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GEOTHERMAL EXPLORATION: AN EVALUATION OF THE MICROSEISMIC GROUNDNOISE METHOD

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ABSTRACT

Geothermal energy can probably make significant contributions to our rapidly increasing energy requirements. Because of this, major exploration programs are underway in many countries at the present time. A variety of geological and geophysical methods has been used to study known geothermal systems and to explore new areas for this valuable resource. Some of these methods have worked better than others, and some have not worked at all. The exploration significance of one of these methods, microseismic groundnoise studies, has been questioned. An attempt to resolve this question and to test the original geothermal-groundnoise hypothesis, that is: geothermal reservoirs are dynamic systems and can be located by measuring ground motion above them, resulted in extensive groundnoise surveys in two areas which appear to have great geothermal potential. These surveys were carried out at Kilauea Volcano, and the East Rift of Kilauea Volcano, on the Island of Hawaii. These areas were selected because of their relatively homogeneous geologic environment, for other surveys indicate that "lithologic noise" tends to be a problem when evaluating groundnoise data. Even in this relatively ideal environment, the surveys tend to indicate changes in the acoustic properties of rocks, rather than

changes in groundnoise activity resulting from dynamic geothermal reservoirs.

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INTRODUCTION

A geothermal exploration method which utilizes anomalous microseismic-groundnoise intensity has received considerable attention and has been extensively applied in delineating geothermal reservoirs in many prospect areas. This prospecting method was introduced by Clacey (1968) with preliminary results of a survey in the Taupo volcanic region of New Zealand. A more comprehensive review of the groundnoise data in this region was presented by Whiteford (1970). Similar surveys have been conducted on the Lipari and Vulcano Islands, Italy (Luongo and Rapolla, 1973), and in the Imperial Valley, California (Goforth et al., 1972).

The microseismic-groundnoise survey by Clacey indicated anomalously-high semismic-noise levels at low frequencies (1 to 10 Hz) in an area of known geothermal potential. These anomalous noise levels are supposedly related to fluid movement or pressure fluctuations within dynamic geothermal reservoirs. Although Clacey did not supplement his empirical measurements with a theoretical model, Douze and Sorrels (1972) have attempted to explain anomalous groundnoise intensities in the Imperial Valley of California with a model geothermal reservoir having dynamic characteristics.

A recent study by Iyer (1974) in the East Mesa area of the Imperial Valley does not substantiate the anomalous groundnoise activity reported by Goforth and others (1972). Iyer found anomalous 2.5 Hz noise levels over the entire survey area but they were a result of a small waterfall on the All American Canal, and not a geothermal system. Direction studies using L-shaped arrays gave many non-unique solutions as well.

Iyer concludes "...we were unable to detect seismic noise over the Mesa thermal anomaly that could not be attributed to cultural noise". He also states that much more work is needed "...to answer the interesting question of whether seismic noise is generated by hydrothermal systems of the type found in that (Imperial Valley) region".

Because the exploration significance of this method is in question, two groundnoise surveys were carried out on the Island of Hawaii. The survey areas included Kilauea summit and the East Rift of Kilauea Volcano. These areas were selected because of their relatively homogeneous geologic environments, for other surveys had indicated that "lithologic noise" tends to be a problem when evaluating microseismic groundnoise data.

ENERGY AND THE ECONOMY

Recognition of geothermal energy as a major source of low cost, relatively pollution free electric power has initiated an unprecedented interest in increased development of this natural resource. Although geothermal energy has been used in many countries for electric power production and space heating, the impending energy crisis in the United States has caused major energy producers and government entities to view geothermal resources as a viable energy source, capable of significant contributions to our rapidly increasing energy requirements.

Analysis of projected demands for electricity indicates a compound growth rate for the rest of the century. Experts agree that the consumption of electric power will double every ten years. If predictions of future electric power consumption are correct, we will require more than 10^9 kilowatt hours (Tkwh) in the year 2000, at least six times the electricity we use today (Chapman et al., 1972). The resources we presently use to produce this energy are limited and tend to present environmental problems, so we must consider alternate methods to supplement conventional energy sources.

Estimates of depletable energy resources within the United States are illustrated in Table 1. The figures in the table indicate

the number of years the resource will be available for commercial exploitation at current rates of use. The resources listed as recoverable, include known reserves. The resources listed as total, include estimates of expected, but not necessarily recoverable, reserves.

RESOURCE	RECOVERABLE (YRS)	TOTAL (YRS)
FOSSIL FUELS		
Coal	125.0	1300
Petroleum	5.0	280
Natural Gas	5.0	110
Oil Shale	-	2500
NUCLEAR FISSION		
Conventional Reactors	2.3	15
Breeder Technology	115.0	750
NUCLEAR FUSION		
Deuterium-deuterium	-	10^9
Deuterium-tritium	-	10^6
GEOTHERMAL HEAT		
Steam and Hot Water	0.2	60
Hot Rock	-	600

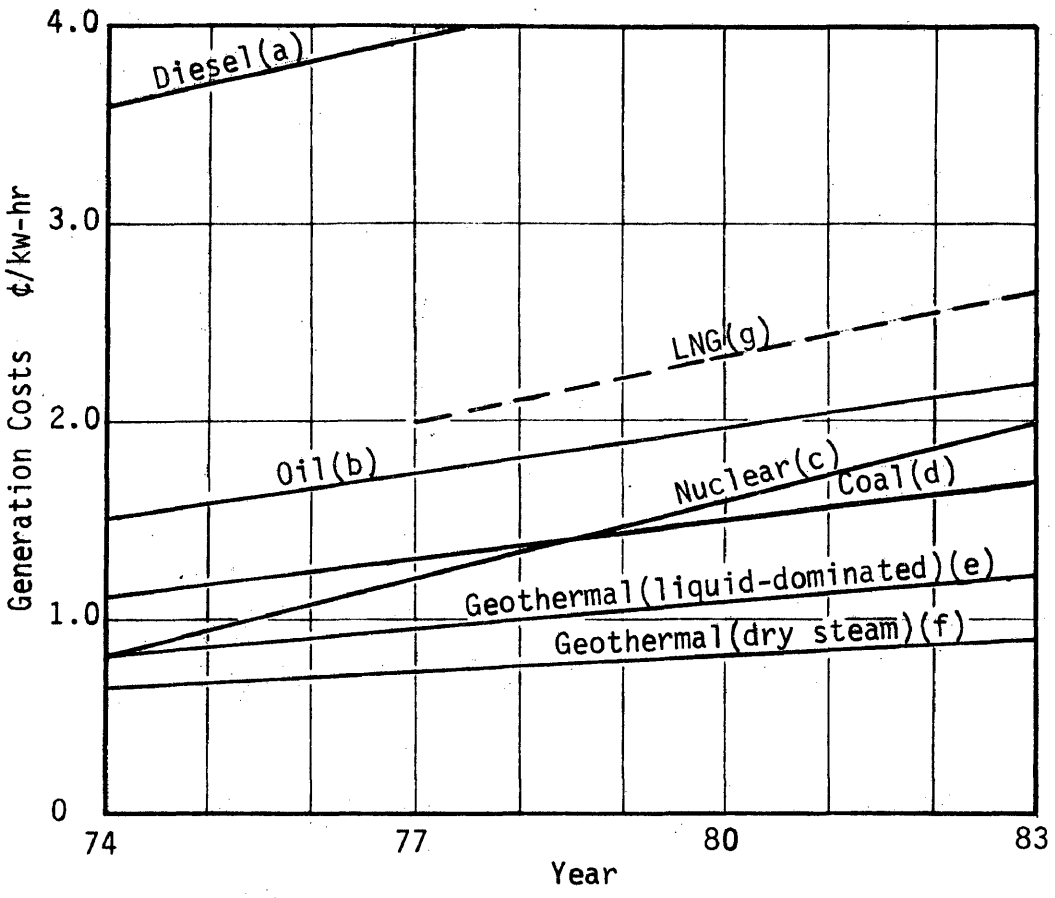
Table 1. Estimates of depletable energy resources in the United States. The figures in the table indicate the number of years the resource would last if all energy came from that source, at current rates of use (from Hammond, 1972).

Fossil fuels will probably continue to be the major source of energy in the near future, although nuclear power plants are rapidly increasing in number. Fossil fuel and nuclear power plants, however, have been criticized by many environmental groups because of air pollution and fear of nuclear accidents. These groups have been lobbying for increased

government support of research into other energy resources. These include solar radiation, wind power, and sea thermal gradients. The only one of these that is practical at the present time is solar radiation for home heating and cooling. The natural resource that can be used to produce electric power at low cost and with a minimum of environmental problems is geothermal energy. This is the reason for the greatly increased interest in geothermal energy.

Although it is difficult to compare average electric power generation costs by various generation methods, an attempt has been made in Figure 1. The difficulty in comparing costs is a result of differences in interest, taxation, amortization, depreciation, and a number of other hidden factors, for each of the generation methods (Koenig, 1973). The average generation costs illustrated in Figure 1 are based upon estimated fuel and operating expenses as noted. The energy sources used in this compilation include diesel fuel,oil, nuclear, coal, liquid-dominated geothermal systems, vapor-dominated geothermal systems, and liquified natural gas. Hydroelectric power generation is generally less expensive than all other electric power sources. However, geothermal power production will be economically competitive with hydroelectric facilities when most of the ideal hydroelectric sites have been occupied.

Average power costs at several geothermal areas of the world are illustrated in Table 2. The geothermal power costs versus the local conventional power costs are compared for both electric power generation, space heating, refrigeration, and drying facilities. In all cases, the geothermal derived energy is less expensive than the average conventional source. These studies indicate that geothermal energy is a natural resource which has,and will



NOTES:

(a) Diesel based power as determined by price of diesel fuel, assumed at $20/bbl.

(b) Oil fired plants: $7/bbl of residual fuel, $340/Kw for an 880-mw installation.

(c) Nuclear plant cost $600/Kw in 1980 for 1000-mw plants; fuel at 3 mills/Kw-hr in 1973 escalated at 5% per year; waste disposal, environmental protection-1 mill/Kw-hr; operation-1 mill/Kw - No escalation in plant costs.

(d) Coal; 11.3 mills/Kw-hr for a plant going on stream in 1972.

(e) Geothermal, liquid dominated, Ahuachapan-type, 7 mills/Kw-hr in 1971, escalated at 5%.

(f) Geothermal, vapor dominated, Geyser-type, 6.2 mills/Kw-hr in 1973, escalated at 5%.

(g) LNG-liquified natural gas: assumed delivery cost of $1.50/1000 CIF in 1977.

Figure 1. Generation Costs (from Meidav, 1974).

continue to have, practical and industrial significance.

GEOTHERMAL APPLICATION	LOCATION	GEOTHERMAL PRODUCTION	LOCAL AVERAGE OF OTHER FUELS
Electric Power, U.S.mills/10^3 watt hours (Kwh)	Namafjall, Iceland	2.5-3.5	---
	Larderello, Italy	4.6-6.0	~ 7.5
	Matsukawa, Japan	4.6	~ 6.0
	Cerro Prieto, Mexico	4.1-4.9	~ 8.0
	Pauzhetsk, U.S.S.R.	7.2	~10.0
	The Geysers, U.S.	5.0	7.0
Space Heating, U.S.$/$10^9$ cal (Gcal) energy	Reykjavik, Iceland	4.0	6.7
	Sweged, Hungary	3.0	11.0
Refrigeration, U.S.$/Gcal energy	Rotorua, New Zealand	0.12	2.40
Diatomite Drying, U.S.$/ton	Namafjall, Iceland	~ 2	~12

Table 2. Selected Comparative Cost Data For Geothermal Energy (from Koenig, 1973).

GEOTHERMAL RESOURCES DEVELOPMENT

Geothermal resources have been utilized throughout historic time. Early uses include thermal spring baths and steam condensation for drinking water in arid climates. Modern exploitation includes industrial and home heating, mineral recovery from hot water condensates, and electric power production. In 1812, a boric acid recovery plant was started in Larderello, Italy. The mineralized water from fumeroles was boiled in cauldrons and boric acid was recovered from the residue. The first continuous electric power generation began in 1913 at Larderello when a 0.25-Mw generator was installed. At the present time, the total power capacity of all Italian geothermal power plants is 390 Mw (Koenig, 1973).

Geothermal resources have been developed for a number of industrial and private uses in the Taupo volcanic region of the North Island of New Zealand. These include the development of a heat-exchanger system for the production of high quality steam for pulp and paper mill processes at Kawerau, the use of natural steam for timber drying at Kawerau, the use of natural hot water in space heating of houses, hospitals, schools and hotels at Rotorua, the development of a geothermal air conditioning system for a hotel at Rotorua, numerous agricultural and experimental forestry applications at Lake Taupo, and the production of electric power at Kawerau and Wairakei. There is a great potential for continued

geothermal utilization in New Zealand.

The primary use of geothermal energy in Iceland has been in municipal heating systems. Approximately 50 percent of the population of Iceland use hot-water wells for home heating. Thermal waters provide the energy needed for greenhouse operations in southwestern Iceland, and dry geothermal steam is used for diatomite processing at Nomafjall. Although geothermal electric power production is limited at the present time, the availability of cheap and abundant geothermal energy will probably be important for Iceland's future power needs.

Hot springs in Japan have been used as therapeutic spas, baths, and resorts for centuries. Recent exploitations of geothermal resources include electric power production, space heating, sulfur and salt recovery, and rice processing. Many volcanic centers have been explored on the Islands of Japan. Continued exploration and future development of geothermal resources is expected to supplement the electric power requirements of this industrial nation.

The completion of a 5-Mw power plant at Pauzhetsk in 1967, the generation of 0.68 Mw from a freon-based heat-exchanger at Paratunka in 1970, and the space-heating operations at Makhoch Kala, are the most significant geothermal developments in the Soviet Union. Continued exploration and drilling on the Kamchatka Peninsula, at the Caspian Sea, in Siberia, and many other areas, indicate a high government priority to develop this abundant natural resource.

Drilling and production tests at Cerro Prieto, Baha California, Mexico, indicated a large geothermal power potential. A 55-Mw power plant was constructed and put on-line at Cerro Prieto in 1974. The results of geophysical exploration and continued drilling indicate a

considerable expansion of this power generation facility.

The most important utilization of geothermal energy in the United States is in the production of electric power at The Geysers, California. The initial 12.5-Mw installation of 1960 has grown to a present 392-Mw plant capacity. Plans to install additional 110-Mw power plants on an annual basis through the 1970's will result in a 1,180-Mw power-generation facility in 1980. Exploration and evaluation of many geothermal areas in the western United States will probably result in the development of more power facilities in the near future, especially if greater advances are made in modest-enthalpy hot-water generation systems. Although electric power generation is the primary concern of many energy producers in the United States, hot geothermal waters have been used for space heating in Klamath Falls, Oregon; Boise, Idaho; and a few other cities within the country for many years.

A review of geothermal resources development would not be complete without mention of the many countries that have active geothermal exploration and development programs in progress. As indicated by Koenig (1973), at least 50 countries are interested, or have active exploration projects, in geothermal resource evaluation. Figure 2 illustrates the worldwide status of geothermal exploration and development.

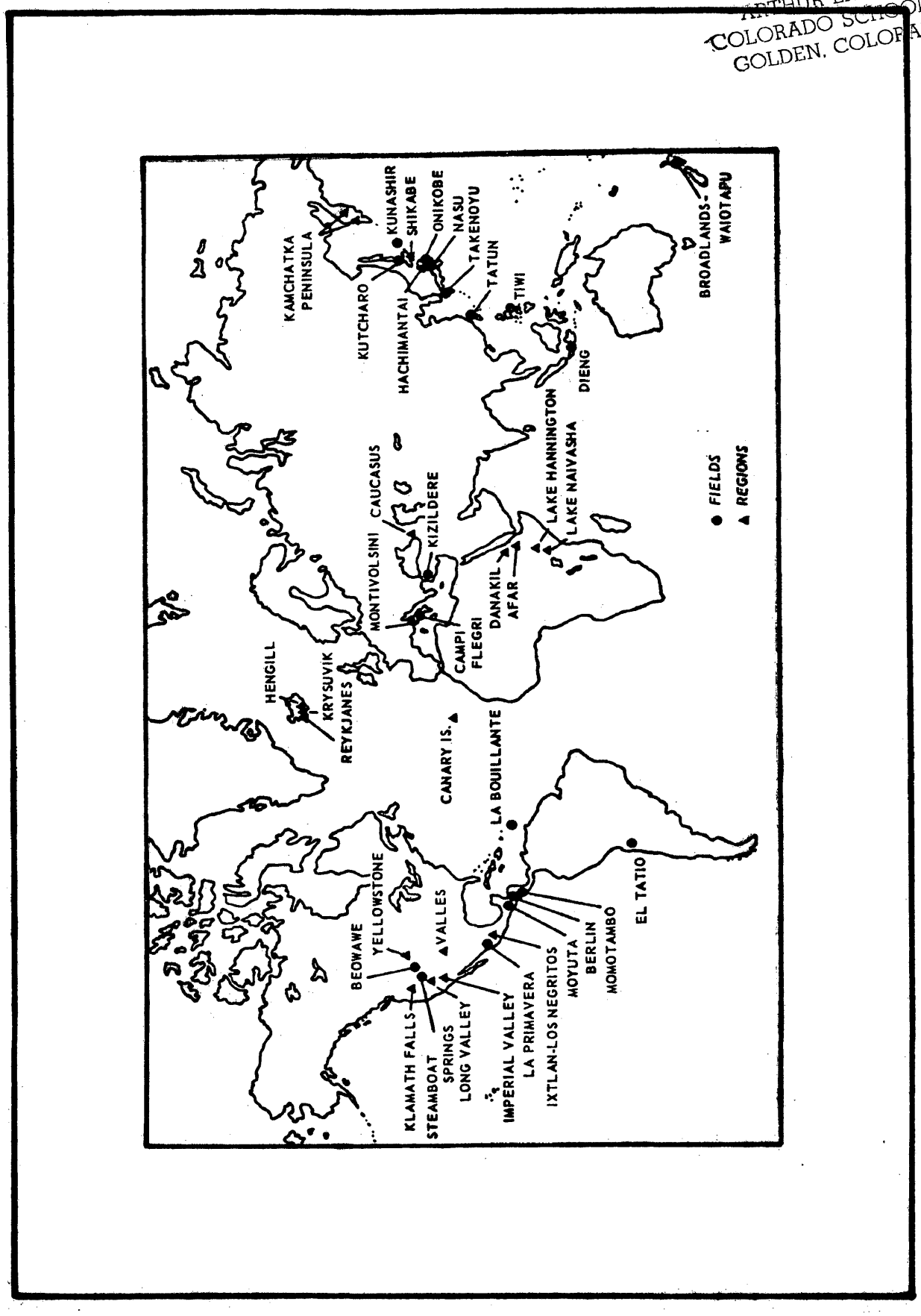


FIGURE 2. Major Undeveloped Geothermal Fields And Regions Under Exploration (Koenig, 1973).

EXPLORATION FOR GEOTHERMAL RESOURCES

The resource that is being sought is the naturally occurring concentrations of thermal energy which are within economic reach of the earth's surface. Exploration for these resources has been guided by surface thermal manifestations throughout the world. Because the energy must be stored in a form which can be extracted (high temperature water or superheated steam), any geological, geophysical, or geochemical method which is based upon the thermal properties of rocks and water can be a potentially useful tool to locate these near surface geothermal reservoirs.

A variety of methods has been used to study known geothermal systems and to explore new areas for this valuable resource. Some methods have worked better than others, and some have not worked at all. The geologic environments in which geothermal systems occur are varied and can be very complex. Because of this, no single exploration method will necessarily provide the required information in all geothermal provinces. It is more likely that a combination of exploration methods will be required to locate, and to eventually develop, all geothermal systems that have yet to be found.

Geology And Hydrology

Exploration for geothermal resources has usually been confined to areas of known geothermal manifestations. These thermal indicators have included hot springs, geysers, fumeroles, steaming ground, and hot groundwater wells. High-temperature shallow geothermal reservoirs are usually associated with tectonically active areas near the margins of crustal plates (Muffler and White, 1972). These areas are characterized by recent igneous activity, the geological phenomenon which transfers deep-seated thermal energy to near surface meteoric water.

The purpose of conventional geological mapping, which is aided by photogeology and other remote sensing techniques (infrared scanning, radar, and microwave) is to provide information on (Grose, 1971; Combs and Muffler, 1973):

1) the tectonic and stratigraphic setting of the area,
2) the geometry of geologic structures,
3) the age and occurrence of volcanic rocks within the area,
4) the nature and extent of hydrothermal alteration and mineralization, and
5) the location, size, and elevation of modern thermal springs.

A detailed hydrological study serves two purposes. It provides basic information for modeling the underground thermal regime and it can be beneficial in planning future exploration programs. The hydrology study should include (Combs and Muffler, 1973):

1) temperature and discharge measurement on both hot and cold springs,

2) determination of the water table if wells are available,

3) determination of groundwater movement, and

4) chemical analyses of available springs.

Although tectonically active areas have been considered prime targets for geothermal exploration, a number of thermal manifestations occur in other geological settings. Scientists in the Soviet Union have realized that many deep sedimentary basins may provide commercial geothermal development in remote areas (Facca, 1970). If our present technology can be adapted to utilize hot water at or near boiling temperatures on a competitive basis with fossil fuel sources, important geothermal developments will occur in a variety of geological settings.

Temperature And Heat Flow

Surface temperature variations have been studied in several thermal areas from aircraft equipped with infrared scanners. These surveys include known geothermal areas in Iceland (Friedman et al., 1969; Palmason et al., 1970), Mexico (Gomez Valle et al., 1970), New Zealand (Hockstein and Dickinson, 1970), and the United States (Moxham, 1969; Lee, 1969).

The infrared scanners which have been used to measure near surface temperature variations are sensitive to radiation which is emitted in the near- and intermediate-infrared range. They normally operate in the 3- to 5-, and the 8- to 14- micrometer spectral bands. Although this type of reconnaissance survey would be most useful in geothermal prospect areas which have limited geological, geophysical, and 'ground truth' data, most of the surveys have been flown in areas where surface thermal manifestations have been previously known and

mapped. This tends to indicate that the airborne infrared technique is being evaluated for exploration significance rather than being used as an applied geothermal exploration tool.

Temperature and temperature gradient measurements at shallow depths (1-150 meters) have been used to outline anomalous areas in many geothermal prospects (Combs and Rex, 1971; Combs, 1971; Lovering and Goode, 1963; Burgassi et al., 1964; Thompson et al., 1964). Because the resource that one seeks is anomalous geothermal heat, a direct measurement of temperatures and temperature gradients is presumably the most logical and direct exploration technique. This exploration philosophy, although sound in basic concepts, can be misleading because of:

1) near surface changes in geology which can result in thermal conductivity variations,
2) lateral groundwater movement which can mask centers of thermal activity, and
3) vertical convection which can result in high thermal gradients in the rock above a convection cell, but very low thermal gradients below it.

Because these temperature and thermal conductivity studies can be misleading, a comprehensive study of the geology and the hydrology of an area is necessary before conclusions can be made about temperature distributions at depth.

Geochemistry

Geochemical methods have been used for geothermal resource exploration, evaluation, and exploitation. White (1970) and White and

others (1971) have summarized the important geochemical properties of both hot-water and vapor-dominated systems. Empirical studies have indicated correlations between chloride content and reservoir type. Surface springs and subsurface water are generally low in Cl (<20ppm) when associated with vapor-dominated systems. The Cl content of near surface waters is generally much greater (>50ppm) when associated with hot-water geothermal systems.

Several empirical geothermometers have been developed for estimating underground temperatures in liquid-dominated geothermal systems. Silica content in surface thermal waters is used to estimate equilibrium temperatures with quartz (Fournier and Rowe, 1966; Mahon, 1966; Truesdell, 1970; White, 1970; White et al., 1971).

Other geothermometers include measurements of Na/K ratios in thermal spring waters and comparing these ratios to known solubility curves. This method of estimating underground temperatures below 300°C has been described by White (1965, 1968, 1970), Ellis and Mahon (1967), Ellis (1969, 1970), and Fournier and Truesdell (1970). This Na/K method is most applicable in areas which have similar types of rocks and minerals because the solubilities of these ions versus temperature are dependent upon the minerals which control solution composition.

An empirical NA-K-Ca geothermometer was developed by Fournier and Truesdell (1973) and appears to give better temperature estimates of water-rock equilibrium. As with other geochemical methods of estimating subsurface temperatures, however, this geothermometer can also be affected by continued water-rock interaction as the thermal fluids ascend and cool, and by mixing of thermal fluids with near surface

ground water.

Geochemical data are not only useful in estimating subsurface equilibrium temperatures, but can also be used (Grose, 1971):

> "...in predicting size, extent, volume, and variations in permeability of the reservoir, and compositional changes, flow patterns and recharge potential of the reservoir fluids. ... Furthermore, geochemical studies during exploitation are very valuable with respect to depletion and energy changes within the reservoir, corrosion and precipitation problems in the well and power plant facilities, effluent disposal at the surface or underground, and economic recovery of chemicals and/or fresh water from the geothermal brines."

Gravity And Magnetic Measurements

Gravity and magnetic measurements have added a third dimension to surficial geologic mapping in many thermal areas. Although geothermal systems are very complex and not very well understood, physical property measurements (rock density and magnetic permeability) obtained in drill holes within geothermal areas, have resulted in more accurate structural modeling.

The primary use of gravity measurements has been to determine the depth to basement rocks where thermal manifestations occur within sedimentary basins. These depth determinations have helped locate buried faults which, in many cases, serve as conduits for geothermal-fluid migration. These measurements have also been used to locate buried igneous rocks, and rocks which have been hydrothermally altered by geothermal fluids (Hockstein and Hunt, 1970; Biehler and Combs, 1972).

Magnetic surveys have also been used to locate buried faults and igneous intrusions. Several surveys have also shown magnetic lows

associated with thermal springs in geothermal areas. These anomalies are caused by the hydrothermal alteration of magnetite to pyrite (Studt, 1964).

Gravity and magnetic potential methods, like any geothermal exploration method used at the present time, are most useful when combined with other geological, geochemical, and geophysical studies. An integrated exploration approach is the only way that the complex geological environments of geothermal systems will be understood

Electrical Resistivity

The conduction of electric current in rocks and minerals can take place by electronic or ionic processes. Because all rocks at the earth's surface are porous, however, ionic conduction is the usual transfer mechanism. Under normal conditions, the pore spaces of rocks are either partially or completely filled with an electrolyte. The amount of the electrolyte and the salt concentration usually control the resistivity of a rock (Keller and Frischknecht, 1966).

Temperature is another parameter which effects the resistivity of an electrolyte within a rock and causes variations in rock resistivity in geothermal areas. To illustrate the temperature dependence of rock resistivity, and thus the importance of resistivity measurements in geothermal exploration, let us assume we have a rock with a resistivity of 100 ohm-meters at a temperature of 20°C. If we take this rock and raise its temperature to 200°C while maintaining sufficient pressure to prevent vaporization of the electrolyte, the resistivity of the rock will decrease to approximately 18 ohm-meters. This thermal dependence

of electrical resistivity has been used extensively in geothermal prospect evaluation.

Galvanic Resistivity Methods

Resistivity measurements are usually made with four terminal array. Current is passed into the ground with a pair of source electrodes and a potential is measured as a result of this current with a pair of receiver electrodes. Several galvanic resistivity methods have been used in geothermal exploration. The most important have been dipole mapping, a reconnaissance resistivity method, and vertical electric sounding, a method which is used to detail vertical resistivity distributions.

Vertical Electric Sounding

The vertical electric sounding method has been used to study many geothermal prospects (Breusse and Mathiez, 1954; Zhody et al., 1973; Banwell, 1970; Banwell and Macdonald, 1965; Hatherton et al., 1966; McQuen, 1970; Meidav, 1970; Keller, 1966; Meidav and Furgerson, 1972). A sounding is done by starting with a very closely spaced electrode array to measure near surface resistivites. The array is then systematically expanded to provide information on resistivity variations at greater depths. Apparent resistivities are calculated for each of the array spacings, and both of these are plotted on log-log paper. The resulting resistivity-versus-depth curve is compared to theoretical curves (such as those by Orellana and Mooney, 1966), or input to automatic computer schemes (Crous, 1971; Zhody, in print) for final interpretation.

Dipole Mapping

The dipole mapping resistivity method (Keller, 1966; Risk et al., 1970; Stanley et al., 1973) has been used extensively in geothermal exploration. It is a reconnaissance method that is used to investigate conductive bodies at depth. The electric field about a long-line source (1 to 4 km) is measured with two orthogonal receiver lines (30 to 100 m) at various locations. The source current, the total electric field at the receiver location, and the source-receiver geometry are used to calculate resistivities which would be measured over a half space. Variations from true resistivities in a half space, or apparent resistivities for a layered earth, are then investigated for geothermal significance.

Stanley and others (1973) indicated several facts which should be considered when using the dipole mapping technique:

1. The depth of investigation at a measurement point depends on:
    a. The distance and azimuth of the measurement point from the source dipole.
    b. The length of the source dipole.
    c. The geology.

2. The definition with which a particular geologic unit can be mapped is dependent in part upon the electrical properties of the surrounding geologic units.

3. The total-field mapping technique is well suited for locating large electrical heterogeneities. The anomaly then should be investigated using an electrical or electromagnetic sounding technique to find the resistivity distribution with depth.

4. In the presence of steeply dipping contacts, apparent resistivities measured over a large conductive body with the source dipole located over resistive material will be higher than the true resistivity of the conductive body. Conversely, the apparent resistivity measured outside the body, with the source dipole located over the conductive body, will be lower than the true resistivity at the measurement point.

Electromagnetic Methods

Electromagnetic methods have not been used as extensively in geothermal exploration as direct-current methods. Recent developments in equipment design and data interpretation, however, have resulted in more numerous applications. The methods that have been used to study the electrical properties of geothermal systems include the audio-magnetotelluric technique, the time-domain electromagnetic sounding technique, and the vertical magnetic dipole technique. A summary of these methods can be found in Keller (1966), Vanyan (1967), Keller (1970).

Audio-Magnetotelluric Method

The audio-magnetotelluric method has been used to study resistivity distributions in several geothermal areas. These include surveys on the North Island of New Zealand (Keller, 1969), at Momotombo Volcano in west-central Nicaragua (Harthill, 1971), and in the Long Valley area of California (Hoover et al., 1973).

Cagniard (1953) developed the theory for the magneto-telluric method. Starting with Maxwell's equations, and assuming that the electromagnetic wave is planar and impinging vertically on the surface of a horizontal, homogeneous, isotropic earth, Cagniard showed that the apparent resistivity is related to the electromagnetic fields and angular frequency of the impinging electromagnetic wave.

The recording procedure consists of measuring orthogonal components of the earth's natural time-varying electromagnetic field. The source of this electromagnetic energy is lightning storms throughout the world. An electric field is measured with a horizontal electric dipole while the magnetic field is measured with a horizontal-axis induction loop or

magnetometer. These two sensors are oriented perpendicular to one another. Ratios of the electric-field intensity to the magnetic-field intensity at various frequencies are used to calculate apparent resistivities. Because the apparent resistivity and the depth to which an electromagnetic wave penetrates the earth are functions of frequency, a sounding curve of apparent resistivity-versus-depth is obtained.

Time-Domain Electromagnetic Method

The time-domain electromagnetic method has been used in several geothermal studies. These include surveys at the summit of Kilauea Volcano, Hawaii (Jackson and Keller, 1972), of the Taupo Volcanic Region of New Zealand (Keller, 1969), and the East Rift of Kilauea Volcano, Hawaii (Skokan, 1974). Published information describing the method and data interpretation includes: Harthill (1969), Jacobson (1969), Silva (1969), Keller (1970), and King (1972).

An electromagnetic field is introduced into the earth with a current step into a grounded source dipole. Transient fields at various receiver locations are recorded with vertical-axis loops or grounded electric dipoles. The time rate-of-change of the recorded fields are related to resistivity distributions at depth.

Vertical Magnetic Dipole Method

The vertical magnetic dipole method has been used to study near surface resistivity variations which are related to deep-seated geothermal sources. A few of the published surveys which discuss field methods and data interpretation include those by Lumb and Macdonald (1970) in the Broadlands Field, New Zealand, and at El Tatio in Chile, by Keller (1970) at Kilauea Volcano, Hawaii, and by Petrick (1974) at

Roosevelt Hot Springs, Utah.

A time-varying electromagnetic field is applied to a vertical-axis source loop. This field produces secondary induced currents in conducting materials in the vicinity of the primary field. If one measures the magnetic field at a receiver location with a similar vertical-axis loop receiver, the measured field is a resultant of the primary and induced secondary fields. The electrical and magnetic properties of the conductive body are determined by comparing the phase and amplitude of the secondary field with those of the primary field. If several frequencies are used at a particular source-receiver location, a sounding curve of apparent resistivity-versus-depth is obtained. If one frequency is used and the source-receiver system is moved horizontally over the surface, a profile is obtained.

This type of system has been most useful for locating conductive ore bodies in high resistivity host rocks. Low input power and minimum frequencies of 10 to 20 Hz have limited the depths of investigation in many low-resistivity geothermal environments.

Self-Potential Method

The self-potential method has not had very much application in geothermal prospecting but it may become more important as more is learned about the nature of geothermal systems. This method has been used in a study of the Mud Volcano area in Yellowstone National Park (Zhody et al., 1973), and a study of Long Valley Caldera, Mono County, California (Anderson and Johnson, 1974).

The self-potentials that one measures on the surface of the

earth can be caused by oxidation-reduction reactions, streaming potentials, and thermoelectric effects. The latter of these, however, can produce only small self-potential effects in the thermal range we are considering in a geothermal system. Early experiments by Poldini (1938 and 1939) showed that a positive potential would be produced at the top of an upward migrating water column. It is hoped that this effect, or an oxidation-reduction chemical reaction within or around a geothermal reservoir, will produce self-potential anomalies which can be used to delineate geothermal systems.

Seismology

Seismic methods in geothermal exploration include the active reflection and refraction techniques that have been used for many years in oil and gas exploration, and the passive microseismic groundnoise and microearthquake techniques which have been developed as specialized geothermal exploration tools.

Reflection and Refraction Seismology

Reflection and refraction seismic methods have had limited use in geothermal exploration. These tools had been developed in the oil industry for subsurface-structure and basement-configuration determinations. Geothermal applications include a seismic reflection study by Hayakawa (1970) in several areas within Japan, and seismic refraction studies by Hockstein and others (1967), Hockstein and Hunt (1970), Grindley (1970), and Thompson and others (1967) in the Broadlands geothermal field in New Zealand and the Dixie Valley geothermal area of

Nevada. Because most geothermal areas have complex geologic structure, advancements in seismic data handling and enhancement techniques which are being applied in the oil industry may result in more frequent applications of these methods to geothermal exploration. Velocity analysis and seismic attenuation characteristics may be very helpful in delineating the depth and areal extent of geothermal reservoirs. Analysis of many more case histories will be required before any conclusions can be made about the geothermal exploration significance of these tools.

Microearthquakes

Geothermal systems generally occur in tectonically active zones around the world. It is within these zones of deep-seated crustal tectonics that thermal energy is transferred to near-surface environments by both igneous-rock and thermal-fluid movements. Fault displacements and their associated microearthquake activity within some of these geothermal areas has been described by many authors. These include Lange and Westphal (1969), Hamilton and Muffler (1972), Ward and others (1969), Ward and Bjornsson (1971), Brune and Allen (1967), Ward and Jacob (1971), and Butler and Brown (1974).

The location of this near surface microseismic activity has been related to the seismic release of thermally-induced or regional stress along fault zones within active geothermal areas. Although microseismicity occurs in non-geothermal environments as well, the correlation of local microseismic activity with other geological, geochemical, and geophysical studies in a known geothermal area, may help to delineate the areas of highest thermally-induced stress and thus the areas of maximum geothermal potential.

THE NATURE OF MICROSEISMIC BACKGROUND NOISE

The normal microseismic background noise of the earth has been referred to as "microseisms" or "earth noise". The average earth noise peaks near seven seconds and has been conveniently divided into three bands (Iyer, 1964). These include the "long-period microseisms" with frequencies of less than 0.1 Hz, the "storm microseisms" with frequencies between 0.1 Hz and 0.5 Hz, and the "short-period microseisms" with frequencies greater than 0.5 Hz. Because the groundnoise surveys on the Island of Hawaii were restricted to frequencies of 0.5 Hz and greater, only the short-period microseisms will be discussed here.

The origin of short-period microseisms has been attributed to (Iyer, 1964):

1) Air-coupled atmospheric vibrations,
2) Volcanic tremors,
3) Wave action and turbulence in bodies of water,
4) Wind noise,
5) Cultural noise, and
6) Residual earth noise.

Another proposed source of microseismic noise in this frequency range was introduced in geothermal exploration methods. This source is related to pressure fluctuations or fluid movements within geothermal reservoirs.

A strong correlation between short-period microseismic activity and micro-barometric pressure fluctuations has been observed by Walsh (1955) and Saha (1962). Some of these variations in intensity have been correlated with storm fronts and thunderstorm activity. Akamatu (1961) has found dominant frequency peaks which depended on the place of measurement. These have been related to variations in geology at the measurement sites for atmospheric coupling to the ground can be strongly effected by the thickness of near surface alluvial cover.

Short-period microseisms have also been related to active volcanoes. Spasmotic tremor and harmonic tremor are discussed by Eaton and Murata (1962). These microseisms are observed at Kilauea Volcano, Hawaii, before and during eruptions. The spasmotic tremors are thought to result from the movement of lava in the conduits under Kilauea before surface eruptions, and harmonic tremors have been caused by the extrusion of lava through surface fissures.

Atmospheric turbulence and wave action in bodies of water are another source of microseismic activity. Although most of the wave action in open water is long period, the breaking of waves on shore and turbulence in shallow bodies of water can generate high frequency microseisms (Press and Ewing, 1952; Longuet-Higgins, 1952; Van Straten, 1952). Wave action is probably the most significant source of groundnoise on the Island of Hawaii in the absence of volcanic activity.

High-frequency seismic energy is also generated by the action of wind on trees and structures. The amplitude spectrum of the source depends upon the strength of the wind and the frequency of oscillation of these obstructions. Observations by Frontti and others (1962) indicate that the frequencies of microseisms caused by the wind can range from

as low as 1 Hz to 25 Hz or greater.

Cultural activity inputs much seismic energy into the earth. This source includes road traffic, aircraft, trains, and town and industrial noise. A wide range of frequencies can result. A study that was carried out in France by Rocard (1962) gives an indication of the attenuation of industrial noise with distance:

> A stone quarry using powerful stonebreaking equipment, located 2 km from a good seismic station mounted on granite, will cause a slightly perceptible trembling effect at 3 or 4 Hz; when the quarry stops operation (Sundays, or during the night), this noise disappears ... A station 2 km away from a mine records the various movements and disturbances of equipment; 5 or 6 km away, nothing except explosive firings is perceptible.

A study by Griffin (1963) indicated that high intensity cultural noise from large cities is significant for about 50 km from the source.

Variations in microseismic noise levels in various geologic environments have also received some attention. Two seismic noise profiles were run by United Electro Dynamics in 1961 and 1962 (Griffin, 1963); one was located in California and the other in Washington. This study indicated a nearly exponential decrease in the 0.25 to 2.22 Hz frequency range with increasing distance from the ocean. Variations in noise levels due to changes in geology were evident for consistently high noise levels were observed on Tertiary sediments.

Studies of "long-period microseisms" and "storm microseisms" have indicated that these earth noises tend to propagate as Rayleigh-type waves. Frantti (1962) points out that this assumption may not be correct for the higher frequency microseisms, for a large percentage of these seismic disturbances probably propagate as body waves.

An analogy has been made between the seismic and electric

properties of a material (Anstey, 1974):

> Just as a material which conducts electrical energy has an impedence, so a material which propagates acoustical energy has an acoustic impedence. Just as a fixed electrical voltage produces a large current in a material of low electrical impedence, so a fixed acoustic pressure produces a large particle velocity in a material of low acoustic impedence (such as Quaternary clay) and a small particle velocity in a material of high acoustic impedence (such as granite).

Because seismic speeds vary according to elastic parameters, groundnoise intensities also vary with differing rock elasticities.

Another important parameter which must be considered in groundnoise measurements is the attenuation of seismic energy with frequency. The loss mechanisms which cause this include the relative motion between a rock matrix and contained pore fluids, and sliding at points of contact in granular rocks. Although very little information is available on seismic wave attenuation at low frequencies, the results of high frequency experiments indicate that seismic attenuation (White, 1965):

1) is directly proportional to frequency,
2) is reduced with high confining pressures, and
3) is increased with the addition of a small amount of moisture to an originally dry rock.

All of these attenuation characteristics must be considered when analyzing microseismic groundnoise variations.

PRELIMINARY FIELD STUDY

A microseismic-groundnoise survey was carried out in the Gregory Rift Valley in Kenya as an early part of this study. The area of interest was adjacent to Lake Naivasha and centered about Orgaria Peak (Figure 3). Numerous thermal manifestations, hydrothermally altered rocks, and recent igneous activity, made this region a prime geothermal target. Local rock types include Quaternary pyroclastics and lacustrine deposits, Upper Pleistocene rhyolites, comendites, and trachytes, and recent obsidians (Thompson and Dodson, 1963). Because the United Nations and the East African Power and Lighting Company were sponsoring extensive geological, geophysical, and geochemical programs with the ultimate hope of developing geothermal power, this area seemed to be ideal for testing the microseismic-groundnoise technique.

Vertical groundnoise intensities were observed at 45 recording locations for periods of 2 hours. Because previous geothermal surveys in other areas had indicated anomalous groundnoise intensities in the 2 Hz region of the spectrum, the recording and playback system were designed to monitor variations in 2 Hz groundnoise intensities. This survey indicated large variations in the amplitudes of 2 Hz microseisms, so only the minimum ground velocities that occurred at each recording location were compared. The relative amplitudes (in db) at each of

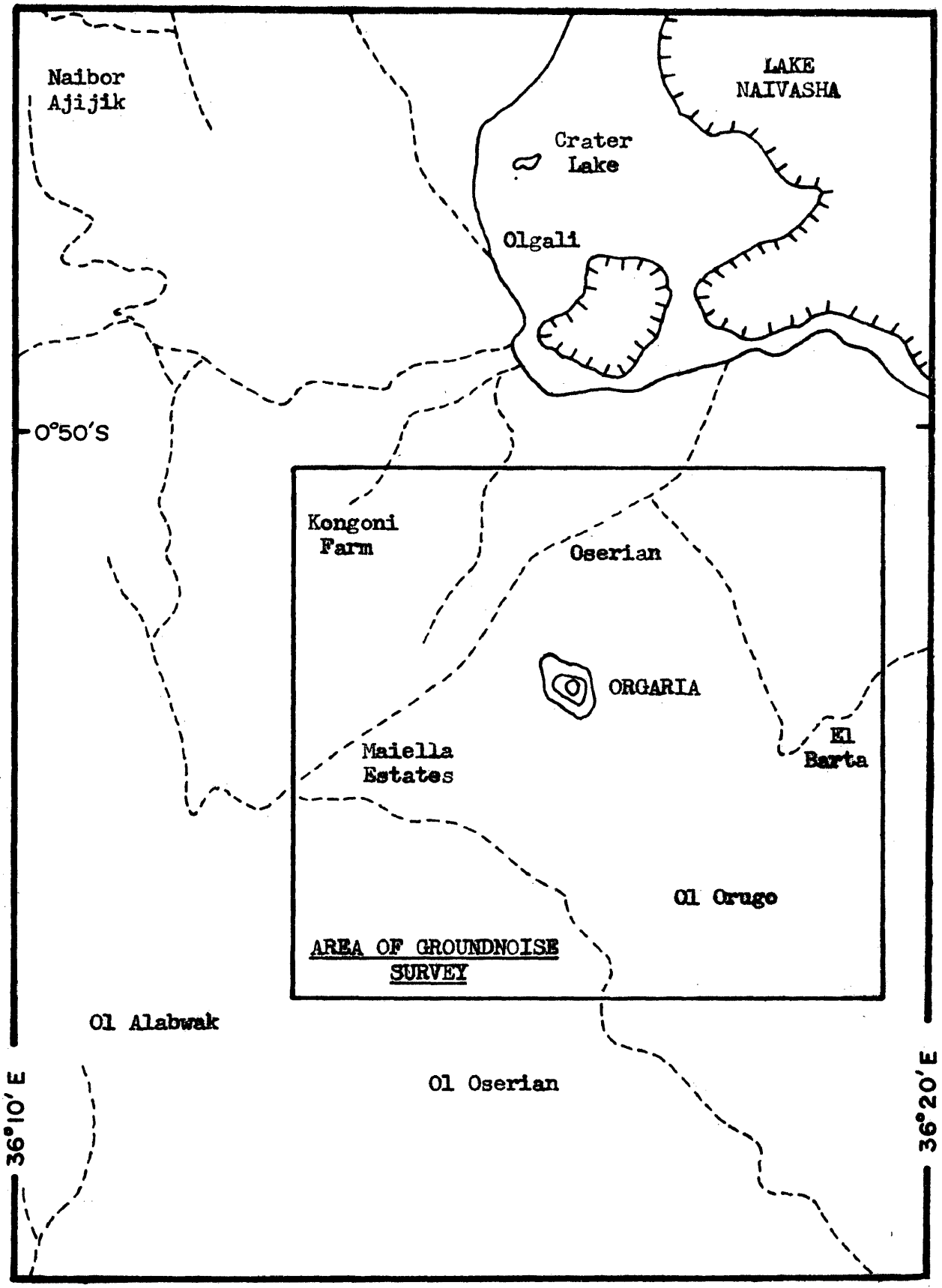


FIGURE 3. Location Map Of The Microseismic Groundnoise Survey Area In The Gregory Rift Valley, Kenya.

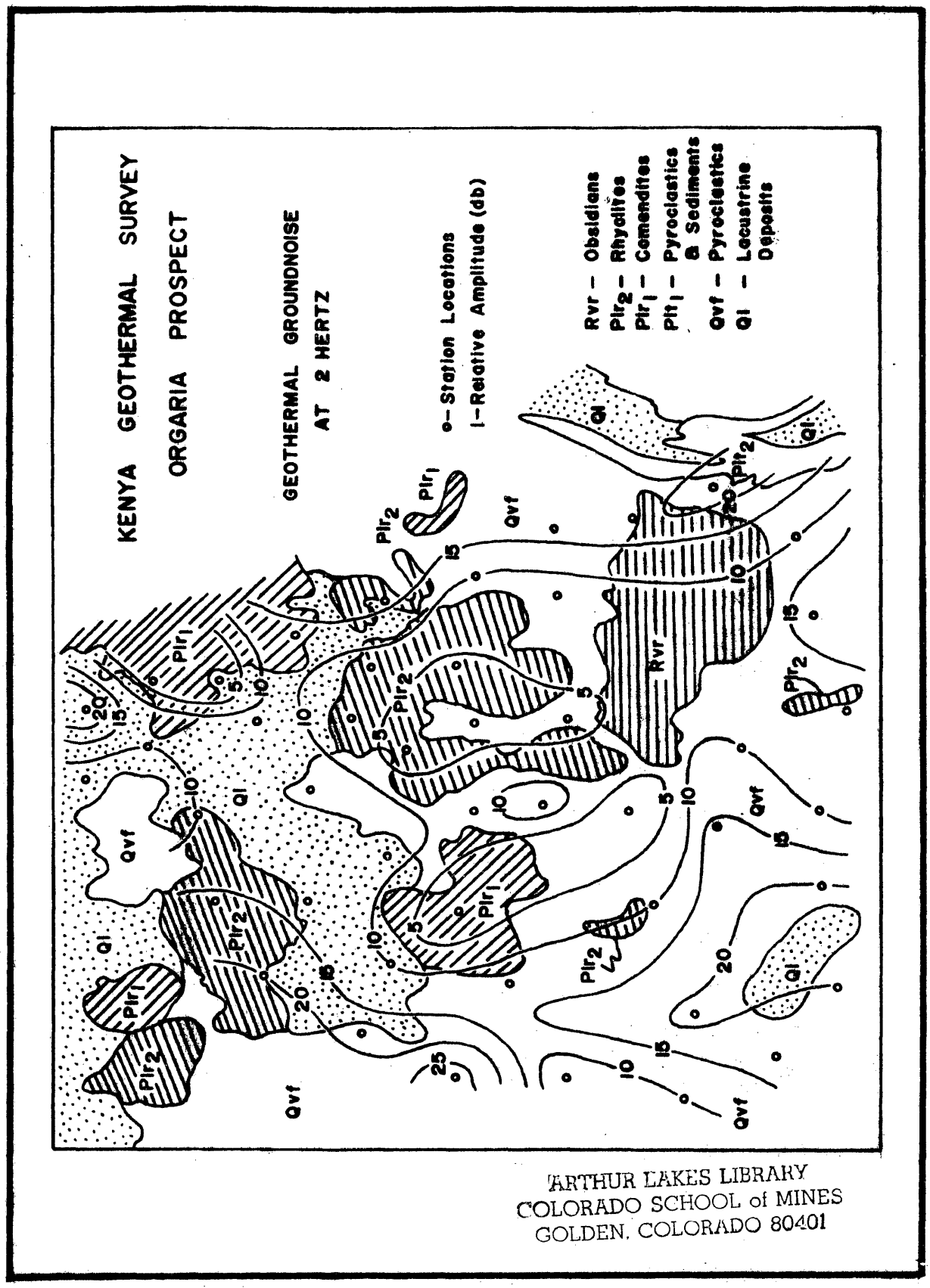


FIGURE 4. Relative Groundnoise Amplitudes (2 Hz) At The Orgaria Prospect, Gregory Rift Valley, Kenya (Geology taken from Thompson and Dodson, 1963).

these locations is contoured in Figure 4. Maximum groundnoise variations of 29 db were observed among the 45 recording stations. Unfortunately, the areas of highest surface thermal manifestations exhibited the lowest particle velocities. Furthermore, the survey indicated a high correlation between groundnoise amplitude and local geology. The recording stations in the vicinity of outcropping rhyolites, comendites, and obsidians measured the lowest vertical ground velocities, whereas the recording stations located in pyroclastics and lacustrine deposits measured the greatest ground motion.

The results of this survey indicated the influence of lithologic variations on microseismic-groundnoise measurements. If the groundnoise method is to be used as originally intended (i.e. for locating areas of maximum ground motion resulting from dynamic geothermal reservoirs), then groundnoise variations resulting from "lithologic noise" must be kept to a minimum. Because most areas of geothermal interest tend to have complex geologic structure and lithology, the exploration significance of this method is in question.

An attempt to resolve this question and to test the original geothermal-groundnoise hypothesis, resulted in extensive groundnoise surveys in two areas which appear to have great geothermal potential. These surveys were carried out at Kilauea Volcano, and the East Rift of Kilauea Volcano, on the Island of Hawaii. These two locations were chosen because, of all possible geothermal prospects, groundnoise variations due to "lithologic noise" would be a minimum.

MICROSEISMIC GROUNDNOISE SURVEYS ON THE ISLAND OF HAWAII

Two areas on the Island of Hawaii were selected for the microseismic groundnoise surveys. These include the summit of Kilauea Volcano and part of the East Rift Zone of Kilauea. A location map of the survey areas is shown in Figure 5. These areas were selected for this study because:

1. recent volcanic activity indicates a near surface source of potential geothermal energy.
2. the existence of a shallow magma reservoir at Kilauea Volcano is indicated by surface deformation studies (Kinoshita and others, 1974), and
3. the relatively homogeneous geologic environments were more suitable for an evaluation of the microseismic groundnoise method than most other geothermal areas.

Geology Of The Survey Areas

Kilauea Volcano lies on the main volcanic rift that passes through Kohala and Mauna Kea and probably originated when faulting on the side of Mauna Loa intersected magma within this rift. Kilauea produces lava flows with volcanic ash. The flows range from a few inches to 15 feet in thickness and vary in length from a few feet to several miles. At least

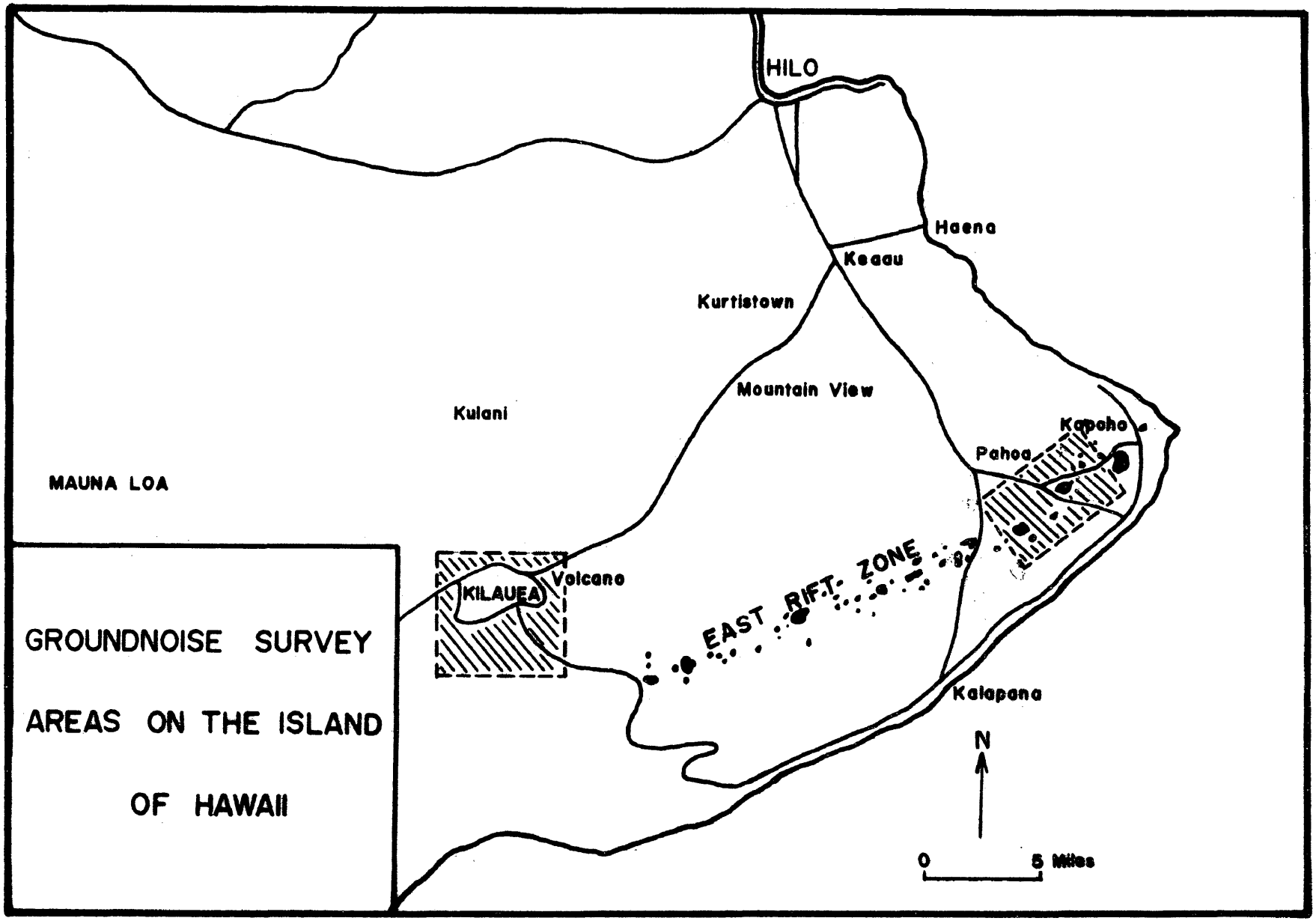


FIGURE 5. Location Map Of The Microseismic Groundnoise Survey Areas On The Island Of Hawaii.

eleven different ash deposits are recognized in the summit area (Stearns, 1966).

Two zones of cones, fissures, and dikes transect Kilauea Volcano. One of these zones, the southwest rift, extends from Kilauea into the Kau District and is characterized by numerous surface fissures. The other zone "... trends southeast from Kilauea for four miles and then turns N. 65° E. and extends to Cape Kumukahi (East Point) in the Puna District" (Stearns, 1966). The "East Rift Zone" is characterized by volcanic cones, fissures, and pit craters.

The volcanic flows and ash deposits of Kilauea Volcano are divided into two geologic units. These include the old Hilani volcanic series which are capped by the Pahola Ash. The Puna volcanic series of late Pleistocene and recent time rest upon the older unit.

Approximately 41 eruptions have occurred at Kilauea Volcano and in the East Rift Zone since 1750. Duration of these eruptions range from a few hours to over a year. Seismic studies have indicated a source of magma activity at about 60 km below Kilauea. The magma works its way to the surface through a system of dikes. The existence of a near surface reservoir is indicated by inflation of the summit area prior to surface eruptions and noticable collapse after the extrusion of lava in a summit or flank eruption. Studies of the Kapoho eruption of 1960 indicate horizontal magma movement from the summit reservoir to the Kapoho vent 30 miles to the east. A generalized diagram of the East Rift Zone of Kilauea Volcano is illustrated in Figure 6 and a generalized geologic map of the area is shown in Figure 7.

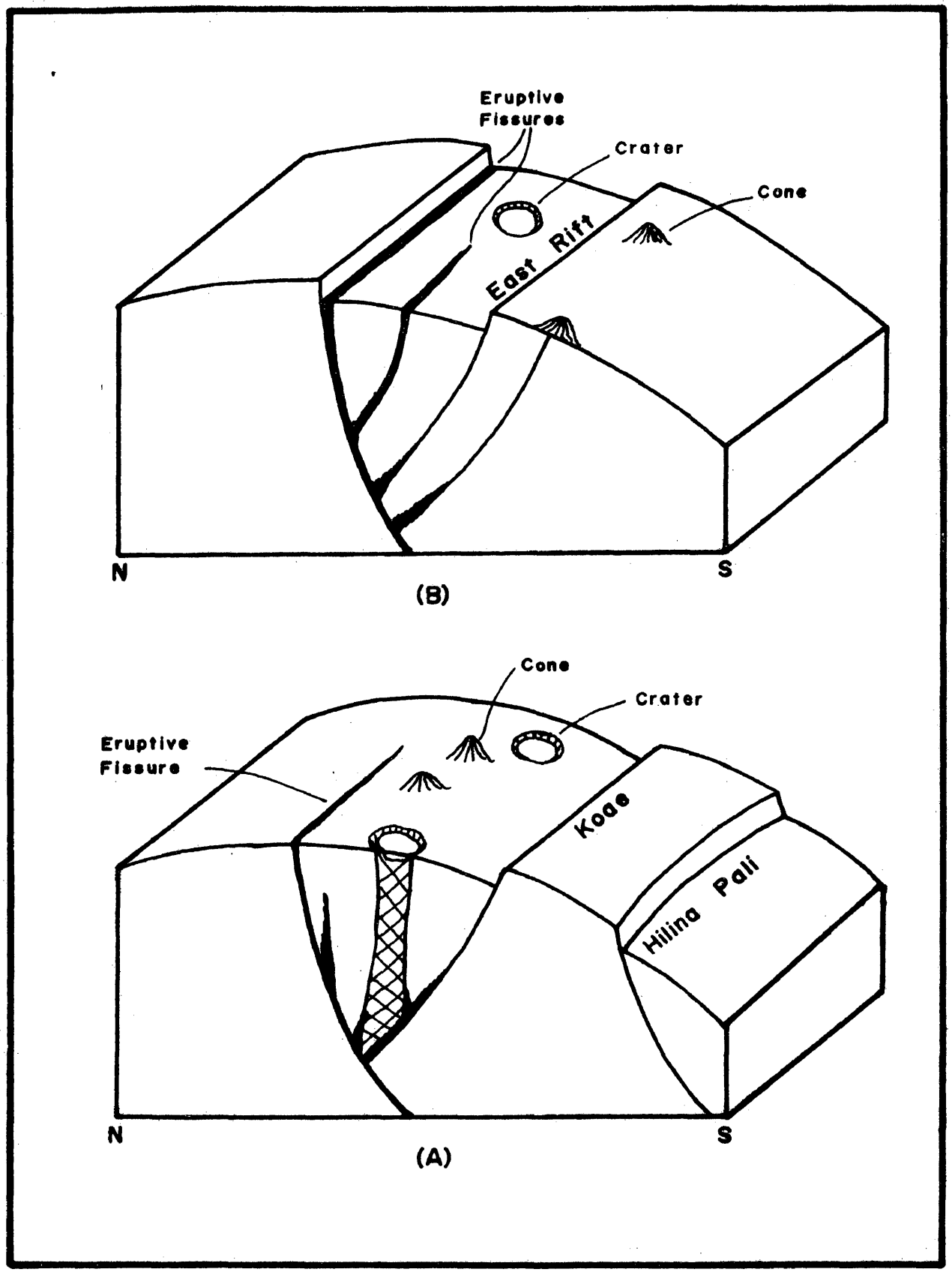


FIGURE 6. Generalized Geologic Structure of: (A) Kilauea Volcano, (B) East Rift Zone.

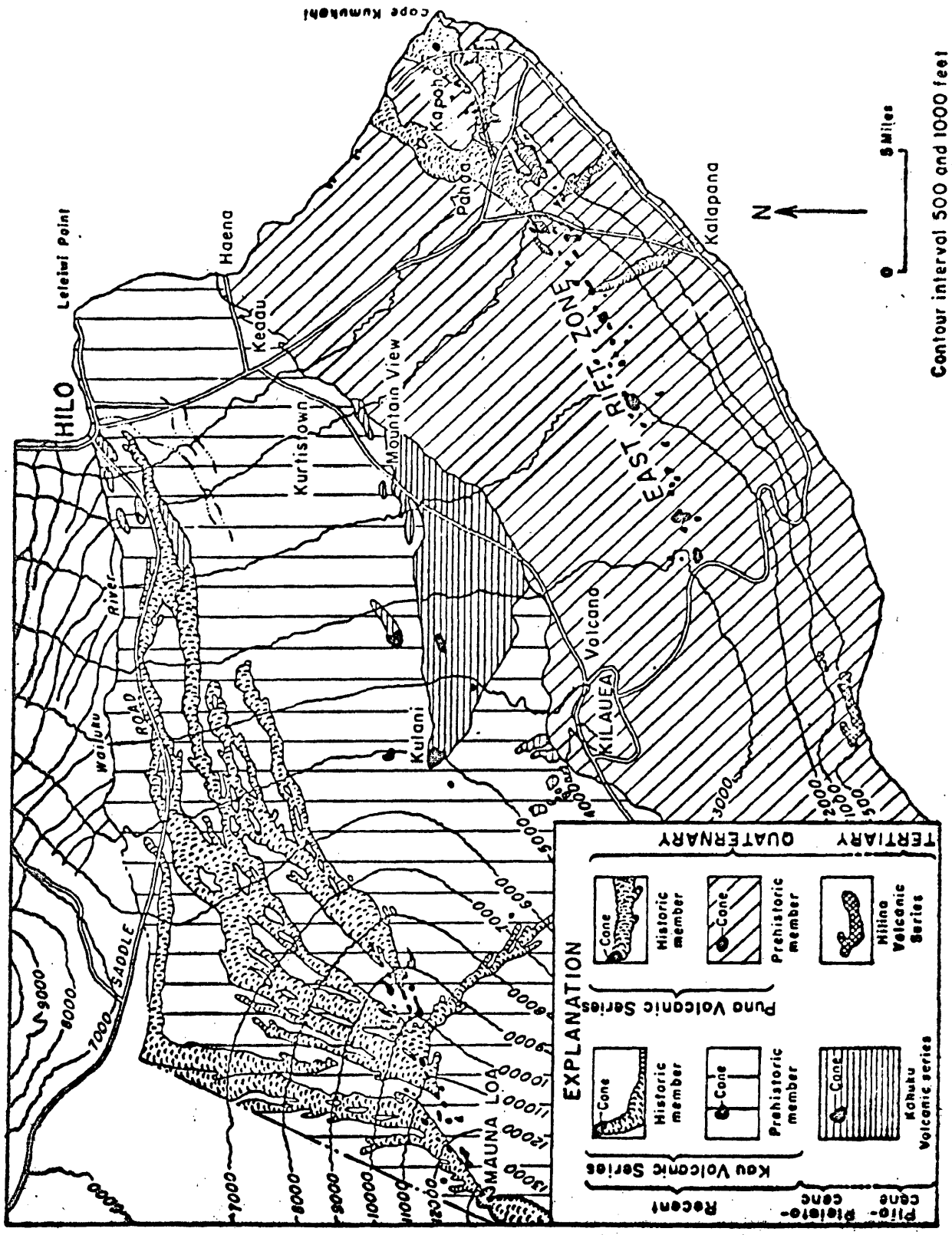


FIGURE 7. Generalized geologic map of the East Rift and surrounding areas (after Davis and Yamanaga, 1968).

MICROSEISMIC GROUNDNOISE DATA COLLECTION AND REDUCTION

Equipment And Field Recording Techniques

Recording Equipment

The three microseismic groundnoise recording systems that were used in this survey are illustrated in Figure 8. One system was used as a base station to monitor diurnal variations in background noise levels while the other two stations measured local variations in seismic groundnoise intensity. Each recording system consisted of a seismometer, a high-gain amplifier, a voltage controlled oscillator (VCO), a stable oscillator, and a magnetic tape recorder for data storage.

The outputs of the seismometers were input to adjustable-gain, solid-state amplifiers. The solid-state amplifiers were used to increase the signal levels of the seismometers to satisfactory input levels for the VCO's. The output frequencies of the VCO's were proportional to the input signal levels. Stable oscillators were used to monitor fluctuations in recording speed. The microseismic groundnoise signals and the tape-speed compensation signals were recorded on magnetic tape recorders for later analysis.

Geotechnical Corporation Model 18300 seismometers were used as the groundnoise transducers. The main-coil generator constants for these seismometers were 620 Volt-sec/meter. Although these seismometers have continuously-adjustable natural undamped frequencies of 0.75 Hz to 1.1 Hz, they were preset and maintained at 1.0 Hz, vertical operation. Internal

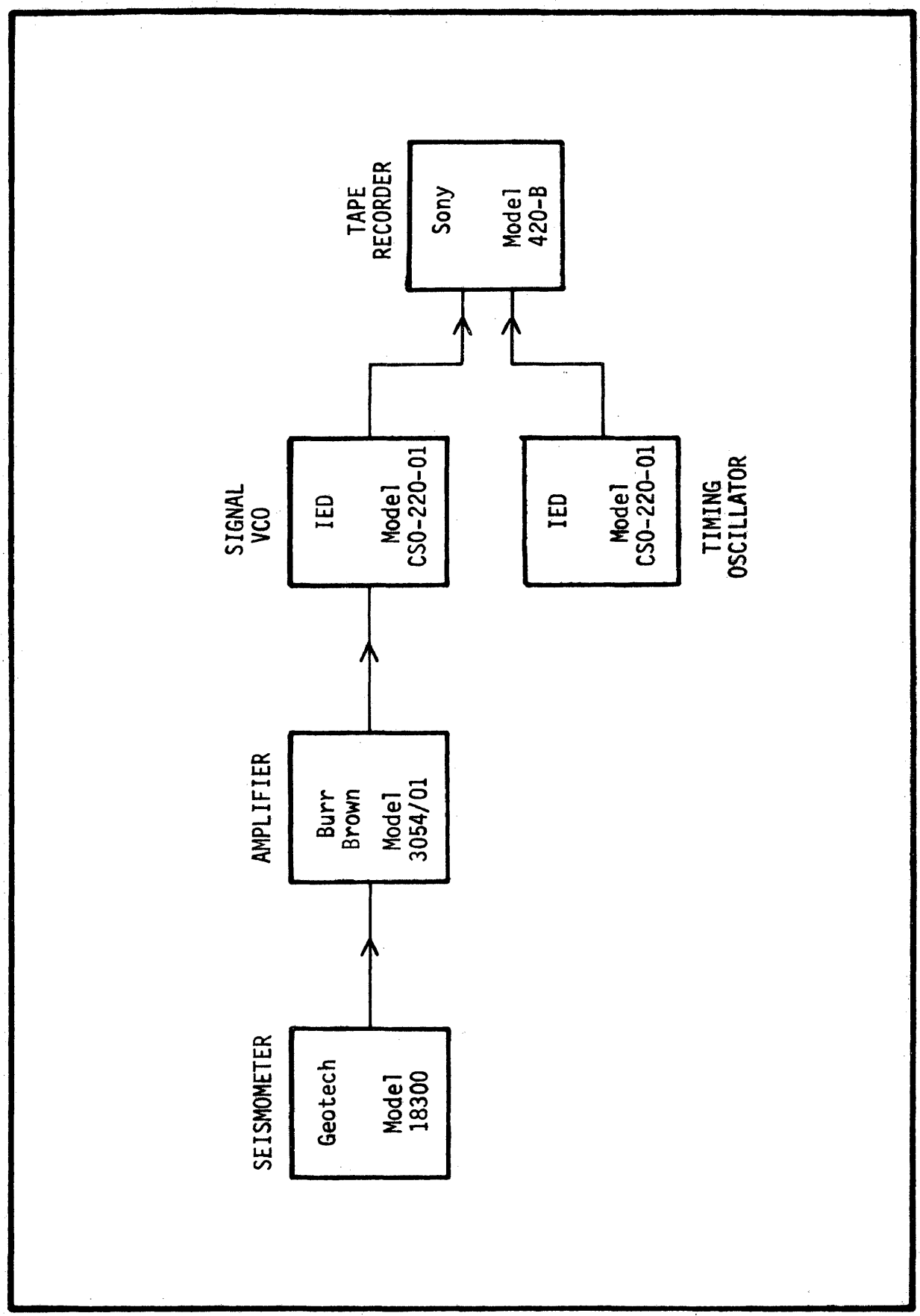


FIGURE 8. Microseismic Groundnoise Recording System.

calibration coils with 0.1975 Volt-sec/meter coil constants were used to determine seismometer response characteristics. These short-period seismometers were used because they are portable, relatively light weight, and are constructed for rugged field operations.

Two-stage, non-inverting, amplifiers were used to increase the output voltage of the seismometers. Burr Brown Research Corporation Model 3054/01 operational amplifiers were selected because of their low-noise characteristics. Each stage of the amplifiers had adjustable gains of 0 to 60db, in 10 db gain steps. These step-wise adjustable gains resulted in a maximum total voltage gain of 120 db. The amplitude response of these amplifiers is flat to several hundred Hertz.

IRIG proportional bandwidth IED Model CSO-220-01 VCO's were driven by the amplifier outputs. Input sensitivities of $\pm$2.5 volts resulted in full bandwidth deviations of $\pm$7.5 percent. These VCO's were selected because of their linearity, low subcarrier distortion, and time and temperature stability. They are linear to 0.25 percent of bandwidth, their subcarrier distortion is less than 1 percent at any frequency in the bond, and after a fifteen minute warmup period, their center frequencies are stable to within 1 percent of bandwidth over a temperature range of -20°C to +85°C. These VCO's are also very insensitive to changes in battery voltage, less than $\pm$0.5 percent of bandwidth for a $\pm$10 percent change in operating voltage, and have low power requirements, 8 ma of current at 28 Vdc. All of these characteristics were very important for prolonged field recording sessions.

The VCO outputs were recorded on Sony Model 420-B magnetic tape recorders at a speed of 19 cm/sec. Speed fluctuations which result in

wow and flutter were very low, less than 0.5 percent. Portability and low power requirements made these tape recorders ideal for this application.

Two examples of the seismic noise that was recorded are illustrated in Figure 9. One of these records, station 18, shows the dominant low frequency microseisms that were seen at most of the recording locations. The record from station 34 shows much higher frequencies near the eruption at Mauna Ulu.

Playback Equipment

The playback equipment for the microseismic groundnoise system is illustrated in Figure 10. The field-recorded tapes were played back through the same Sony tape recorders that were used to record the data. To decrease the playback time, however, the recorders were run at eight times the recorded speed. This increase in playback speed resulted in time-compressed data records.

Two Sonex distriminators, Models S-25, were driven by the tape recorder outputs. One of the discriminators was used to demodulate the FM microseismic groundnoise signal while the other was used to demodulate the FM tape-speed conpensation signal. The outputs of the two discriminators were summed in a differential amplifier to provide a tape-speed error improvement of approximately 30 db for the speed variations encountered.

Sonex discriminators were selected because of their high amplitude modulation rejection, their high attenuation of adjacent IRIG channels, and their low output-noise levels. The output-zero stability

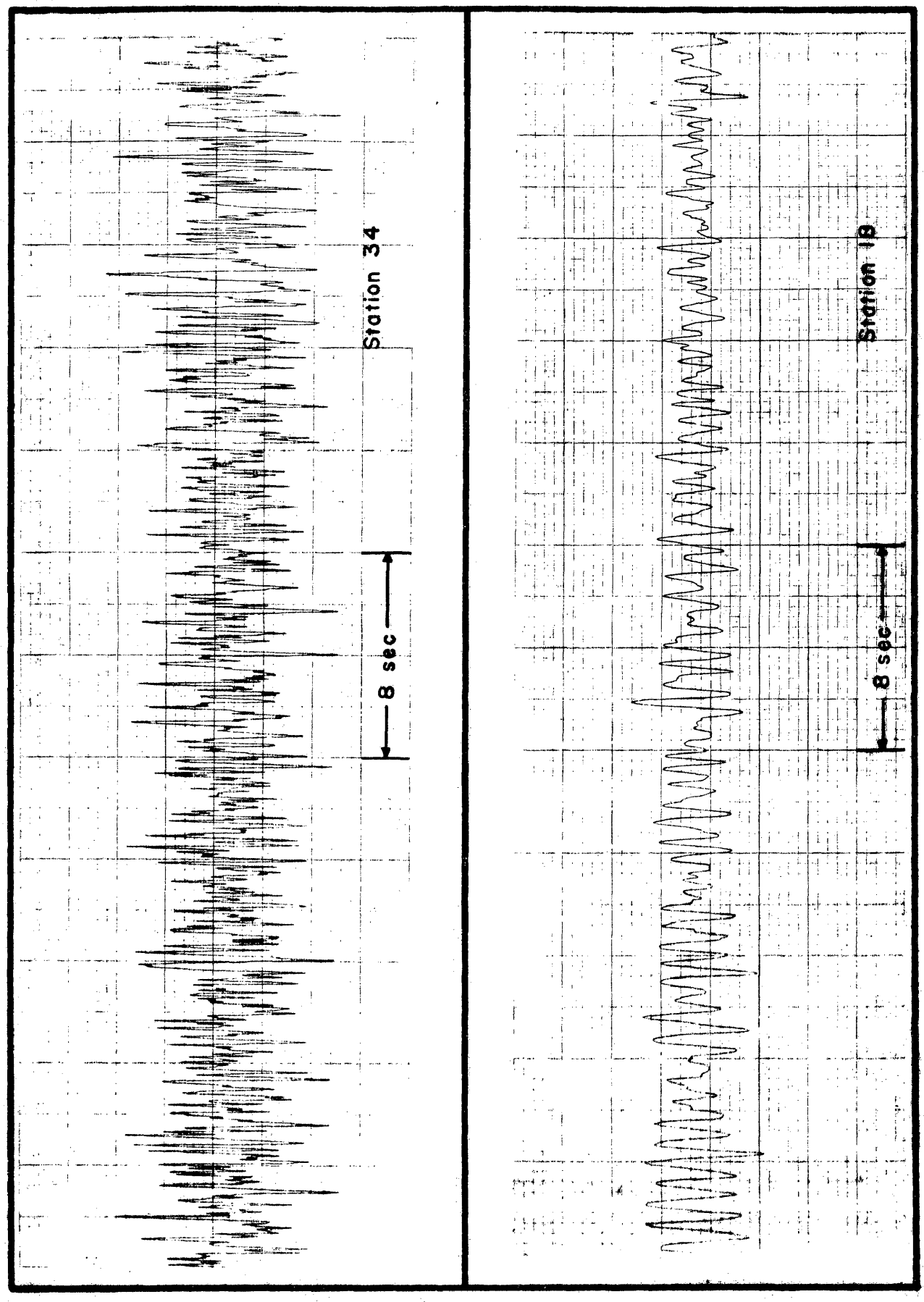


FIGURE 9. Seismic Noise Records.

and the deviation-sensitivity stability of these discriminators was less than ±0.1% in a 15-hour period after a 10-minute warmup. The discriminators also have an 18 db/octave, linear phase, low-pass Bessel filter in the output stage. Because the groundnoise system was designed for low frequency analysis, 0 to 20 Hz, the low-pass output filters of the discriminators had a cutoff frequency of 160 Hz, or eight time the highest frequency of interest because of frequency compression in the playback equipment. The velocity response of the recording and playback equipment is illustrated in Figure 11.

Analysis Equipment

The tape-speed compensated, amplitude-modulated groundnoise signals were the inputs to several bandpass filters for frequency analysis. Because the frequencies of interest were 0.5 to 10 Hz, and because of frequency compression in the playback system, Burr-Brown, Model 5715-BP1P, with center frequencies of 4.0 Hz, 8.0 Hz, 16.0 Hz, 32.0 Hz, and 64.0 Hz were selected. This resulted in the non-frequency compressed amplitude response illustrated in Figure 12.

The bandpass-filtered, groundnoise signals were the inputs to precision, full-wave rectifiers. These rectifiers, which use two operational amplifiers in an absolute value circuit, are ideal for low-amplitude signals because the forward-conduction, turn-on voltage of the rectifier is decreased by a factor which is equivalent to the open-loop gain of the amplifier. The output of the full-wave rectifier was smoothed with a low-pass, active filter which provided a DC output voltage equal to the RMS value of the full-wave rectified, sine-wave input.

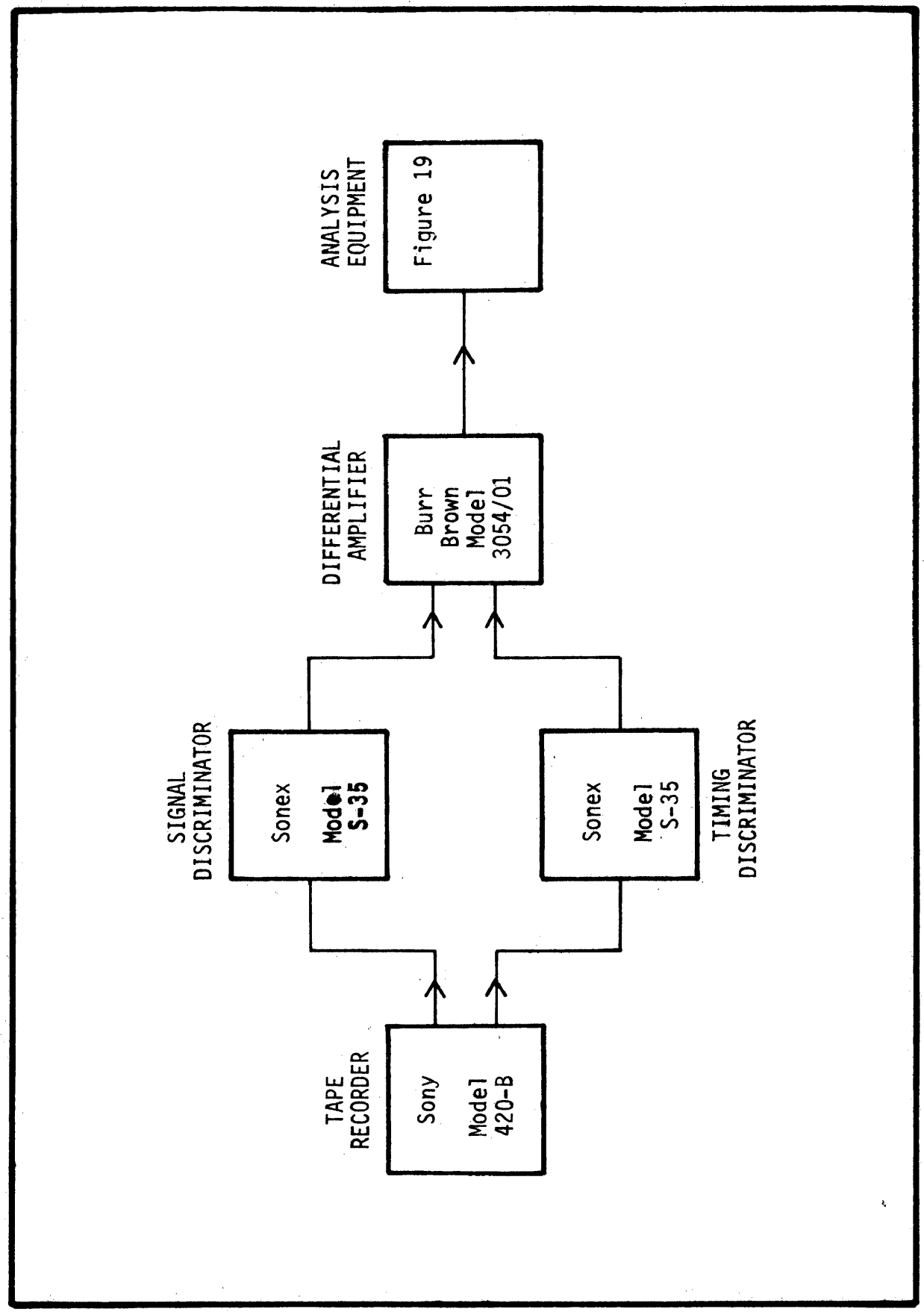


FIGURE 10. Microseismic Groundnoise Playback System

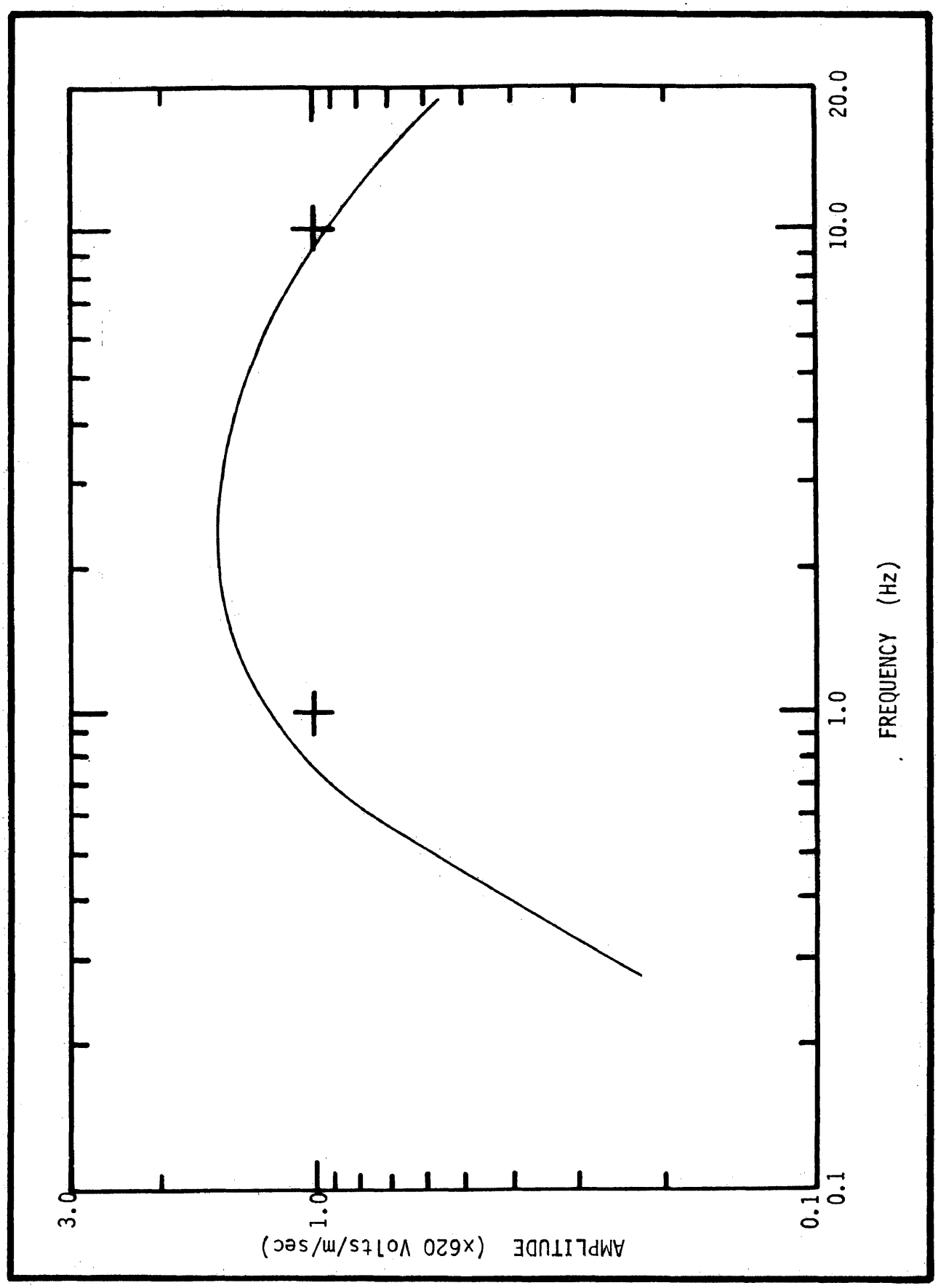


FIGURE 11. Velocity Response Of Recording And Playback Equipment.

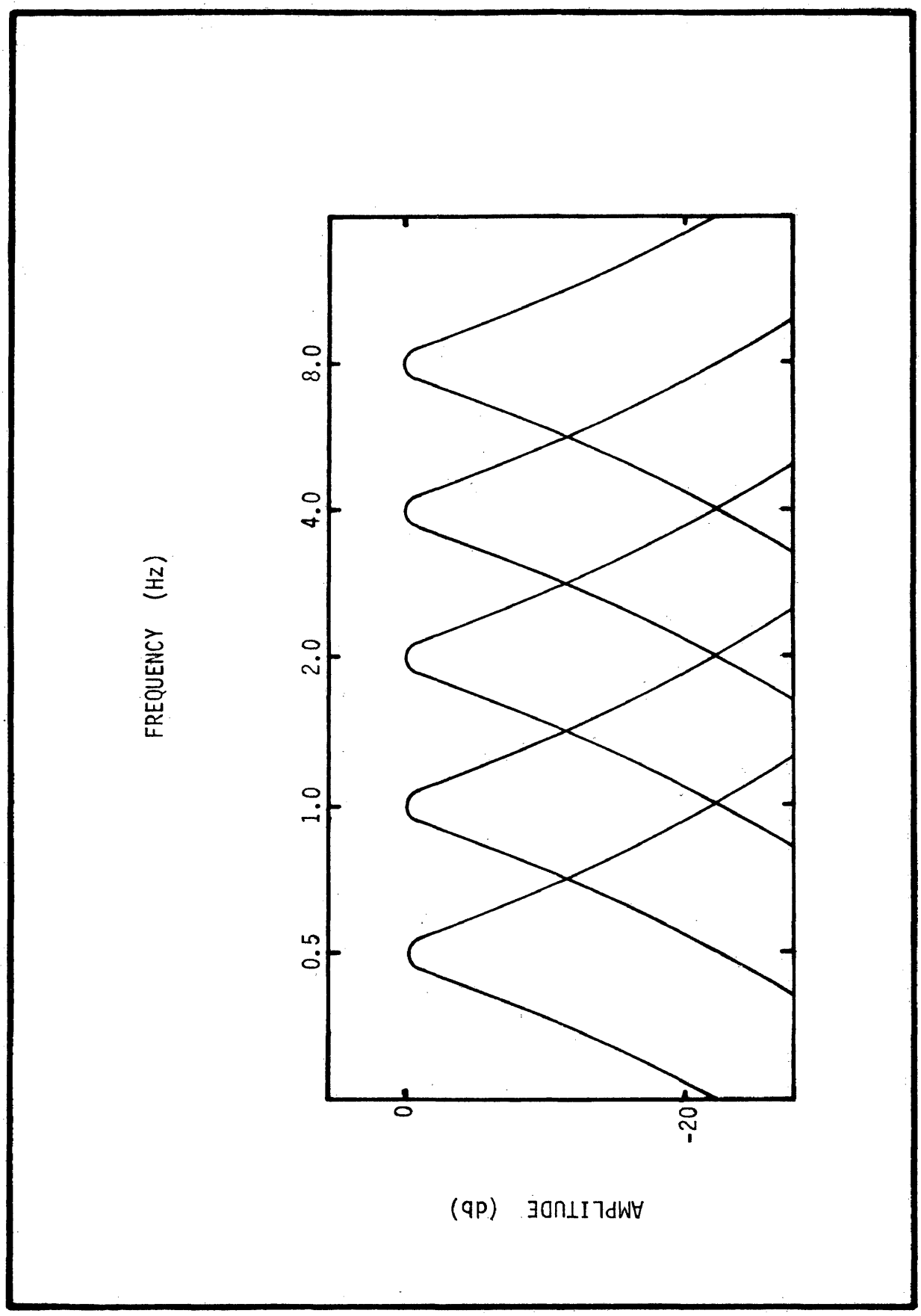


FIGURE 12. Band Pass Characteristics Of Analysis Equipment

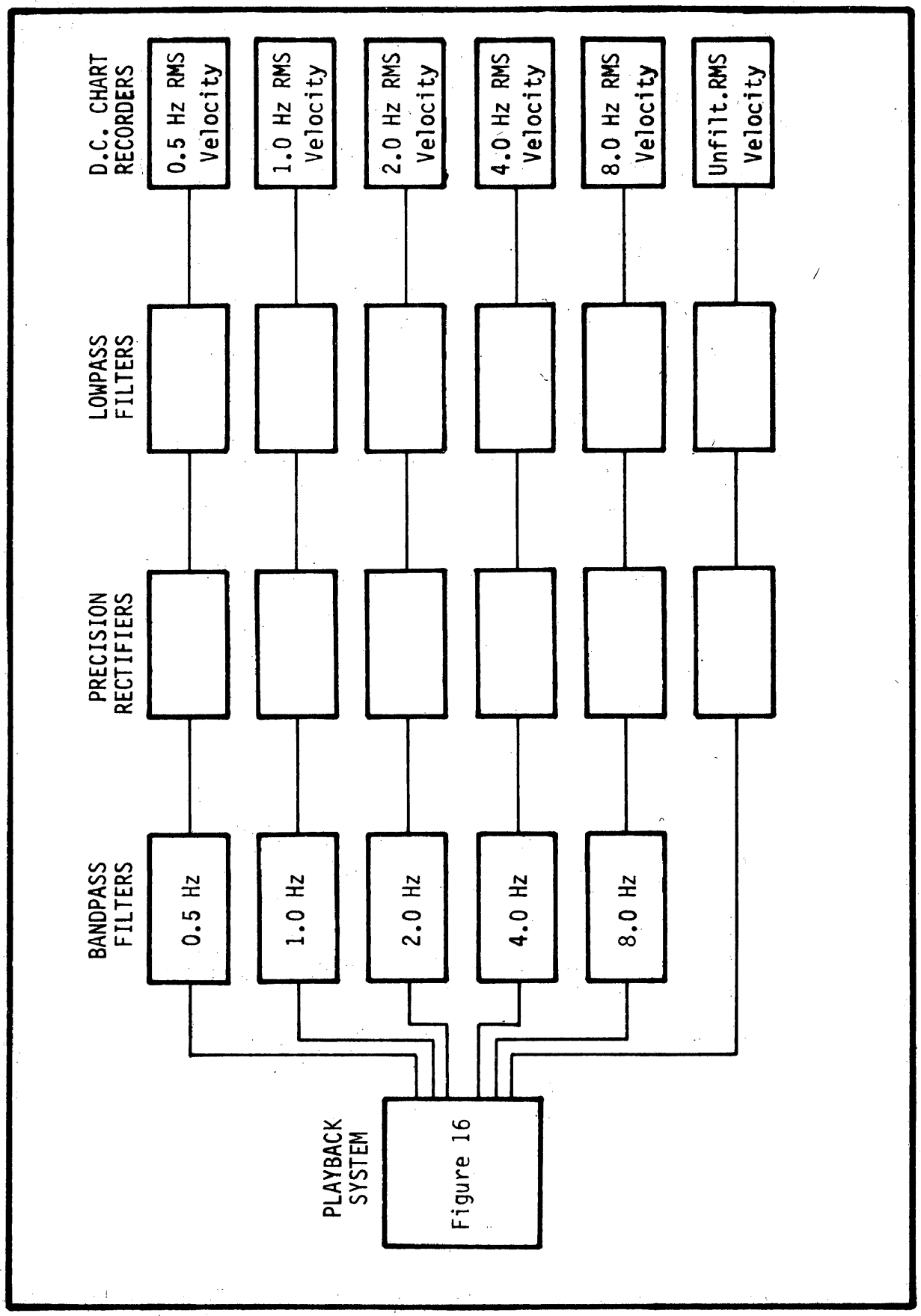


FIGURE 13. Groundnoise Analysis Equipment.

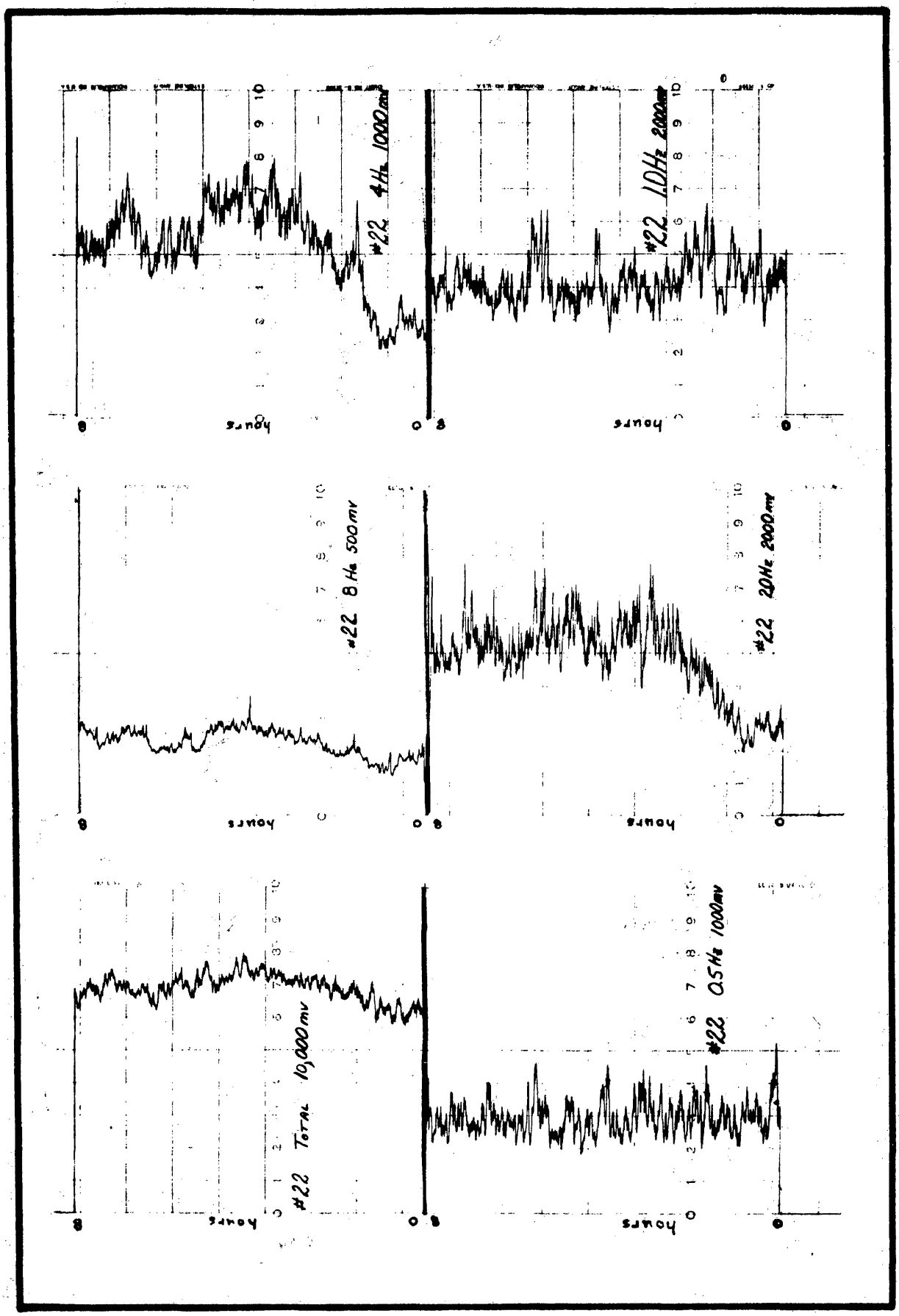


FIGURE 14. Time-Compressed, Eight Hour Record Of Seismic Noise From Station 22.

These DC voltages were recorded on Esterline Angus, Model T-171B, DC chart recorders. The groundnoise analysis equipment is illustrated in Figure 13.

The output of the groundnoise analyzer for station number 22 at the summit of Kilauea Volcano is illustrated in Figure 14. This figure includes time-compressed, eight-hour records of seismic noise for the 0.5 Hz to 8.0 Hz frequency range. The additional record, indicated as "total", is a non-bandpass filtered record of groundnoise activity at that station. Similar records were obtained for each of the stations occupied. This figure illustrates the variability in seismic noise amplitude over a period of eight hours. Some of these variations are a result of microseisms and teleseisms. Other variations are a result of cultural noise, the primary one being vehicles, and natural local disturbances caused by wind and rain. In the final groundnoise analysis, only the lowest levels of groundnoise activity were assigned to each recording station.

Field Procedure

The seismic groundnoise recording systems were used to study two active volcanic areas on the Island of Hawaii. While monitoring the first area of interest, the summit of Kilauea Volcano, one recording system was located at a base station and the other two recording systems were periodically moved from one location to another in order to monitor local variations in seismic groundnoise activity. The base station and each field recording station were operated during the night for a period of six to eight hours. A total of forty-two stations was recorded in

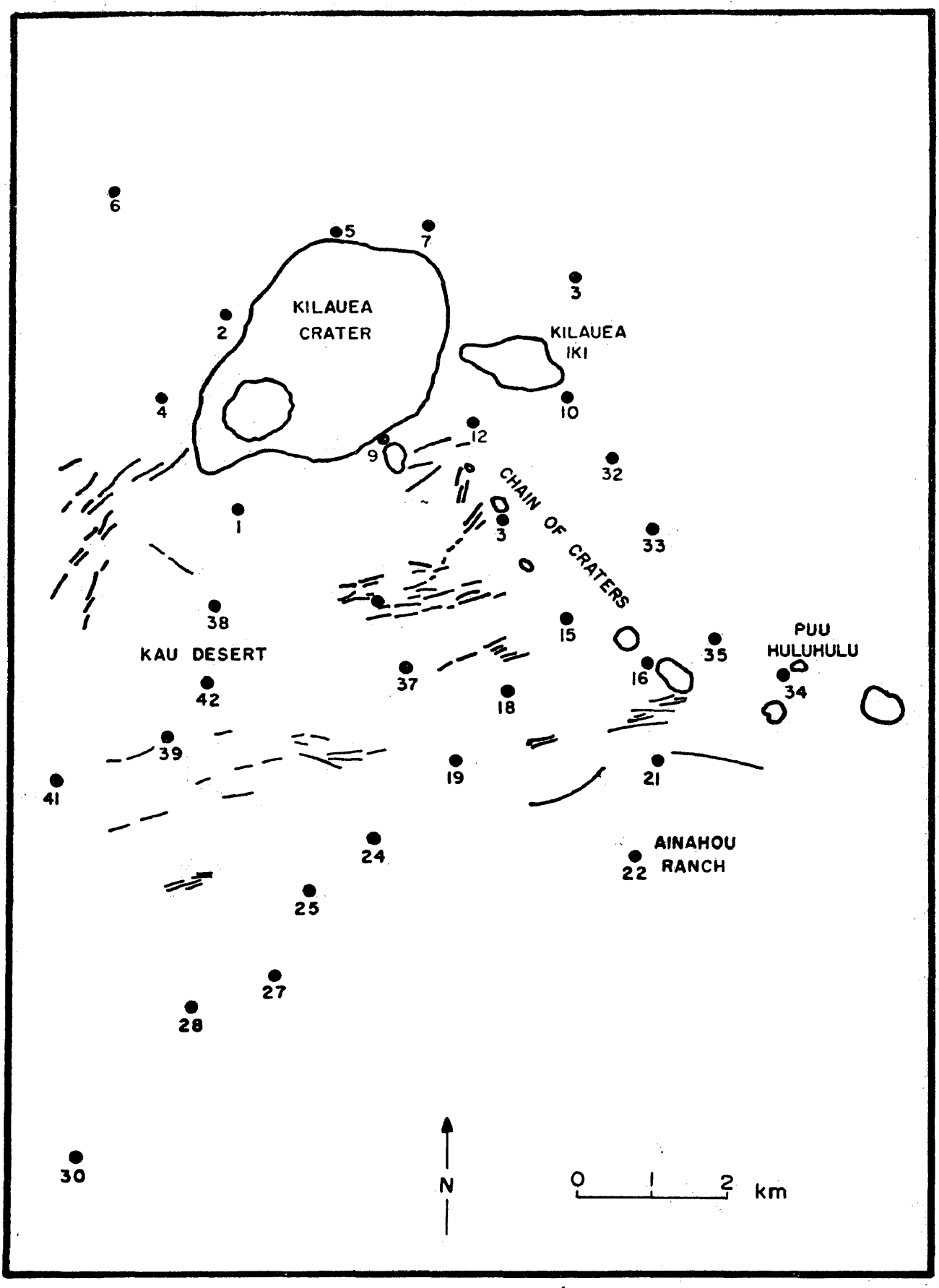


FIGURE 15. Groundnoise Recording Stations At The Summit Of Kileaua.

this area. Although the station locations were limited to accessible roads and trails within Kilauea Volcano National Park, a good areal density was obtained. Station locations are illustrated in Figure 15.

The procedure that was used to monitor groundnoise intensity in the other area of interest, the East Rift of Kilauea Volcano, was slightly different. Analysis of the Kilauea Volcano data indicated a small correlation between changes in base station groundnoise intensity and changes at either of the other two recording systems. Local variations in groundnoise intensity, which were a result of culture, wind, or rain, were not easily recognized at all recording stations. These analyses indicated that groundnoise measurements could be obtained in short periods of time if the observer was aware of local noise conditions and adjusted the survey locations to minimize unwanted natural and cultural noises. Because of this, the recording systems in the East Rift were operated for a period of 30 minutes to 1.5 hours. A total of forty-five locations was occupied in the East Rift Zone. Station locations are illustrated in Figure 16.

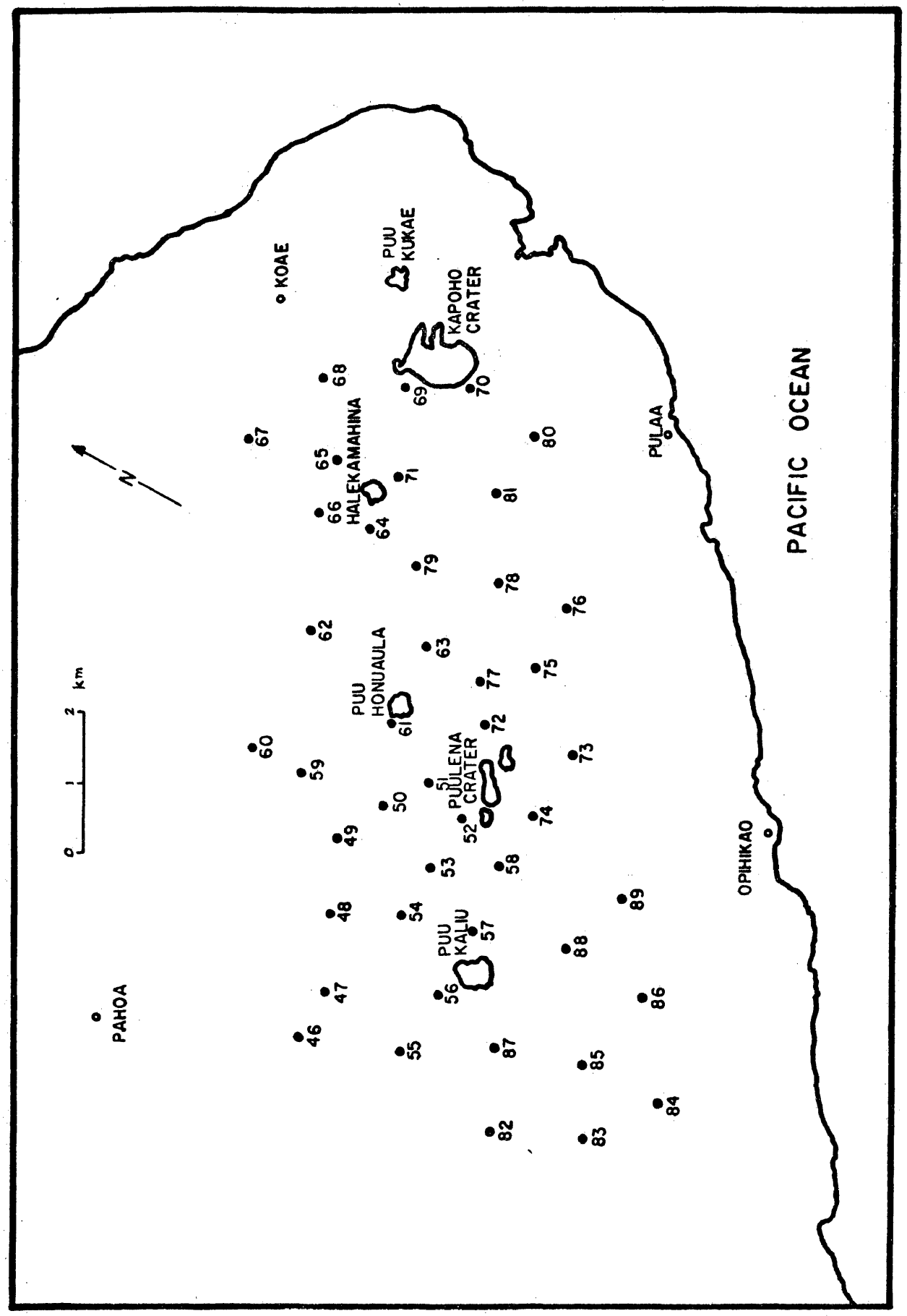


FIGURE 16. Groundnoise Recording Stations At The East Rift Zone.

Groundnoise Data From The Summit Of Kilauea Volcano

While conducting the microseismic groundnoise survey at Kilauea Volcano, a base station was operated to monitor daily variations in groundnoise activity. A summary of this vertical ground velocity information for eleven consecutive days is illustrated in Figures 17 through 19. Because the groundnoise variations for each record are typical of those illustrated in Figure 14, only the lowest levels of groundnoise for the entire recording period were assigned to each recording location.

Figure 17 shows the variations in the 0.5- and 1.0 Hz data records. The plot of 0.5 Hz ground velocity versus recording day shows a relatively constant minimum velocity of approximately 23 mμ/sec for the first five days. The amplitude then decreases to a minimum of 16 mμ/sec on the eighth day, increases to its normal level on the ninth and tenth days, and then decreases again. The ratio of maximum to minimum ground velocity for this period of time is 1.44 or about 3.1 db. Average ground velocity at 1.0 Hz is 16 mμ/sec, with a maximum deviation of about 6 db. Non-coincidence in ground-velocity variations from day to day is evident in this figure.

Figure 18 shows the variations in ground velocity for the 2.0- and 4.0 Hz data records. The average 2.0 Hz ground velocity is 14 mμ/sec. The maximum deviation for the eleven day period is approximately 6 db. The 4.0 Hz record has an average RMS velocity of 5 mμ/sec with maximum variations of about 10 db.

Figure 19 shows the daily variations in ground velocity for 8.0 Hz and the unfiltered record. The average 8.0 Hz velocity for the eleven

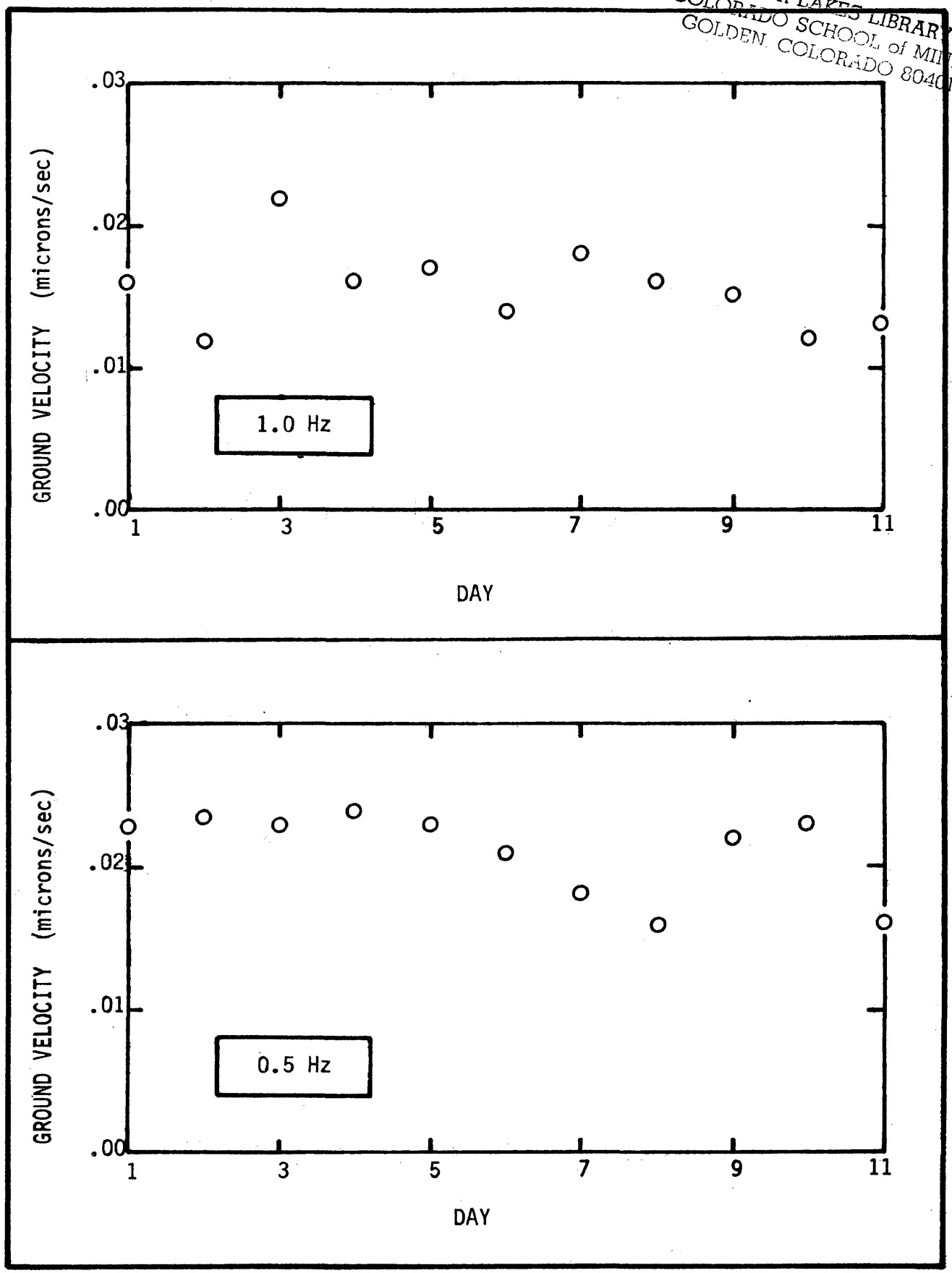


FIGURE 17. Base Station Variations In Ground Velocity At 0.5 Hz And 1.0 Hz.

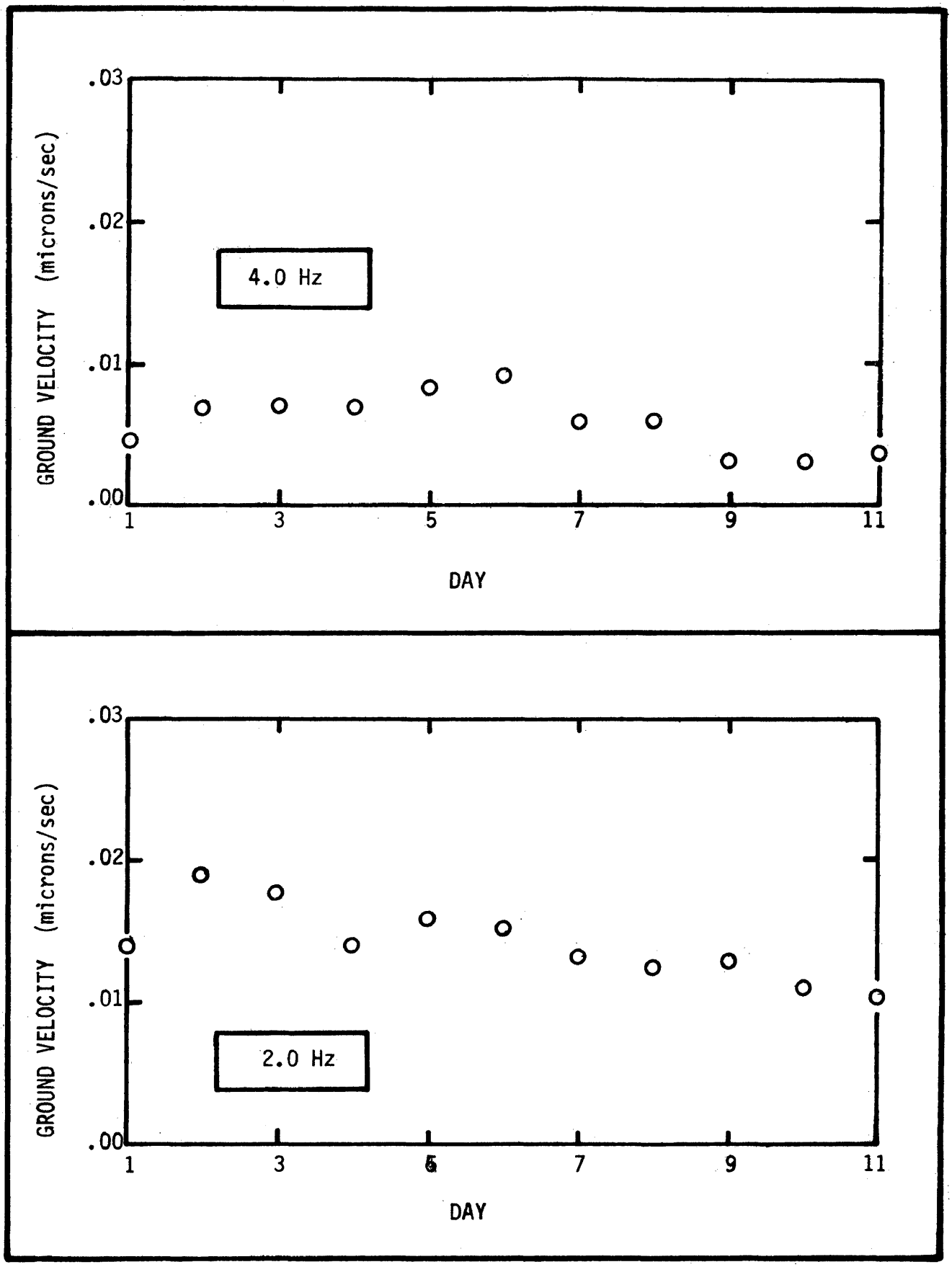


FIGURE 18. Base Station Variations In Ground Velocity At 2.0 Hz And 4.0 Hz.

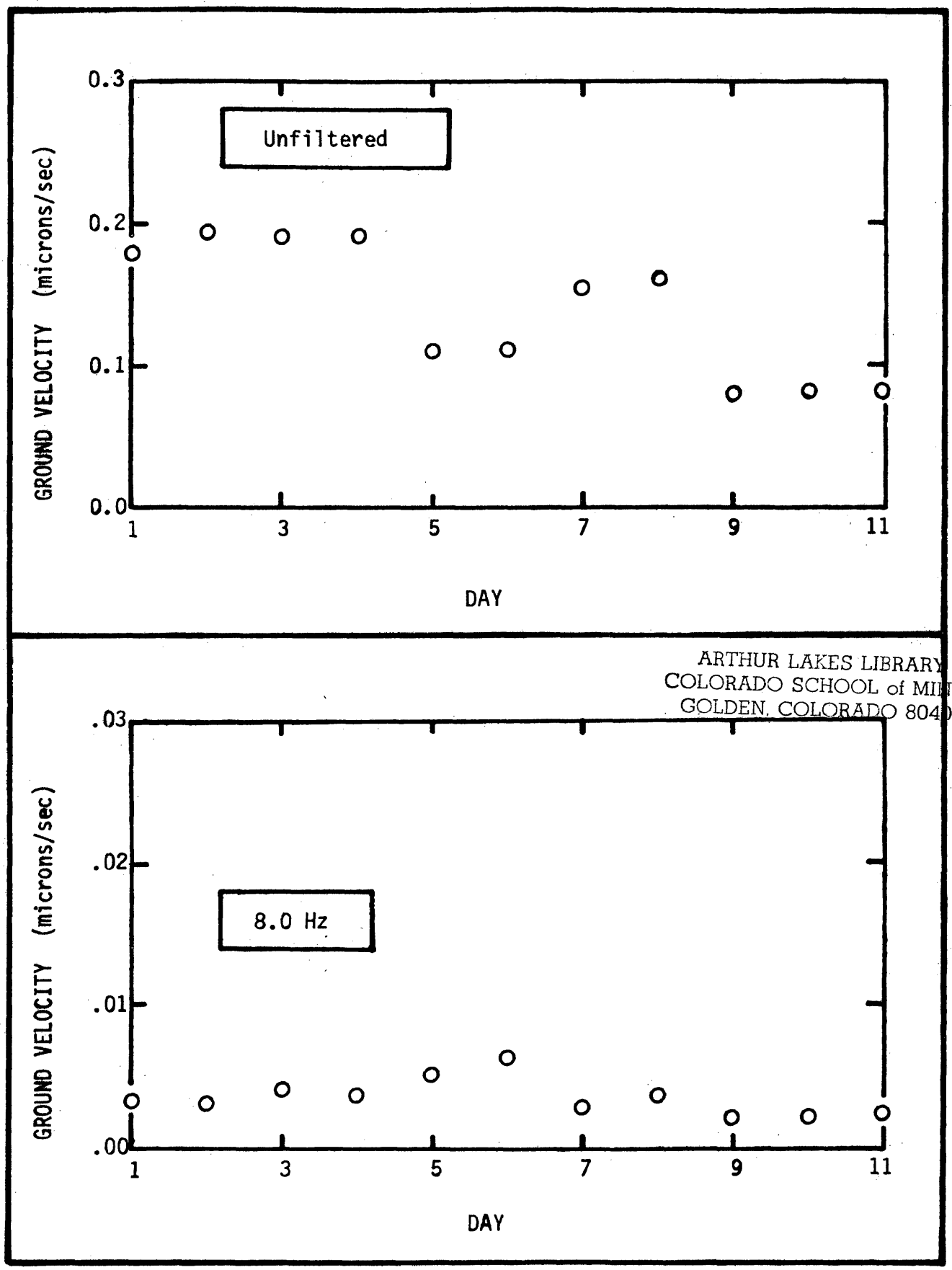


FIGURE 19. Base Station Variations In Ground Velocity At 8.0 Hz And The Unfiltered Data Records.

day period is approximately 3 mμ/sec with maximum deviations of about 7 db. The unfiltered record indicates much more erratic changes with an average ground velocity of 500 mμ/sec and maximum deviations of about 7 db.

A summary of minimum ground velocity versus frequency for the Kilauea base station is illustrated in Figures 20 through 22. These figures indicate generally decreasing groundnoise amplitudes with increasing frequency. Minimum ground velocities from day to day, however, show considerable variation.

The operation of a base station was to serve two purposes. The first was to monitor groundnoise variations at the same location for many days during the survey and the second was to use this information to normalize the amplitudes of the substations. Analysis of the records for simultaneous recordings indicated, however, that changes in ground velocity could seldom be seen at all three recording stations. For this reason, the substations were not normalized to the base station.

Figures 23 through 28 illustrate minimum microseismic groundnoise velocities for the Kilauea Volcano region. Figure 23 (ground velocity at 0.5 Hz) indicates two groundnoise highs on the northwest and the southeast side of Kilauea Crater. The only other velocity contrast of any significance is a small local high at station number 18 just south of Chain of Craters Road. This may be of some significance because a short time after the completion of this survey, an eruption occurred in the vicinity of this recording location.

The ground velocities at 1.0 Hz are illustrated in Figure 24. A velocity high extends across Kilauea Crater and closes to the east.

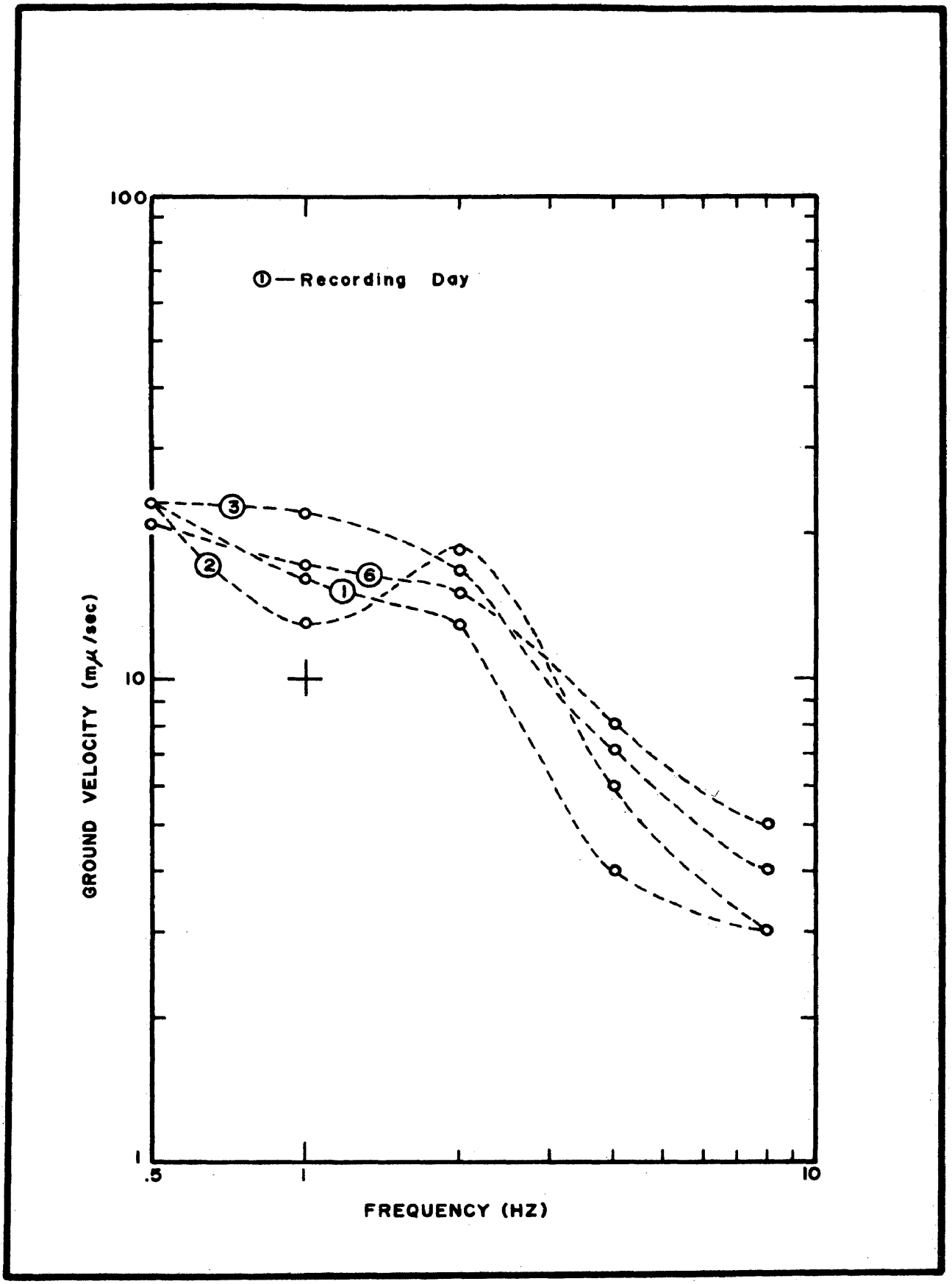


Figure 20. Ground velocity versus frequency at the Kilauea base station.

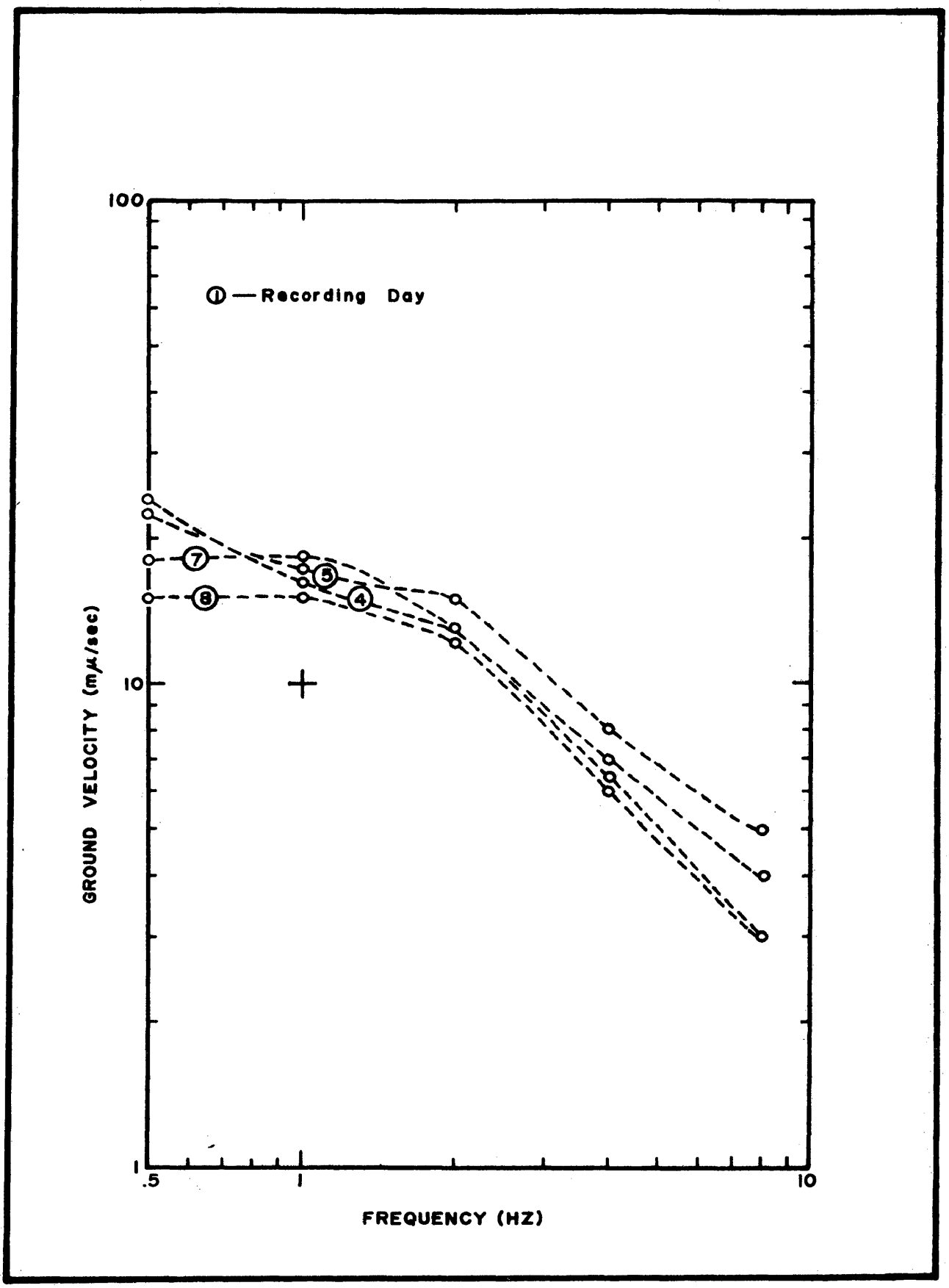


Figure 21. Ground velocity versus frequency at the Kilauea base station.

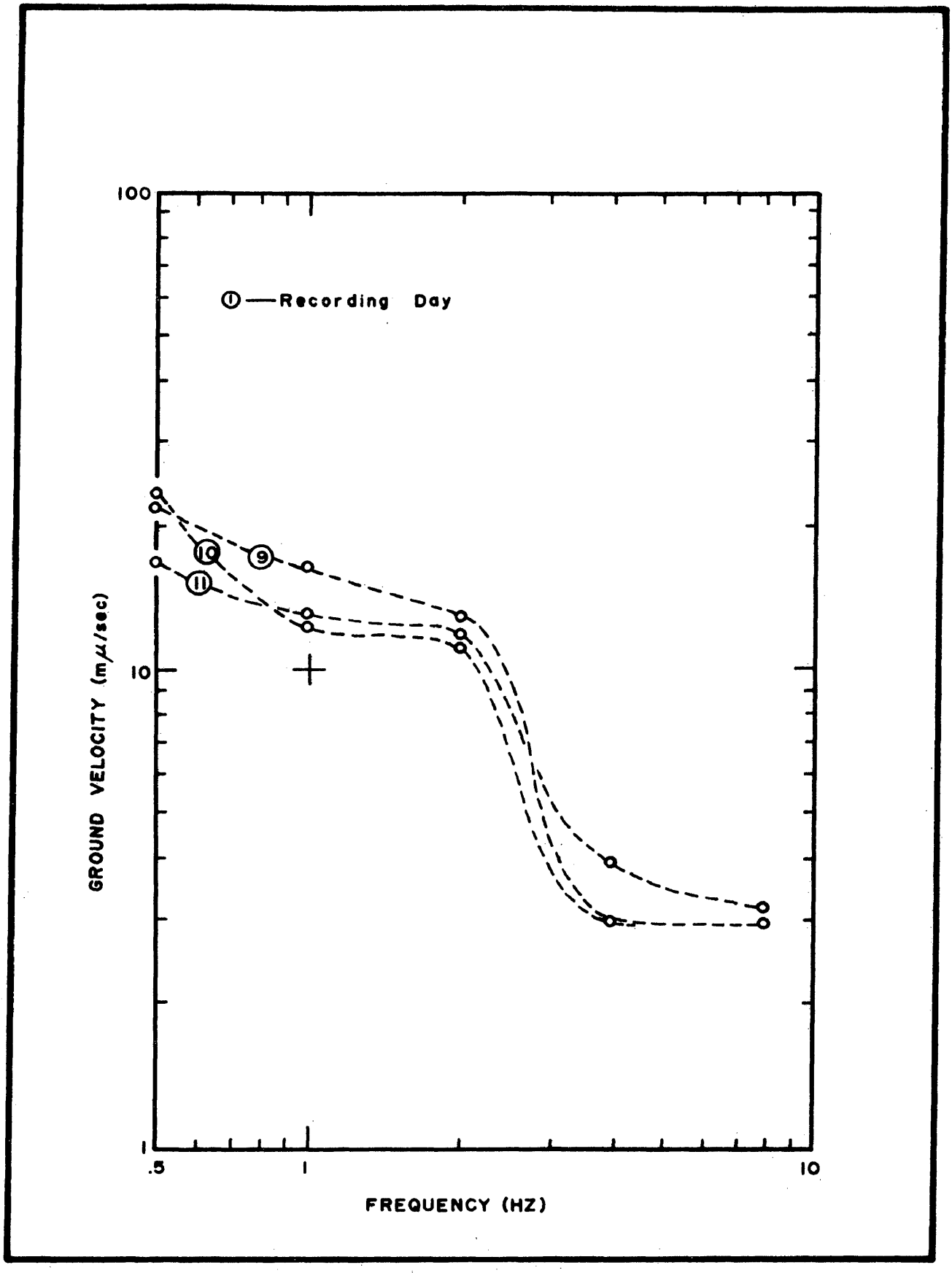


Figure 22. Ground velocity versus frequency at the Kilauea base station.

There is also an increase in intensity to the southeast in the Ainahou Ranch area. Very low ground velocities are indicated in the Kau desert region. Ground velocities for 2.0 Hz are shown in Figure 25. A major increase in vertical ground velocity is evident as one proceeds from the Kau Desert in the west to Pauahi Crater in the east.

Figure 26, with 4.0 Hz velocities, also shows a velocity trend of low to high from west to east. In this frequency range, however, we begin to see the effect of an eruption east of Puu Huluhulu. The effect of this eruption is most pronounced in the higher frequencies, 8.0 Hz band, as the velocity analysis indicates in Figure 27.

The map of unfiltered ground velocities for Kilauea summit is illustrated in Figure 28. This presentation, although weighted by the low frequency end of the spectrum because of high frequency attenuation, gives one of the best representations of groundnoise in the area. A velocity high is centered at the east side of Kilauea Crater. It extends across Kilauea and then trends south into the Kau Desert. Another velocity high of equal magnitude trends northwest and parallels the "Chain of Craters" which extends to the southeast.

A summary of dominant groundnoise frequencies in the Kilauea area is shown in Figure 29. As indicated, most of the Kilauea region is characterized by 0.5 Hz groundnoise. The only significant change in dominant frequency occurs southeast of Kilauea Crater near Mauna Ulu. An eruption was taking place in this area while the survey was being carried out. The increase in dominant frequency results from harmonic tremor which occurs when magma is forced out of the surface fissures.

The results of other geophysical measurements will be presented

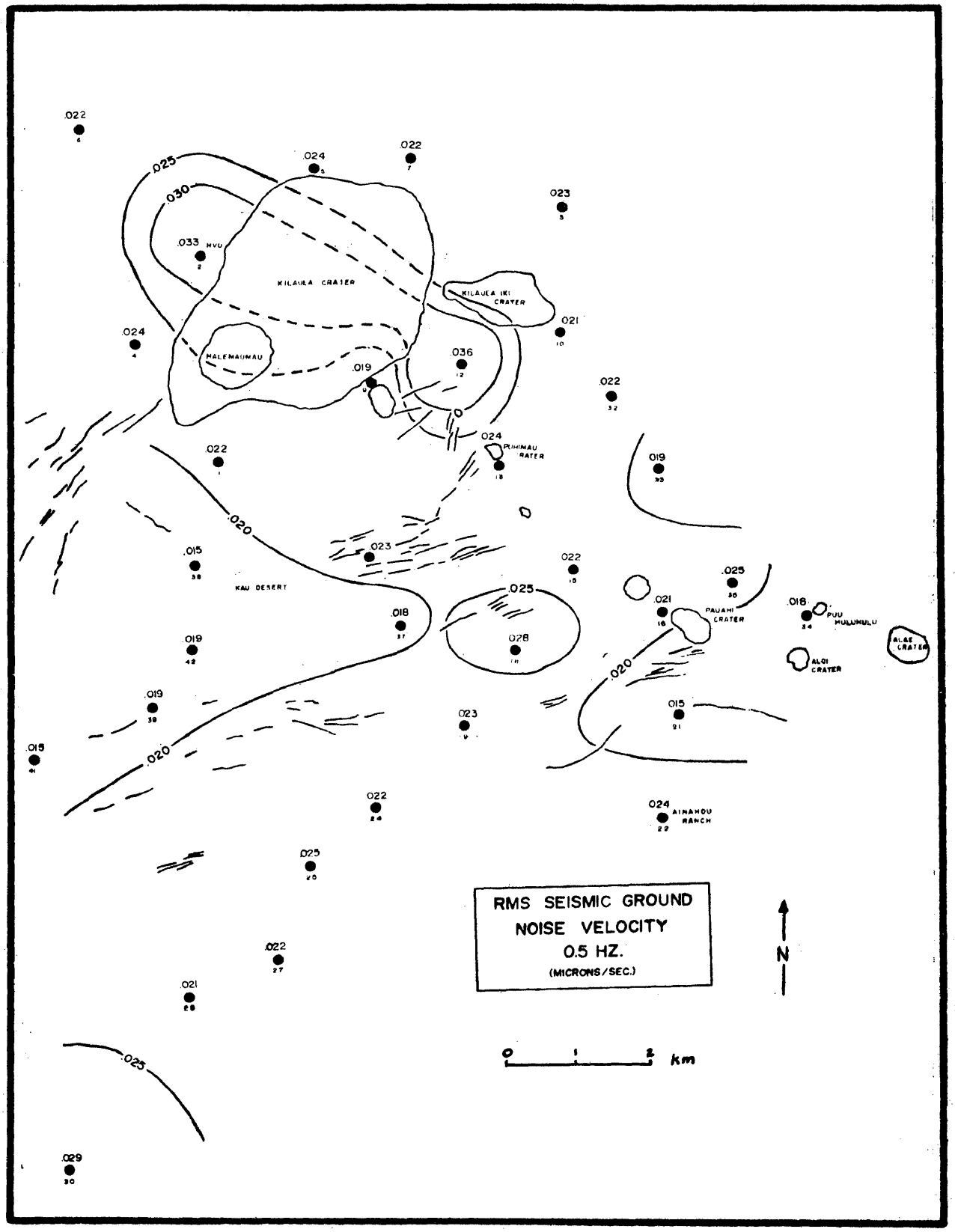


FIGURE 23. Microseismic Groundnoise At Kilauea Volcano (0.5 Hz).

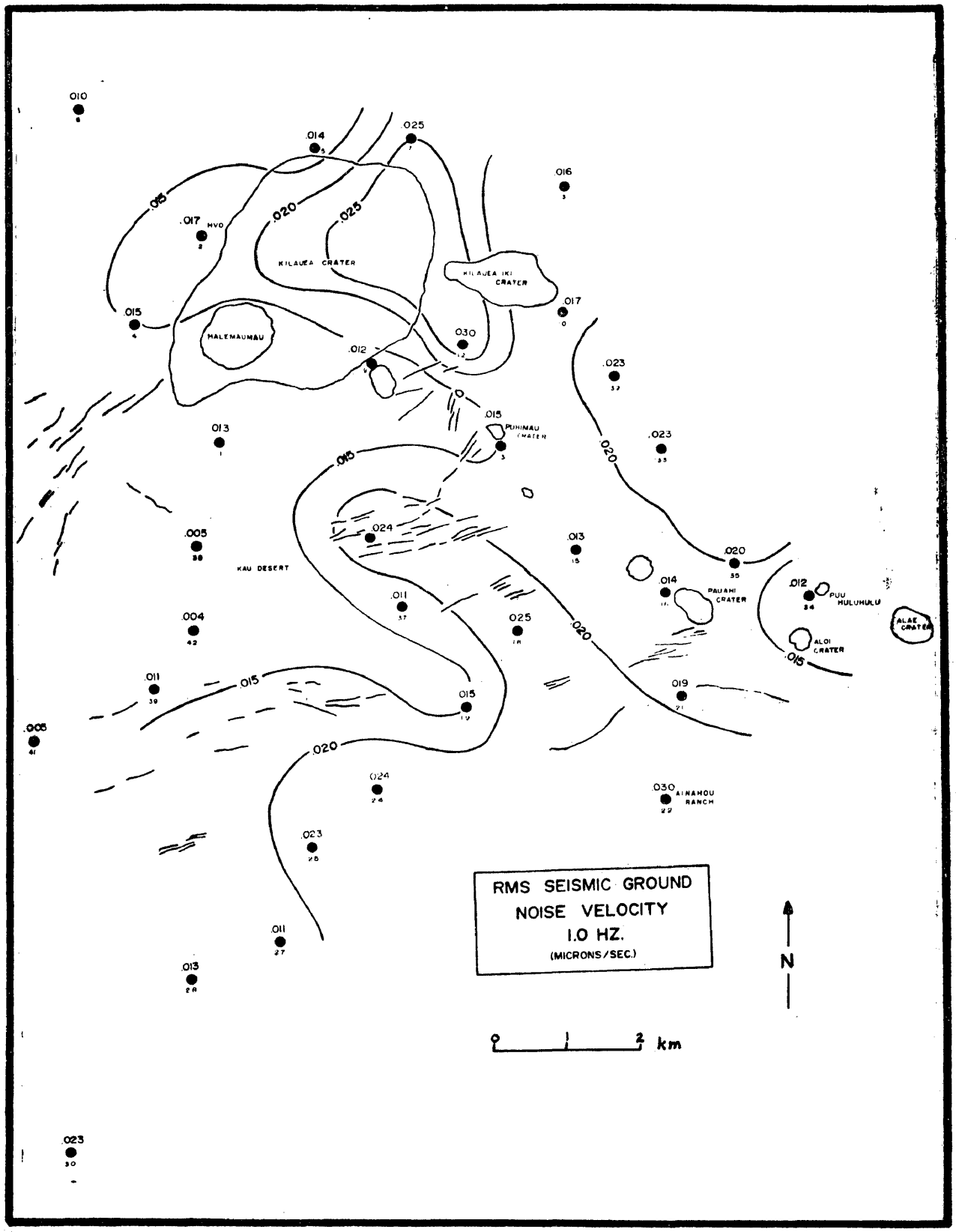


FIGURE 24. Microseismic Groundnoise At Kilauea Volcano (1.0 Hz).

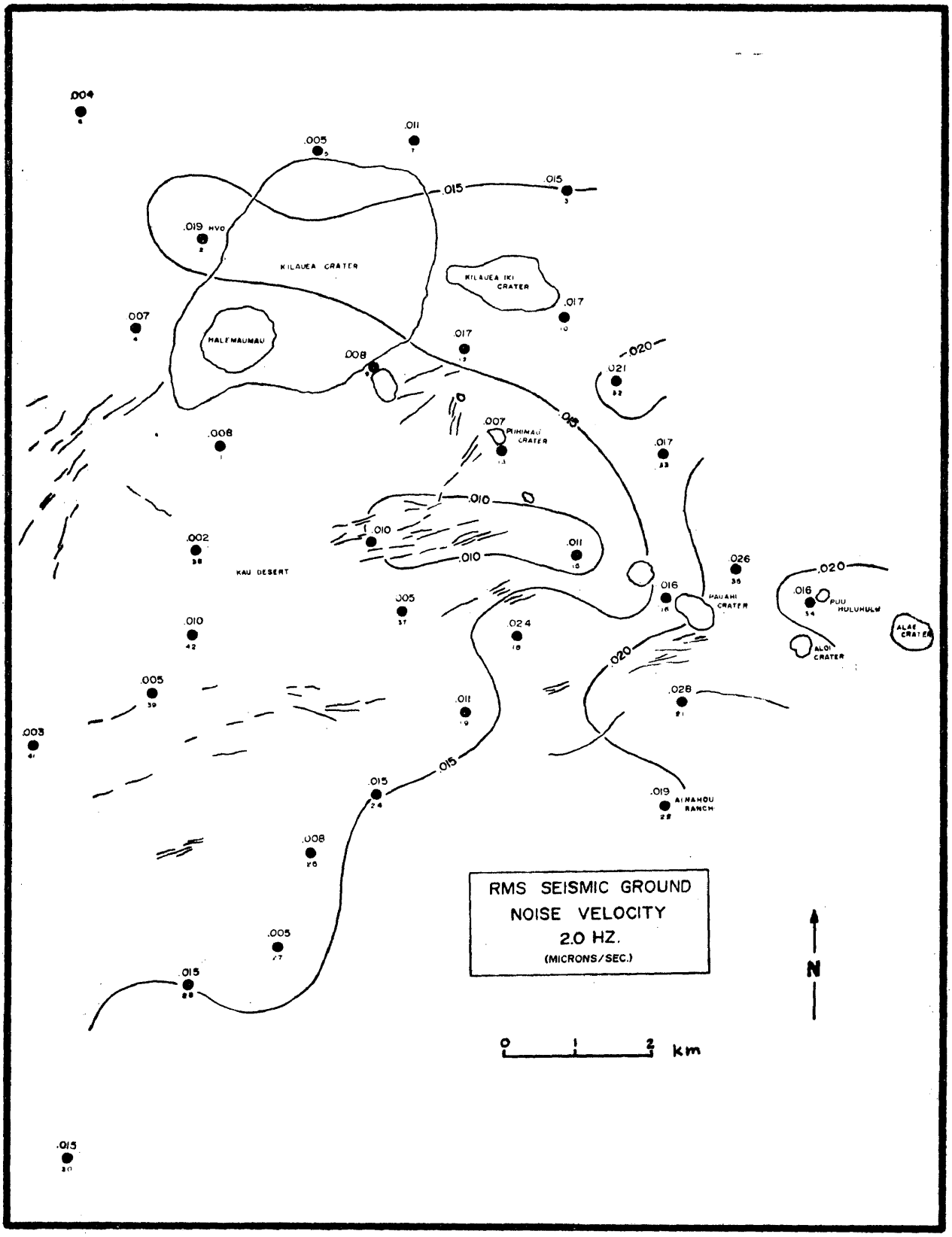


FIGURE 25. Microseismic Groundnoise At Kilauea Volcano (2.0 Hz).

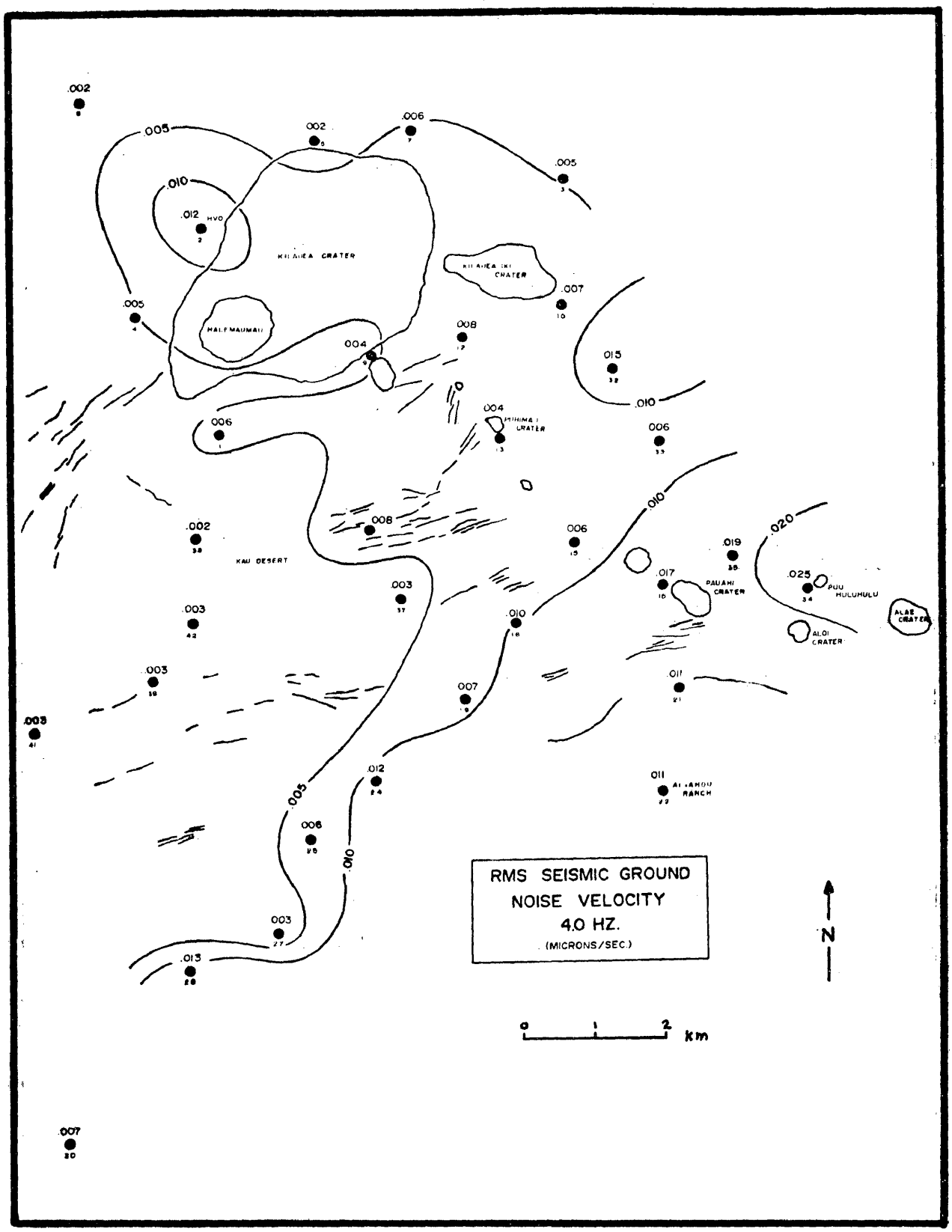


FIGURE 26. Microseismic Groundnoise At Kilauea Volcano (4.0 Hz).

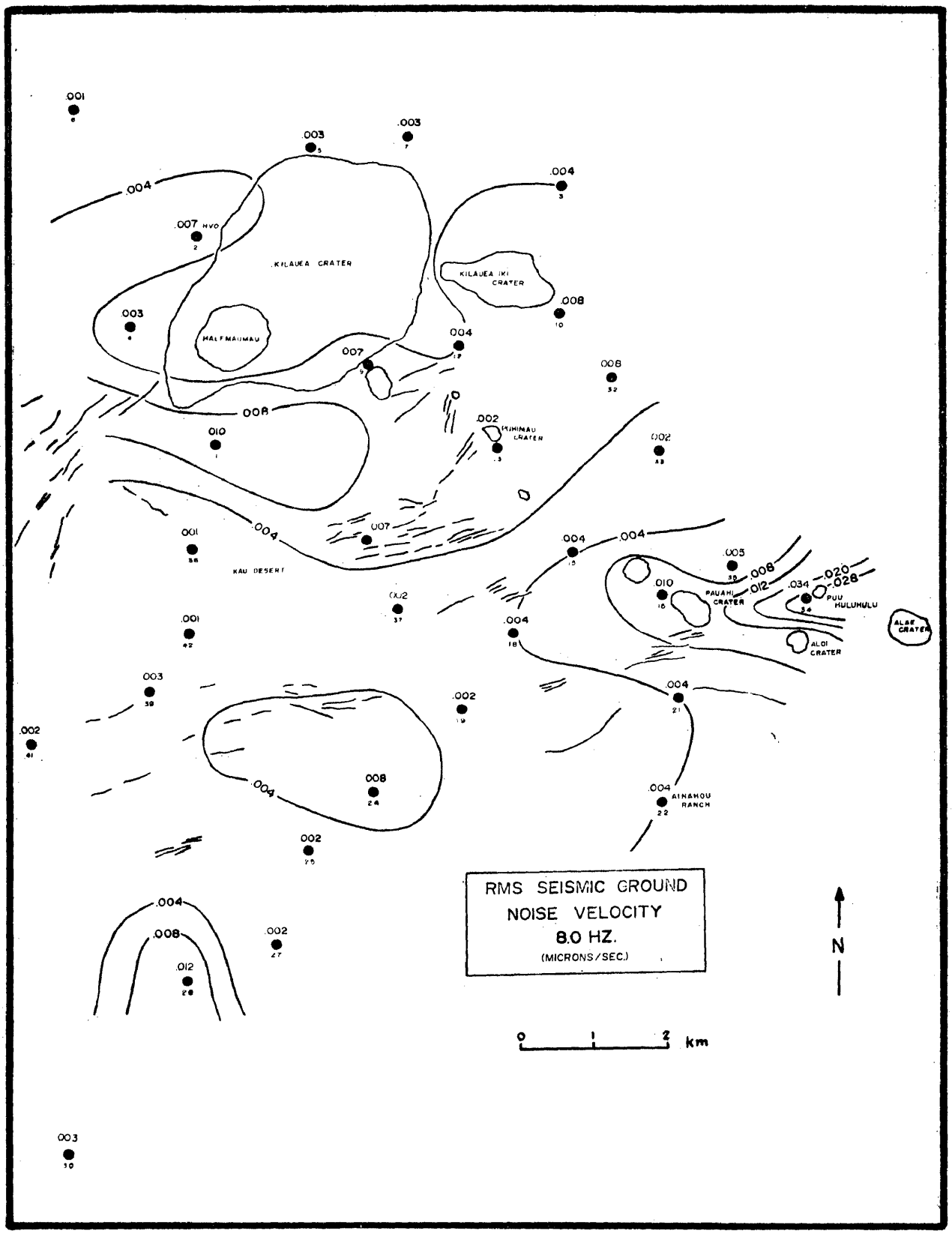


FIGURE 27. Microseismic Groundnoise At Kilauea Volcano (8.0 Hz).

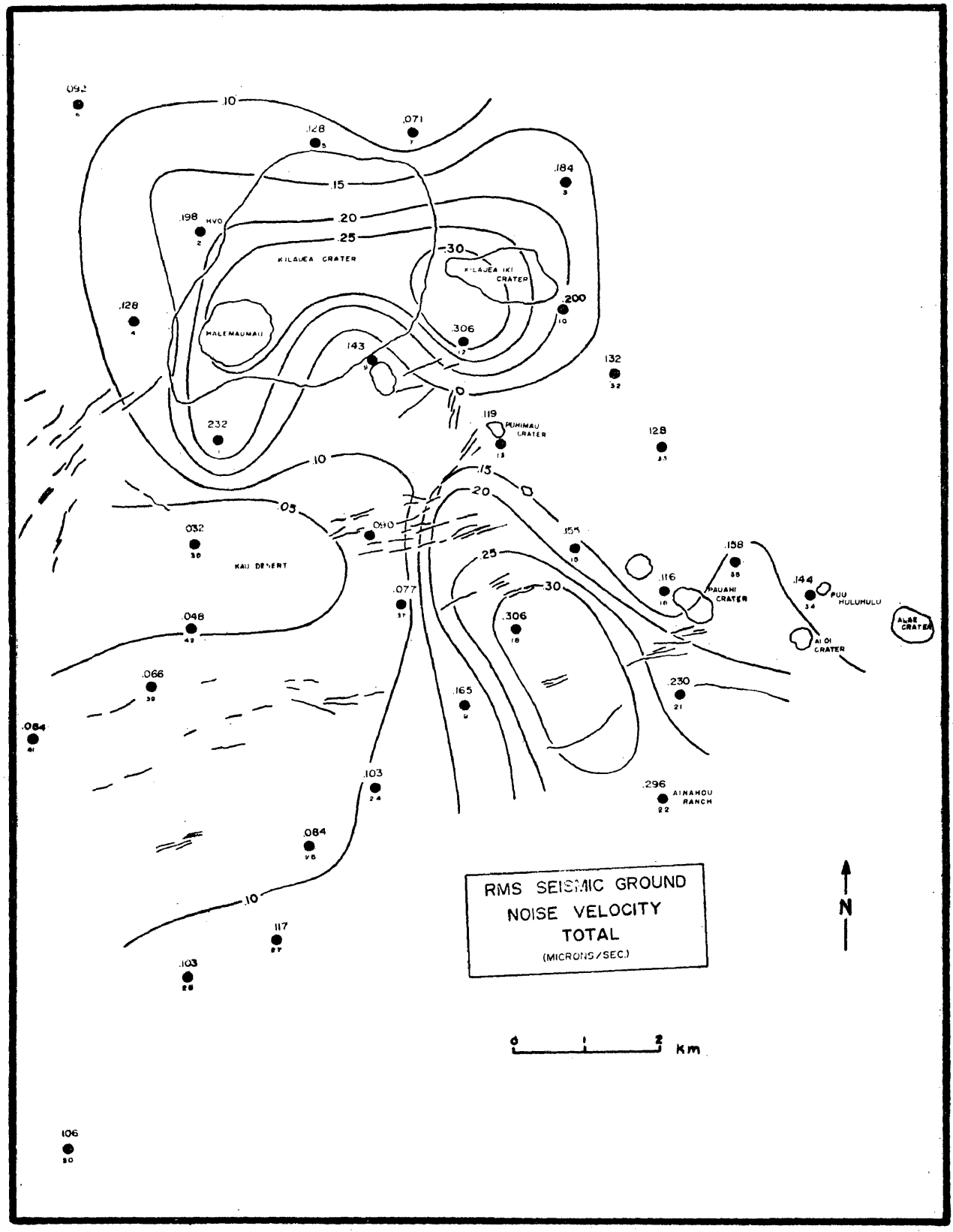


FIGURE 28. Microseismic Groundnoise At Kilauea Volcano (Unfiltered).

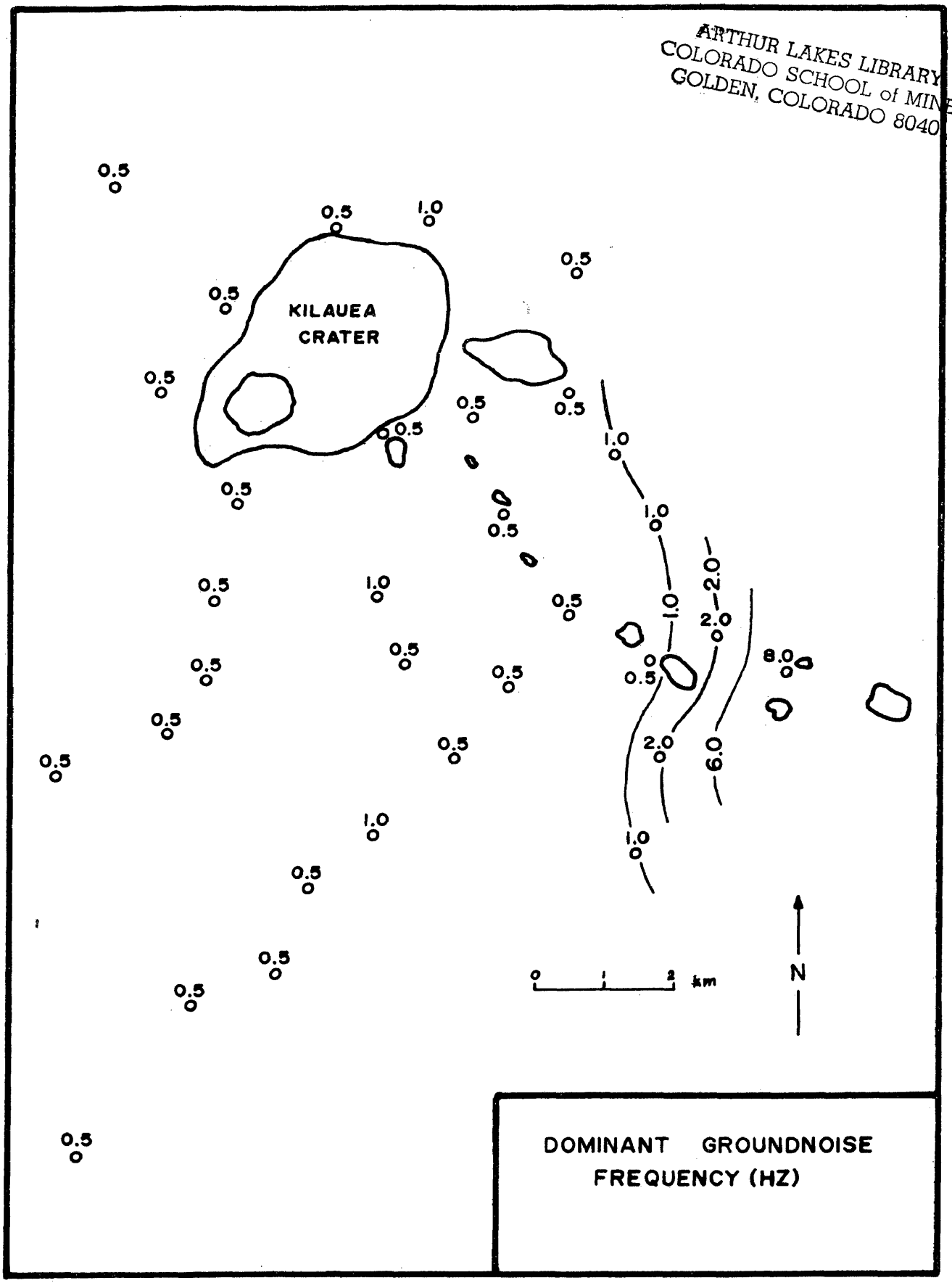


Figure 29. Dominant Groundnoise Frequencies at Kilauea Crater.

before these microseismic groundnoise measurements are evaluated.

Groundnoise Data From The East Rift Zone

Vertical ground velocities for the forty-five recording locations in the East Rift Zone are illustrated in Figures 30 through 36. The 0.5 Hz analysis in Figure 30 indicates a decrease in velocity with increasing distance from the shoreline. A few low velocity zones within the Rift Zone indicate minor variations in local geology. An increase in frequency to 1.0 Hz (Figure 31) indicates a velocity dependence which is also parallel to the shoreline except in the area between Puu Kaliu and Puulena Crater.

Ground velocities at 2.0 Hz (Figure 32) have very little variation except in the Puulena Crater area. As the analysis frequency increases to 4.0 Hz (Figure 33), more complex velocity variations are evident. We continue to observe a groundnoise high around Puulena Crater but there is also a velocity high located two miles south of Halekamahina.

The 8.0 Hz velocity analysis, Figure 34, continues to indicate a groundnoise high near Puulena Crater but it has an east-west elongation at the south edge of the crater. The major trend in maximum ground velocity is parallel to the Rift Zone but is shifted to the south from the surface expressions of the Rift. The groundnoise illustration for the unfiltered data (Figure 35) also indicates a major east-west high velocity trend with the greatest amplitude in the Puulena Crater area.

A summary of dominant groundnoise frequencies for the East Rift Zone is illustrated in Figure 36. As indicated, there are two areas which are characterized by high frequency groundnoise. One of these areas is

located near Puulena Crater and the other is located 3 km southwest Kapoho Crater.

The results of other geophysical measurements will be considered before any conclusions are made on the significance of these groundnoise measurements.

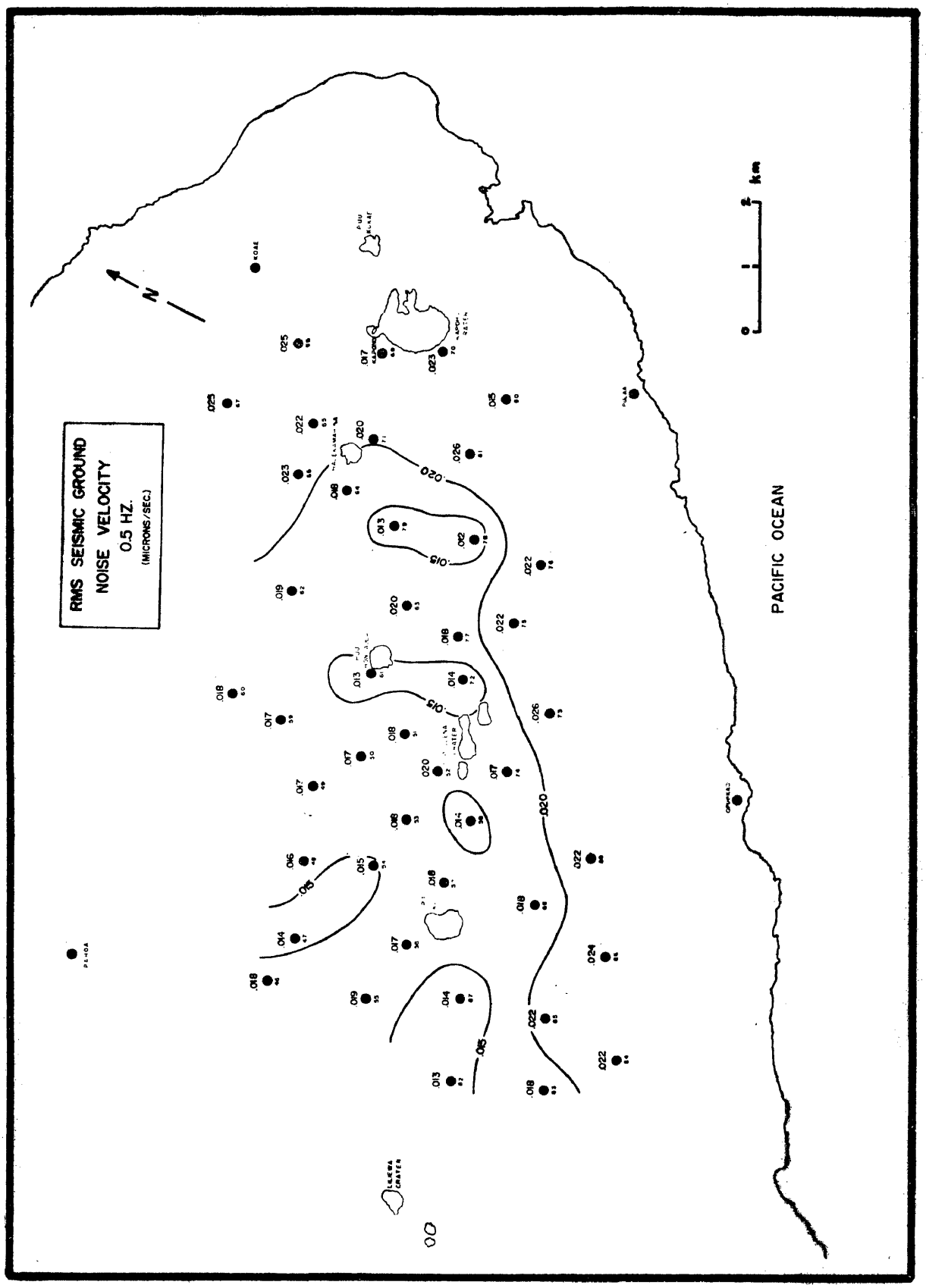


FIGURE 30. Microseismic Groundnoise At The East Rift (0.5 Hz).

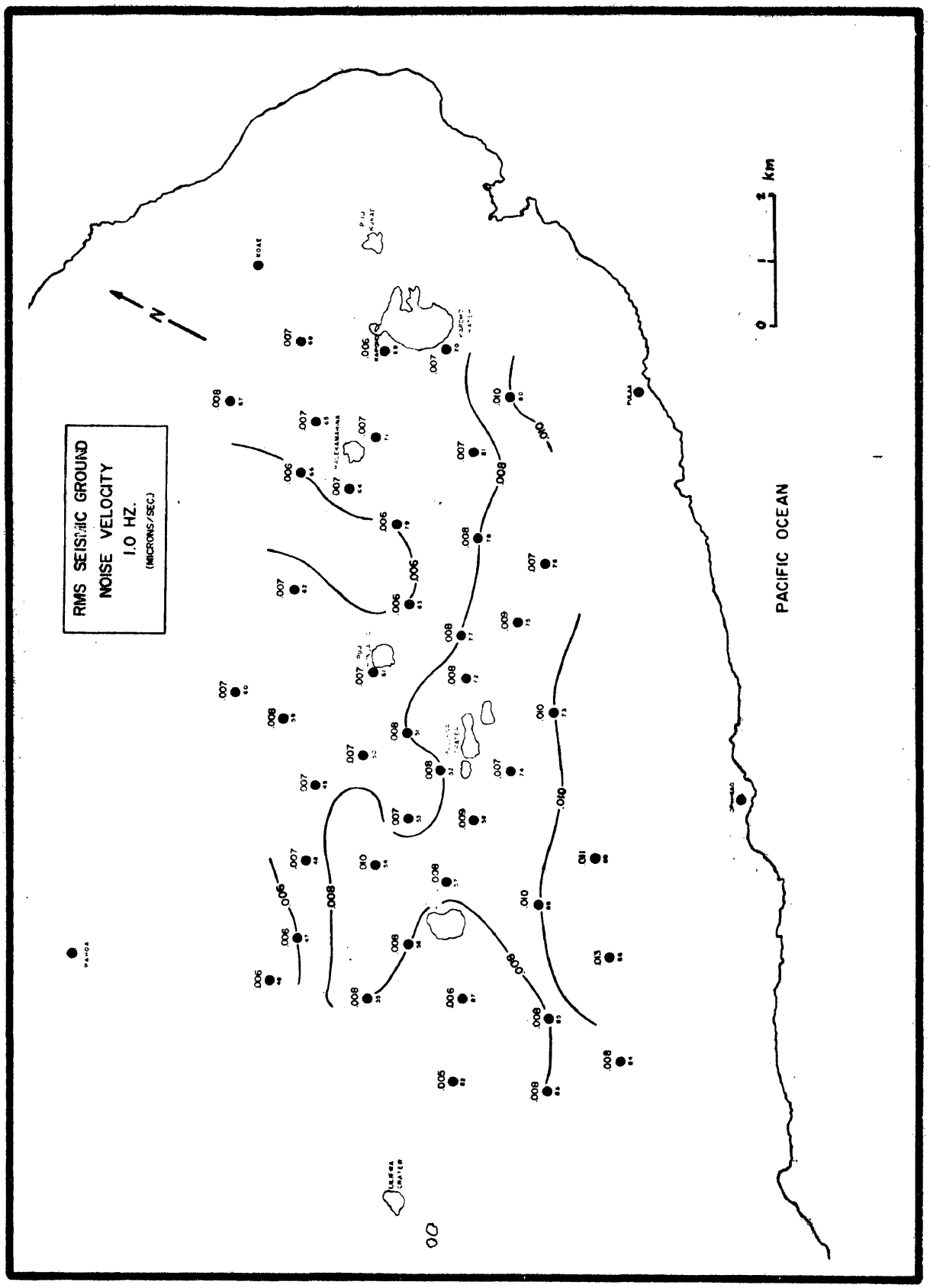


FIGURE 31. Microseismic Groundnoise At The East Rift (1.0 Hz).

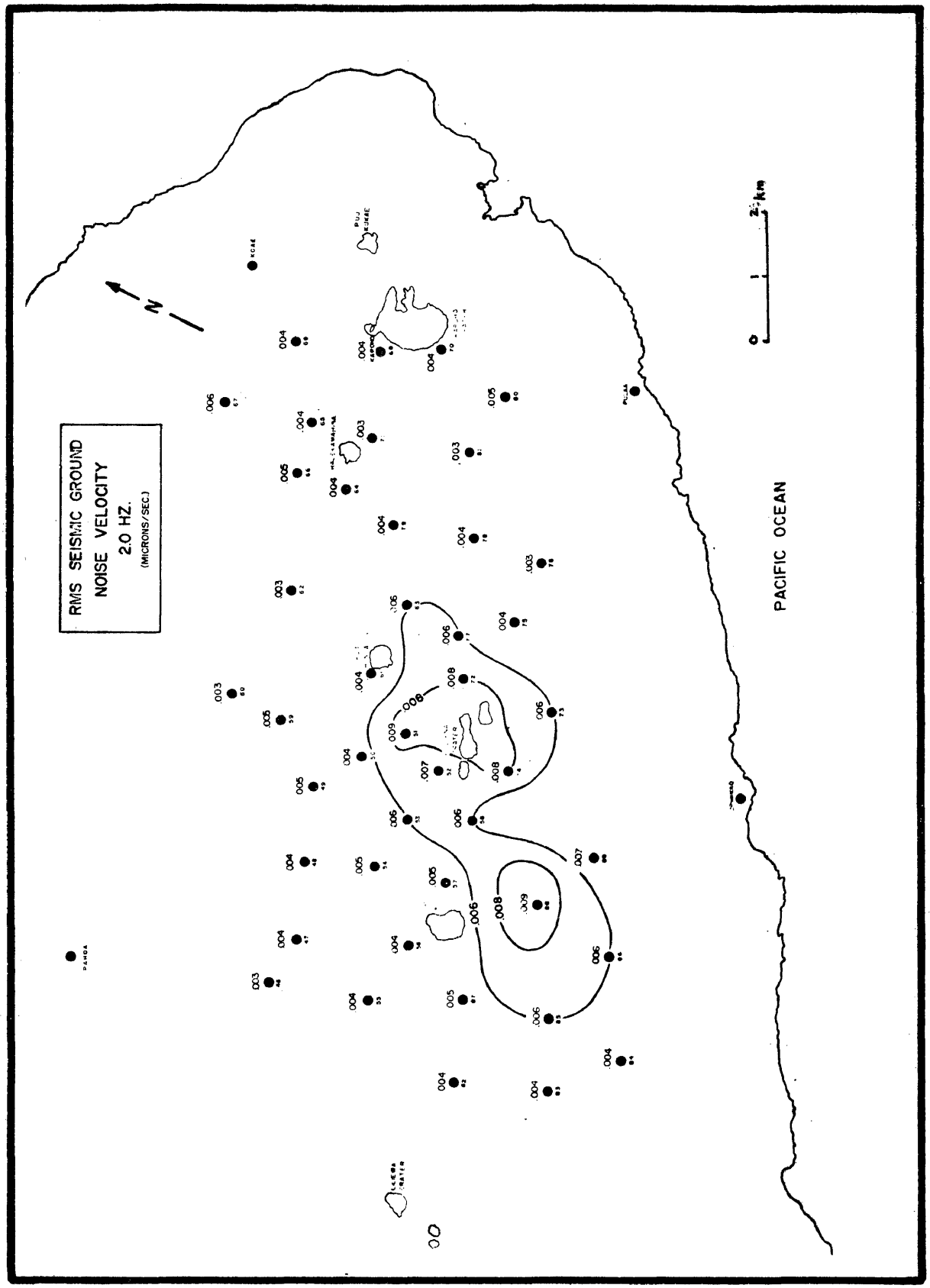


FIGURE 32. Microseismic Groundnoise At The East Rift (2.0 Hz).

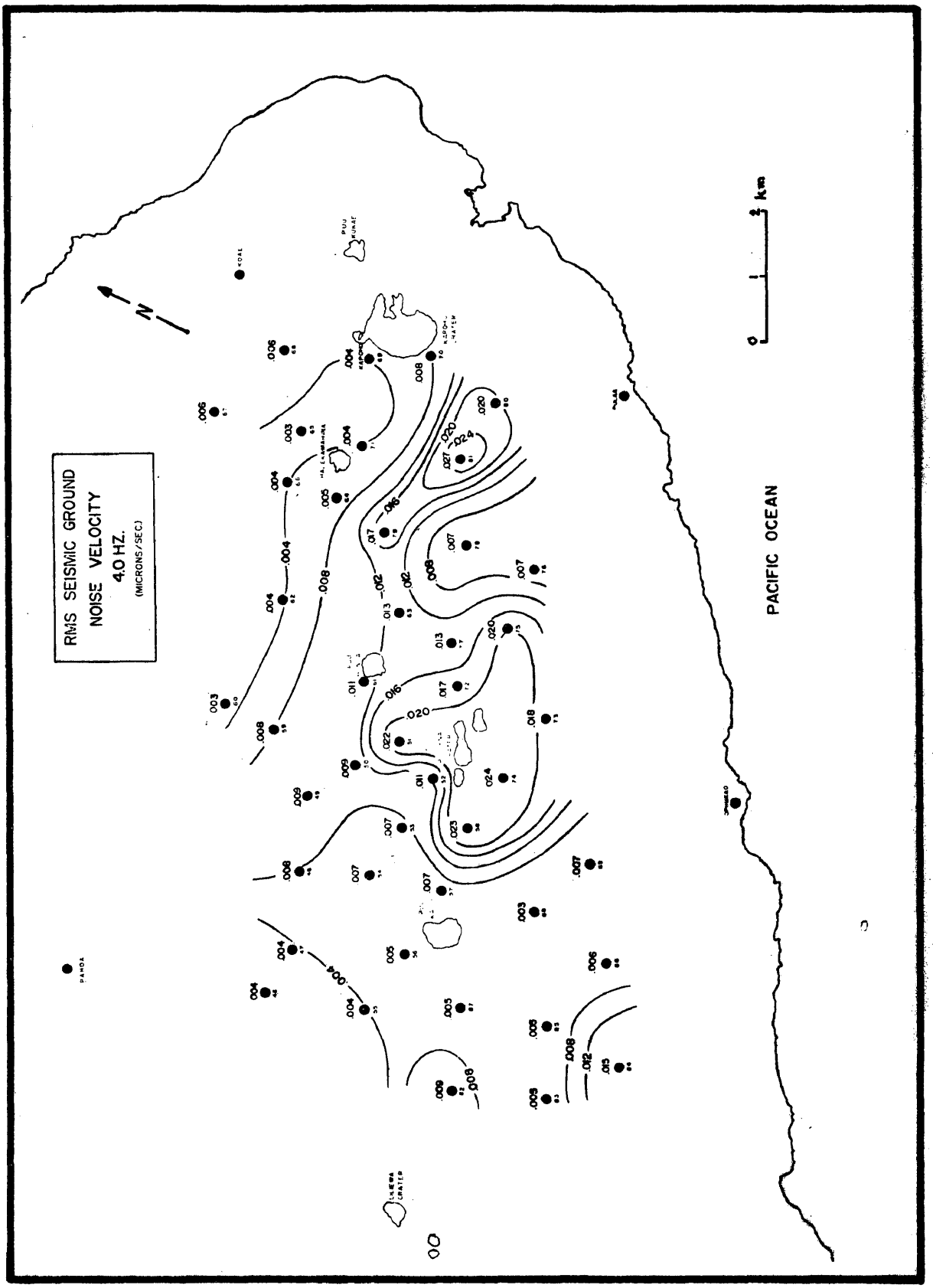


FIGURE 33. Microseismic Groundnoise At The East Rift (4.0 Hz).

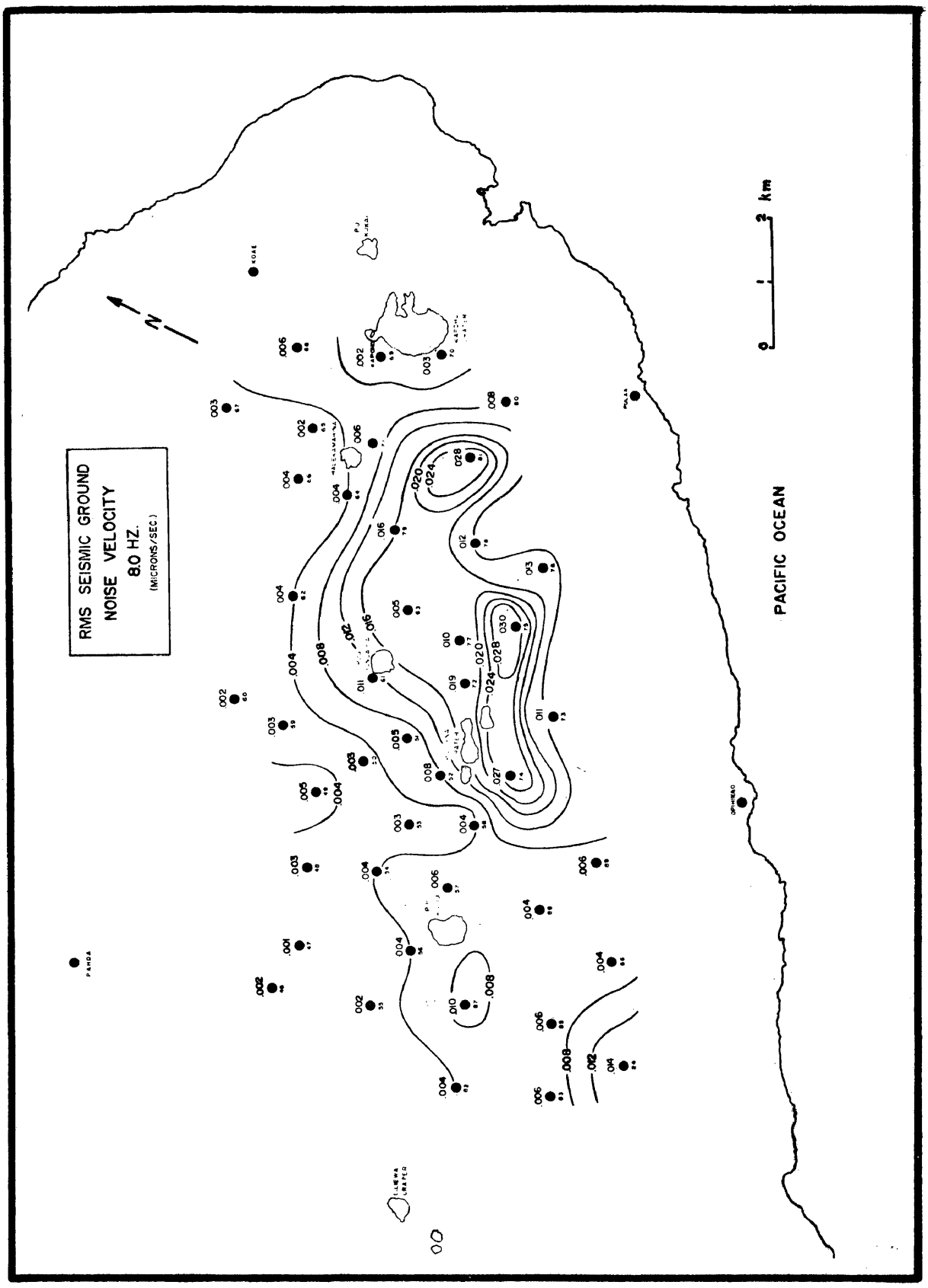


FIGURE 34. Microseismic Groundnoise At The East Rift (8.0 Hz).

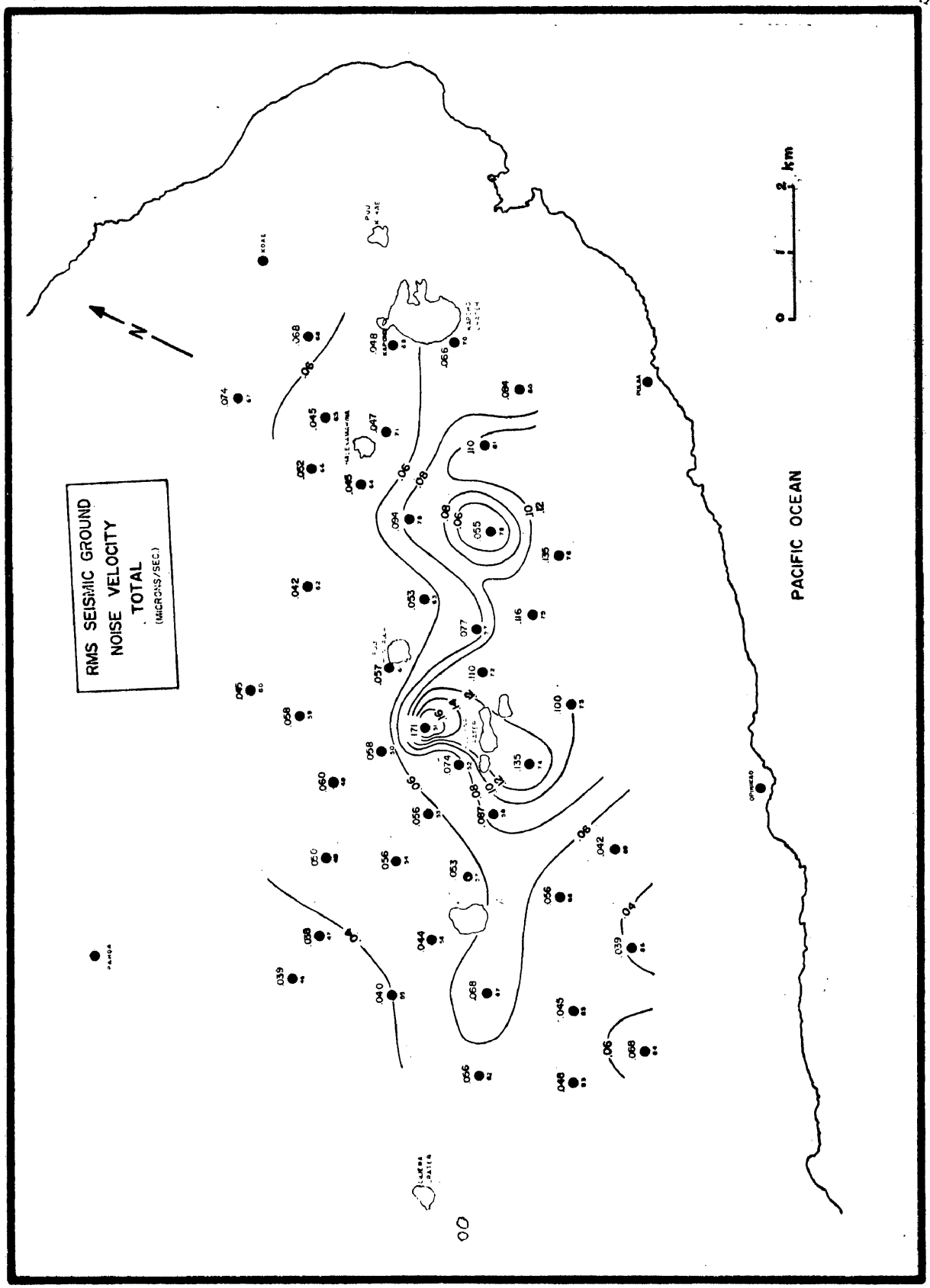


FIGURE 35. Microseismic Groundnoise At The East Rift (Unfiltered).

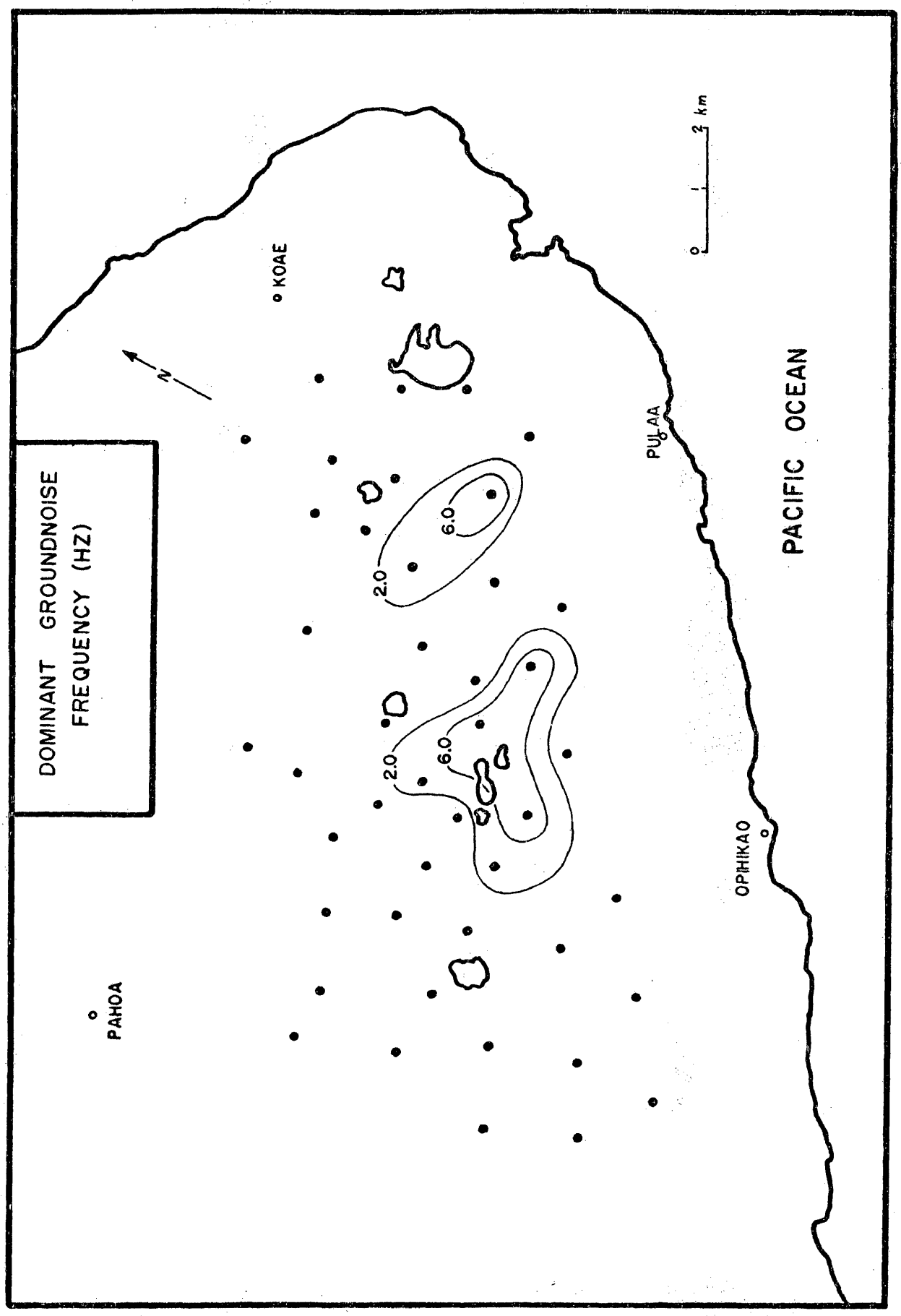


FIGURE 36. Dominant Groundnoise Frequencies In The East Rift Zone of Kilauea Volcano.

RESULTS OF OTHER GEOPHYSICAL SURVEYS AT KILAUEA VOLCANO AND THE EAST RIFT ZONE

Several geophysical surveys have been conducted at Kilauea Volcano and in the East Rift Zone. The United States Geological Survey has operated the Hawaiian Volcanic Observatory (HVO) for many years. This organization, with its geological and geophysical studies, has made Kilauea Volcano a field laboratory of recent volcanic activity. Most of the observations of HVO have been related to local microseismic activity and ground deformation studies. Other geophysical surveys have included gravity, magnetic, and electrical resistivity. Detailed scientific information has also been gained with the completion of a borehole at the summit of Kilauea.

Kilauea Volcano

Electrical Resistivity

A time-domain electromagnetic survey was used to study conductivity variations with depth at the summit of Kilauea Volcano (Jackson and Keller, 1972). The resistivity soundings were interpreted with two-layer earth modeling. Resistivities varied from 8- to 24 ohm-meters for the upper layer and were about 2 ohm-meters for the lower layer. A contour map of the depth to the top of the low resistivity zone is shown in Figure 37.

A complex dike and fault system is believed to support a high-level water table in the Kilauea area. The high resistivity upper

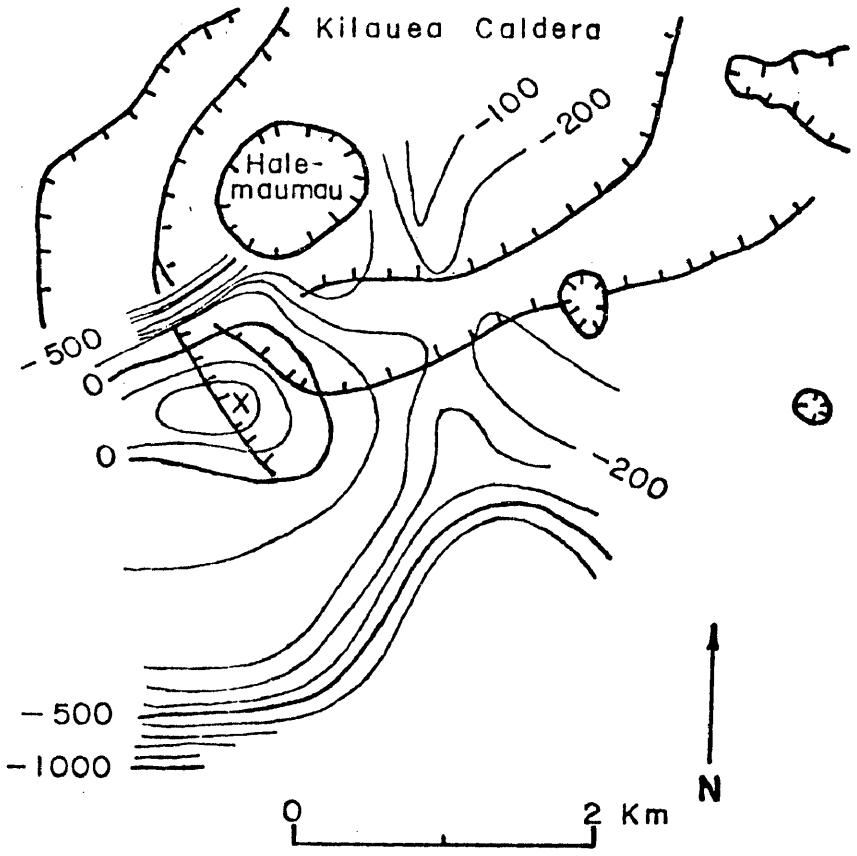


FIGURE 37. Contour Map Of The Top Of A Low Reststivity Zone At The South End Of Kilauea Crater (From Keller, 1974).

layer is probably related to this dike-impounded fresh water. The low resistivity zone, which is nearest the surface of the earth (less than 900 meters deep) at the south edge of Kilauea Crater, is thought to be related to a thermally-induced upwelling of saline water (Jackson and Keller, 1972).

A dipole mapping survey was also conducted at the summit of Kilauea (Keller, written communication). One of the sources, bipole source 1, used in this survey was located one mile south of Halemaumau Crater (see Figure 38). Very low resistivities are apparent on the east side of Kilauea Crater. This area of low resistivity is the northeastern extension of the southwest rift of Kilauea. Apparent resistivities in the "Chain of Craters" area to the southeast of Kilauea are only moderately low and do not indicate any anomalous regions. If the resistivities mapped from source 1 are compared to the resistivities mapped from source 11 (Figure 39) on the uphill side of Kilauea Crater, however, an important contrast is apparent. Resistivities outside of the Rift Zone are larger by an order of magnitude. This tends to indicate that the Rift is highly faulted and fractured and has much greater porosity than the surrounding area.

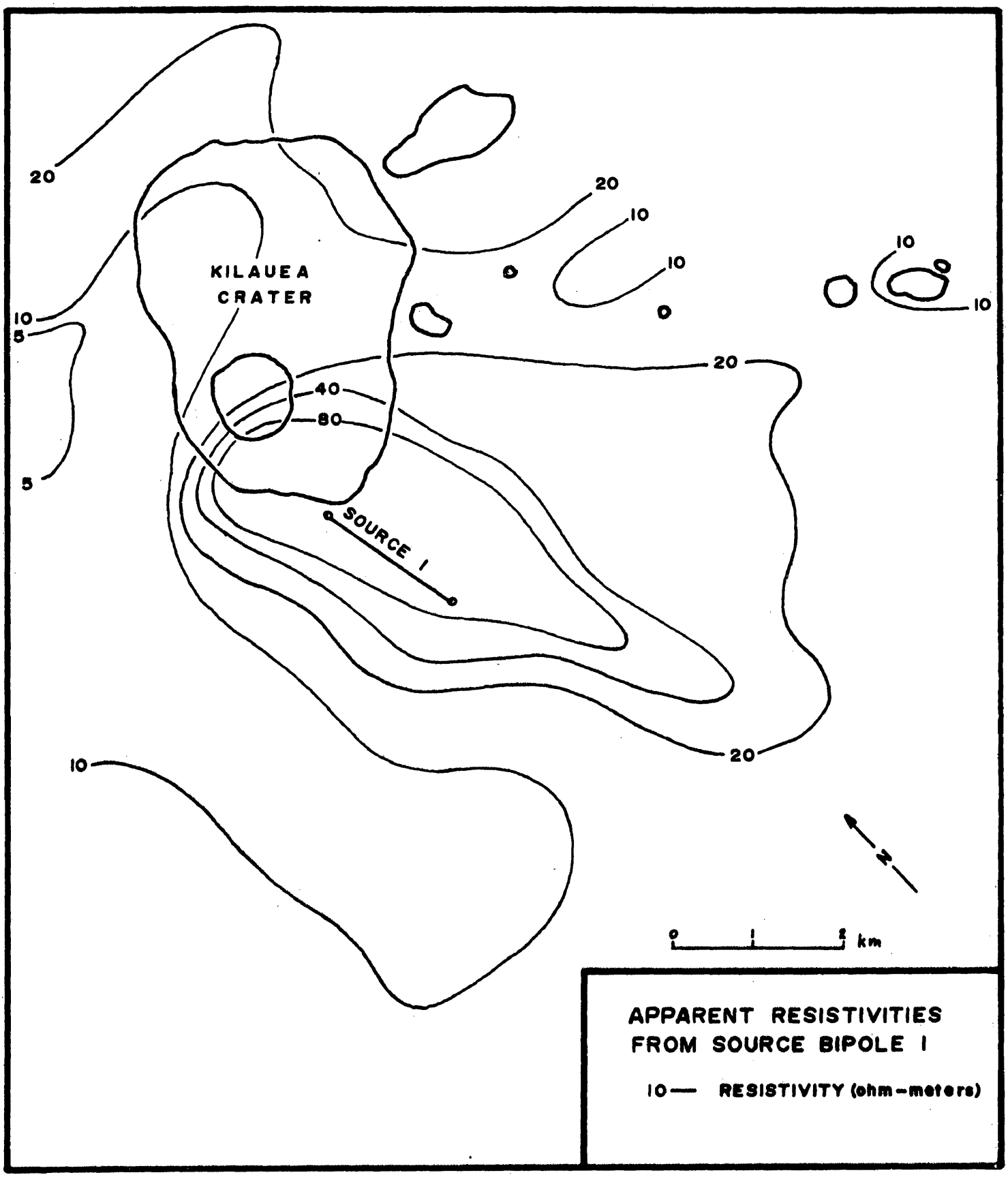


FIGURE 38. Apparent Resistivities Mapped From Source Bipole 1.

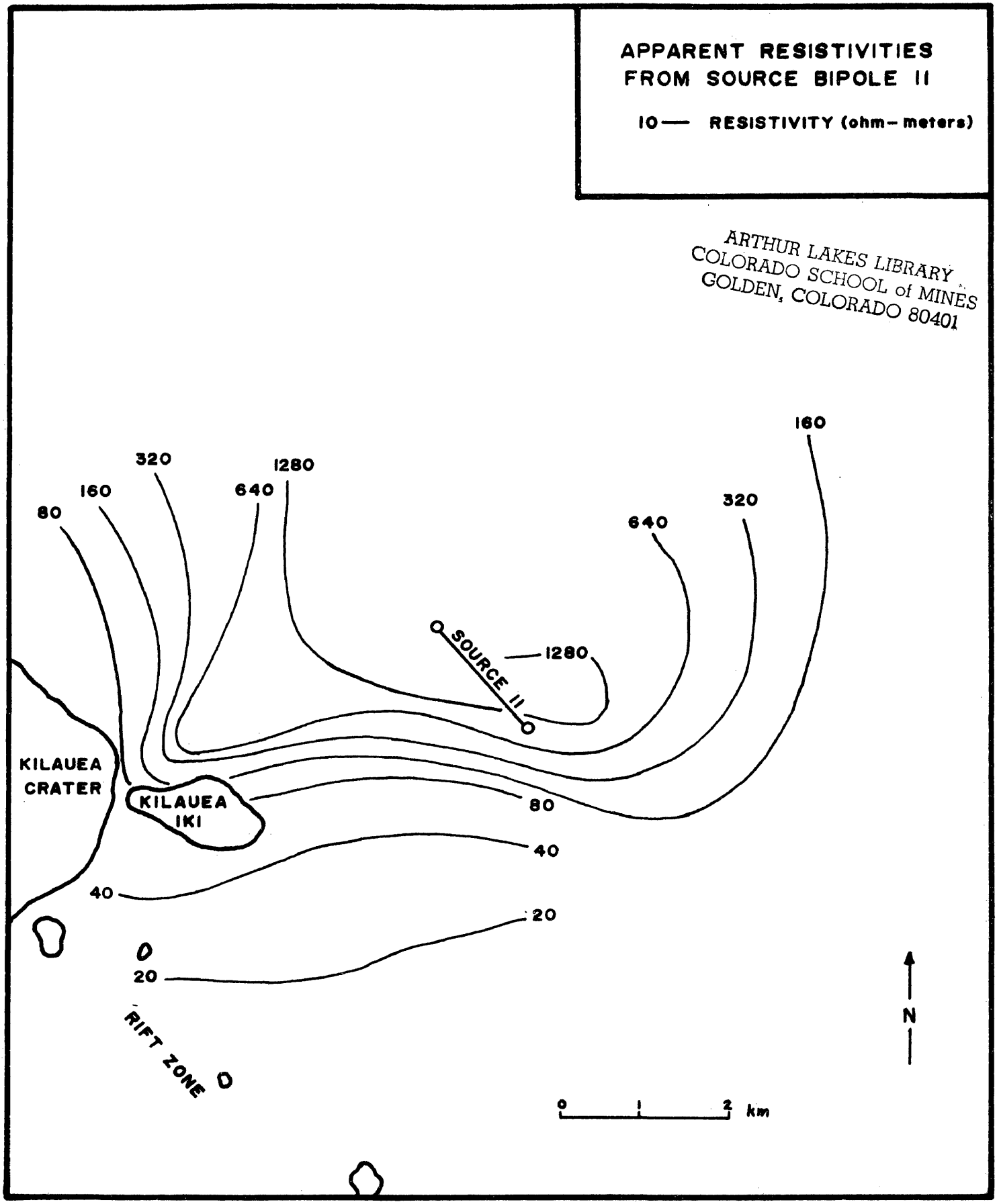


FIGURE 39. Apparent Resistivities Mapped From Source Bipole II.

Earthquake Seismicity

A summary of Hawaiian seismic activity for 1969 was presented by Koyanagi and Endo (1971). It is evident from this particular study and many others like it that the area of highest earthquake concentration and highest recurrence rate for the Island of Hawaii was at Kilauea Volcano and the southwest rift of Mauna Loa. Swarms of small earthquakes have been confined to areas of eruption and to an area just south of Kilauea Crater. These earthquakes had focal depths of a few kilometers. During periods of non-eruption, earthquakes were scattered over a much larger area and occurred at much greater depths.

Figure 40 is a detailed location map of the Kilauea Crater area and a summary of earthquake activity within this region. All earthquakes within 8 km of a baseline A-A' on the location map were projected onto a vertical plane along that line. The trend of shallow earthquake activity beneath Kilauea Crater and deeper activity to the southeast is clearly evident. The electrical conductivity anomaly (from Figure 37) is also indicated in this figure and appears to correlate well with the highest occurrence of near surface earthquake activity.

A recent earthquake study by Ward and Gregersen (1974) at the summit of Kilauea Volcano indicates that the crust under Kilauea Crater and in the rift zones tends to have slightly higher seismic wave velocities: and that the crust under the Kaoiki fault zone just west of Kilauea and in the region south of Kilauea Crater, has slightly lower seismic wave velocities. These results indicate the probable intrusion of higher density rocks into the rift zones at depth.

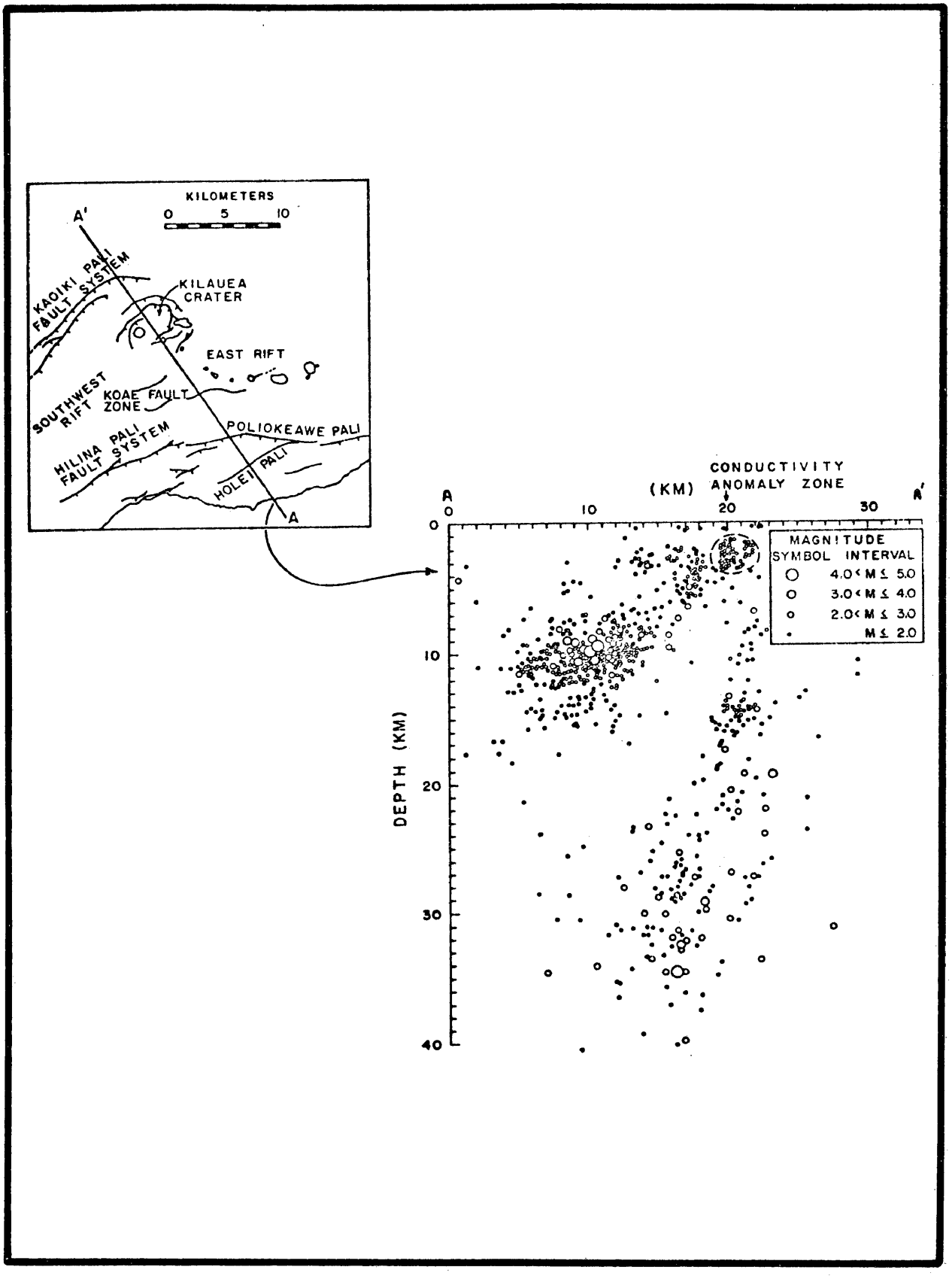


FIGURE 40. Earthquake Activity Versus Depth At Kilauea Volcano (From Koyonagi and Endo, 1971).

Crustal Deformation

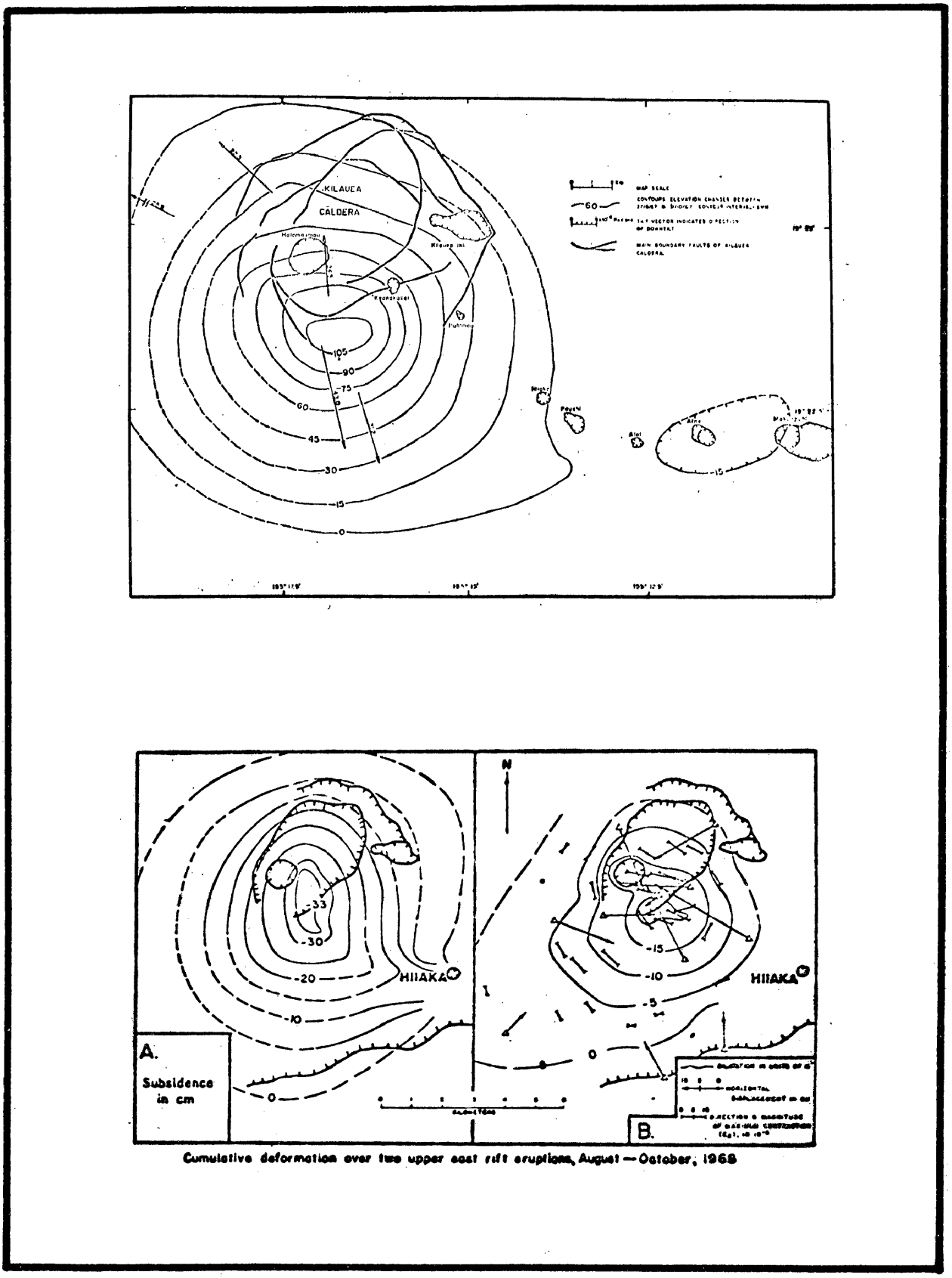
The Hawaiian Volcanic Observatory has operated several tiltmeter systems at the summit of Kilauea Volcano to monitor crustal deformation. These measurements indicate that the ground surface expands upward and outward before any surface volcanic activity begins, and then subsides and contracts after eruption (Kinoshita, et al., 1974).

The results of these vertical ground deformation studies for two periods of activity (September to October, 1967, and August to October, 1968) are illustrated in Figure 41. Both of these studies indicated that the maximum vertical uplift and subsidence occurred at the south edge of Kilauea Crater. These vertical displacements were 105 mm and 33 mm, respectively. Spirit level and water tube tilt measurements for a period of moderate inflation (November, 1969 to February, 1970) are indicated in Figure 42. The area of maximum ground tilt for this period is also at the south edge of Kilauea Crater.

Analyses of these and other deformation studies indicate the probable existence of a near surface magma reservoir at a depth of 2 to 4 km below the surface (Eaton, et. al., 1971). Many cracks and fissures in the vicinity of Kilauea Crater are evidence that displacements of this magnitude have exceeded the elastic properties of the near surface rocks. It is probable that this highly fractured zone extends to great depths below the summit of Kilauea.

Kilauea Borehole

A borehole was drilled at the summit of Kilauea Volcano in 1973. This hole was located one kilometer south of Halemaumau in the area of



Cumulative deformation over two upper east rift eruptions, August — October, 1968

FIGURE 41. Ground Deformation At Kilauea Crater (From Kinoshita et al, 1974).

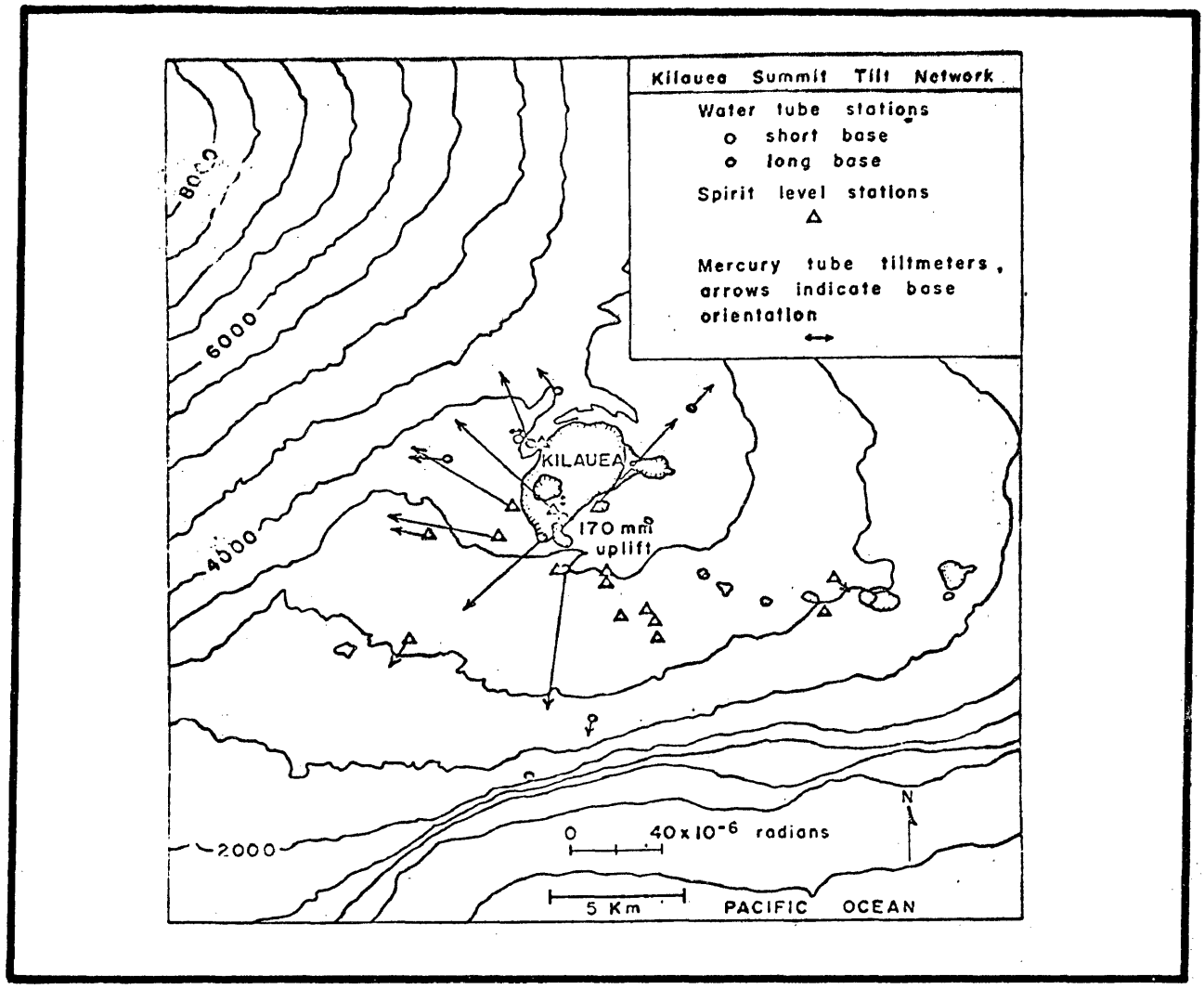


FIGURE 42. Ground Tilt From Kilauea Summit Tilt Network (From Kinoshita, et al., 1974).

maximum ground deformation, lowest resistivity, and highest occurrence of shallow earthquakes. The purpose of the borehole was to test the possible occurrence of geothermal energy at depth and to obtain scientific information on the internal nature of Kilauea Volcano (Keller, 1974).

The test hole was drilled to a depth of 1262 meters. The upper 490 meters were above the water table and the temperature remained nearly constant at 22°C for this part of the borehole. Below 490 meters, a complicated temperature profile was observed. The bottom hole temperature was 137°C and the temperature gradient in the last several hundred meters of the hole was about 400°C per kilometer.

Of particular interest are the geophysical measurements that were taken in the upper part of the borehole. The rocks to a depth of 323 meters are characterized by thin flows with densities that vary from 1.75 to 2.50 grams/cc, and water contents that ranged from 10 percent to 45 percent. This interval was above the water table and had resistivities of several thousands of ohm-meters. Unfortunately, full-waveform acoustic logs and interval transit time logs were only run at depths greater than 315 meters and no acoustical properties of the near-surface rocks were measured. The fact that the near surface rocks are undersaturated and have relatively low densities, however, may be of significance in the final analysis of the microseismic groundnoise data.

East Rift Zone

Geophysical studies of the East Rift Zone include time-domain electromagnetic (EM) surveys, a gravity survey, a magnetic survey, dipole mapping surveys, and earthquake monitoring. All of these scientific investigations were conducted to gain more information on the relationship of the East Rift Zone to the summit of Kilauea Volcano. The resistivity and earthquake studies are of particular importance.

Electrical Resistivity

The results of a preliminary time-domain electromagnetic survey in the East Rift Zone near Kapoho Crater, another time-domain electromagnetic survey near Puulena Crater, and a dipole mapping survey which included both of these areas are of particular interest because they include the area that was investigated with the microseismic groundnoise method.

Two source bipoles of the dipole mapping survey illuminated the area of maximum groundnoise activity at Puulena Crater (see Figures 43 and 44). One source (bipole source 2) was located along the road from Pahoa to Kaimu, 5 km west of Puulena Crater. The other source (bipole source 7) was located along the road to the University of Hawaii Experimental Agricultural Station, 1 km east of Puulena Crater. Bipole-dipole resistivity soundings from these two sources in the direction of Puulena Crater are illustrated in Figure 45.

The resistivity sounding from source 2 indicates a high near-surface resistivity (>200 ohm-meters), a 50 ohm-meter second layer

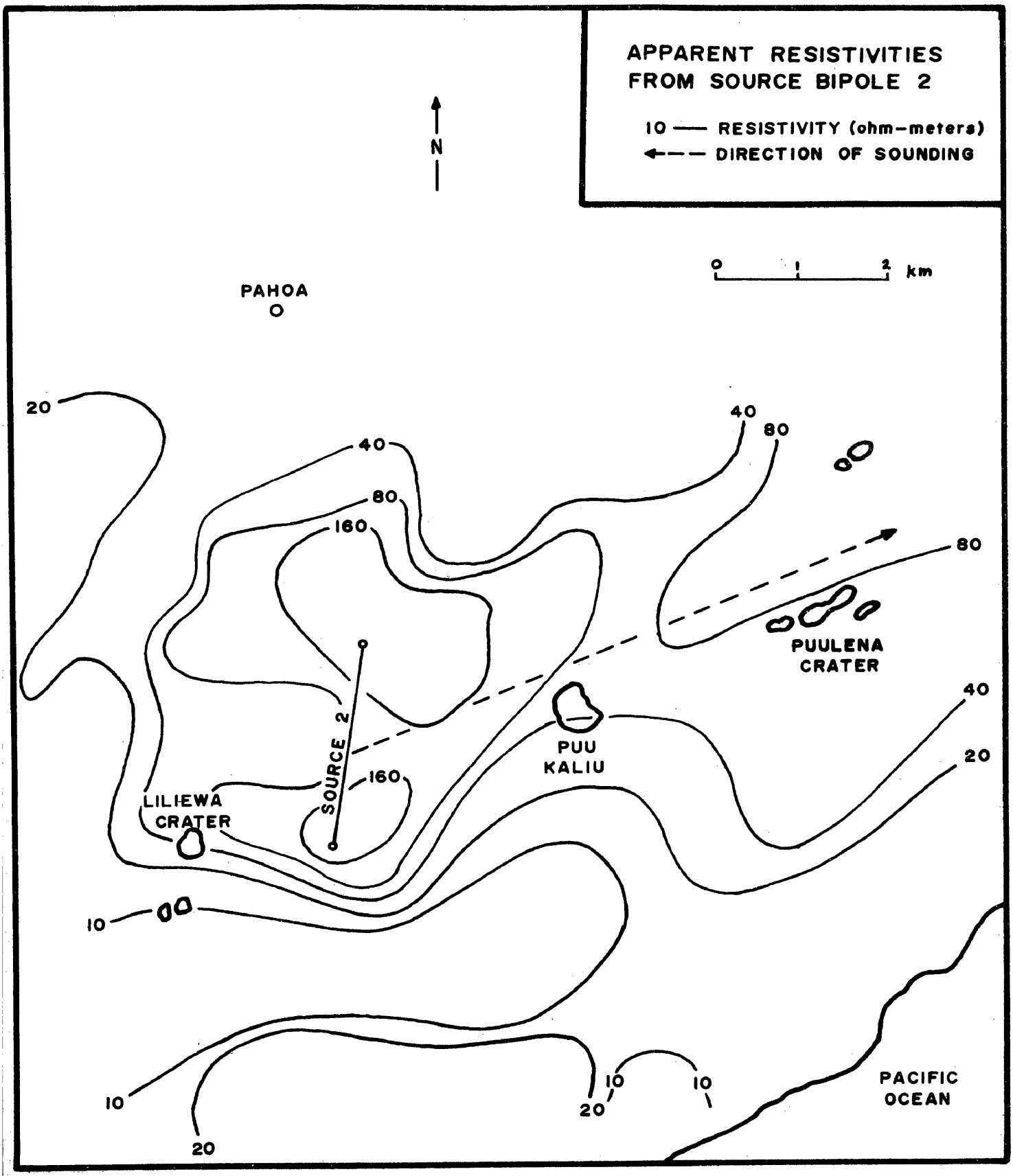


FIGURE 43. Apparent Resistivities Mapped From Source Bipole 2.

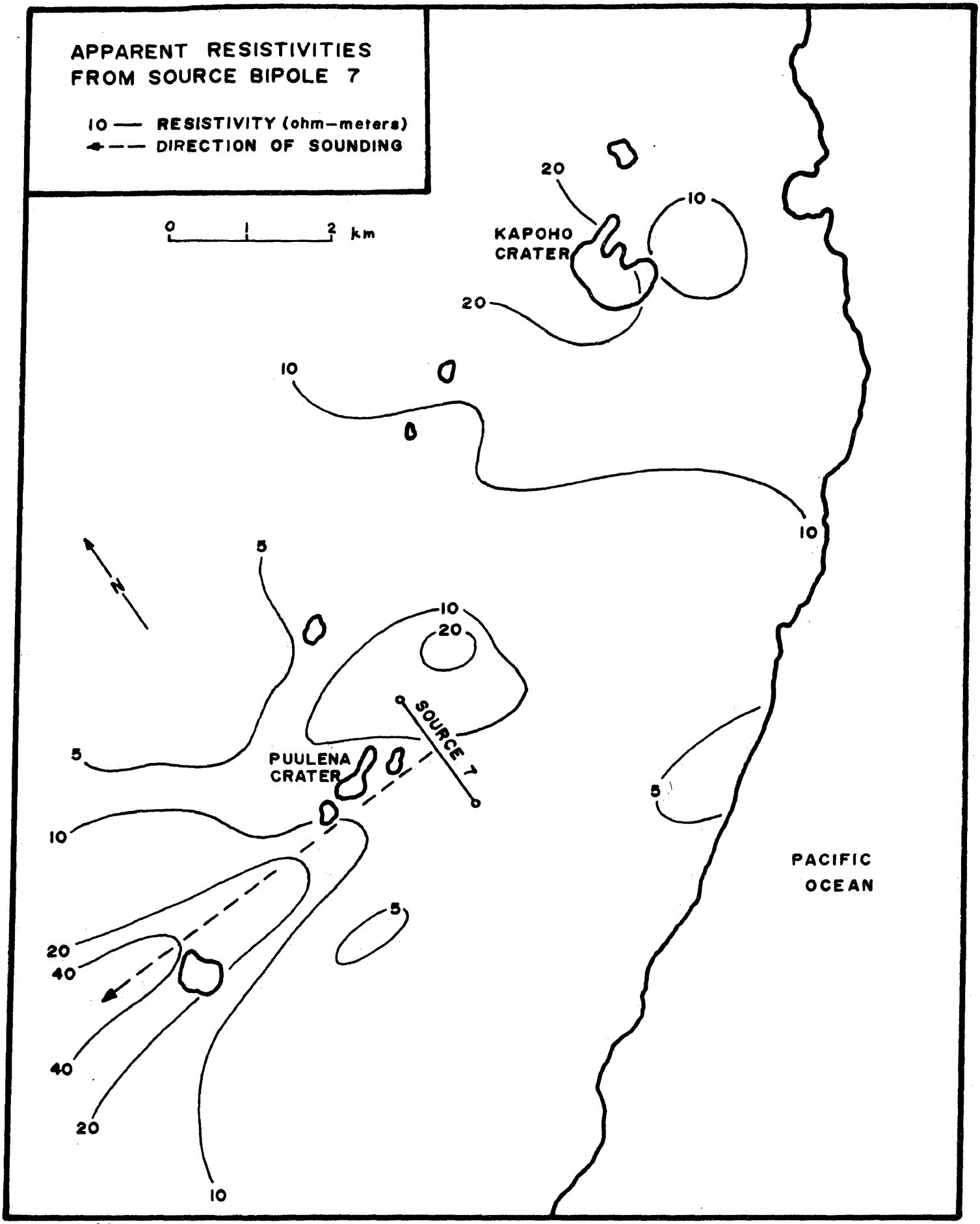


FIGURE 44. Apparent Resistivities Mapped From Source Bipole 7.

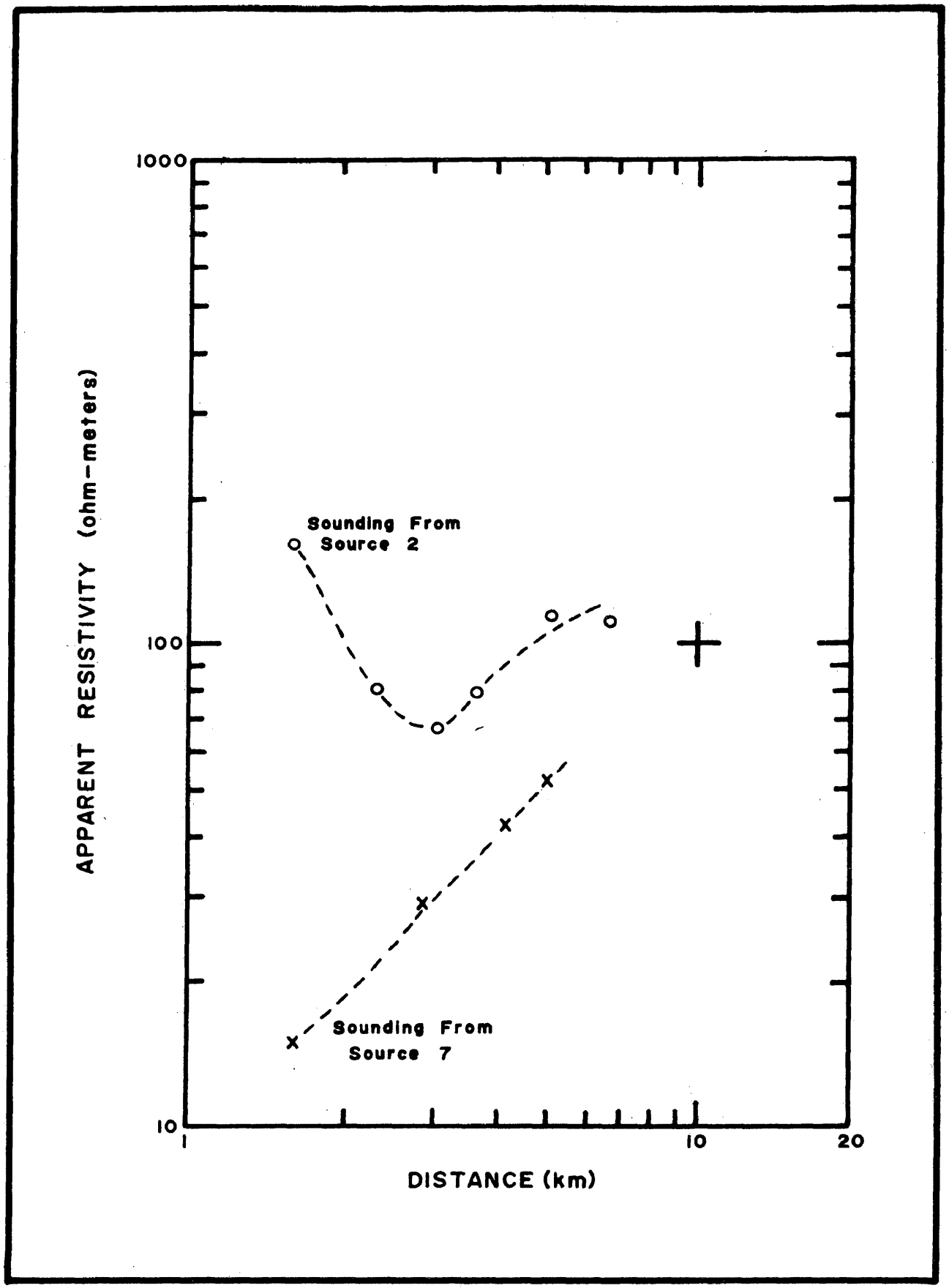


FIGURE 45. Bipole-Dipole Soundings From Sources 2 And 7.

extending to a depth of about 2.5 km, and a high third layer resistivity. The sounding from source 7 indicates a near-surface resistivity of about 10 ohm-meters and a high second layer resistivity (>100 ohm-meters) extending to a depth of at least 5 km.

Analyses of these bipole-dipole soundings in the Puulena Crater area of the East Rift indicate that resistivities tend to increase with depth. This region is either very permeable and surface waters migrate downward and away from this zone, or there are dense rocks with little interconnected porosity at depth. The near-surface rocks at Puulena Crater, however, have low resistivities and probably have high permeability and water saturation.

The electromagnetic sounding data also shows a zone of high resistivities at depth near Puulena Crater (Skokan, C. K., personal communication). Furthermore, this part of the East Rift Zone has a major structural offset which is apparent in the electromagnetic data. The surface expression of the Rift Zone on the east side of Puulena Crater is offset about 1.5 km in a northwest direction.

It is probable that a complex dike system has developed at depth in this major structural break and that the near-surface rocks are highly faulted and fractured as they are at the summit of Kilauea.

The results of a preliminary time-domain electromagnetic survey in the area around Kapoho Crater are illustrated in Figure 46. The apparent resistivities that are mapped in this survey indicate a resistivity high which is located between Puulena and Kapoho Craters. This may also be an area of near-surface dike intrusions.

The higher resistivity zones at depth which are located at Puulena

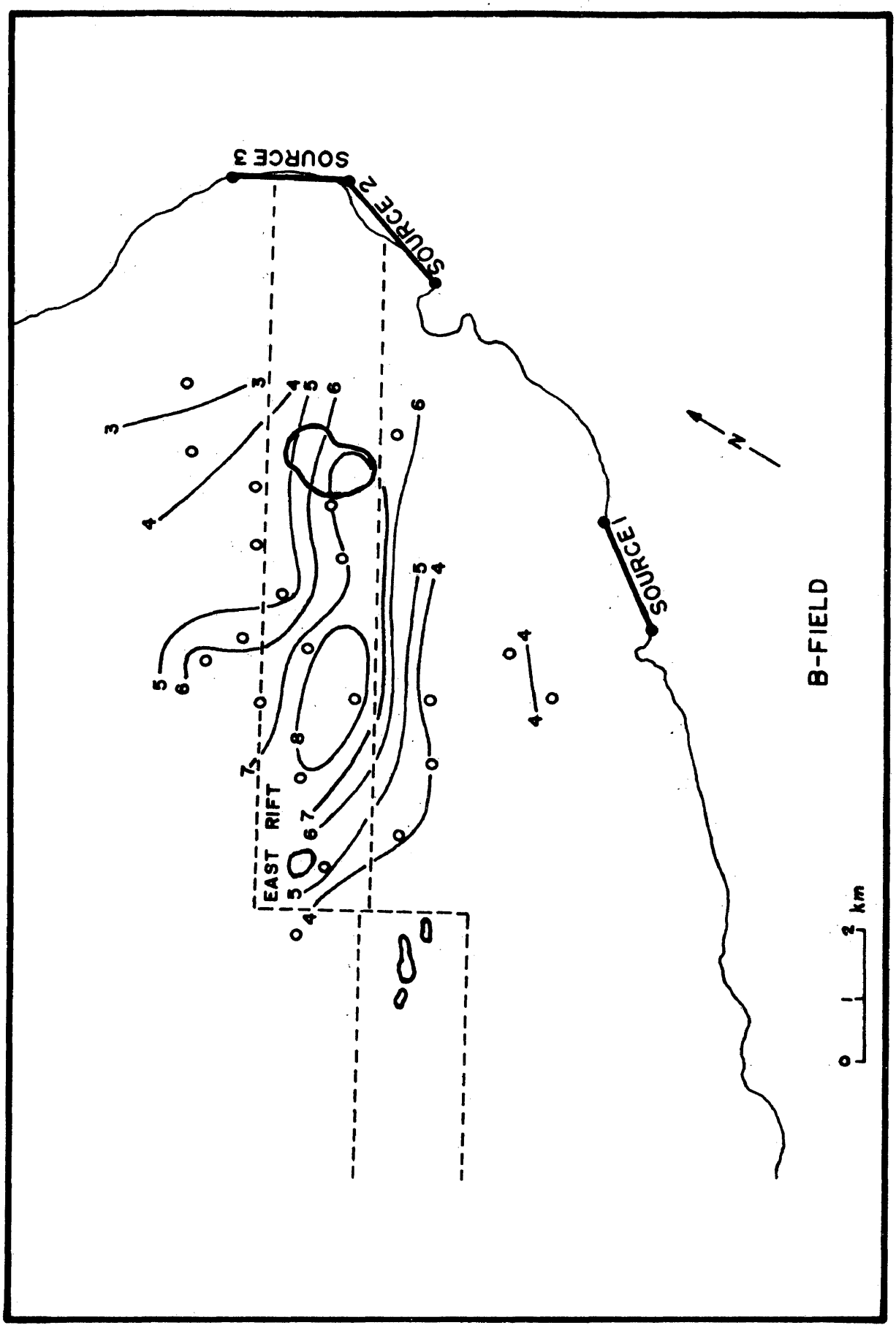


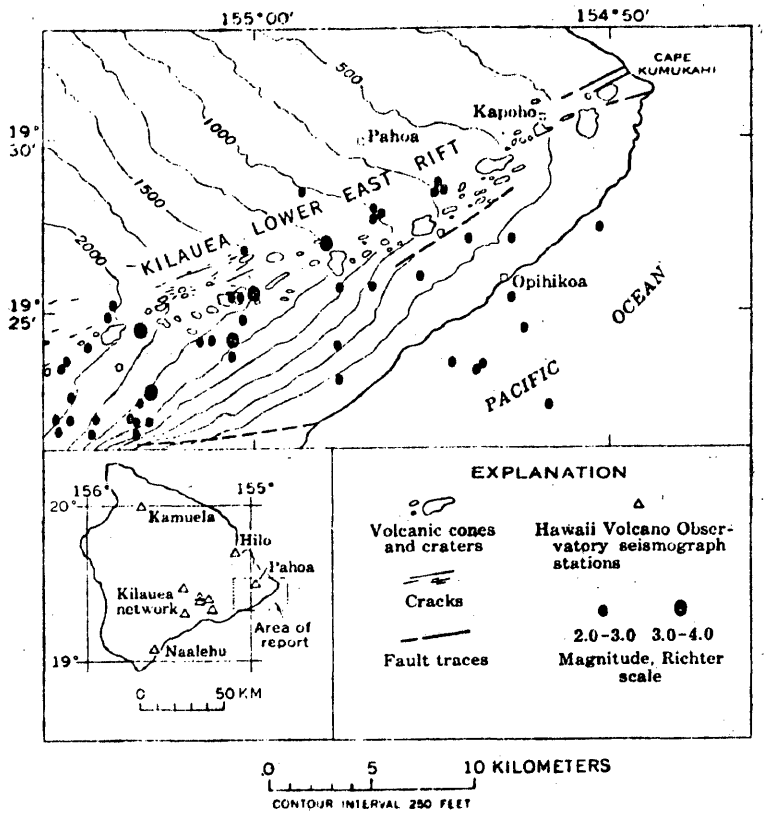
FIGURE 46. Maximum-Voltage Resistivities In The East Rift Zone From A Time-Domain Electromagnetic Survey.

Crater and between Puulena and Kapoho Craters are of particular interest because microseismic groundnoise anomalies are also located in these two areas.

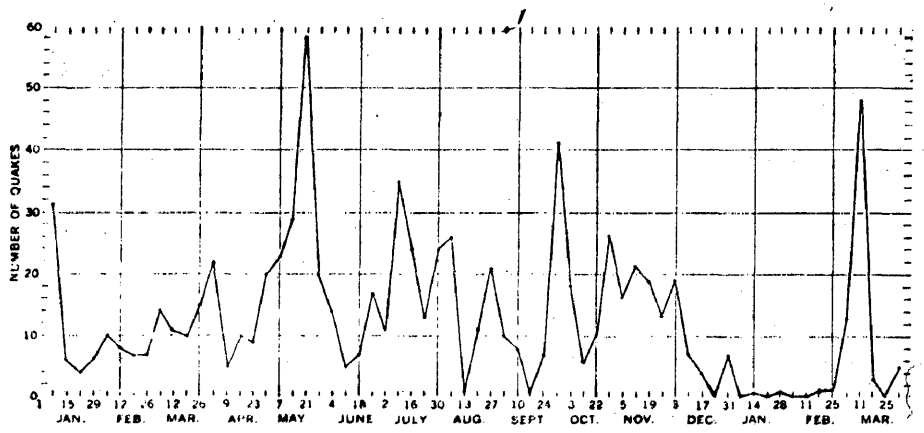
Earthquake Seismicity

Studies of earthquake occurrence in the East Rift Zone indicate that most of these events are confined to the upper 12 km and do not extend into the mantle. Concurrent observations of seismicity and crustal deformation indicate lateral transfer of magma from the summit of Kilauea to areas of eruption in the East Rift Zone.

A seismicity study from January, 1962, to March, 1963, indicated that 840 earthquakes occurred within, and south of, the East Rift. These were recorded on the Pahoa seismograph and they ranged from 0.5 to 4.0 on the Richter Scale. Epicenters of earthquakes with a magnitude of 2.0 or greater are indicated in Figure 47a. One of the striking features of this study was the relationship between Kilauea summit eruptions and seismic activity in the East Rift. Very low seismicity was recorded in the East Rift Zone after a brief eruption in and near Aloi Crater at the summit. Figure 47b illustrates the weekly frequency of earthquake occurrence and the low seismicity period (December to February) after the summit eruption. Other studies such as this indicate that eruptions at the summit reduce reservoir pressures and thus relieve stresses in the eastern flank of Kilauea (Koyanagi, 1964).



(a)



(b)

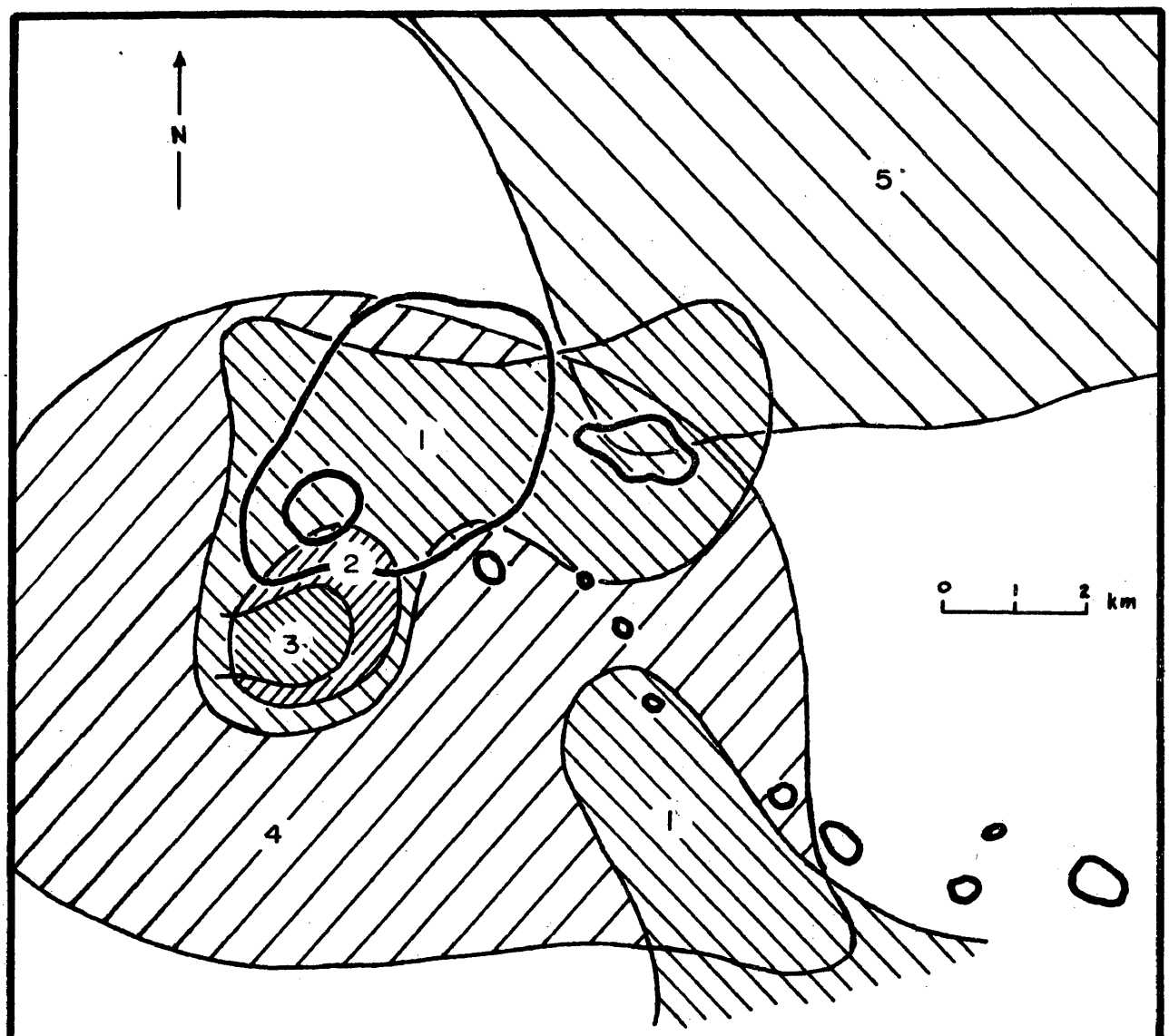
FIGURE 47 . Earthquake Seismicity Of The East Rift (From Koyanagi, 1964)

DISCUSSION OF RESULTS

Kilauea Crater

It is tempting to look at the Kilauea Crater unfiltered groundnoise data (Figure 28) and to speculate that the anomalous groundnoise intensities are directly related to a geothermal system at depth. If we consider the dominant frequencies (Figure 29), however, we see that the high frequency harmonic tremor resulting from the eruption at Mauna Ulu decreases very rapidly in frequency and intensity with increasing distance from the source. Three kilometers from the eruption, the dominant frequency decreases to 0.5 Hz, that of oceanic microseisms. If the groundnoise anomaly is associated with a geothermal system, then the system appears to be characterized by very low frequency (0.5 to 1.0 Hz) microseisms.

If we review the groundnoise data at Kilauea Crater with the idea that variations in rock properties may have caused the anomalous groundnoise intensities, we must look at the results of other geophysical surveys in the area. Figure 48 is a summary of the results from electrical resistivity measurements, shallow earthquake determinations, crustal deformation measurements, and anomalous groundnoise intensities. This



1. Area of maximum groundnoise intensity.

2. Area of maximum shallow earthquake occurrance.

3. Area of conductivity anomaly.

4. Area of maximum ground deformation.

5. Area of high resistivity.

FIGURE 48. Summary of Geophysical Results at the Summit of Kilauea Volcano.

figure shows that the area at the south edge of Kilauea Crater has the highest occurrence of near-surface earthquake activity, a deep electrical conductivity anomaly, a surface expression of greatest crustal deformation, and high microseismic groundnoise intensities.

The shallow earthquakes and surface deformation studies tend to indicate that the near surface rocks are highly faulted and fractured. The Kilauea borehole data suggest that the near surface rocks are undersaturated and have relatively low densities as well. Many well-documented studies in earthquake engineering have shown that local changes in soil profiles at recording sites can produce significant changes in the amplitude and spectral composition of ground motion. If the surface deformation at Kilauea and along the "Chain of Craters" has been very intense, this could have resulted in significant changes in the seismic properties of the rocks and anomalous groundnoise measurements at the summit of Kilauea Volcano.

East Rift Zone

The microseismic groundnoise measurements in the East Rift Zone of Kilauea Volcano indicate anomalous intensities which are probably related to variations in the acoustic properties of rocks. Shallow earthquake activity tends to indicate highly faulted and fractured rocks at shallow depths at the surface expression of the Rift. A summary of the electrical resistivity surveys and the microseismic groundnoise measurements is illustrated in Figure 50. The two areas of anomalous groundnoise intensities and high dominant groundnoise frequencies seem to be associated with near-surface, high-resistivity zones. The higher

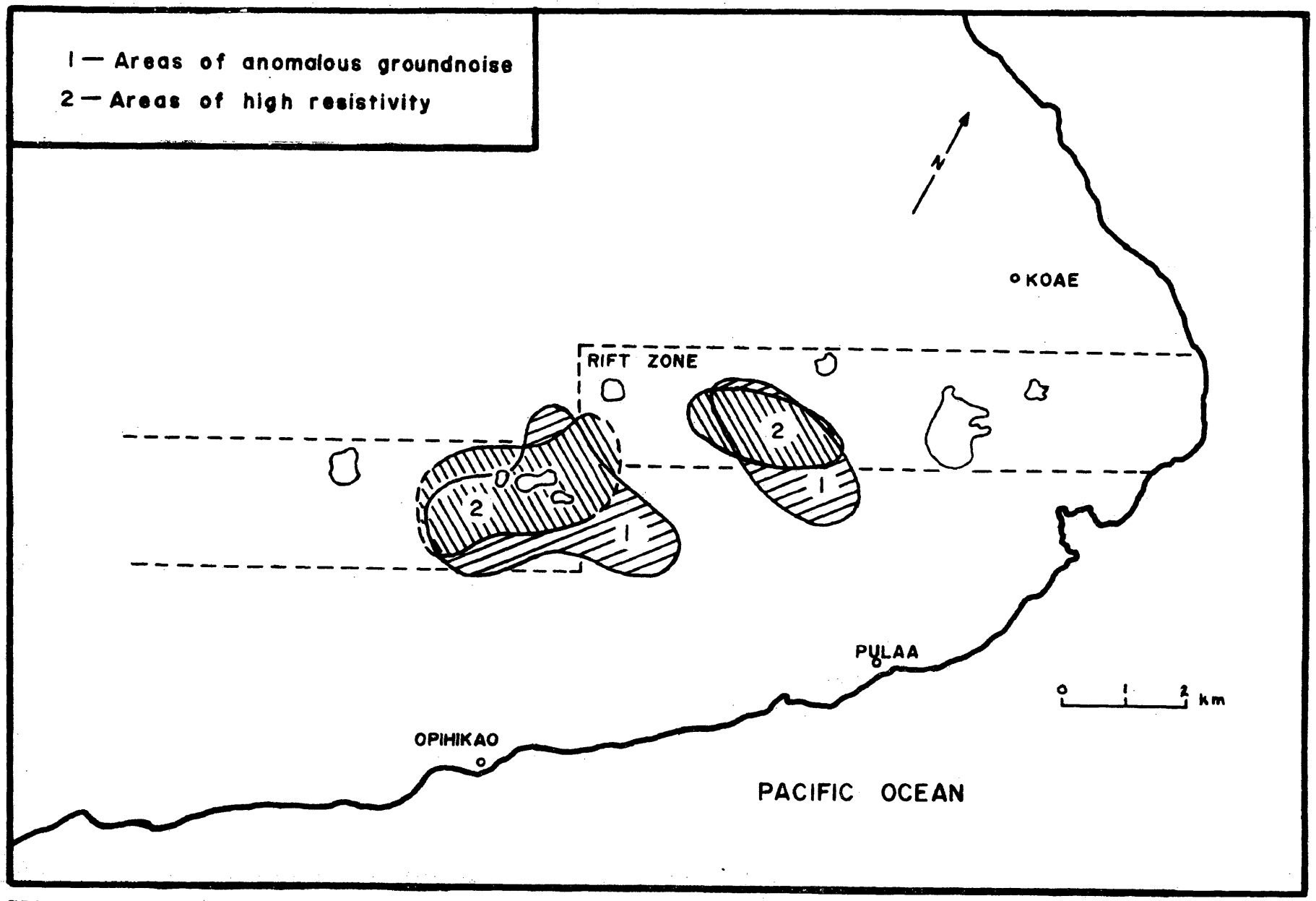


FIGURE 50. Summary of Geophysical Results in the East Rift Zone.

resistivities are probably a result of the intrustion of high-density, low-porosity material into the highly fractured rocks within the Rift. This would explain the reduced attenuation of high frequency microseisms in these areas.

Because the groundnoise amplitudes are relatively high, however, the near surface rocks must have low acoustic impedences. The very-low, near-surface apparent resistivities that were mapped around Puulena Crater (Figure 44) tend to indicate that the near surface rocks do have high, water-filled porosity. If a higher density of faulting and fracturing has caused this, the acoustic impedence of the near-surface rocks would also be reduced. This has probably resulted in higher groundnoise intensities in both of these areas.

The microseismic groundnoise method was used on the Island of Hawaii because variations in groundnoise intensity due to "lithologic noise" would be a minimum. These results indicate that the microseismic groundnoise exploration method, even when used in a relatively homogeneous geologic environment, can give anomalous results which may relate to changes in the acoustic properties of rocks rather than the dynamic characteristics of a geothermal system at depth.

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