

GROUND RADON SURVEYS FOR GEOTHERMAL
EXPLORATION IN HAWAII

A THESIS SUBMITTED TO THE GRADUATE DIVISION OF THE
UNIVERSITY OF HAWAII IN PARTIAL FULFILLMENT
OF THE REQUIREMENTS FOR THE DEGREE OF

MASTER OF SCIENCE
IN GEOLOGY AND GEOPHYSICS

DECEMBER 1980

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Nay, if I understand anything, greater wealth
now lies hidden beneath the ground in the mountain-
ous parts of your territory than is visible and
apparent above ground.

Agricola de Re Metallica, 1556.

ACKNOWLEDGEMENTS

This thesis describes part of the research work which I have had the good fortune to be able to conduct in such an interesting environment as Hawaii. There are several to whom I am indebted for providing me with the opportunity to be able to pursue and to complete this work: Dr's Ralph Moberly, Frank Peterson, John Sinton and Donald Thomas.

I especially thank Dr Thomas, who as supervisor of the geothermal investigations was both colleague and friend and with whom I had many productive discussions.

I thank the members of the committee for wading through this thesis and for their useful comments and criticisms, especially those of the chairman, Dr Sinton.

Thanks are due to Mr. Barry Frankhauser for helpful discussions of the techniques involved as well as permission to use some of his data, and to Mr Kevin Cuff for assistance with parts of the field and analytical work.

To my wife, Marcie, who typed the manuscript and lived through it all, I can only say "vinaka vaka bibi noqu lewa".

ABSTRACT

Exploration for geothermal resources in Hawaii has required adaptation of conventional exploration techniques as well as the implementation of relatively new techniques because of the complexities introduced by the oceanic island environment and the basaltic lavas which form the islands. One exploration technique previously untried in Hawaii, and used to only a very limited extent elsewhere, is the measurement of the concentration of the radioactive gas, radon, within ground gas at shallow depth. The measurement procedure applied utilizes alpha-particle sensitive film housed within an inverted 250 ml container. The device is exposed by burial in the ground for periods of 4 to 5 weeks. The films are easily developed in the laboratory using a constant temperature bath of NaOH which enhances perforations caused by alpha particle tracks enabling them to be counted under magnification. The establishment of numerous measurement stations within an area enables patterns of the concentration of radon to be mapped. To confirm whether this technique could be successfully applied in Hawaii, surveys were conducted in four areas of different characteristics: the summit of Kilauea volcano and the lower part of its active east rift zone, on the island of Hawaii; the lower part of the extinct north rift of Haleakala volcano, Maui and within the extinct eroded Waianae caldera in western Oahu.

The technique was demonstrated to have been successful, and enabled the delineation of anomalous zones associated with the volcanic structures and resulting from increased ground permeability and anomalous subsurface temperatures. The data further indicate the common development of ground gas convection systems within and around these structures. To further assist in determining the controls over subsurface radon, gamma-ray spectrometry analysis of rocks, soils and volcanic sublimates was also made and other properties of soils were examined. These data indicate that elevated concentrations of uranium, the main parent of radon, occur within secondary mineralization within Hawaii hydrothermal systems, and which appear to enhance measured concentrations of radon. Results of the study overall indicate that the radon measured is commonly derived from depths of several 10's of meters, but that in areas of increased permeability and heat measureable concentrations of radon can be transported from depths of several 100's meters. As an exploration technique in Hawaii, such surveys were found to be cost-effective, being both rapid and inexpensive.

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I. INTRODUCTION

This thesis describes the application of surveys measuring the concentration of radon in ground gas for geothermal exploration in Hawaii. The work was carried out as part of the U.S. Department of Energy funded Geothermal Resource Assessment Program of the Hawaii Institute of Geophysics. The surveys herein described were conducted during the period from June, 1978 to December, 1979.

A. Definition of "Geothermal"

It is important at the onset to define the meaning and use of the term "geothermal". This can be satisfactorily done using the following explanations from Ellis and Mahon [1977]:

geothermal energy - refers to the potentially useful energy stored as hot water or steam in favorable geological situations within the upper several kilometers of the Earth's crust;

geothermal system - refers to the particular rock-hydrological situation of the hot water or steam, and which is contained within a geothermal area of finite surface boundaries;

geothermal development - refers to the harnessing of steam and hot water flows, either natural or induced, through drilling into the rock system.

These terms are therefore used specifically, as opposed to the term hydrothermal. Hydrothermal system is a general

term referring to a rock-water system containing high-temperature fluids.

B. Requirements for a Geothermal System

A geothermal system requires several basic properties to enable energy production. Healy [1976] notes that although most of the stored heat is in the rock "... a geothermal system suitable for energy production should contain rocks with adequate permeability, saturated with hot water or steam, within economic drilling distance". This requires geological and hydrological conditions which can form a fluid reservoir and the presence of hot rock to heat the fluids either within or before entering the reservoir. Drilling into the reservoir, or into structural channel-ways (e.g. faults) transporting thermal fluids, and the consequent migration of the thermal fluid to the surface provides the mechanism by which the heat energy is transported.

The most common type of geothermal system is one which is dominated by liquid water (hot water-dominated). Typically this type of system has the following characteristics: recharge of meteoric water over a broad area which then percolates to large depths (often up to 10 km) where it is heated by conduction during migration through hot rock. A variety of heat sources are possible, the most common being, elevated geothermal gradients associated with crustal thinning or mantle upwelling, plutonism or intrusion of igneous bodies.

After heating and the consequent decrease in density, the water rises usually through fractures and fault planes aided by convective forces. If it encounters a permeable zone (lithological or structural) it may then accumulate under reservoir conditions. For this to occur the permeable zone must be confined by impermeable material. This is often an impermeable lithological unit (e.g. shales, indurated tuff) but can also be of similar lithology to the reservoir rocks but of lower permeability.

In many cases, notably in continental environments, faults have penetrated the overlying impermeable strata thus allowing a natural discharge of thermal fluids at the surface (Fig. 1). This creates the typical surface manifestations of geothermal systems such as, hot springs, steaming ground and fumaroles. Summarizing the above, the formation of a geothermal reservoir requires adequate recharge, hot rocks at depth and a zone of permeability which is in some way sealed.

C. Geothermal Environment in Hawaii

In Hawaii, because of the basaltic lavas, the young volcanic terrain and the oceanic island environment geothermal systems are somewhat different from those of areas possessing continental characteristics. One of the most obvious differences is the lack of any widespread surface features of geothermal activity such as hot springs and areas of steaming ground. These features do occur to a limited degree in Hawaii

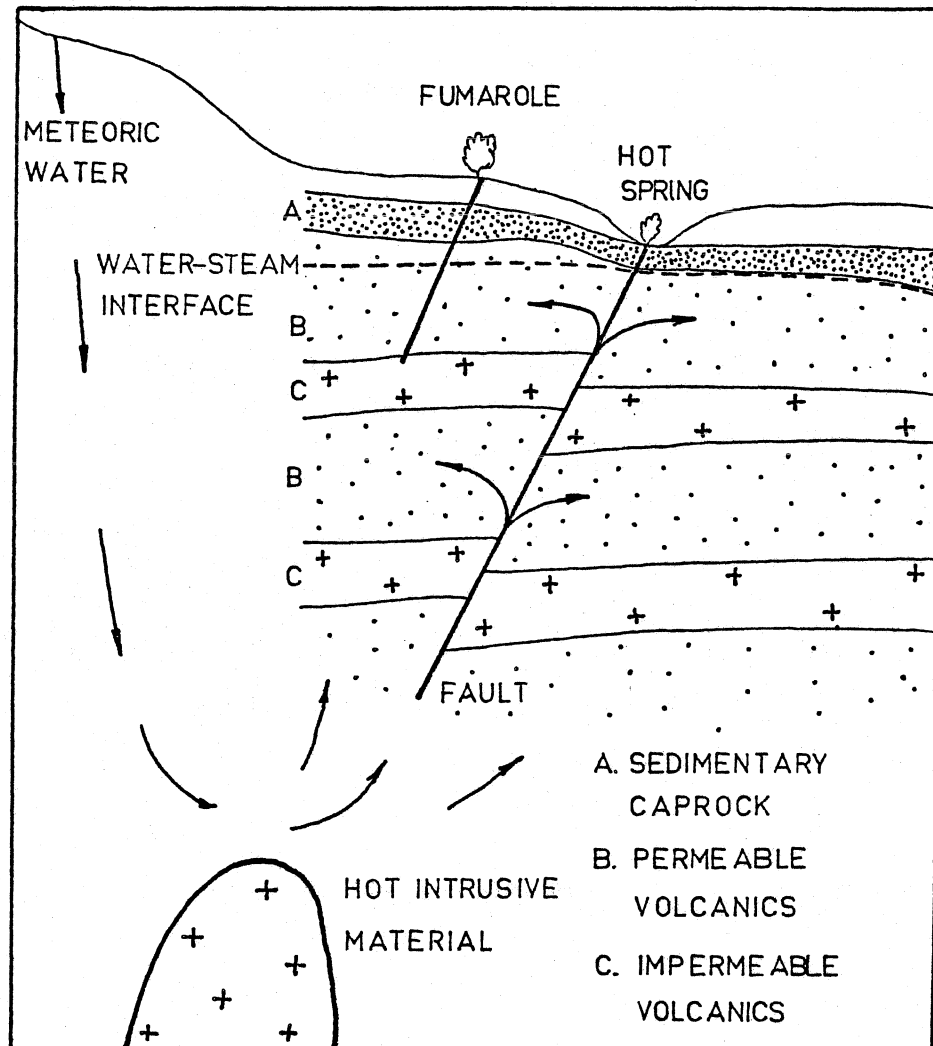


Figure 1. Sketch of main features of a high temperature geothermal system associated with recent volcanism in a continental area [after Ellis and Mahon, 1977].

as part of volcanic hydrothermal systems, but as such are limited to the summits or rift zones of the active volcanoes (Kilauea and Mauna Loa).

The main reason for the lack of widespread surface features of this type, especially in the areas of active volcanism is the very high permeability of the basaltic lavas and the poor development of, or lack of, a soil horizon. Consequently in most cases the recharging groundwaters move rapidly to the basal aquifers and even if heated at depth cannot migrate surfaceward. Those hot springs and thermal water seepages that do occur are limited to coastlines and are a result of the thermal groundwaters rising towards sealevel along the Ghyben-Herzberg lens (the basal saline-freshwater transition zone) (Fig. 2). Thermal water has also been reported to occur within cracks on the east rift of Kilauea [Macdonald, 1973]. As a consequence of the form of migration of these thermal waters they are highly mixed with other groundwaters and are of low temperature (≈ 28 to 35°C).

To date one geothermal area has been determined to exist in Hawaii. It is at Puna on the lower east rift of Kilauea Volcano, and has been confirmed by the drilling of geothermal well HGP-A [Helsley, 1977; Kroopnich et al, 1978]. The characteristics of this, or other potential geothermal systems, have not yet been unequivocally determined, however,

several basic features can be postulated for such systems in Hawaii:

(i) the source of heat to the groundwaters may be shallow in comparison to many continental systems, perhaps on the order of 3 to 5 km;

(ii) the rocks containing the heat are probably exclusively intrusive, although the immediately surrounding ancient flows may obtain some heat by conduction;

(iii) the geological conditions which enable a reservoir to form are probably largely structural, not lithological and essentially the same as those features which cause entrapment of shallow groundwater, such as dike complexes and impermeable layers. Impermeable beds which could act as sealing layers over permeable zones saturated with hot water are most likely to be massive, unfractured ancient flows, ponded lavas or perhaps ash layers. The possibility also exists of an associated decrease in permeability from chemical sealing by the deposition of secondary minerals (e.g. silica, anhydrite, calcite) from thermal fluids. A "caprock" in the sense of that found in many continental type geothermal systems is not likely to occur in Hawaii;

(iv) because of the common high recharge and the high ground permeability, water-dominated systems are the most likely to develop;

(v) the geothermal fluids are most likely to be basal and dike-confined groundwaters of meteoric origin, although a

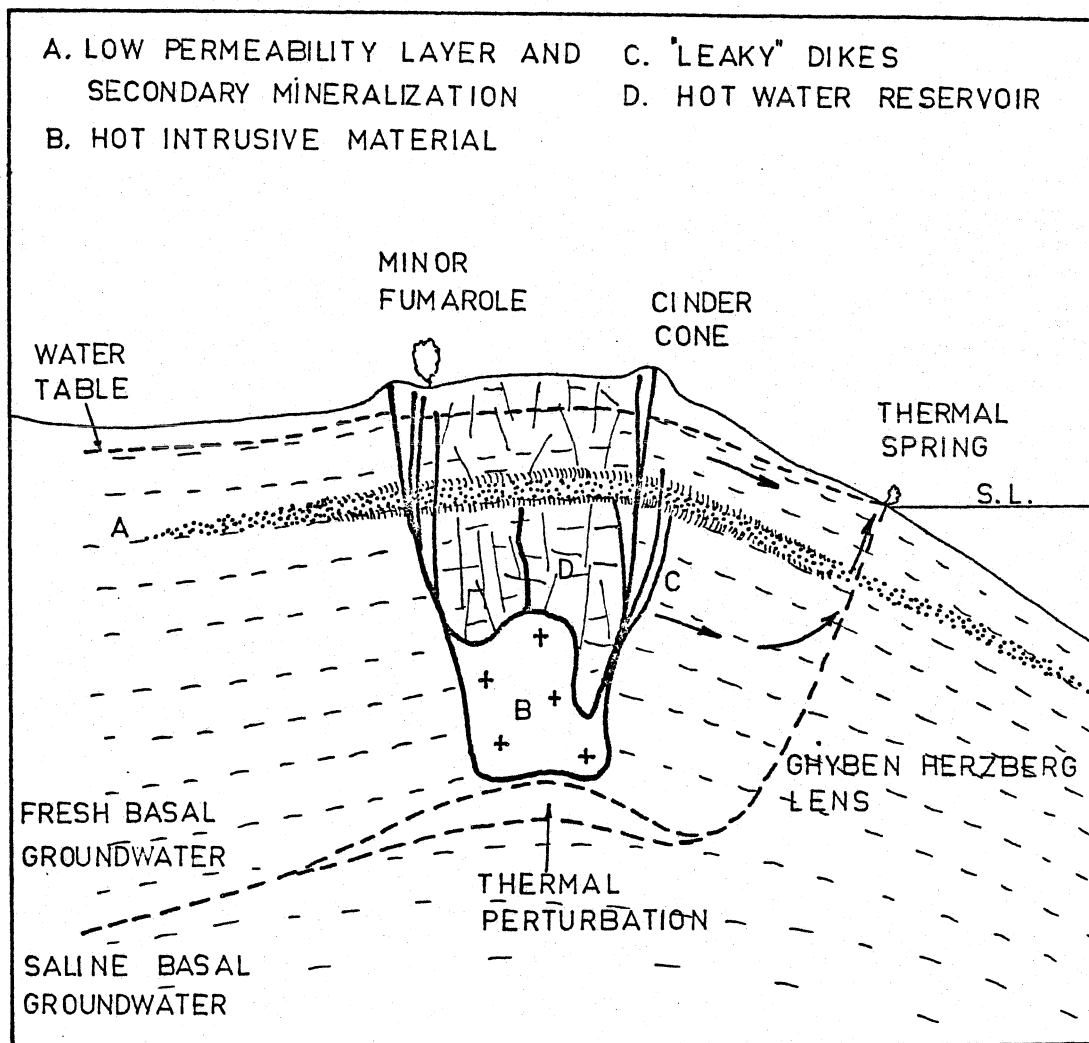


Figure 2. Sketch of main features of a rift zone geothermal system in Hawaii (cross-section). Permeability of reservoir rocks is largely due to fracturing. Recharge of groundwater is both locally and from higher elevation up-rift. Arrows indicate leakage of thermal groundwater.

significant saline water component could be expected from encroachment of deep groundwater of marine origin due to thermal perturbation of the Ghyben-Herzberg lens and certainly after development has begun. An increase in the saline water component of up to 50 to 70% would not appear unreasonable after prolonged development.

D. History of Geothermal Exploration in Hawaii

Previous to 1978 geothermal exploration in Hawaii was largely concentrated on the lower east rift zone of Kilauea Volcano. This area was considered to be highly prospective from several criteria: (a) the occurrence of several historic eruptions such as those in 1790, 1840, 1955 and 1960; (b) numerous dike complexes presumed to occur within the rift; (c) several coastal thermal springs, warm water in cracks and minor fumaroles; and (d) relatively flat topography and easy access.

During 1961 four shallow wells were drilled within the rift zone by Hawaii Thermal Power Company. Two wells (1 and 2) were located close to steaming fissures produced by the 1955 eruptive activity, but were abandoned because of loss of drilling tools in the holes: well 2 produced low pressure steam and had a maximum downhole temperature of 105°C. Wells 3 and 4 extended below sea level, but bottom water temperatures were only 93° and 43°C, respectively [Macdonald, 1973].

From 1972 to 1977 the Hawaii Geothermal Project

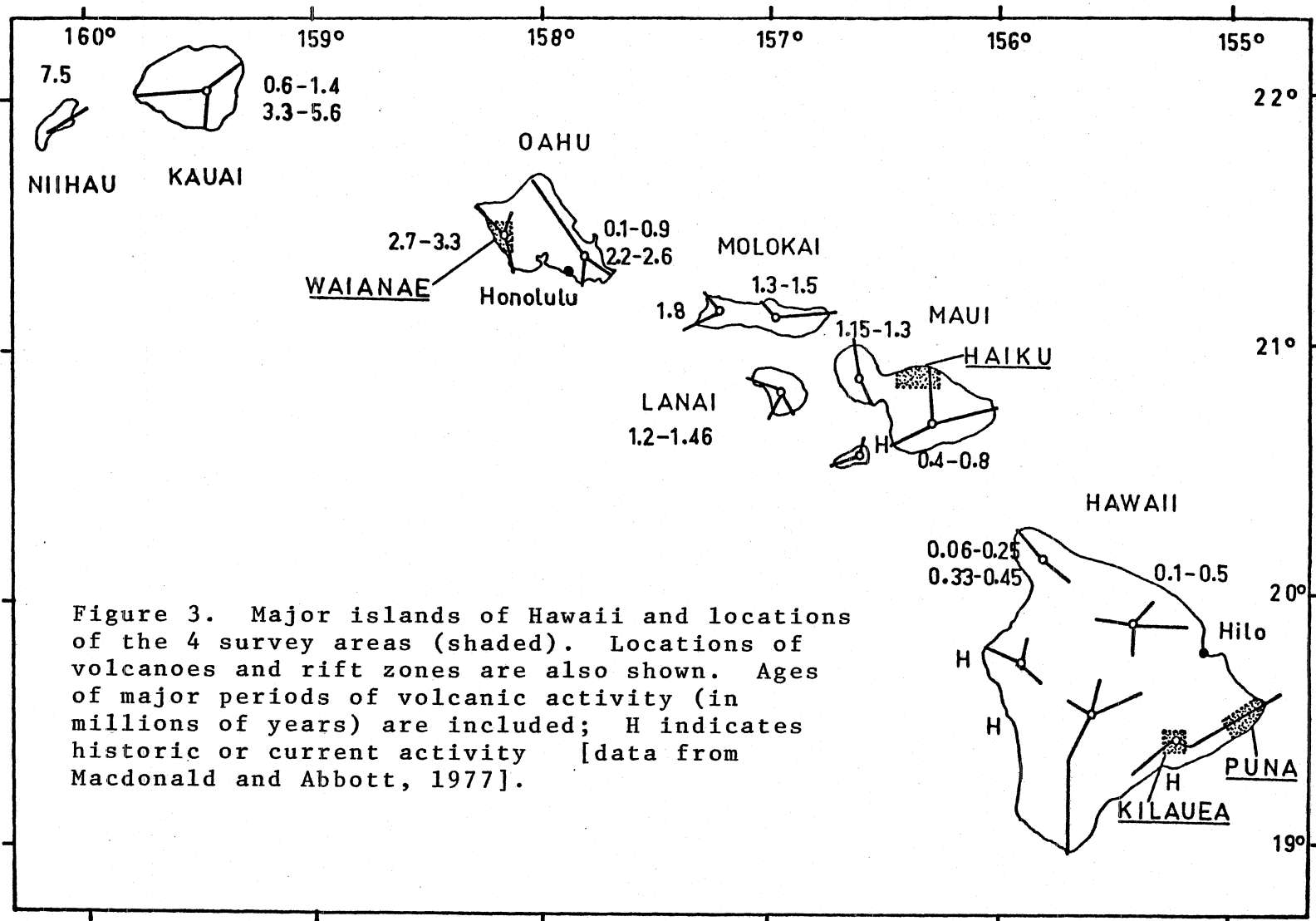
(funded by the State of Hawaii, National Science Foundation, Energy Research and Development Administration and the Hawaiian Electric Company) of the Hawaii Institute of Geophysics carried out geothermal exploration surveys. The majority of these surveys were in the Puna area and largely utilized geophysical techniques. The earliest conducted was an airborne infra-red imagery survey over the rift zones of Kilauea, Mauna Loa and Hualalai [Abbott, 1974]. Other survey techniques included electrical resistivity and self-potential [e.g. Klein and Kauahikaua, 1975; Keller et al 1977; Kauahikaua and Klein, 1978] micro-seismic [Suyenaga et al, 1978], ground magnetics [Norris, 1976], and gravity surveys [Furumoto et al, 1976]. Groundwater temperature profiles were measured in shallow groundwater wells in the area [Epp and Halunen, 1979] and chemical analyses and stable isotope determinations were made of shallow groundwaters [MCMurtry et al, 1977].

The data gathered by the above surveys, overall, showed anomalous results within the Puna area, however, correlation of the anomalies determined was not always good and led to different interpretations of the structure and extent of the indicated geothermal reservoir. Assessment of the results obtained, consideration of geological and hydrological conditions and land availability led to the selection of a drilling site within the rift structure [Macdonald, 1976]. The well, HGP-A, was completed in April, 1976, to a depth of 1953 m. Downhole temperatures were surprisingly high, on the

order of 300 to 320°C below about 1100 m, with a maximum temperature of 358°C. A 3.0 MW generator will be installed on this well during 1981. In comparison to these temperatures, other developed geothermal systems in the world have reservoir temperatures commonly in the range of 240 to 280°C. Reservoir temperatures in Icelandic geothermal areas (a similar active basaltic environment to Hawaii) are also higher than most systems and from around 280° to nearly 300°C [Kristmannsdottir, 1975].

A limited number of other surveys were also carried out in other parts of the state from 1972 to 1976. The majority of these were on the island of Hawaii with a concentration of geophysical surveys on the summit and upper rift zones of Kilauea [e.g. Jackson and Keller, 1972; Zablocki, 1976] as well as reconnaissance surveys on the lower rift zones of Mauna Loa and Hualalai. Some of these studies, however, were more of a volcanological nature, and not specifically directed towards geothermal exploration. Several preliminary investigations were made during that time in the eroded caldera complexes on the island of Oahu: Koolau in eastern Oahu [Furumoto, 1976; Souto, 1978] and Waianae in western Oahu [Tasci, 1975].

To date, exploratory drilling for geothermal in Hawaii has been limited to the island of Hawaii. In addition to the four shallow 1961 holes and HGP-A, a research hole funded by the National Science Foundation was drilled in 1973 on the



summit of Kilauea Volcano to a depth of 1262 m (hole bottom 160 m B.S.L.). The highest downhole temperature measured was 137°C at the bottom, where the temperature gradient reached 370°C/km [Zablocki et al, 1974]. The hole has subsequently been abandoned. During 1979 two deep holes (1676 and \approx 1828 m) were drilled by a private combine in the Puu Waawaa area on the north flank of Hualalai volcano, west Hawaii. Neither hole encountered groundwater of above ambient temperatures and consequently do not possess geothermal potential. During 1980 another private group has begun drilling a deep geothermal well at Opihikao in the Puna area, 4.3 km west-south-west of HGP-A. No information about this well is as yet, available. Several other industrial concerns are currently negotiating with land owners in the Puna area and further drilling is anticipated there over the next two years.

In addition to the above investigations, the U.S. Department of Energy funded Geothermal Resource Assessment Program of the Hawaii Institute of Geophysics, began a state-wide assessment in 1978. The aim of that investigation was to define all potential geothermal areas within the State of Hawaii. These areas have now been classified according to indicated potential [Thomas et al, 1979; Cox and Thomas, 1979,a] and investigations are currently being conducted in the most prospective [e.g. Cox et al, 1979; Mattice and Lienert, 1980; Thomas et al, 1980; Cox and Cuff, 1980]. Another aspect of this program has been continued studies of both geothermal

well HGP-A and the Puna reservoir [Thomas, 1980; Kauahikaua et al, 1980; Cox, 1980,a]. The current emphasis of this program, especially for investigations planned for 1981, is to carry out more detailed investigations on the island of Maui.

E. Criteria for Using Radon and Objectives

During exploratory investigations in Hawaii an attempt has been made to adapt techniques to, or utilize others which would be suitable for local conditions. Two such techniques were believed to be the determination of the concentration of radon in ground gas at shallow depth and the concentration of mercury in soil. The basis of using these two elements is that they are known to occur in association with geothermal and hydrothermal systems, are capable of migration within a vapor phase and can be measured at the surface. Surveys utilizing these two elements are considered to be complementary and were consequently carried out in conjunction.

As discussed below, elevated concentrations of radon relative to surrounding values are caused by increased ground permeability which allows a greater vertical migration of ground gas. This migration is indicated to be further enhanced by the presence of elevated temperatures. Therefore, surveys in which the areal concentrations of radon in ground gas are mapped provide the potential of locating permeable zones associated with anomalous subsurface temperatures.

Mercury is well documented as being associated with both terrestrial and submarine hydrothermal systems [e.g. Bostrom and Fisher, 1969; U.S. Geological Survey, 1970; Weissberg and Rohde, 1978; Cox 1980, b] and has been applied successfully as an exploration tool for locating geothermal systems [Matlick and Buseck, 1975; Klusman and Landress, 1979]. Consequently, elevated concentrations of soil mercury in areas in which radon anomalies have been measured further substantiate the existence of anomalous subsurface temperatures.

The objective of this study was to utilize the above properties and to develop and apply a geothermal exploration technique that was inexpensive, easy to implement and mobile, and that was not greatly influenced by the inherent difficulties of working in areas of young basaltic rocks and complex hydrological conditions. This study describes the radon measurement technique used, its application to geothermal exploration in Hawaii, as well as the controls over the release of radon and its subsurface migration.

II. AREAS OF FIELD SURVEYS

A. State-wide Assessment and Selection of Study Areas

The state-wide regional assessment [Thomas et al, 1979] utilized all available on-file geological, geophysical and geochemical data to define areas which have geothermal potential and classify them in order of priority. An important part of this assessment was the use of on-file groundwater chemical data from various State and Federal agencies. A large amount of such data had been collected over the years due to the dependence of Hawaii on groundwater for water supply purposes.

Interpretation of these chemical data in terms of geothermal indicators was not, however, straightforward. As these groundwaters are relatively shallow and mixing of waters from different aquifers has often occurred they could not be assessed in the same manner as hot spring discharges. This consequently precluded the use of the standard quantitative chemical geothermometers such as SiO_2 , Na-K and Na-K-Ca [e.g. Fournier and Truesdell, 1973; 1974] by which estimates of reservoir temperatures (or the temperature of the last water-rock equilibrium) can be made. Use was therefore made of SiO_2 concentrations, groundwater temperatures and the ratio of Cl/Mg ions as qualitative indicators [Cox and Thomas, 1979, a; 1979, b].

Based on conclusions reached from the above assessments four areas were selected in which to carry out the radon (and mercury) surveys for this study. The selection of these areas also considered the criteria for geothermal reservoir formation previously described, and the need for a diversity of environments in which to assess the success of the technique. The areas selected were (Fig. 3):

(a) Puna area lower east rift of Kilauea Volcano, island of Hawaii - an active rift zone within which a geothermal system had been determined;

(b) Haiku area, lower north rift of Haleakala Volcano, island of Maui - an extinct(?) rift zone;

(c) Summit of Kilauea Volcano - an active caldera; and

(d) Lualualei Valley, Waianae Volcano, island of Oahu - an extinct, eroded caldera.

B. Description of Areas Studied

To assist in understanding the environment of, and conditions existing in each of the survey areas, a summary of the main features follows:

1. Puna Area

This area, in the southeast of the island of Hawaii experiences a mean annual precipitation of from 185 to 375 cm. Mean annual temperatures range from 20 to 24°C. The area is of relatively low relief. The highest part of the area is around 280 m A.S.L. on the crest of the rift zone, which

trends roughly eastward from the summit of Kilauea, through the center of the area surveyed. The center of the area is 25 km from the summit. Immediately inland, to the north and northwest of the rift the topography remains relatively flat but in the south and east decreased fairly rapidly to sea level. Along the rift zone itself, soil development is variable but is generally poor or non-existent due to the extensive covering of young (and historic) lava flows. Over areas of older flows soil development increases, reaching a thickness of 0.5 m of organic soil in some areas away from the rift.

The Puna area is built up of flows of the Puna Volcanic Series, composed of tholeiitic basalt containing varying amounts of olivine phenocrysts, which erupted from fissures along the rift. The source of these lavas is believed to be a magma chamber below the summit of Kilauea. This lower section of the east rift is still active and experienced historic eruptions in 1790, 1840, 1955 and 1960, and along the trace of the rift ejecta has accumulated around eruptive vents forming spatter cones. The Puna Volcanic Series lavas are typical pahoehoe and a'a flows of varying permeability.

The water table in the area occurs at depths commonly in the order of 80 to 320 m, and in several shallow wells warm temperatures (32 to 93°C) have been measured. Several naturally occurring areas of steaming ground and minor fumaroles occur along the trace of the rift and there are

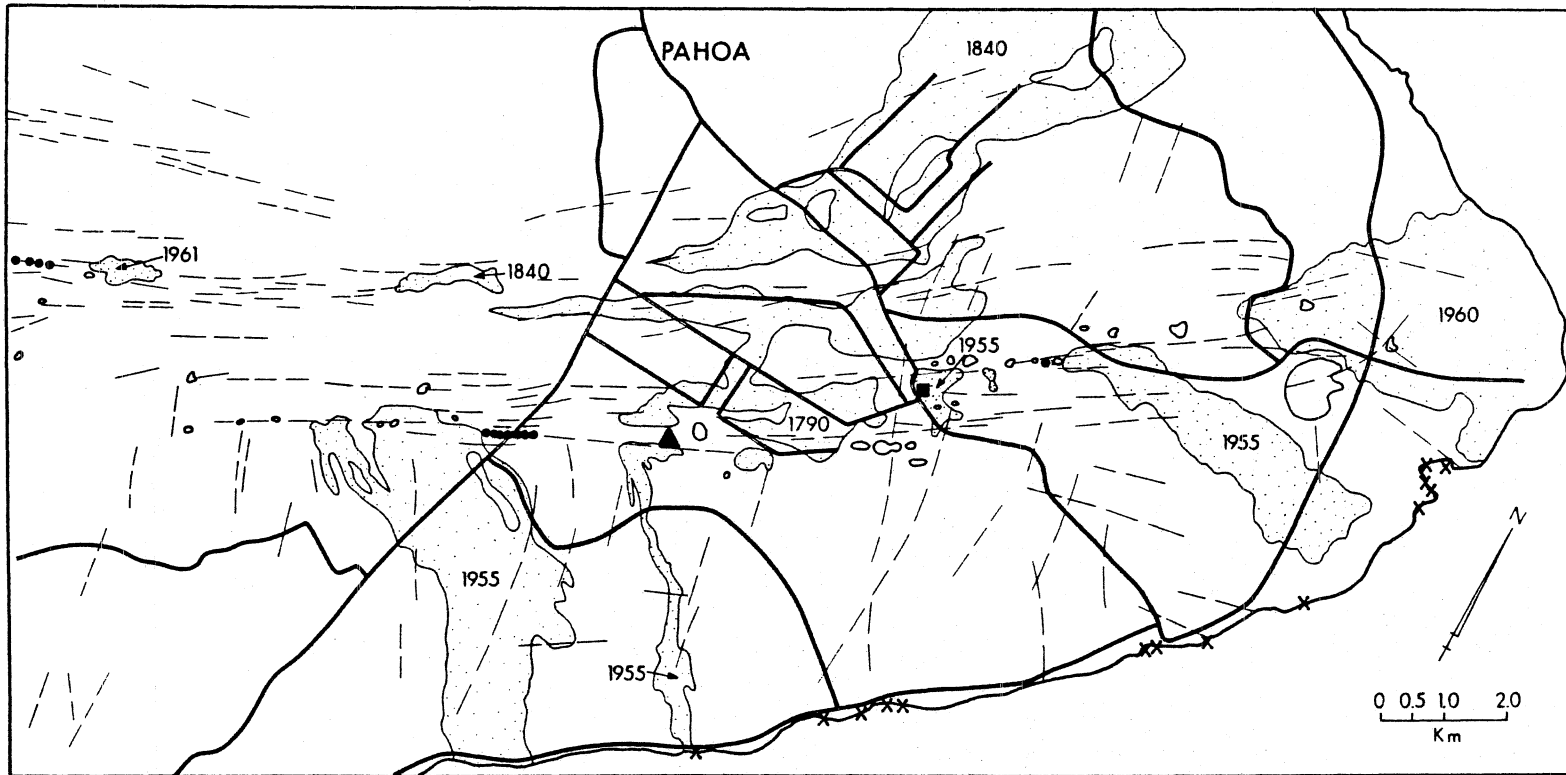


Figure 4. Puna survey area, showing air photo lineations, eruptive features and historic flows. Steaming areas (closed circles) indicate the active part of the rift structure. Geothermal well HGP-A is shown by a square; the approximate location of a well currently being drilled, by a triangle. Thermal springs, or airborne infra-red anomalies [Fischer et al, 1966] are shown by crosses; main roads by a double line.

several seepages of warm freshwater along the southern coast.

The locations of eruptive features suggest the rift is 3 to 4 km wide at the surface. Fractures and cracks are common within the rift, being largely parallel to it but also transverse (Fig. 4). Near the center of the survey area the trace of the eruptive features is offset approximately 0.8 km, immediately east of which the HGP-A well was drilled. This offset was also expressed in some of the geophysical surveys, notably self-potential, and may indicate transverse faulting [Zablocki and Koyanagi, 1979]. The feature may be related to the "buttressing" effect of Mauna Loa volcano and the possible existence of an ancient rift zone of Mauna Loa underlying the younger flows in the eastern part of Puna [Madconald, 1973]. Of interest is the occurrence of an apparent right-lateral offset approximately 12 km to the west, up-rift. Overall, expansion along the east rift largely takes place towards the south, with down-dropping of fault blocks in that direction. This expansion may be by a series of en echelon fractures with the major structure of the rift dipping southward and graben-type structures forming within the rift between zones of faulting [Moore and Krivoy, 1964].

2. Haiku Area

The Haiku survey area is in the northern part of east Maui and is centered over the lower part of the north rift of Haleakala, approximately 25 km from the summit. The area is

on the margin of the windward side of east Maui and has an average annual rainfall of 102 to 203 cm. The western part is relatively flat ranging in elevation from 60 to 150 m but elevation increases rapidly eastward to 350 m at the topographic high over the rift zone. Much of the coastline is rugged, with lava flows forming steep cliffs. Within the south of the area the elevation increases rapidly forming the upper flank of Haleakala. The western section is largely under sugarcane cultivation, and the higher eastern part, is covered by virgin forest. Air temperatures are variable depending on altitude but are commonly 22 to 24°C. Weathering of the lavas has produced a mature soil profile in much of the area, which reaches 4 m in thickness, and weathering of the underlying lavas reaches depths of 5 to 10 m.

Haleakala is considered a dormant volcano having no historic eruptions from the summit or the north and east rift zones, but experiencing a fissure eruption on the lower southwest rift in about 1790. The age of the youngest activity on the northern rift was estimated to be mid-Pleistocene by Stearns and Macdonald [1942], however, subsequent radiometric dating [McDougall, 1964] provided an age of approximately 0.8 to 0.4 m.y. for the Kula Volcanic Series, the last flows erupted from that rift. The Kula lavas cover the area and are predominantly hawaiite a'a flows with thick clinker beds, with lesser amounts of alkalic olivine basalt. The total thickness of this

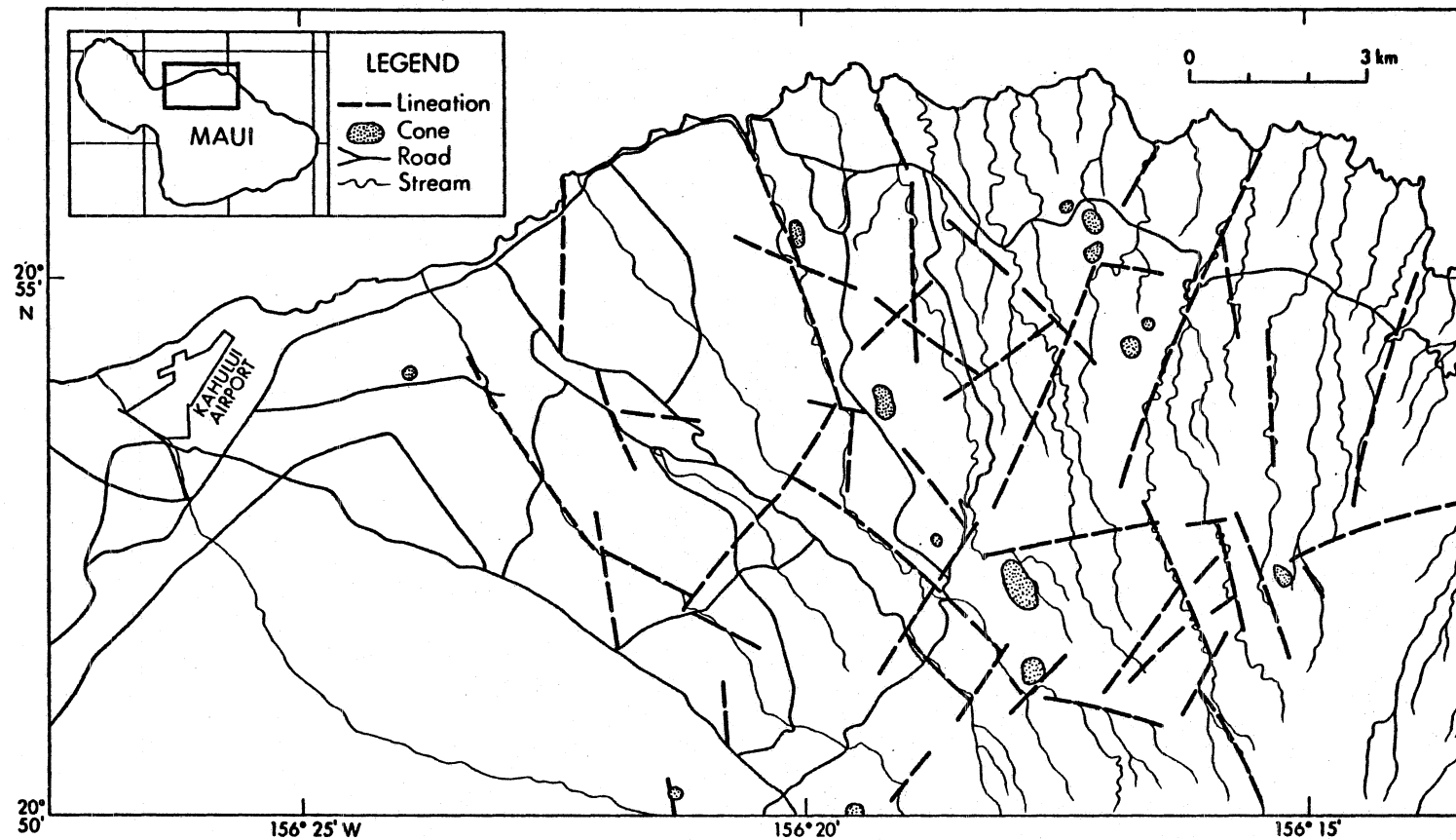


Figure 5. Haiku survey area, showing main roads and streams. The location of the rift zone is defined by two parallel trends of eroded cinder cones. Broken lines represent major faults and air photo lineations.

formation in the area is in the order of 50 to 100 m. These lavas are relatively permeable and overlie the highly permeable Honomanu Volcanic Series that forms the basal rocks of east Maui. The Honomanu lavas are dominated by thin pahoehoe flows of tholeiitic olivine basalt, 3 to 22 m thick and provide much of the groundwater to east Maui [Stearns and Macdonald, 1942]. These lavas are exposed in some of the deeper stream valleys.

The surface expression of the rift is of two trends of eroded cinder cones along a broad topographic high (Fig. 5), which is dissected by numerous north-south stream valleys. Many of these valleys appear to be controlled by faults or zones of fracturing. Minor extrusions of Kula lavas through isolated fissures also formed several small cones off the main trend of the rift. Air photo lineations and field inspection suggest that this lower part of the rift zone is bounded by well-developed faults and that the intervening area is relatively highly fractured. Some vertical movement may have occurred on several of the prominent faults. The formation of the cinder cones generally appears to have been associated with the faults.

3. Kilauea Summit

Kilauea is one of the two currently active volcanoes in Hawaii, and has developed on the southern flank of the other, Mauna Loa. The summit area of Kilauea is around 1160 m A.S.L. and as a consequence distinct windward and leeward climatic

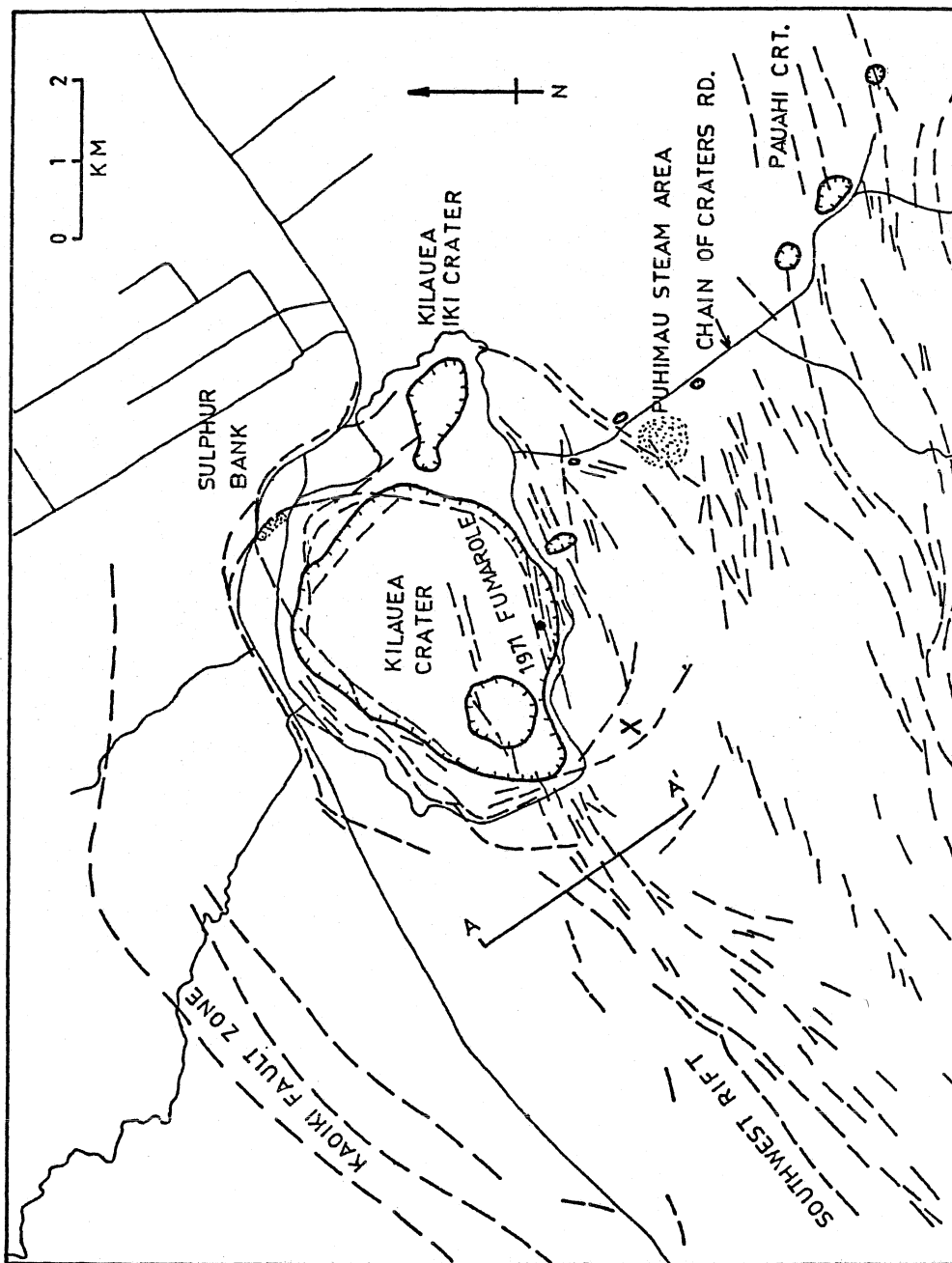


Figure 6. Kilauea summit survey area, showing main faults and cracks. The upper part of the east rift follows the line of craters along Chain of Craters Rd. X denotes the location of the 1973 drillhole; A-A' the traverse in Section VI. B. Roads are shown by unbroken lines.

conditions exist on both the upper slopes and the summit. Annual rainfall on the summit is 100 to 130 cm and the mean temperature 15 to 16°C. Due to both climatic conditions and the range of age and type of eruptive material the "soil" on the summit varies widely. To the northeast of the caldera thick rich soils give rise to tropical rainforest whereas to the south and southwest, barren ash and historic flows cover large areas. Between these extremes a gradation of soil development exists.

The rocks of Kilauea [from Macdonald and Abbott, 1977] are divided into the older Hilina Volcanic Series and the younger Puna Volcanic Series, separated by approximately 10 m of Pahala ash. Both series consist of pahoehoe and a'a flows of tholeiitic basalt, mainly olivine basalt and oceanite, and associated cinder and spatter cones and ash deposits. The caldera of Kilauea is 4.8 x 3.2 km in dimension adjacent to which is the smaller crater of Kilauea Iki. Much of the inner crater of Kilauea is bounded by abrupt fault scarps. The collapse crater of Halemaumau occurs in the southwest of the crater floor and has been the main focus of surface activity of Kilauea over the last 150 years. The southwestern part of the caldera merges with the southwest rift zone at an area of complex faulting and the east rift zone is linked to the caldera by a chain of pit craters. This chain of pit craters is crossed by the outer boundary faults of the caldera (Fig. 6) along which both horizontal and vertical movement has been relatively common.

Non-eruptive surface activity on the summit largely consists of areas emitting various quantities of steam and gas. The most active of these are within the caldera, notably Halemaumau, and fissures on the caldera floor, especially those formed in 1972 and 1974 on which solfataras have developed. Significant deposits of sulfur are also formed at the steaming area of Sulphur Bank to the northeast of the caldera. Steam also discharges from fissures on the upper southwest rift and from several of the pit craters on the upper east rift. Of note is the large area of steaming ground at Puhimau, which is the result of intrusive activity in about 1938.

Little is known about the hydrology of the summit area, largely due to the lack of drillholes. Geophysical logging of a hole drilled to the southwest of the caldera (Fig. 6) in 1973 located a brackish water table at approximately 488 m depth [Zablocki et al, 1974]. Considering the likelihood of complex hydrology this may, however, have indicated a perched body of water.

4. Waianae Area

The surveys in this area were centered on Lualualei Valley in western Oahu (Fig. 3) to the west of the north-northwest to south-southeast running Waianae Range. This leeward part of Oahu receives an annual rainfall of 50 to 75 cm and has a mean temperature of 24 to 25°C. The Waianae Range forms the western boundary of the central plateau of Oahu, and has

elevations to 850 m. Resistant ridges trending southwest from the range to the coast have formed four large amphitheatre-type valleys, the largest of which is Lualualei. Erosion of the surrounding ridges has formed talus slopes at their bases and significant alluvium and colluvium infilling of valleys, which is indicated to be over 300 m thick in parts of the larger valleys [Macdonald and Abbott, 1977]. The coastal zone is flat and is largely covered by calcareous sedimentary material of marine origin and mid-late Pleistocene age. Numerous seasonal streams drain westward from the Waianae Range.

Lualualei Valley roughly corresponds to the location of the caldera of the ancient Waianae volcano at its present erosion level. The western part of the ediface has been removed by erosion, and collapse subsidence of the central part of the caldera relative to the boundaries, may also have occurred. Evidence (geological and geophysical) suggests that the central vent area was located between Mauna Kuwale and Kolekole Pass (Fig. 7;8) where extensive dike formation, brecciation and highly resistant topography are evident, and flows generally dip away from this center [Stearns and Vaksvik, 1935; Cox et al, 1979].

The rocks of the area have been classified as the Waianae Volcanic Series and subdivided into lower, middle and upper members [Stearns 1967]. The lower member formed the mass of the volcano and is comprised of tholeiitic lavas, largely as thin-bedded pahoehoe and associated pyroclastics.

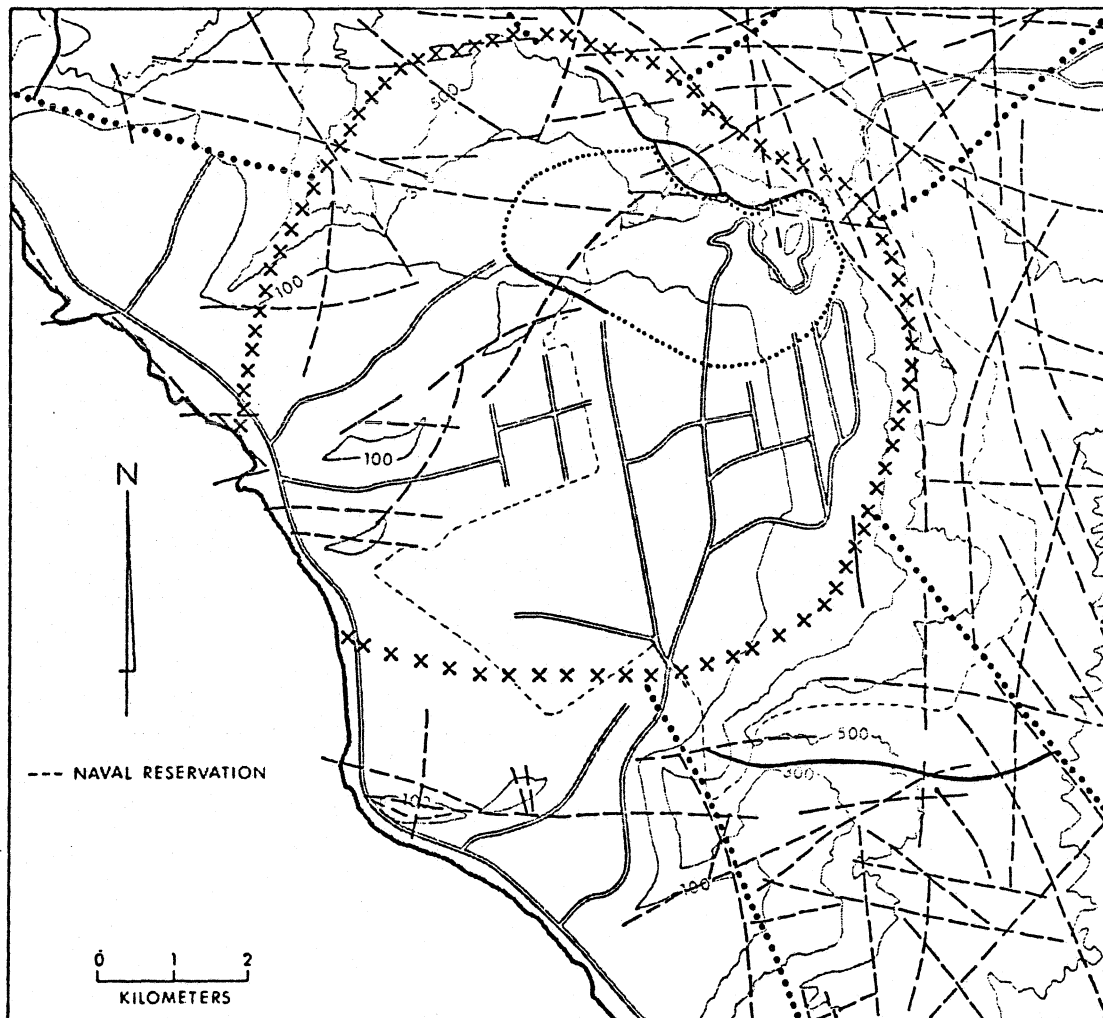


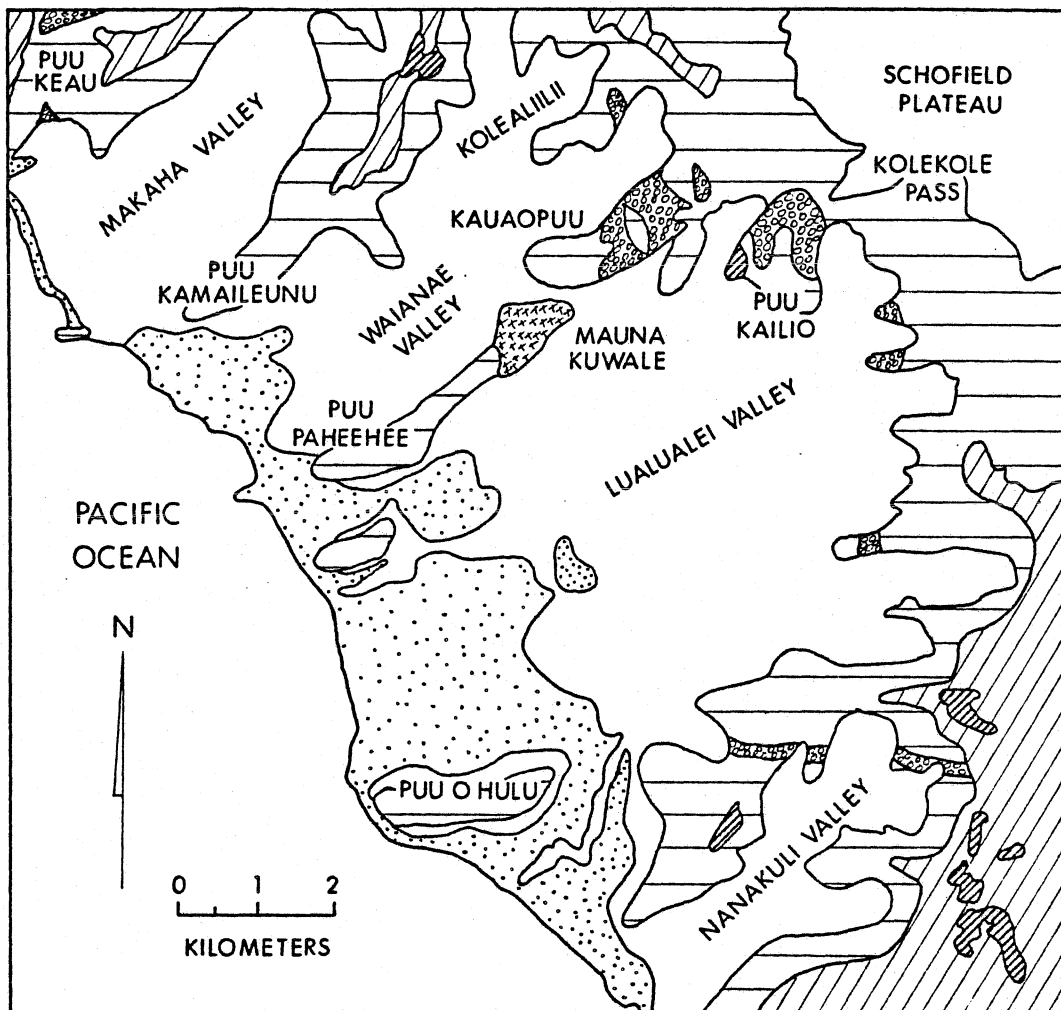
Figure 7. Waianae survey area. Air photo lineations are shown by broken lines; faults mapped by Stearns et al [1940] by unbroken lines. Crosses indicate the inferred outline of the boundary of the eroded caldera; dots outline the main eruptive zone; closed circles outline the indicated locations of the NW, SE and NE rift zones. The 100, 300 and 500 m contours are shown.

The massive sub-horizontal lavas of the middle member formed and filled the caldera, also building up the height of the volcano by flows down its slopes. This member is largely of tholeiitic composition, but becomes more alkalic towards the top. The largely alkalic flows of the upper member unconformably overlies the middle member and although absent in several parts are up to 700 m thick in the southeastern part of the area.

The bulk of the subaerial activity of the Waianae Volcano has been dated at 3.6 to 2.4 m.y. ago [Doell and Dalrymple, 1973]. The lower and middle members were formed penecontemporaneously and the change to alkalic composition of the upper member occurred over 0.2 m.y. [McDougall, 1964]. A post-erosional period of renewed activity may have taken place during the Pleistocene and formed lava and cinder near Kolekole Pass [Stearns, 1967].

Three rift zones are indicated to have developed (NW, SE and NE). The southeast is evidenced by a series of later stage cinder and spatter cones, probably representing the last vents of the upper member activity. Major lineations from air photographs (Fig. 7) suggest that the northwest-southeast systems were 4 to 5 km in width, and that of the northeast 2 to 3 km.

The lower and middle member rocks act as a reservoir for fresh basal groundwater; the numerous dikes also cause the gradient to be stepped from near sealevel at the coast to



GENERALIZED GEOLOGY MAP

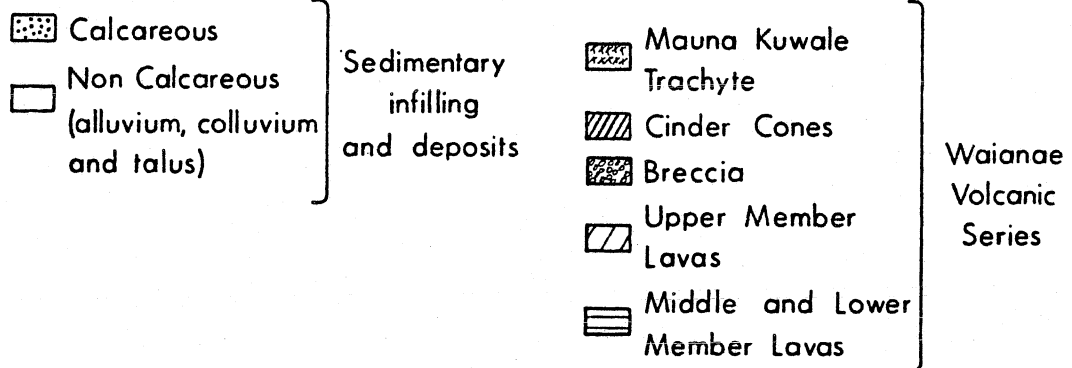


Figure 8. Generalized geology of the Waianae area [after Stearns et al, 1940].

more than 500 m A.S.L. on the Waianae Range. Limited fresh groundwater also occurs in the alluvium and sedimentary material, but seawater encroachment into this material is common due to the continuous pumping of many groundwater wells in this shallow aquifer [Takasaki, 1971].

III. DESCRIPTION OF RADON AND ITS SUBSURFACE OCCURRENCE

A. Chemistry of Radon

Radon, atomic number 86, occurs naturally as an almost inert, soluble gas. It is the heaviest noble gas, with a boiling point of -61.8°C . Twenty-seven unstable isotopes of radon (Rn-200 to Rn-226) have been determined to exist, most of which have extremely short half-lives ($t_{\frac{1}{2}}$). The most commonly occurring isotopes of radon in the natural environment are Rn-222 ("radon"; $t_{\frac{1}{2}} = 3.825$ day), Rn-220 ("thoron"; $t_{\frac{1}{2}} = 54.5$ sec) and Rn-219 ("actinon"; $t_{\frac{1}{2}} = 4.0$ sec). Of these, Rn-222 is the most abundant due to its longer half-life, and the abundance of its parent U-238. The activity due to Rn-220 in the natural environment is less than about 5% of that due to Rn-222. These three main isotopes are formed by the radioactive decay of the radium isotopes Ra-226 ($t_{\frac{1}{2}} = 1622$ yr), Ra-224 ($t_{\frac{1}{2}} = 3.64$ day) and Ra-223 ($t_{\frac{1}{2}} = 11.70$ day), respectively within the radioactive decay chains of U-238, Th-232 and U-235. Their disintegration energies are listed in Tables 1-3 and as shown the range of energies of the radon alpha particles within these three series is 5.49 to 6.82 MeV. The majority of alpha-active species, however, emit alpha particles with energies over the broader range of 4 to 8 MeV. Consequently, there is a wide range of half-lives, as overall, a greater disintegration energy produces a shorter half-life. The

TABLE 1
U-238 (Uranium) Decay Series

Element	Symbol	$t_{1/2}^1$	Energy (MeV)		
			α	β	γ
92 Uranium	U-238	4.51×10^9 y	4.19	..	0.048
90 Thorium	Th-234	24.1 d	..	0.19	0.09
91 Protactinium	Pa-234	1.17 m	..	2.29	1.0
92 Uranium	U-234	2.48×10^5 y	4.77	..	0.05
90 Thorium	Th-230	8.0×10^4 y	4.68	..	0.068
88 Radium	Ra-226	1622 y	4.78	..	0.186
<u>86 Radon</u>	<u>Rn-222</u>	<u>3.825 d</u>	<u>5.49</u>	..	<u>0.51</u>
84 Polonium	Po-218	3.05 m	6.00
		(β decay (0.02%) to At-218, $t_{1/2}^1 = 1.3$ s)			
		(α decay (99.98%) to Pb-214)			
82 Lead	Pb-214	26.8 m	..	0.65	0.24
83 Bismuth	Bi-214	19.7 m	5.50	1.5	0.61
		(α decay (0.04%) to Tl-210, $t_{1/2}^1 = 1.32$ m)			
		(β decay (99.96%) to Po-214)			

TABLE 1 - continued

Element	Symbol	$t_{1/2}$	Energy (MeV)		
			α	β	γ
84 Polonium	Po-214	1.6×10^{-4} s	7.80	..	0.8
82 Lead	Pb-210	22.0 y	..	0.016	0.046
83 Bismuth	Bi-210	5.0 d	..	1.16	..
84 Polonium	Po-210	138.4 d	5.30	..	0.80
82 Lead	Pb-206	stable isotope			

[after Friedlander et al, 1966 and Adams and Gasparini, 1970].

TABLE 2
Th-232 (Thorium) Decay Series

Element	Symbol	$t_{1/2}$	Energy (MeV)		
			α	β	γ
90 Thorium	Th-232	1.39×10^{10} y	4.01	..	0.06
88 Radium	Ra-228	6.7 y	..	0.054	..
89 Actinium	Ac-228	6.13 h	..	1.11	0.90
90 Thorium	Th-228	1.90 y	5.42	..	0.08
88 Radium	Ra-224	3.64 d	5.68	..	0.24
<u>86 Radon</u>	<u>Rn-220</u>	<u>54.5 s</u>	<u>6.28</u>
84 Polonium	Po-216	0.158 s	6.78
82 Lead	Pb-212	10.6 h	..	0.36	0.238
83 Bismuth	Bi-212	60.6 m	6.05	2.20	0.04
		(α decay (33.7%) to Tl-208)			
		(β decay (66.3%) to Po-212)			
84 Polonium	Po-212	3.0×10^{-7} s	8.87
81 Thallium	Tl-208	3.1 m	..	1.79	2.62
82 Lead	Pb-208	stable isotope			

[after Friedlander et al, 1966 and Adams and Gasparini, 1970].

TABLE 3
U-235 (Actinium) Decay Series

Element	Symbol	$t_{\frac{1}{2}}$	Energy (MeV)		
			α	β	γ
92 Uranium	U-235	7.13×10^8 y	4.39	..	0.185
90 Thorium	Th-231	25.5 h	..	0.30	0.25
91 Protactinium	Pa-231	3.48×10^4 y	5.00	..	range
89 Actinium	Ac-227	21.6 y	4.95	0.046	range
(α decay (1.2%) to Fr-223, $t_{\frac{1}{2}} = 22$ m) (β decay (98.8%) to Th-227)					
90 Thorium	Th-227	18.17 d	5.98	..	range
88 Radium	Ra-223	11.7 d	5.71	..	range
<u>86 Radon</u>	<u>Rn-219</u>	<u>3.92 s</u>	<u>6.82</u>	..	<u>0.27</u>
84 Polonium	Po-215	1.83×10^{-3} s	7.37
82 Lead	Pb-211	36.1 m	..	1.36	0.83
83 Bismuth	Bi-211	2.15 m	6.62	0.59	0.35
(β decay (0.32%) to Po-211, $t_{\frac{1}{2}} = 0.52$ s) (α decay (99.68%) to Tl-207)					
81 Thallium	Tl-207	4.79 m	..	1.44	0.90
82 Lead	Pb-207	stable isotope			

[after Friedlander et al, 1966 and Adams and Gasparini, 1970].

disintegration of a radium atom (atomic mass 226, 224 or 223) produces an alpha particle and a radon atom (atomic mass 222, 220 or 219); the atom is stripped of its outer electrons and expends its kinetic energy. The neutral atom may then diffuse until it disintegrates or escapes from the rock or mineral grain.

The radioactive decay of radon nuclides is by the spontaneous emission of alpha particles from their nuclei. Alpha particles are composed of two protons and two neutrons with a resultant charge of +2. The emission of an alpha particle subsequently reduces both the atomic number and the neutron number by two, and the mass number by four, forming a daughter isotope of a different element. Due to the appreciable mass of an alpha particle, its ejection will impart a certain amount of recoil energy to the nucleus. The total alpha decay energy (E_{α}) is the sum of the kinetic energy of the alpha particle, the recoil energy it imparts to the product nucleus, and the energies of any gamma rays that are emitted [Faure, 1977].

The ground state decay of Rn-222 [Faure, 1977] emits a single set of alpha particles with a kinetic energy of 5.4897 MeV and forms Po-218. The alpha decay energy of Rn-222 is:

$$E_{\alpha} = 5.4897 + \frac{5.4897 \times 4}{218} = 5.5904 \text{ MeV}$$

(α particle energy) (recoil energy)

In addition to the alpha particle, a gamma ray of 0.51 MeV is observed, however, this accounts for only 0.07% of the disintegration of the nuclide, and indicates that some of the alphas are emitted with lower kinetic energies and leave the product nucleus in an excited state. This emission of the energetic alpha particle and the amount of recoil energy (0.10 MeV) which is imparted to the product nucleus, cause damage to the surrounding area of the crystal.

B. Concentrations of Parent Radionuclides in Hawaii Lavas

Possibly one of the major controls on determining the amount of the different isotopes of radon available for release to rock voids is the concentration of the parent radionuclides within the rocks or minerals. Of major importance, then, are the concentrations of the primary radionuclides of the three main radioactive decay series: U-238, Th-230 and U-235. Of these, U-235 is the least abundant as U-238 constitutes 99.27% of the naturally occurring uranium on earth. In the Hawaiian islands, which are largely formed of basaltic lava flows ($\text{SiO}_2 \approx 50\%$) of low radioactivity, the concentrations of uranium and thorium are low relative to those rocks of higher silica content found in other localities, especially continental environments.

The basaltic lavas found in Hawaii can be classified into two broad petrochemical groups, tholeiitic basalts and alkalic basalts. The tholeiitic basalts are relatively rich

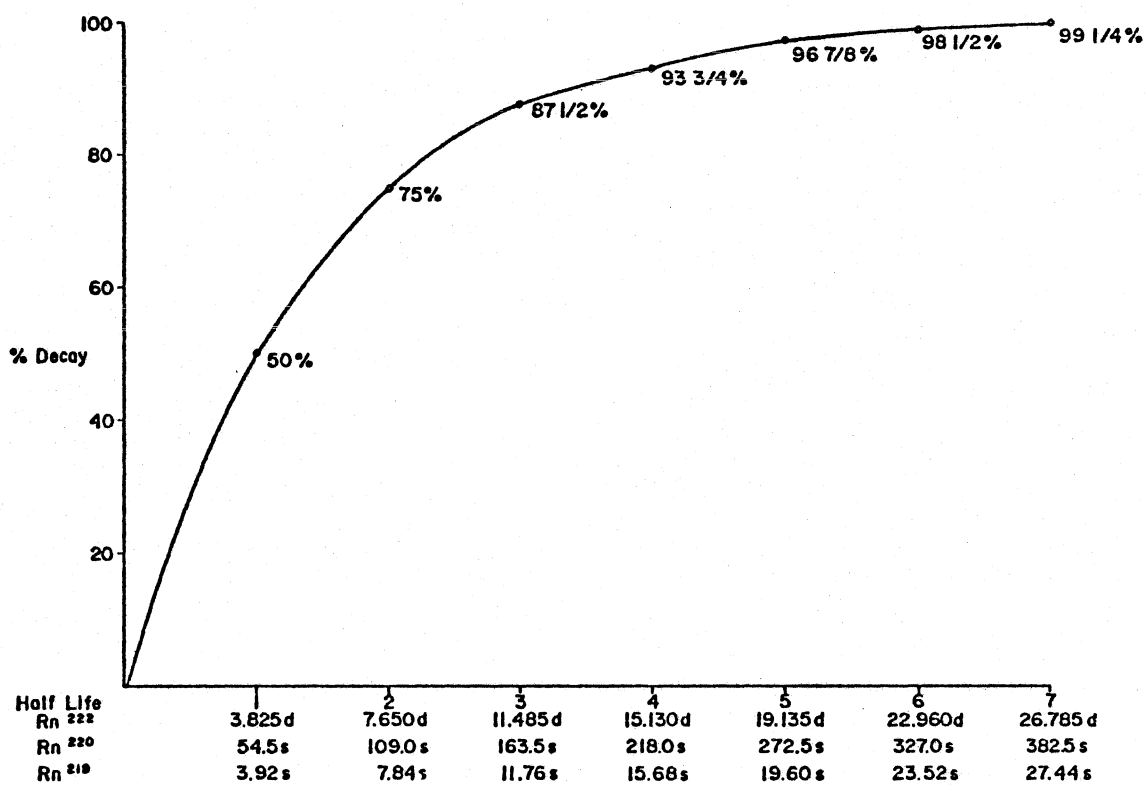


Figure 9. Percent decay as a function of time for Rn-222, Rn-220 and Rn-219. It can be seen from the graph that only 75% of Rn-222 has decayed over a period of 7.6 days. [from Nielson, 1978].

in silica and poor in the alkalies (sodium and potassium) and the contained pyroxene is largely augite (and minor pigeonite) containing little calcium or aluminum. The alkalic basalts are comparatively poor in silica and rich in alkalies, and pyroxene is generally augite containing calcium and aluminum [Macdonald and Abbott, 1977]. The tholeiitic basalts constitute the mass of the shield volcanoes which form the islands of Hawaii, and are produced throughout much of their life. During the final eruptive stages, however, the more viscous alkalic basalts are produced, and form flows which overlie the tholeiitic lavas. The still active volcanoes of Kilauea and Mauna Loa on the island of Hawaii have not yet reached this stage of activity and consequently have not erupted alkalic flows. In some of the older areas of extinct volcanism, however, post-eruptive erosion has removed a large part of the overlying alkalic basalts. In some cases renewed post-erosional volcanism has occurred and the lavas erupted are largely silica-deficient nephelinites and basanites. Nephelinites contain nepheline in place of feldspar and usually contain olivine; basanites are transitional between nephelinites and ordinary alkalic basalt, and contain both nepheline and plagioclase feldspar [Macdonald and Abbott, 1977].

Between these lava types distinguishable variations in the concentrations of uranium and thorium exist, although all concentrations are well below those of the more felsic

rocks (greater than $\approx 52\% \text{SiO}_2$). Table 4 lists reported uranium and thorium concentrations for Hawaii lavas. Using these figures a mean value for tholeiitic basalt is 0.24 ppm uranium and 0.74 ppm thorium ($\text{Th/U} = 3.09$); for alkalic basalts 0.95 ppm uranium and 3.06 ppm thorium ($\text{Th/U} = 3.22$) and for typical silica-deficient post-erosional lavas, 1.59 ppm uranium and 5.30 ppm thorium ($\text{Th/U} = 3.33$).

The formation of radon isotopes in nature is closely related to the concentrations of their immediate parent, radium. Consequently, as for uranium, the occurrence of radium in secondary minerals may also be important; this aspect is discussed below, and here concentrations of radium within lavas are presented. Few determinations of radium in Hawaii lavas have been reported; Poole and Joly [1914 in Piggot, 1931] list three basalt samples from the Hawaiian Islands as having an average content of 1.10 pg/g radium, one of which was a "ropy lava from Kilauea" (presumably tholeiitic) with a radium concentration of 0.96 pg/g ($\text{pg/g} = 10^{-12} \text{g/g} \approx 0.976 \text{pCi/g}$; Ci = Curie, the quantity of a radionuclide in which the number of disintegrations per second is 3.700×10^{10}). The data presented by Piggot [1931] show a mean radium concentration in tholeiitic basalts from Hawaii of 0.94 pg/g (range 0.79 to 1.09) and interestingly a similar, mean concentration in alkalic basalts of 0.85 pg/g (range 0.75 to 0.90).

TABLE 4
Mean Composition of Reported U and Th in Hawaii Lavas

Lava Type	U ppm		Th ppm		Th/U	N ^o of Samples	Reference
	Mean	Range	Mean	Range			
Tholeiitic basalt	0.23					1	A
" "	0.28	0.17-0.49	0.69	0.11-1.38	2.46	6	B
" "	0.18	0.03-0.33	0.62	0.45-0.86	3.46	8	C
" "	0.30	0.15-0.57	0.95	0.55-1.69	3.17	11	D
" "	0.21	0.08-0.49	0.71	0.25-1.32	3.38	17	E
Alkalic basalt	0.57	0.38-1.22	1.90	1.17-3.34	3.33	3	A
" "	0.99	0.97-1.00	4.23	4.01-4.44	4.27	2	B
" "	1.34	1.00-1.68	3.16	2.92-3.39	2.36	2	C
" "	0.56	0.29-0.87	1.96	1.50-2.60	3.51	3	D
" "	1.28	0.22-2.57	4.06	1.46-7.07	3.17	11	E
Melilite, nephelinite, basanite	2.16	2.11-2.26	7.18	6.74-7.48	3.32	2	A
" " "	1.02	0.50-2.24	3.42	1.62-6.74	3.35	10	C
Average of all Determinations:							
Tholeiitic basalt	0.24		0.74		3.09		
Alkalic basalt	0.95		3.06		3.22		
Melilite, nephelinite, basanite	1.59		5.30		3.33		

A: Tatsumoto, 1966; B: Berry, 1973; C: Fankhauser, 1976; D: Tatsumoto, 1978; E: May, 1979.

He notes however, that because these concentrations are similar to those in granites which are more silicic the concentrations may be too high. Evans and Goodman [1941] also consider that Piggot's values are too high and suggest a mean value for mafic igneous rocks of 0,31 pg/g.

Wilkening [1974] determined a radium content of Hawaii tholeiitic lavas of 0.24 pg/g, which is in general agreement with that of Evans and Goodman. Approximations of Ra-226 content calculated from uranium activities in the samples analysed are in agreement with Wilkening's data and gave mean values (in pCi/g) of 0.16 for lavas, 0,23 for soils and 0,14 for sublimate (Table 8; Appendix 5).

Radium concentrations of 0.09 to 0.18 pg/g have been for "volcanic rocks" (largely basaltic and alkalic lavas) in Japan have been measured [Kimura et al in Iwasaki et al, 1956]. Therefore, as the radium content of Hawaii lavas is low, up to an order of magnitude lower than that in rocks of intermediate and felsic composition, the potential of radon to emanate from these lavas is also low. Consequently, from the aspect of radon entering the ground gas, migrating, and being of sufficient concentration to be measureable near the ground surface, other conditions become important.

The potential availability of the different isotopes of radon can then be discussed in terms of the relative abundances of their parent radionuclides. Due to both the very low abundance of naturally-occurring U-235 and the

short half-life of Rn-219 (4.0 sec) in this decay series, this isotope makes little contribution to the total radon being released to rock voids. The contributions of the isotopes Rn-220 and Rn-222 are largely dependent on the relative amounts of thorium and uranium respectively, in the rocks. Fleischer and Mogro-Campero [1978] note that rocks with Th/U ratios of 3 or 4 will contribute a significant fraction of Rn-220, and that with higher ratios the Th-232 decay series will tend to be dominant. A common Th/U ratio for Hawaii lavas is in the order of 3.3, which suggests that a significant Rn-220 contribution can be released from the rocks. Using data from Evans and Goodman [1941], Fleischer and Mogro-Campero [1979, a] present a Th/U ratio for mafic igneous rocks of 4.0, a Rn-220/Rn-222 activity in the ground of 1.28, and the fraction of Rn-220 activity in the ground of 0.56. By using a Th/U ratio for Hawaii lavas of 3.3, a Rn-220/Rn-222 activity ratio of 1.05 and a fraction of Rn-220 activity of 0.51 can be determined. Any changes in the relative amounts of the isotopes that are subsequently measured near the surface are most likely to be due to the conditions controlling or existing during migration.

C. Radon Emanation From Rocks and Minerals

Positive alpha particles can only travel short distances in matter before they are reduced to thermal

energies. In passing through matter they mainly lose energy by interaction with electrons but lose only a very small fraction of their energy in a single collision. Thus they are not appreciably deflected in the collision and alpha particle paths are almost straight lines. The range of the alpha particle is dependent on its initial energy [Friedlander et al, 1966]. However, as alpha particles are of relatively large mass and charge they produce a large amount of ionization along their paths, therefore in practice, the most energetic alpha rays from radioactive sources will be stopped by 0.034 mm of lead or 0.13 mm of water [Moore, 1962]. Diffusion of radon isotopes from minerals is slight, and Tanner [1964,a] concluded that "... any appreciable emanations of Rn-222, Rn-220 or Rn-219 atoms come from radium isotopes distributed in secondary crusts or films, or in the shallow surface layers of intact crystals of the host minerals." He also noted that radon isotopes deeper in crystals will not be released without the development of large internal surfaces, such as from chemical corrosion, weathering, or intense microscopic fracturing. Enhanced emanation of radon has been noted from various secondary phases. Films of hydrous or gelatinous iron oxide deposited on grain surfaces may produce a large fraction of the Rn-222 emanating from Ra-226 trapped in them [Hahn, 1936] and Rn-222 should be readily released from Ra-226 adsorbed on (or cementing)

clay particles or on the exterior surfaces of clay lenses [Tanner, 1964,a]. The emanation of radon is also enhanced by hydrothermal alteration and temperature increases, which are further discussed below.

The principal mechanisms by which radon isotopes escape from mineral grains and enter rock pores, capillaries or microfractures are by radioactive recoil into the liquid-containing spaces, or by diffusion from solid material of appreciably smaller dimensions than the diffusion length of the most short-lived isotope [Tanner, 1964,a]. With the decay of radium, energy is released as an alpha particle and by the recoil of the radon nucleus. The amount of radon generated by recoil is variable and is dependent on the grain size of the rock (or other radium containing material). Andrews and Wood [1972] calculated for sediments of marine origin that for 1.0 μm diameter particles 4.9% of the radon generated can escape by the recoil mechanism, and for 100 μm diameter particles only 0.049% can escape. With a recoil energy of 85 KeV, the Rn-222 atom has an average recoil range of 0.1 μm in water and 64 μm in air [Andrews and Wood, 1972] and 0.03 to 0.04 μm in rocks [Stoker and Kruger, 1975]. Further, as radon is a noble gas it is not adsorbed on rocks [Tokarev and Shcherbakov, 1956 in Tanner, 1964,b], however, Andrews and Wood [1972] consider the readsorption of recoil radon atoms on adjacent surfaces can occur if these atoms have

not substantially lost their kinetic energy before collision. They believe that rock voids saturated with water (as opposed to air) can more effectively absorb the released kinetic energy and may be expected to enhance the release of radon from a solid into the adjacent voids.

The radon emanating power of a specific material describes the percentage of radon which is released from the material grains relative to that which is produced from the radioactive decay of radium. Both the emanating power and the concentrations of the parent radionuclides determine the amount of radon which enters the rock void or the fluids within those voids. The emanating power is partly dependent on the specific surface area ($1/\sqrt{d}$; d = diameter) of the mineral grains [Andrews and Wood, 1972]. In naturally occurring materials the emanating power may vary from nil to nearly 100%, although it normally ranges between the narrower limits of 1 to 30% [Tokarev and Shcherbakov, 1956 in Tanner, 1964, a], Barretto et al [1974] measured the emanating power of crushed basalts to be in the order of 9% of the Rn-222 being formed. Wilkening [1974] determined the emanating power of tholeiitic basalts from the island of Hawaii to be 2% and also refers to other work by Barretto [1973] in which a similar 2% value for Hawaii lavas was determined. As noted by Wilkening [1974] the low emanating power of the Hawaii lavas indicated that although radium is present radon isotopes are limited from

escaping due to being trapped within the crystalline and glassy grains and their short half-lives. The lavas measured, however, were surface collected historic flows and it is likely that under in situ conditions the older lavas would experience macro- and micro- fracturing, weathering (and hydrothermal alteration) that should increase the emanating power. It would also be noted that a determination of radon emanating power of a rock assumes that the rock particles have uniform radium distribution and crystal structure, as crystals or grains with structural imperfections are likely to enhance the diffusional movement of radon isotopes.

D. Mechanisms of Radon Migration in the Ground

There are two major mechanisms by which subsurface radon can migrate: diffusion and convection. The controls over these determine the distance radon atoms can move before undergoing radioactive decay.

1. Diffusion

Diffusional movement is one of the main mechanisms of radon release from mineral grains. Movement is due to a concentration gradient, the radon moving from an area of greater to lesser concentration. Diffusion essentially occurs where the radon moves relative to fluids within the rocks, and coefficients of diffusion (which are dependent on such parameters as porosity, permeability, fluid content

and temperature) have been determined experimentally for various media. Tanner [1964,a] notes that the diffusion coefficients for the different major isotopes of radon in identical media are almost equal, and that the diffusion distances vary. He calculated that in dry sand (using a diffusion coefficient of $5 \cdot 10^{-2} \text{ cm}^2/\text{sec}$) the diffusion lengths of Rn-222 and Rn-220 are 160 cm and 2 cm, respectively; Fleischer and Mogro-Campero [1978] determined diffusion lengths of 155 cm and 2 cm, respectively, in dry porous soil. Tanner [1964,a] further calculated that a decrease in concentration of 100-fold will take place over distances of 730 cm and 9 cm, respectively, in dry sand. Bogoslovskaya et al [1932] calculated a rate of movement of radon from an artificial source, in sand, of $7 \cdot 10^{-4} \text{ cm/sec}$ and concluded that under optimum conditions, the mean migration distances for Rn-220 and Rn-219 in dry soil are very small and in the order of 2 cm and 6 mm respectively. Schroeder et al [1965] considered that a value of $0.03 \text{ cm}^2/\text{sec}$ is a fair approximation of the diffusion coefficient for moderately dry, sandy soils, which agrees well with that used by Tanner [1964,a].

The diffusion coefficient of radon in the open air is reported by Budde [1958] to be approximately $10^{-1} \text{ cm}^2/\text{sec}$, and approximately $10^{-5} \text{ cm}^2/\text{sec}$ in water. He utilized various grain size materials (over 30 day periods) to experimentally show that in the dry state the diffusion

TABLE 5

Mean Diffusion Distances of Radon Isotopes in Different Media

Medium	Mean Distance		Diffusion Coefficient
	Rn-222 (cm)	Rn-220 (cm)	(cm ² /sec)
Air	220	2.85	10 ⁻¹
Porous soil	155	2.0	5 x 10 ⁻²
Water	2.2	0,0285	10 ⁻⁵
Saturated Porous Soil	1.55	0,020	5 x 10 ⁻⁶

[after Tanner, 1964, a and Fleischer and Mogro-Campero, 1978].

coefficients of all the materials used (sands and clays) was similar to that in the open air, but that in a state of natural moisture there were great differences directly related to grain size. The use of the diffusion coefficient for open air can, however, be used to obtain an indication of the maximum distance a radon atom will travel by diffusion after release from the mineral grains.

Several generalizations can be made about diffusion coefficients [Tanner, 1964,a]:

- (i) diffusion coefficients for the several radon isotopes in identical media are practically equal;
- (ii) the diffusion coefficient is not sensitive to the pore diameter in the usual range of geological materials;
- (iii) increasing moisture in a porous medium causes a reduction of coefficient greater than can be accounted for by the pore space occupied by water.

2. Convection

Convection is the situation where the subsurface fluids (liquid or gas) themselves move, and act as the transport mechanism by "carrying" the radon. In this case the radon may be in such a small quantity relative to the carrier that its diffusive properties are overwhelmed by the motion of the carrier, such as in the movement of groundwater or ground gas. The latter case is emphasised in these surveys, specifically upward movement of ground

gas induced by above ambient subsurface thermal conditions.

The basic condition that is being considered is that recoil and diffusional processes are important in releasing radon from the rock matrices and crystal structures to the surrounding fluids and that dynamic movement (largely convective) of these fluids will transport the contained radon over distances considerably greater than would be possible by diffusion alone. The amount of radon made available to the convective process is, as described above, related to the concentration of the parent nuclides and the emanating power of the rocks, and the subsequent convective mobility of radon is due to both the gaseous form of this element and its virtually inert chemical nature. These properties provide radon with the potential to be an indicator of permeable ground and above ambient temperatures.

E. Radon migration by Convection Processes

For radon to migrate significant distances a convective transport mechanism must predominate. In simple terms subsurface convective migration of radon can be considered as having lateral and vertical components which can be represented by lateral flow of groundwater and the vertical movement of ground gas.

In the case of lateral (and also downward) migration of groundwater the distance the radon will travel in solution is dependent on the velocity of the groundwater flow, which

is essentially gravitational. In the case of diffusional transport, the diffusion coefficients of radon isotopes in water-saturated porous media are extremely small. If the diffusion coefficient for water (10^{-5} cm²/sec) is used as an upper limit, Rn-222 is indicated to undergo 100-fold diminution in 10.7 cm [Tanner, 1964,a]. Groundwater flow within the thinly bedded (commonly 0.5 to 3 m) and shallow dipping (7 to 8°) lavas of Hawaii can be considered anisotropic with the greatest flow along the lateral component. The thinness of these lavas causes the formation of many flow boundaries and the common occurrence of "clinkery" a'a lava and subsequent fracturing make the lavas, overall, highly permeable. Assuming a typical natural lateral flow velocity of 0.2 to 2.0 m/day [F.L. Peterson, pers. comm., 1980] and the mean life of Rn-222 of 5.5 days, it can be seen that radon migration by this mechanism is limited to maximum distances in the order of 11 m.

The limited ability of radon to migrate in this fashion further suggests that under normal conditions in Hawaii the radon content in the groundwater at a location is controlled by the availability of radon within the immediate area. This general condition is borne out by other examples. Okabe [1956] concluded that radon in hot spring waters in Japan was derived from sources near the points of discharge rather than from the sources of the waters themselves. From studies in New Zealand geothermal

areas, Belin [1959] considered that both steam and water collected radon from near-surface rocks and so the radon measured in surface discharging fluids is derived largely from the last rocks through which the fluids passed, As a consequence of this he noted a higher radon concentration in discharging fluids within rocks of felsic composition than those of intermediate composition.

From groundwater studies in Utah, Tanner [1964,b] determined that in those waters less than 0.5% of the Rn-222 within the water could have been derived from the accompanying Ra-226, as the radioactivity due to Rn-222 was 4-times that due to Ra-226. He found that the concentrations of Rn-222 over the survey area (several km²) were relatively uniform, and concluded that the Ra-226 source of the Rn-222 is a dispersed one rather than a concentrated one. He further considered that Rn-222 in groundwater well samples probably came Ra-226 quite close to the well but that it was immobile enough not to be present in the water sample. From the many studies of radon in groundwaters reported in the literature the one aspect that does become evident is that the dissolved radon is rarely in equilibrium with its parent or daughter radionuclides in solution.

The vertical transport condition that is being considered in this study is that of radon moving within upward migrating ground gas. This movement of the ground gas being induced and enhanced by heating and the consequent decrease

in density. The velocity of this movement is dependent on both temperature and ground permeability, and will determine how far the radon can be transported. Mogro-Campero and Fleischer [1977] have considered cases of a "normal" thermal gradient of $30^{\circ}\text{C}/\text{km}$; they calculated that for air or water in sand (hydraulic conductivity, K , of $> 3 \cdot 10^{-7}/\text{cm}^2$) convection over a vertical distance of 100 m could occur, and for $K > 3 \cdot 10^{-8}/\text{cm}^2$ over a vertical distance of 300 m. The latter value of K applies to sand and soil and sandstone of high permeability, and is within the range for permeable basalts (10^{-5} to $10^{-9}/\text{cm}^2$) [Freeze and Cherry, 1979]. Mogro-Campero and Fleischer concluded that "... in regions of higher than usual permeability, thermally induced subterrestrial fluid convection is a potential transport mechanism". Gasparini and Mantovani [1978] concluded that radon transported with convective fluids can remain in detectable amounts over distances greater than 100 m, as long as the fluid velocity is higher than 100 m per day (0.12 cm/sec). The assumption then, is that if vertical convective transport is possible over distances of several 100's of meters under suitable conditions of permeability at a normal thermal gradient, greater transport distances should be possible in areas of steeper thermal gradients and increased permeability.

In the situation of Hawaii lavas, while the horizontal component of permeability probably exceeds the vertical

from a hydrological viewpoint, permeability in both directions is so great and so subject to local variations that any difference between them is difficult to assess [Peterson, 1972]. Of prime importance in vertical movement potential is the development of secondary fracturing and faulting, which could enable high velocity gas flow to occur. In these situations convective transport of radon would be vastly dominant. As an example, the concentration of radon in ground gas measured at 2 m depth in a drillhole (Geothermal Test Well 3, depth 210 m) in the Puna area of Hawaii (air temperature in well 25.5°C at sampling point) was 85% greater than that measured in the ground nearby, at 30 cm depth [Cox, 1980,a]. The radon concentration of 62.4 R.U. measured in the well is approximately equivalent to $6.43 \cdot 10^{-3}$ nCi/l (Section IV.F). In this particular well the velocity of the upward moving ground gas was estimated to be 5 cm/sec, and that in another hole in the same area, which was vigorously discharging steam (Geothermal Test Well 2) to be 50 cm/sec. These velocities, although higher than would be expected within the lava pile indicate that well-developed faults or fracture systems could give rise to significant flow velocities. As an approximation, if a vertical flow velocity of 0.1 cm/sec is used for fractured lavas, ground gas could migrate in the order of 86 m/day. Knowing that in 7.5 days only 75% of an initial quantity of radon will decay (Fig. 8) detectable amounts of radon could travel in the order of 650 m.

Of importance, however, is that available radon will be continually picked up along the path of the ascending ground gas, which will presumably produce an enhanced concentration near the surface. Another consequence of fractured ground is that emanation of radon will probably be higher than in similar, but unfractured ground, due to the increase in surface area. To this end, various workers have shown higher concentrations of radon in ground gas over faults or faulted areas [e.g. Nielson, 1978; Soonawala and Telford, 1980]. The surveys by Nielson at Roosevelt Hot Springs, Utah, successfully located mapped faults that communicate with the structurally controlled geothermal reservoir. This included faults buried by alluvium.

In Hawaii the radon being measured near the ground surface is indicated by this study to be derived largely from the thickness of lavas above the water table. The thickness of these unsaturated lavas is commonly 30 to 300 m. This is supported by the fact that in surveys in areas where alkalic lavas overlie tholeiitic lavas, the radon concentrations measured are significantly higher (see below) than those determined in areas of solely tholeiitic lavas. Other studies, in which the variations of concentration of radon in ground gas in association with eruptive or intrusive activity are measured [Gasparini and Mantovani, 1978; Cox et al, in press] suggest however, that during periods of magma migration and the accompanying increase in heat, the vapor given off at depth may pass through groundwater saturated lavas removing

much of the radon from the liquid phase in addition to transporting that radon formed above the water table.

F. Vertical Variations in Radon Concentration

As the technique used in this survey measures the concentration of radon in ground gas at shallow depth (≈ 30 cm) it is necessary to consider the vertical variations that may occur and their causes. Vertical gradients of radon concentrations are, however, dependent upon many factors including ground conditions as well as phenomena that can cause short-term fluctuations in radon concentration.

Kovach [1945] determined an increase of radon concentration to 200 cm depth that was linearly proportional to the depth. The total concentration of radon isotopes were also determined by Jaki and Hess [1958] to increase linearly with depth, with the Rn-220 isotopic content approaching a maximum at about 75 cm depth, whereas the Rn-222 continued to increase linearly to about 6 m, then increased asymptotically towards a maximum concentration. Depletion of radon occurs in the upper layers (< 50 cm) of the ground and is due to radon exhalation into the atmosphere, and consequently the concentration of radon decreases to nearly zero at the ground surface [Hatuda, 1953; Fleischer and Mogro-Campero, 1979,b]. Diffusion of Rn-222 in a uniform medium with a free surface produces a concentration profile in which the concentration at depths of 50 to 70 cm is approximately one-third the

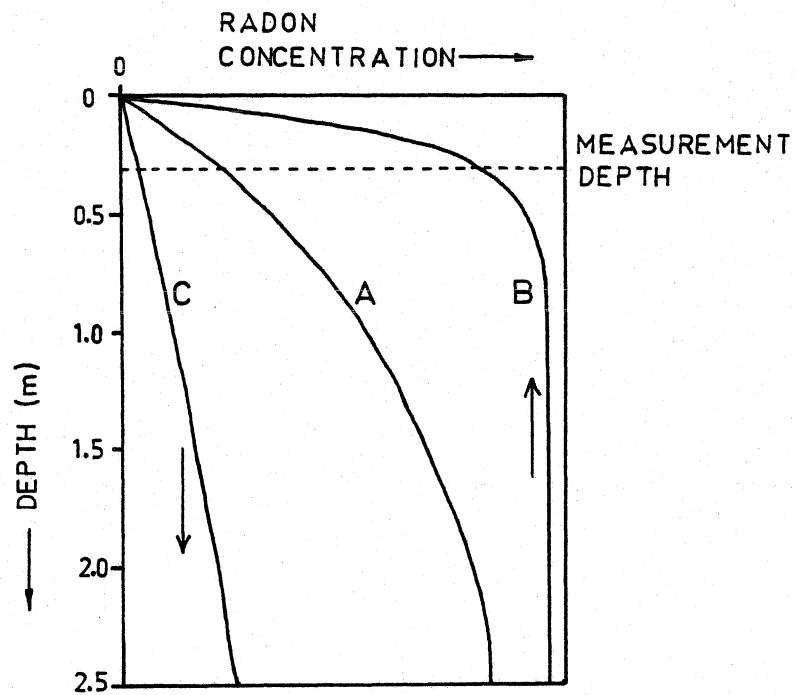


Figure 10. Radon concentration with depth: A, represents the Rn-222 concentration with depth for a dry homogeneous earth in the absence of vertical flow (i.e. movement by diffusion only); B, represents Rn-222 concentration in the presence of vertical upflow; and C, Rn-222 concentration in the presence of downflow [from Fleischer and Mogro-Campero, 1979, a].

concentration at greater depth [Fleischer and Likes, 1979]. A comparison between profiles in which no convective flow occurs (diffusion only) and profiles in which upflow and downflow are active is shown in Fig. 10. Convective transport can be seen to significantly alter the concentration profile at shallow depths. In (unsaturated) soil, it is generally assumed that the diffusion mechanism predominates over the transport mechanism (in the form of moving soil water) as the factors affecting transport are more transitory than those affecting diffusion [Tanner, 1964,a]. This however may not necessarily be the case for areas where transport by vapor occurs, and especially where elevated subsurface temperatures exist.

The type of overburden can also have an affect on the radon concentration in the near-surface ground gas. In Hawaii especially, soil conditions can vary considerably between, and within areas. This variation is dependent on both the climatic conditions and the age of the lavas, and soil thicknesses can range from 3 to 4 m in older eroded areas to non-existent in areas of historic lava flows. Variations in radon emanation from soil such as caused by differences in organic content and grain size are discussed in Section V.

G. Additional Influences on the Concentration of Radon at Shallow Depth

Numerous factors can effect both the concentration of radon in ground gas at shallow depth and its exhalation at

the ground surface. Short term variations are not uncommon and can be caused by such influences as changes of barometric pressure with rainfall, changes in atmospheric temperature and wind velocities. A consequence of these fluctuations is that radon measuring techniques that utilize non-integrating instantaneous or short-period measurements can provide inconsistent results.

Rainfall apparently has one of the greatest short-term effects, especially in areas of high rainfall [Rosen, 1957]. It is generally agreed [Kovach, 1945; Hatuda, 1953; Tanner, 1964,a] that heavy rainfall causes an increase in near-surface radon concentration, which is followed by a decrease in radon emanation from mineral grains as the near-surface ground becomes water saturated, reducing the diffusion coefficient and the permeability. Under saturated or near-saturated conditions, although there is an increase of radon concentration within the ground, a substantial part of the radon becomes dissolved in the water phase and as well as a reduction of diffusion there is an accompanying reduction of convection. Botset and Weaver [1932] measured higher Rn-222 concentrations in shallow ground gas in the same holes in wet weather than in dry weather, and concluded that during heavy rainfall the Rn-222 is sealed in the ground and tends to approach equilibrium with the contained radium. It has

also been observed that during periods of heavy rainfall the radon concentration in closely spaced holes tended towards the same value [Kovach, 1945]. A series of radon detectors were exposed at Kilauea Volcano during a period in which rainfall was exceptionally heavy and continuous, causing the ground to become saturated and surface sheet flooding to occur. The measured radon concentrations during that time were 25 to 35% lower than immediately subsequent measurements made in that area during a period of more characteristic intermittent heavy rainfall. Under those conditions ground gas transport appears to have been reduced due to both the dissolution of radon in the water and complete saturation of the ground.

Changes in atmospheric pressure can increase or decrease the volume of gas in the ground. A minor increase in radon exhalation at the surface was observed with a decrease in barometric pressure [Rosen, 1957] and Kovach [1944], and Hatuda [1953] observed that a decrease in barometric pressure produced an increase of the radon content of ground gas. The latter is due to a greater amount of radon given off to the atmosphere, and the subsequent upward movement of ground gas with a higher radon concentration from below, to replace this. The effects due to barometric pressure changes are greatest at shallow depth, being most pronounced at depths of less than 75 cm; the time-lag for this effect increases with depth [Hatuda, 1953]. It has also been postulated that subsurface

radon can be moved primarily by the pumping action of the atmosphere through barometric air pressure changes and the diurnal changes in temperature [Gingrich, 1975]. Diurnal variations in radon exhalation have been observed and were considered to be most likely related to diurnal temperature changes [Rosen, 1957]; maximum exhalation of radon isotopes has been determined to occur at night and the minimum in the evening [Malakhov et al, 1966].

High velocity wind for several hours can reduce the radon content in the top layers of ground and by Bernoulli's Principle (a wind at the ground surface causes a lowering of pressure and consequent emission of soil gas into the atmosphere) can cause a mixing of near-surface ground gas with free atmosphere [Kovach, 1945]. This condition, however mainly applies to a case in which subsurface movement of radon is by diffusion and not by convection.

Reported changes in radon concentration due to the above phenomena are variable: Tanner [1964,a] considers that such short-term factors contribute to less than 10% of the changes in soil aeration; Gingrich [1975], however, notes changes in ground radon concentrations by factors of 3 to 10 over a 24 hour period. Consequently, the use of a 4 to 5 week measurement period can average out short-term fluctuations of this type so that they have negligible effect on the concentration measured.

H. Thermal Effects on Radon Diffusion and Convection

Under conditions of increased temperature within the ground two parameters of radon are affected, release from mineral grains and conditions of transport.

Changes of diffusion coefficient with increased temperature have been reported for different media. Starik and Melikova [1961] noted that the diffusion coefficients in nearly saturated muds are about an order of magnitude smaller than in water and that the effect of temperature is to double the diffusion coefficient with each rise of 10° to 20° in the ground. Gasparini and Mantovani [1978] consider that reported empirical observations indicate that no more than 10 to 12% of the produced Rn-222 can be removed from a rock by heating to temperatures as high as 260°C. They also consider that for significantly increased diffusion to occur much higher temperatures are required, and that because of the very low thermal conductivity (in carbonate and silicate rocks) diffusion is unlikely to be significantly increased by direct heating by a magma or by hot gases. It is, however, possible that in the case of rocks of basaltic composition, and with the high permeability of most Hawaii lavas, more enhanced diffusion will result from the higher temperatures associated with volcanically active, or recently active areas.

Convective movement of radon within ground gas can be expected with increasing temperature of the gas. By the Boyle-Charles Law ground gas can expand approximately 1/273 of

its volume for each degree ($^{\circ}\text{C}$) increase in temperature (under constant pressure) [Kovach, 1945] resulting in a decrease in density and a tendency to rise. Radon migration is, however, also influenced by the phase of the pore fluids as there is a temperature-dependent distribution of radon among the phases [Tanner, 1964,a]. The solubility coefficient of radon in water (the ratio of radon concentration in water to that in a vapor phase, at equilibrium) decreases with increasing temperature, i.e. as the temperature increases more radon is dissolved in the vapor phase. Solubility coefficients in water at various temperatures have been determined:

$0^{\circ}\text{C} = 0.51$	$60^{\circ}\text{C} = 0.13$
$20^{\circ} = 0.26$	$80^{\circ} = 0.112$
$40^{\circ} = 0.16$	$100^{\circ} = 0.107$

[Rogers, 1958; Sedlet, 1966; D'Amore, 1975]. Consequently, radon tends to remain in solution in water, unless it is "flushed out" by a gas. This preferential extraction of radon from groundwater by gas migrating through it has been postulated to occur during eruptive events on Kilauea [Cox et al, in press] and is greatly enhanced with increasing temperature in the 0° to 60° range [Gasparini and Mantovani, 1978]. The same occurrence has been observed in discharges from producing geothermal wells in New Zealand where the radon content of the steam is appreciably higher than that in the

wellhead water [Whitehead, 1978]. The decreasing solubility of radon in water with increasing temperature has also been noted [Stoker and Kruger, 1975] but it continues to remain high (10's cm^3/ℓ) at temperatures of several 100°C [Gasparini and Mantovani, 1978]. Radon solubility in water does, however, decrease with an increasing amount of dissolved material (electrolytes), e.g. for seawater ($\text{Cl} \approx 19500$ ppm) at 18°C the solubility coefficient is 0.165 [D'Amore, 1975]. This suggests that in the saline geothermal waters in Hawaii ($\text{Cl} \approx 2000$ to 3000 ppm) radon would have a very low solubility and a consequent greater tendency to enter the vapor phase.

Increased subsurface temperatures, then, can be considered to increase radon release from mineral grains, increase radon migration within convecting ground gas and in some cases increase the amount of radon within the gas by flushing the radon from groundwaters. These factors are obviously important in considerations of thermally-induced migration of radon.

I. Radon in Geothermal Systems

1. General

Elevated levels of radioactivity have been commonly reported to be associated with geothermal (and hydrothermal) systems. Previous studies have considered not only radon but also the other radionuclides present, their concentration and their source [e.g. Belin, 1959; Mazor, 1962; Whitehead, 1976; Leonard and Janzer, 1978]. Recently, increasing attention

has been given to the environmental-health aspect of this association, and especially to radon in cases where the geothermal fluids are utilized [e.g. O'Connell and Kaufman, 1976; O'Connell and Gilgan, 1978; Whitehead, 1978].

A common conclusion reached from these studies is that thermal groundwaters in an area have lower radioactivity than the cool groundwaters which recharge the geothermal system, and also that the vapor discharged often has a higher radon content than the thermal water. This suggests that in some cases radioactive elements may be deposited within geothermal systems, in addition to radon being preferentially absorbed in the vapor phase. Of consideration to this study is whether the radon being detected at shallow depth in the ground is from deep within the systems and partly derived from secondary deposition, or is collected from closer to the surface by the rising ground gas during migration, or whether some contribution of both occurs.

The amount of radon potentially available to geothermal fluids is dependent on the concentration and distribution of the immediate parent, radium, within the rocks through which these fluids pass. Within the basaltic lavas of Hawaii radium is relatively uniformly distributed (Section III.B) but a concentrated source in secondary mineralization is possible. Radium has similar chemistry to the other alkaline earth elements and so is subject to the same geochemical processes as calcium, barium and strontium, and is known to undergo

adsorption and co-precipitation in hydrological systems [Stoker and Kruger, 1975; Kruger et al, 1977]. Geothermal investigations in Nevada, [Wollenberg et al, 1974; 1975] showed varying abundances of radioelements within hot spring systems in which CaCO_3 is the predominant deposit. Radioactive anomalies from both field gamma radiometry and laboratory gamma spectrometry, in those hot spring areas appeared to be associated with low-flowing CaCO_3 - rich systems. For thermal spring systems in Montana, Leonard and Janzer [1978] considered that the high radon content of the thermal water and spring gases was derived from Ra-226 deposited on rock surfaces or co-precipitated with CaCO_3 . They concluded that in the systems studied dissolution of radioactive chalcedony at depth probably accounted for most of the radioactivity.

Uranium is commonly known to be enriched in material deposited in reducing environments and conversely, depleted in materials deposited in oxidizing environments. It is highly soluble as the species UO_2^{2+} , and as the uranyl ion (U^{6+}) it is relatively mobile under oxidizing conditions and complexes readily with OH^- , CO_3^{2-} and SO_4^{2-} . Uranium can be introduced into a geothermal system as uranyl carbonate and sulfate in groundwater. Due to their retrograde solubilities these minerals may then be deposited at high temperatures [Blake et al, 1956; Wollenberg, 1975], or if reducing conditions exist within the system due to the presence of H_2S , the uranyl ion may be reduced to the relatively insoluble

uranous ion (U^{4+}).

The concentration of uranium by precipitation under reducing conditions associated with hydrothermal activity has also been established in areas of submarine volcanism [e.g. Lalou and Brichet, 1980]. In the case of the Puna geothermal system, $CaSO_4$ as anhydrite and gypsum (also retrograde solubilities) is depositing within the casing of the geothermal well HGP-A (determined by XRD on down-hole scrappings); abundant H_2S occurs in well head discharges (≈ 750 mg/kg of steam [Thomas, 1980]) and the pH of the geothermal water is around 3.5. These conditions theoretically could enable the co-precipitation of uranium and radium salts with calcium salts within the geothermal reservoir. Whitehead [1978], however, concluded that in New Zealand geothermal systems the converse is occurring and that those systems are depleted in uranium and radium. He suggested that uranium may be deposited in a halo around a geothermal system, causing ground radon concentrations to rise peripheral to the system and to fall directly over it.

In contrast to uranium, Th^{4+} remains relatively insoluble as it does not change its oxidation state during rock weathering. This potentially greater loss of uranium relative to thorium can produce higher Th/U ratios within strongly weathered basalts and possibly also within those that have experienced hydrothermal alteration. The latter, however, may not be evident if uranium enrichment in secondary minerals

within the rocks has occurred. The above could influence the relative abundances of uranium derived Rn-222 and thorium derived Rn-220 available within geothermal reservoir rocks.

2. Geothermal Reservoir Engineering

Measurements of the concentration of radon in fluids discharged from wells in geothermal fields have also been successfully applied to hydrological engineering. The specific application has been in assessing the flow characteristics within the reservoir, the recharge to the reservoir and the source of the recharge. This use of radon as an engineering tracer has also been applied to evaluating reservoir stimulation techniques [Stoker and Kruger, 1975; Kruger et al, 1977]. Apart from the characteristics of the reservoir rocks, the chemical and physical characteristics of the geothermal fluids are important controls on radon release, as is the concentration and distribution of radium. Monitoring of geothermal wells has shown marked variations in radon concentration with time, both in individual wells and among wells. Measurements of radon concentration at several flow rates indicate a possible dependence of the effective emanating power of reservoir rocks on pressure [Stoker and Kruger, 1975]. Measurements on well discharges in the vapor-dominated Larderello geothermal field in Italy show that almost all the radon is contained within the gas phase (compared to the condensate phase) and that all the measured radioactivity is due to

TABLE 6
Radon Concentration In Various Geothermal Fluids

Location	Fluid Type	Range Radon (nCi/l) *	Reference
<u>Water</u>			
Badgastein, Austria	Spring water	0.5-120.0	A
Bristol, England	" "	1.38-2.39	B
Tiberias, Israel	" "	1.4-7.2	C
Japan	" "	1.8-49.0	D
Broadlands, New Zealand	Well water	<0.006-0.012	E
North Island, New Zealand	Pool/spring water	0.8-320.0	F
South Island, New Zealand	Spring water	0.14-1.83	G
East Mesa, USA	Well water	0.034-0.069	H
Nevada, USA	Spring water	0.073-1.29	I
Salton Sea, USA	Well water	1.4	H
Utah, USA	Spring water	1.35	J
Western USA	Spring water	0.003-14.0	K
<u>Steam - Gas</u>			
Surtsey volcano, Iceland	Fumarole gas	0.52.0,171	L
Larderello, Italy	Well gas	0.20-68.46	M
" "	Well condensate	0,23-12.7	M
Tuscany, Italy	" "	18.9-63.0	H

TABLE 6 - continued

Location	Fluid Type	Range Radon	References
Tuila & Kirchurich volcanoes, Kamchatka	Fumarole gas	0.005-0.011	N
Broadlands, New Zealand	Well condensate	0.77-4.62	E
North Island, New Zealand	Fumarole gas	0.32-340.0	F
" " " "	Fumarole condensate	7.0-340.0	F
The Geysers, USA	Well gas	8.77-31.40	O
" " "	Well condensate	16.3-31.40	H
Salton Sea, USA	Well steam	3.8	H
HGP-A well, Puna Hawaii	Well steam	0.76-2.40	P

* nCi = 10^{-9} curies

A: Pohl-Rüling & Scheminzky, 1972

B: Andrew & Wood, 1972

C: Mazor, 1962

D: Iwasaki et al, 1956

E: Whitehead, 1978

F: Belin, 1959

G: Whitehead, 1976

H: Kruger et al, 1977

I: Wollenberg, 1975

J: Rogers, 1958

K: O'Connell & Kaufman, 1976

L: Björnsson, 1968

M: D'Amore, 1975

N: Shavrov in Iwasaki et al, 1956

O: Stoker & Kruger, 1975

P: Kruger, 1977

Rn-222 [D'Amore, 1975]. In water-dominated areas within that field (or at least near secondary liquid bodies) low gas/steam ratios and high radon content in the gas are observed. In the vapor-dominated areas high gas/steam ratios and low radon contents are observed.

Two short flow tests measuring radon concentration in discharged fluids have been conducted on geothermal well HGP-A [Kruger, 1977; 1978]. In those tests the concentration of radon in the steam did not change significantly with changes in flow rate. This agrees with the model of a liquid reservoir [Stoker and Kruger, 1975] and suggests a radial flow pattern over a large reservoir in which the concentration of radon is somewhat independent of flow rate [Kruger, 1977]. These flow tests indicate the existence of a large uniform deposit of radium in the reservoir which Kruger believes suggests that the radium is essentially that contained within the lavas. The range of radon concentrations measured was 0.76 to 2.4 nCi/l (mean of 1.03) which is relatively low in comparison to many other geothermal fields (Table 6). This and the similarity of the HGP-A radon concentrations to those measured in Iceland and Kamchatka demonstrate the characteristically low values found associated with hydrothermal systems in basaltic terrains.

IV. FIELD AND LABORATORY METHODS

A. Common Types of Radon Surveys and Measurement Techniques

From the 1920's to the present numerous studies which measured a variety of forms of and controls over radon within the ground and at the ground-air interface have been reported. Studies of radon concentrations in ground gas were common. Most of which used some adaptation of driving metal pipes (some with internal sampling tubes) different depths into the ground. The ground gas in these pipes was subsequently drawn off into collection flasks and the radon content determined by ionization chamber [e.g. Botset and Weaver, 1932; Kovach 1944, 1945; Hatuda, 1953; Budde, 1958; Morse, 1976]. Another method used was the collection of radon decay products on a buried aluminum cylinder, charged to -600 V, which attracted the positively charged alpha-particles [Jaki and Hess, 1958]. The radon content of groundwaters has also been studied using water samples collected from springs and drill-holes [e.g. Tanner, 1964,b; Andrews and Wood, 1972; Morse, 1976].

Most of the analytical techniques for measuring radon have used some form of ionization chamber and electrometer. Tanner [1964,b] used a method for measuring the radon concentration in water by boiling the dissolved gas from the water sample into an evacuated ionization chamber. The ionization current was measured with a vibrating-reed electrometer and recording potentiometer during radioactive equilibrium between

Rn-222 and its short-lived decay products. After storage of the sample for a week or more the Rn-222 produced by the Ra-226 in the sample was measured by the same method. A method that is now more commonly used for water samples is to pass a stream of nitrogen (or helium) through the sample and over an activated charcoal trap (-80°C); the trap is then evacuated and heated to 200°C , after which the desorbed radon is transmitted to a ZnS coated scintillator flask by a nitrogen stream. The alpha particle emissions from Rn-222 and its daughters are determined by placing the scintillator flask on the photocathode window of a photomultiplier tube, operated with a scintillation spectrometer [Andrews and Wood, 1972; Kruger et al, 1977].

Studies which involved radon exhaled at the ground-air interface usually consisted of collecting the gas within some form of tank on the ground surface. The collected ground gas could then be drawn directly into an ionization chamber [Wilkening and Hand, 1960; Kraner et al, 1964]. Rosen [1957] measured radon activity within the tank by induction, applying a negative potential to a wire within the tank, which caused the positively charged decay products to be deposited on the wire. The set-up to the wire was then connected to an electrometer so that it acted as an integrating ionization chamber. Simply, the ionization chamber contains an insulated metal rod, to which is connected a two-wire electrometer. The wires are charged, producing a potential, and the introduced air is

ionized by alpha radiation in proportion to the radon concentration. The velocity of discharge of the electrometer is in proportion to the radon concentration, and measures the amount of ionization produced. As ionization chambers record as ratemeters, they are insensitive to low count rates.

More recent advances have produced several types of solid state detectors and portable alpha scintillometers. There are several types of digital alpha scintillometers, which record each pulse as single events (nonintegrating type detectors). In the type commonly used the gas sample is drawn into a chamber whose inside walls are coated with ZnS (Ag) and viewed with a photomultiplier tube [Morse, 1976]. Another unit with similar components is in the form of a probe inserted into the ground, removing the need for an air pump to withdraw ground gas. A recent solid state detector, combines an electronic counter that records all pulses above a specified threshold energy (typically 1 or 2 MeV) with a silicon detector within a housing cup [Warren, 1977]. This type of detector gives an instantaneous or short period (\approx 3 days) reading but may accumulate an instrumental background count due to deposition of radon daughter radionuclides (mainly Po-210) after repeated use. The use of films coated with an alpha-particle sensitive emulsion and alpha-sensitive plastics is now becoming more common for a wide variety of purposes, including field surveys. One commercially available system [Fleischer et al, 1972;

Gingrich, 1975] consists of a similar technique to that used in the surveys described here, but utilizes a different detector material and configuration within the housing cup. Both methods utilize relatively long field exposures (2 to 5 weeks) and are considered to be integrated measuring techniques as they average background "noise" and allow for short-term positive and negative fluctuations.

B. Radon Surveys With Specific Applications

Measurements of radon are being used increasingly in the field of applied research, notably in volcanic-eruption and earthquake "prediction" and in the exploration for uranium mineralization. Other studies have been concerned with environmental aspects such as radiation from nuclear power plants, the effects of nuclear explosions [Buddemeir et al, 1978; Zielinski and Rosholt, 1978] and the health physics aspect of the presence of radon in air in underground mine workings [Duggan et al, 1968].

Measurements of radon concentrations have been made in a variety of media on, and peripheral to volcanoes and volcanic rift zones. These media include groundwater, thermal spring water and gas, fumarole gas, eruption clouds and volcanic ejecta, as well as ground gas [e.g. Iwasaki et al, 1956; Iwasaki, 1976; Lambert et al, 1976; Chirkov, 1976; Sato and Sato, 1977; Gasparini and Mantovani, 1978; Polian and Lambert, 1979; Cox et al, 1980]. Similarly, a variety of techniques are utilized in the application of radon monitoring

for prediction of earthquakes of non-volcanic origin. The most common of these have been measurements of radon concentrations in groundwaters [e.g. Hatuda, 1953; Wakita et al, 1980] and over the last several years solid detectors have become increasingly used in the monitoring of changes in the concentration of radon in ground gas [e.g. King, 1978,a; Shapiro et al, 1980; Birchard and Libby, 1980]. Variations in local stress fields produced by earthquakes derived from ground movements can produce measurable changes in subsurface concentrations of radon. This is believed to be largely due to enhancing radon emanation from mineral grains by thermal effects, increasing surface areas by fracturing and deformation of crystals. These forms of stresses also act during volcanically-derived earthquakes, but in such cases an appreciable effect is also associated with increased upflow of gases due to increased temperatures and the increased release of radon associated with the movement of magma. Migration of magma, as occurs on such volcanoes as Karymsky and Kilauea produces increases in stress and temperature and a greater upflow of ground gas. These changes produce measurable increases in radon concentration before and during an eruptive (or intrusive) event [Chirkov, 1976; Gasparini and Mantovani, 1978; Cox et al, 1980]. Because of the volcanically and seismically active environment of Hawaii and the potential importance of temporal variations in radon concentration to radon mapping surveys as an exploration technique, several

stations were monitored for longer periods of time (Section IV.A).

A similar technique of measurements as used here is also applied to uranium exploration. In those surveys Rn-222 (a decay member of the U-238 series) is the radon isotope being sought and several techniques have been developed for distinguishing that isotope from the other isotopes of radon [Fleischer and Mogro-Campero, 1978 ; Fleischer and Mogro-Campero, 1979,a; Fleischer and Likes, 1979]. Basically, the distinction is possible by the use of membranes over the detector air-space entrance, which retard the passage of the shorter-lived isotopes long enough for them to decay. Such membranes were not required in this study but a discussion of the induced lag of the passage of Rn-222 both into and out of the cups through these membranes can be found in Fleischer et al [1979]. Other applications of radon measurements to uranium exploration, and case histories are described Budde [1958], Gingrich [1975] and Morse [1976]. However, in the case of uranium exploration, where the primary source of radon is uranium mineralization, it is not always clear what mechanisms control or cause the migration of radon in the ground from the source mineralization to the ground surface. An attempt has been made to quantitatively analyze conditions of radon movement in overburden by using various mathematical models [Soonawala and Telford, 1980]. That study and others, which show the limited range of radon within the ground, suggest

that in most cases where radon is used as a "pathfinder" for uranium mineralization some form of dispersion "halo" of secondary mineralization is required to have developed around the parent deposit. This halo acting as a source closer to the surface than the uranium deposit itself. "False" radon anomalies in ground gas have been determined in the pursuit of uranium deposits [D.L. Nielson, pers. comm. 1980] but such occurrences are rarely reported in the literature. It is unlikely that diffusional transport of radon alone would enable the detection of uranium mineralization at depths much greater than 10 to 20 m and this suggests that other mechanisms must be involved in the migration of radon in association with uranium ore bodies. That topic is discussed by Mogro-Campero and Fleischer [1977].

C. The Radon Measuring Device Used in this Study

The alpha-particles emitted by the decay of radon can be detected by alpha-particle sensitive film. The films used in this study are the commercially produced Kodak LR 115 Type II which consist of a 100 μm thick inert polyester base coated on one side with a 12 μm thick dark red emulsion of cellulose nitrate. The cellulose nitrate emulsion has been treated to increase its response to ionizing particles. The films were primarily designed for dosimetry of weak concentrations of ionizing particles or neutrons and can record protons with energies lower than 100 KeV and alpha-particles with energy

between several hundred KeV and 4 MeV. Costa-Ribeiro and Lobaõ [1975] however, consider that these films are mainly suitable for alpha-particles with energy ranges of from ≈ 0.9 to 1.5 MeV. The films can be used to record alpha-particles with energies above 4 MeV but those particles need to be decelerated by a screen of inert material of several μm thickness or (as in this study) an air space. These films convert each alpha-particle impact to a separate physical image which can be observed as a perforation of the colored emulsion. The films are highly insensitive to photons (e.g. light or χ or γ radiation) but are sensitive to cosmic rays.

Small strips of the film (2.0 x 2.5 cm) are attached with adhesive tape to strips of thin cardboard (which removes the necessity to handle the actual film) and are then suspended vertically from a plastic clip glued inside the bottom of a 250 ml polypropylene cup. The cup is buried inverted in the field as depicted in Fig. 11 a. After the cup is in place the hole is refilled with soil to further reduce surface influences. Refilling also had the added effect of discouraging interference, both human and animal, but inevitably approximately 5% of cups are lost or disturbed in the field. The 30 cm depth was maintained throughout the study and a constant approach was also used for all aspects of exposing, developing and counting the films, to aid in reducing the number of variables as the aim of these surveys is to seek

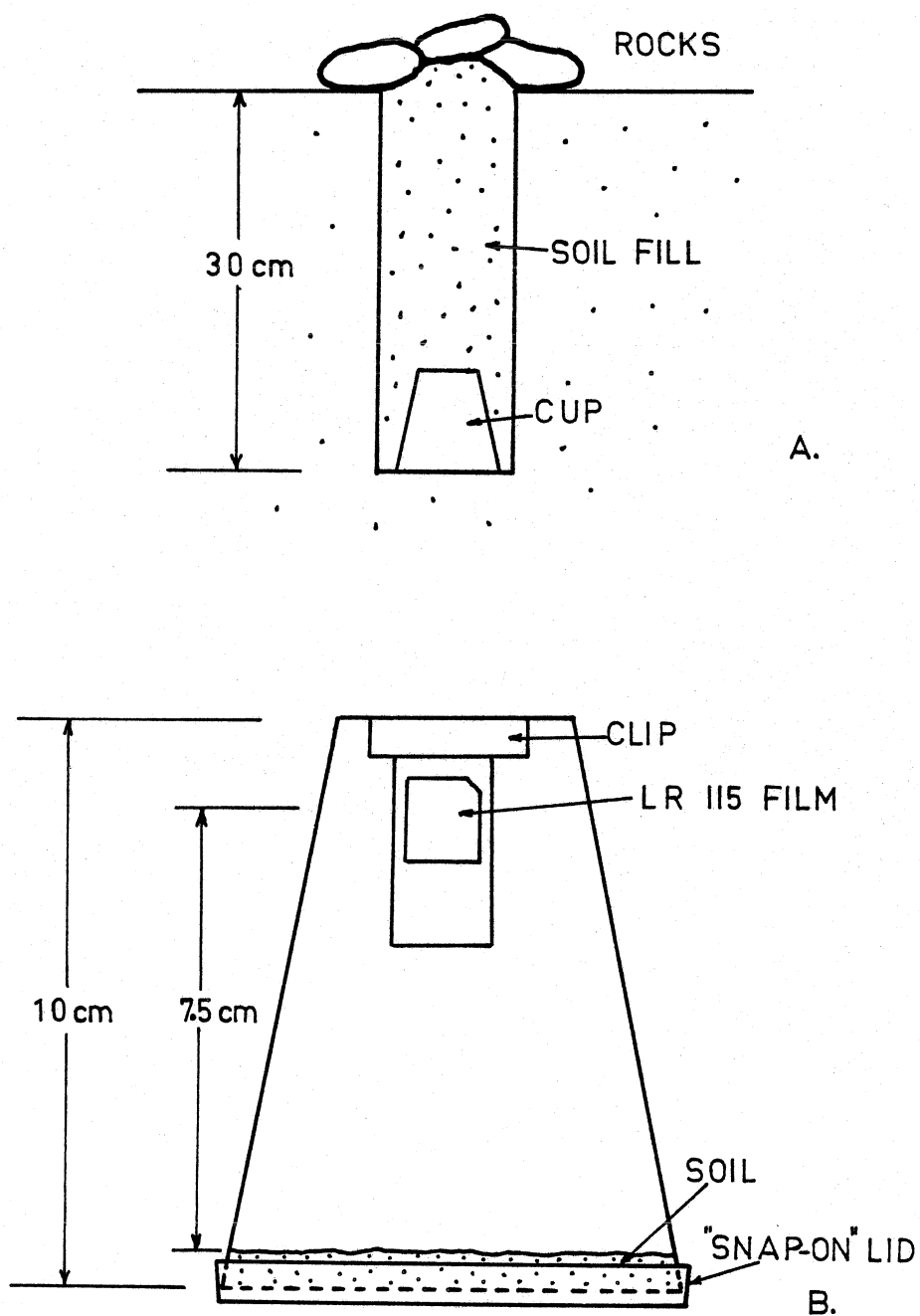


Figure 11. Radon measuring device. A: Field set-up of buried, inverted open cup. B: Set-up for laboratory measurements of radon emanation from soil; 30 cc of soil placed in inverted cup with lid affixed. Thickness of soil layer is approximately 0.8 cm.

relative differences in radon concentration within an area. After field exposure for a period of 4 to 5 weeks, the cups are retrieved. The film and cardboard strip are immediately removed from the cup and wrapped individually in sheets of clean paper and sealed in plastic bags. The use of the 4 to 5 week exposure time is both to enable an adequate number of tracks to be detected in the low radioactivity basaltic terrain and to compensate for the variations in radon concentration due to short-term phenomena. Integrated measurements of this type provide a more realistic radon concentration characteristic of a site than is possible by short-term or spontaneous measurements.

By suspending the film within the inverted cup an air-space is created and the film is maintained approximately 7.5 cm above the material in the base of the measurement hole. This configuration essentially allows only those alpha-particles attributable to radon gas to reach the film (although alpha-emitting daughter isotopes of radon may also be detected). What is being detected therefore, is the concentration of radon within the ground gas measured as alpha-particle tracks per area of film per unit time.

D. Film Developing

On return to the laboratory, the film strips are unwrapped and rinsed with distilled water. The identification number of each film, which is written on the cardboard

backing, is then scratched onto the upper part of the film. The film is detached from the cardboard strip, and any remaining adhesive material from the fixing tape is removed. The constant temperature bath used for film development is shown in Fig. 12. The film strips are clipped into plastic paper clips which are mounted on a plexi-glass platform suspended at the top of the NaOH developing solution. The set-up used enabled 12 films to be developed at one time. It was also convenient in that it enabled all 12 films to be emersed and removed from the solution together ensuring the same degree of emulsion etching on each.

As a control to determine that each set of films was developed satisfactorily one standard film was included in each set. These standards were exposed to an Am-241 reference source for 5 minutes. The source used was a 24 mm diameter aluminum planchet (New England Nuclear) on to which Am-241 was electrodeposited on a platinum disc forming a 10 mm diameter active area, and had a nominal activity of 4.48×10^4 dpm. The source was mounted on a plastic cylinder, with a 10 mm diameter internal hole, so that it would be 22 mm above the film producing deceleration of the emitted alpha particles.

The bath set-up used was a Blue M "MagniWhirl" constant temperature jar bath. Heating of the liquid in the bath was by stainless steel tubular elements of low watt density. A circulator plate at the base was periodically automatically

pulsated to circulate the contained water. A 1000 ml bell jar containing the NaOH solution was suspended within the larger bath. To improve the temperature stability, the outer water jacket was encased in aluminum foil and thermometers were affixed into both baths. The temperature of the solution in the inner bath could be maintained within 0.2°C during each developing and within a range of 59.95 to 60.35°C between developing different film sets. The temperature of the water in the outside bath was maintained between 60.35 and 60.55°C .

A small, low speed, magnetic stirring wheel was fitted into the inner bath (Fig. 12) to circulate the NaOH solution. Although this procedure disagrees with the developing instructions provided with the films, it was found that when the solution was not circulated uneven developing of the film resulted. Costa-Ribeiro and Lobão [1975] also found that without agitation, the cellulose nitrate emulsion occasionally peeled off, or became badly damaged. This developing of the films in an alkaline solution intensifies the tracks produced in the emulsion by the passage of alpha-particles. The solution used was a 10% (2.5 N) solution of reagent grade NaOH in distilled water. The period of developing time was 65 minutes for most films used, however, the optimum developing time varied slightly between different batches of the same type of film from the manufacturer. Consequently, for each new batch of film purchased, the

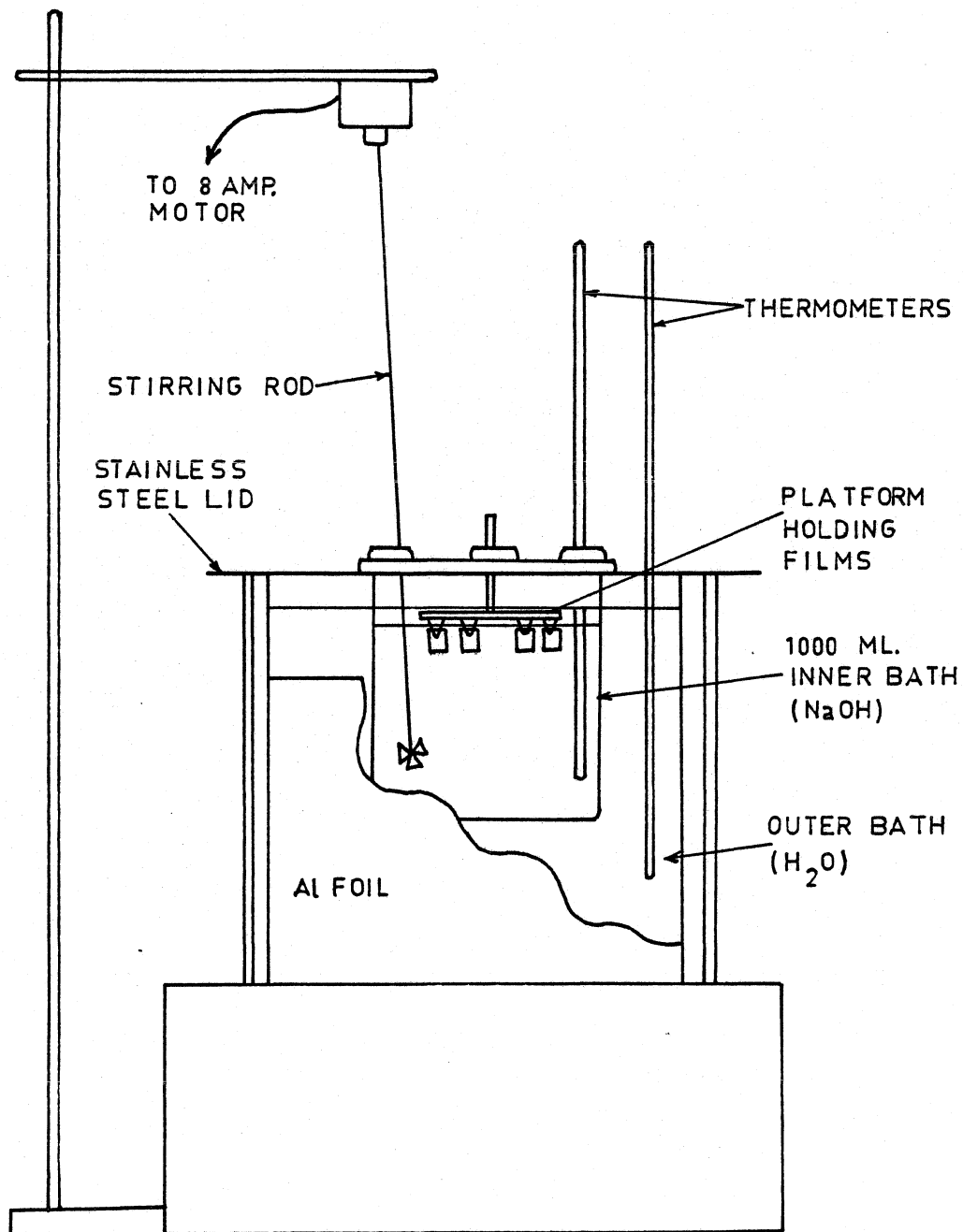


Figure 12. Constant temperature bath used for film development.

optimum developing time needed to be determined. The usual procedure for this was to use 6 films all exposed in the same fashion to the Am-241 source, and during developing remove one at each one minute interval during a period of from 60 to 65 minutes. The only batch of film that required significantly different developing, was an early one used which needed 90 minutes, due to the 13 μ m emulsion thickness.

It was found that because of the large volume of solution used (1000 ml) six or seven sets of films could be developed before a fresh solution was required. Some loss of solution did occur however due to evaporation, but was satisfactorily rectified by returning the solution to its origin level by adding distilled water after each period of development. Film development was satisfactorily timed by a household cooking timer. The platform holding all 12 films was then removed and placed in a beaker of water. The films were individually rinsed under running water, removing all solution and partly-dissolved emulsion. After this each film was patted dry with laboratory tissues and mounted, emulsion side up, on glass microscope slides by adhesive tape.

The perforations (tracks) in the films were counted manually with a binocular microscope at 100 x magnification. To enhance the tracks and enable easier counting a green -polyester filter (Kodak Wratten Filter N^o 40) was emplaced below the stage of the microscope. The green provided a

high contrast against the red of the emulsion enabling easier counting and providing a greater distinction between tracks, and perforations due to scratches or blemishes. Tracks are however quite distinct and appear as even-edged circular holes under magnification. To improve the counting statistics, tracks were counted over one square centimeter of film. The most convenient way to do this was by placing a template with a hole of that size over the film and counting across the film along lines 1 mm apart. The template was always placed over the same area on each different film to maintain uniformity of procedure. Every tenth film was recounted as a confirmation of uniformity in track selection.

E. Determination of Background Emanation from Soil

The emphasis of these surveys is to measure that concentration of subsurface radon that is related to anomalous conditions of temperature and permeability. In the measurement configuration used, however, a certain contribution to the total alpha-tracks detected is derived from the soil immediately below the cup. To distinguish this from the contribution of deeper radon, the radon emanation from the soil was determined in the laboratory and subtracted from the total count detected in the field [Fleischer and Mogro-Campero, 1978; Cox, 1980]. As an optimum field procedure in these types of surveys it was found not to be necessary to collect a soil sample for this background determination at

each measurement location, but to collect a suite that is representative of the soil types within an area. Typically, approximately 15% of the locations were sampled and it was found that the areal extent of surface material types could be satisfactorily determined for this purpose by the use of field notes and air photographs.

The resultant value corrected in this manner may be positive or negative and demonstrates the broad characteristics of ground gas movement within an area. Negative values occur where the concentration of radon in the ground gas at a location is less than the amount of radon emanating from the soil. This condition indicates a lack of upward movement of ground gas, and may be caused by low permeability ground. There is, however a common occurrence of negative values in areas peripheral to zones of positive values. This pattern strongly suggests the existence of ground gas convection systems in which the negative values are caused by lateral or downward movement of ground gas. If downflow does occur the radon concentrations may be further lowered due to an intrusion of atmospheric air and subsequent dilution of the radon content. The concept of ground gas convection has also been proposed by Mogro-Campero and Fleischer [1977] who consider that the existence of such convection cells implies downdrafts as well as updrafts. They also suggest that the location of updrafts could shift with time if the cells were not fixed by geological non-conformities.

The samples of soil for background determination were collected from the base of the holes in which the cups were placed and immediately sealed in plastic bags. In the laboratory, 30 cc of the total sample was placed in a capped, inverted cup containing a detecting film and set up as in (Fig. 11 b). The samples were not dried, in an attempt to more closely approximate the in situ conditions. The use of dry soil can produce a more efficient emanation of both Rn-222 and Rn-220 and may differ from in situ conditions. (Experiments by Fleischer and Mogro-Campero [1978] showed that dry soil produced 100% of the Rn-222 and 92% of the Rn-220 expected, and that saturated soil produced only 85% of the Rn-222 and 2% of the Rn-220 expected.) The cup was tapped on a hard surface to evenly distribute and compact the soil producing a soil layer on the average 8 mm thick. The cups were then left for periods of 4 to 5 weeks before the films were removed and processed in the usual fashion. The suitability of this method is that it directly measures the radon emanation from the soil. Results from indirect methods such as determination by measuring uranium or thorium contents or the radium parent radionuclide tend to be dependent on the distribution of these elements as well as the specific suite of minerals in the sample [Fleischer and Mogro-Campero, 1978].

F. Units of Radon Concentration

After developing the field exposed films the total number of tracks per square centimeter (T/cm^2) was counted.

The mean of background tracks from unexposed films was then subtracted from this value. This background is the mean number of parasitic tracks and on unexposed films which have been carried into the field was determined to be 5 T/cm^2 (from 5 individual films). It is possible that some of these background tracks are derived from cosmic radiation, and their possible existence is noted by the film manufacturer [Kodak]. The resultant value is then divided by the period of field exposure (hours). As the number of tracks (T) detected is relatively low in the environment of Hawaii, the results are presented in units of $\text{T} \cdot 10^{-2} / \text{cm}^2 / \text{hr}$. It was determined that the mean background of the films used for soil emanation, which have remained in the laboratory, was 3 T/cm^2 (from 17 individual films).

In summary, the corrected value representing the contribution of deeper radon is determined by:

$$\frac{[\text{T/cm}^2 - 5 \text{ T/cm}^2]}{\text{hours}} - \frac{[\text{T/cm}^2 - 3 \text{ T/cm}^2]}{\text{hours}} = \text{T} \cdot 10^{-2} / \text{cm}^2 / \text{hr (R.U.)}$$

Field exposed film	Soil emanation	Corrected units
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The conversion of the radon units derived from alpha-track counting to the volumetric units used in direct analyses of Rn-222 gas should be mentioned. As facilities were not available to carry out direct comparisons an attempt has been made to use other available data. It should be noted,

however, that in the surveys as applied, no membranes were used to isolate Rn-222, and this type of conversion can provide erratic results due to (1) alpha-emitting aerosols being deposited on or near the film and (2) moisture condensation [R.L. Fleischer writ. comm., 1980]. Calibration exposures of one type of alpha-sensitive detector (ally diglycol carbonate, used by Terradex Corporation) have been made by A.C. George [in Fleischer et al, 1979] in a controlled radon atmosphere; those comparisons provided a conversion of 13,000 R.U. = 1 nCi/l ($0.91 \text{ T/mm}^2/30 \text{ days} = 1 \text{ pCi/l}$) and for a cellulose nitrate detector previously used by Terradex, a conversion of 1611 R.U. = 1 nCi/l [R.L. Fleischer, writ. comm., 1980]. Fleischer considers that the calibration for the film used in this study (LR-115) would lie between that of the above detectors; an intermediate value is provided using the conversion in Mogro-Campero and Fleischer [1977] of $0.7 \text{ T/mm}^2/30 \text{ days} = 1 \text{ pCi/l}$. Converting this to the units used, an approximate conversion of 9700 R.U. = 1 nCi/l (of Rn-222) is obtained. This value is a rough approximation only but can be used for very generalized comparisons. For example, using data from Table 6, the measured radon concentration in wellhead steam discharges from HGP-A is on the order of 7372 to 23280 R.U.

A calculation of the conversion of alpha-tracks detected on the films from the Am-241 standard reference source to Curies was made. Allowing for radioactive decay, the

the activity of the source was 4.47×10^4 dpm, which equals 20.135 nCi. Using 17 different films, the mean number of tracks detected was 96924 T/cm²/hr. This produces an equivalent of 1 nCi = 4814 R.U., however, it cannot necessarily be directly compared to the above conversions due to the lack of a volumetric factor.

G. Gamma Spectrometry

To provide a comparison to the values of radon emanation measured from soils concentrations of uranium and thorium were analysed by gamma spectrometry. This was also carried out to establish whether the concentrations of these parent radionuclides in soils are similar to those in the lavas from which they are derived. Uranium and thorium analyses of several sublimates deposited from hydrothermal fluids and precipitate from HGP-A well were also made.

Total samples of soil were oven dried at 110°C for 12 hours, then tightly packed into 20 cc polypropylene, screw-cap, liquid scintillation vials (approximately 35 gm of sample). The vials were made air-tight by further sealing the caps with a high vacuum resin sealant. The vials were then left for at least 21 days before analysis to allow the radon isotopes to reach equilibrium with their parent nuclides; Rn-222 reaches 99% of its equilibrium activity with Ra-226 after approximately 26 days. This equilibrium between Ra-226 and Rn-222 is established by the following

equation:

$$A_t = A_e (1 - e^{-\lambda t})$$

where A_t = activity due to radon at time t

A_e = equilibrium activity due to radon

and λ = radioactive decay constant for Rn-222

Gamma-ray detection was performed using a 7.6 x 7.6 cm well-type [thallium-activated sodium iodide (NaI (Tl)) crystal] detector were fed into an Ortec 113 preamplifier, amplified by a Tennelec TC 205 amplifier and then accumulated in 410 channels of a 1024 channel Hewlett Packard HP 5422B pulse height analyser. The pulse light spectrum stored in memory could be displayed on a cathode tube, graphed on a strip chart recorder and recorded digitally on a typewriter. This spectrometry technique is discussed in detail in Adams and Gasparini [1970] and elsewhere.

To ascertain that no excessive drift of the spectral peaks occurred during detection, probe standards of Cs-137 and Co-60 with reference peaks of 0,662 MeV and 1,33 MeV, respectively, were counted before and after each sample; the Cs-137 standard for 2 minutes and Co-60 standard for 5 minutes. Soil samples were counted for periods of between 4000 and 5000 minutes. If the drift of the peaks was greater than two channels, the instrumental gain was adjusted to obtain the correct peak positions using the probe standards,

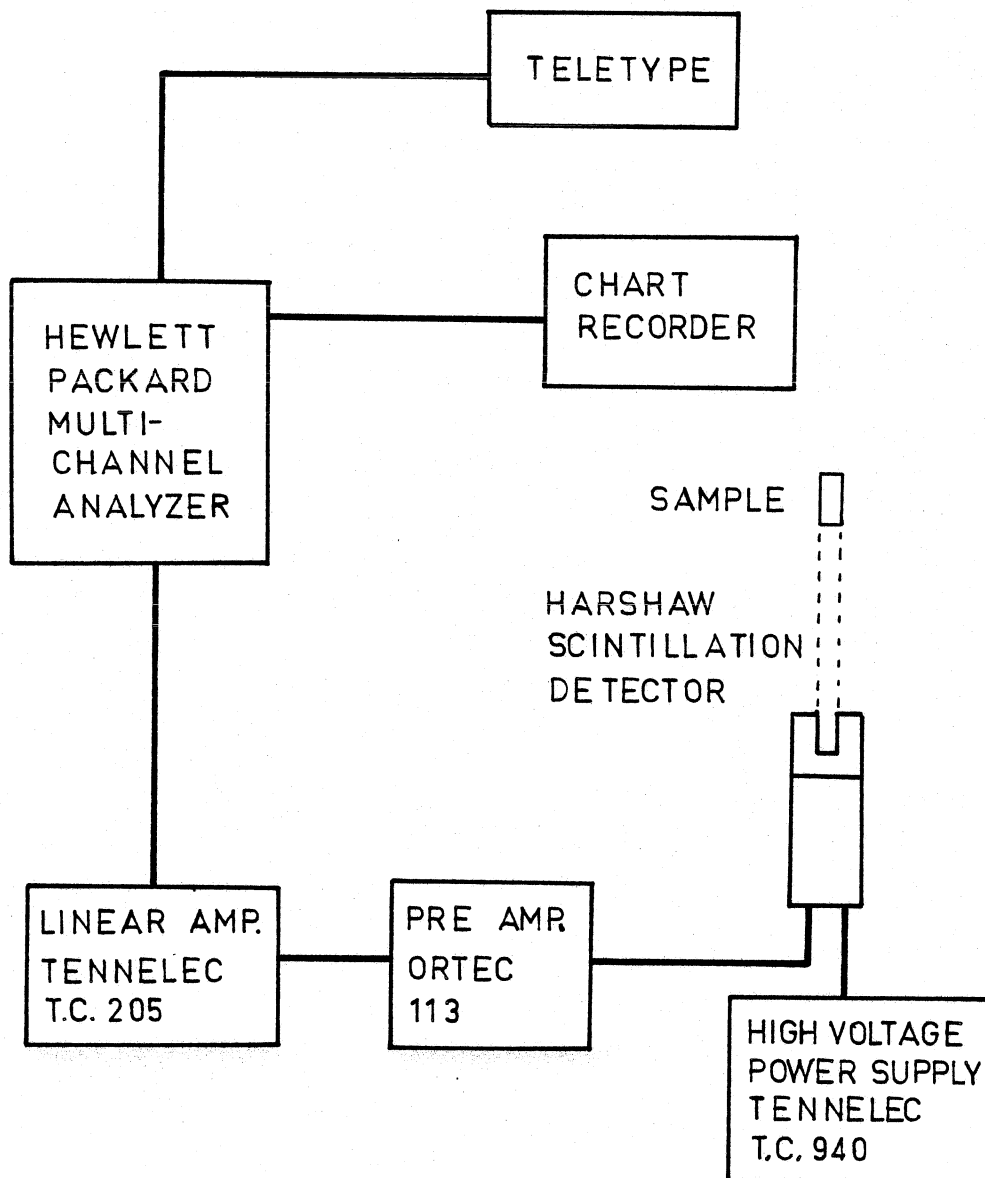


Figure 13. Sketch diagram of gamma-ray spectrometry system [after Fankhauser, 1976].

and the sample rerun.

To calculate the uranium and thorium concentrations in the samples, calibration constants were first determined using ball-milled standards of uranium and thorium ore (NBL 76-B: U = 101 ± 2 ppm; NBL 82-A: Th = 201 ± 2 ppm, U = 8 ppm; New Brunswick Laboratory, Atomic Energy Commission, New Jersey). The background value of the gamma spectrometer was determined by using a NaCl sample. The background and ore standards were packed with the same geometry as the samples and counted for periods of between 4500 and 5000 minutes. After counting, the specific activities of uranium and thorium in the samples were calculated, and these data and the calibration factors were inserted in a computer program ("Analyze", R. Lynch, 1975) to calculate the concentrations of uranium and thorium in parts per million (ppm).

H. Determination of Soil Properties

To obtain a better understanding of some of the conditions of radon emanation from surface material several properties were determined for a set of 40 soil and sublimate samples collected from different environments within Hawaii (Fig. 35). The results of these determinations are tabulated in Appendices 4 and 5.

1. Radon Emanation and Uranium and Thorium Concentration

The methods used to determine these, are described above.

2. pH

The pH of the samples was measured using a pH meter with a combination electrode. A split of the total sample which had been oven dried (110° for 12 hours) was sieved through a 1.00 mm size mesh to remove rock fragments and coarse organic material. The sieved material was then mixed into a slurry of 2 : 1 distilled water to sample by volume. In each case the electrode was allowed to remain in the slurry for two minutes before the reading was made. Each measurement was repeated, and the mean of the two used; the error of all repeat measurements was within $\pm 0,2$ pH units.

3. Percent Organic Content

The organic content of the soil was determined using the method described by Hesse [1972]. A split of the total sample was oven dried for 24 hours at 110°C to remove the contained moisture, and the weight of the dried sample determined. The sample was then placed in a Pyrex glass beaker in a high temperature muffle oven for 16 hours at 400°C . The percent organic content was determined by the weight loss on reweighing. This method is considered to provide an estimate of organic matter (in non-calcareous soil) that is sufficiently accurate for most purposes. Keeling [1962 in Hesse, 1972] determined that the heating of clays at 375°C for 16 hours destroyed 90% of the carbonaceous material without loss of structural water. Further, it has been determined that the greatest part of weight loss due to

clay mineral water occurs in the temperature range of 450 to 600°C [Ball, 1964 in Hesse, 1972]. In the case of the sublimate samples it was considered after microscopic examination that there was negligible organic content, but the use of this method produced false results apparently due to inorganic volatiles being driven off.

4. Percent Grain Size Fractions

A split of the total sample was dried by heating in an oven for 24 hours at 110°C. The weighed dry sample was shaken through a set of brass sieves. The percent by weight was then determined for each of the following size fractions: > 1.0 mm, 1.0-0.5 mm, 0.5-0.25 mm and < 0.25 mm. The organic material was not removed from the samples before sieving as it was desired to measure the in situ size distribution of the total sample.

I. Mercury Concentration of Soil

As mentioned, surveys measuring the concentration of mercury in the soil were carried out in conjunction with the radon measurements. These mercury studies are not discussed in detail here, but the results are referred to where they are relevant to the radon surveys.

Soil samples were taken at spacings of 0.5 to 1 km at depths of 20 to 30 cm, and sealed in plastic bags. In the laboratory the samples were air dried and sieved to < 0.5 mm.

Analysis was by a Gold Film Mercury Detector [M^cNerney et al, 1972] which measures total mercury collected as Hg⁰ on gold films after high temperature combustion. The concentration of mercury is reported as parts per billion (ppb) and the precision of analysis was determined to be $\pm 5\%$. Background concentrations of mercury in analysed soil ranged from 20 to 150 ppb; the lowest background values were found in immature soils derived from young lava flows, and the highest values (usually > 80 ppb) from mature soils with a high organic content.

V. ANALYTICAL RESULTS

A. Precision of Radon Results

1. Film Counting

The reproducibility of alpha track counts on the films can be determined from the standard films which were exposed to the Am-241 source. The mean number of tracks per field of view (1.3 mm^2 ; 100 x magnification) for a random 40 films was 112.5, or 86.5 tracks per mm^2 . The standard deviation of tracks counted on these films was 6.8%.

2. Radon Emanation From Soil Samples

To determine the precision of measurements of radon emanation from soil samples duplicate measurements on 5 samples (all 30 cc) were made. The results are included in Appendix 1. The average deviation for these measurements was determined to be approximately 6.9%.

A possible source of additional error in the procedure used to compensate for the background emanation of radon from the soil is in the method of classifying the soil type at each measurement location. As described, this emanation is not determined for each field measurement location, but representative soils are taken for the different types occurring within an area. It is difficult to quantify a possible degree of error in selecting the background value in this manner. However, in 8 cases (2 in each area surveyed) where a value for soil emanation had been estimated on this

basis, samples were subsequently taken for determination. In these cases the measured values were within $\pm 5\%$ of the estimated value. This possible error becomes more significant in the zones of low total field counts where total tracks measured are less than a factor of about 2 above the emanation from the soil. In the anomalous zones with high radon concentrations the total field counts are in the order of 4 to 20 times the soil emanation and so any such error is not significant. Overall, as the areal pattern of radon concentration is being sought, any differences introduced by this variation have negligible effect on the mapping of those patterns.

3. Total Error in Radon Measurements

Because of the difficulty in assessing the precision of the final soil background-corrected values due to the numerous variables, an attempt was made to determine the degree of error empirically. This was done by setting up quadruplicate films at one location (P1) which had been monitored for the previous five months. This approach was used as the temporal variations in radon concentration that can occur do not always allow a direct comparison with subsequent measurements at a site. The method used was to set up two films back-to-back, in each of two cups, buried together in the same hole for the same period of time. These results are presented in Appendix 2. The mean deviation was approximately 6%. The maximum error at a typical measurement

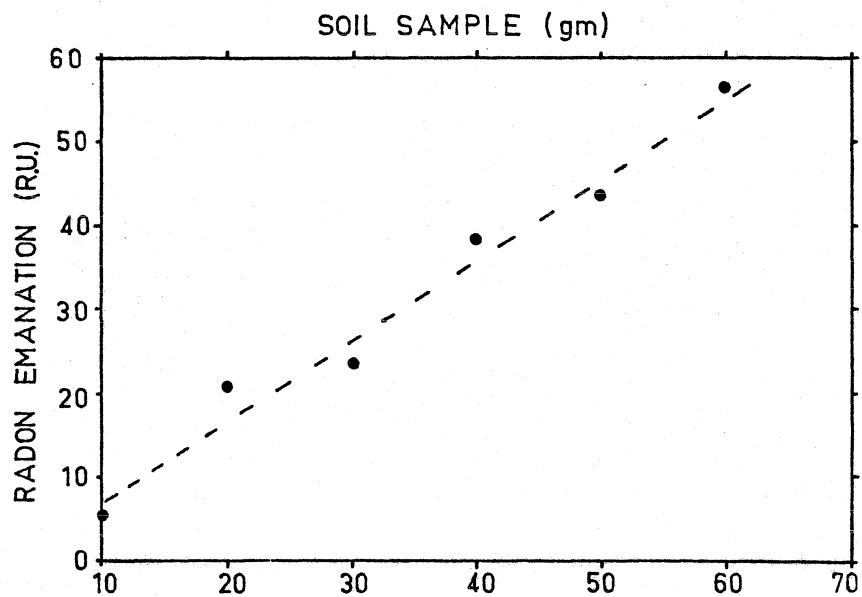
station is indicated to be on the order of 19%, but is more realistically considered to be less than 10%,

4. Variations in Emanation With Soil Volume

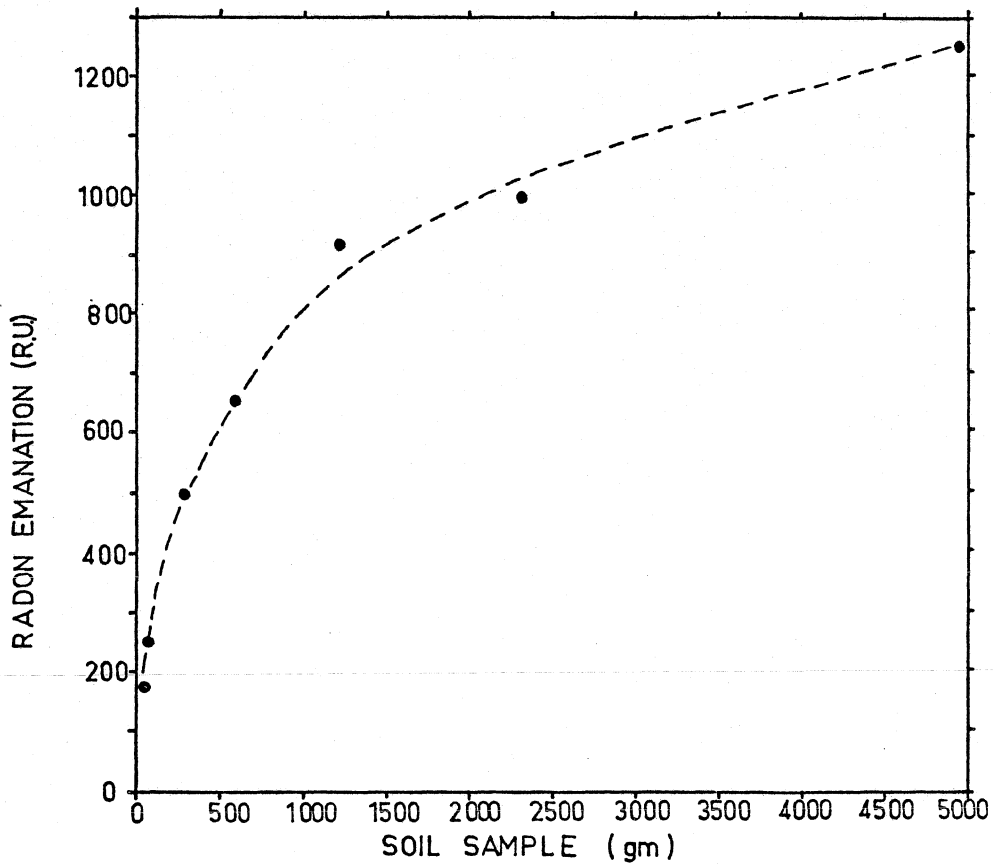
The mean precision of radon emanation determined from soil duplicates of a constant volume (30 cc) was 6.9%, as stated above. The constant volume was used to further standardize the overall method, as it was determined that different values of emanation were measured with different volumes of soil. Broadly, this constituted a linear increase in emanation with volume (Fig. 14,a) for the volumes used (10 to 60 cc).

Fleischer and Mogro-Campero [1978] report an increase in emanation with increases in a large volume of soil for sample sizes of 50 to 4981 g; the data from those authors has been converted to R.U. and plotted in Fig. 14,b. As both of these experiments were in a sealed environment any transport of radon would be by diffusion and not convection. Under these conditions and because of the difference in the diffusion lengths of Rn-222 and Rn-220, increasing the volume of soil can also be used to distinguish between these two isotopes [Fleischer and Mogro-Campero, 1978]. Consequently, because of the short diffusion length of Rn-220 the contribution of this isotope to the air space in the cup is from the volume of soil within one diffusion distance. The diffusion distance, $\ell = (D/\lambda)^{\frac{1}{2}}$ where D = diffusion coefficient and λ = radon isotope decay constant. The Rn-220 contribution,

Figure 14. Radon emanation as a function of soil volume; emanation increases with soil volume. 14, A: linear increase with a small volume of soil (U content approximately 0.9 ppm). 14, B: exponential increase with a large volume of soil (U content of soil, 2.2 ppm) [from Table 4, Fleischer and Mogro-Campero, 1978, with radon values converted to R.U.]. The Rn-222 contribution increases with soil volume, but the Rn-220 contribution remains relatively constant and from within one diffusion distance.



A.



B.

then, should remain relatively constant, while that of Rn-222 increases with volume.

Under in situ conditions a similar model can be used of a relatively constant contribution of Rn-220 from immediately below the cup and a contribution of Rn-222 that is the same as its value in the nearby soil, but controlled by the drop-off in concentration due to the vertical gradient effect (Section III. F) and the depth of the cup relative to the ground surface [Fleischer and Mogro-Campero, 1978]. In measurements at an approximately constant depth, the ratio should be approximately constant. This model, however, assumes that diffusion is the only mechanism by which radon is transported. In those cases where vertical convection is dominant the Rn-222/Rn-220 ratio would be likely to increase as although more Rn-220 would be carried into the cup a relatively greater amount of the longer lived Rn-222 would be transported from below.

Consequently, one question for the field application of these measurements regards the effects due to different thicknesses of soil. This is somewhat difficult to ascertain as the difference in the properties of the soil controlling the diffusion coefficients and emanating power relative to the underlying lavas have to be considered. Field observations indicate that areas with thicker soil produce higher radon concentrations, however, within a particular survey area appreciable differences in thickness (e.g. greater than

≈ 3 m) are required to produce measureable effects on the radon concentration, a situation that was not experienced. These higher values are also a function of weathering, and Wilkening [1974] determined that on the island of Hawaii the Rn-222 emanation from organic soils over old lavas is 6.7 times that from barren lava flows. The variation of radon emanation (from tholeiitic soils in the Puna area) was also indicated by this study to be a function of weathering and alteration. Typical values were, 0.92 R.U. for fine-grained lava and spatter fragments, 1.24 R.U. for very thin soils with a high proportion of fragments, 2.15 R.U. for thin soils with minor organics, 5.12 R.U. for dark organic-rich soils and 8.17 R.U. for "clayey" hydrothermally altered soil.

B. Precision and Accuracy of Gamma Spectrometry

The precision and accuracy are best measured using standard samples, as a determination of precision based on counting statistics may be misleading due to instrumental drift. The standards of uranium and thorium ore (NBL-76 and NBL-82 A, respectively) against which samples are calibrated, were each analyzed three times, but calibrated against a concentration previously determined. The mean deviations for these triplicate analyses were 1.01% and 0.40%, respectively. A uranium standard (NBL-75) of known concentration (510 ± 10 ppm) was also analyzed; the value determined, 509 ppm, was well within the range of the standard.

The precision of the uranium and thorium analyses of the soil samples was determined by duplicate measurements on five different samples (Appendix 5). For uranium the mean deviation was 17.9% (range 9.6 to 32.9%) and for thorium the mean deviation was 31.9% (range 14.5 to 71.6%). Some of these deviations are rather high, and are believed due to U-238 series disequilibrium in some of the samples, and instrumental drift during some of the earlier analyses. Consequently, in those cases the mean of the duplicate analysis has been used for discussion. Considering possible disequilibrium in the U-238 series the results for the uranium contents are probably accurate only to $\pm 15\%$ for the post-erosional and alkalic basalts, and to 25 to 50% for the tholeiitic basalts [Fankhauser, 1976].

C. Discussion of Analytical Results

The results of the various determinations carried out on the samples of soil and sublimate are listed in Appendix 4. These results, especially the correlations of different properties are discussed here.

Grain size fractionation of the soil was conducted to observe if any relationship existed between particle size of the total soil and radon emanation from it. Of the four size fractions separated, that less than 0.25 mm by weight-percent is used for discussion on the assumption that the finest grain size material would have the greatest effect

on emanation. Figure 15 is a plot of the percent of soil < 0.25 mm against radon emanation from the total sample. No systematic relationship is evident between these two parameters for the samples used, however, if those points with the greatest scatter are ignored there is some suggestion of a correlation with negative slope. The lack of an obvious trend is possibly a function of the size fraction used not being fine enough, and the large inhomogeneity of the samples. The samples included volcanic ash, young "gritty" soils from historic flows, dry soils from arid leeward areas and organic-rich soils from high-rainfall areas. It may, however, be of importance that experiments by Starik and Melikova [1957 in Tanner, 1964,b] and F. J. Davis [pers. comm. in Tanner, 1964,b] revealed that the amount of Rn-222 made available to passing fluids may be the same or less from a rock after it is crushed than before. The distribution of points in Figure 15, however, suggests that both the type of soil and its chemistry also influence the amount of radon emanation.

The conclusion reached from other studies [Andrews and Wood, 1972] was that the release of radon from rock particles is proportional to $1/d^{1/2}$ or $(Rn) = C_1 d^{-1/2}$, where (Rn) is the radon release per gram of rock and C_1 is a constant. Expressing the radon release as a percent release and putting the equation in logarithmic form, gives

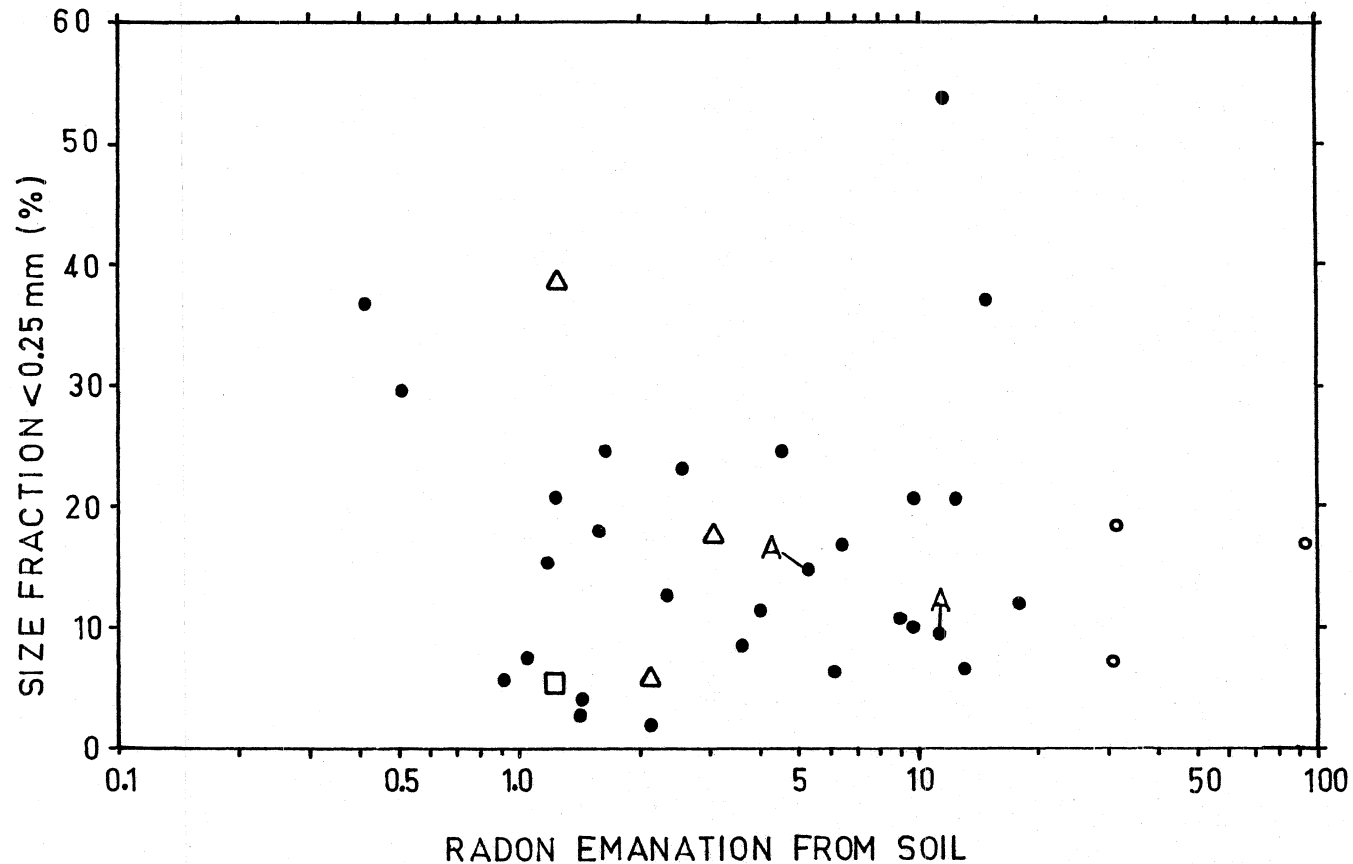


Figure 15. Radon emanation from soil against percent fraction of soil < 0.25 mm. Closed circles: tholeiite and olivine basalt soils (A = hydrothermally altered); open circles: alkali basalt soils; open triangles: basanite and nephelinite soils; open square: volcanic sublimates. No definite correlation is apparent between these properties although there is some suggestion of a negative trend.

$$\log (\% \text{ Rn release}) = -0.5 \log d + C_2$$

where, C_2 is a constant. Therefore, plotting the logarithm of the percent radon release and the particle diameter should produce a slope of -0.5 (i.e. a smaller diameter grain will have a higher percent radon release). Those authors also believe that the percent radon release for a given particle size should be independent of the uranium (or radium) content of the material, but may be influenced by differences in its structure and physical properties. They also noted that differences in mineralogy (of the sedimentary material used) did not cause greatly different percent radon release-particle size relationships. They concluded that the radon content of groundwaters (and by extension, ground gas) is determined by the physical nature of the rock to a much greater extent than by the uranium content. Marsden and Watson-Munro [1944] concluded that the radioactivity of soil is dependent on the rocks from which the soils are derived rather than the type of soil. To observe if this may be the case, averaged values of uranium and thorium concentrations in soils are compared with the concentrations in lavas from which they are derived (Table 7). These data are in reasonable agreement, considering the limited ranges of uranium and thorium concentrations found in basaltic rocks and the varied effects of weathering on the soils.

In figure 16 the weight percent of organic material in the soil is plotted against the radon emanation from the

TABLE 7

Comparison of Average U and Th Concentrations in Lavas and Derived Soils

Type	Lava				Soil			
	U	Th	Th/U	No. Samples Averaged	U	Th	Th/U	No. Samples Averaged
Tholeiitic basalt	0.396	1.178	3.0	6	0.579	1.221	2.1	7
Olivine basalt	0.233	0.618	2.7	5	0.359	0.746	2.1	2
Alkalic basalt	1.217	3.324	2.7	3	1.414	5.301	3.8	3
Basanite, nephelinite	0.532	2.013	3.8	2	0.909	2.344	2.6	1

Values in ppm

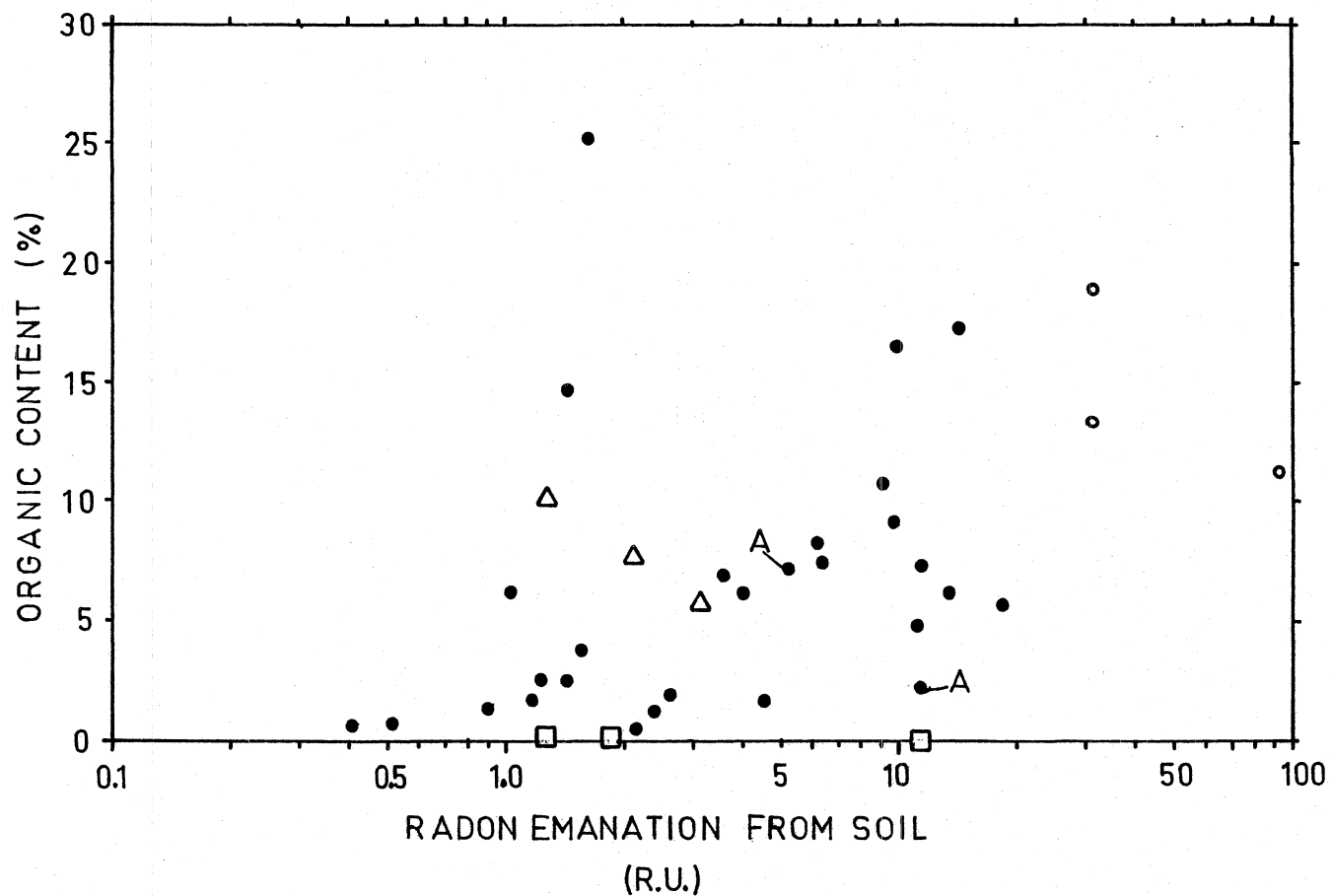


Figure 16. Radon emanation from soil against percent organic content. Closed circles: thoeliite and olivine basalt soils (A = hydrothermally altered); open circles: alkali basalt soils; open triangles: basanite and nephelinite soils; open squares: volcanic sublimates. No real trend is apparent, but some increase in emanation with organic content may occur.

total sample. No real trend is apparent but the data suggest that radon emanation from the soil may be partly influenced by its organic content. This could possibly be a function of radon absorption by organic matter; charcoal is well known to be a radon absorber, and as noted in Section IV. A is commonly used as a trap in collection and analytical procedures. Studies of radon at the ground-air interface [Pearson and Jones, 1965] have also shown that radon is absorbed by organic matter, specifically that radon dissolved in soil water is absorbed and transpired by plants. However, due to the time lapse after sample collection and the time period of emanation measurement, the source of the radon must be within the sample itself. The scatter of points in Figure 16 is believed to be partly a function of the variations in soil type, to which the amount of vegetation-derived organic material is often related. Higher contents of organic matter are more often found in more weathered and finer grain size soils.

Figure 17 plots the radon emanation from total soil samples against the concentration of uranium + thorium for those samples. The distribution of points suggests that radon emanation from soils is significantly greater where the uranium and thorium concentrations are significantly greater. In cases where the uranium and thorium concentrations are relatively uniform other factors appear to have a predominant affect on radon emanation. This indicates

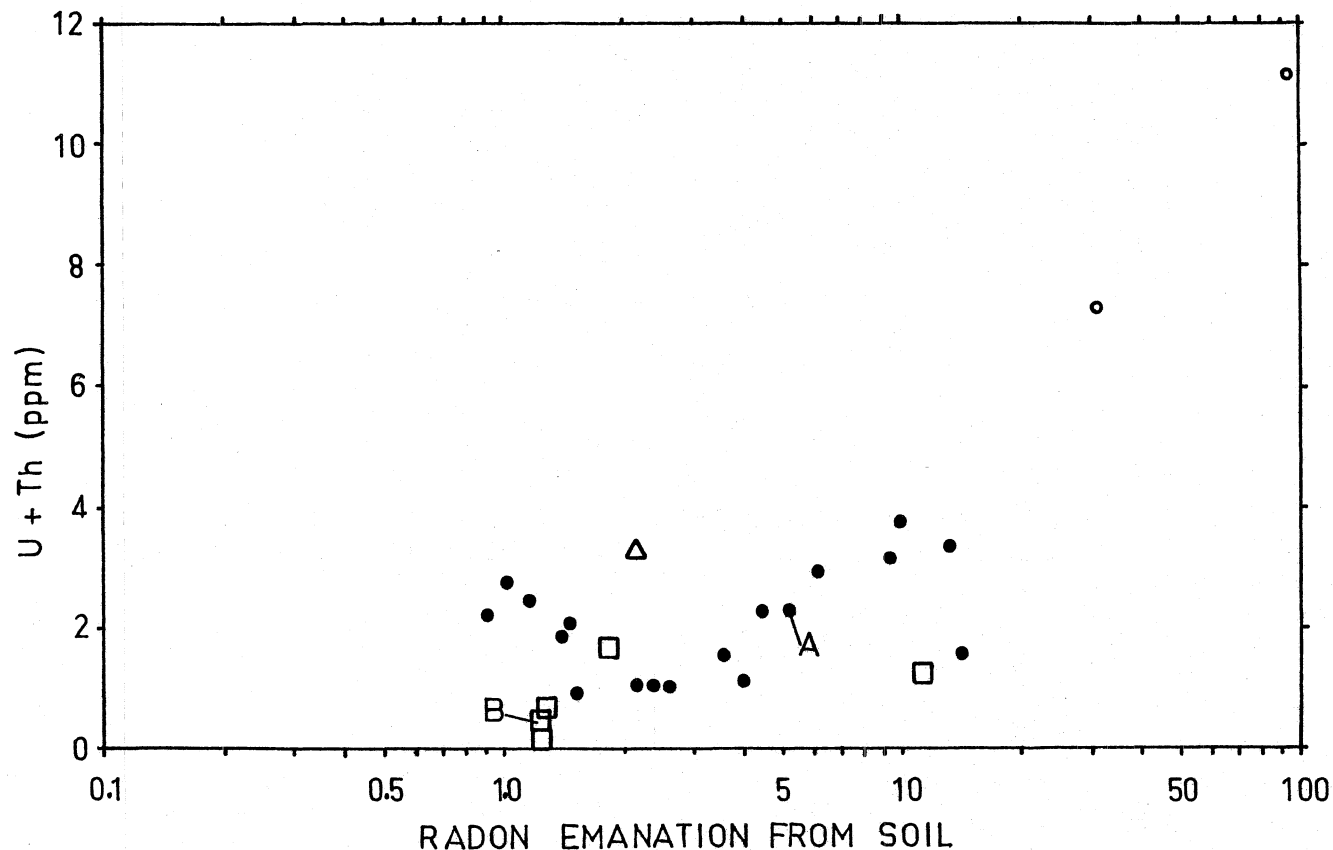
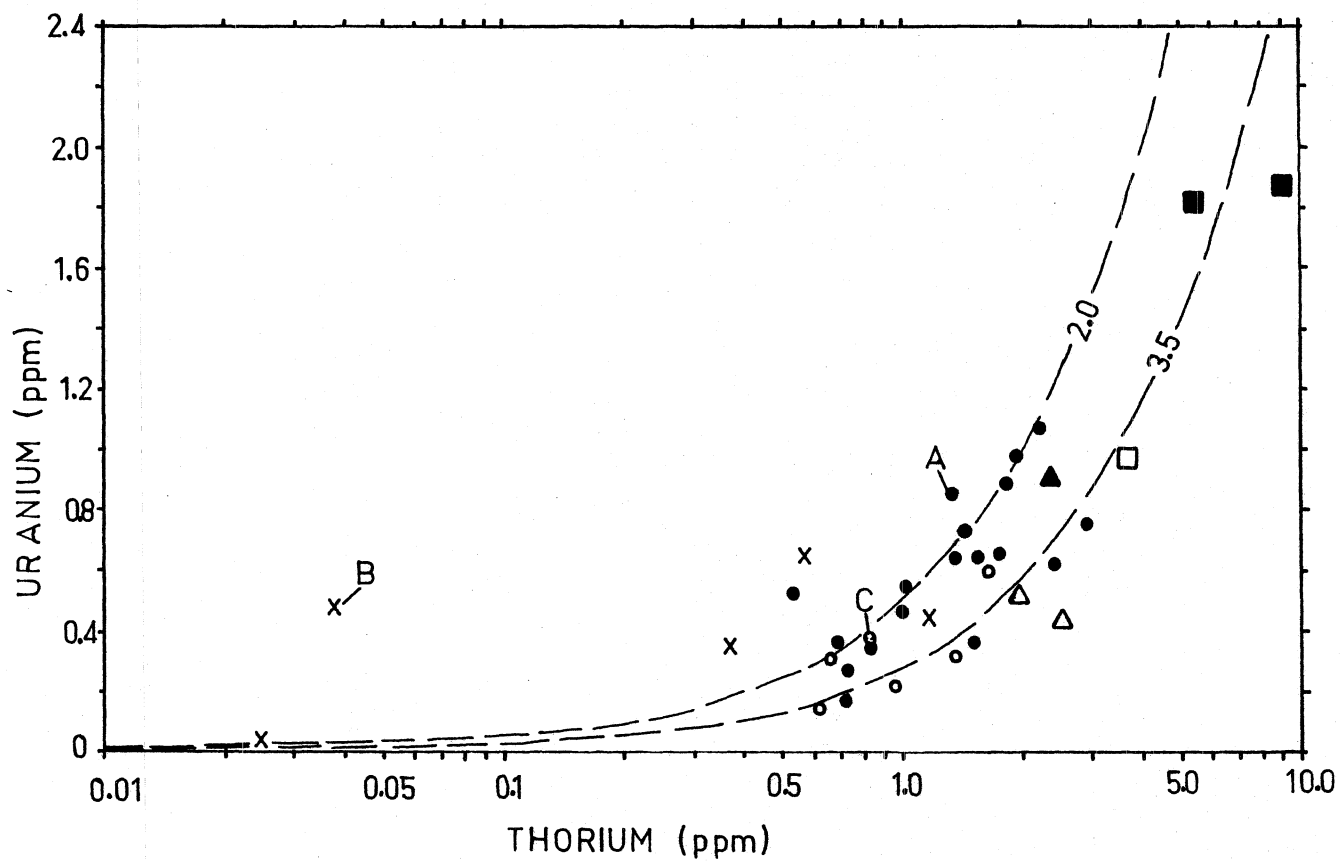


Figure 17. Radon emanation from soil against the concentration of uranium + thorium. Closed circles: tholeiite and olivine basalt soil; open circles: alkali basalt soil; open triangles: basanite and nephelinite soils; open squares; volcanic sublimates (A = hydrothermally altered soil; B = CaSO_4 scale, HGP-A). A greater emanation of radon with higher U + Th concentrations is suggested; however in the case of relatively uniform U and Th concentrations, other factors influence radon emanation.

that overall the amount of radon emanation measured is related to the concentration of the parent nuclides. The very short half-lives of the daughter radon isotopes relative to the age of the parent nuclides demonstrate that they must be solely a result of the decay of those nuclides. A comparison was also made of the uranium + thorium concentrations of the soils and the corrected radon measurements to determine if concentrations of the parent nuclides in the near surface material influenced the shallow radon concentration. No correlation was apparent. Wollenberg [1975] made a similar comparison (uranium to total radon), likewise finding no correlation, and a range of alpha-track densities that was much greater than that of the uranium concentrations. The radon emanation was also compared to uranium and thorium concentrations separately. These showed similar distributions to radon against uranium + thorium, and to each other. This suggests that the Rn-220/Rn-222 ratio of radon emanating from the soils is consistent with the Th/U ratio of the soils.

The Th/U ratio of soil, sinters and lavas is demonstrated by the plot of uranium concentration against thorium (Fig. 18). The sample ratios can be compared to the line representing a Th/U ratio of 3.5, a value characteristic of Hawaii lavas. The analyzed lavas and the means of values reported in the literature all fall along, or close to, this ratio line. The soil and sinter samples show a broad

Figure 18. Plot of uranium concentration against thorium (A = hydrothermally altered soil; B = CaSO_4 scale HGP-A; C = tholeiitic basalt HGP-A chloritic alteration). Lavas approximate the Th/U ratio line of 3.5 and most soils are scattered below this line. Samples with a Th/U ratio of well below 2.0 are associated with hydrothermal activity, and indicate uranium enrichment.



	SOILS.	LAVAS.
Tholeiite - olivine	●	○
Alkali	■	□
Basanite - nephelinite	▲	△
Sublimite	x	

TABLE 8

U and Th Concentration and Rn

Emanation From Rocks and Sinters

Sample N ^o	Location	Material	U (ppm)	Th (ppm)	Th/U	Rn (R,U.)	Ra ϕ (pCi/g)
K1	1971 fumarole, Kilauea	*S(minor CaSO ₄)	0,358	0,366	1,02	1,24	0,13
SS II	near " "	*CaSO ₄ + S	0,439	1,189	2,71	1,88	0,16
SS V	Sulphur Bank, Kilauea	*S + clay	0,034	0,025	0,74	1,23	0,01
SS VIII	Puhimau, Kilauea	*CaSO ₄	0,653	0,583	0,89	11,36	0,23
	Scale HGP-A, Puna	*CaSO ₄	0,492	0,039	0,08	1,24	0,18
	HGP-A, 1850 m	Thermal groundwater (\approx 300°C)	B,L,D,	B,L,D,
1960	Puna, 1960 flow	Tholeiite	†0,612	1,683	2,75	..	0,22
FJ56	HGP-A, 1636 m	Tholeiite (chloritic alteration)	0,382	0,828	2,17	..	0,14

TABLE 8 - continued

Sample N ^o	Location	Material	U (ppm)	Th (ppm)	Th/U	Rn (R.U.)	Ra ϕ (pCi/g)
1932	Byrons Ledge, Kilauea	Olivine basalt	†0.313	0.687	2.19	..	0.11
1974	Kilauea	Tholeiite	†0.321	1.333	4.15	..	0.12
1926	Mauna Loa	Olivine basalt	†0.157	0.626	3.99	..	0.06
M.K.	Mauna Kea	Alkalic basalt	†0.970	3.660	3.77	..	0.35
FJ30	Kula Volcanic Series	Hawaiite	0.226	0.954	4.22	..	0.08
	Honolulu Volcanic Series	Basanite	†0.517	1.935	3.74	..	0.19
	Honolulu Volcanic Series	Nephelinite	†0.427	2.505	5.87	..	0.15

ϕ approximation only; calculated from activities ratio Ra-226/U-238 = 2.78×10^6

* X-Ray Diffraction

† B,L, Fankhauser, 1976; pers. comm., 1979

B.L.D. Below limit of detection

scatter, but the majority have a Th/U ratio of less than 3.5. Of significance in this plot are the low Th/U ratios of surface deposited sublimates and the very low ratio (0.08) of a CaSO_4 scale sample from HGP-A. These low ratios suggest enrichment of uranium relative to thorium within these secondary minerals. Interestingly, a sample of HGP-A core from 1636 m depth and which had experienced chloritic alteration, also had a low Th/U ratio (2.17). Also of interest is that a sample of water from the Puna geothermal reservoir (1859 m, HGP-A; Table 8) contained no detectable gamma-emitters, which may suggest precipitation at depth.

A Th/U ratio of 2.0 is also represented on Fig. 18, and those samples with an appreciable relative enrichment of uranium fall above this line. It is interesting to note that several soil samples are included in this category, and are from areas in which anomalous concentrations of radon, or steam discharges occur. This suggests that radioactive minerals may be deposited at the surface or shallow depth, by circulating thermal fluids. If such near surface enrichment does occur, it would provide an additional source of radon and enhance the use of this technique in geothermal exploration.

VI. RESULTS OF FIELD SURVEYS

A. Temporal Variations in Radon Concentration

As has been mentioned above temporal variations in the concentration of radon in ground gas do occur and can be measured. Variations can be very short-term changes over periods of several hours or over slightly longer periods, such as diurnal variations. To determine the characteristics of longer term variations a series of measurement stations was established on the summit of Kilauea and along the east rift, and have been monitored for over sixteen months [Cox and Cuff, in prep]. The data obtained for one of these stations in the Puna area, (T21), are presented (Fig. 19). This station is located adjacent to an area of steaming ground resulting from the formation of an eruptive fissure in 1955 (Fig. 36).

Variations of greater than a factor of 10 were measured at this station and regularly preceded specific seismic events. The frequency of these seismic events is due to the volcanically active environment of the area in which this station is located. This suggests that within an area of this type radon mapping as applied to geothermal exploration should be carried out during a limited period of time, as the purpose of such mapping is to assess the areal distribution of radon concentrations within an area. The temporal variations determined suggest that the inclusion of further

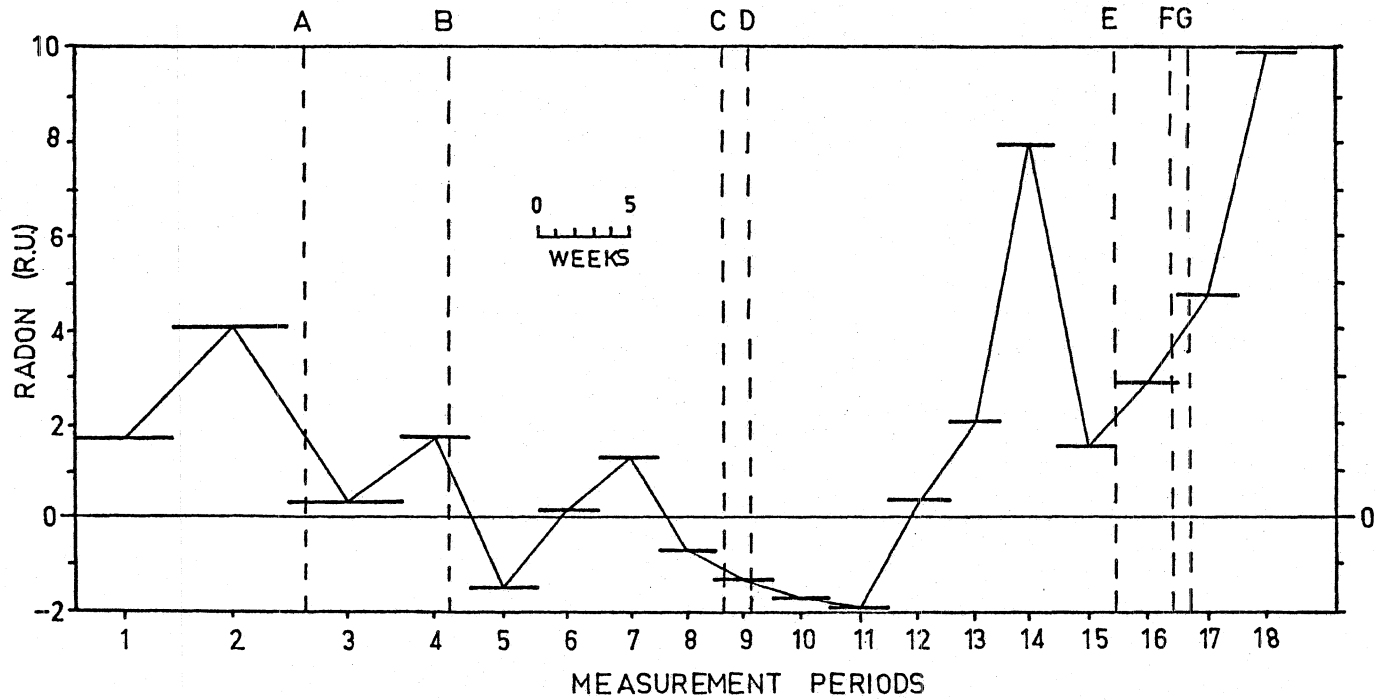


Figure 19. Temporal variations in ground radon concentration, measured at station T21, Puna. Solid bars show time period of measurement; broken lines represent occurrence of seismic events (Appendix 6); location of station is shown in Fig. 36.

measurements made in an area at a later date may tend to distort the pattern of mapped concentrations. This was found to be the case only in the volcanically active areas, and in other areas patterns were readily reproducible. A possible method of adding such extra stations would be to "tie" them into the previous survey by relocating several adjacent stations, and several others throughout the area initially surveyed and normalizing the new measurements to these.

The correlation between increases in radon concentration and volcanic-seismic events will not be elaborated here, although one aspect is relevant and that is that the events producing the increases generate or increase subsurface heat. This in itself is further confirmation of thermal enhancement of radon migration. Briefly, the data at station T21 show that all peaks where the radon concentration increased over a certain value (for this station the zero value) precede an event producing anomalous seismic activity. Based on the recorded behavior at this station the most recent values are significantly anomalous and suggest further seismic activity on the lower part of the rift may be imminent. The events occurring during the measurement periods shown include a magnitude 5.4 earthquake, several seismic 'swarms' indicating intrusions and the Pauahi Crater eruption of 16-17 November, 1979. (The subsequent measurement period, 19, has a value of 6.1 R.U. and a seismic swarm on the

upper east rift zone occurred during that period). It should also be noted, that the period of time between the measured radon increase and the event itself is in part dependent on the location of the measuring station relative to the location of the event.

B. Traverse on Upper Southwest Rift, Kilauea

In order to obtain a better understanding of variations in radon concentration that can occur over close distances, a 3.1 km, northwest-southeast traverse was established on the the upper part of the southwest rift of Kilauea (Fig. 6). The traverse was located in an area where the "soil" is relatively uniform, being largely ash with negligible organic content, except for the northernmost 500 m in which a poor soil formation and sparse vegetation occur. Stations were placed at 200 m spacings, and in addition to radon measurements samples of soil were taken and their mercury content and pH determined.

The results from this traverse are shown in Fig. 20, including the locations of cracks and fissures which were exposed at the surface. The low concentrations of radon reflect the low emanation from the tholeiitic lavas in this area. The concentrations increase locally over cracks and associated fracturing and the overall trend of the values increases southward towards an eruptive fissure formed in

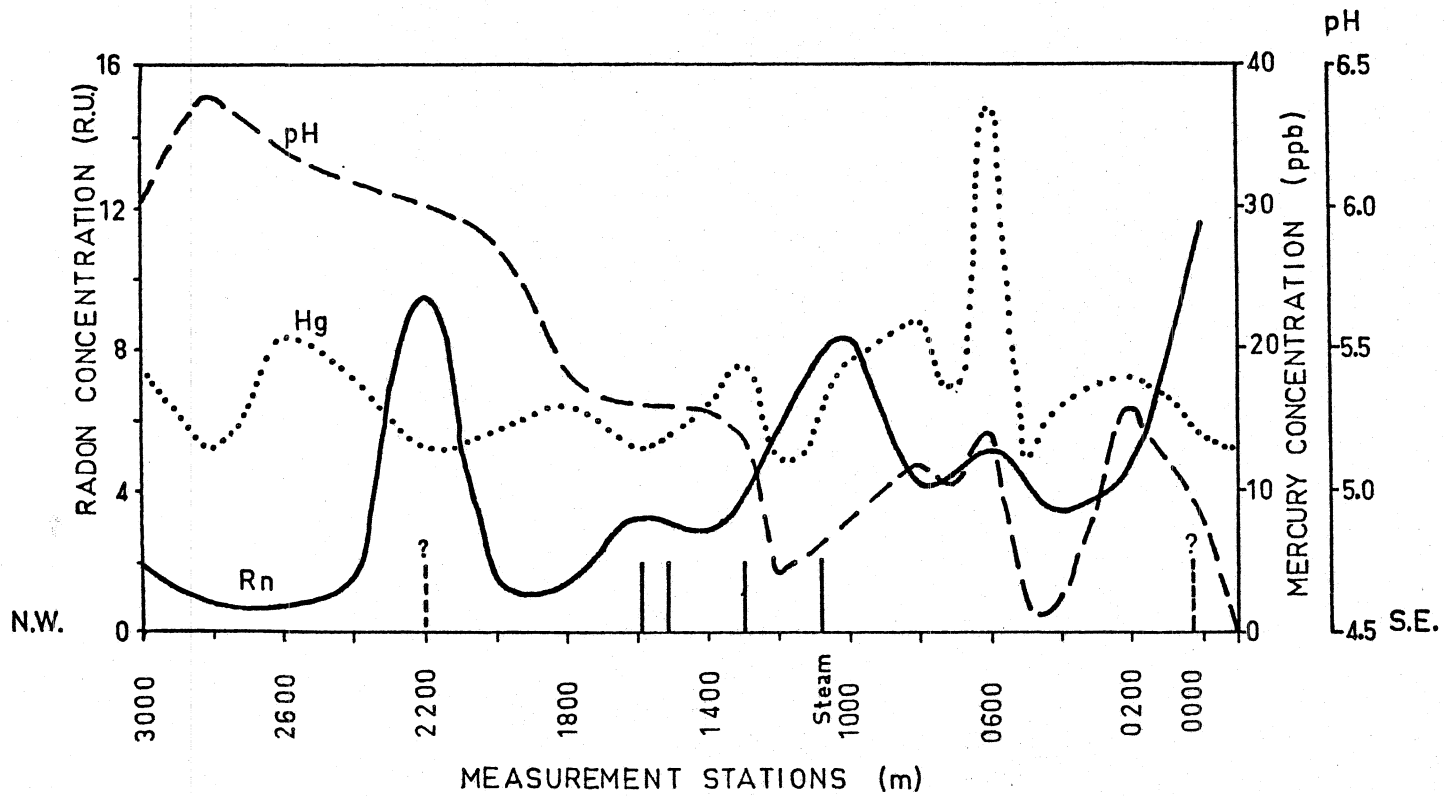


Figure 20. Traverse on S.W. rift of Kilauea; radon, mercury and pH profiles are shown; cracks are denoted by straight lines. Location on Fig. 6.

1974. The localized radon peak at station 2200 is believed to reflect the existence of a buried crack.

In contrast to the radon values, mercury concentrations demonstrate a more uniform trend but also show localized increases. These increases do not directly coincide with the radon highs and location of the cracks, but tend to occur peripherally to them. This effect was also observed in other areas (Figure 26; 31) and appears to be a function of the greater volatility of mercury at higher temperatures, although this is somewhat limited due to the overall low temperatures near the surface. Consequently over outgassing structures (indicated by radon highs) the mercury is partially remobilized and tends to be deposited in peripheral areas of relatively lower temperature. Similarly, a deposition "halo" of mercury has been determined to exist in the cooler outer portions of the Roosevelt Hot Springs geothermal system, Utah [O.D. Christensen, pers. comm., 1980]. The pH of the soil broadly follows a converse trend to the radon concentrations, becoming more acid over the cracks and the areas in which outgassing is both observed and indicated.

These data show that lateral variations in radon concentration are partially dependent on the spatial distribution of permeable structures and consequently the density of sampling. This traverse demonstrates that the wider sampling spacing commonly used in these surveys (0.4 to 0.6 km) will detect lateral trends in radon concentration and outgassing,

but will not necessarily display the detail. This, it should be noted, is the purpose of the surveys in this exploration application and more detailed follow-up mapping can be subsequently carried out to define specific structures if required.

C. Puna Radon and pH Mapping

Contoured values of radon concentrations in the Puna area (Fig. 21) show positive values in association with the rift zone. In the western part of the area positive values extend both north and south of the rift. Within the area of positive values the 5 R.U. contour outlines a well-defined elongate higher value zone coincident with the surface expression of the rift. Another high value zone occurs to the north near the town of Pahoa. The axis of the elongate high follows the trend of those areas in which surface heat occur, from the steaming area of the 1955 fissure in the west (location of geothermal test well 2), through several minor fumaroles near geothermal test well 3 central to the area, to the cinder cone from which the 1960 flow erupted (an area of warm ground) in the extreme northeast of the anomaly and where it terminates. Also of interest is the zone \approx 700 m west of HGP-A where more detailed measurements outline a narrow, low positive-negative zone. On the basis of self-potential surveys, Zablocki and Koyanagi [1979] have postulated a left-lateral fault between this feature and HGP-A. The cause of this low and negative value radon zone is not

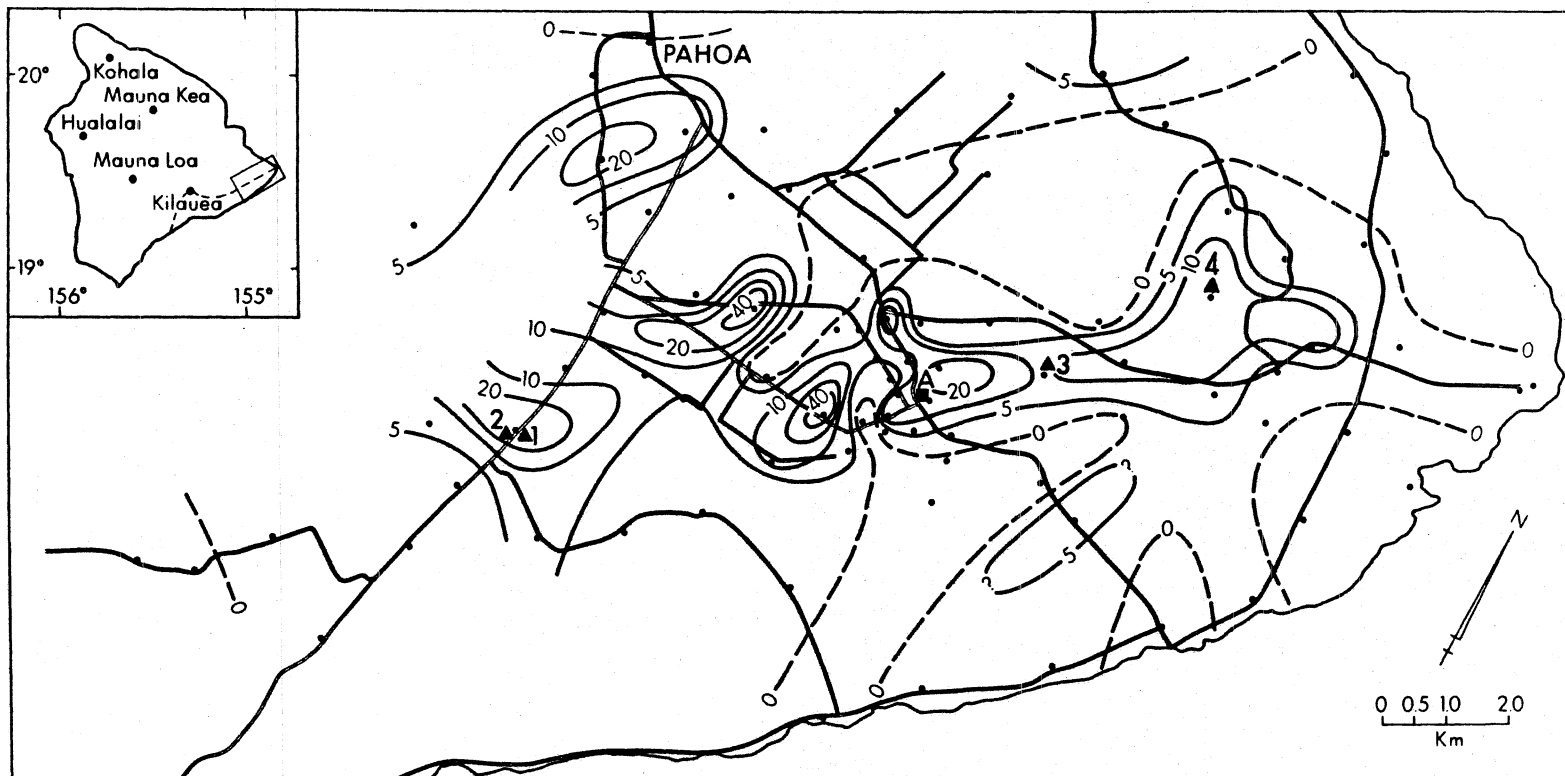


Figure 21. Contours of ground radon concentrations, Puna area, Hawaii. Contour intervals are 0, 5, 10, 20 and 40 R.U. A: location of HGP-A well; triangles 1, 2, 3 and 4 are locations of the 1961 geothermal test wells. The broken line shows the 0 R.U. contour. Solid dots show measurement locations.

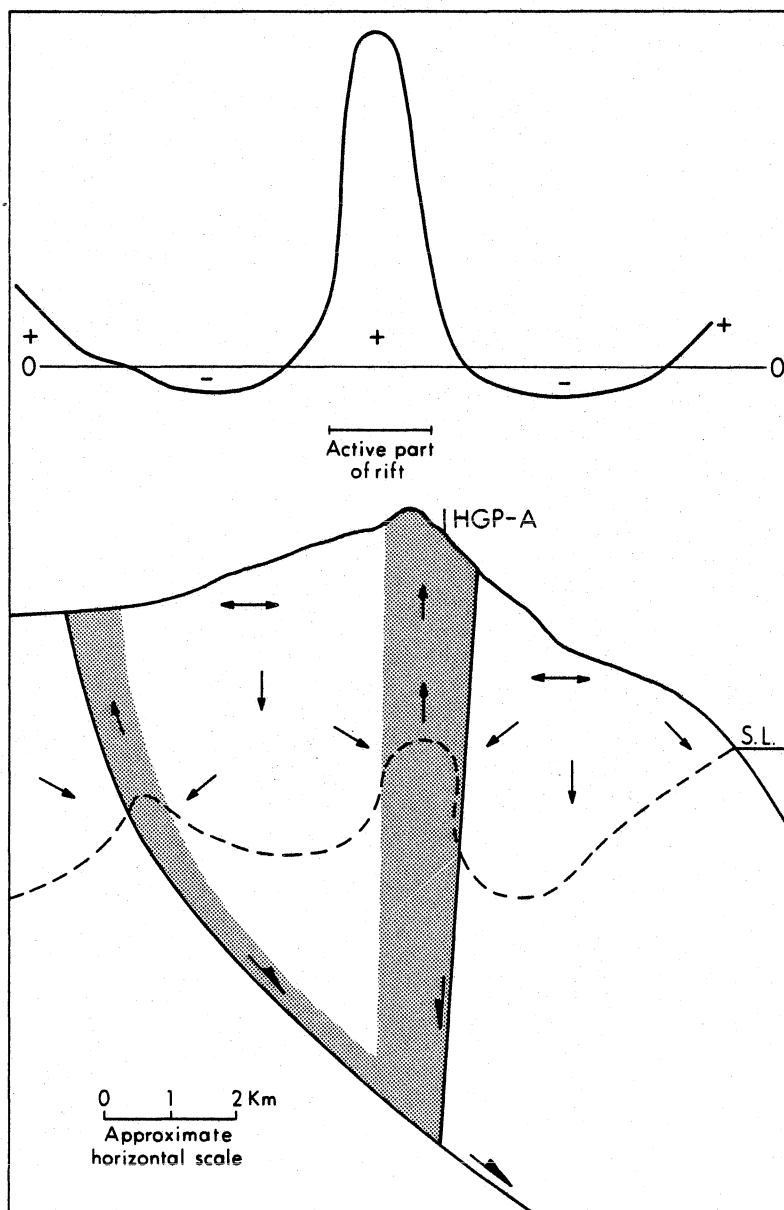


Figure 22. Cross-sectional sketch (NNW-SSE) of lower east rift of Kilauea at Puna. An interpretation of the main rift structure [after Moore and Krivoy, 1964] is shown, with the location of well HGP-A on the most recently active part. The shaded areas represent zones of high fracture permeability. Small arrows represent movement of groundgas and the broken line the Ghyben-Herzberg lens. The upper section of the figure is a profile of the radon concentration from Fig. 21.

fully understood but presumably results from a decrease in permeability below these stations. This may be due to the presence of low permeability intrusive material at relatively shallow depth below this zone, and electrical resistivity surveys by Kauahikaua et al [1980] show a resistivity boundary in that area that they consider may be due to the presence of dike complexes. Significant in the overall distribution of radon concentrations is that peripheral to the central high zone are negative and low positive values. These are considered to be part of a thermally-induced convection system which has developed around the rift structure, and to indicate areas in which lateral or downflow of ground gas occurs, as well as a probable decrease in ground permeability (Fig. 22).

As with the summit of Kilauea the surface material in the Puna area is highly variable with the consequence that the surface pH is also highly variable. The contoured pH values (Fig. 23) are somewhat difficult to interpret in terms of indicating geothermal activity, especially because significant variations can occur over small distances. Several features are, however, evident: isolated zones of low pH over the trace of the rift structure and in areas where outgassing is observable; zones of low pH along the south coast that are possibly related to thermal water leaking from the rift and discharging at sealevel, and low

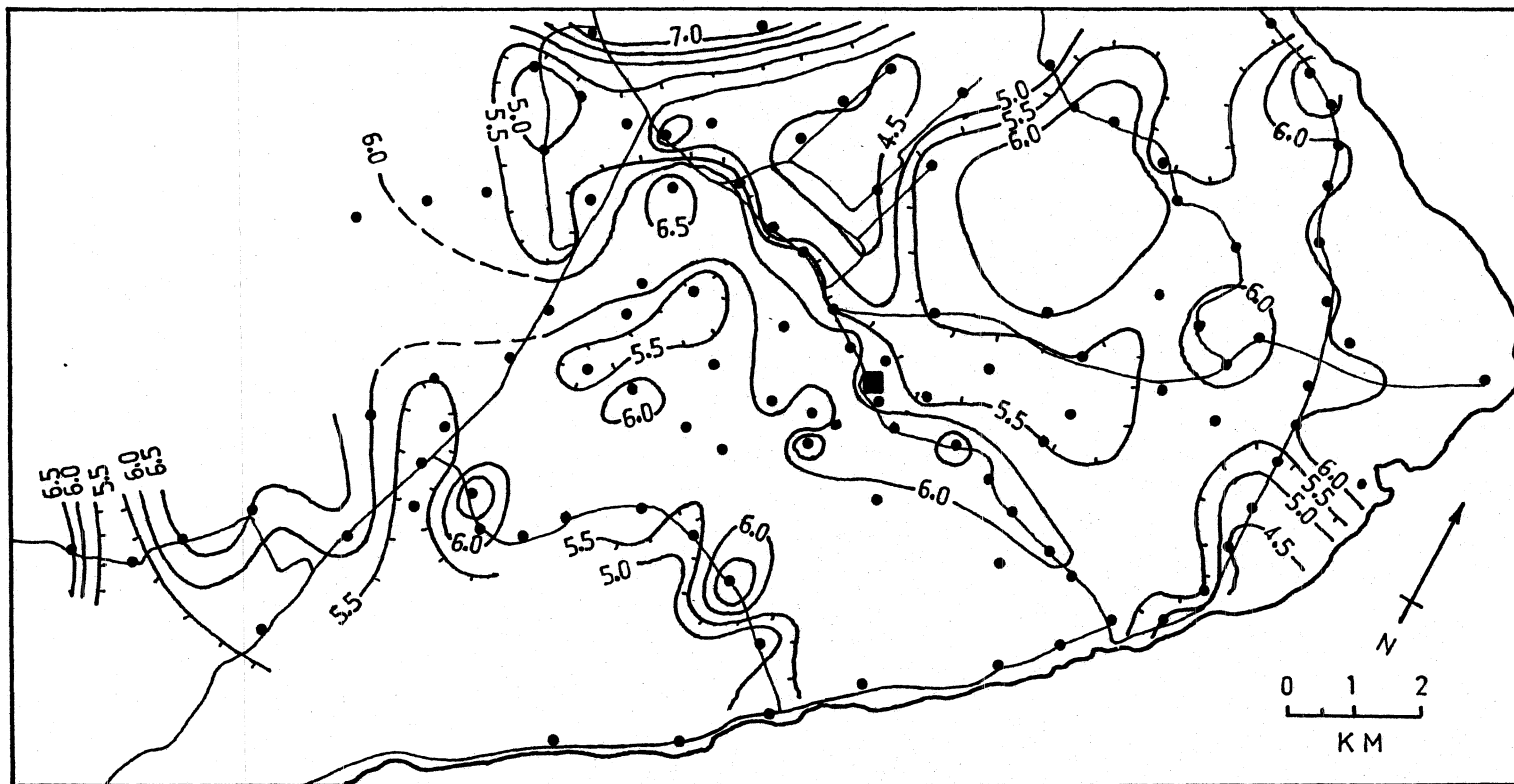


Figure 23. Soil pH map of Puna area. Contour interval is 0.5 pH units; values of 5.5 or less are considered anomalous from a geothermal aspect. Well HGP-A is denoted by a closed square; closed circles are sample locations.

values to the north which suggest the possibility of a further zone of outgassing in association with the rift structure (Fig. 22). The radon data also provides some suggestions of the latter which is in broad agreement with the concept in which the major rift structure is dipping seaward (southward). Obviously this interpretation of soil pH is tentative at best and ignores some of the non-thermal influences.

Basically, the survey results are consistent with a model of a geothermal reservoir that is elongate and rift-confined by both high fracture intensity and dike-impoundment of groundwater.

D. Haiku Radon and pH Mapping

Within this area, the topography and location of cinder cones indicate the rift zone to be approximately 5 km in width. Surveys were not carried out in the most eastern part of the area because of the rugged terrain and lack of access. Elevated radon concentrations were measured over, and 1 to 2 km west of the western boundary of the rift structure. Low positive and negative radon values occur peripheral to the central elongate high and medium positive values were measured along the coastal region (Fig. 24). The measurements indicate that the main anomaly does not continue upright to the south as the values decrease in this direction.

The background value of radon concentration in this area is higher than in the other survey areas due to the

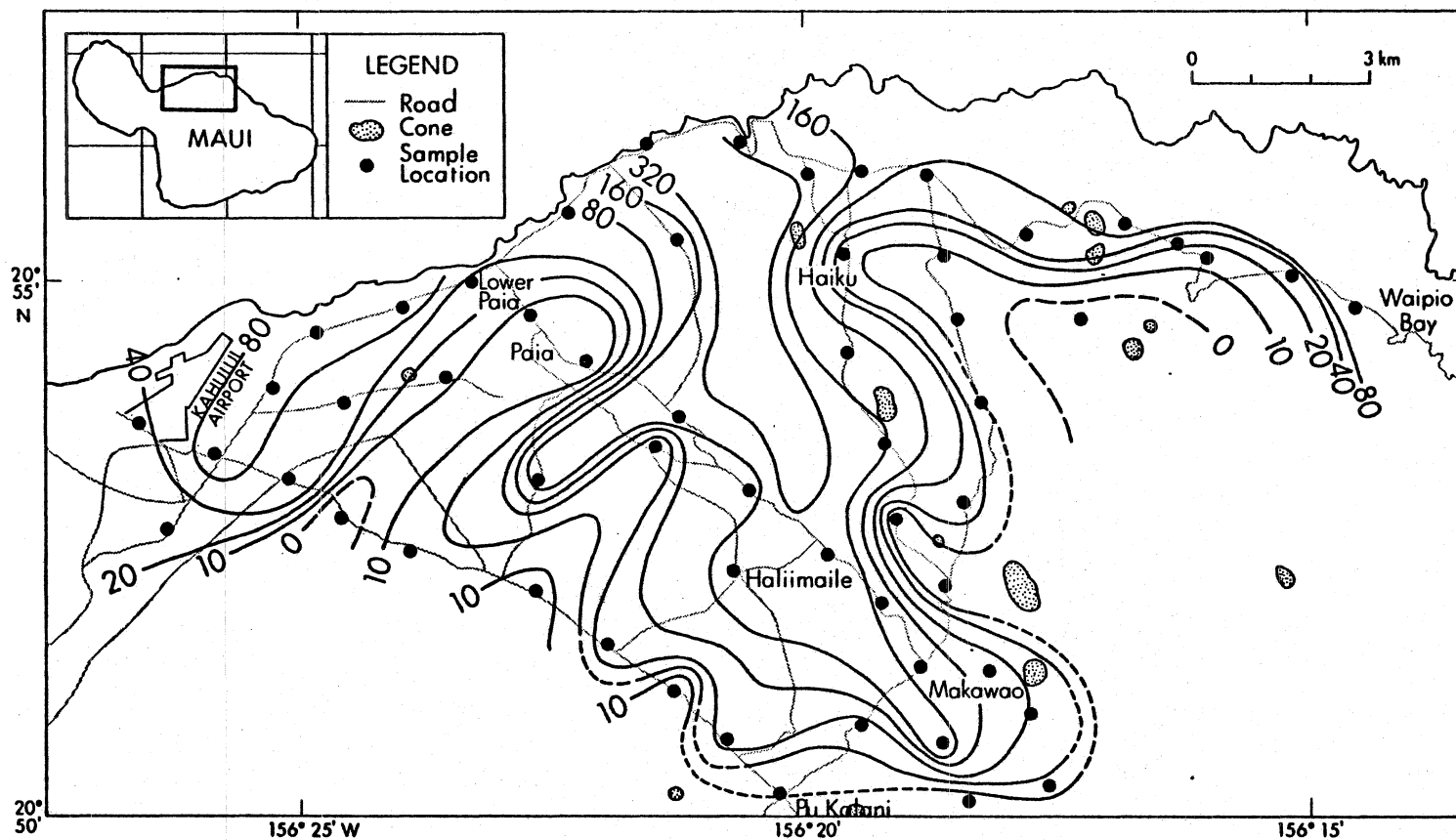


Figure 24. Contours of ground radon concentration, Haiku area, Maui. Values are contoured geometrically. The elongate high value zone falls parallel and to the west of the rift zone boundary.

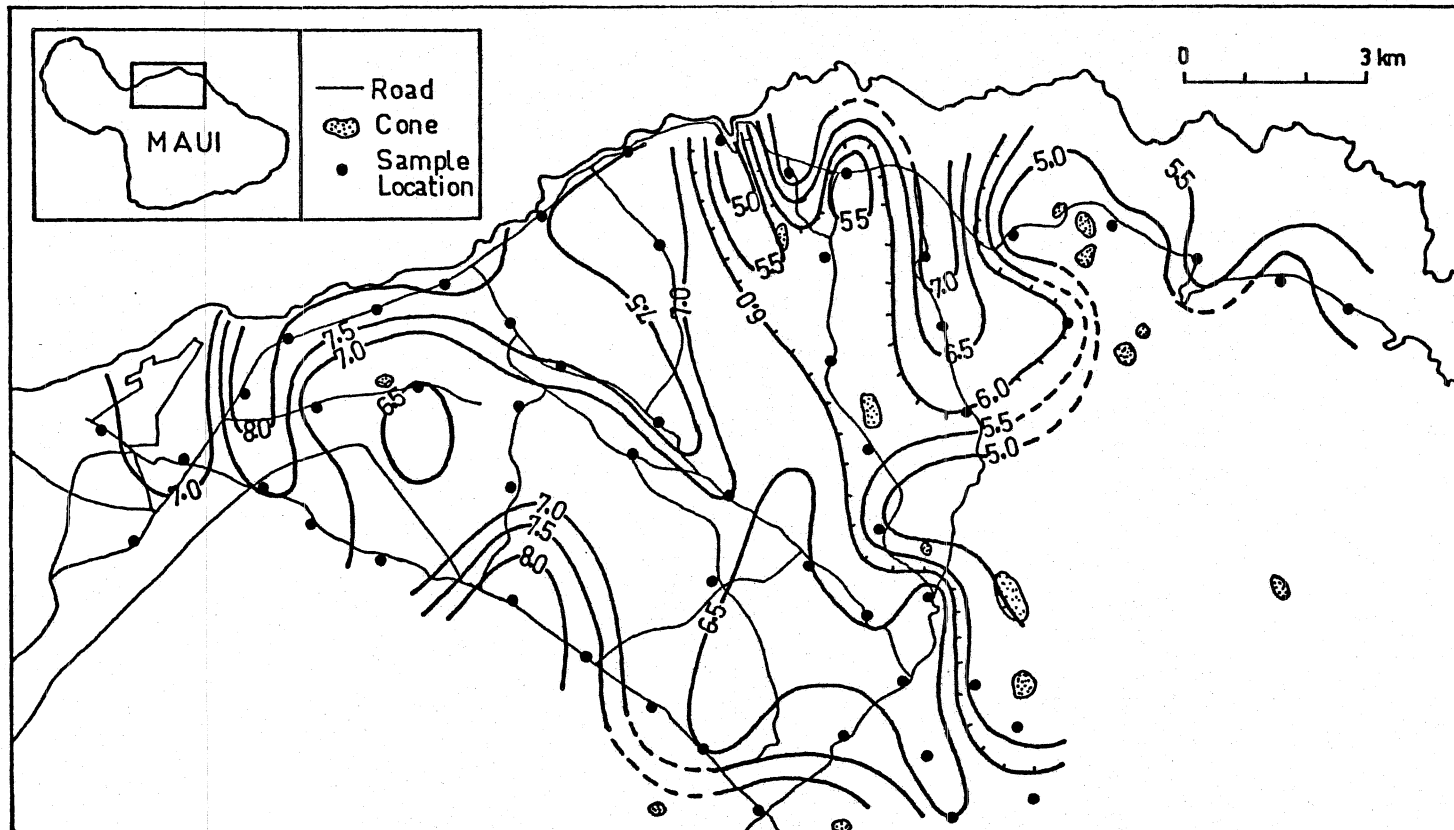


Figure 25. Soil pH map of Haiku area. Contour interval is 0.5 pH units; values of 6.0 or less are considered anomalous from a geothermal aspect. It is significant that the lowest pH (central north and northeast) correspond to the location of faults, and highly anomalous mercury concentrations [Cox and Cuff, 1980].

alkalic lavas and greater weathering; concentrations of greater than 80 R.U. are here considered significant. The limited peripheral negative zones to the east and west of the central high again suggest the existence of a ground gas convection system, which considering the distribution of concentrations appears to have developed around several major faults. The medium values along the coast and the "open end" on the central anomaly are significant and are believed to be associated with shallow thermal groundwaters. This tends to be confirmed by anomalous groundwater chemistry in that area.

The distribution of soil pH values (Fig. 25) is less variable than in the other areas due to the presence of relatively uniform soil. The values decrease from a background of 7.0 to 8.0 pH units in the west, to 4.5 to 6.0 over the rift structure. An east-west profile of the survey results (Fig. 26) shows the relative positions of the radon, pH and mercury anomalies to each other and the rift structure. The mercury high is again displaced relative to the radon high. The apparent association of these anomalies with the rift zone is significant, and is believed to indicate the existence of anomalous subsurface temperature conditions.

E. Kilauea Radon and pH Mapping

Fig. 27 presents the distribution of corrected radon concentrations on the summit of Kilauea during the period

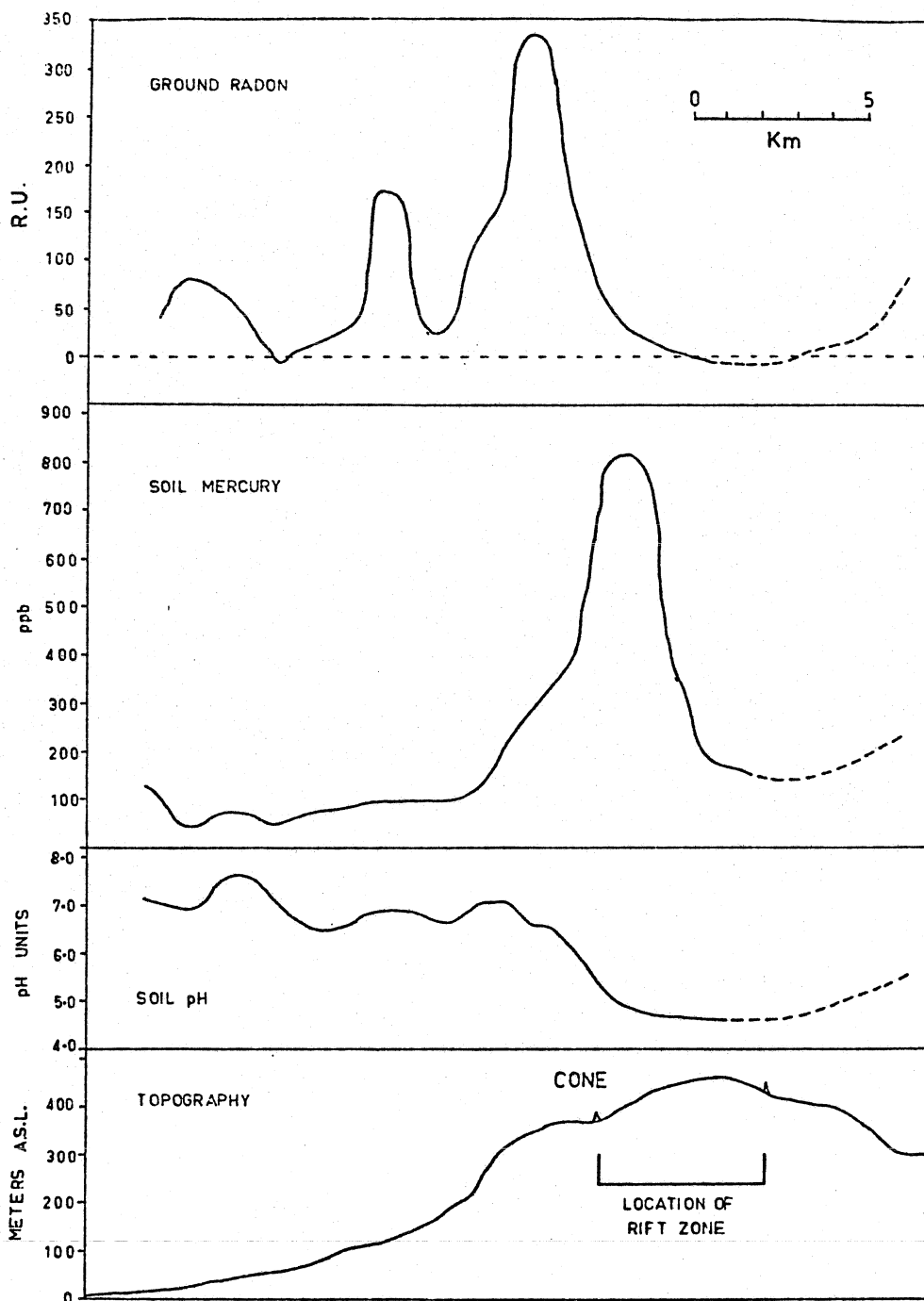


Figure 26. East-west profile of results through center of Haiku area. The elevated topography and width of the rift zone are shown. The anomalous values of radon, mercury and pH occur in association with the rift structure.

August to September, 1978. The monitoring currently being conducted indicates that the concentrations measured during that earlier period are slightly higher than a "characteristic" period of no specific events or significant inflation. This is especially so for many stations along the Chain of Craters Road in which negative values appear to be characteristic. Overall, the relative pattern of concentrations remains basically the same.

Contouring of the measured concentrations can only be done in a generalized manner due to the limitations in coverage, but a map showing the main features can be produced. The concentrations ranged from -3.24 to 155.6 R.U. and except for an area of negative values to the northeast of the caldera, in which no volcanic activity occurs, are positive. This negative zone may at other times extend to the central part of the Chain of Craters Road. Other zones of low positive values (e.g. < 2.5 R.U.) also occur, such as immediately northwest and south of the caldera, and appear to have developed adjacent to or between zones of high positive values.

The main feature of the map is the zone of high concentration (and thus outgassing) associated with the caldera. It is interesting that an extension of the zone of high concentration to the northeast of the caldera is indicated, and that the highest value measured was at the Sulphur Bank

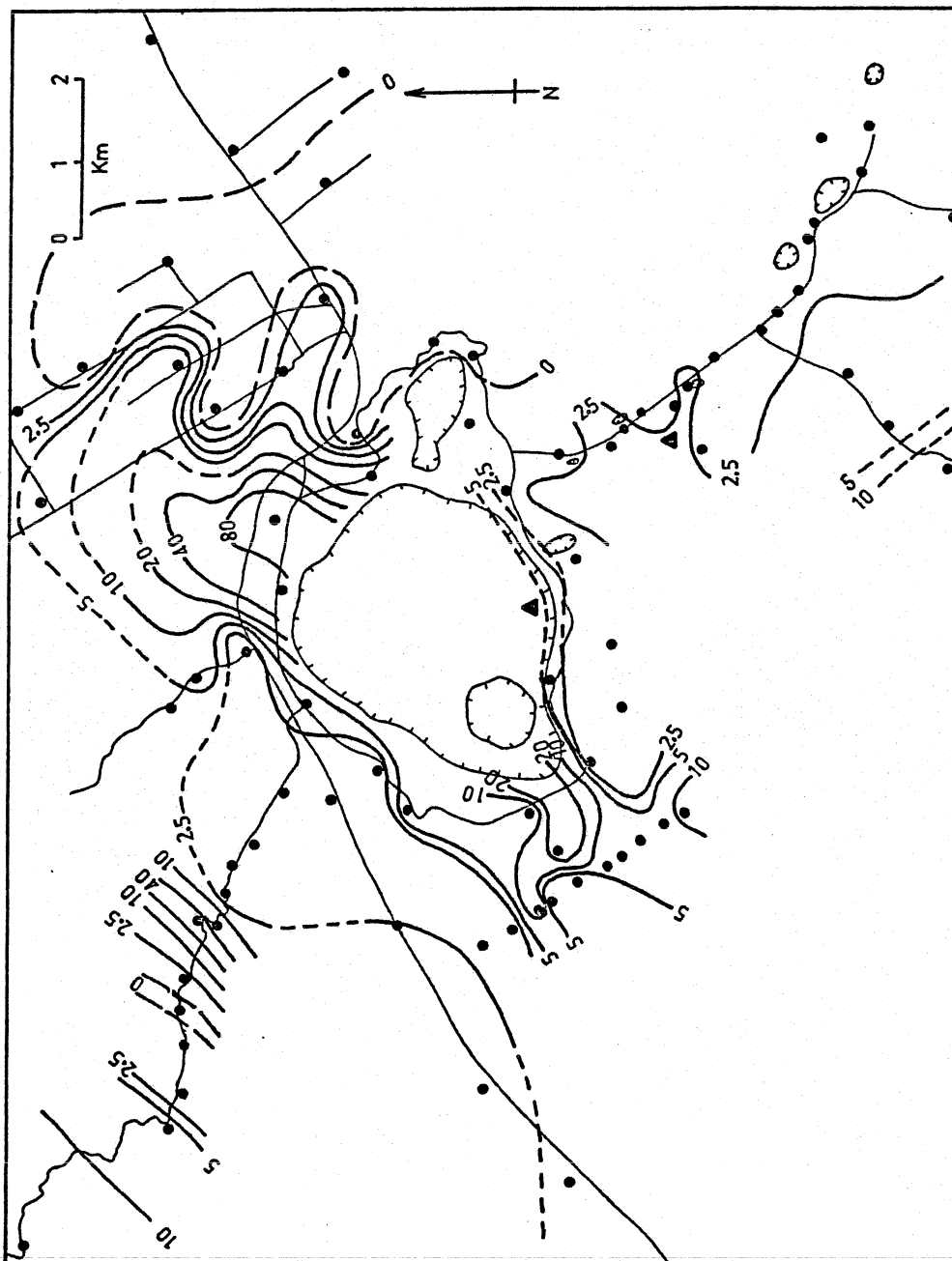


Figure 27. Contours of ground radon concentration, Kilauea summit, Hawaii. Contour intervals are 0, 2.5, 5, 10, 20, 40 and 80 R.U. The highest concentration measured was 155.6 R.U. at Sulphur Bank. Large broken line is the 0 R.U. contour; small broken lines, inferred contours. Triangles, locations where film was destroyed by high temperature; solid dots, show measurement locations.

steaming area which is associated with well-developed boundary faults. Monitoring of a station at Sulphur Bank has shown extreme variations but consistently high positive values. Several measurements were made on the floor of Kilauea caldera, and produced low positive values (0.6 and 1.6 R.U.). It is believed that these values are not representative of radon concentrations in that area, because of the complexities introduced by the existence of numerous recent flows and ponded lavas. All indications are, however, that higher values could be expected on the floors of both Kilauea and Kilauea Iki immediate to areas where outgassing is occurring and that in those environments marked and very localized variations occur. At two stations, 5 m from the fumarole (147°C) of the 1971 fissure in Kilauea Crater and at the center of the Puhimau steaming area, the films were destroyed by the high ground temperatures (75° and 95°C, respectively).

Also of interest is the isolated high value measured over one of the faults of the Kaoiki Fault system to the northwest. This value is made more significant by the existence of very low positive values peripheral to it (one over an adjacent fault) which suggest that these relative decreases in outgassing are related to the higher outgassing from the fault. This fault system is considered to be inactive volcanically but is apparently an old and well-developed structure between the volcanic edifices of Kilauea

and Mauna Loa and one along which tectonic adjustment occurs.

Areal mapping of this type in an environment such as this active summit area is complicated by the extreme variations in the type and permeability of surface materials. A further consideration is the sudden change in elevation such as between caldera floors and the surrounding area, and consequently mapping has been confined to the area surrounding the calderas. There is also the possibility of some dilution of the radon concentration in stations immediately next to the caldera rim due to the entrance of atmosphere air into the caldera wall. It is also of note that background concentrations in this area are similar to those at Puna (Figure 21, 33) which is also formed of tholeiitic basalts, and that there is an elevation difference between the two areas on the order of 900 to 1000 m. This suggests negligible effects result from elevation differences of that order.

In summary, higher radon concentrations were measured in areas in which outgassing (steaming) was observed at the surface and in areas in which fracturing or faulting was known to exist. The overall trend on the summit was of high positive values associated with the caldera structure (and other well-developed structures) surrounded by zones of low positive and negative values. (Some of the latter vary between positive and negative values depending on such conditions as seismicity and inflation during the period of measurement).

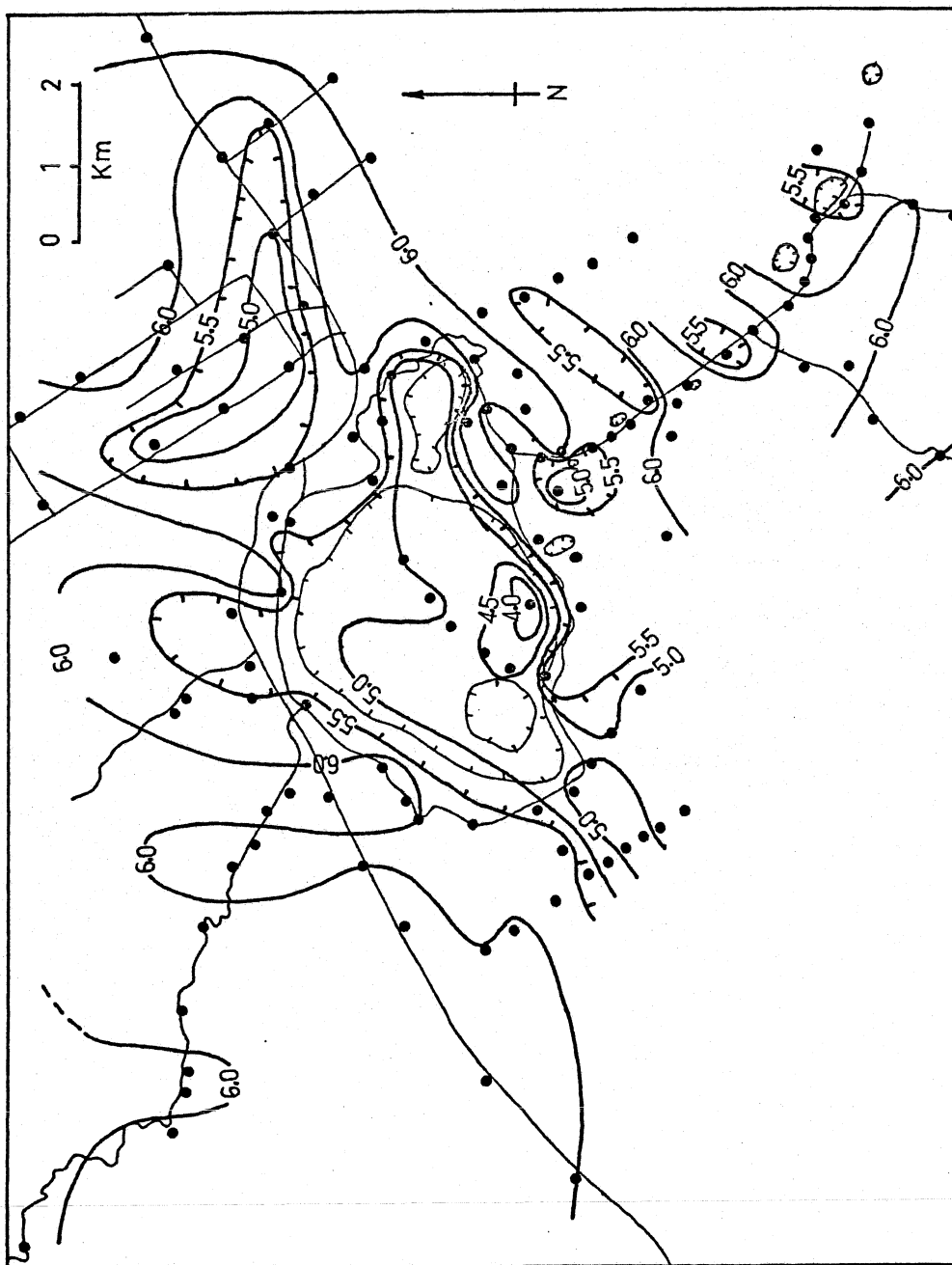


Figure 28. Soil pH map of Kilauea summit. Contour interval is 0.5 pH units. Areas with a pH of 5.5 or less are shown. Closed circles are soil sample locations.

Areal mapping of soil pH values in the area is similarly affected by the variations in the surface material, but a map of generalized trends is included (Fig. 28). This shows that the lowest pH values occur in those zones in which the greatest outgassing occurs. The most acidic value measured (pH of 3.6) was at the 1971 fissure fumarole in Kilauea Crater, where there is a significant component of H_2S in the steam. In some steaming areas the soils are much less acidic; e.g. at Sulphur Bank where although sulfur is deposited, the H_2S content of the steam is very low, and at Puhimau where the H_2S content of steam is negligible [D.M. Thomas, pers. comm., 1980]. There were several small zones of pH less than 5.5 detected along the Chain of Craters Road, which appear to occur in association with specific fractures.

The area of pH of less than 5.5 in the vicinity of Volcano town to the northeast of the crater partly coincides with a zone of positive radon concentrations. Anomalous values in this area are not fully understood and due to the well-developed soil cover and dense vegetation identification of any faulting is not possible. The partial correlation of anomalies and the fact that they do not occur in adjacent areas of similar soil-vegetation characteristics suggests that fracture-related outgassing may be possible in that area. Further, considering the location, it is not unlikely that faulting has developed peripheral to the caldera.

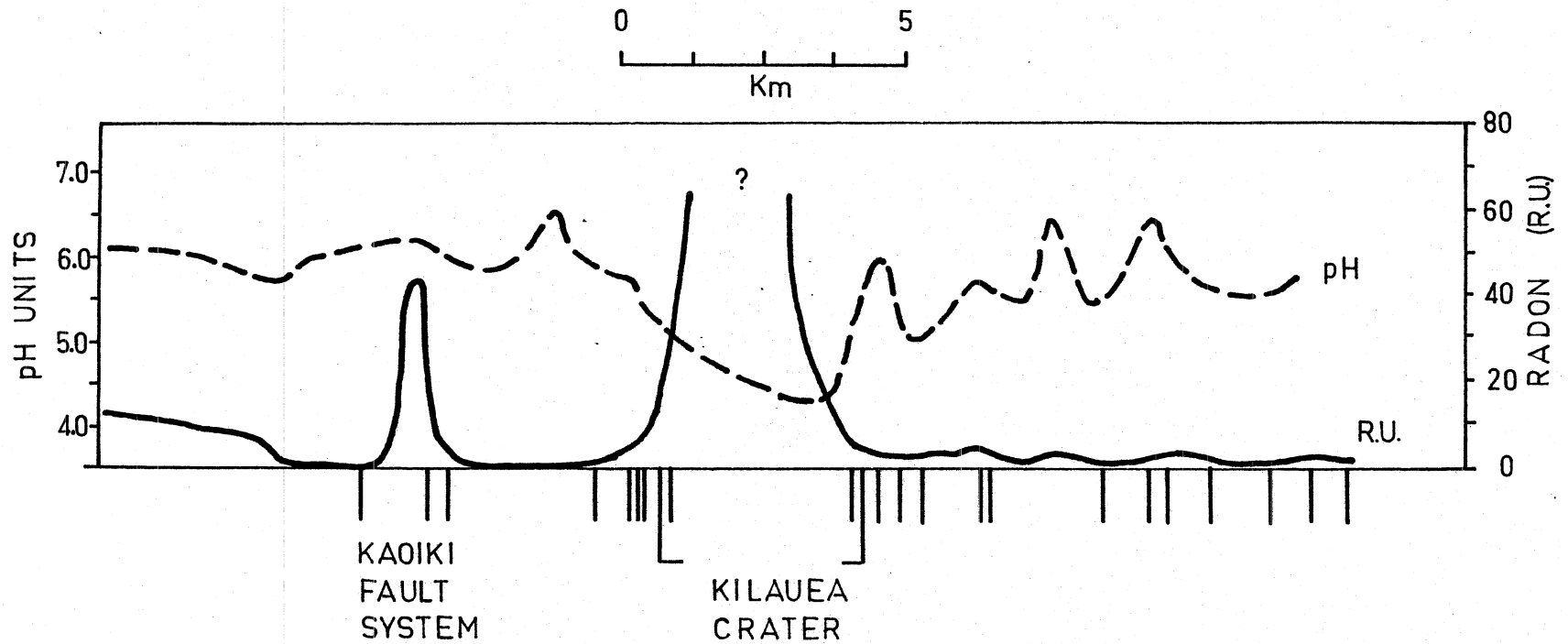


Figure 29. NW-SE profile of results across Kilauea summit, showing radon and pH values. Lines at base represent locations of faults and main cracks. Radon concentrations within Kilauea crater have not been measured in detail.

A northwest-southeast profile across the summit (Fig. 29) shows the broad variations in radon and pH values. The basic trend is of an inverse relation, low pH accompanying high radon concentrations, and the radon highs commonly associated with observed fractures. Localized variability of values is evident and typifies the complexity of carrying out such surveys in an area of this type.

F. Waianae Radon and pH Mapping

The measured radon concentrations (Fig. 30) show consistent patterns within the area, which (in comparison to Kilauea) demonstrates not only the relatively uniform surface material, but also the lack of extreme spatial differences in temperature.

The three main features in the pattern of radon concentrations are evident:

(i) a negative zone central to Lualualei Valley and roughly central to the inferred ancient caldera, and a smaller negative zone to the south of the inferred boundary of the caldera;

(ii) an approximately circular zone of high positive values peripheral to the larger negative zone; and

(iii) a trend to extremely high values with high elevation toward the Schofield Plateau in the northeast of the area.

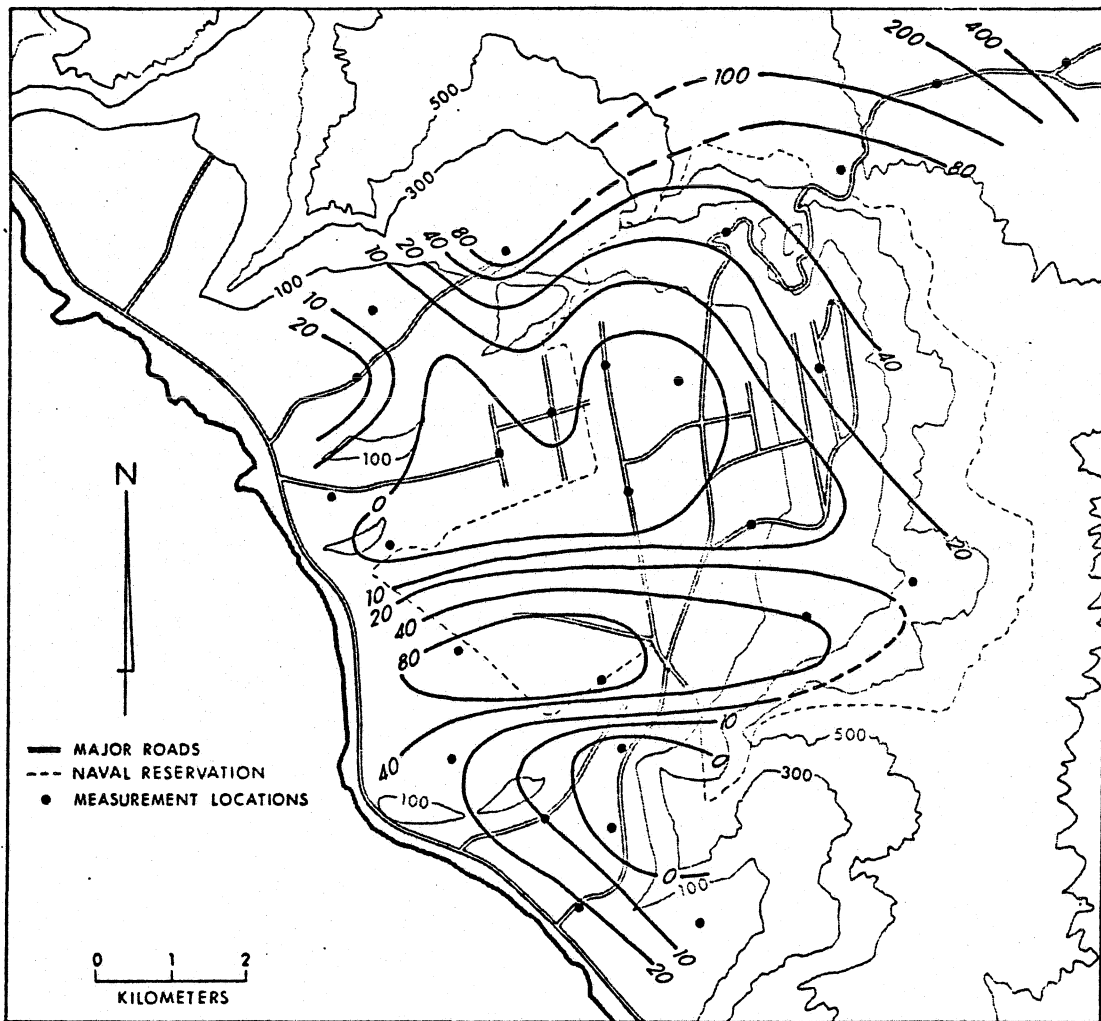


Figure 30. Contours of ground radon concentration, Waianae area, Oahu. Contoured at intervals of 0, 10, 20, 40, 80, 100, 200 and 400 R.U. Higher concentrations surrounding a main central negative value area are evident.

The interpretation given these data is that a greater upward movement of ground gas is occurring within the peripheral areas which represent the structural boundary of the ancient caldera at its present erosion level and a zone of relatively greater fracturing. The low positive and negative values central to Lualualei Valley may be due in part to a greater thickness of sedimentary material and a lower degree of fracture permeability of the underlying lavas, but the overall pattern of radon values suggests a relationship to the peripheral high value zones. The same conditions are also considered to apply to the smaller negative zone to the south, also considered part of a ground gas convection system. Although it is not known whether intrusive activity younger than the age of the last flows has occurred within the caldera complex, there are indications of the existence of at least a low order thermal anomaly below the area: several groundwater wells have above ambient temperatures and anomalous chemistry; electrical resistivity surveys indicate the presence of warm saline water saturated rock; and anomalous concentrations of soil mercury coincide with radon anomalies [Cox et al, 1979].

The cause of the large increase in radon concentrations to the northeast has not been determined. Although the radon emanation from soil and rock in that area is higher than within the valley, due to the existence of more alkalic soils, the values measured in the field are

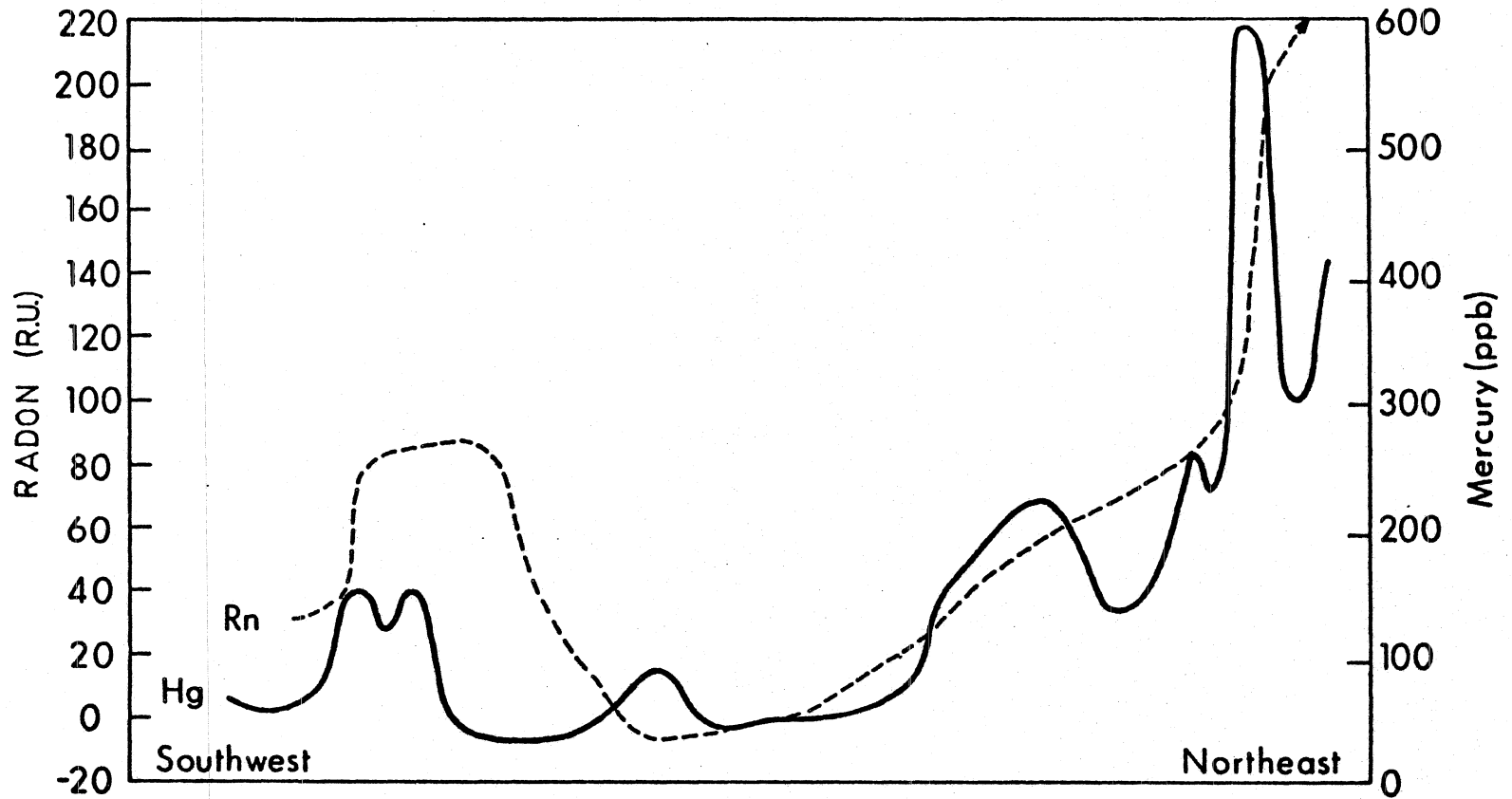


Figure 31. SW-NE profile of results across Waianae area, showing radon and mercury concentrations.

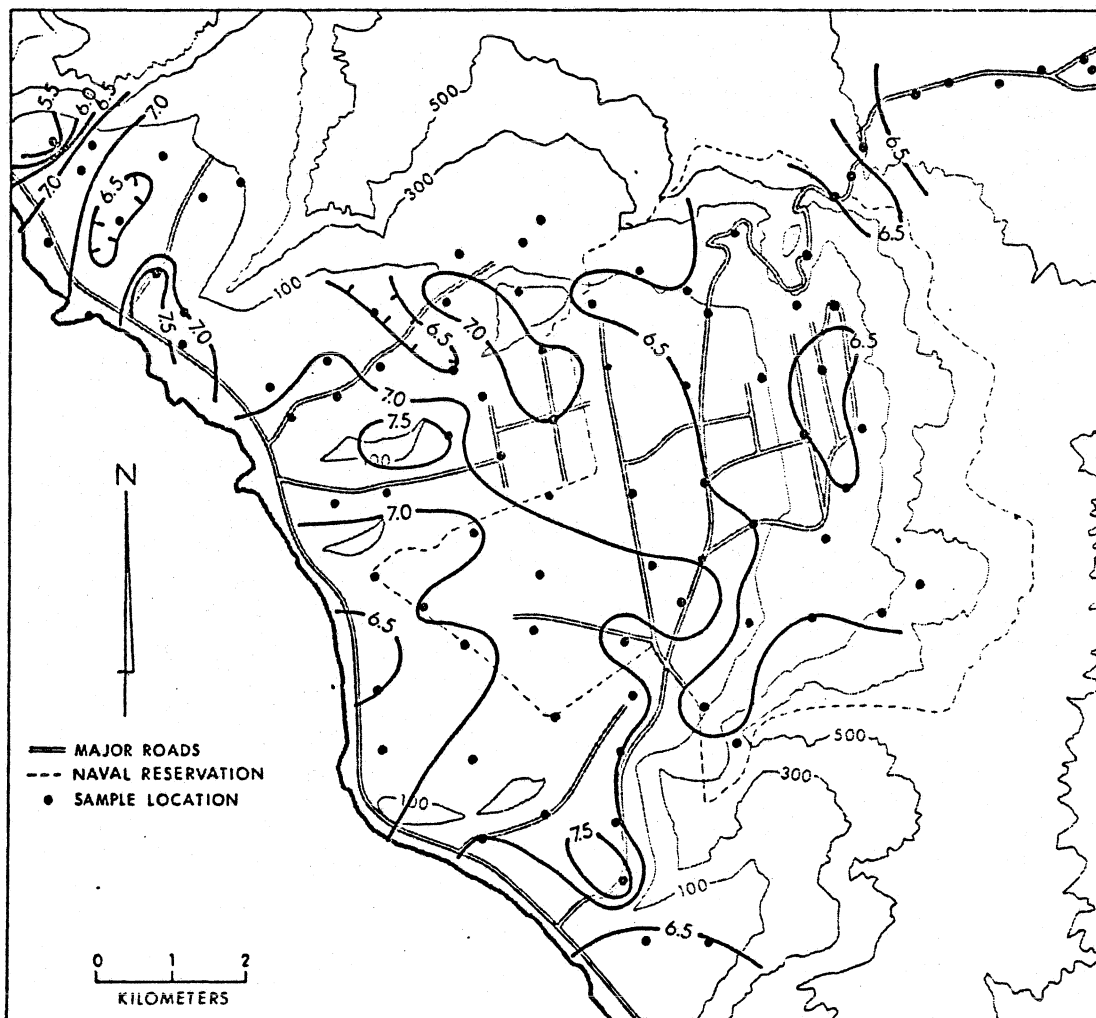


Figure 32. Soil pH map of Waianae area. Contour interval is 0.5 pH units.

excessively high. The area is within a Military training area, but no artificial source for the elevated concentrations is reported to be present. The gradual increase is noteworthy, as is the location within the inferred northeast rift zone and the coincident increase in soil mercury concentrations (Fig. 31). The values consequently are considered to be real, and a natural phenomenon. The only feasible explanations that can be forwarded at this time are a possible relation to late stage intrusive material, the existence of very high fracture permeability or some influence related to the rapid increase in elevation.

The soil pH in the area is largely neutral to slightly acid (Fig. 32) and only general trends are discernable. The main trends are a decrease in pH (< 6.5) towards the east and the inferred caldera boundary and an arcuate zone of higher pH (> 7.0) in the western part of the area. The low pH over the Waianae Range could possibly be related to the different climatic environment at that higher elevation. The pH data are not considered to show anomalous patterns and the variations that do exist are probably derived from surface phenomena, possibly changes in soil chemistry.

G. Frequency Distribution of Radon and pH

The percentile frequency distributions of the corrected radon concentrations for the four areas studied are presented in Fig. 33. These data reflect the characteristic

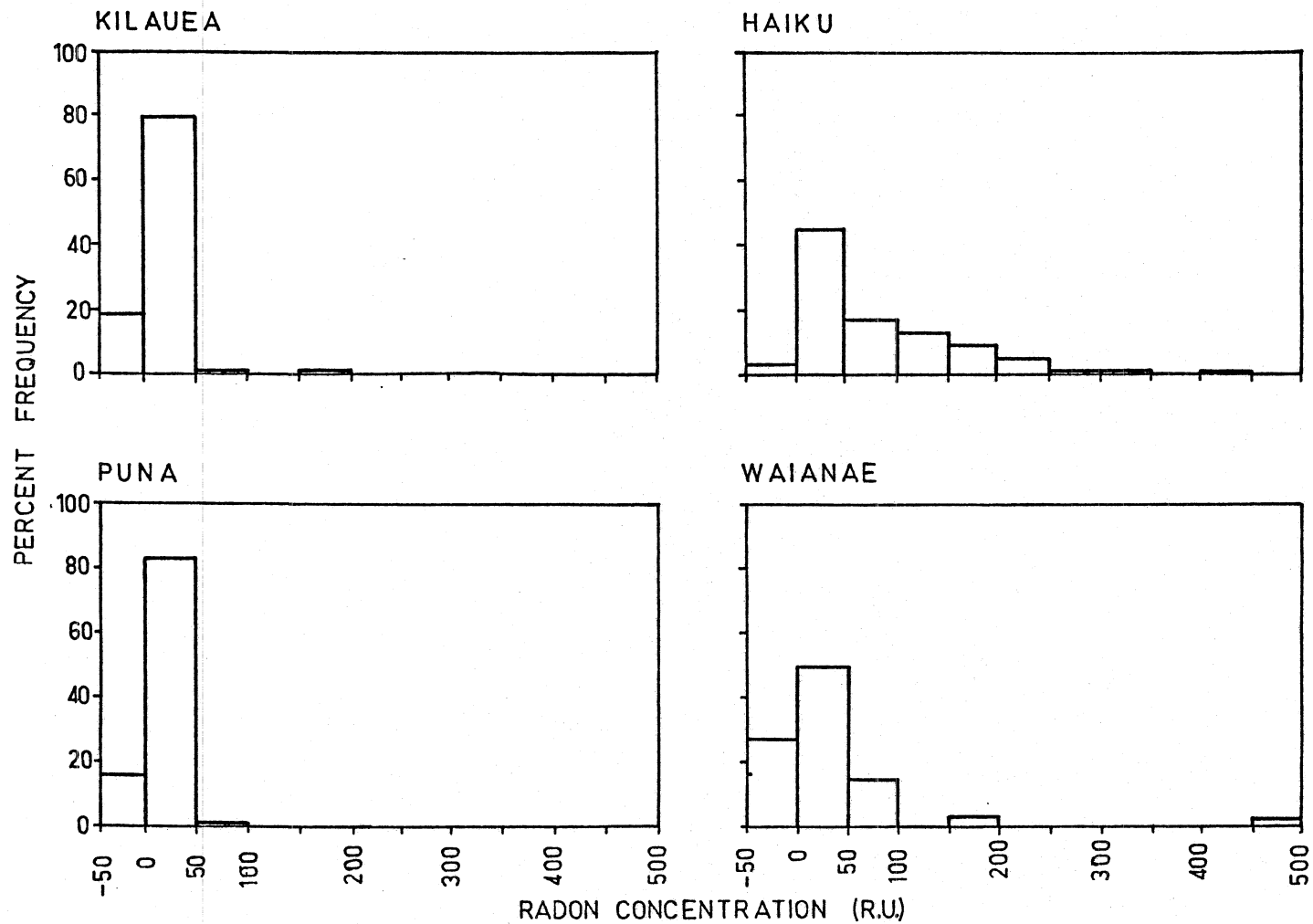


Figure 33. Percent frequency distribution plots of ground radon concentration for the four survey areas. Divisions are 50 R.U.

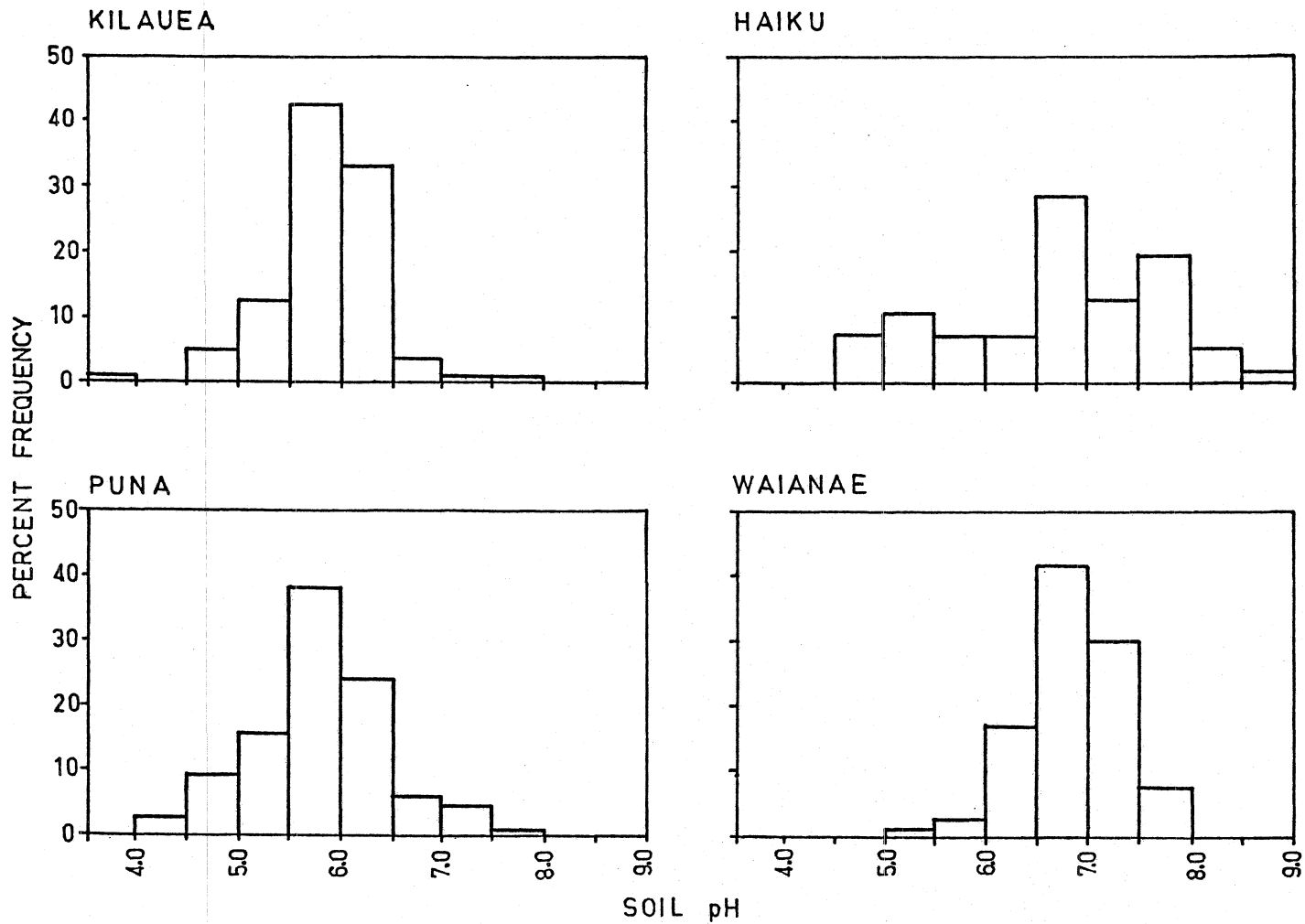


Figure 34. Percent frequency distribution plots of soil pH values for the four survey areas. Divisions are 0.5 pH units. Vertical scale is to 50% only.

differences in the concentrations of the parent radionuclides, uranium and thorium, in the rocks occurring in the areas. The volcanically active areas of tholeiitic basalts, Kilauea and Puna, have the lowest radon concentrations, with approximately 80% of the measurements falling between 0 and 50 R.U. Conversely, the similarity of the values in the Haiku and Waianae areas partly reflect the occurrence of alkalic basalts, with only 45 to 50% of the measurements in the 0 to 50 R.U. range. The differences in the distribution of concentration between each of the above sets of areas is also indicated to be related to their ages and the development of weathering and soil profiles. These are more advanced in Haiku and Waianae than in the two still volcanically active areas.

The frequency distributions of pH show the values characteristic of each area. The greatest occurrence of the low pH, more acidic conditions (pH of < 6.0), is within the active areas of Puna and Kilauea, both of which show approximately symmetrical distributions (Fig. 34). The soil pH in the dry leeward Waianae area is overall near neutral with few values below 6.0. At Haiku the pH has a wide range of values and although the larger proportion of measurements are near neutral, a significant proportion of acid values also occur producing a multi-modal distribution.

VII. GENERAL DISCUSSION

A. Association of Elevated Radon Concentration and Heat

The above information indicates that within much of Hawaii the major portion of the radon measured is likely to be derived from material within tens of meters of the detector, and that under normal conditions a uniform distribution of radium within the lavas can be assumed. However, in zones where permeability and convection are greater an increased contribution of radon originating at greater depth (up to several 100 meters) can be expected, and it is highly likely that enrichment of the parent nuclides also occurs.

Simplest case in which there is vertical migration of radon from an infinite lateral source by diffusion alone has been expressed [Tanner, 1964,a] by:

$$C_m = C_o \exp \left(-\sqrt{\frac{\lambda}{D}} Z \right)$$

where C_m = measured concentration

C_o = original concentration

Z = migration distance

λ = decay constant of radon

D = diffusion coefficient of the material

Using data from the Puna area an upper distance for this

type of diffusive transport (for Rn-222) of 215 cm in dry porous material and 2.15 cm in saturated porous material can be calculated by:

$$C_m = 5 \text{ R.U. (assumed as lower limit of anomalous concentrations)}$$

$$C_o = 1.24 \text{ R.U. (for tholeiitic lavas)}$$

$$\lambda = 2.097 \cdot 10^{-6} / \text{sec (Rn-222)}$$

$$D = 5 \cdot 10^{-2} \text{ cm}^2 / \text{sec (porous material)}$$

$$D = 5 \cdot 10^{-6} / \text{sec (saturated porous material)}.$$

The addition of a transport component to a one-dimensional solution shows that for a source of uniform concentration but at increasing depths an increase in velocity of flow or diffusion coefficient is required to maintain the measured radon at a certain concentration. This solution may be expressed [Grammakov, 1936] as:

$$C_m = C_o \cdot \exp \left[-Z \left(\frac{V}{2D} - \sqrt{\frac{V^2}{4D^2} + \frac{\lambda}{D}} \right) \right]$$

where, the factor V = velocity of flow, is added.

Using a vertical migration distance of 150 m and the values from the diffusion case, a mean velocity of flow of 0.23 cm/sec can be approximated within the bounds of the Puna anomaly. To obtain the highly anomalous concentrations

measured in that area (e.g. 40 R.U.) and assuming the same source depth an increase in velocity or diffusion coefficient by a factor of approximately 8 is required. However, a higher measured concentration (C_m) can also be obtained by increasing the concentration of the source (C_0). This introduces the possibility that enrichment of parent radionuclides within the reservoir or at shallower depth can also increase the radon concentration measured near the surface. If the higher measured concentration of 40 R.U. is used, and it is assumed that the greatest contribution of radon is derived from the unsaturated lavas above the water table (i.e. depths of 75 to 250 m) the solution shows that this measured concentration can be produced with the lower velocities of $11.3 \cdot 10^{-3}$ to $37.6 \cdot 10^{-3}$ cm/sec. It is interesting to note that on the summit of the extinct volcano Mauna Kea (4200 m) upward flow rates of ground gas have been measured (in 12 m deep holes) over a range of $3 \cdot 10^{-3}$ to $25.3 \cdot 10^{-3}$ cm/sec with an average of $8.1 \cdot 10^{-3}$ cm/sec [Woodcock and Friedman, 1979]. This migration of ground gas is believed by those authors to be in response to pressure changes in the atmosphere and a source of heat within the mountain.

It can thus be seen that the concentration measured (C_m) can be increased by increases in any or all of, diffusion coefficient, velocity of flow, source concentration or a decrease in the depth to the source. These changes can be

produced by increases in permeability and heat. It should also be noted that overall, in the areas surveyed there was no correlation between radon concentration and depth to the water table. In instances where there was some indication of this, further investigation showed it to be fortuitous and more related to the location of groundwater wells from which the data were obtained.

To further confirm that increased permeability and temperatures produce the elevated concentrations of radon as measured in geothermal areas in Hawaii, the following points can be noted:

(i) ground permeability is increased by faulting and associated fracturing, which are known to be well-developed within rift zones and in association with caldera structures. The existence of these features can be observed in the field and on aerial photographs (e.g. Macdonald and Abbott, 1977; Cox et al, 1979];

(ii) an increase in the velocity of ground gas migration will be caused by an increase in permeability but also can be induced by anomalous thermal conditions (e.g. Nielson, 1978). The relation between higher radon concentrations in ground gas and higher subsurface heat, although assumed on the basis of theoretical and empirical concepts, is further confirmed by the following direct and indirect evidence;

(a) increases in radon concentration in association with heat producing subsurface events (e.g. Figure 19;

Appendix 6; Birchard and Libby, 1980; Cox et al, in press);

(b) higher radon values in areas where steam is discharging at the surface, than in surrounding areas where it is not (e.g. Figure 20, 21 and 27);

(c) common association of higher radon concentration and low soil pH. This especially evident in volcanically active areas and indicates an elevated amount of H_2S or SO_2 in ground gas implying above ambient thermal conditions (D.M. Thomas. pers. comm., 1980);

(d) the common existence of elevated mercury concentrations in soil, often peripheral to radon highs. Elevated concentrations of mercury are known to be associated with anomalous temperatures due to the temperature-related volatility of mercury (e.g. Matlick and Buseck, 1975; Klusman and Landress, 1979);

(e) in areas of significant radon anomalies, data gathered from other techniques are in most cases anomalous: these include groundwater temperatures and chemistry, electrical resistivity and self-potential measurements (e.g. Keller et al, 1977; Cox and Thomas, 1979,b);

(iii) the available evidence indicates that uranium is concentrated in secondary mineralization in hydrothermal systems (Wollenberg, 1975; Lalou and Brichet, 1980). The occurrence of a very low Th/U ratio in $CaSO_4$ mineralization which has been deposited in HGP-A during the 3.5 years since it was drilled and uranium concentrations higher than in the

tholeiite reservoir rocks suggests relatively rapid deposition of a uranium bearing mineral phase in this geothermal reservoir (Table 8). It is therefore possible that such deposition may occur in other geothermal systems in Hawaii. It is also significant that both surface sublimate deposits and several soil samples from areas of high radon concentration (most of which were steaming) had low Th/U ratios. This may suggest that some increase in uranium concentration in material above, as well as within, geothermal systems may occur from the circulating thermal fluids. In areas where this relative enrichment of uranium occurs, it is also reasonable to assume that radium would also be enriched. An important consideration is that the anomalous radon concentrations measured are significantly greater than would be produced solely by the indicated enrichment of radium and uranium.

The above data are believed to support the postulate that elevated radon concentrations are produced by a combination of several parameters that control the availability and migration of radon. Further support for this is the Puna survey, which was conducted retrospectively, after the well HGP-A had confirmed the existence of a high temperature reservoir. The results of that survey correspond well to a model based on realistic geological and hydrological constraints.

B. Radon Mapping as a Geothermal Exploration Technique

In terms of suitability of the described technique for geothermal exploration in Hawaii, the examples presented in this study are believed to demonstrate a high degree of success. This is especially so considering the diversity of environments of the survey areas. In this respect, it is emphasized that what is sought with this type of radon mapping are relative high concentration zones within an area of the same lava type. Consequently, absolute concentrations measured cannot be related to surveys conducted in other areas of different characteristics. Distinct anomalies were determined, which are in agreement with the majority of results from other techniques utilized, notably geochemical. Although it is obvious that surface exploration of this type requires drilling to confirm the apparent success (and the real potential for economic development) the determination of anomalous values enclosing the HGP-A location is highly significant.

The surveys further demonstrated that the technique can be applied in areas in which appreciable cultural activity precludes the use of other methods, both geochemical and geophysical. It was also found that the films were not affected by ground temperatures of up to 15°C above the ambient temperature ($\approx 24^\circ\text{C}$) and Fleischer and Likes [1979] consider that for alpha-tracks to be affected the films need to be exposed to temperatures above 60°C for at least four

weeks. The technique proved to be highly mobile, enabling surveys in areas of limited access, and could be carried out by one or two people. The technique is rapid, each station could be set up and a field description recorded within fifteen minutes. Thus, depending on conditions of transport, 15 to 20 stations could comfortably be established in a day. The obvious limitations here, are that ideally one area should be completed within a maximum period of 4 to 6 weeks because of temporal variations. This, however, is not required in areas which are not volcanically active. Another aspect is that the detectors need to be retrieved, and that approximately 5 percent are disturbed preventing the use of that measurement. Considering the above, a competent operator can establish approximately 350 stations in a month. Preparation and development of films is also easy and with the method described at least 44 films can be developed in a day. Counting is also easy, but requires some practice and tends to become tedious. Typically, a film can be counted in 15 minutes.

As an example of cost, the survey expenses of a typical area of 150 km^2 with reasonable vehicular access are given. A suitable sampling density for exploration purposes would be one station per 1.25 km^2 i.e, 120 stations:

detector cost at 0.50c each	=	60.00
survey labor at \$4.50 per hour	=	432.00
labor cost, developing and counting at \$4.50 per hour	=	315.00
labor cost, data reduction to tabulated form	=	120.00
cost chemicals, minor items	=	<u>20.00</u>
		\$947.00

The above estimate does not include accomodation, travel to or from the survey area or the cost of equipment (microscope, temperature bath, standard radioactive source etc). Based on these figures the approximate cost per measurement is \$8.00 or roughly \$6.30 per km² with the station interval used. Although two field personnel are preferable in some rugged areas and the cost is consequently increased, overall, the method was found to be cost-effective and highly suited to the conditions encountered in Hawaii.

Several other considerations are noted in this application of radon surveying. Of importance is the relationship between ground permeability and subsurface heat in producing anomalous measurements near the surface, and the possibility that the surface anomaly may be offset in respect to the source of heat at depth. Such would be the case with an inclined fault within rocks of low permeability, which is continuous with a geothermal reservoir (e.g. Nielson, 1978). In Hawaii, however, the structures that provide permeability

increases relative to surrounding areas are in most cases vertical or near vertical. A further consideration is the width of the zone of high permeability (such as a specific crack or fault, as opposed to a zone of shattering), its permeability relative to surrounding areas and its possible continuity to the source of heat (either hot rock, or hot groundwaters). Thus in Hawaii the spacing and distribution of measurement stations may need to be considered in cases where the normally high permeability of the lavas is increased in such limited zones. Conversely, it is also important to relate the known structure of an area to the radon concentration patterns mapped.

The results of these surveys indicate that in most cases higher concentrations of radon are found in association with faults. However, in some cases, more obviously in the volcanically active areas, low positive and negative values can occur over faults adjacent to one over which high values are occurring (e.g. the Kaoiki fault system on Kilauea, Figure 27). In these cases it is possible that greater outgassing through a more permeable and perhaps deeper fault has created a localized convection cell in which downflow has developed within the adjacent faults.

In regard to the four areas presented in this study, it should be noted that all were selected on the basis of possessing both known or indicated temperature anomalies

and structural features of volcanic origin. Consequently, they constitute a biased data set in terms of producing anomalous results.

The surveys further show that radon mapping, as with all exploration techniques, should be utilized in conjunction with other methods and the results evaluated on the basis of all possible available data, especially geological.

VIII. CONCLUSIONS

The objectives of this study are believed to have been attained by demonstrating the successful application of ground radon surveys to geothermal exploration in Hawaii. The study also produced further data on natural occurring radioactivity in Hawaii in general and more specifically within hydrothermal systems.

The gamma spectrometry data shows that in soils in Hawaii the concentration level of uranium and thorium is similar to that of the lava from which it was derived. The soils commonly, however, have a Th/U ratio slightly lower than the lavas, and assuming (as shown by other studies) that the thorium is less affected by surface weathering processes, indicates an enrichment of uranium. It is possible that with weathering at the surface some uranium after being leached from the lavas is deposited from solution becoming absorbed within the soils or contained clay minerals. Significant decreases in the Th/U ratios are evident in hydrothermal deposits and in some hydrothermally altered soils, suggesting that a greater enrichment of uranium can occur within and above hydrothermal (and geothermal) systems. These hydrothermal deposits also experienced an increase of uranium concentrations by a factor of about 2 of the concentrations in the parent lavas. Although the measured concentration is still low the enrichment that does occur in these secondary

hydrothermal minerals would further enhance the use of radon surveys in investigations for geothermal systems.

The release of radon from rocks or soils is apparently controlled by a combination of factors. Although some disagreement exists in the literature on the controls, this study suggests that a major factor is the concentration of the parent radionuclides within the material. From the limited work carried out it could not be determined if grain size has a major influence on radon emanation from soil, but the degree of weathering is a factor. This could partly be construed as a grain size effect but also constitutes chemical and other physical changes. A greater emanation from some examined soils may also be a function of the enrichment of uranium. The amount of radon emanation from soil may be partly related to the amount of organic material within the soil, but this is probably indirectly, or partly, a function of soil chemistry and grain size.

The ratio of Rn-222/Rn-220 emanating from soil is probably similar to the ratio of their parent radionuclides, uranium and thorium, within the soil or the parent lavas. Where radon migrates through greater thicknesses of soil or from increasing depth there is probably a significant increase of Rn-222 over Rn-220 under the effects of convective transport and the longer half-life of Rn-222. The ratio of the isotopes of radon within the ground gas is,

however, not of prime importance in this application, as relative variations of total radon within an area are being sought.

The surveys as applied can define anomalies in the form of zones of positive values which are high relative to surrounding areas. Typically, concentrations of radon that are considered to be anomalous are on the order of a factor of 4 greater than the mean radon emanation from the soil cover in an area. Of further importance is the indication from these surveys that thermally-induced ground gas convection systems have developed within and around some permeable geological structures. The data suggest that in areas where no such structures occur measured radon values are variable and commonly less than a factor of 2 of the emanation from the surface material in an area. The data presented also confirm the relationship between high radon concentrations and subsurface heat, by both theory and observation. In all four field examples the results obtained could be explained by reasonable geological models. The one possible exception was the extremely high values measured to the northeast of Lualualei Valley, Oahu.

The study suggests that in much of Hawaii the radon measured near the surface is derived from relatively shallow depths, in the order of 10's of meters. A further indication is that where there are increases in both permeability

and heat, there is also an increase in the maximum potential depth from which the radon can be transported. Under such conditions the relatively high velocity ground gas may tend to cumulatively pick-up radon being emanated from the material through which it passes, increasing the concentration measured near the surface.

Although errors of measurement exist in the technique used, they are minimized by the adoption of a constant procedure and the total error of the final corrected concentration is considered to be less than 10%. The application of the technique as described, using measurement spacings of 0.5 to 1.0 km, is considered to be optimum for exploration purposes where patterns within an area are sought, and cost and time are important. More precise studies with greater control can readily be implemented for other purposes if required. The use of field measurements made over periods of about 4 weeks appears to successfully "integrate" short-term variations in radon concentration, however, in volcanically and seismically active areas longer term changes of greater than a factor of 10 were measured. This indicates that subsequent measurements in such an area should be used with caution, even when "tied" into previous stations. This effect is minor in non-active areas and in those areas radon concentration patterns are readily reproducible. Work subsequent to these surveys, however, shows that remapping of

active areas will produce the same relative patterns.

As an exploration method in Hawaii such radon mapping was found to be both cost-effective and rapid. It can be conducted in a variety of environments, including areas in which cultural activities may limit the use of other techniques. Ideally, in Hawaii, concurrent surveys measuring mercury concentrations and pH of soils should be conducted as they not only provide useful information but augment the radon survey results and interpretation.

The following notice from the manufacturer of the alpha-sensitive films used in this study should be noted:

NOTICE

"Kodak does not recommend use of its products in any manner that would infringe valid patents of others. Customers have the responsibility of determining for themselves whether any particular use they might make of these products would require a license under patents of others. In this connection, they may wish to consider one or both of the following patents of Terradex Corporation:

— U.S. patent 3,303,085 dated February 7, 1967,

— U.S. Patent 3,665,194 dated May 23, 1972.

and any corresponding patents currently in force in other countries.

Copies of the U.S. patents can be obtained from the U.S. Patent and Trademark Office, Washington, D.C. 20231.

For a complete list of the Terradex patents in other countries, or information on specific countries, contact Terradex Corporation, 1900 Olympic Boulevard, Walnut Creek, California 94596."

APPENDICES

APPENDIX 1
 Duplicate Measurements of Radon Emanation
 From Soil Samples

Sample No.	T/cm ²	Minus Film Background	Exposure (hours)	R.U. (T · 10 ⁻² /cm ² /hr)	Mean Percent Deviation
W9/1	248	245	763.39	32.09	5.8
W9/2	147	144	476.22	30.24	..
I1/1	758	755	786.27	96.02	13.8
I1/2	533	530	475.92	111.36	..
L39/1	55	52	506.54	10.27	1.1
L39/2	112	109	1073.00	10.16	..
L60/1	97	94	506.54	18.56	1.4
L60/2	205	202	1073.00	18.83	..
P123/1	11	8	477.92	1.67	12.6
P123/2	10	7	478.33	1.46	..

Average deviation = 6.9

All samples 30 cc volume.

APPENDIX 2

Quadruplicate Field Measurements and Temporal
Variations of Radon at Station P1, Puna

Film N ^o .	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U. (T · 10 ⁻² /cm ² /hr)
P1/A	16	11	67.40	16.32
P1/B	192	187	959.85	19.48
P1/C	150	145	796.90	18.20
P1/D	382	377	1559.90	24.17
P1/E	188	183	888.50	20.59
P1/F/1a	205	200	970.15	20.62
P1/F/1b	218	213	970.15	21.96
P1/F/2a	180	175	971.20	18.06
P1/F/2b	194	189	971.20	19.46

Mean of quadruplicate set (P1/F) = 20.02

APPENDIX 3

Variation of Radon Emanation From Soil
With Sample Size

Sample N ^o .	Sample Mass (g)	T/cm ²	Minus Film Background	Exposure (hours)	R.U ₂ (T · 10 ⁻² /cm ² /hr)
SL1	10	42	39	743.81	5.24
SL2	20	158	155	743.79	20.84
SL3	30	176	173	743.75	23.26
SL4	40	204	201	526.54	38.17
SL5	50	233	230	526.54	43.68
SL6	60	300	297	526.58	56.40

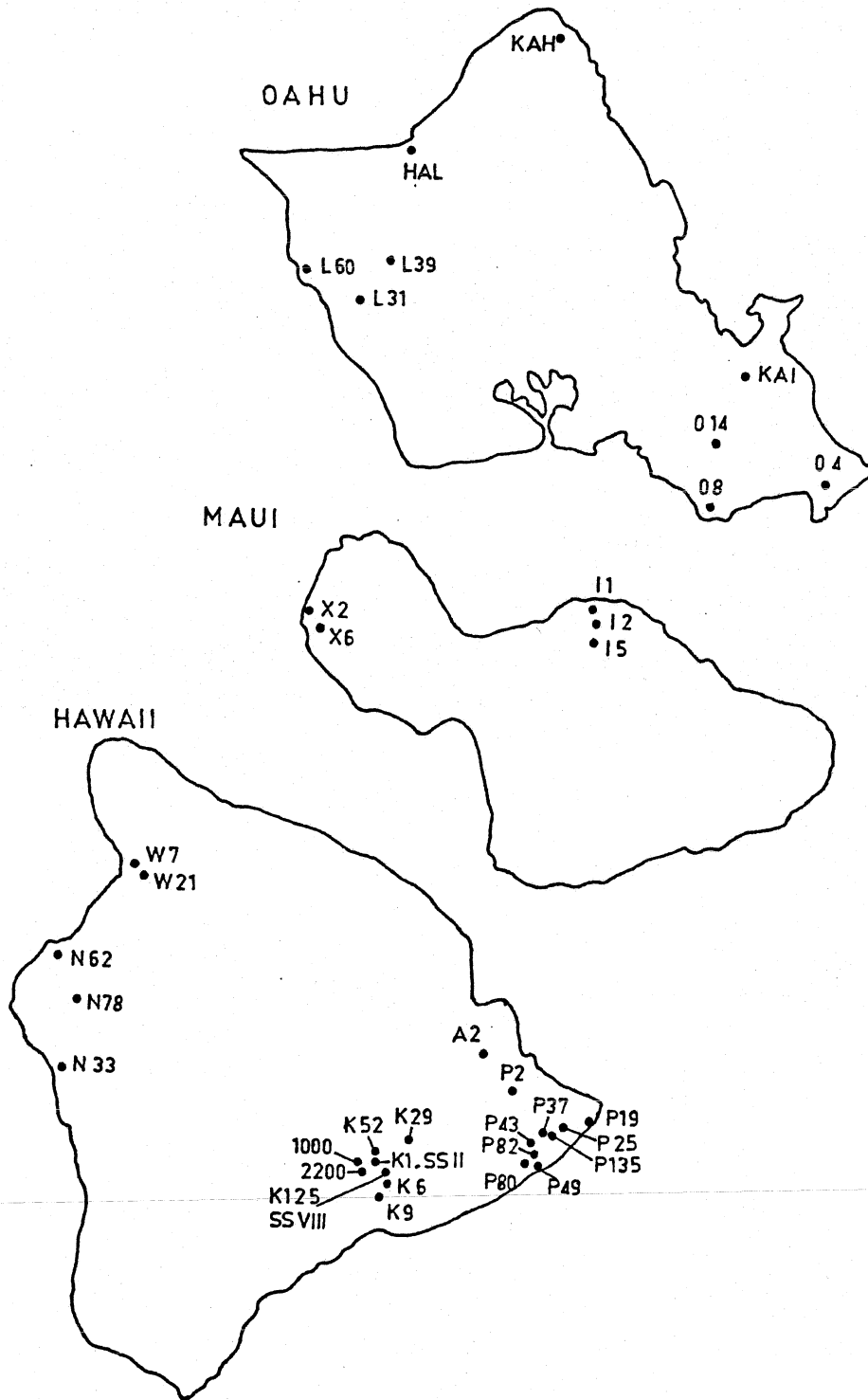


Figure 35. Location of soil samples in Appendix 4.

APPENDIX 4
Properties of Soil Samples

Sample N ^o .	pH	Weight % Organics	Size Fraction (mm)			
			> 1.0	1.0-0.5	0.5-0.25	< 0.25
P2	5.3	25.3	38.5	26.7	10.4	24.4
P19	6.2	1.6	50.7	16.0	8.6	24.7
P25	5.2	7.1	63.8	12.9	8.8	14.5
P37	5.2	2.5	90.0	5.2	1.9	2.9
P43	6.3	1.3	73.9	7.2	6.2	12.7
P49	5.7	16.4	41.0	19.7	29.4	9.9
P80	5.9	0.5	95.6	1.7	0.9	1.8
P82	7.0	1.4	84.8	6.7	3.1	5.4
P135	5.9	14.7	73.0	18.0	5.4	3.6
K6	5.7	2.4	77.9	14.4	2.2	5.5
K9	6.4	1.9	54.9	14.2	7.9	23.0
K29	5.9	6.1	69.7	13.6	5.2	11.5
K52	5.9	2.5	53.1	16.4	9.9	20.6
K125	6.0	2.0	71.2	12.6	6.2	10.0
1000	4.9	0.5	36.6	16.4	10.3	36.7
2200	6.0	0.6	31.4	26.6	12.6	29.4
L31	5.2	6.2	71.1	16.4	5.8	6.7
L39	6.6
L60	1.5	5.6	62.1	18.2	7.7	12.0
N33	..	6.8	75.3	11.8	4.5	8.4
N62	..	1.7	77.0	5.0	2.8	15.2
N78	..	6.1	84.7	5.4	2.7	7.2
W7	..	17.4	36.2	16.1	10.7	37.0
W21	..	7.6	17.1	15.3	14.0	53.8
I1	..	11.2	60.7	15.4	6.9	17.0
I2	8.0	13.4	81.8	7.7	3.1	7.4

APPENDIX 4 - continued

Sample N ^o .	pH	Weight % Organics	> 1.0	Size Fraction (mm)		
				1.0-0,5	0,5-0,25	< 0,25
I5	..	18.9	60.1	15.2	6.6	18.1
X2	..	9.2	52.4	18.7	8.5	20.4
X6	..	4.9	60.8	10.5	8.4	20.3
04	7.0	10.1	43.7	9.7	7.8	38.8
08	7.5	5.8	49.2	24.2	9.2	17.4
014	6.1	7.8	76.2	13.3	4.7	5.8
A2	..	3.8	71.0	6.1	5.1	17.8
Hal	7.5	7.5	44.3	29.1	9.9	16.7
Kah	7.3	8.2	79.1	10.3	4.4	6.2
Kai	5.6	10.7	73.5	11.2	5.1	10.2

Locations are shown on Fig. 35.

APPENDIX 5
U and Th Content and Rn Emanation
From Soil Samples

Sample N ^o .	U (ppm)	Th (ppm)	Th/U	Rn (R.U.)	Ra-226* (pCi/g)
P2	1.67	..
P19	0.643	1.556	2.42	4.57	0.23
P25	0.849	1.369	1.61	5.30	0.31
P37	0.395	1.531	3.88	1.45	0.14
P43	0.515	0.532	1.03	2.40	0.19
P49	9.98	..
P80	0.284	0.723	2.55	2.15	0.10
P82	0.726	1.471	2.03	0.92	0.26
P135	0.705	1.490	2.11	1.46	0.25
K9	0.364	0.679	1.87	2.60	0.13
K29	0.353	0.813	2.30	4.05	0.13
K52	1.23	..
K125	11.36	..
1000	0.41	..
2200	0.51	..
L31	1.075	2.288	2.13	13.23	0.39
L39	0.758	2.979	3.93	10.16	0.27
L60	18.83	..
N33	0.453	1.029	2.27	3.67	0.16
N62	0.648	1.777	2.74	1.18	0.24
N78	0.898	1.856	2.07	1.05	0.32
W7	0.556	1.036	1.86	14.07	0.20
W21	11.32	..
I1	1.874	9.314	4.97	96.02	0.68
I2	31.40	..
I5	1.813	5.551	3.06	31.86	0.65
X2	9.85	..
X6	11.26	..

APPENDIX 5 - continued

Sample N ^o .	U (ppm)	Th (ppm)	Th/U	Rn (R.U.)	Ra-226 * (pCi/g)
014	0.909	2.344	2.58	2.17	0.33
A2	0.181	0.716	3.96	1.57	0.07
Hal	6.40	..
Kah	0.975	1.954	2.00	6.20	0.35
Kai	0.612	2.494	4.08	9.20	0.22

*approximation only; calculated from activities ratio
 $Ra-226/U-238 = 2.78 \cdot 10^6$ [Wollenberg, 1975].

	<u>Duplicate Measurements</u>		(average used in table)
P135A	0.766	1.612	2.10
P135B	0.643	1.367	2.13
P25-2	0.915	1.706	1.86
P25	0.782	1.032	1.32
L39	0.826	3.293	3.97
L39	0.689	2.664	3.87
P80	0.340	0.666	1.96
P80	0.228	0.779	3.42
P37-2	0.375	2.385	6.36
P37-2	0.415	0.677	1.63

APPENDIX 6

Radon Monitoring Data Station T21, Puna

Ref N ^o .	Measurement Period	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U. (T · 10 ⁻² /cm ² /hr)	R.U. Corrected
1979						
1	6/29-8/3	37	32	843.22	3.79	1.64
2	8/3-9/15	70	65	1051.98	6.18	4.03
3	9/15-10/29	30	25	1036.77	2.41	0.26
4	10/29-11/25	30	25	645.88	3.87	1.72
5	11/25-12/21	9	4	620.35	0.64	-1.51
6	12/21-1/15	19	14	601.46	2.33	0.18
1980						
7	1/15-2/6	23	18	528.18	3.41	1.26
8	2/6-2/27	12	7	506.12	1.38	-0.77
9	2/27-3/20	9	4	526.86	0.76	-1.39
10	3/20-4/10	7	2	503.60	0.40	-1.75
11	4/10-5/3	6	1	554.50	0.18	-1.97
12	5/3-5/26	19	14	550.64	2.54	0.39
13	5/26-6/15	25	20	480.02	4.17	2.02
14	6/15-7/6	56	51	505.19	10.10	7.95
15	7/16-7/29	25	20	550.70	3.63	1.48

APPENDIX 6 - continued

Ref No.	Measurement Period	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U. (T · 10 ⁻² /cm ² /hr)	R.U. Corrected
16	7/29-8/21	32	27	547.93	4.93	2.78
17	8/21-9/13	43	38	552.47	6.88	4.73
18	9/13-10/4	66	61	504.09	12.10	9.95

Description of Seismic Events

- A: Magnitude 5.4 earthquake to south of central east rift, 21 September, 1979
- B: Eruption of Pauahi Crater, Kilauea, 16-17 November, 1979
- C: Seismic swarms indicating intrusion on upper east rift, 2 March, 1980
- D: " " " " " " " " 10-11 March, 1980
- E: " " " " " " " " 30 July, 1980
- F: Earthquake in Puna region, 6 August, 1980
- G: Seismic swarms, indicating intrusion on upper east rift, 27 August, 1980

APPENDIX 7

Rn, pH and Hg Data Southwest Rift, Kilauea

Station No	Hg (ppm)	pH	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U. (T · 10 ⁻² /cm ² /hr)	Soil Background (T · 10 ⁻² /cm ² /hr)	R.U.
-0100	13	4.5
0000	14	4.9	145	140	1178.15	11.88	0.41	11.47
0200	18	5.3	62	57	1178.20	4.84	0.41	4.43
0400	16	4.6	49	44	1177.90	3.74	0.41	3.33
0600	37	5.2	71	66	1177.45	5.61	0.41	5.20
0700	17	5.0
0800	22	5.1	58	53	1177.45	4.50	0.41	4.09
1000	19	4.9	107	102	1177.75	8.66	0.41	8.25
1100	14	4.8
1200	12	4.7	74	69	1177.60	5.86	0.41	5.45
1300	19	5.2
1400	16	5.3	43	38	1176.97	3.23	0.41	2.82
1600	13	5.3	48	43	1176.85	3.65	0.41	3.24
1800	16	5.4	25	20	1176.72	1.70	0.41	1.29
2000	14	5.9	25	20	1176.20	1.70	0.41	1.29
2200	13	6.0	123	118	1176.05	10.03	0.51	9.52

APPENDIX 7 - continued

Station N ^o .	Hg (ppm)	pH	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U. (T · 10 ⁻² /cm ² /hr)	Soil Background (T · 10 ⁻² /cm ⁻² /hr)	R.U.
2400	18	6.1	26	21	1154.05	1.82	0.51	1.31
2600	21	6.2	23	18	1153.95	1.56	0.51	1.05
2800	13	6.4	22	17	1153.85	1.47	0.51	0.96
3000	19	6.0	33	28	1153.40	2.43	0.51	1.92

Traverse was established 19 January-10 March, 1979

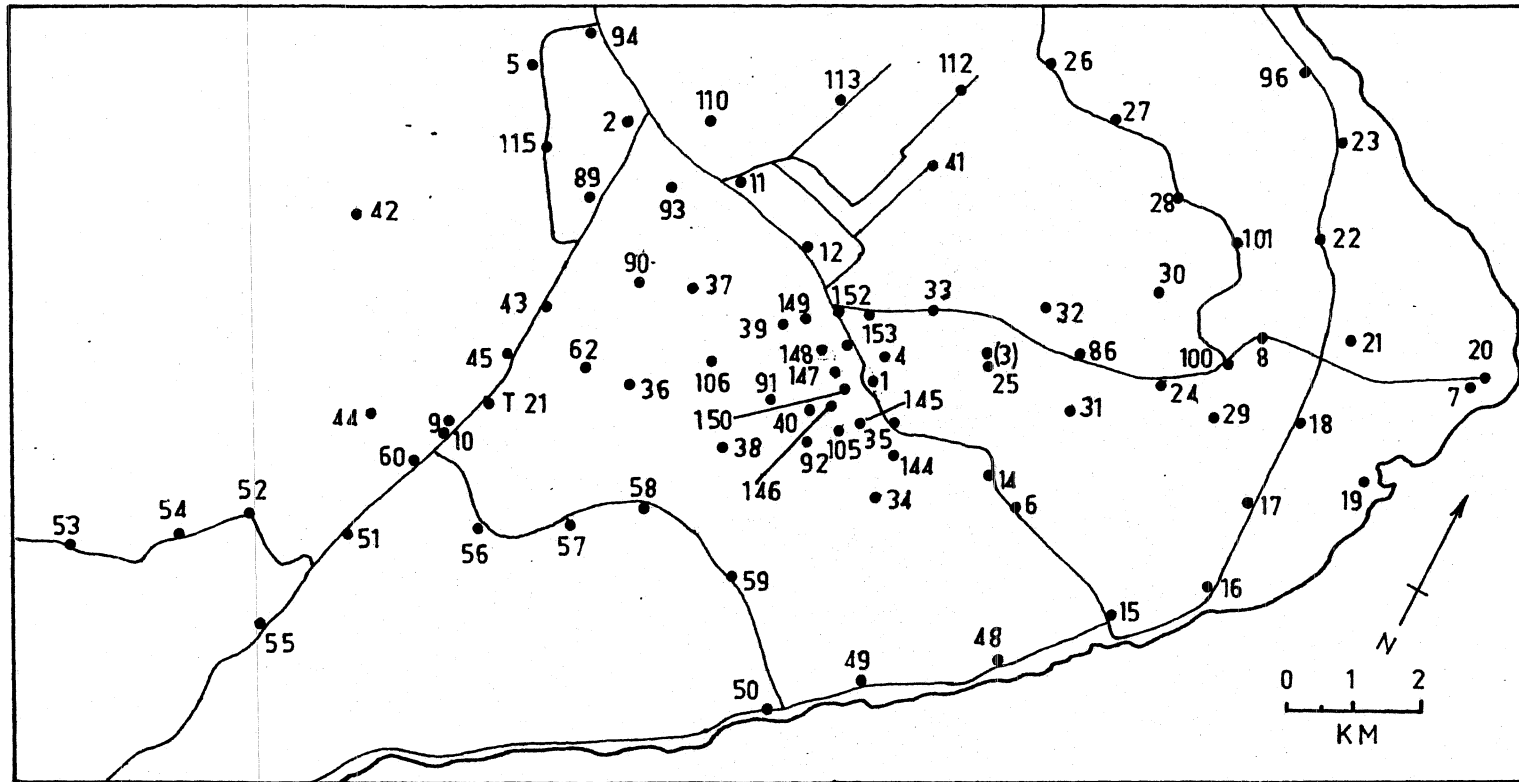


Figure 36. Numbers and locations of radon measurement stations in the Puna area (prefix P). T21 is the location of the monitored station.

APPENDIX 8
Rn and pH Data Puna Area

Station N ^o	pH	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U.	Soil Background (R.U.)	R.U.
P1	5.7	192	187	959.85	19.48	1.24	18.24
P2	5.3	199	194	739.20	26.24	5.12	21.12
P3	..	650	645	1034.45	62.35
P4	5.8	235	230	957.30	24.03	1.24	22.79
P5	5.0	60	55	739.15	7.44	2.15	5.29
P6	6.2	92	87	1009.70	8.62	1.24	7.38
P7	..	44	39	1011.45	3.86	0.92	2.92
P8	6.2	145	140	1012.95	13.82	1.24	12.58
P9	5.4	312	307	1061.10	28.93	2.15	26.78
P10	Film destroyed by heat						
P11	6.0	43	38	793.00	4.79	2.15	2.64
P12	5.5	25	20	792.00	2.52	2.15	0.37
P13	6.2	103	98	797.65	12.29	5.12	7.17
P14	6.3	17	12	795.10	1.51	1.24	0.27
P15	5.8	16	11	794.00	1.39	2.15	-0.76
P16	5.5	40	35	793.00	4.41	1.24	3.17
P17	4.7	16	11	792.70	1.39	1.24	0.15
P18	6.0	13	8	792.05	1.01	1.24	-0.23
P19	6.2	9	4	791.10	0.51	0.92	-0.41
P20	6.1	13	8	790.70	1.01	0.92	0.09
P21	6.3	20	15	790.05	1.90	1.24	0.66
P22	6.1	18	13	789.55	1.65	0.92	0.73
P23	5.9	27	22	789.00	2.79	5.12	-2.33
P24	5.7	74	69	787.00	8.77	1.24	7.53
P25	5.2	114	109	791.00	13.78	5.12	8.66
P26	4.6	85	80	769.92	10.39	1.24	9.15
P27	6.2	8	3	769.25	0.39	1.24	-0.85
P28	6.0	66	61	769.00	7.93	1.24	6.69
P29	5.7	22	17	767.95	2.21	2.15	0.06
P30	5.6	160	155	767.20	20.20	1.24	18.96
P31	5.1	20	15	771.80	1.94	1.24	0.70
P32	6.0	11	6	766.22	0.78	1.24	-0.46
P33	5.7	21	16	765.90	2.09	1.24	0.85
P34	5.9	34	29	749.87	2.53	2.15	0.38
P35	6.1	55	50	748.75	6.68	1.24	5.44
P36	6.4	57	52	766.20	6.79	1.24	5.55
P37	5.2	422	417	766.80	54.38	1.24	53.14
P38	5.9	107	102	765.95	13.32	1.24	12.08
P39	6.2	24	19	766.00	2.48	1.24	1.24
P40	5.8	8	3	765.05	0.39	1.24	-0.85
P41	5.8	6	1	740.70	0.01	1.24	-1.23

APPENDIX 8 - continued

Station No	pH	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U.	Soil Background (R.U.)	R.U.
P42	6.1	50	45	738.95	6.09	2.15	3.94
P43	6.3	145	140	761.20	18.39	0.92	17.47
P44	6.0	57	52	762.10	6.82	0.92	5.90
P45	5.9	79	74	761.35	9.72	2.15	7.57
P48	5.5	42	37	747.05	4.95	2.15	2.80
P49	5.7	88	83	746.75	11.12	5.12	6.00
P50	5.4	55	50	746.10	6.70	5.12	1.58
P51	6.4	27	22	743.60	2.96	0.92	2.04
P52	6.5	40	35	743.10	4.71	1.24	3.47
P53	6.5	13	8	743.10	1.08	5.12	-4.04
P54	6.5	13	8	742.80	1.08	2.15	-1.07
P55	5.6	36	31	742.75	4.17	1.24	2.93
P56	6.4	78	73	742.10	9.84	1.24	8.60
P57	5.6	40	35	746.15	4.69	1.24	3.45
P58	5.9	41	36	741.80	4.85	1.24	3.61
P59	6.6	40	35	741.25	4.72	2.15	2.57
P60	5.5	25	20	739.65	2.70	1.24	1.46
P62	5.1	71	66	737.40	8.95	1.24	7.71
P86	5.5	155	150	915.00	16.39	1.24	15.15
P89	5.8	51	46	918.05	5.01	2.15	2.86
P90	6.1	51	46	917.00	5.02	1.24	3.78
P91	6.4	433	428	916.65	46.69	1.24	45.45
P92	6.7	70	65	915.90	7.10	0.92	6.18
P93	7.2	42	37	909.95	4.07	1.24	2.83
P94	7.2	34	29	850.40	3.87	2.15	1.72
P96	7.2	26	21	867.20	2.74	5.12	-2.38
P100	6.3	38	33	866.30	4.31	1.24	3.07
P101	5.7	26	21	866.20	2.74	0.92	1.82
P105	6.0	20	15	866.65	1.96	0.96	1.04
P106	5.6	14	9	866.60	1.17	1.24	-0.07
P110	4.6	26	21	845.60	2.82	2.15	0.67
P112	4.6	13	8	845.60	1.07	0.92	0.15
P113	4.6	21	16	845.60	2.15	0.92	1.23
P115	5.0	171	166	845.05	22.29	1.24	21.04
P144	..	6	1	342.65	0.29	1.24	-0.95
P145	..	30	25	342.35	7.30	1.24	6.06
P146	..	69	64	342.35	18.70	1.24	17.46
P147	..	16	11	342.60	3.21	1.24	1.97
P148	..	27	22	342.55	6.42	1.24	5.18
P149	..	12	7	341.97	2.05	1.24	0.81
P150	..	62	57	341.90	16.70	1.24	15.46
P152	..	95	90	342.05	26.31	1.24	25.07
P153	..	14	9	341.44	2.64	1.24	1.40

Appendices only include pH data from radon measurement location.

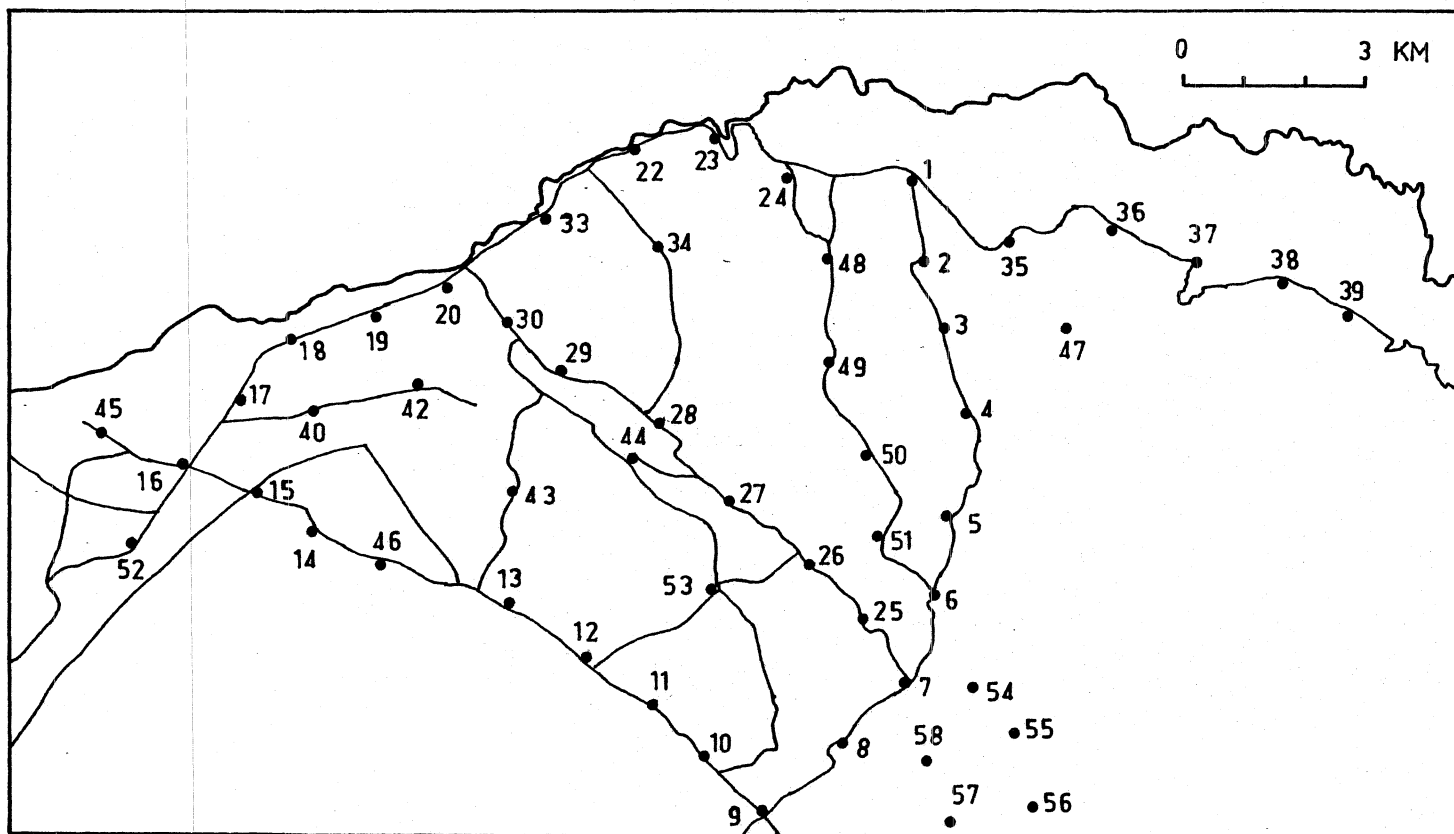


Figure 37. Numbers and locations of radon measurement stations in the Haiku area (prefix I).

APPENDIX 9
Rn and pH Data, Haiku Area

Station No	pH	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U.	Soil Background (R.U.)	Corrected R.U.
I1	..	1046	1041	698.07	149.10	96.02	53.08
I2	8.0	273	268	697.95	38.40	31.40	7.00
I3	6.7	259	254	697.81	36.40	31.40	5.00
I4	6.0		Detector removed				
I5	..	438	433	696.78	62.14	31.85	30.29
I6	6.2	284	279	696.57	40.05	31.85	8.20
I7	6.4	963	958	695.85	137.70	23.59	114.11
I8	6.9	381	376	695.31	54.08	23.59	30.49
I9	7.8	239	234	695.80	33.63	23.59	10.04
I10	6.5	498	493	695.60	70.87	23.59	47.28
I11	6.6	179	174	694.80	25.04	23.59	1.45
I12	7.9	552	547	694.65	78.74	23.59	55.15
I13	8.2	200	195	694.12	28.09	23.59	4.50
I14	7.3	89	84	694.88	12.09	12.18	-0.09
I15	7.5	493	488	694.20	70.30	12.18	58.12
I16	6.9	796	791	688.60	114.90	12.18	102.72
I17	8.5	667	662	672.77	98.40	12.18	86.22
I18	7.8	1152	1147	672.65	170.50	12.18	158.32
I19	7.7	1014	1009	672.20	150.10	12.18	137.92
I20	8.4	328	323	672.00	48.07	12.18	34.89
I21	7.5		Detector removed				
I22	7.5	2881	2876	670.79	429.70	12.18	416.52
I23	5.0	1474	1469	670.55	219.10	5.84	213.26
I24	7.1	2199	2194	669.61	327.70	12.18	315.52
I25	6.8	1240	1235	693.60	178.10	23.59	154.51
I26	6.5	1485	1480	693.10	213.50	12.27	201.23

APPENDIX 9 - continued

Station No	pH	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U.	Soil Background (R.U.)	Corrected R.U.
I27	6.8	1042	1037	705.92	146.90	12.18	134.72
I28	7.8	1824	1819	692.80	262.60	12.27	250.33
I29	7.5	137	132	692.15	19.07	12.18	6.89
I30	7.6	106	101	691.43	14.61	12.18	2.43
I31	5.0		Detector disturbed				
I32	4.7		"	"			
I33	7.5	726.5	721.5	791.1	91.20	12.18	79.02
I34	7.3	1066.5	1061.5	791.0	134.20	12.18	122.02
I35	4.8	667.0	662.0	790.82	83.71	31.40	52.31
I36	4.9	1097.0	1092.0	786.76	138.80	31.40	107.40
I37	5.9	226.0	221.0	786.21	28.11	12.18	15.93
I38	4.7	313.0	308.0	786.01	39.19	12.18	27.01
I39	5.2	1081.0	1076.0	785.33	137.0	12.18	124.82
I40	6.9	489.0	484.0	772.04	62.69	12.18	50.51
I41	6.5	225.0	220.0	771.90	28.50	12.18	16.32
I42	6.9		Detector removed				
I43	6.8	1453.0	1448.0	771.20	187.8	23.59	164.21
I44	6.6	397.0	392.0	769.00	50.98	12.18	38.80
I45	7.1	377.5	372.5	765.56	48.66	12.18	36.48
I46	6.8	308.0	303.0	745.15	40.66	23.59	17.07
I47	6.0	131.5	126.5	716.26	17.66	31.40	-13.74
I48	5.8	244.0	239.0	699.00	34.19	23.59	10.60
I49	5.8	868.0	863.0	698.85	123.50	23.59	99.91
I50	5.9	1299.0	1294.0	698.73	185.2	23.59	161.61
I51	5.0	214.5	209.5	698.2	30.01	23.59	6.42
I52	7.2	282.5	277.5	701.23	39.57	12.18	27.39
I53	6.6	1561.0	1556.0	723.38	215.10	12.27	202.83
I54	5.3	539.0	534.0	675.43	79.06	12.27	66.79

APPENDIX 9 - continued

Station No	pH	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U.	Soil Background (R.U.)	Corrected R.U.
I55	5.3	358.0	353.0	675.33	52.27	12.27	40.00
I56	6.9	209.0	204.0	674.99	30.22	12.27	17.95
I57	6.5	133.0	128.0	674.72	18.97	12.27	6.70
I58	6.2	1293	1288	674.17	191.05	12.27	178.78

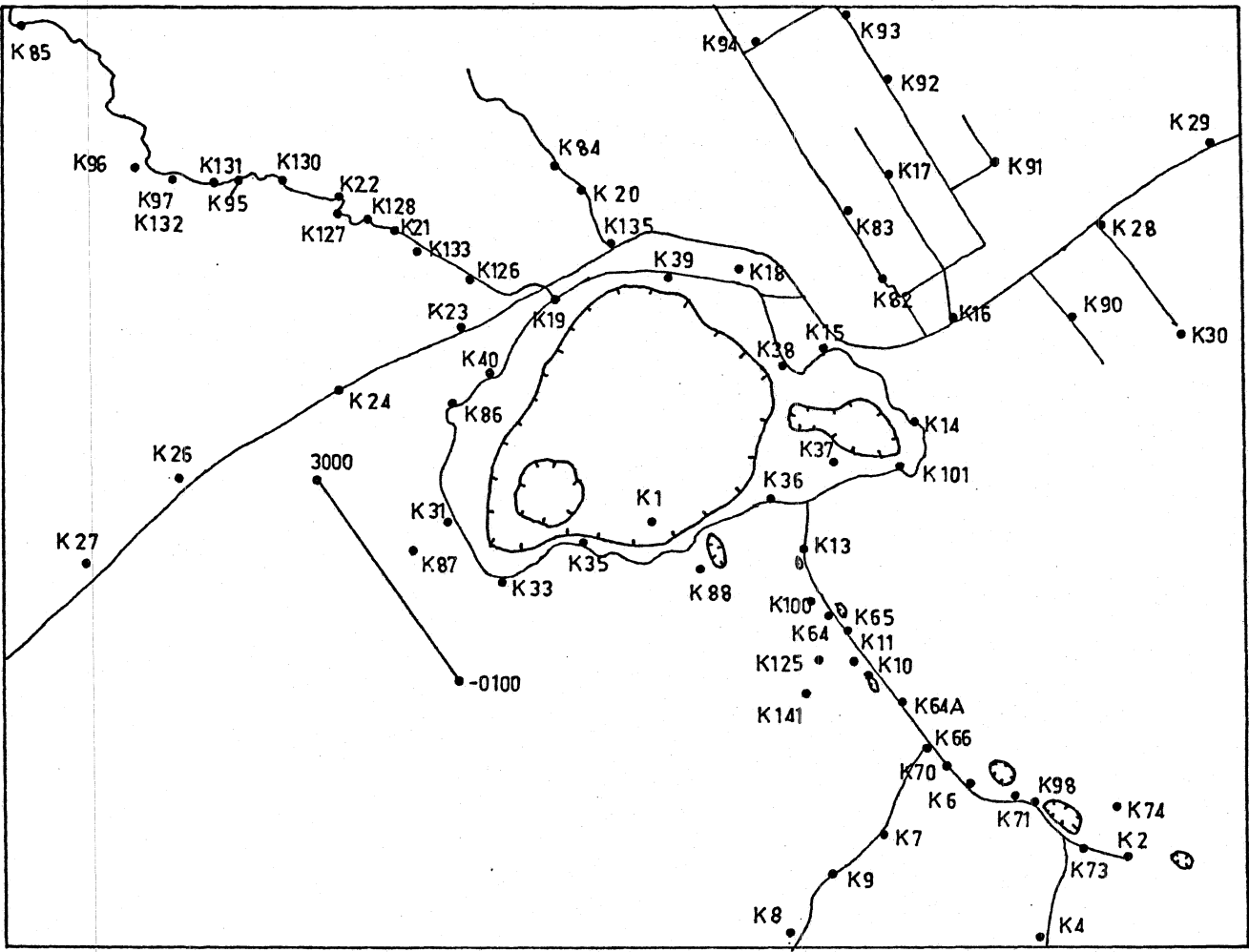


Figure 38. Numbers and locations of radon measurement stations, Kilauea summit. Also shown is SW rift traverse (data Appendix 7).

APPENDIX 10
Rn and pH Data, Kilauea Summit

Station N ^o	pH	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U.	Soil Background (R.U.)	Corrected R.U.
K1	3.6		Film destroyed by heat				
K2	5.7	22	17	770.20	2.21	1.24	0.97
K4	5.7	28	23	769.60	3.00	1.24	1.76
K6	5.7	31	26	768.10	3.39	1.24	2.15
K7	6.3	44	39	767.20	5.08	1.24	3.84
K8	6.0	97	92	767.50	11.99	0.92	11.07
K9	6.4	47	42	767.05	5.48	2.60	2.88
K10	6.6	45	40	765.80	5.22	2.15	3.07
K11	6.2	32	27	764.60	3.53	2.15	1.38
K13	6.0	54	49	763.45	6.42	5.12	1.30
K14	5.6	26	21	743.80	2.82	5.12	-2.30
K15	5.9	19	14	743.65	1.88	5.12	-3.24
K16	5.7	42	37	721.45	5.13	2.15	2.98
K17	5.7	136	131	721.30	18.16	5.12	13.04
K18	5.6	1136	1131	716.85	157.77	2.15	155.62
K19	5.5	46	41	714.75	5.74	0.94	4.82
K20	5.4	58	53	715.30	7.41	1.24	6.17
K21	5.6	20	15	714.20	2.10	1.24	0.86
K22	6.3	368	363	714.10	50.83	5.12	45.71
K23	6.2	23	18	714.00	2.52	1.24	1.28
K24	6.2	32	27	714.00	3.78	1.24	2.54
K26	6.2	42	37	713.90	5.18	0.92	4.26
K27	6.0	20	15	713.80	2.10	0.92	1.18
K28	5.8	28	23	678.90	3.39	1.24	2.15
K29	5.9	27	22	678.30	3.24	1.24	2.00
K30	5.7	21	16	678.60	2.36	2.15	0.21
K31	5.2	75	70	679.70	10.03	0.92	9.11

APPENDIX 10 - continued

Station No	pH	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U.	Soil Background (R.U.)	Corrected R.U.
K33	5.2	24	19	697.65	2.72	0.92	1.80
K35	5.7	32	27	697.20	3.87	0.92	2.95
K36	6.2	28	23	698.25	3.29	0.92	2.37
K37	6.0	34	29	698.10	4.15	2.15	2.00
K38	5.9	118	113	698.05	16.19	1.24	14.95
K39	6.0	500	495	695.15	71.21	0.92	70.29
K40	6.1	18	13	692.75	1.88	0.92	0.96
K64	5.7	40	35	736.72	4.75	1.24	3.51
K64A	..	27	22	748.30	2.94	1.24	1.70
K64	5.2	26	21	735.80	2.85	1.24	1.61
K66	5.7	22	17	748.90	2.27	1.24	1.03
K70	6.1	31	26	753.62	3.45	1.24	2.21
K71	5.8	20	15	738.92	2.03	1.24	0.79
K73	5.8	22	17	726.50	2.34	1.24	1.10
K74	5.9	21	16	727.27	2.20	1.24	0.96
K82	4.9	93	88	941.85	9.34	5.12	4.22
K83	4.9	49	44	941.15	4.68	5.12	-0.44
K84	6.4	55	50	939.00	5.33	1.24	4.09
K85	5.5	143	138	938.00	14.71	2.15	12.56
K86	6.2	76	71	935.95	7.59	0.92	6.67
K87	5.9	292	287	935.70	30.67	0.92	29.75
K88	6.4	35	30	934.50	3.21	0.92	2.29
K90	6.3	32	27	746.15	3.62	5.12	-1.50
K91	6.4	12	7	746.10	0.94	2.15	-1.21
K92	6.2	39	34	746.00	4.56	5.12	-0.56
K93	5.8	45	40	745.95	5.36	5.12	0.24
K94	6.1	84	79	745.87	10.57	2.15	8.44
K95	6.7	19	14	746.00	1.88	2.15	-0.27

APPENDIX 10 - continued

Station No	pH	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U.	Soil Background (R.U.)	Corrected R.U.
K96	6.3	109	104	745.85	13.94	5.12	8.82
K97	5.8		Detector disturbed				
K98	7.5	23	18	746.20	2.41	2.15	0.26
K100	6.8	46	41	745.45	5.50	1.24	4.26
K101	6.4	29	24	742.25	3.23	5.12	-1.89
K125	6.0		Film destroyed by heat				
K126	6.1	22	17	794.55	2.14	1.24	0.90
K127	..	126	121	794.75	15.23	2.15	13.08
K128	..	23	18	794.50	2.27	1.24	1.03
K129	..		Detector disturbed				
K130	..	23	18	793.55	2.27	2.15	0.12
K131	..	23	18	793.85	2.27	2.15	0.12
K132	5.7	30	25	624.88	4.00	2.15	1.85
K133	6.8	17	12	623.95	1.92	1.24	0.68
K135	..	15	10	622.91	1.61	1.24	0.37
K141	..	45	40	522.75	7.65	5.12	2.53

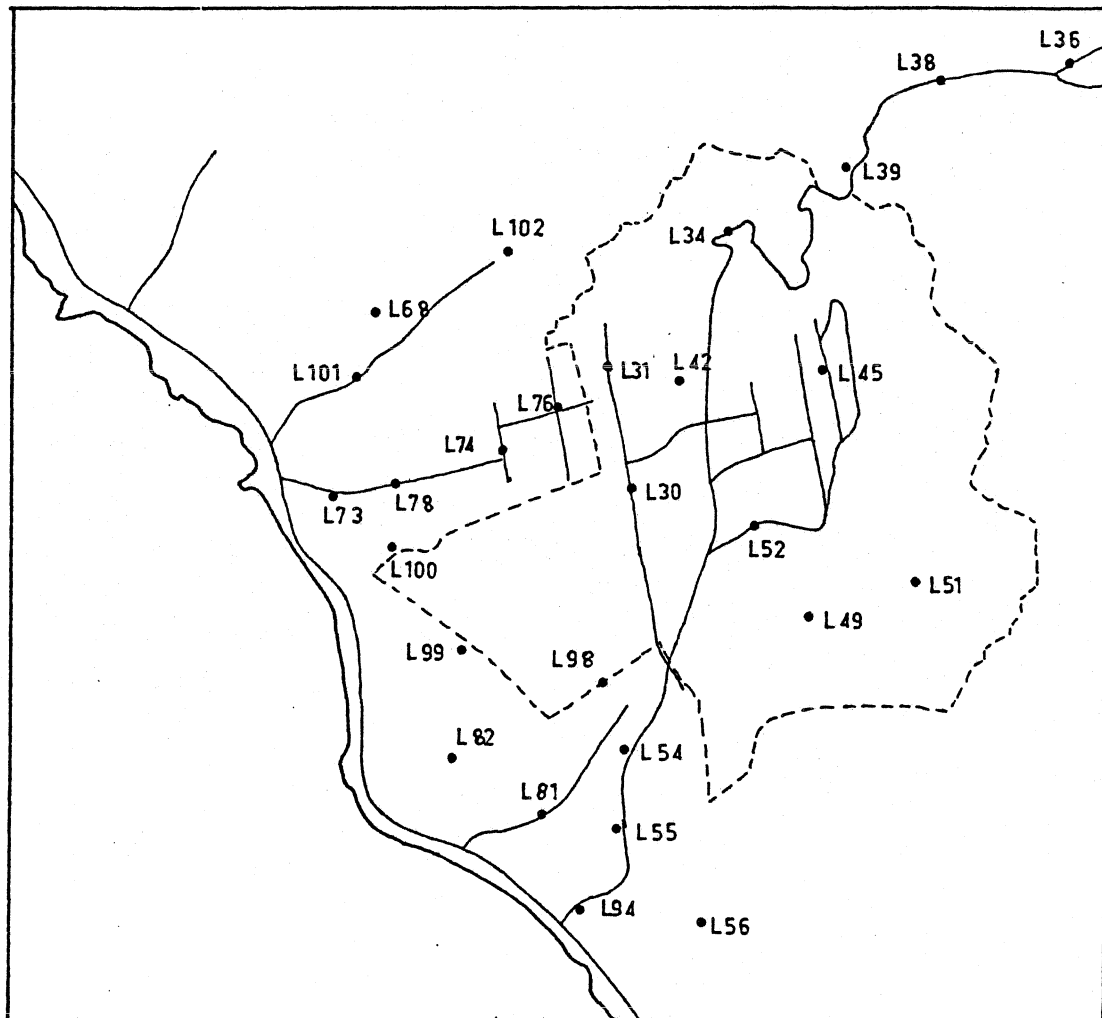


Figure 39. Numbers and locations of radon measurement stations in the Waianae area.

APPENDIX 11
Rn and pH Data, Waianae Area

Station No	pH	T/cm ²	Minus Film Background	Field Exposure (hours)	R.U.	Soil Background (R.U.)	Corrected R.U.
L30	6.7	55	50	722.00	6.93	12.83	-5.91
L31	6.9	75	70	721.90	9.70	12.83	-3.13
L34	6.3	316	311	722.20	43.06	18.39	24.67
L36	6.3	3445	3440	721.95	476.50	18.39	458.11
L38	5.9	1223	1218	721.30	168.90	18.39	150.51
L39	6.6	621	616	721.15	85.42	18.39	67.03
L42	6.5	63	58	719.25	8.06	12.83	-4.77
L45	6.6	228	223	719.03	31.01	9.92	21.09
L49	6.5	386	381	718.9	53.00	12.83	40.17
L51	6.4	196	191	718.4	26.59	9.92	16.67
L52	6.5	118	113	718.2	18.52	9.92	8.60
L54	7.0	108	103	1031.72	9.98	12.83	-2.85
L55	7.0	118	113	1031.62	10.95	12.83	-1.88
L56	6.5	148	143	1031.85	13.86	12.83	1.03
L68	6.5	129	124	698.70	17.75	12.83	4.92
L73	7.4	118	113	696.00	16.21	12.83	3.38
L74	6.7	39	34	696.95	4.88	12.83	-7.95
L76	7.0	116	111	696.85	15.93	9.92	6.01
L78	7.3		Detector removed				
L81	7.2	242	237	1031.15	22.98	12.83	10.15
L82	..	489	484	1031.20	46.94	12.83	34.11
L94	..	361	356	1031.75	34.50	12.83	21.67
L98	..	1060	1055	1031.00	102.30	12.83	89.47
L99	..	1014	1009	1030.57	97.91	12.83	85.08
L100	..	106	101	1030.95	9.80	12.83	-3.03
L101	..	491	486	1030.80	47.18	9.92	37.26
L102	..	1122	1117	1030.25	108.40	18.39	90.01

Only includes pH data from radon measurement locations.

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