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Geoelectric-Geothermal Exploration on Hawaii Island: Preliminary Results

by

Douglas P. Klein and James P. Kauahikaua

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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.

3,500 or 35,000 next pg

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.

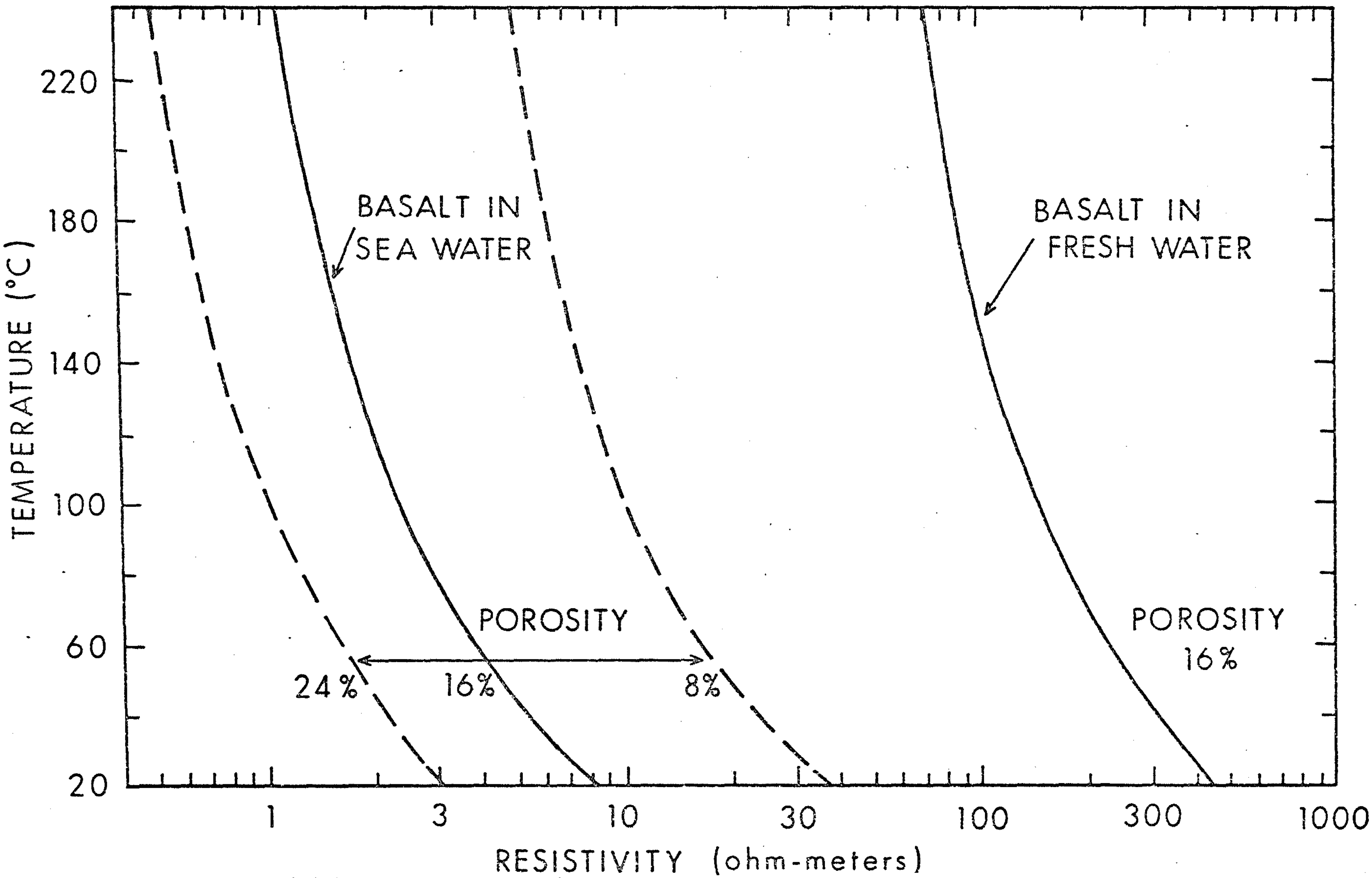


Fig. 1. Resistivity relationships. Solid lines show resistivity versus temperature for water-saturated basalt with 16% porosity. Dashed lines show the resistivity variation expected for porosity changes of ± 50% from a reference porosity of 16%. Fresh and sea water are taken as 50 ppm and 35,000 ppm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

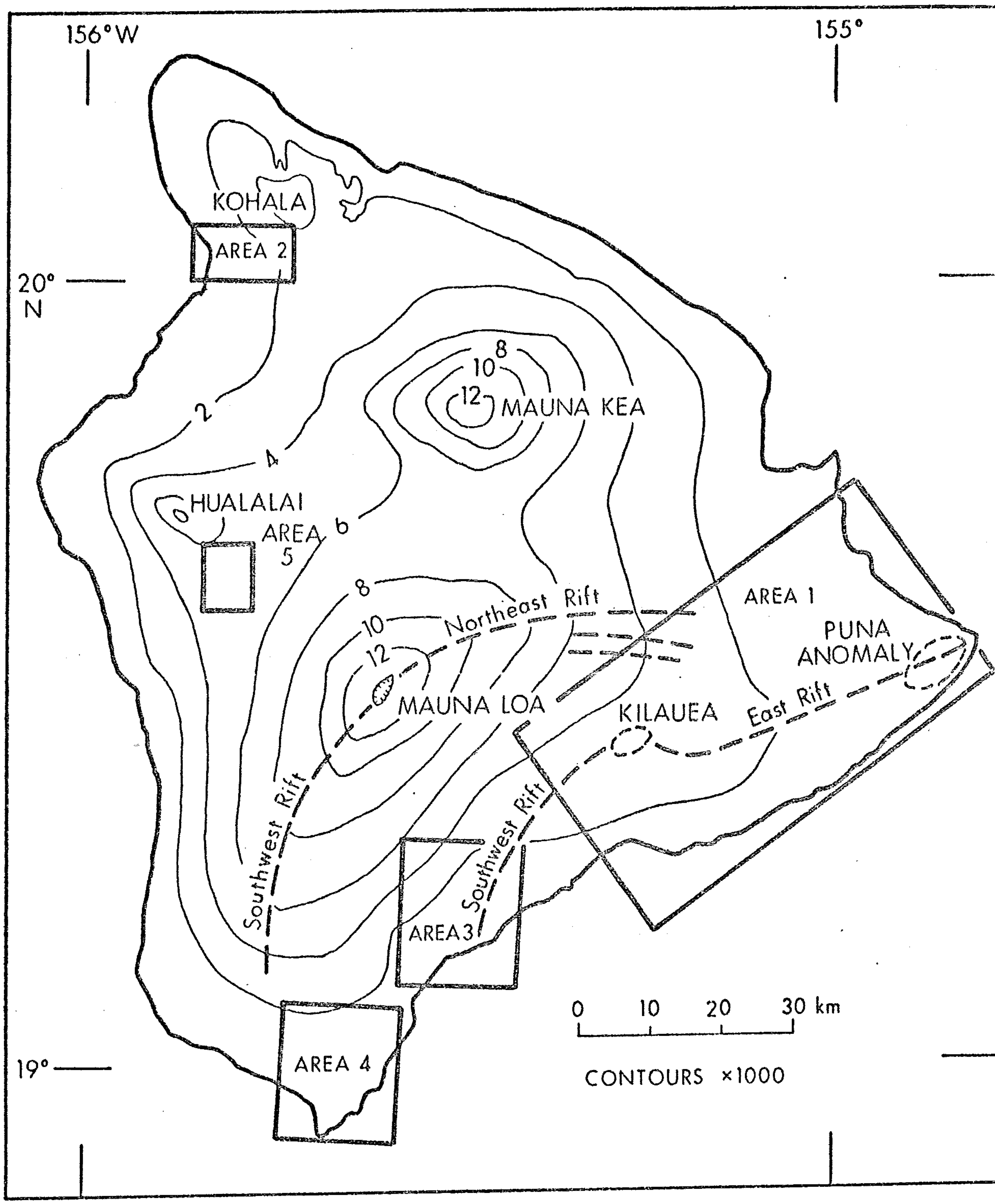


Fig. 2. Areas of resistivity reconnaissance surveys on Hawaii Island. Elevation contours are feet x 1000.

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

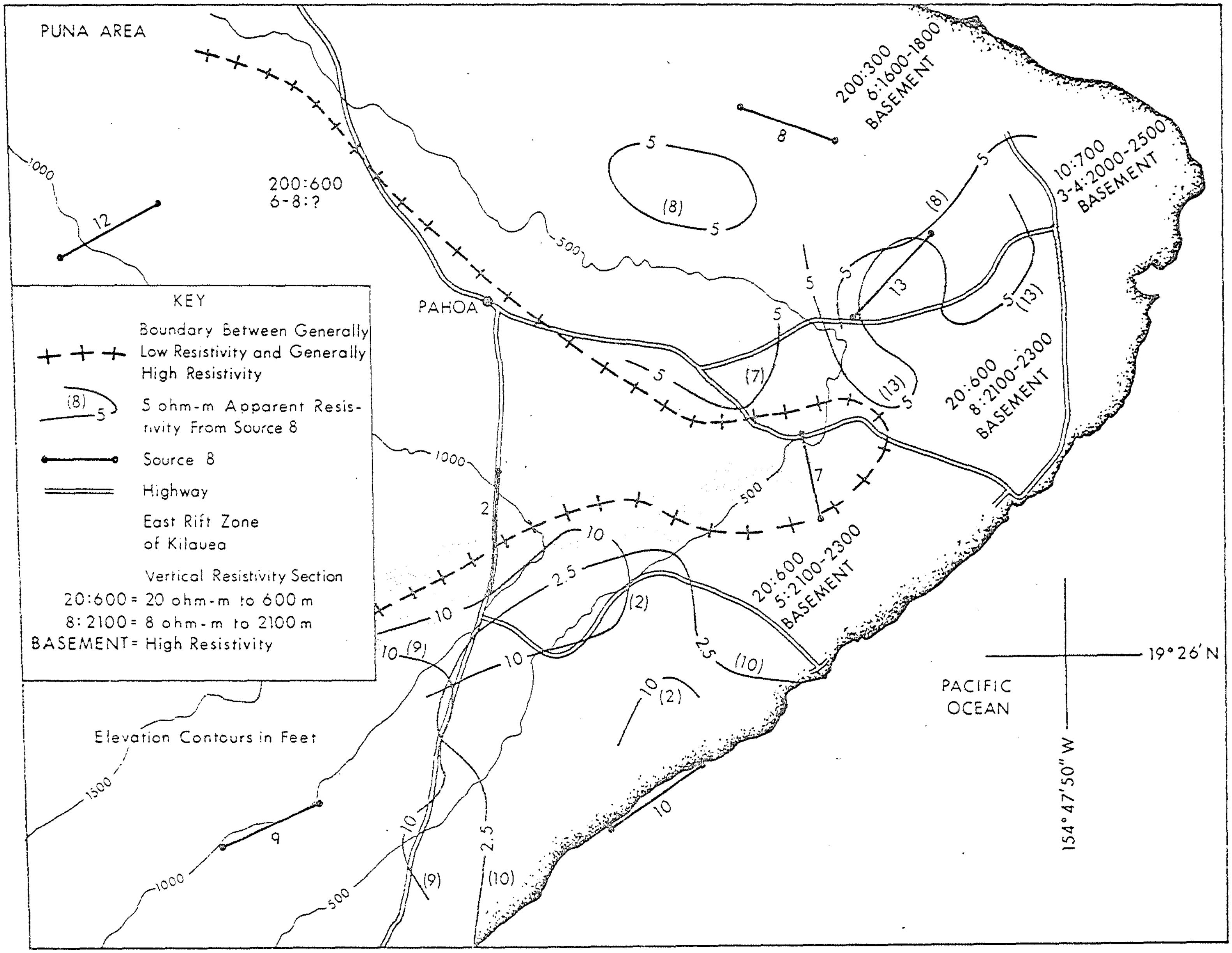


Fig. 3. Results of dipole-dipole mapping in the lower east rift of Kilauea, the Puna Anomaly (from G. V. Keller, 1973). See Figure 2 for general index of location.

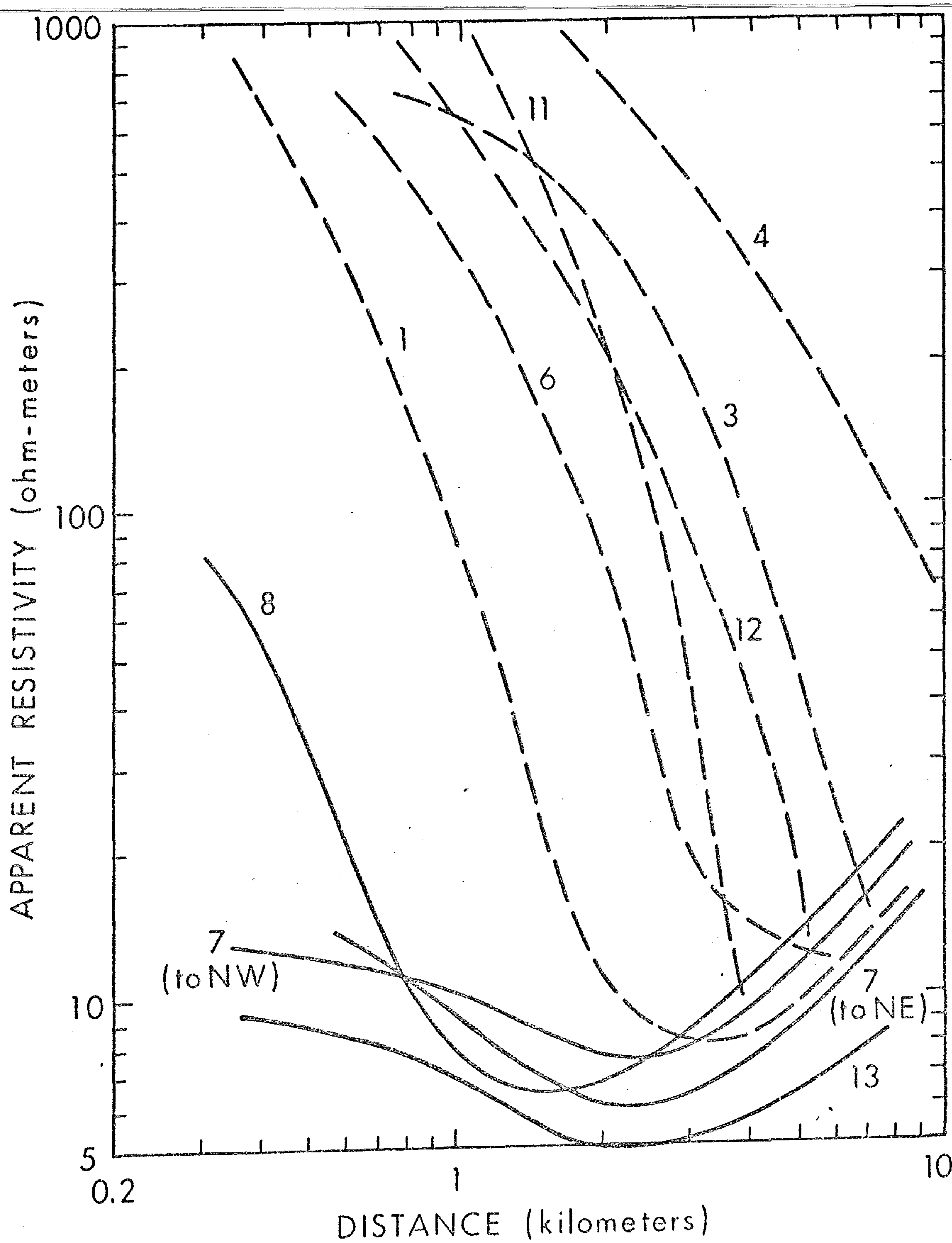


Fig. 4. Generalized trends of apparent resistivity versus source-receiver separation of the dipole-dipole resistivity survey. The solid lines are the trends from stations in the Puna Anomaly as numbered in Figure 3. The data from stations of sources 2 and 10 (omitted) showed such great scatter that distance-resistivity 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:

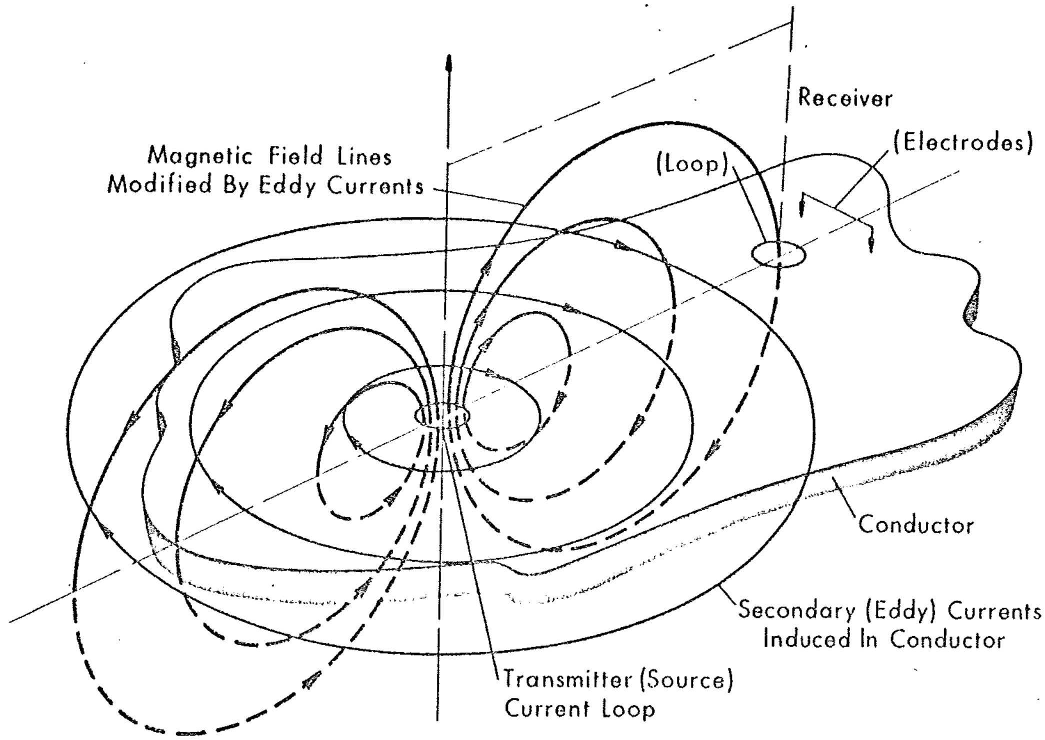
(1) Reconnaissance surveys of areas not covered by Keller's group, and;

(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.

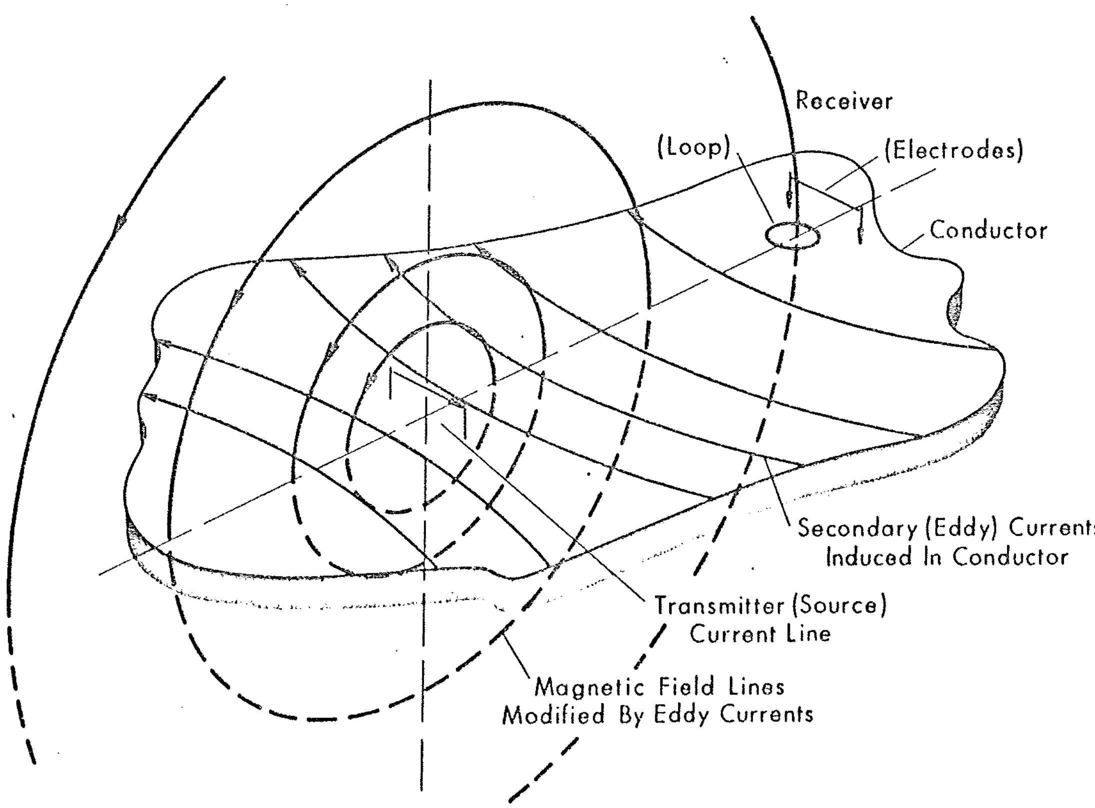
Reconnaissance induction soundings were made in the lower regions of the southeast rifts of both Kilauea and Mauna Loa volcanoes (areas 3 and 4 respectively, Fig. 2). Soundings were also made in the saddle area between the Hualalai and Mauna Loa domes (area 5, Fig. 2).

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



CASE I: LOOP SOURCE WITH LOOP AND/OR ELECTRODES RECEIVER



CASE II: LINE SOURCE WITH LOOP AND/OR ELECTRODES RECEIVER

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.

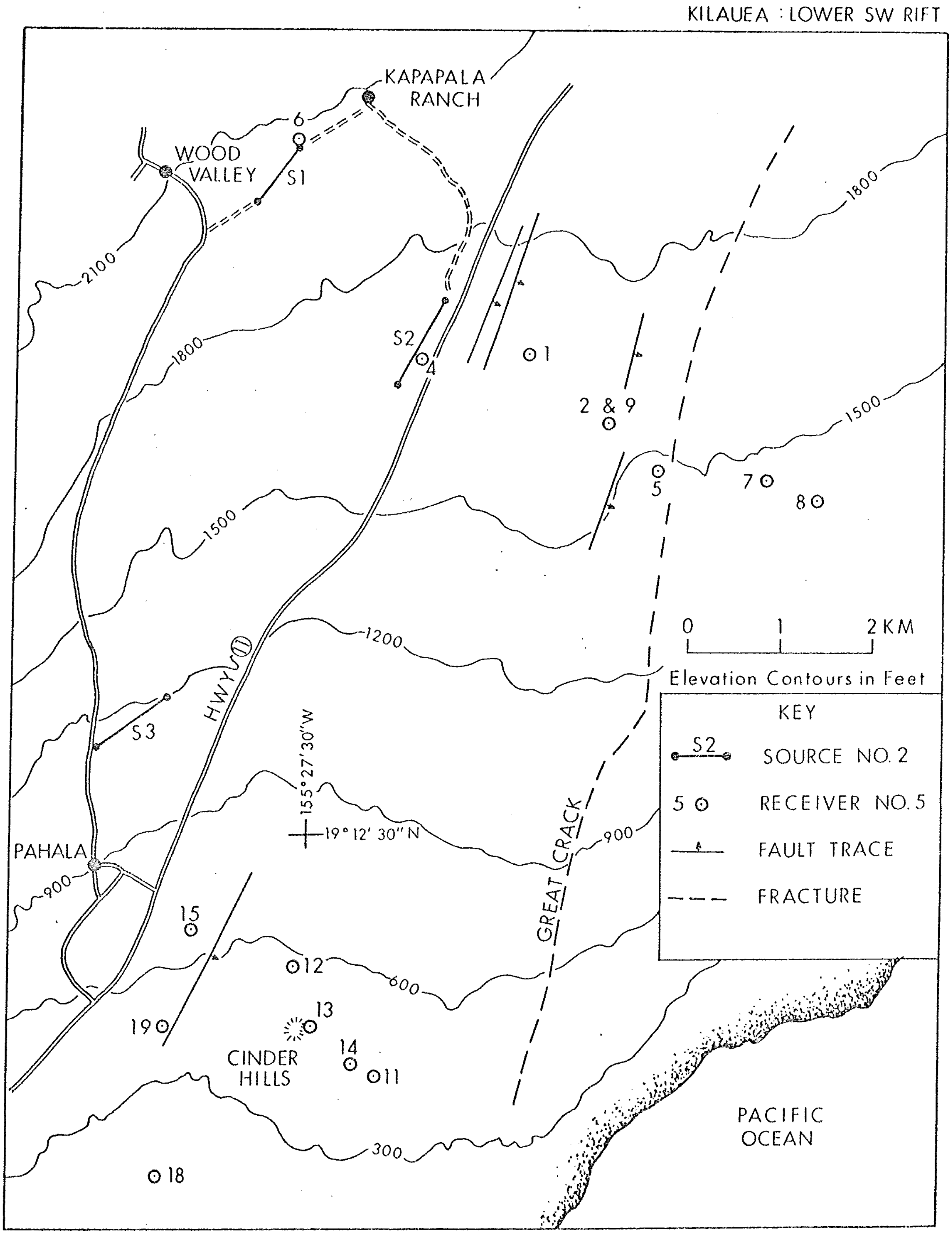


Fig. 6. Line-loop induction stations in the area of the southwest rift of Kilauea (area 3).

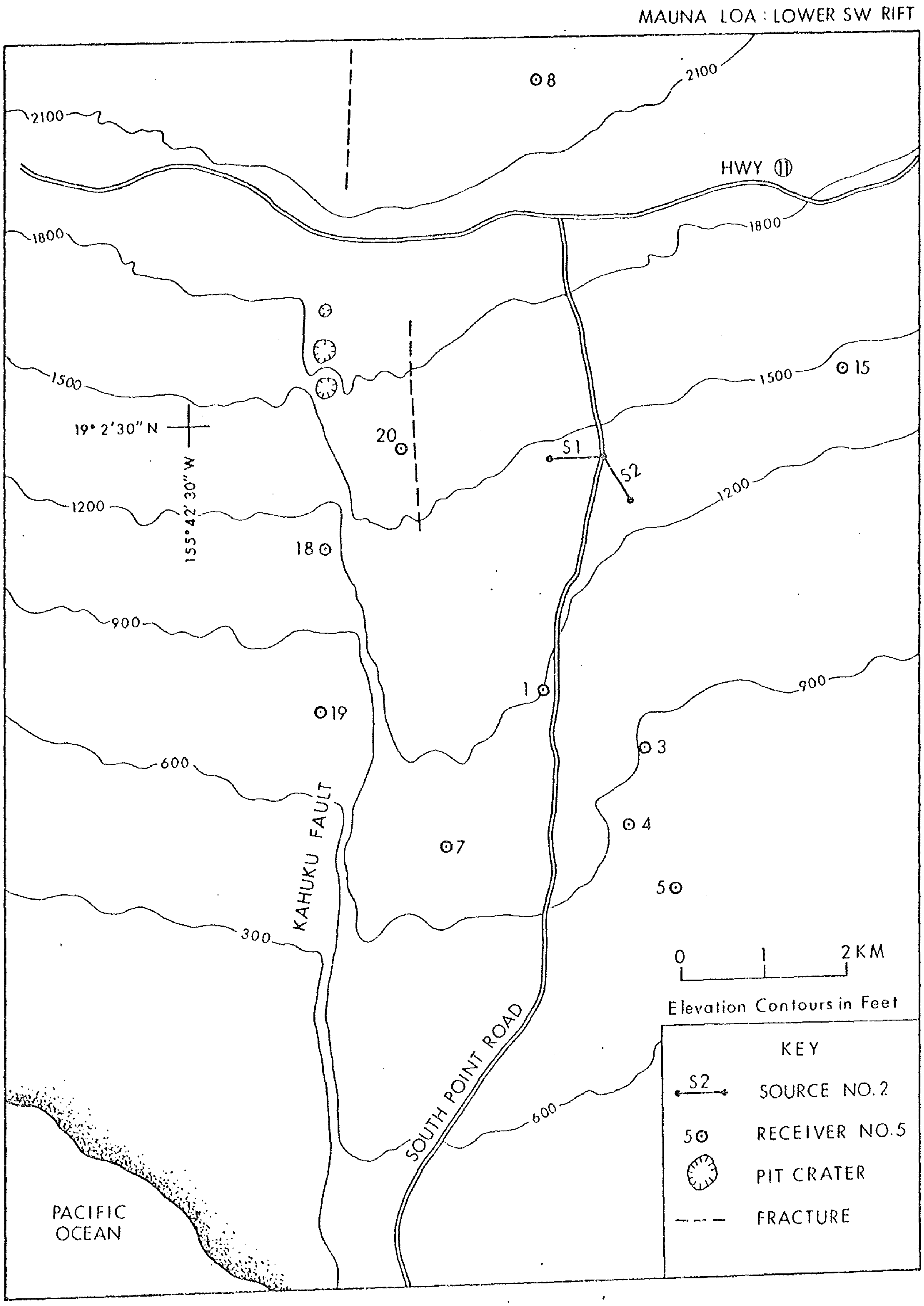


Fig. 7. Line-loop induction stations in the area of the southwest rift of Mauna Loa (area 4).

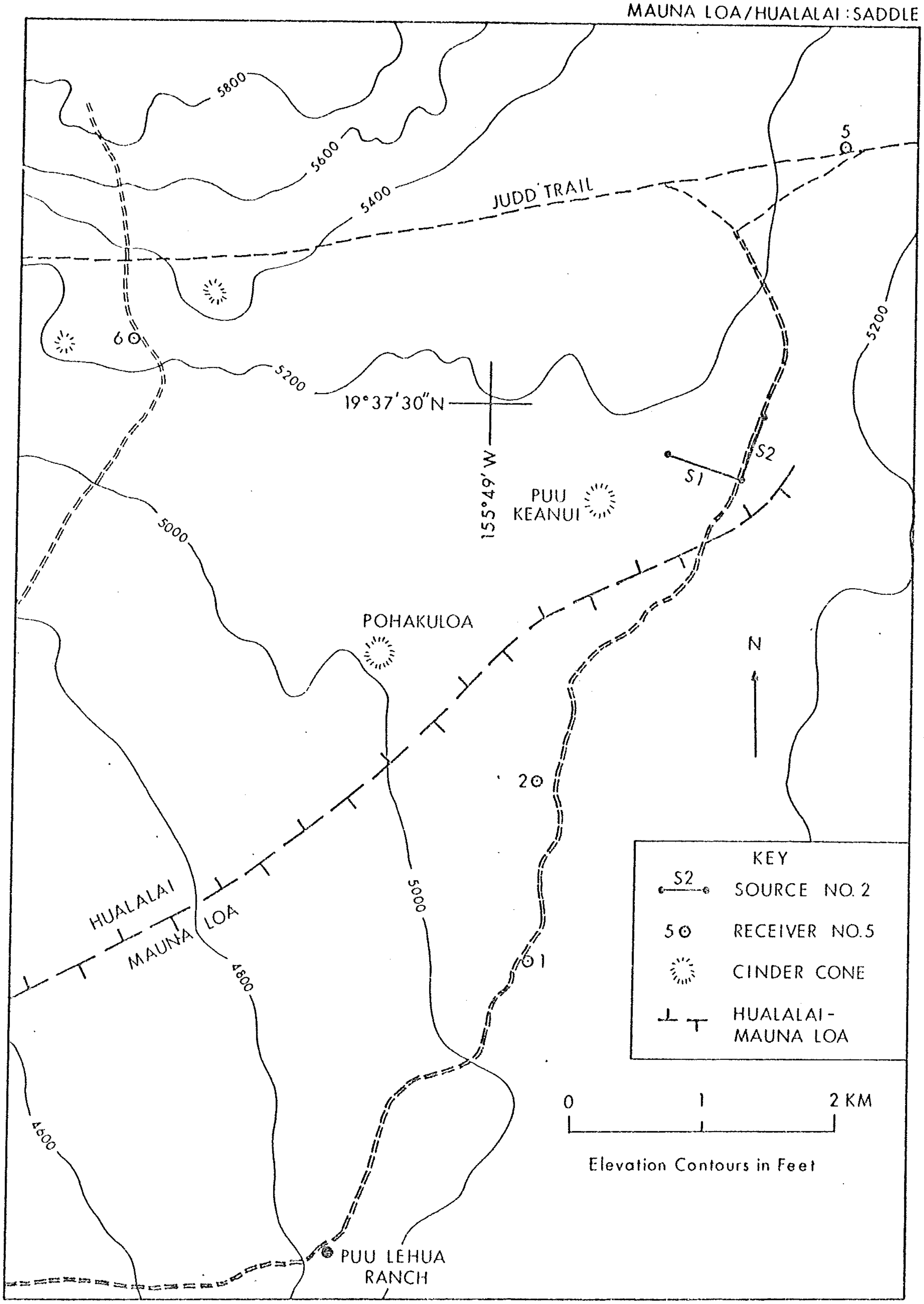
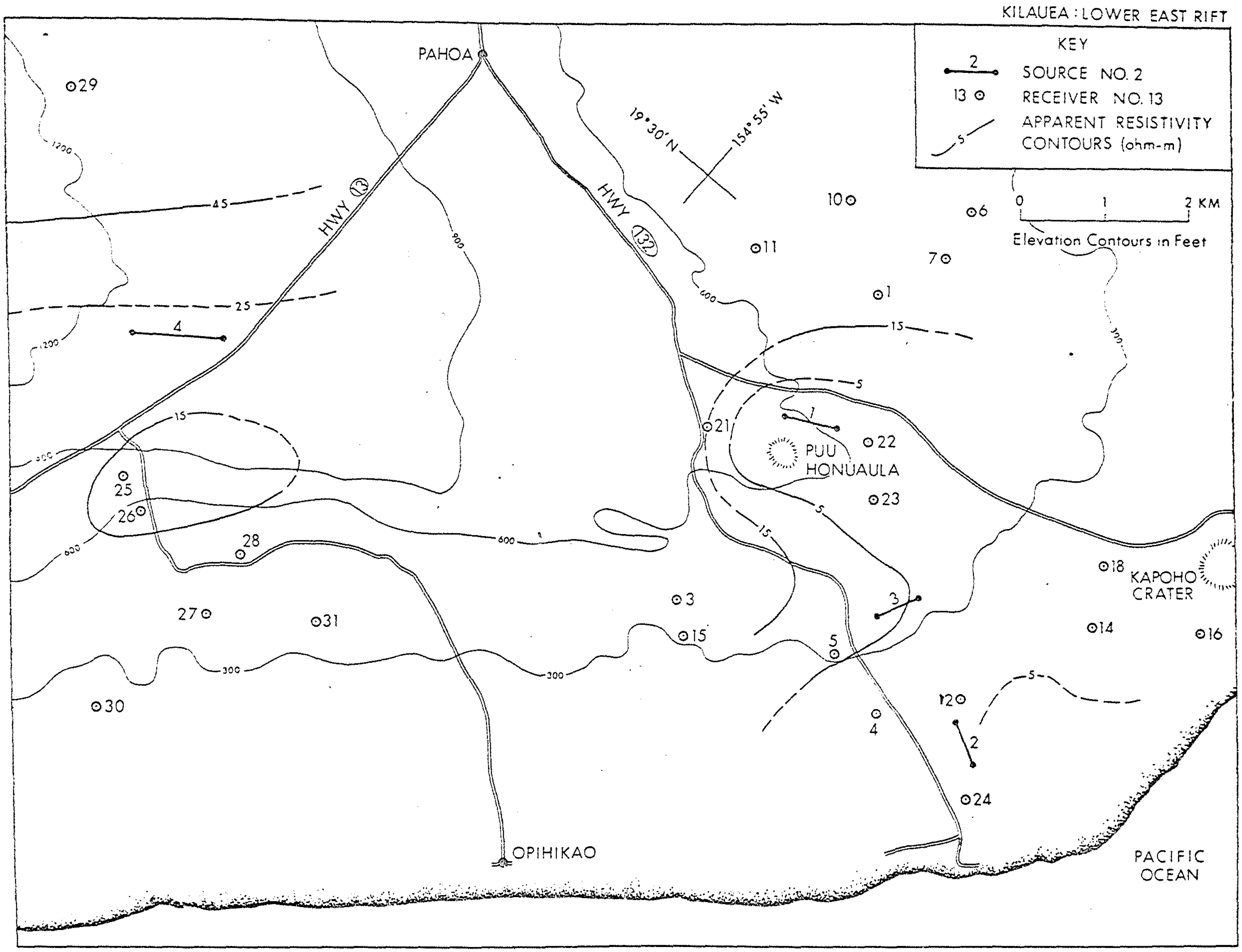


Fig. 8. Line-loop induction station in the saddle area between Hualalai and Mauna Loa (area 5).



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4 Tx
31 Rc

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

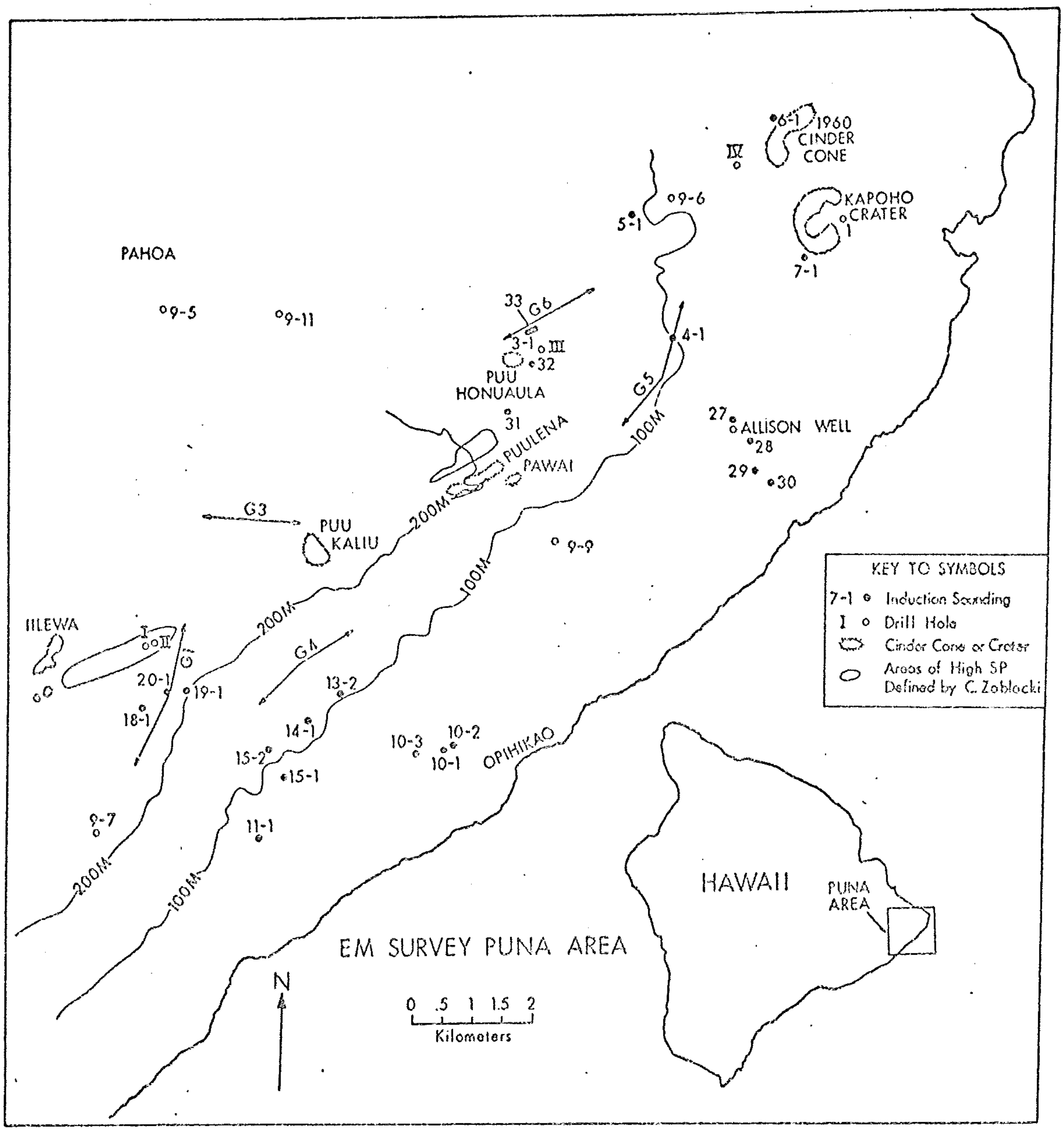

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Fig. 10. Two-loop induction and Schlumberger galvanic stations in the lower east rift of Kilauea. The galvanic soundings are designated by a line and G. (From manuscript in preparation, D. Klein and J. Kauahikaua.)

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)

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/∞

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

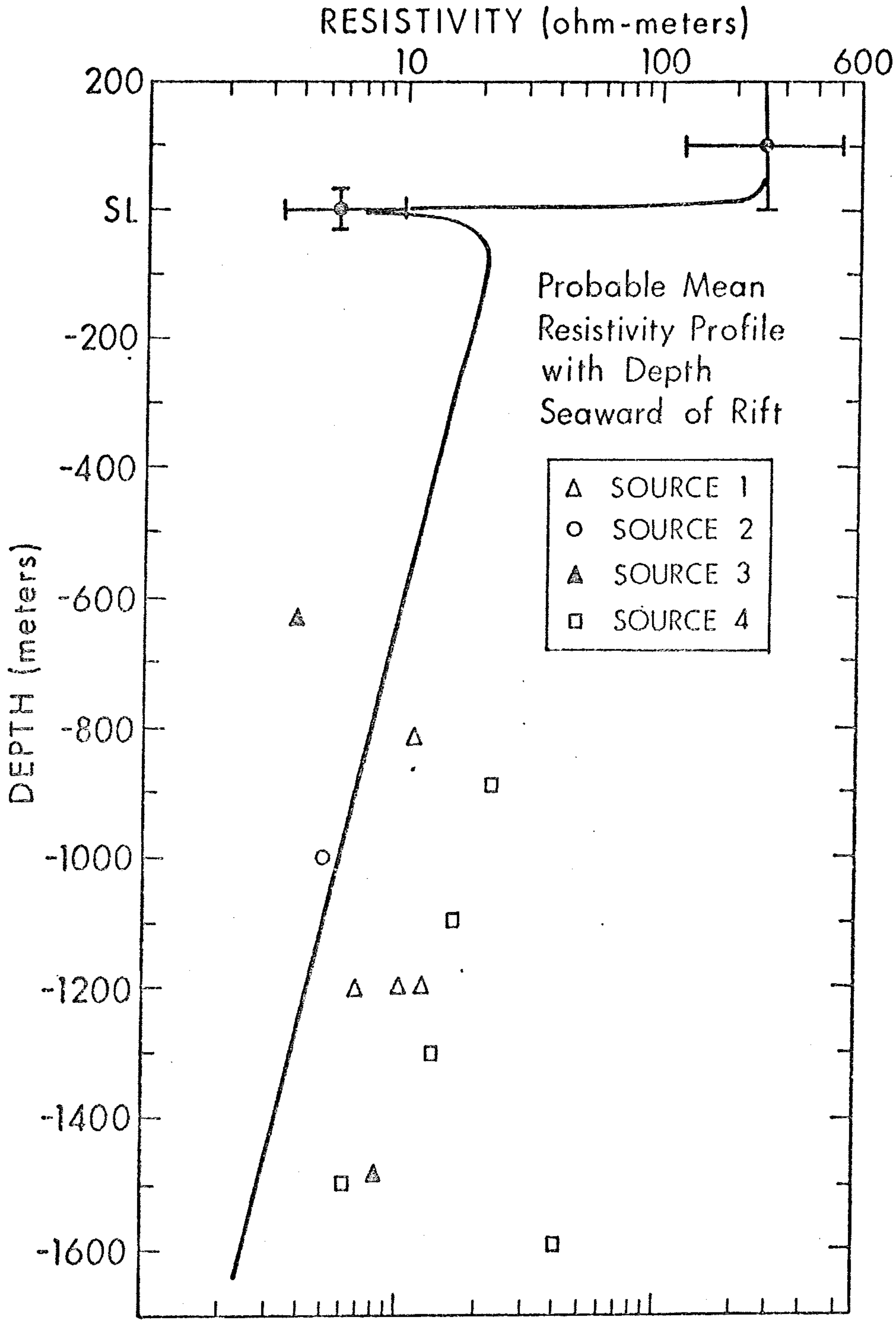

Fig. 11. Preliminary depth-resistivity profile seaward of the Kilauea East Rift in Puna. The basic control points are based on induction data (see text). The horizontal bars refer to the spread of data and have no statistical significance. The deep control points (from line-loop induction data with sources 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):

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

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.

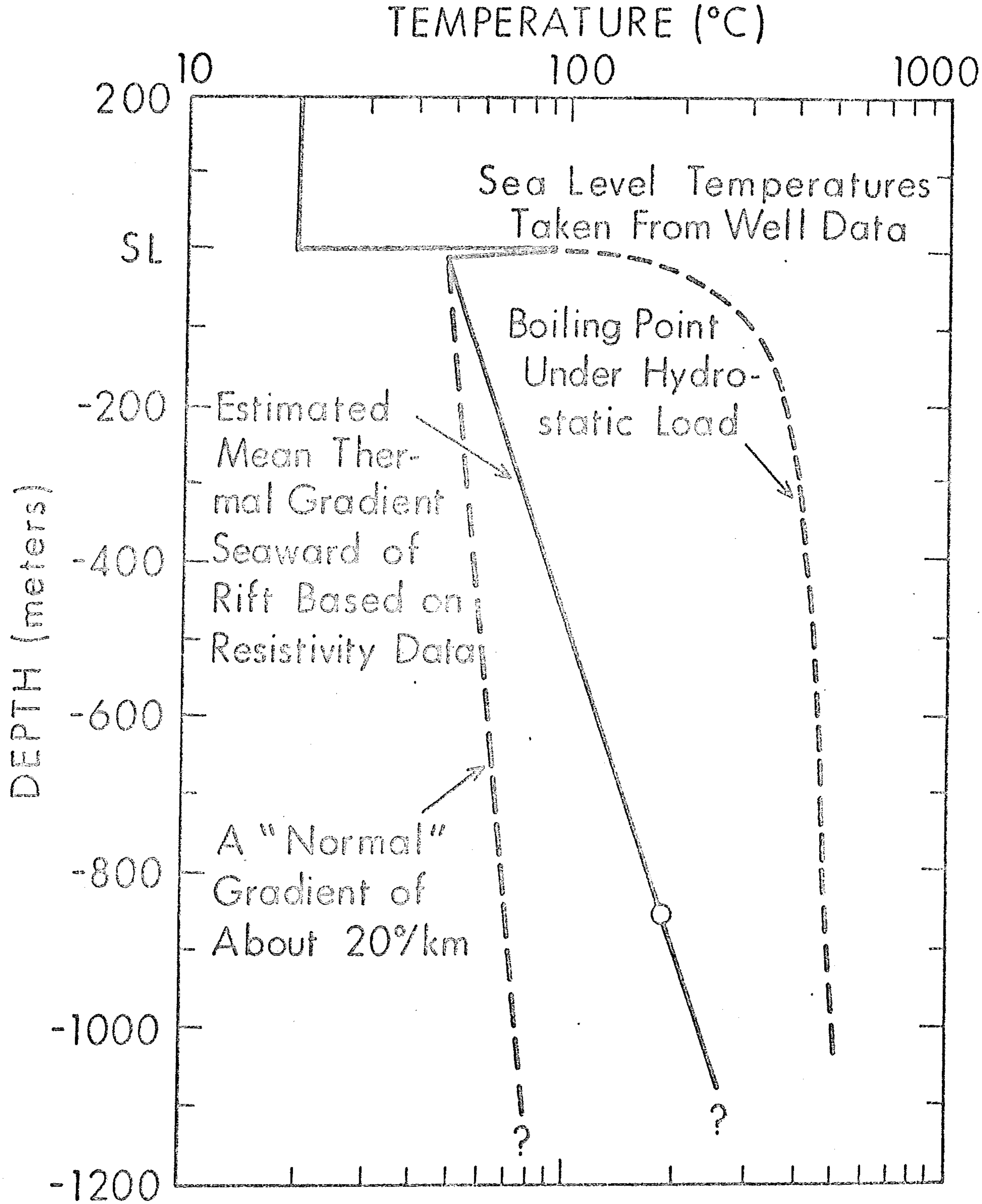

Fig. 12. Preliminary depth-temperature profile seaward of the Kilauea East Rift in Puna. The circle at 850 meters depth is the calculated temperature for a 5 ohm-m resistivity (see text). The temperature "gradients" indicated are to be considered in a qualitative sense only.

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.

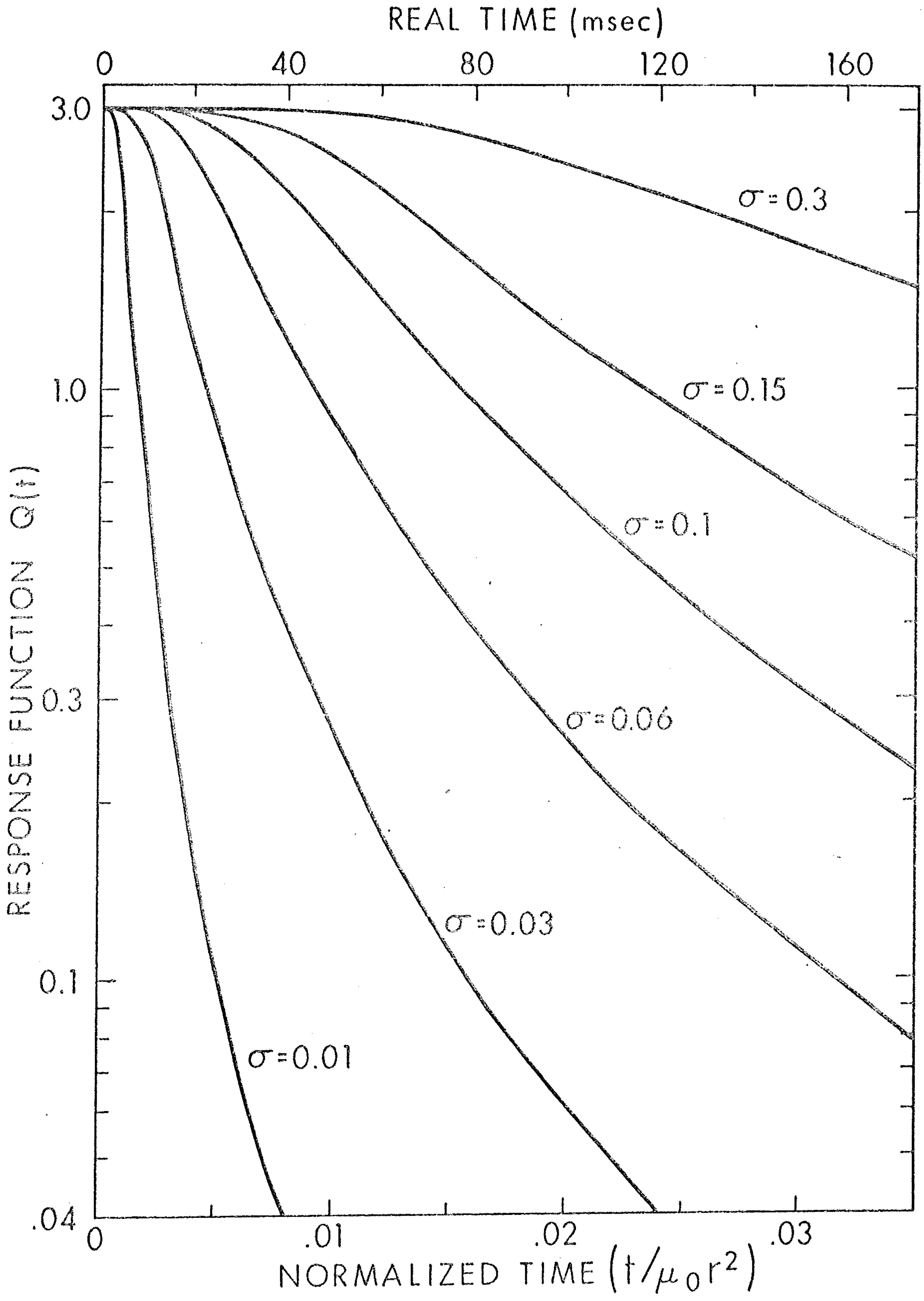

Fig. 13. Time-decay response of a line-loop inductive system. Q(t) is the receiver decay function for a sudden change in source current over an infinite half-space of uniform conductivity, σ; μ_o is the magnetic permeability in free space (mks units); t is time in seconds; and r is source-receiver separation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

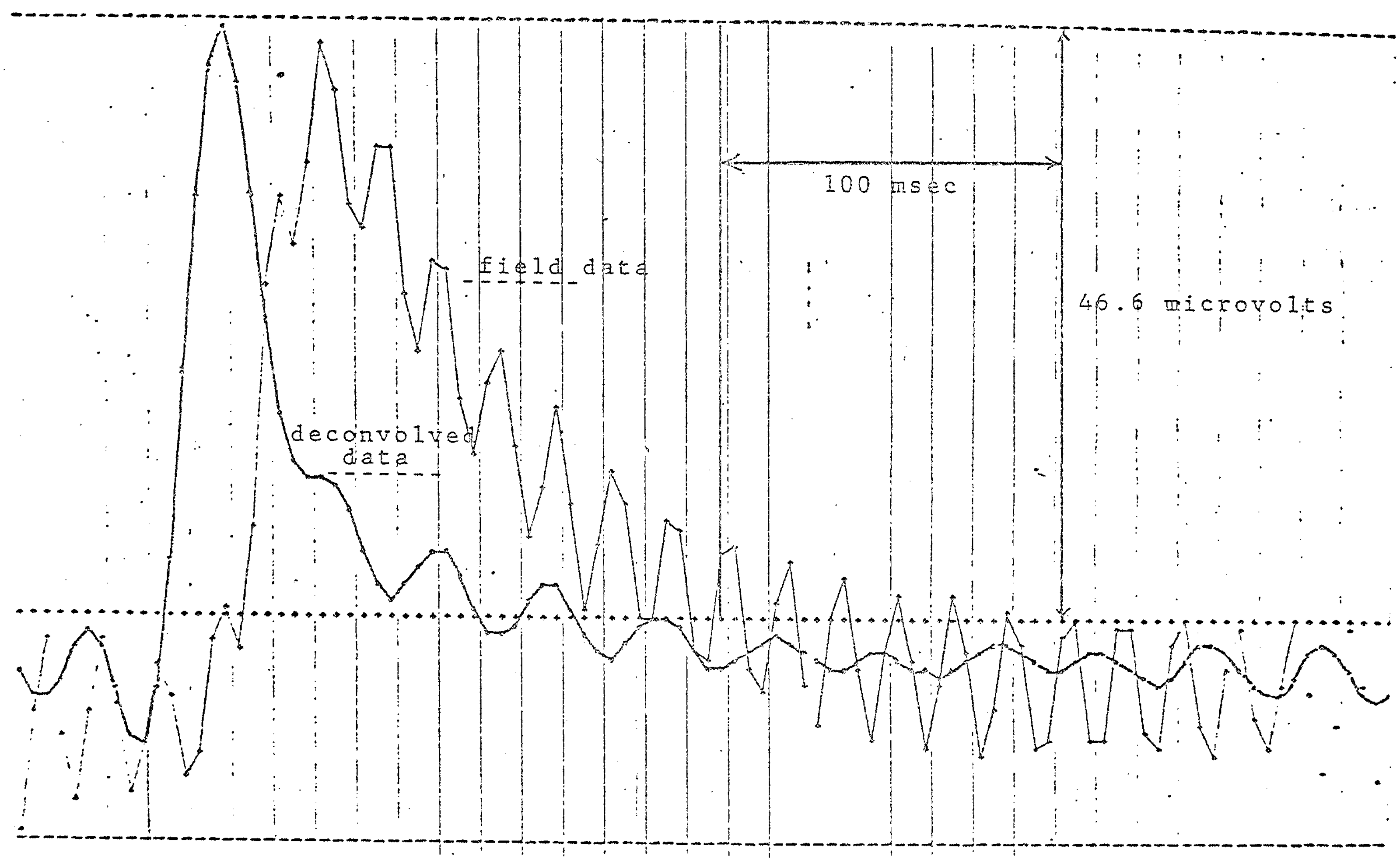

Fig. 14. Time-domain field data example. Digitized field data (light curve) and deconvolved data (heavy curve) for voltage transient observed at station 29 (Puna).

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