivities are low, and therefore the temperatures may be high. The actual values obtained are similar to those determined in the East Rift Geothermal area. Further work is warranted here.

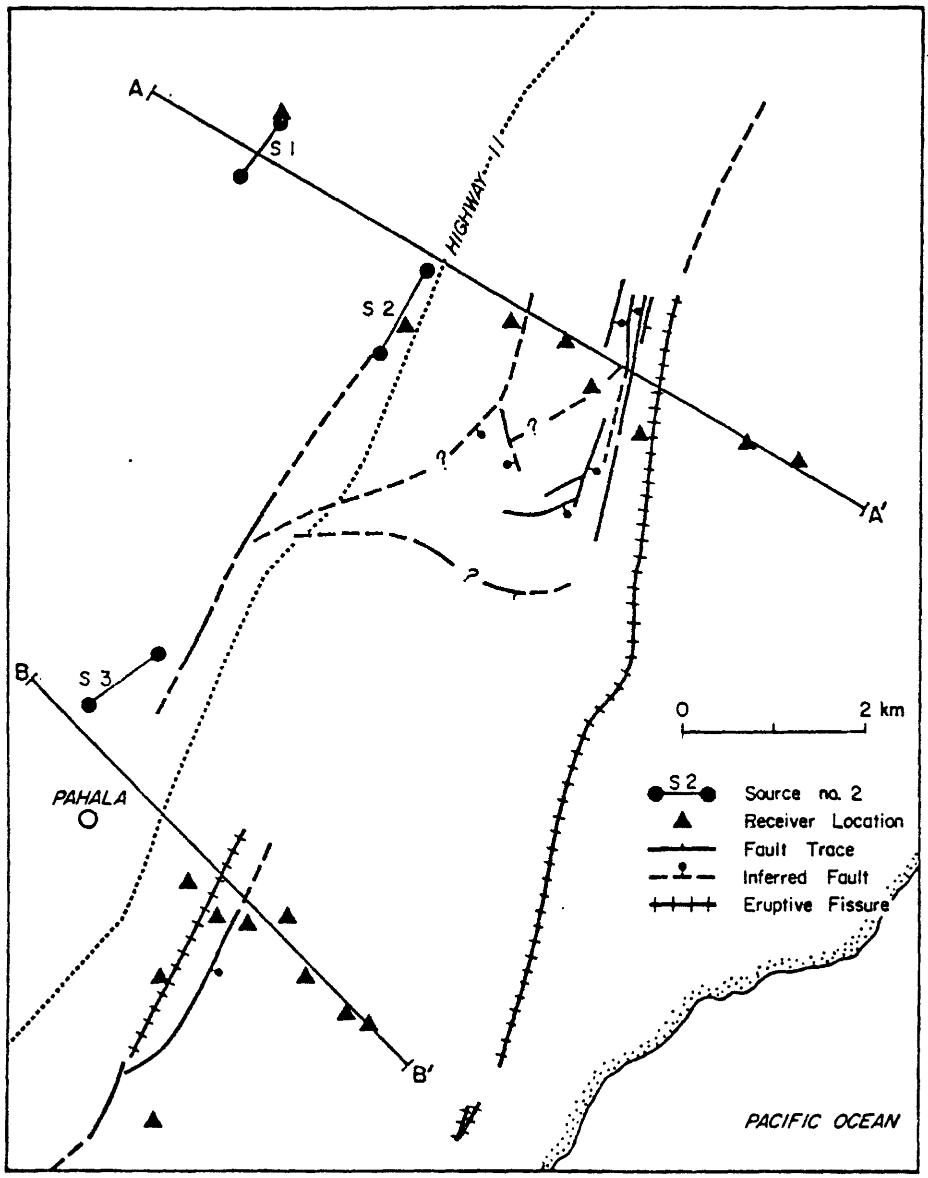



Figure 16: Map of Kilauea's southwest rift zone showing prominent volcanic features (from Holcomb, 1980) and the location of the two bipole lines discussed in the text.

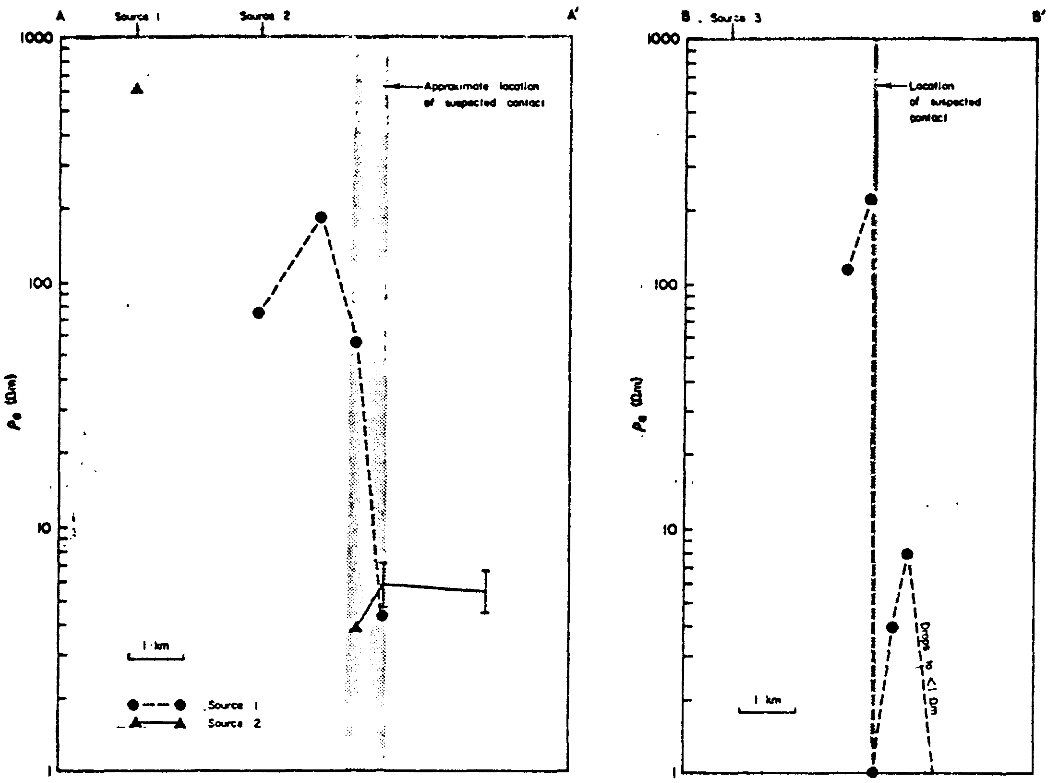



Figure 17: Apparent resistivities measured along the two profiles shown in Fig. 16. Also shown are the locations of an interpreted subsurface contact.

### 3.6 KEAAU

At first inspection, the Keaau area seems an unlikely geothermal prospect; however, two features may make it an area possibly worth investigation. First, the area is on the projected trace of Mauna Loa's northeast rift zone. If it exists, the rift would now be buried by younger Kilauea lavas. This portion of the rift may still be hot at depth even though it has not been active for at least 1000 years (the oldest overlying flow; Holcomb, 1980). Second, shallow wells tap groundwater which is somewhat anomalous geochemically (Thomas and others, 1979; p. 35).

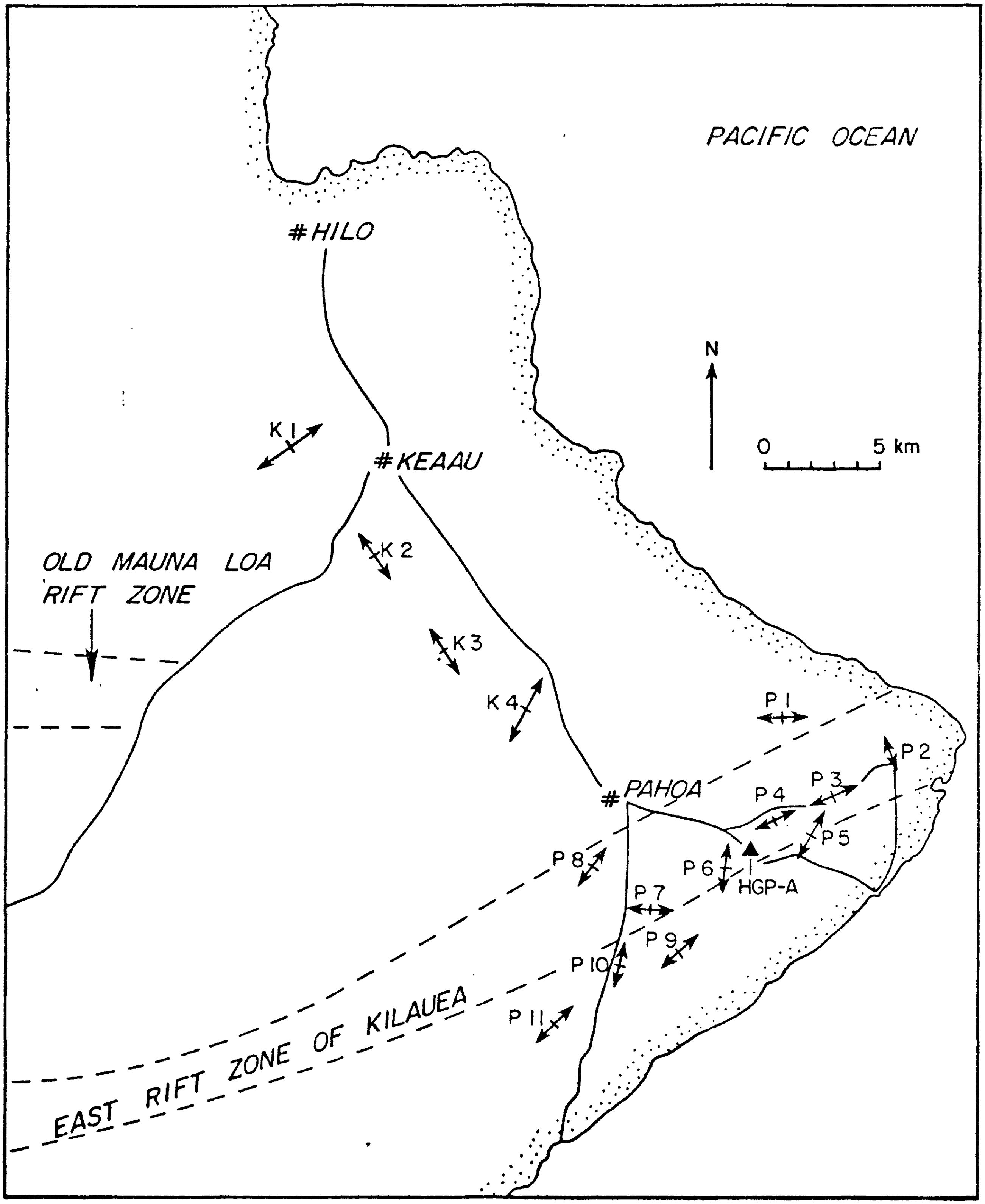



Figure 18: Map of the eastern cape of Hawaii showing the locations of Keaau and Puna VES soundings.

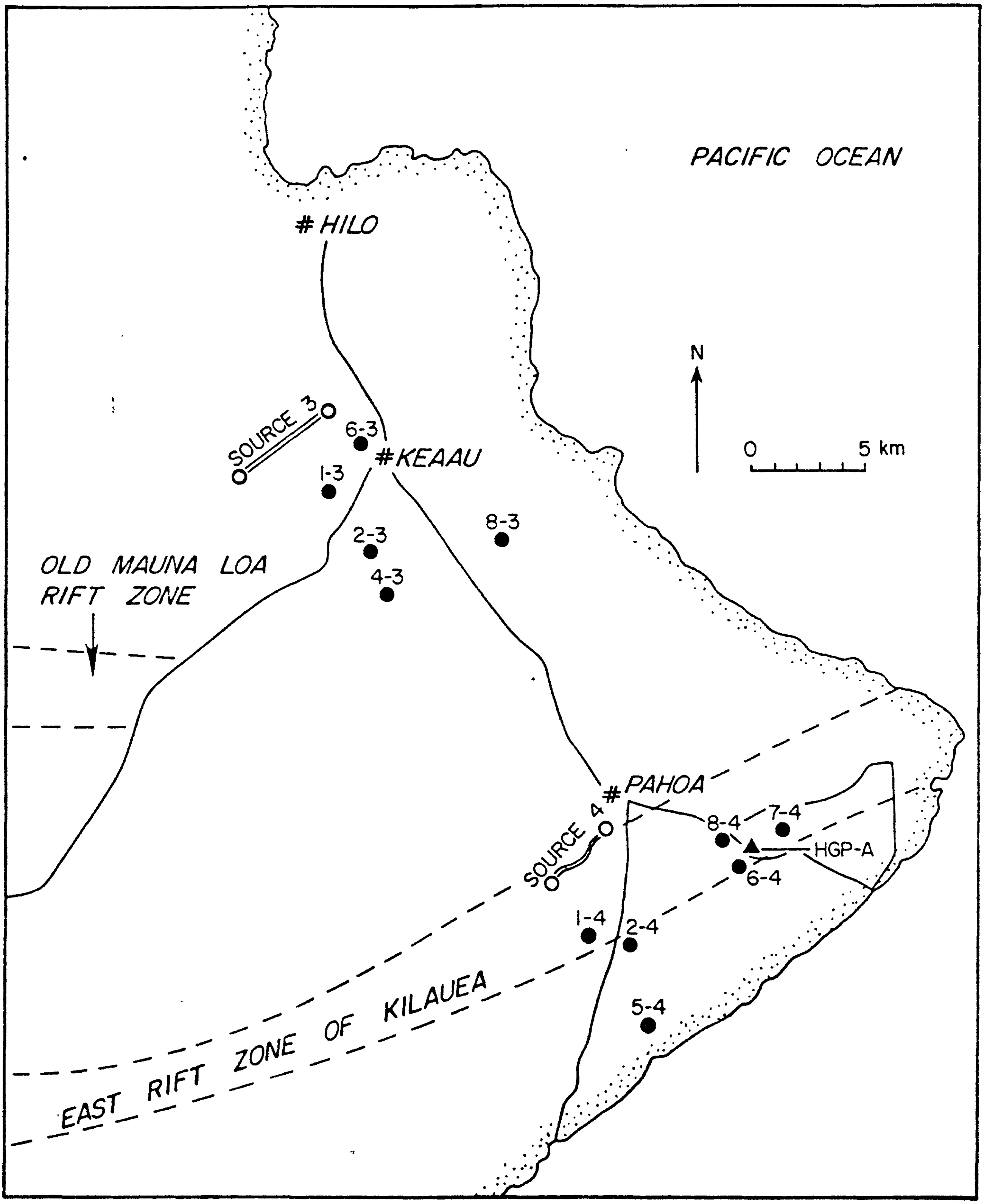



Figure 19: Map of the eastern cape of Hawaii showing the locations of Keaau and Puna TDEM soundings.

Figures 18 and 19 show the locations of VES and TDEM work completed in the Keaau area during the 1979 survey. The VES data are plotted in Figs. 20 and 21 . The data are all consistent with a model containing highly-resistive, undersaturated lavas (3000-20,000 ohm-m) overlying cold, freshwater-saturated lavas (500-900 ohm-m), which in turn overlie seawater-saturated lavas (less than 100 ohm-m). The depths to these various interfaces are not well resolved, but are not important to this assessment. For evaluating the possibilities of deep residual heat, it is the resistivities of seawater-saturated rocks that are the most important. Three of the four sets of transformed TDEM data in Fig. 22 show that these resistivities are greater than 30 ohm-m between the approximate depths of 1 to 3 km. Sounding 2-3 suggests that this value may be as low as 13 ohm-m; however, this value may be suspect because this sounding is flanked by two others that both show more resistive strata at depth.

The prospects are not good for a large geothermal resource beneath this area. A value of 30 ohm-m is typical for cold, seawater-saturated lavas on Oahu (Zohdy and Jackson, 1969) and is the highest resistivity yet determined for these depths on Hawaii. The lower resistivity value of sounding 2-3 may be explained as an area of higher porosity.

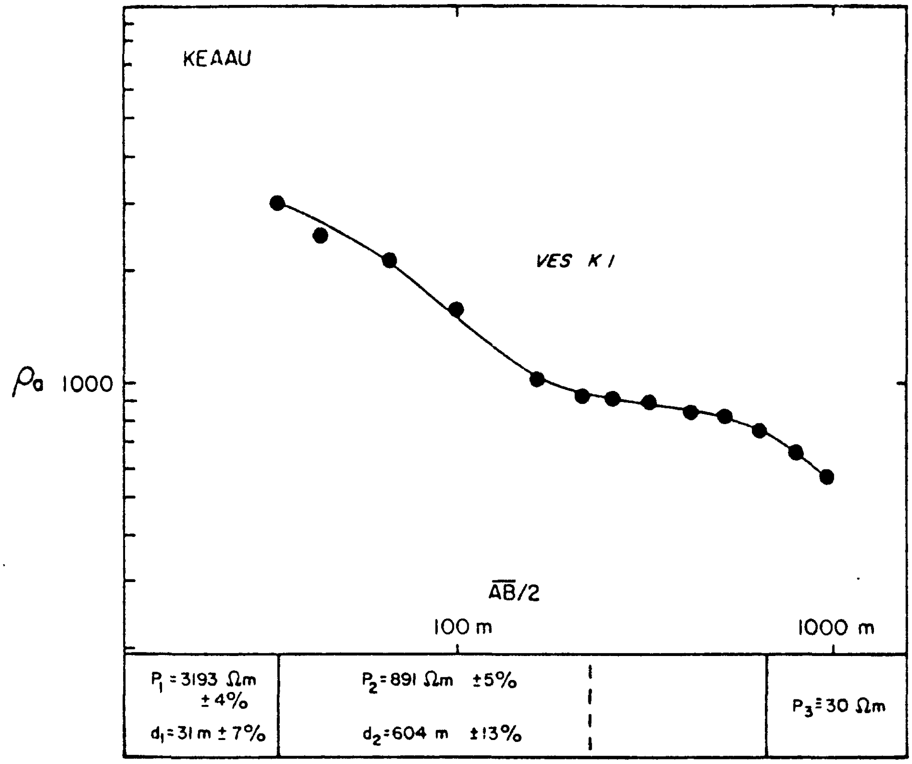



Figure 20: Data and interpretation for Keaau VES K1.

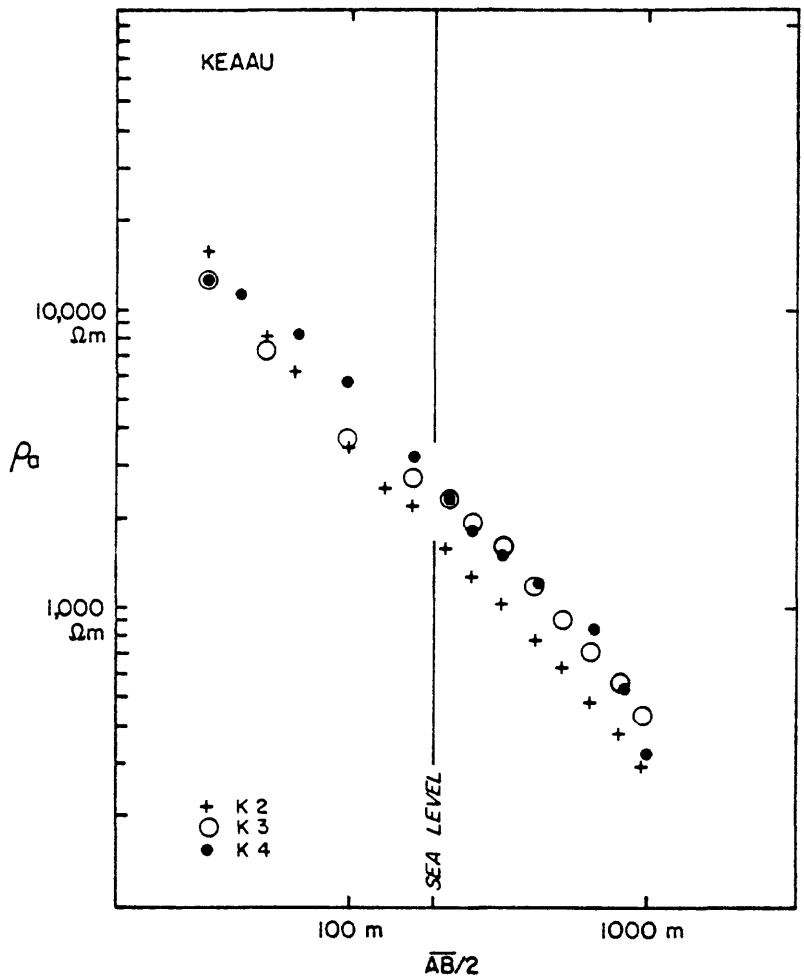



Figure 21: Composite plot of VES K2, K3, and K4.

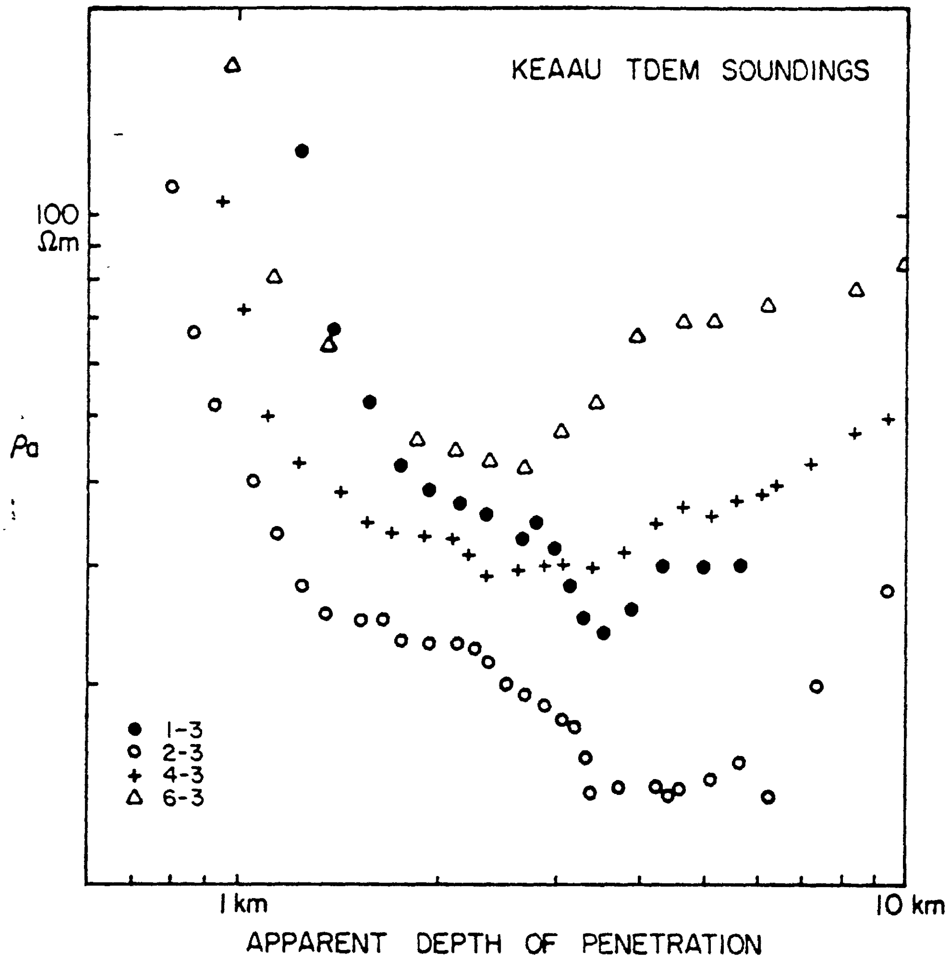



Figure 22: Keaau TDEM data converted to apparent resistivities versus apparent depth of penetration.

### 3.7 THE EAST RIFT GEOTHERMAL AREA OF KILAUEA (PUNA)

With the exception of its summit, the east rift zone of Kilauea, now a Known Geothermal Resource Area (KGRA), has been studied to a much greater extent than any other volcanic feature in the state of Hawaii. The earliest geophysical studies clearly showed that the lowest fluid-saturated-rock resistivities on the island were found in Puna (Klein and Kauahikaua, 1975; Keller and others, 1977). More detailed studies located several anomalies within the rift and beneath the southeastern rift flank (Keller and others, 1977; Kauahikaua and Klein, 1977a and 1977b; Zablocki, 1977; Kauahikaua, 1981b). HGP-A, the successful geothermal test hole, was located near one of the self-potential anomalies within the rift structure.

Follow up work during the 1979 survey consisted of six deep TDEM soundings southwest of HGP-A and six VES along the rift. The purpose was to map the lateral and vertical extent of the resource. Locations of all soundings are shown in Figures 18 and 19 .

Detailed interpretations of VES P10, P7, P9, P5, and P4 have already been published (Kauahikaua and Klein, 1977a; listed as G1, G3, G4, G5, and G6, respectively). With the exception of P7, each of the interpretations had a highly variable surface layer over a relatively uniform, 6000-7000 ohm-m layer that extended down to sea level. For AB/2 spacings greater than the soundings' elevation, the apparent resistivities always dropped dramatically, signifying a resistivity too low to resolve with this data. Sounding P7 was different because its interpretation required material of moderate resistivity (2500-600 ohm-m) below the 6000-7000 ohm-m layer; additionally, the apparent resistivities for this sounding do not decrease as dramatically for AB/2 spacings greater than the sounding elevation.

Analysis of the additional soundings obtained during the 1979 survey seemed to confirm that VES obtained in the vicinity of the east rift would be of two basic types. The first kind is exemplified by VES P3: a thin, variable surface layer of resistivity greater than 4000 ohm-m, over a thick, rather uniform layer of 6000-7000 ohm-m that extends to sea level, over a halfspace where the resistivity was too low to resolve.

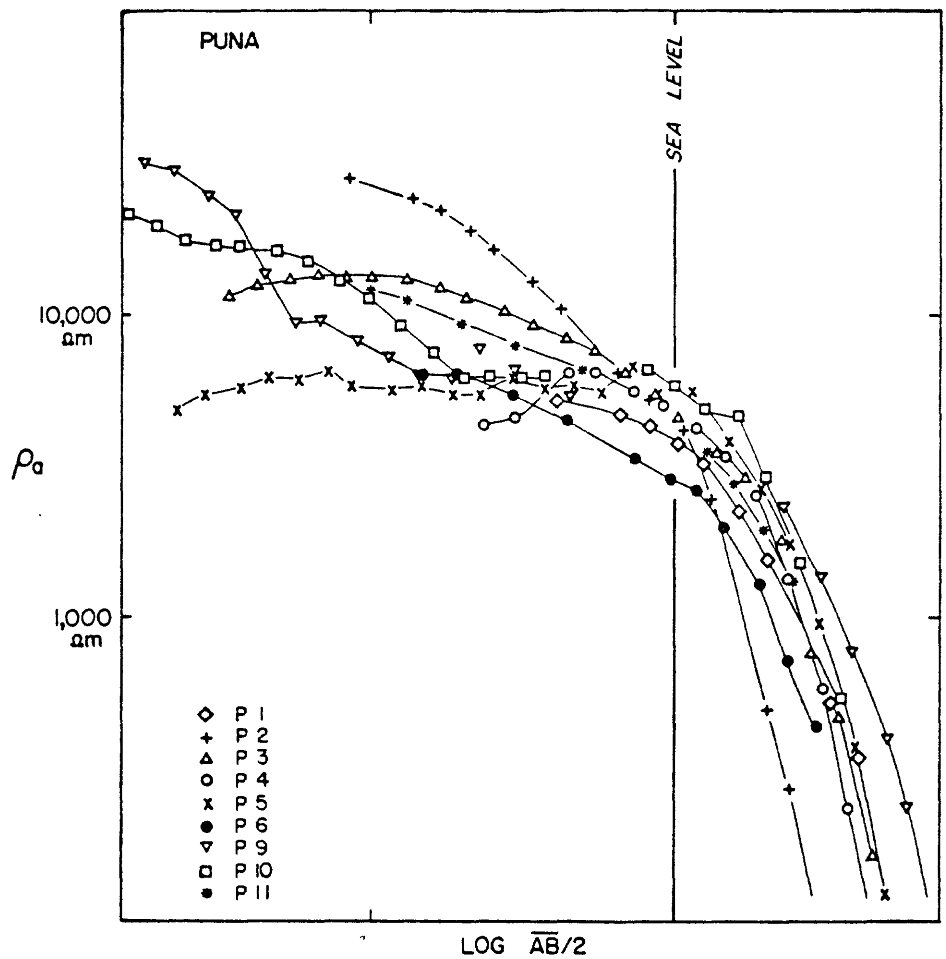



Figure 23: VES data in area I laterally shifted along the AB/2 axis so that the spacings corresponding to sea level were coincident.

Fig. 23 is a composite plot of data from seven soundings in this category aligned horizontally so that the AB/2 corresponding to each of the scundings' elevation are coincident. The data from two soundings that are considered transitional to the second class of soundings are also plotted (P6 and P1). With the exception of P2, which was measured perpendicular to a graben structure and may therefore be distorted, the deeper portions of each of the soundings in this first class are remarkably similar to each other when plotted in this way.

The second class of soundings is exemplified by P7: a thin, surface layer of less than 4000 ohm-m resistivity over the 6000-7000 ohm-m layer which, in turn, overlays material in the range of 2500-600 ohm-m resistivity. Fig. 24 is a composite plot of the two soundings definitely in this category (P7 and P8) as well as the two transitional soundings (P6 and P1) already plotted in Fig. 23 .

The primary distinction between the two groups of sounding interpretations is whether moderate resistivity material can be found between the 6000-7000 ohm-m layer and sea level. Using this classification, Keaau VES K2, K3, and K4 also fit into the second category, and the sounding categories begin to fit a spatial pattern. Category I soundings are located either within the rift east of HGP-A, or on the rift's southeastern flank. Category II soundings are located within the rift west of HGP-A or on the northwest flank of the rift.

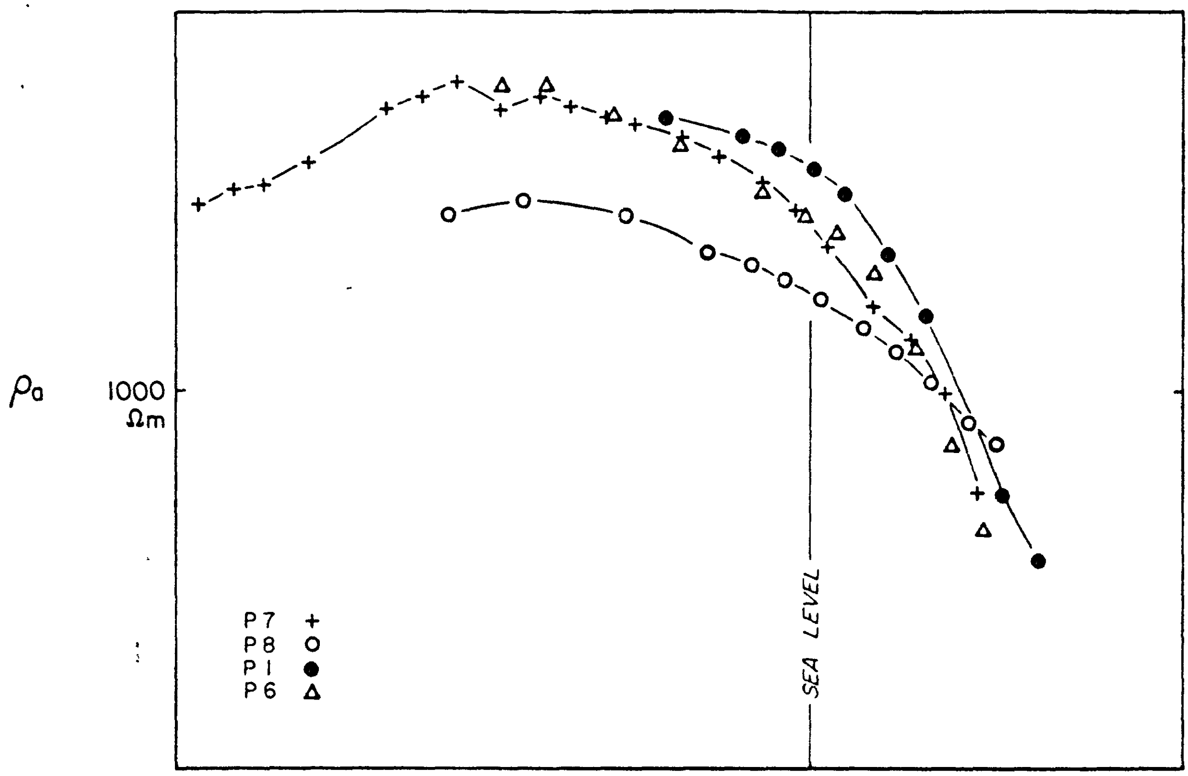



Figure 24: VES data in area II laterally shifted along the AB/2 axis so that the spacings corresponding to sea level were coincident.

The pattern of soundings bears a striking resemblance to the pattern of basal-water occurrences in this area. Wells drilled within area I (the category I area) generally tap brackish-to-saline water with a hydraulic head of less than 1 m and an abnormally high temperature. Wells drilled within area II (the category II area) tap a well-developed basal lens of fresh, cold groundwater with a hydraulic head of about 3 to 7 m. The distinction between area I and area II soundings appears to be a distinction between the presence or absence of a significant thickness of freshwater. The distinction has theoretical grounds in that rock saturated with fresh, cold water is up to 20 times more resistive than rock saturated with warm, brackish-to-saline water. The detection of mcderate resistivity material above sea level may be detection of freshwater directly.

Because there are no wells drilled within the rift structure west of HGP-A, we must use the VES results to make inferences about the hydrology there. The interpretation of sounding P7 (Kauahikaua and Klein, 1977a) and P8 are consistent with the presence of cold, freshwater at an elevation of at least 9 to 17 m above sea level; if so, this is too high for basal water and must be dike-impounded water. Moving east towards HGP-A, VES P6 results suggest that the water table is still elevated, but not as high as farther west. Moving farther east past HGP-A, the transition to the situation of virtually no basal lens is complete at the location of VES P4 and P5. The decline in water-table elevation is mirrored in the seismic refraction results west of HGP-A (Suyenaga, 1978) and is coincident with an L-shaped self-potential anomaly near HGP-A (Zablocki, 1977; Zablocki and Koyanagi, 1979).

The early controlled-source EM data also show systematic differences between soundings in the two areas. Small-separation audio-frequency soundings detected resistivities of less than 7 ohm-m just below sea level within area I. These low resistivities were interpreted to be warm, seawater-saturated rocks (Kauahikaua and Klein, 1977a). Deeper resistivity structures were delineated by TDEM surveys. Beneath most of the rift (east of HGP-A) and southeast flank, a layer of 2 ohm-m or less has been mapped generally at depths greater than 1 km below sea level (Skokan, 1974; Kauahikaua, 1981b). The exception to this occurrence is the area within the rift east of HGP-A. The low-resistivity layer comes to within 500 m below sea level and is thought to represent high-temperature fluids at depth (Kauahikaua, 1981b). The few soundings in area II do suggest that the area is one of generally higher resistivity.

The distinction between areas is also clear in the 1979 TDEM data. If we look at the minimum apparent resistivities for each curve plotted in Fig. 25 , we see that soundings 1-4 and 8-4 have the highest values of 28 and 120 ohm-m ; these soundings are near the boundary between the two areas. The rest of the soundings are well within area I and have minimum apparent resistivities between 13 and 23 ohm-m.

The use of larger spacings between source and TDEM sounding locations offers better resolution of deeper structures at the expense of resolution at shallow depths. Each of the apparent resistivity curves in Fig. 25 begins to increase below 3 km indicating a resistive basement layer at those depths. This interface seems to be resolved much better than the first 1.5 km in these soundings, contrary to the 1974 TDEM soundings that could only resolve resistivities down to 1.3 km. The presence of the resistive basement may prevent the TDEM apparent resistivity values from reach-

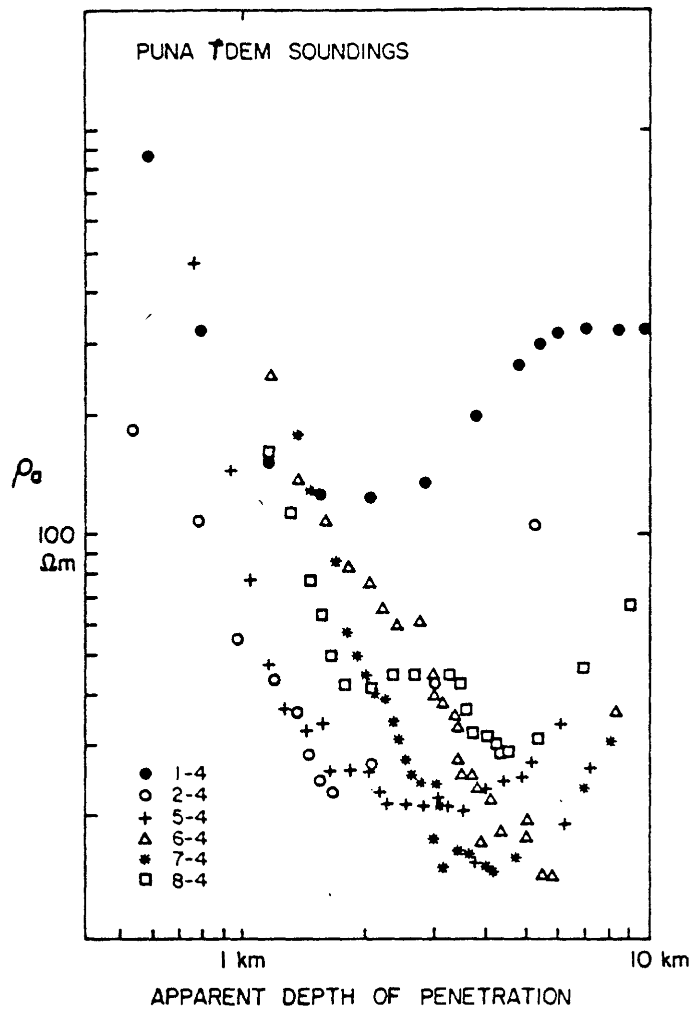



Figure 25: Puna TDEM data presented as apparent resistivities versus apparent depth.

ing the apparent resistivity asymptote for those depths shallower than basement (analogous to the middle-layer resolution problems in a type-H VES curve; see Keller and Frischknecht, 1966, p. 135).

The geologic interpretation of this compilation of electrical geophysical data is rather simple. The shallow subsurface electrical differences seen between areas I and II are readily accounted for by differences in hydrology;

little or no basal lens of freshwater exists beneath area I, whereas a very thick basal lens of cold, freshwater is maintained beneath area II (Davis and Yamanaga, 1973, p. 34-35). The lack of an appreciable lens in area I is attributed, in part, to mixing of seawater and freshwater caused by the thermal effects of the rift.

The deeper electrical structure is attributed primarily to the effects of heat on water-saturated rocks. An anomalously conductive stratum mapped at rather uniform depths of 1000-1300 m below sea level beneath the flanks of the rift and 250-500 m below sea level beneath the portion of the rift downrift of HGP-A has been interpreted to represent water temperatures greater than 200° C (Kauahikaua, 1981b). The 1979 TDEM data strongly suggests that the area I-area II distinction is still seen at depth.

The resistive basement observed below 3 km depth in the 1979 TDEM soundings is probably due to a decrease in porosity caused by ambient pressure conditions at the time and depth of eruption. A similar layer has been delineated beneath the summit of Kilauea (Kauahikaua and others, 1979), Hualalai, and Keaau, and probably does not have much geothermal significance except that it marks the lower boundary of any exploitable resource.

Area I is the most promising part of the east rift for further geothermal exploitation. Temperatures are expected to be greater than 200° C at depths greater than 1 km below

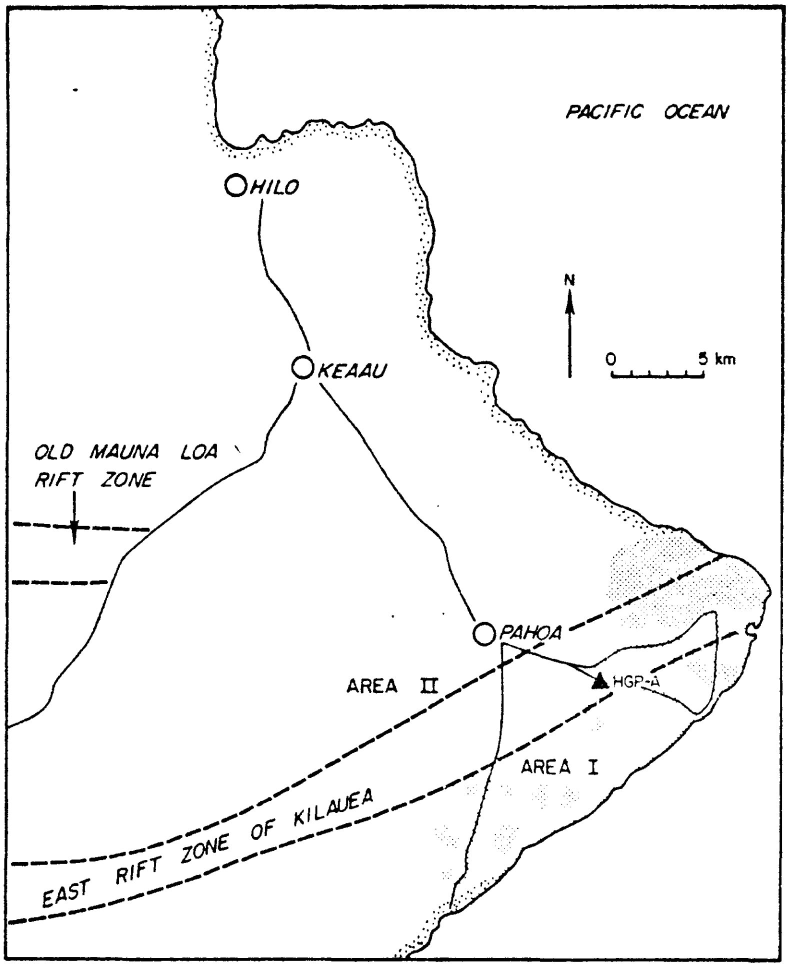



Figure 26: Locations of area I and II on the east rift zone of Kilauea. Area I is the most promising for geothermal exploitation.

sea level beneath the flanks and 250-500 m below sea level within the rift east of HGP-A (Kauahikaua, 1981b). Temperatures may also be high 1 km deep beneath the rift west of HGP-A; however, the chances of high temperatures are probably greater towards the southeastern edge of the rift zone. Subsurface temperatures are much cooler north of the rift zone structure and are probably unsuitable for geothermal development.

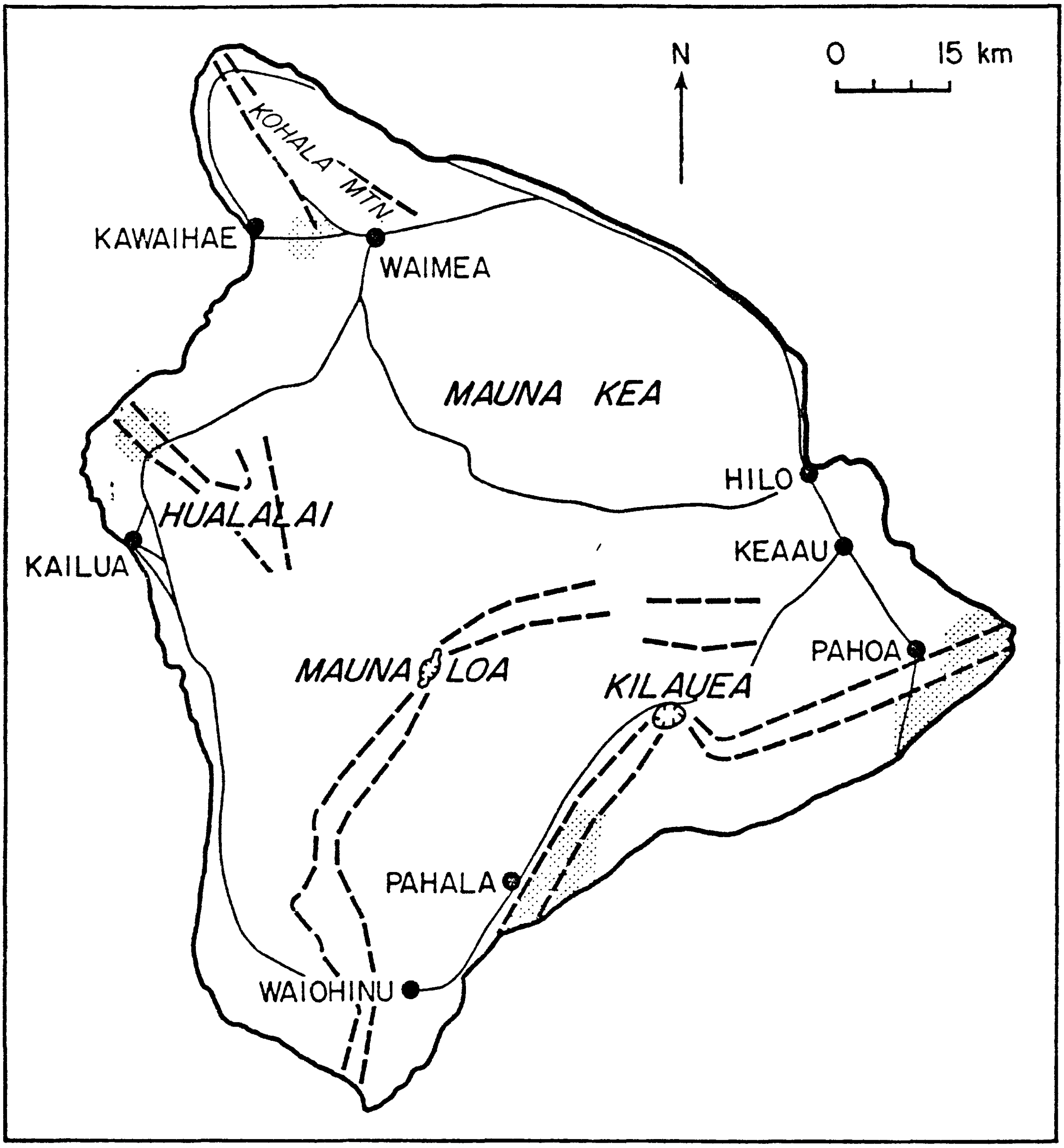



Figure 27: Summary map of exploitable areas on the island of Hawaii.

# 4

# CONCLUSIONS

The Keaau area has been ruled out as a geothermal resource area based on four VES and four TDEM soundings. These results indicate that the seawater-saturated rock unit has a resistivity range from 13 to 30 ohm-m and can be accounted for by typical porosities in Hawaiian basalt. From ten TDEM, three VES and two equatorial DC soundings located in the South Point area, no anomalously low resistivities were found. Although much of the data may be distorted by a extensive pipe-network, there is no indication of a geothermal resource in this area.

In the Kawaihae area, low apparent resistivities (less than 5 ohm-m) are observed near the coastline, while a high resistivity basement coincides with a magnetic anomaly near a warm-water well. The resistive basement may represent a heat source for a moderate-temperature resource in this area. A similar structure to that observed in Kawaihae is observed near Puu Mau on the northwest rift zone of Hualalai. A low resistivity layer (less than 9 ohm-m) is underlain by a more resistive basement. This structure may represent a warm fluid-saturated layer underlain by a resistive heat source.

By far the most data have been collected on the east rift zone of Kilauea Volcano. A specific area of very low resistivity (less than 3 ohm-m) has been outlined, where high-temperature fluids are expected to exist at depth (Fig. 26). A structure similar to the east rift zone has been observed on the southwest rift zone of Kilauea, implying a high-temperature resource is located in this area as well. Future work is suggested in the southwest rift area in order to better assess the size cf the resource. The areas with geothermal promise are presented in Figure 27.

## ACKNOWLEDGEMENTS

The 1979 field work could not have been accomplished without the full-time efforts of Craig Crissinger of the Hawaiian Volcano Observatory (HVO). We also thank the HVO staff, especially Dallas Jackson who generously loaned equipment and time to our effort. Sounding VES 8 at the summit of Hualalai was done in cooperation with Dallas Jackson. Charlie Zablocki and Frank Frischknecht of the USGS in Denver, Colorado, also helped immeasureably with equipment loans and general support.

We thank the following organizations for granting access to their lands: Puu Waawaa Ranch, Hualalai Ranch, Huehue Ranch, Bishop Estate, Parker Ranch, Kahuku Ranch, Kawaihae Ranch and Puna Sugar Co. Thanks also to Dallas Jackson, Barry Leinert, Don Thomas, Walt Anderson, Adel Zohdy and especially Doug Klein for their reviews of the manuscript.

A special note of thanks to Dick and Jane Webb of Hilo, Hawaii, for their warm hospitality during this survey period.

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## REFERENCES

Abbott, A.T., 1975, Photogeologic survey: imagery from infrared scanning of the rift zones of Kilauea and Mauna Loa, <u>in</u> Summary Report for Phase I: Hawaii Geothermal Project Report, p. 29-32.

Adams, W.M., 1968, An electrical resistivity profile from Mahukona to Kawaihae, Hawaii: Water Resources Research Center technical report no. 23, 32 p.

Adams, W.M., Peterson, F.L., Mathur, S.P., Lepley, L.K., Warren, C., and Huber, R.D., 1969, A hydrogeophysical survey from Kawaihae to Kailua-Kona, Hawaii: Water Resources Research Center technical report no. 32, 156 p.

Adams, W.M., Mathur, S.P., and Huber, R.D., 1970, Aeromagnetic, gravity, and electrical resistivity exploration between Pahala and Punaluu, Hawaii: Water Resources Research Center technical report no. 28, 62 p.

Anderson, W.L., 1979, Program MARQDCLAG: Marquardt inversion of DC-Schlumberger soundings by lagged-convolution: U.S. Geological Survey Open-File Report 79-1432, 58 p.

Davis, D.A., and Yamanaga, G., 1973, Water resources summary: island of Hawaii : U.S. Geological Survey Report R47, 42 p.

Fischer, W.A., Moxham, R.M., Polcyn, F. and Landis, G.H., 1964, Infrared surveys of Hawaiian volcanoes: Science, v. 146, p. 733-742.

Fornari, D.J., Peterson, D.W., Lockwood, J.P., Malahoff, A., and Heezen, B.C., 1979, Submarine extension of the southwest rift zone of Mauna Loa, Hawaii: visual observations from U.S. Navy deep submergence vehicle DSV Sea Cliff: GSA Bull., v. 90, p. 435-443.

Group Seven, 1978, The Puu Waawaa geothermal prospect: unpublished report, 24 p.

Holcomb, R.T., 1980, Preliminary geologic map of Kilauea volcano, Hawaii: U.S. Geological Survey Open-File Report 80-796, 2 sheets.

Hussong, D.M. and Cox, D.C., 1967, Estimation of ground-water configuration near Pahala, Hawaii using electrical resistivity techniques: Water Resources Research Center technical report no. 17, 35 p.

Inman, 1975, Resistivity inversion with ridge regression: Geophysics, v. 40, p. 798-817.

Jackson, D.B. and Keller, G.V., 1972, Electromagnetic sounding survey of the summit of Kilauea volcano, Hawaii: JGR, v. 77, p. 4957-4965.

Kauahikaua, J., 1980, Program MQLVTHXYZ: Computer inversion of three-component, time-domain, magnetic-field sounding data generated using an electric wire source: U.S. Geological Survey Open-File Report 80-1159, 109 p.

Kauahikaua, J., 1981a, Interpretation of Time-Domain Electromagnetic soundings in the Calico Hills Area, Nevada Test Site, Nye County, Nevada: U.S. Geological Survey Open-File Report 81-437, 36 p.

Kauahikaua, J., 1981b, Interpretation of Time-Domain Electromagnetic soundings in the East-Rift geothermal area of Kilauea volcano, Hawaii: U.S. Geological Survey Open-File Report 81-979, 24 p.

Kauahikaua, J. and Klein, D.P., 1977a, Electromagnetic induction sounding measurements in the Puna district: in Geoelectric Studies on the East Rift, Kilauea volcano, Hawaii island: Hawaii Institute of Geophysics technical report HIG-77-15, p. 91-119.

Kauahikaua, J. and Klein, D.P., 1977b, Interpretation of electromagnetic transient soundings made on the east rift of Kilauea volcano: in Geoelectric Studies on the East Rift, Kilauea volcano, Hawaii island: Hawaii Institute of Geophysics technical report HIG-77-15, p. 121-173.

Kauahikaua, J., Zablocki, C.Z. and Jackson, D.B., 1979, Controlled-Source Electromagnetic mapping at the Summit of Kiluaea Volcano, Hawaii (abs.): Am. Geophy. Union EOS, v. 60, p. 811-812.

Keller, G.V. and Frischknecht, F.C., 1966, Electrical Methods in Geophysical Prospecting: Pergamon Press, Oxford, 517 p.

Keller, G.V. and Rapolla, A., 1976, A comparison of two electrical probing techniques: IEEE Trans. on Geoscience Electronics, v. GE-14, p. 250-256.

Keller, G.V., Skokan, C.K., Skokan, J.J., and Daniels, J., 1977, Electrical resistivity and time-domain electromagnetic surveys of the Puna and Ka'u districts, Hawaii: in Geoelectric Studies on the East Rift, Kilauea volcano, Hawaii island: Hawaii Institute of Geophysics technical report HIG-77-15, p. 1-89.

Klein, D.P. and Kauahikaua, J., 1975, Geoelectric-geothermal exploration on Hawaii island: Preliminary results: Hawaii Institute of Geophysics technical report HIG-75-6, 23 p.

Macdonald, G.A., 1973, Geological prospects for development of geothermal energy in Hawaii: Pacific Science, v. 27, p. 209-219.

Macdonald, G.A., 1977, Geology and ground-water hydrology: in Preliminary studies for geothermal exploration in Hawaii, 1973-1975: Hawaii Institute of Geophysics technical report HIG-75-5, p. 19-28.

Macdonald, G.A. and Abbott, A.T., 1970, Volcanoes in the Sea: University of Hawaii Press, Honolulu, 441 p.

Malahoff, A. and Woollard, G.P., 1965, Magnetic surveys over the Hawaiian Ridge: Hawaii Institute of Geophysics technical report HIG-65-11, 80 p.

Malinowski, M.J., 1977, Geology of the Kawaihae quadrangle, Kohala mountain, Island of Hawaii: Master's thesis, Univ. of Hawaii, 156 p.

Mundry, E. and Worzyk, P., 1979, On the coastal effect on geoelectric soundings: Journal of Geophysics, v. 45, p. 329-336.

Mundry, E., 1980, The effect of a finite distance between potential electrodes on Schlumbrger resistivity measurements-a simple correction graph: Geophysics, v. 45, no. 12, p. 1872-1875.

Roy, A. and Apparao, A., 1971, Depth of investigation in Direct current methods: Geophysics, v. 36, no. 5, p. 943-959.

Skokan, C.K., 1974, A time-domain electromagnetic survey of the east rift zone, Kilauea volcano, Hawaii: Ph.D. dissertation no. 1700, Colorado Sch. Mines, 150 p.

Suyenaga, W., 1978, Crustal structure of the east rift zone of Kilauea, Hawaii from seismic refraction: 1. near-surface structure: in Seismic Studies on Kilauea volcano, Hawaii island: Hawaii Institute of Geophysics technical report HIG-78-8, p. 77-92.

Thomas, D., Cox, M., Erlandson, D., and Kajiwara, L., 1979, Potential geothermal resources in Hawaii: A preliminary regional survey: Hawaii Institute of Geophysics technical report HIG-79-4, 103 p.

Zablocki, C.J., Tilling, R.J., Peterson, D.W., Christianson, R.L., Keller, G.V., and Murray, J.C., 1974, A deep research hole at the summit of an active volcano, Kilauea, Hawaii: Geophysical Research Letters, v. 1, p. 323-326.

Zablocki, C.J.,1976, Mapping thermal anomalies on an active volcano by the self-potential method, Kilauea, Hawaii: in Proc. of the second UN Symp. on Develop. and Use of Geothermal Resources, v. 2, p. 1299-1309.

Zablocki, C.J., 1977, Self-potential studies in east Puna, Hawaii: in Geoelectric Studies on the East Rift, Kilauea volcano, Hawaii island: Hawaii Institute of Geophysics technical report HIG-77-15, p. 175-195.

Zablocki, C.J. and Koyanagi, R.Y., 1979, An anomalous structure in the lower east rift zone of Kilauea volcano, Hawaii, inferred from geophysical data (abs.): Hawaii Symp. on Intraplate Volcanism and Submarine Volcanism, p. 177.

Zohdy, A.D.R. and Jackson, D.B., 1969, Application of deep electrical soundings for groundwater exploration in Hawaii: Geophysics, v. 34, p. 584-600.