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Geoelectric-Geothermal Exploration on Hawaii Island: Preliminary Results

by

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HAWAII INSTITUTE OF GEOPHYSICS
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ABSTRACT

Geoelectric reconnaissance surveys were performed on Hawaii Island to locate areas of low resistivity that might have the potential for commercial geothermal development. The lower northeast end of the Kilauea East Rift Zone has low resistivities (less than 10 ohm-m) that indicate anomalous geothermal conditions. Preliminary results of both AC and DC resistivity surveys indicate that the most promising zone of high temperature waters is at depths of 500 to 1500 meters beneath the 1955 eruptive vents in this region. The temperature in this zone may be as great as 200°C. Further analysis is continuing on the data and additional data are being gathered to attempt to outline the probable extent of the material with low resistivity.

INTRODUCTION

Exploration of the electrical resistivity structure within the upper 2 kilometers of the crust on Hawaii Island has been undertaken as part of a general investigation of the potential of utilizeable geothermal resources in Hawaii. Both controlled-source AC (electromagnetic induction) and DC methods were used in the exploration program which was undertaken in two phases: (1) reconnaissance work to locate potential geothermal targets; (2) detailed surveys to assess the possible development of potential geothermal areas.

This report deals primarily with the results of the initial electrical reconnaissance surveys and to a lesser degree with the preliminary results of the more detailed surveys that are still underway.

Horizontal mapping and vertical probing of the earth's electrical properties are primary tools in prospecting for anomalous geothermal regions (Keller 1970, 1971; Meidav, 1970). The basis for this is the experimental evidence that temperature variations markedly alter the electrical resistivity of water-bearing rocks (Keller and Frischknecht, 1967; Keller, 1970, 1971; Hermance et al., 1972). The resistivity of wet rocks decreases roughly as a negative exponential function of the temperature increase if other factors remain unchanged. Since the resistivity of crustal rocks is determined primarily by pore waters, the rock porosity and ionic concentration of the pore waters are also major factors in determining resistivity. Figure 1 illustrates the effect of these factors on the resistivity of water-saturated basalt having 16% porosity. The solid lines plot resistivity against temperature for the rock saturated in water having an NaCl concentration equivalent to seawater (3,500 parts per million) and fresh water (500 parts per million). The dashed lines indicate the range of resistivity variations that can be expected for a ±50% variation in porosity. Note that either an increase in porosity or an increase in ionic concentration decreases the resistivity of a rock.

It is also known, though less well understood, that thermal regions can be associated with anomalous static electrical potentials in the earth (Zohdy et al., 1973; C. Zablocki, personal communication). This last relationship is the basis for applying the self-potential mapping surveys to geothermal prospecting.

The resistivity reconnaissance surveys were performed in the general areas indicated in Figure 2, chosen largely on the basis of recently active volcanism or known existence of anomalous thermal conditions (Macdonald, 1973).

The geoelectric methods applied to the present surveys were:

1. Dipole-dipole probing carried out by G. V. Keller and his associates from the Colorado School of Mines with support from the University of Hawaii.

2. Self-potential mapping, carried out by C. Zablocki of the U.S.G.S. with support from the Hawaii Institute of Geophysics.

3. Schlumberger vertical profiling, carried out cooperatively by C. Zablocki of the U.S.G.S. and the Hawaii Institute of Geophysics.

4. Horizontal-loop/loop induction soundings.

5. Horizontal-line/loop induction soundings.

The last two surveys were performed by personnel from the Hawaii Institute of Geophysics. The descriptive results relating to each method that follow are not necessarily in chronological order, but are in the sequence with which our present thinking about the geoelectric physics of the island of Hawaii has developed. Only brief reference will be made to the self-potential results, as a detailed report is in preparation by C. Zablocki.

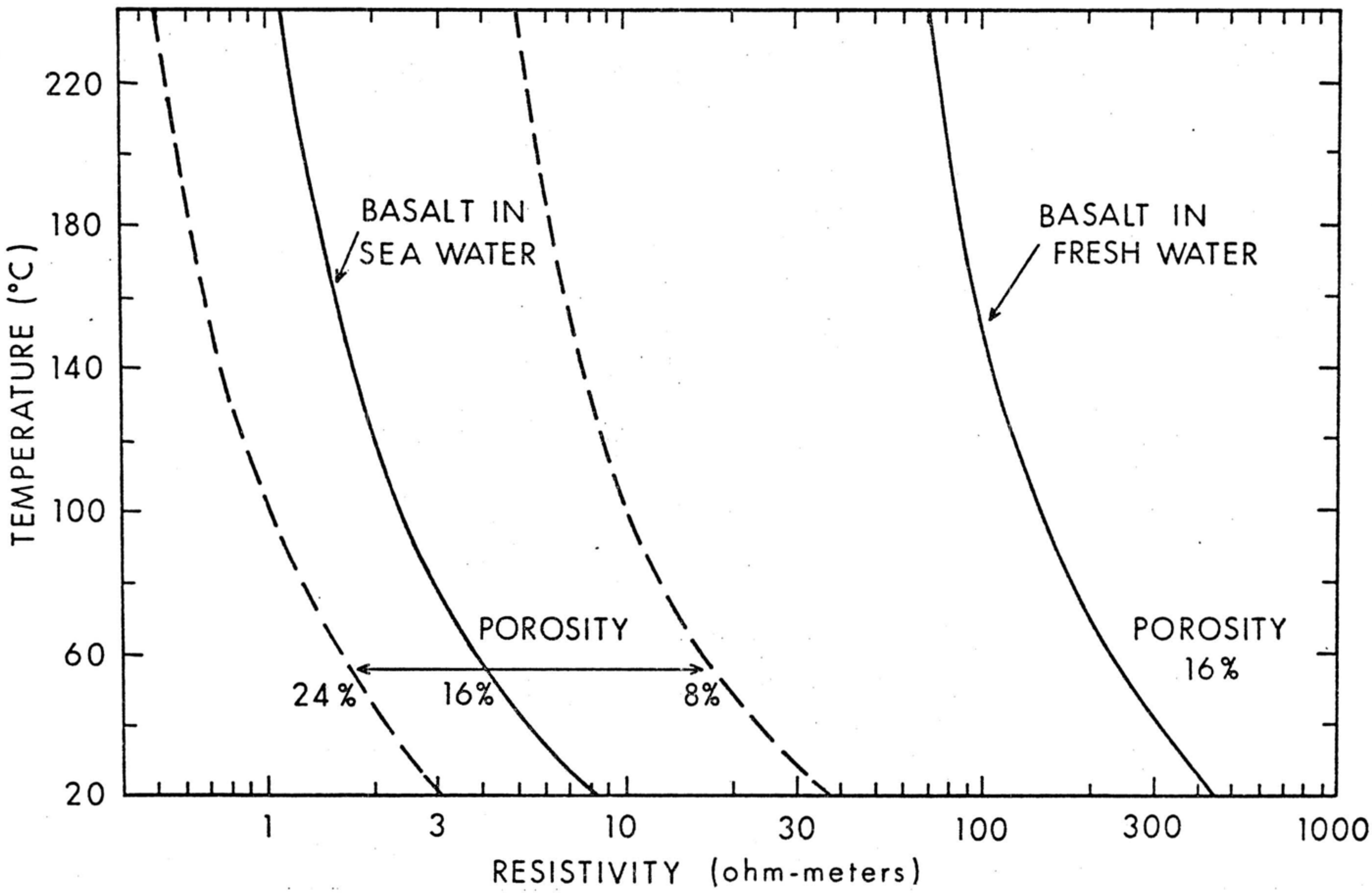

Fig. 1. Resistivity relationships. Solid lines show resistivity versus temperature for water-saturated basalt with 16% porosity. Dashed lines show the resistivity variation expected for porosity changes of ± 50% from a reference porosity of 16%. Fresh and sea water are taken as 50 ppm and 35,000 ppm equivalent NaCl concentration, respectively. Based on Meidav's nomogram (1970).

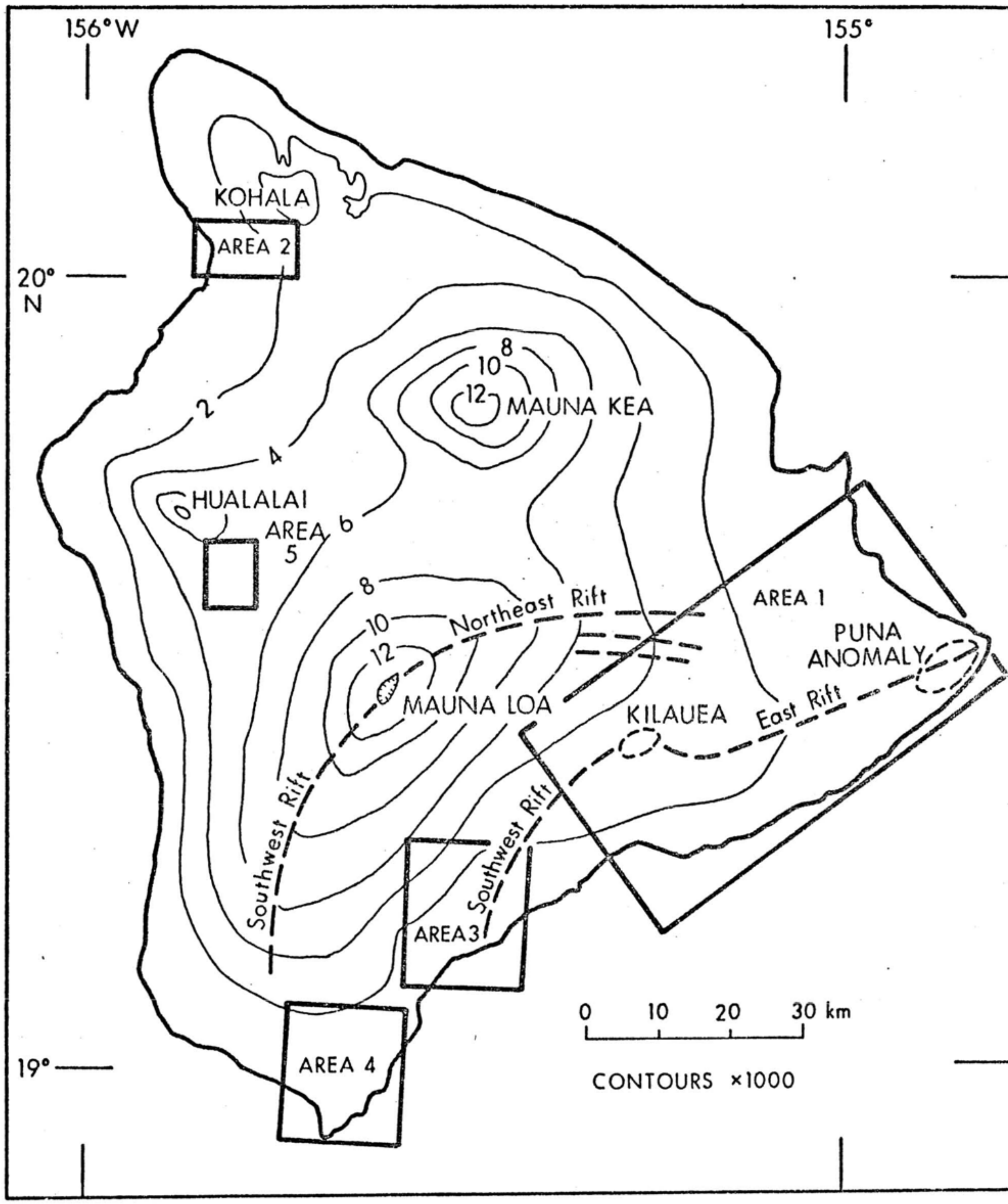

Fig. 2. Areas of resistivity reconnaissance surveys on Hawaii Island. Elevation contours are feet x 1000.

DIPOLE-DIPOLE PROBING

"Dipole-dipole" probing results described here are a condensation of work done by G. V. Keller, J. Daniels, J. Skokan and K. Skokan (Keller, 1973). The survey technique, a variant of the "dipole-dipole" method described in detail by Keller (1966), and Alpin et al. (1966), uses a pair of current electrodes to establish a static voltage potential in the earth. The separation of these source electrodes is commonly taken as infinitesimally small for analytic calculations (dipole approximation); however, in the present case, the finite separation between source electrodes was taken into account (as a "bipole", Keller, 1973). The gradient of the source potential distribution is characteristic of the earth's electrical resistivity structure and when mapped by a set of passive electrodes set up at various locations about the active electrodes can be interpreted as an "apparent" resistivity (Keller and Frischknecht, 1966). The apparent resistivities are displayed on a map to provide an indication of horizontal resistivity structure (dipole - mapping) or on a graph as a function of separation between the current and voltage electrodes, in which case it is possible to interpret the mean vertical resistivity profile (dipole-sounding).

Dipole-dipole surveys were applied primarily in the Puna District of Hawaii which includes the active Kilauea Volcano and the area to the northeast (area 1, Fig. 2) (Keller, 1973). The northeast rift zone of Kilauea Volcano traverses this area and is the locus of recent (1955-1962) volcanic extrusions and steam seeps, especially along the lower portion near the eastern point on Hawaii Island (Macdonald, 1973). Abnormal hydrothermal conditions are also found in scattered warm water wells and pools seaward of the lower part of the rift (Davis and Yamanaga, 1968).

Keller's group (1973) also made measurements in a small region on the northwest coast of Hawaii (area 2, Fig. 2) where water wells show slightly anomalous temperatures of a few degrees (°C) above normal. These latter measurements did not provide evidence of low resistivities that could be geothermally generated, and will not be considered further.

The dipole-dipole survey isolated an anomalous region on the lower part of the rift zone of Kilauea Volcano, which will be called the "Puna Anomaly" in this report (see Fig. 2). Apparent resistivities in the area of this anomaly range from 5 to 20 ohm-meters (ohm-m) compared to values of greater than 200 ohm-m measured elsewhere outside the immediate volcanic edifice of Kilauea. Figure 3 shows the generalized results from this area. The lowest resistivities, 2.5 to 10 ohm-m, occur as a few localized anomalies identified as the areas where contours of low apparent resistivity from different source electrodes overlap.

Plots of the mean apparent resistivity against source-receiver separation for the Puna District dipole survey are reproduced in Figure 4. These data provide a rough estimate of the vertical resistivity distribution and illustrate the anomalous low apparent resistivities (solid lines) of the lower regions of the East Rift Zone in comparison to data in other parts of Puna. A typical section in the anomalous area has a resistivity of 9 to 80 ohm-m to a depth of about 500 meters, 5 to 8 ohm-m maximum in the depth range of roughly 500 meters to 2000 meters, and a very resistent material below 2 kilometers.

Resistivities of 5 to 8 ohm-m are of the correct order of magnitude to be associated with geothermal waters. However, the anomalous area in Puna is at low elevation (generally below 300 meters elevation) and the effect of saline waters at depths greater than sea level also would lower the resistivity irrespective of temperature controls.

The low-resistivity layer from 500 to 2000 meters depth is apparent in the data from all of the sites in the anomalous area (Fig. 3), and if associated with hot water represents a significant resource.

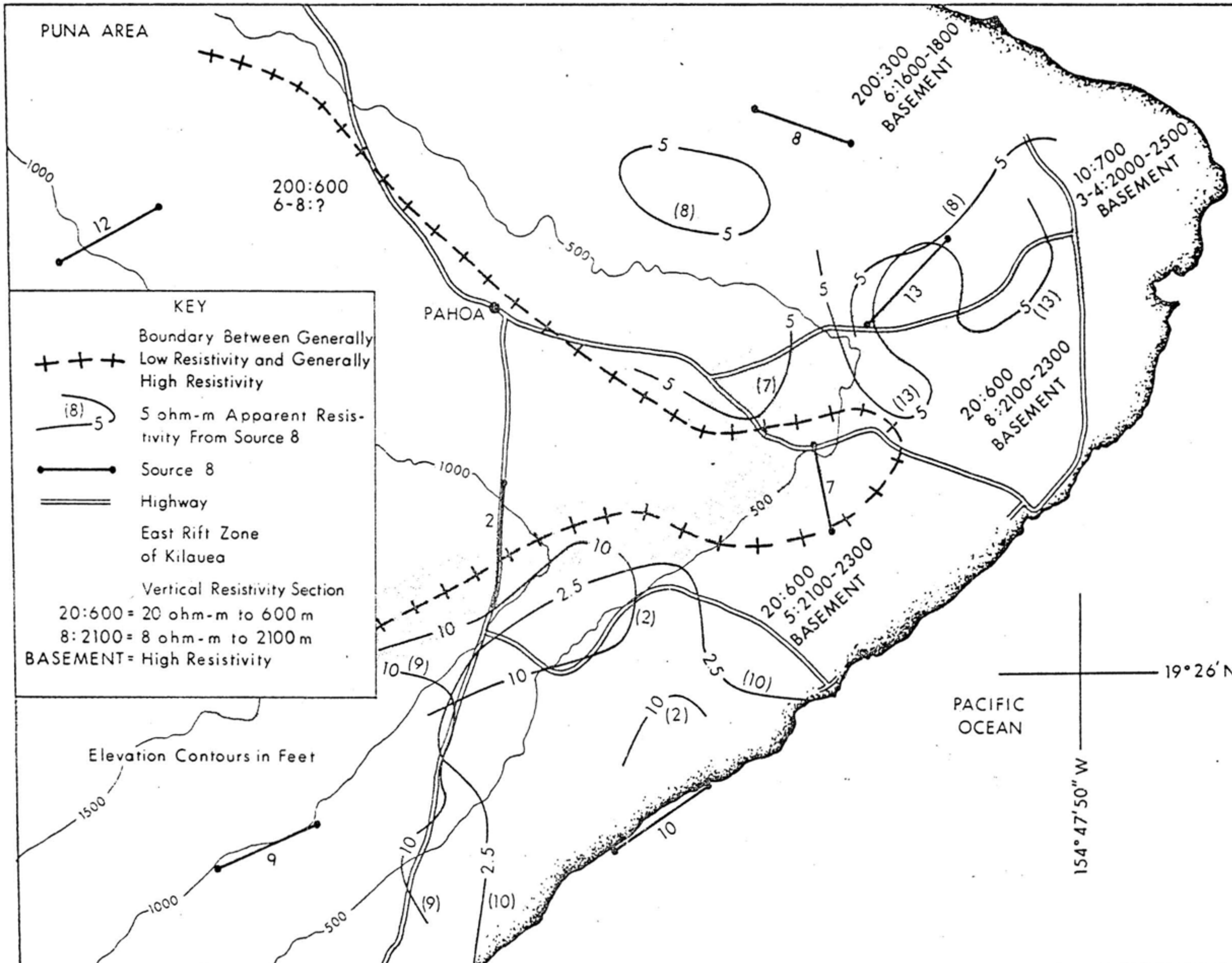

Fig. 3. Results of dipole-dipole mapping in the lower east rift of Kilauea, the Puna Anomaly (from G. V. Keller, 1973). See Figure 2 for general index of location.

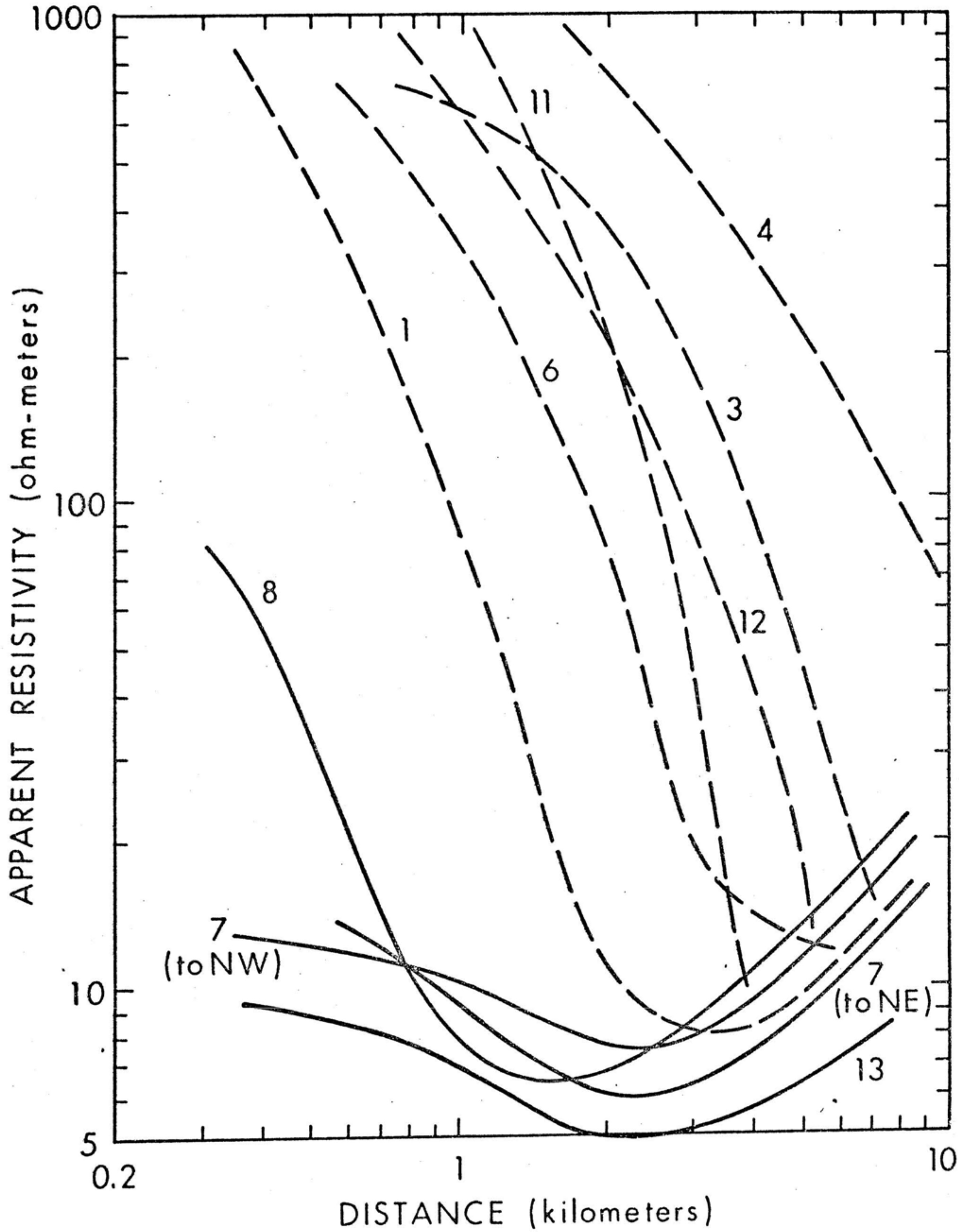

Fig. 4. Generalized trends of apparent resistivity versus source-receiver separation of the dipole-dipole resistivity survey. The solid lines are the trends from stations in the Puna Anomaly as numbered in Figure 3. The data from stations of sources 2 and 10 (omitted) showed such great scatter that distance-resistivity trends were not meaningful (Keller, 1973). The dotted lines are the trends of selected data from stations outside the Puna Anomaly (from G. V. Keller, 1973).

LINE-LOOP INDUCTION SOUNDINGS

Inductive methods use time-varying magnetic fields to develop electric voltages in the earth according to Faraday's law (Keller and Frischknech, 1966). A controlled magnetic source field is generated from a current line grounded to the earth or from a closed current loop (see the schematic in Fig. 5). This magnetic field is modified by secondary magnetic fields associated with electric currents in the earth, the latter controlled by the induced voltages and earth resistivity according to Ohm's law. Thus both the magnetic fields and ground potentials are diagnostic of earth resistivity. Induced ground voltages can be measured directly with a set of passive voltage electrodes or indirectly by measuring the total magnetic flux of both the induced earth currents and the source currents. For the present measurements, the source field was produced by a 0.5 to 1.5-kilometer grounded wire carrying a time-varying current with maximum amplitude of 1 to 10 amperes. The total magnetic flux was measured using an induction coil placed on the ground. Details of this technique can be found in Keller (1970, 1971).

Since earth currents for a given voltage gradient are inversely proportional to resistivity (Ohm's law), inductive methods are best suited to prospecting for conductors (Keller, 1971). Also, the magnetic flux of all electric currents is additive, thus the method generally tends to smooth out the effects of lateral inhomogeneities. The combined result is that the method is generally not very precise in determining fine detail in resistivity structures. Our instrumentation was designed along the lines of the system described by Jackson and Keller (1972) and is specifically suited for deep (2-kilometer) penetration with the objectives of:

(1) Reconnaissance surveys of areas not covered by Keller's group, and;

(2) Further study of the low resistivity zone found by Keller (1973) below 500 meters depth on the east rift of Kilauea with a technique that would be less sensitive to lateral inhomogeneities in the shallower resistivity structure.

Reconnaissance induction soundings were made in the lower regions of the southeast rifts of both Kilauea and Mauna Loa volcanoes (areas 3 and 4 respectively, Fig. 2). Soundings were also made in the saddle area between the Hualalai and Mauna Loa domes (area 5, Fig. 2).

The rift zone areas are prospective geothermal areas because they have been the locus of recent magmatic activity and display local hydrothermal activity in the form of warm surface rocks or warm water seeps (Macdonald, 1973). Specific areas chosen for investigation were largely determined by aerial photographic infrared anomalies (A. T. Abbott, personal communication). The decision to survey the Hualalai region was based primarily on the consideration that older volcanoes if underlain by residual magma pockets may have had time to form hydrothermally sealed reservoirs of geothermal fluids. For instance, Hualalai, as compared to the still-active volcanoes of Mauna Loa and Kilauea (Macdonald and Abbott, 1970), last erupted in 1801.

Nineteen soundings were made in the regions of the southwest rift of Kilauea, 22 in the southwest rift of Mauna Loa, and eight on the Hualalai-Mauna Loa saddle. Maps of the areas are given in Figures 6, 7, and 8. The data are not all yet formally processed, as reduction priorities have been biased toward the Kilauea east rift data discussed below. Nevertheless, based on field reduction that gives a minimum apparent resistivity (see Appendix) it is fairly certain that none of the above areas contain significant geothermal prospects. As Table 1 shows, the apparent resistivities in these regions are higher than expected for a hot-water-saturated rock which is expected to have a resistivity of less than about 10 ohm-m. Although more precise analysis is in progress, it is unlikely that the results will be encouraging for geothermal purposes.

The above reconnaissance data, including those of the dipole-dipole survey, have directed our primary effort to the lower part of the East Rift Zone of Kilauea, the Puna Anomaly. Thirty-one line-loop induction soundings have been performed in this area (Fig. 9). Table 2 gives the apparent resistivities calculated in the field as well as "formal" apparent resistivities made as a preliminary step in "formal" reduction. The final analysis, still underway, is largely based on interpretation curves developed by Silva (1969). Our "formal" apparent resistivities given here were determined by fitting our reduced data to a homogeneous-earth curve. This gives some indication of the precision of our field interpretation. Work still underway will determine whether the data warrant further resolution of the conductivity structure into models having two or more layers.

The data generally agree with the dipole-dipole results for the low resistivity zone below 500 meters depth. The mean inductive resistivity (using "formal" resistivity values and excluding station 29 because it is outside the anomalous area) is 10.4 ohm-m, slightly higher than Keller's estimate

Table 1. Minimum Apparent Resistivities (ρ_a, in ohm-m), from Reconnaissance Line-Loop Induction Soundings

Area	Station - Source	Half-Separation (meters)	ρ_a
Kilauea, Southwest Rift	1 - 1	1690	18
	2 - 1	2205	53
	4 - 1	1190	19
	5 - 1	2525	50
	6 - 2	1260	14
	7 - 2	1990	48
	8 - 2	2275	56
	9 - 2	1090	13
	12 - 3	1565	28
	15 - 3	1120	13
	18 - 3	2355	14
	19 - 3	1600	20
Mauna Loa, Southwest Rift	1 - 1	1385	13
	3 - 1	1745	12
	4 - 1	2215	17
	5 - 1	2625	19
	7 - 1	2425	17
	8 - 1	2210	33
	15 - 2	1475	13
	18 - 2	1775	29
	19 - 2	2215	23
	20 - 2	1280	19
Mauna Loa-Hualalai Saddle	1 - 1	1925	38
	2 - 1	1325	13
	5 - 1	1295	19
	6 - 2	2355	40

Table 2. "Minimum" and "Formal" Apparent Resistivities (ρ_a, in ohm-m) from Line-Loop Induction Soundings in Puna

Magnetic Induction Station and Source		R/2 (meters)	ρ_a Field Reduction	ρ_a Formal Reduction
1	(I)	839	4.0	11.
3	(I)	1310	8.7	6.7
4	(I)	1765	11.0	-
5	(I)	1375	7.1	10.
6	(I)	1570	12.0	-
7	(I)	1250	7.7	13.0
10	(I)	1330	1.3	-
12	(I)	1850	6.7	-
14	(II)	1020	3.7	5.
15	(II)	1750	5.0	-
16	(II)	1550	4.8	-
18	(II)	1330	3.7	-
21	(III)	1495	7.7	7.7
22	(III)	975	3.6	-
23	(III)	645	2.8	3.8
24	(III)	1225	3.5	-
25	(IV)	890	5.9	23
26	(IV)	1040	6.3	15
27	(IV)	1645	6.7	6.2
28	(IV)	1325	9.1	13
29	(IV)	1595	17.0	48
30	(IV)	2230	10.0	-
31	(IV)	1850	5.6	-

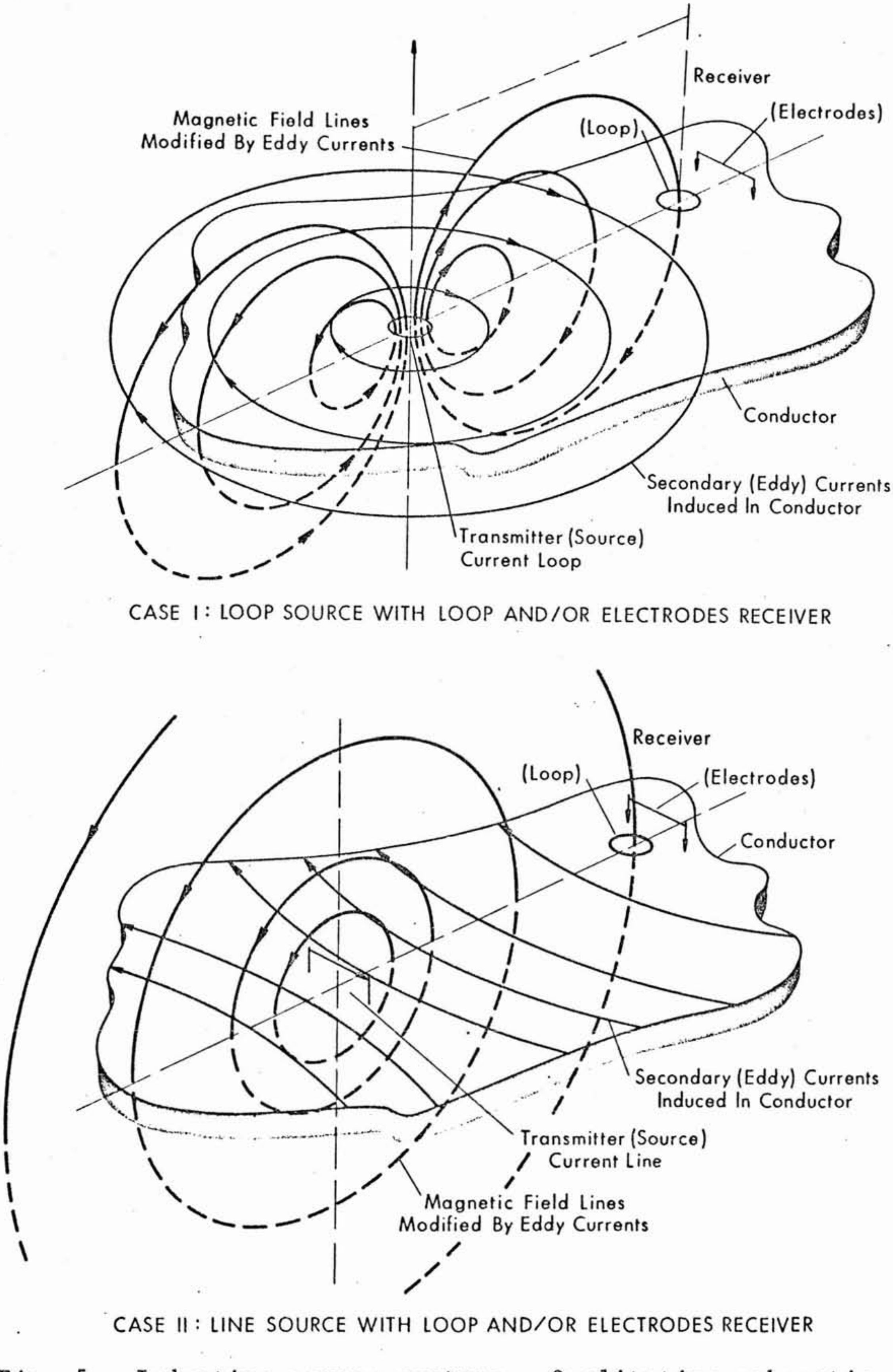

Fig. 5. Inductive survey systems. Qualitative schematic illustrating the relationships between magnetic fields and induced earth currents for various inductive source-receiver configurations on a uniform horizontal conducting layer.

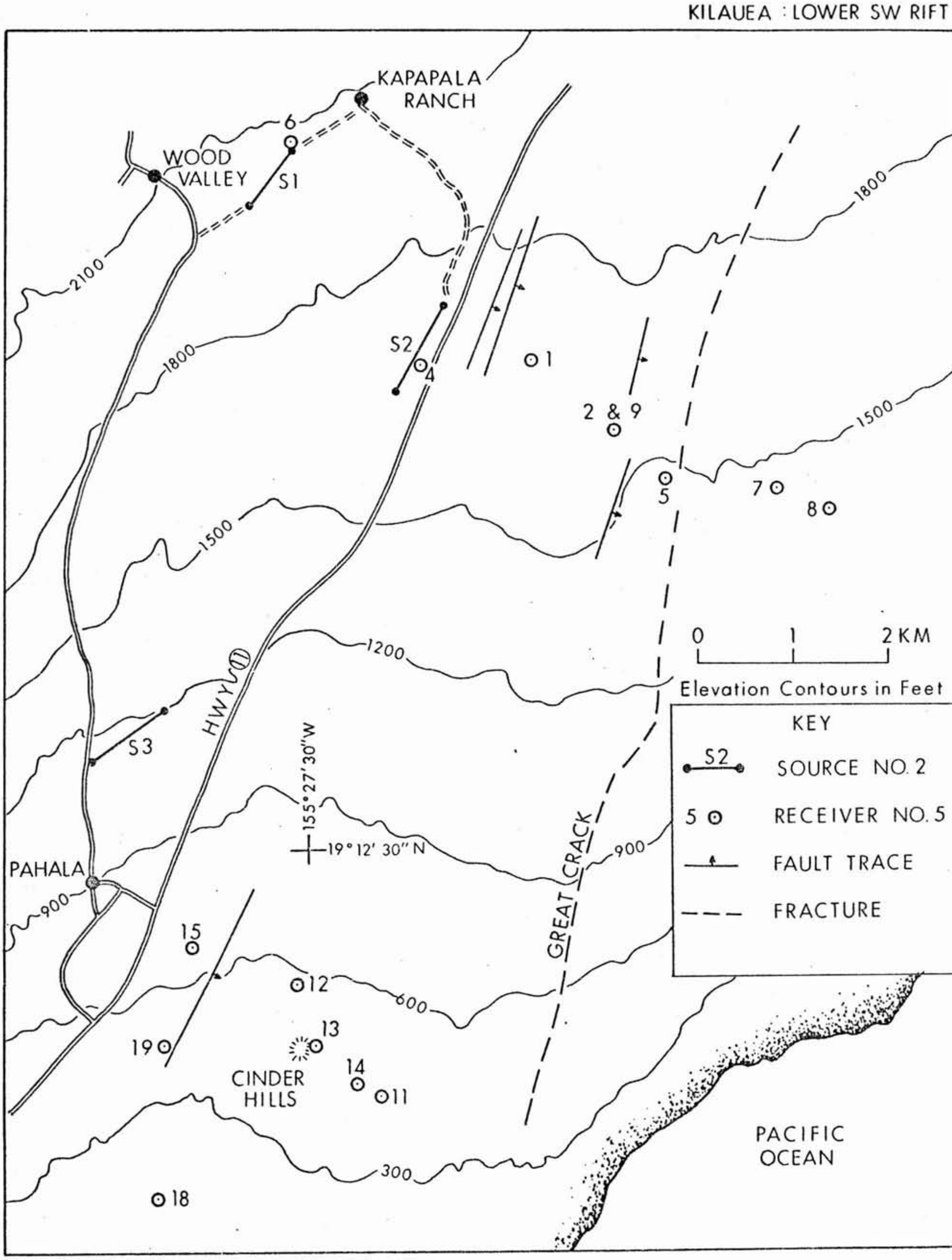

Fig. 6. Line-loop induction stations in the area of the southwest rift of Kilauea (area 3).

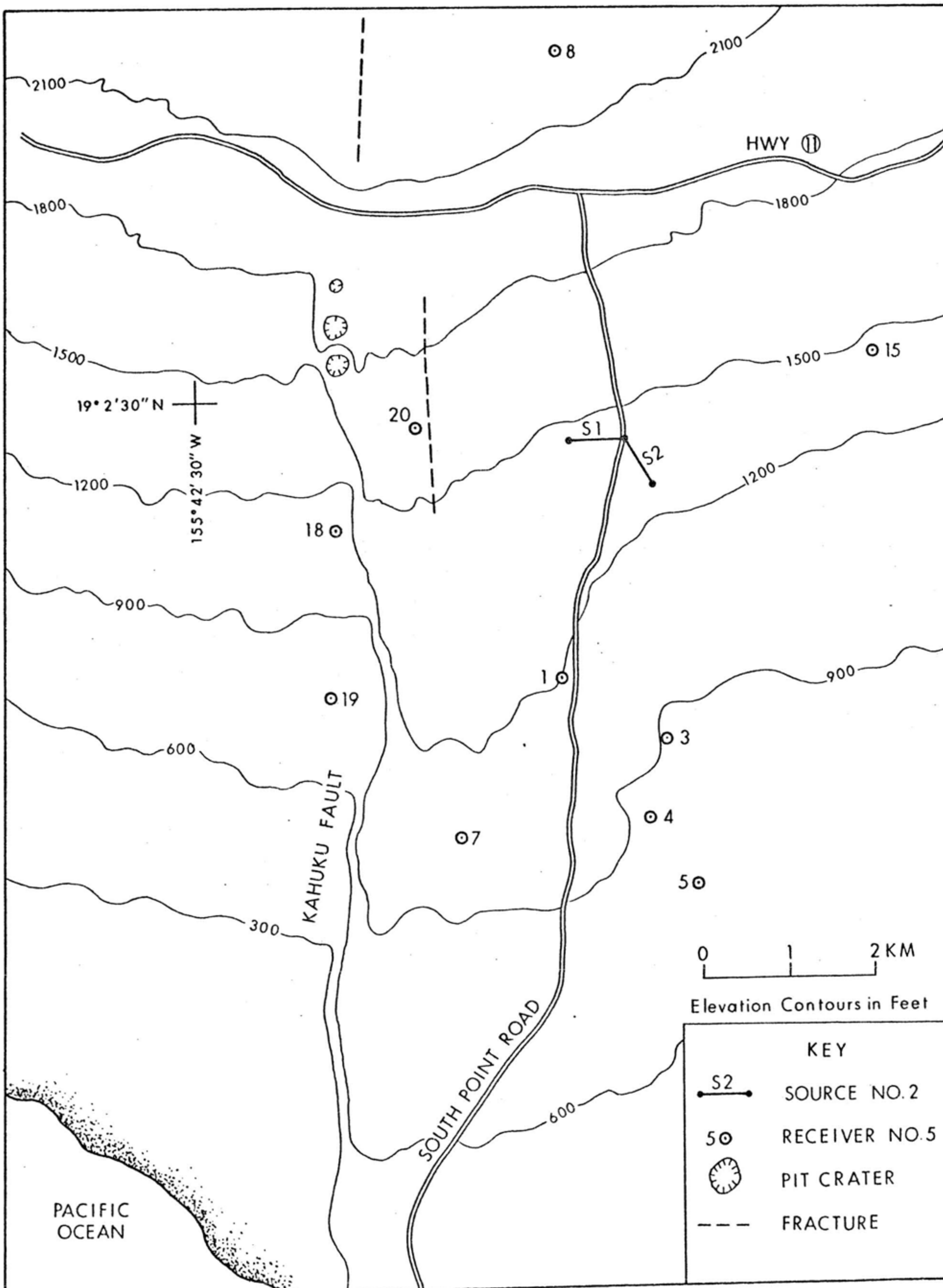

Fig. 7. Line-loop induction stations in the area of the southwest rift of Mauna Loa (area 4).

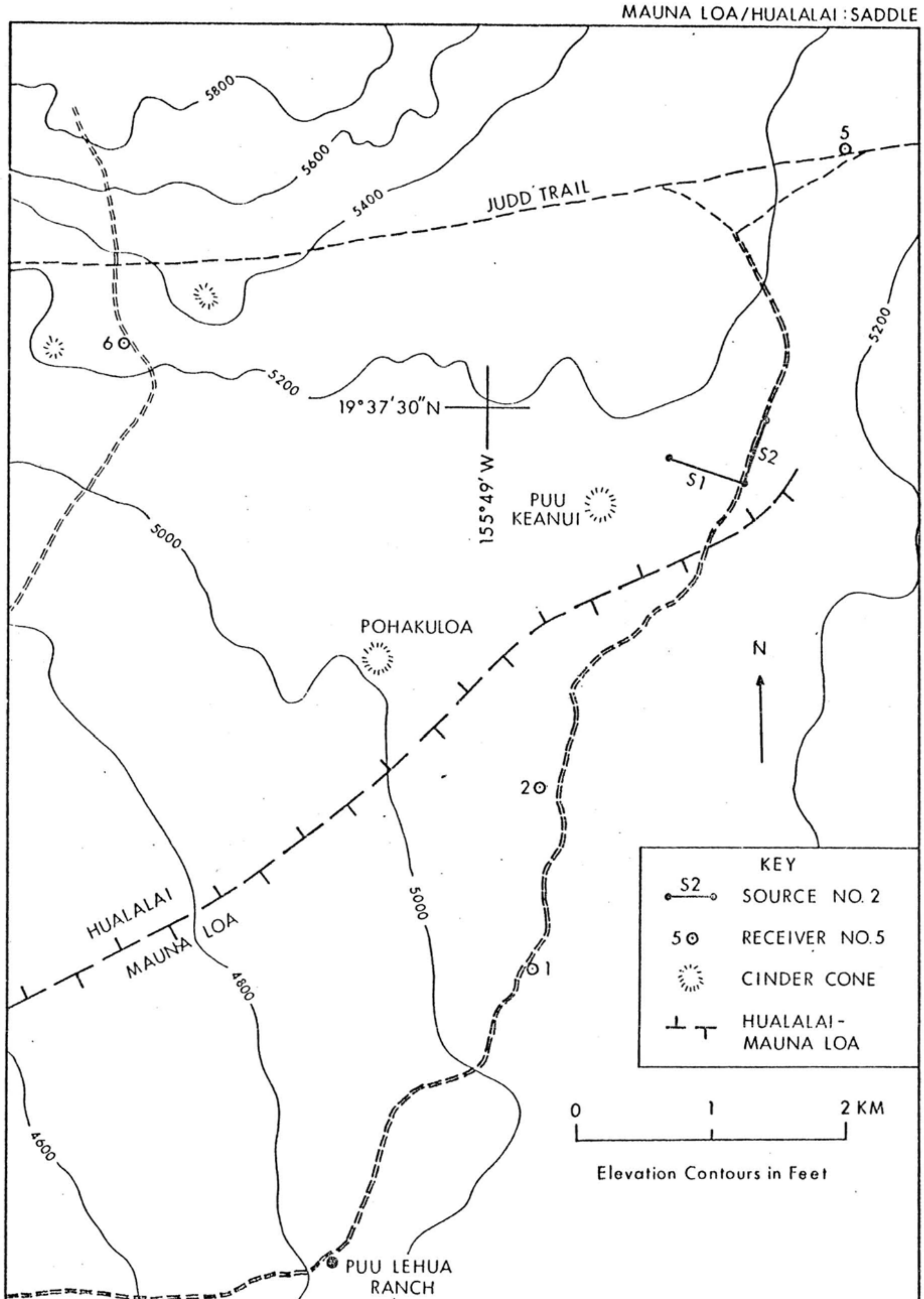

Fig. 8. Line-loop induction station in the saddle area between Hualalai and Mauna Loa (area 5).

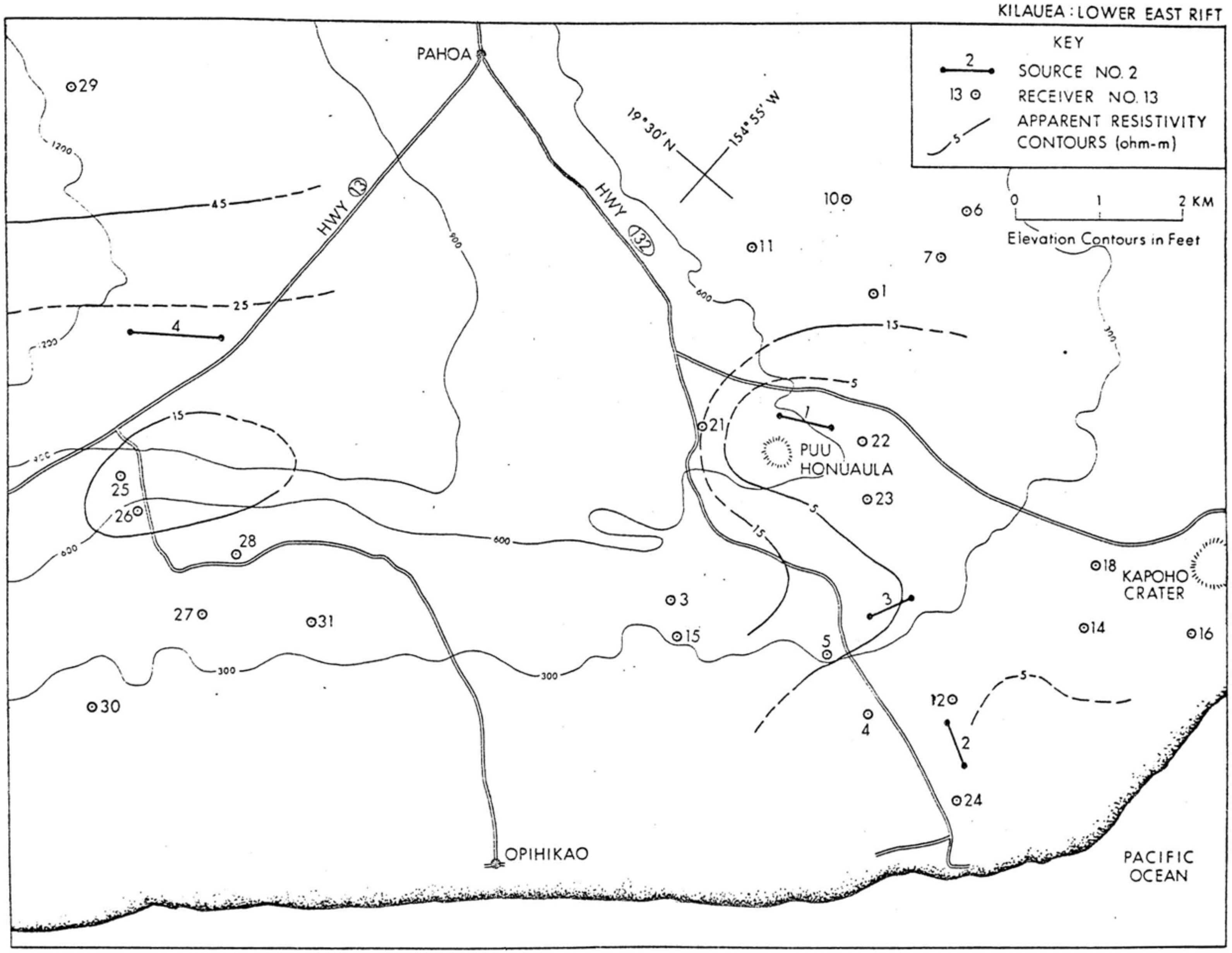

Fig. 9. Line-loop induction stations in the lower east rift of Kilauea (Puna Anomaly). The generalized contours of apparent resistivity are controlled by placing the data at the half-separation points between source and receivers. The rift zone roughly trends from the northwest, just above the intersection of Highway 13 and the Opihikao Road to the northeast through Kapoho (see Fig. 3).

of about 5 ohm-m, based on dipole data. This difference may be due to the fact that the inductive results have integrated the effect of higher resistivities at shallow depth due to our preliminary reduction technique. According to the spatial distribution of data, it appears that the lowest resistivities are found seaward of the rift along the most eastern portion of the Kilauea East Rift Zone between Puu Honuaula and Kapoho (see Fig. 9). There is also a low to the southwest.

These data, combined with dipole-dipole results, give a fairly unambiguous picture of a low-resistivity mass at depths of 500 to 1500 meters below the Puna Anomaly. Its possible geothermal significance, however, must be weighed with respect to both the effect of seawater saturation of rocks at depth and temperature effects as illustrated in Figure 1.

LOOP-LOOP INDUCTION SOUNDINGS

Dipole-dipole results from the Puna Anomaly indicated significant lateral changes in apparent resistivity (see Fig. 3). The centers of lowest resistivity, if thermally generated, imply that there are isolated near-surface thermal vents, conceivably local upwelling of geothermal fluids from a deeper zone of enhanced temperature.

To provide more detailed information regarding the near-surface electrical structure and the extent of possible near-surface thermal waters, we made several shallow soundings using the loop-loop induction method and the Schlumberger DC method. Station locations for these soundings are shown in Figure 10.

Horizontal loop-loop induction soundings discused in this section are similar in principle to the line-loop method discussed earlier (Keller and Frischknecht, 1966). The spacing between source and receiver in two-loop sounding is only a few hundred meters, thus the depth of penetration would be about sea-level for the elevations (below 200 meters height) encountered in the lower Kilauea East Rift Zone.

Many of the soundings proved to be of negligible value for geothermal purposes, i.e., the system response could not be distinguished from the response of the system in "free space". Practically speaking, as the loop-system was calibrated in a resistive area (Pohakuloa saddle area, Zohdy and Jackson, 1969), this means that for the radius of resolution of the system the earth was homogeneously resistive. Other soundings provided data that were uninterpretable in terms of a stratified earth. These stations were generally located in the near-rift zone, which we assume by inference from the dipole-dipole results, to be affected by large lateral variations in resistivity.

Table 3 lists the basic facts resulting from the loop-loop soundings. A final interpretative report of these data is in preparation and only the main geothermal implications will be stressed here.

Using the two-layer model curves of Frischknecht (1967), these data were interpreted as showing a very resistive overburden overlying a conductive substratum. The top of the conducting material was generally close to sea-level and is considered to be a marker between conductive water-saturated basalts below and dry basalts above. Based on well-data control (Davis and Yamanaga, 1968) and the calculated low resistivity below this interface, we conclude that the high conductivity layer is saline water. Apparently, the fresh-water-saturated zone (Stearns and Macdonald, 1946) between the lower saline region and upper dry zone is too thin and/or resistive to be resolved.

The resistivity of the second layer ranges from 2 to 10 ohm-m, which is lower than expected for cool-water-saturated rocks and would be more typical of rocks in brackish to saline water having temperatures of 50 to 100°C. Well-hole temperature profiles (D. Epp and J. Halunen, personal communication) indicate that, below the rift (i.e., seaward), temperatures are indeed encountered in this range. Commonly however this temperature regime is only a few meters thick and then cools again. Apparently the shallow resistivity measurements did not in general penetrate through this thin layer.

The implication of the combined shallow-resistivity and well-hole data is that seaward of the rift the area is underlain by a thin zone of hydrothermal fluids that are heated by rift intrusives as they migrate seaward through a highly permeable rock. These hydrothermal waters are neither hot enough nor voluminous enough to provide a geothermal power source, leaving as a candidate for potential power only the relatively massive low-resistivity region below a few hundred meters depth.

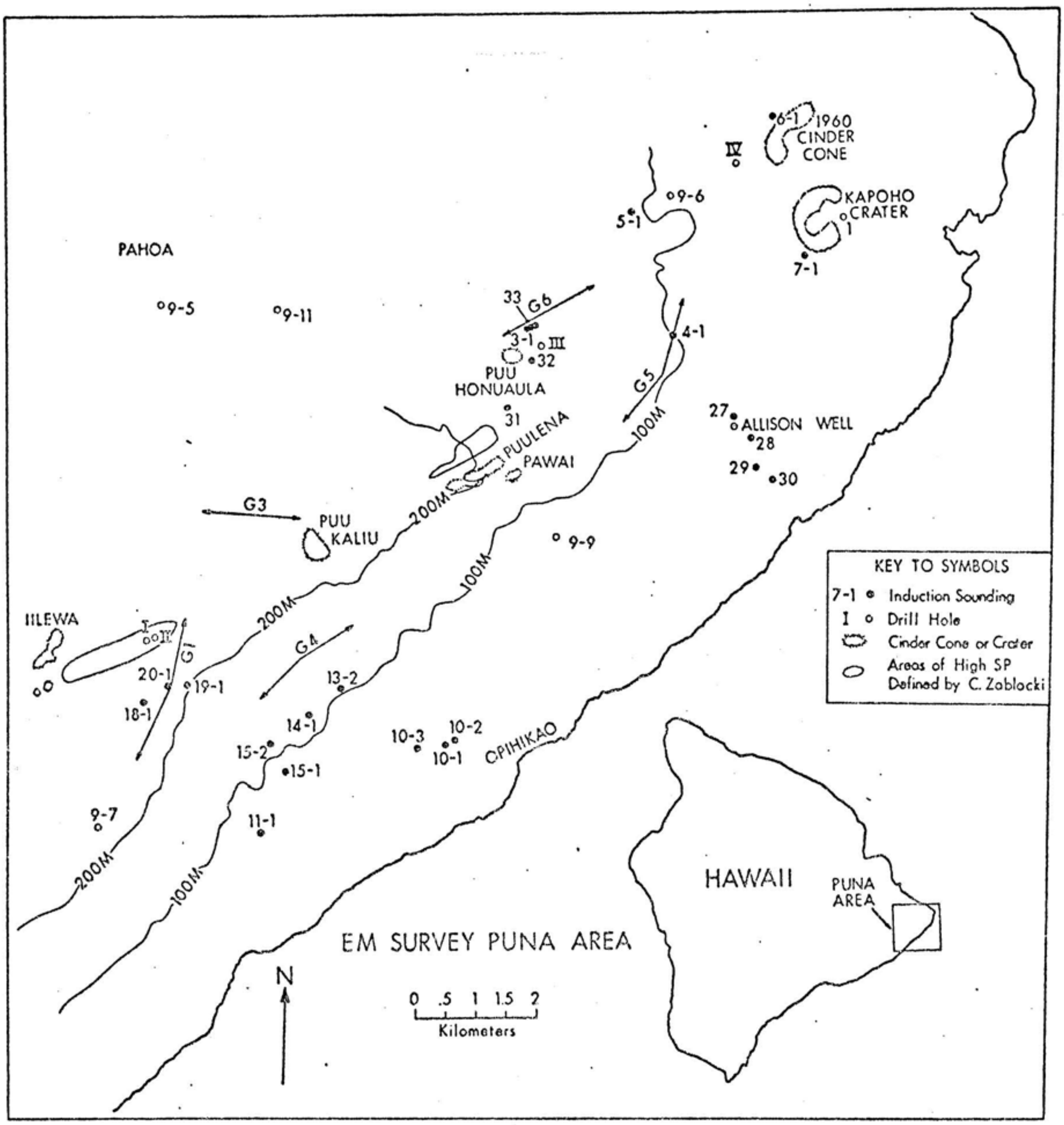

Fig. 10. Two-loop induction and Schlumberger galvanic stations in the lower east rift of Kilauea. The galvanic soundings are designated by a line and G. (From manuscript in preparation, D. Klein and J. Kauahikaua.)

Table 3. Results of Loop-Loop Inductive Soundings in Puna

Station	Elevation (meters)	Source Receiver Separation (meters)	Elevation Layer 2 (meters)	Resistivity Layer 2 (ohm-m)	Comment
3-1	182	305	-	-	(free space)
4-1	98	488	-24	5.6	-
5-1	103	424	-	-	(free space)
6-1	47	341	-36	5.9	-
7-1	34	524	-62	(6.7-24.0)	-
10-1	46	356	+2	1.6	-
10-2	49	634	+9	4.9	-
10-3	56	521	+4	2.6	-
11-1	75	335	(-9 - 33)	(2 - .4)	-
13-2	101	491	(22?)	(9.1?)	poorly determined
14-1	110	450	-	-	indeterminate
15-1	82	533	(-8?)	(2.6?)	poorly determined
15-2	113	462	-31	6.3	-
18-1	262	671	-	-	(free space)
19-1	197	594	-	-	(free space)
20-1	244	671	-	-	(free space)
27	38	366	(0)	1.8-2.2	-
28	35	366	(0)	1.2-1.5	-
29	28	366	0	2.7-3.3	-
30	18	366	0	2.2-2.7	-
31	181	366	-	-	(free space)
32	201	366	-	-	(free space)
33	177	400	-	-	(free space)

SCHLUMBERGER SOUNDINGS

The previously discussed soundings have been relatively insensitive to the electrical structure above sea level; thus to obtain more precise data on this resistive material, several Schlumberger soundings were made. This method is fundamentally a "dipole" DC technique as described earlier. The electrode configuration in a Schlumberger sounding, however, is essentially a single placement of the passive voltage-sensing electrodes while the active current electrodes are spread out in-line symmetrically from the sensing electrodes. The method is widely applied for determining resistivity variations with depth, (e.g., see Meidav, 1970) and the interpretation theory is well established (Orellana and Mooney, 1966; Keller and Frischknecht, 1966; Rijkwaterstaat, 1969).

Table 4 summarizes the results of the Schlumberger soundings (see Fig. 10 for station locations). The resistivity of the bulk of rocks above sea level is about 6000 ohm-m. The implication of this high value is that the rocks contain negligible free water above sea level. This result combined with the loop-loop induction results generally indicates that all water saturation is at or below sea level with no anomalously high water levels (thermal or non-thermal) in the region seaward of the lower Kilauea east rift.

One exception is sounding G3, where lower resistivities are encountered at about 80 feet above sea level. This (above rift) sounding is just north of one of the local anomalous areas of the dipole data (see Fig. 3). This result may reflect dike-impounded high water, as this sounding was just uphill from the rift-zone surface expression and may have been over a portion of a dike swarm associated with the rift (Stearns and Macdonald, 1946). However, this anomaly correlates with a self-potential anomaly discussed later and there is good evidence that it is associated with a local heating effect. The rift zone in this general locality extruded a voluminous mass of lava in 1955; steam seeps and warm ground are still in evidence. Keller's group (1973) made a detailed profile across the rift in this area and defined the width of the anomalous zone of low resistivity as possibly 1 to 2 kilometers.

This anomaly was drilled in the early 1960's with the hope of finding steam (e.g., see Stearns, 1966; Koenig, 1970). While temperatures close to the boiling point were encountered, the holes did not produce steam and were abandoned. The depth of the holes did not reach sea level, and thus there is no direct evidence of hot water potential.

Table 4. Results of Schlumberger Soundings. Resistivities, ρ(ohm-m) and Thicknesses h(meters)

Station	Elevation (meters)	ρ/h Layer 1	ρ/h Layer 2	ρ/h Layer 3
G1	244	20,000/8	6,660/238	<100/∞
G3	275	2,900/5	5,800/195	<150/∞
G4	131	31,000/3	6,200/131	<100/∞
G5	105	-	6,300/104	<10/∞
G6	165	1,300/2	7,800/171	<100/∞

SELF-POTENTIAL MAPPING

Positive natural voltage potentials in the earth have been correlated with thermal zones at Kilauea Volcano by field surveys performed by C. Zablocki (personal communication). He also discovered two anomalies of large voltage (100's of millivolts, mV) in the East Rift Zone roughly near the resistivity anomalies delineated by Keller (1973). Although the exact explanation for these large anomalies is uncertain, the correlations observed on Kilauea Volcano strongly support the supposition that they are related to high-temperature conditions.

Cooperative arrangements were made between the U.S. Geological Survey and the Hawaii Institute of Geophysics to expand Zablocki's survey of the lower east rift. The basic data analysis and interpretation is being handled by Zablocki, and the detailed results will be described in a separate article. Briefly, however, no significant potential anomlies were found that were not already outlined. These anomalies are now well mapped, however, and they seem to closely follow the surface expression of the 1955 eruption in the lower east rift. Potential theory sets the depth of generation of these anomalies at about sea level, although it is possible that the level of the ultimate source of sea level temperature disturbances is deeper.

SUMMARY

Geoelectric reconnaissance surveys on Hawaii Island have located one area of generally low resistivity that can be considered to have possible potential for geothermal development, the Puna Anomaly on the lower east rift of Kilauea Volcano.

Dipole-dipole results provide the basic geoelectric picture of Puna. In addition to showing that the lower east rift of Kilauea has generally low resistivity, the data indicate several local spots of exceptionally low resistivity (less than 5 ohm-m). A generalized resistivity-depth profile was obtained of 20 ohm-m to 600 meters depth, 5 ohm-m from 600 to 2000 meters depth, and high resistivity below 2 kilometers. Deep inductive surveys to about 1.5 kilometers depth in general verified this result. Less deeply penetrating surveys established that anomalous resistivities at sea level are associated spatially with the rift zone and particularly with the 1955 eruptive loci. A generally shallow and low-resistivity layer believed to be caused by rift-intrusive warmed water exists seaward of the rift. Temperatures in this zone can range from about 30° to 90°C. Although there is no geoelectric evidence of the thickness of this zone, well data indicate it to be only a few meters thick, and we believe it is of negligible potential for commercial power utilization. The conductive layer at greater depth (below 500 meters) is the most promising indication of a large mass of hot water.

A preliminary geoelectric model for the region seaward of the rift is summarized in Table 5 and Figure 11. In Figure 11, the upper control point, above sea level, and the sea level control point are based on loop-loop induction data, although Schlumberger soundings indicate the true resistivity above sea level is more likely to be about 6000 ohm-m. The deep control points at roughly 850 meters depth are based on line-loop induction data. The limiting horizontal and vertical bars refer respectively to the spread of calculated resistivities and estimated depth of penetration. The data points in the deep zone are apparent resistivities plotted versus half-spread distance of the line-loop system. The data indicated by squares are those found generally in the more southwestern boundary areas of the Puna Anomaly. It seems that the deep region of lowest resistivity is to the northeast in the anomalous area; however, the significance of the resistivity differences is uncertain at this time. The anomalous resistivity

Table 5. Preliminary Depth-Resistivity Model Seaward of Rift in Puna Area

Zone	Depth (meters)	Resistivity (ohm-m)	Comments
I	Above sea level	6000 (based on Schlumberger soundings)	Dry porous and permeable aa lava flows.
II	Sea level	2-10	Porous and permeable lava saturated with brackish water; measured temperature is 30° to over 90°C. The high temperature is only in a thin layer based on well measurements.
III	Sea level to -500	35 (estimated value just below thin, low resistivity layer, poor control in this region; Keller estimates a mean resistivity in this region of 20 ohm-m)	Porous and permeable lava saturated with brackish water, temperature about 50°C based on well measurements; 35 ohm-m is the expected resistivity for 10% porosity, 10,000 ppm salinity, and 50°C.
IV	-500 to -2000	5 (inductive soundings place a 5-50 ohm-m range on this value; Keller's results indicate the resistivity is on the low side)	Saturated porous lavas extruded extruded below sea level. For an assumed 10% porosity and sea water salinity (35,000 ppm) the estimated temperature is 180°C.
V	Below -2000	High resistivity	Dehydrated and/or dense basalt; fracture and pore porosity partially closed.
VI	Rift Zone (near surface)	Locally anomalous electrical characteristics, in narrow zones and locally high self-potential gradients. The depth expression of these anomalies is not yet defined.	Fractured and warm rocks. Possible upwelling of geothermal fluids in local areas. Probably hot intrusives at depth.

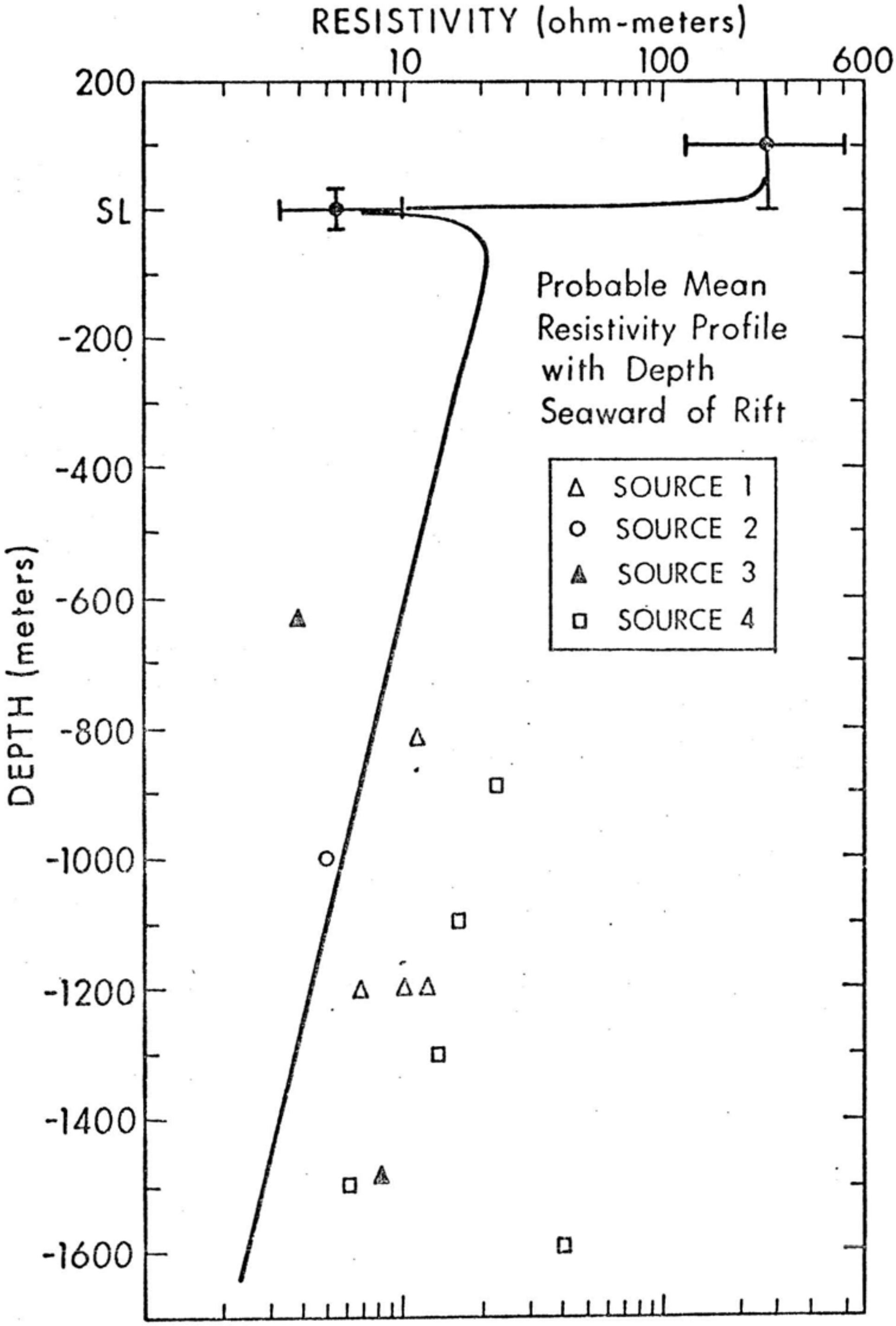

Fig. 11. Preliminary depth-resistivity profile seaward of the Kilauea East Rift in Puna. The basic control points are based on induction data (see text). The horizontal bars refer to the spread of data and have no statistical significance. The deep control points (from line-loop induction data with sources numbered corresponding to Fig. 9) are "apparent resistivity" plotted against "1/2" source-receiver separation. The solid line is plotted to take into account the dipole data results. The resistivity inversion at sea-level (SL) is inferred from temperature probes.

inversion at sea level below the rift reflects the thin layer of warm water that is generally observed by well data and thermal probes rather than electrical data.

The "probable" mean resistivity curve is drawn toward the lower limit of the induction data to take into account the dipole-dipole results, which indicated a slightly lower resistivity than did the inductive data.

Although the above results are still tentative, the deeper resistivity data deserve comment in terms of probable temperature and geothermal potential. As indicated in Figure 1, it is theoretically possible to estimate the temperature of saturated rock strata if the bulk resistivity is known. However, such an estimate depends on the rock porosity and the equivalent salinity of the pore liquid. In application, these parameters must be established independently, and in the absence of drill-hole data, estimates of the values for these parameters are subject to error; thus are the temperature estimates also.

The empirical relationship that has found application in relating bulk rock resistivity, ρ_b; pore liquid resistivity, ρ_w. and fractional porosity, ϕ is (Keller and Frischknecht, 1966; Keller, 1970):

$$\rho_b = \rho_w \cdot A\phi^{-m}$$

where A and m are experimentally determined constants. Brace and Orange, (1968) demonstrated that for a wide variety of rocks under a pressure load sufficient to close fracture porosity (a few kilobars) the constants A and m are 1.0 and 2.0, respectively. These are the assumed constants that were used to construct Figure 1, and that will be used here to derive the in situ pore-fluid resistivity.

Taking a bulk resistivity of 5 ohm-m and a porosity of 10% for Hawaiian rocks at 500 to 1500 meters depth, the fluid resistivity in situ is .05 ohm-m, according to the above relationship. For temperatures of less than about 300°C, the relationship between temperature and fluid resistivity is approximately (see Meidav, 1970; Brace, 1971) given by:

$$\rho_w(T) = \frac{\rho_w(18°C)}{1+ .025(T-18°C)} .$$

Assuming seawater saturation (ρ_w = .25 at 18°C), the above equation solved for T gives an expected temperature of about 180°C at depths of 500 meters or more.

If porosities are closer to 25% as found in the deep drill hole on Kilauea (Keller, 1974; Zablocki et al., 1974), the temperature would be considerably lower. There is, however, no real basis for direct comparison between the rocks of Puna at 800 meters depth and those of the Kilauea Summit. The Puna rocks at this depth were probably extruded beneath the sea, which would tend to make them less porous (Moore, 1965).

We have chosen not to apply Keller's estimates of the constants A and m (A = 3.5, m= 1.8, Keller, 1973). These values were experimentally determined by Keller using 75 surface basalt samples from Hawaii. Such constants would result in a ρ_w in situ of 0.016 ohm-m and a temperature considerably in excess of 180°C. For the depth being considered (0.5 to 1.5 kilometers) the rocks may bear little resemblance to surface samples. They were probably extruded beneath sea-water and are also subject to moderate overburden pressure that we believe will alter the fracture porosity to result in a relationship between ρ_w, ρ_b, ϕ more toward that indicated by Brace and Orange (1968).

The fluid salinity is also unknown; Puna well data all show fluids with salinities less than half that of sea-water. If salinities of less than that of sea-water extend to depth, the temperature estimate could be considerably increased. However, sea-water salinity is considered a good approximation since well-water samples on Oahu from about 400 meters depth indicate roughly equivalent ionic concentration as sea water (Hufen, 1974).

Only drilling data can resolve the above ambiguities. From the standpoint of our present position, an average temperature profile in Puna is hypothesized in Figure 12. Also shown on the figure is the boiling point curve. Although the mean temperature curve is not close to the boiling point curve, there is the optimistic possibility that in the anomalous regions the temperature gradient would be considerably higher.

Future plans include several intermediate depth soundings over the specific areas of the anomalous geoelectric data to attempt to provide more quantitative estimates of the above possibility and to better define the region above 500 meters depth.

Based on these preliminary results, we conclude that the Puna area near and seaward of the rift zone is geothermally anomalous. However, in our opinion, the probability of encountering a large mass of water with temperatures above 200°C does not seem particularly high. The most likely candidate for hot water is the region from 500 to 1500 meters depth below the 1955 eruptive loci.

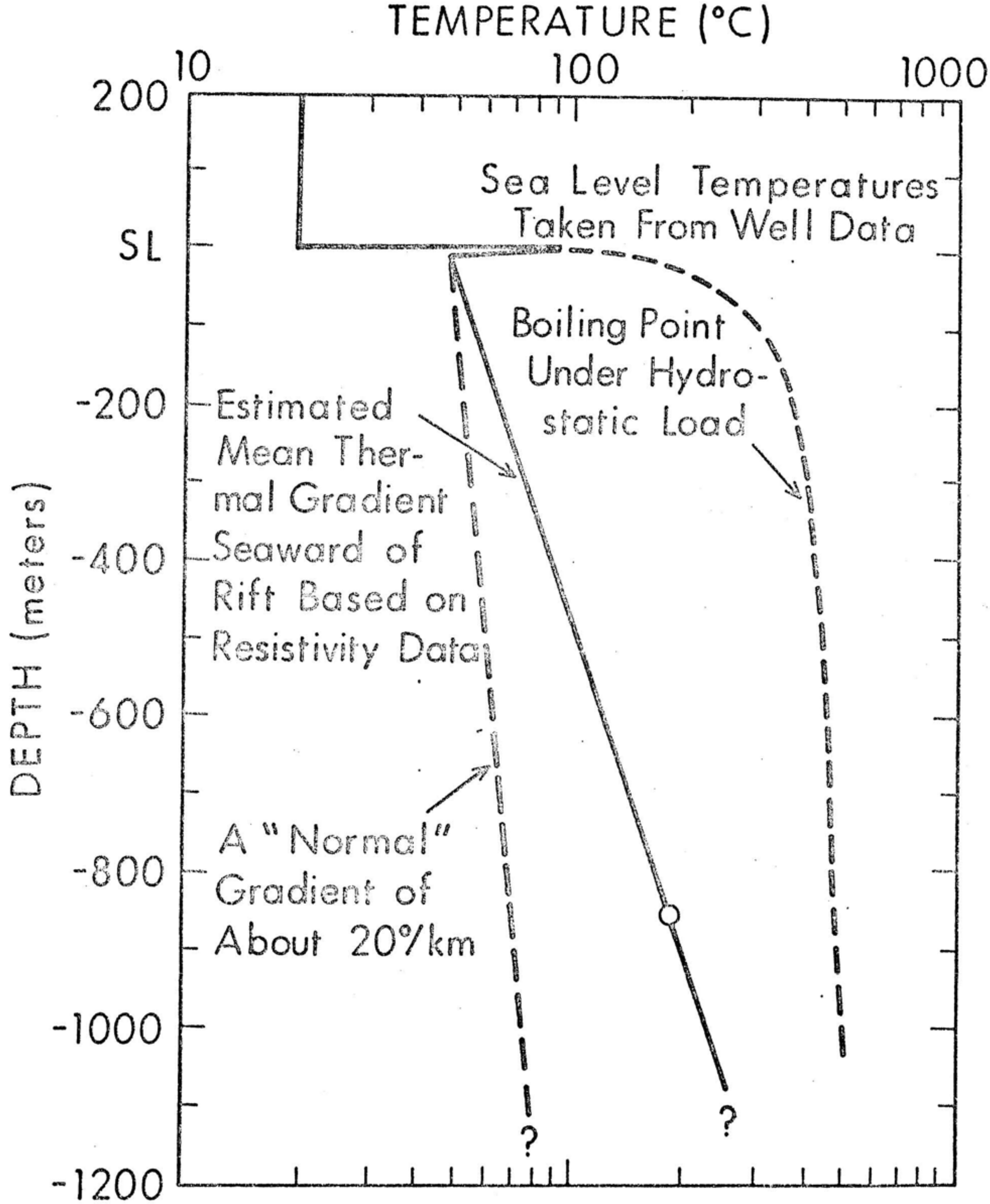

Fig. 12. Preliminary depth-temperature profile seaward of the Kilauea East Rift in Puna. The circle at 850 meters depth is the calculated temperature for a 5 ohm-m resistivity (see text). The temperature "gradients" indicated are to be considered in a qualitative sense only.

APPENDIX: FIELD INTERPRETATION OF TIME-DOMAIN INDUCTION DATA

Line-loop induction soundings were made in the "time-domain", i.e., wide-frequency-band transient signals were generated rather than discrete frequency signals. Formal analysis of such transients is fairly involved (Silva, 1969; Jackson and Keller, 1972); however, a rapid estimate of the mean earth-conductivity is possible using the method described here. To our knowledge such determinations have not been previously described in the geophysical literature. C. Zablocki suggested this approach to us.

Our physical system consisted of a long, grounded wire excited by an 8-second half-period square-wave voltage with an amplitude of about 800 volts. Thus the inductive excitation of the earth was a magnetic spike variation produced at each change in the source current. The time-variation of this signal, modified by induction in the earth, was recorded with an oscillograph microvolt recorder connected in series with an induction coil placed horizontally on the ground.

The theoretical inductive response of such a system in the case of a homogeneous conducting plane earth is given by Wait (1951) as:

$$V(t) = \frac{L_1 A_2 I}{2\pi\sigma r^4} \; Q(t) \sin\theta \quad \text{(mks)} \qquad (1)$$

where $V(t)$ is the induced voltage in a receiving loop of effective area A_2; I is the amplitude of the current step in the source line of length L_1; the earth has conductivity, σ; the separation between source and receiver elements is r; and θ is the angle between the source line and the connecting line r between source and receiver. The response function $Q(t)$ is given by:

$$Q(t) = 3\,\mathrm{erf}\left(\frac{\alpha}{2\sqrt{t}}\right) - \left[\frac{3\alpha}{2\sqrt{t}} + 2\left(\frac{\alpha}{2\sqrt{t}}\right)^3\right] \mathrm{erf}'\left(\frac{\alpha}{2\sqrt{t}}\right) \qquad (2)$$

where $\alpha = r\sqrt{\sigma\mu_o}$, μ_o is the magnetic permeability ($4\pi \times 10^{-7}$).

According to eq. (1) the time-dependent decay, $Q(t)$, of the inductive response is determined by σ and r, independent of other system parameters. This is illustrated in Figure 13 where $Q(t)$ is plotted against $t/\mu_o r^2$ for various values of σ. Thus the decay time and r can determine σ.

We constructed a graph of the "half-decay time", $t_{\frac{1}{2}}$, (defined as the time for the recorded signal to decay to one-half of its maximum amplitude), for various σ and r. This we used to rapidly estimate the mean conductivity ("apparent conductivity") of the earth for each sounding. Such apparent conductivities can be interpreted as the conductivity of an infinite half-space with the same $t_{\frac{1}{2}}$ response as the earth, or in a more physical sense, as the mean earth conductivity within the inductive radius of the system assumed to be r/2.

One complication is that the receiver system response has an inherent decay-time independent of the earth. For instance, when using a low-pass filter to attenuate commercial or natural magnetic noise fields with frequencies of about 40 Hz or more, the filter introduces its own intrinsic decay constant into the recorded transient. For our system the intrinsic $t_{\frac{1}{2}}$ constant was about 50 milliseconds and the effect is to lengthen the apparent $t_{\frac{1}{2}}$ decay of the earth response. Equivalently, this makes the apparent conductivities higher than they should be (if recorded with a system of perfect response). Thus our field-determined apparent conductivities should be considered as maximum values (or minimum apparent resistivities). This effect limited the highest resistivities determineable in the field to about 40 to 50 ohm-m. For highly conducting areas, however, the intrinsic system response is probably a negligible factor.

"Formal" reduction in any case removes the known instrumental response by deconvolution to obtain a more precise estimate of the true earth response. Compare, for instance, the "field" and "formal" apparent resistivities of Table 2. Figure 14 illustrates an example of a digitized field signal and the estimated earth response as determined by deconvolution.

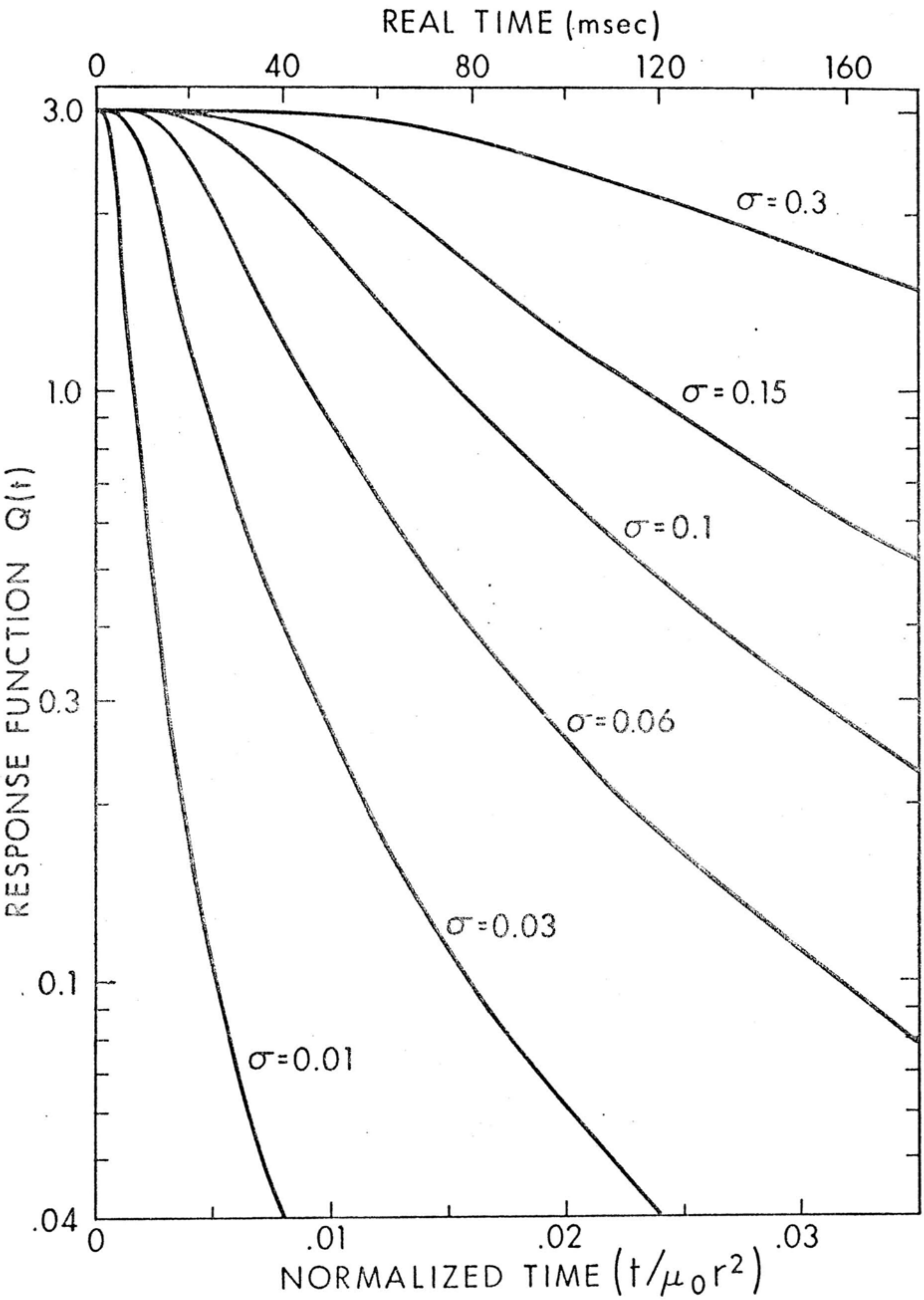

Fig. 13. Time-decay response of a line-loop inductive system. Q(t) is the receiver decay function for a sudden change in source current over an infinite half-space of uniform conductivity, σ; μ_o is the magnetic permeability in free space (mks units); t is time in seconds; and r is source-receiver separation in meters. The top scale gives real-time (msec) for 5 = 2 kilometers.

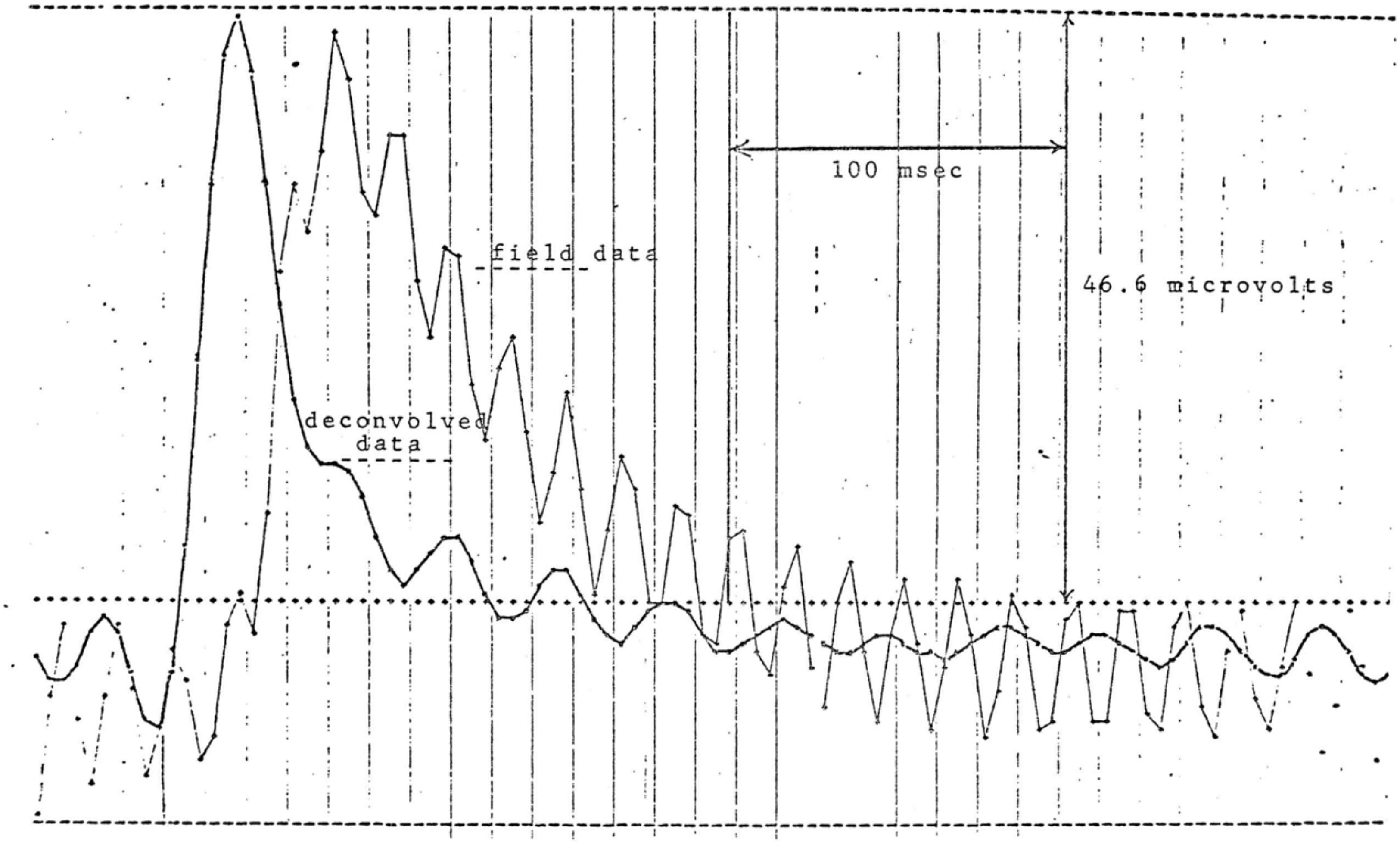

Fig. 14. Time-domain field data example. Digitized field data (light curve) and deconvolved data (heavy curve) for voltage transient observed at station 29 (Puna).

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