The Geology of North America
Vol. N, The Eastern Pacific Ocean and Hawaii
The Geological Society of America, 1989

Chapter 12

The Hawaiian-Emperor Chain

INTRODUCTION

Intraplate volcanism within the Pacific Plate not generated at spreading plate margins is most obvious in Hawaii and the Hawaiian-Emperor volcanic chain. This chain forms a global relief feature of the first order. This chapter consists of five separate sections that summarize the volcanism and geology of Hawaii and the Hawaiian-Emperor chain.

Less obvious but probably greater in overall volume are other seamounts and seamount chains scattered across the northern and eastern Pacific basin. Some of these appear to owe their origin to intraplate volcanism, but many probably formed at mid-ocean ridges. Batiza (this volume, Chapter 13) discusses these other, largely submarine, volcanoes.

The Island of Hawaii lies at the southeastern end of the Hawaiian-Emperor volcanic chain—a dogleg ridge, largely submarine, stretching nearly 6,000 km across the north Pacific Ocean basin. From Hawaii the chain extends northwestward along the Hawaiian Ridge to a major bend beyond Kure Atoll. North of the bend the chain continues in a northward direction as the submarine ridge of the Emperor Seamounts. Volcanoes are active at the southeast end of the chain and become progressively older to the northwest, reaching ages of 75 to 80 million years at the north end of the Emperor Seamounts. Most of this volcanic chain, with an estimated area of 1,200,000 km^2, lies beneath the ocean. Only the Hawaiian Islands and a few atolls of the Hawaiian Ridge, totaling some 16,878 km^2, rise above the sea (Plate 5).

The first section of this chapter, by David Clague and Brent Dalrymple, discusses the overall tectonics, geologic history, and origin of the Hawaiian-Emperor volcanic chain. The petrology and ages of the rocks dredged from the seamounts and collected from the islands indicate that the entire chain has evolved from volcanic activity similar to that presently occurring on the island of Hawaii plus some minor late-phase volcanism occurring up to a few million years after the main volcanic activity.

The second section, by Thomas Wright and David Clague, presents a comprehensive summary of the petrology of Hawaiian lavas. The mineral and chemical composition of these lavas and the sequence in which they are erupted indicate a preshield stage of alkalic basalt, a main stage of tholeiitic basalt, a postshield stage of alkalic basalt, and a rejuvenated stage of strongly undersaturated alkalic lava.

The third section, by Fred Klein and Robert Koyanagi, discusses the seismicity and tectonics of the Island of Hawaii where active volcanism and subsidence make that island the focus of most of the dynamic processes occurring along the Hawaiian-Emperor volcanic chain. Most of the many earthquakes appear to be directly or indirectly related to active volcanism.

The fourth section, by Robert Decker, discusses the physical processes by which magma of the main stage of Hawaiian shield building may form, ascend, and erupt. The historical activity of Kilauea and Mauna Loa volcanoes provides the main evidence for inferring these dynamic processes.

The fifth section, by Donald Thomas, presents a synthesis of the hydrothermal systems formed at the summit and along the east rift zone of Kilauea Volcano. Interactions of fresh and saline ground water with recurring intrusions of basaltic magma at depths of only a few kilometers produce complex geothermal reservoirs.

Many topics about Hawaiian volcanism and the geology of the Hawaiian-Emperor chain are not discussed in these five sections. It is hoped that the extensive references provided at the end of this chapter will help to cover these omissions.

Clague, D. A., and Dalrymple, G. B., 1989, Tectonics, geochronology, and origin of the Hawaiian-Emperor Chain;
Wright, T. L., and Clague, D. A., 1989, Petrology of Hawaiian lava;
Klein, F. W., and Koyanagi, R. Y., 1989, The seismicity and tectonics of Hawaii;
Decker, R. W., 1989, Magma and eruption dynamics;
Thomas, D. M., 1989, Hydrothermal systems in Hawaii;
in Winterer, E. L., Hussong, D. M., and Decker, R. W., eds., The Eastern Pacific Ocean and Hawaii: Boulder, Colorado, Geological Society of America, The Geology of North America, v. N.

TECTONICS, GEOCHRONOLOGY, AND ORIGIN OF THE HAWAIIAN-EMPEROR VOLCANIC CHAIN

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INTRODUCTION

The Hawaiian Islands—the seamounts, banks, and islands of the Hawaiian Ridge—and the seamounts of the Emperor Seamounts (Fig. 1) include more than 107 individual volcanoes with a combined volume slightly greater than 1 million km^3 (Bargar and Jackson, 1974). The chain is age progressive, with still active volcanoes at the southeastern end; the volcanoes at the northwestern end are about 75 to 80 million years old. The volcanic ridge is surrounded by a symmetrical deep as much as 0.7 km deeper than the adjacent ocean floor (Hamilton, 1957). The deep is in turn surrounded by the broad Hawaiian Arch (see Plate 5).

At the southeast end of the chain lie the eight principal Hawaiian Islands. Place names for the islands and seamounts in the chain are shown in Figure 1 or listed in Table 2. The Island of Hawaii includes the volcanoes of Mauna Loa, which last erupted in 1984, and Kilauea, which erupted in 1987. Loihi Seamount, located about 30 km off the southeast coast of Hawaii, is also active and considered to be an embryonic Hawaiian volcano (Moore and others, 1979, 1982). Hualalai Volcano on Hawaii and Haleakala Volcano on Maui have erupted in historical times. Between Niihau and Kure Island, only a few of the volcanoes rise above the sea as small volcanic islets and coral atolls. Beyond Kure the volcanoes are entirely submerged beneath the sea. Approximately 3,450 km northwest of Kilauea, the Hawaiian chain bends sharply to the north and becomes the Emperor Seamount chain, which continues northward another 2,300 km.

It is now clear that this remarkable feature was formed during approximately the past 70 m.y. as the Pacific lithospheric plate moved first north and then west relative to a melting anomaly called the Hawaiian hot spot, located in the asthenosphere. According to this *hot-spot hypothesis*, a trail of volcanoes was formed and left on the ocean floor as each volcano was progressively cut off from its source of lava and a new volcano formed behind it.

Wilson (1963a, c) was the first to propose that the Hawaiian Islands and other parallel volcanic chains in the Pacific were formed by movement of the sea floor over sources of lava in the asthenosphere. Although the Emperor chain was recognized as a northward continuation of the Hawaiian chain by Bezrukov and Udintsev (1955) shortly after the Emperor Seamounts were first described by Tayama (1952) and Dietz (1954), Wilson confined his hypothesis to the volcanoes of the Hawaiian Islands and the Hawaiian Ridge. Christofferson (1968), who also coined the term "hot spot," extended Wilson's idea to include the Emperor Seamounts and suggested that the Hawaiian-Emperor bend represents a major change in the direction of sea-floor spreading, from northward to westward. Morgan (1972a, b) proposed that the Hawaiian and other hot spots are thermal plumes of material rising from the deep mantle and that the worldwide system of hot spots constitutes a reference frame that is fixed relative to Earth's spin axis.

Although experimental testing of the various hypotheses proposed to explain hot spots has so far proven unproductive, the hot-spot hypothesis has several important corollaries that can and have been tested to varying degrees. Foremost among these is that the volcanoes should become progressively older to the west and north as a function of distance from the hot spot. This progressive aging should be measurable with radiometric methods and should also be evident in the degree of erosion, subsidence, and geological evolution of the volcanoes along the chain. A second important corollary is that the latitude of formation of the volcanoes, as recorded in the magnetization of their lava flows, should reflect the present latitude of the hot spot rather than the present latitude of the volcanoes. Third, because the active mechanism is beneath the lithosphere, the Hawaiian-Emperor chain should not be related to the structure of the sea floor. Finally, the volcanic rocks of the volcanoes should be similar in both chemistry and sequence of eruption along the chain or should change in a systematic and coherent way.

In this section we describe the Hawaiian-Emperor volcanic chain. We review the evidence that indicates that all of the corollaries mentioned above are true and that the hot-spot hypothesis is therefore a viable explanation of the origin of the chain. We will also describe the various hypotheses that have been proposed to explain the hot-spot mechanism and discuss their strengths and weaknesses. This section is a condensed version of a paper by Clague and Dalrymple (1987) that includes (1) a more detailed description of the petrology and ages of the individual sampled volcanoes that compose the chain, and (2) a section on petrology of Hawaiian lava.

STRUCTURE AND AGE OF THE UNDERLYING CRUST

The volcanoes of the Hawaiian-Emperor chain were formed by eruption of lava onto the floor of the Pacific Ocean without regard for the age or preexisting structure of the ocean crust or for the presence of preexisting volcanoes. The precise age of the ocean crust beneath much of the chain is poorly known because of the paucity of magnetic anomalies in the area (Fig. 2). The Hawaiian Islands and Ridge east of about Midway Island lie on crust older than anomaly 34 but younger than anomaly M0. In a

HYDROTHERMAL SYSTEMS IN HAWAII

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INTRODUCTION

One consequence of the volcanism that formed the islands of the Hawaiian archipelago is the evolution of hydrothermal activity. Surface manifestations of active hydrothermal systems are evident on the young, eruptively active volcanoes on the island of Hawaii; even on the older, long dormant volcanoes of Maui and Oahu, geophysical and geochemical evidence has been found for lower temperature hydrothermal systems. The physical and chemical characteristics of these systems span a broad range and depend heavily upon the geologic and hydrologic conditions of their formation and evolution.

ELEMENTS OF A HYDROTHERMAL SYSTEM

The geologic conditions necessary for the formation of a hydrothermal system include a persistent source of heat and sufficient fluid recharge, and rock permeability, to allow the formation of a steam- or water-dominated convection system. Thus, because effusive lavas cool almost immediately after eruption, hydrothermal systems form only in proximity to long-lived heat sources such as shallow magma chambers and dike complexes within the rift systems that radiate out from these magma chambers. During periods of intense volcanic activity, a hydrothermal system will be driven by heat provided by the influx of fresh magma from deep within the earth. As this recharge declines, hydrothermal activity begins to draw on the heat stored in the crystallizing intrusive complexes present in the summit and rift systems.

The physical and chemical characteristics of the hydrothermal systems associated with the intrusive complexes will depend on the chemical characteristics of the heat source and reservoir rock and will be strongly influenced by the volume and chemical composition of the fluid recharge to the hydrothermal system. Thus a magma chamber dike complex that receives little meteoric recharge is likely to have a high-temperature vapor-dominated hydrothermal system with a fluid chemistry controlled by the acid gases discharged from recently ascended magma, whereas a system on the terminal submarine ridge of an active rift system will have a chemistry controlled by seawater-basalt reactions.

The fluid recharge to a hydrothermal system will be governed by the porosity and permeability of the intrusives in a hot dike complex as well as that of the surrounding country rock. At shallow depths, fluid permeabilities in extrusive lavas will be extremely high, on the order of several thousands of millidarcies, but at increasing depths, fracture permeability will be a major factor in the reservoir circulation patterns; as temperatures increase, plastic deformation of the rocks will not permit fractures to persist and thus inhibit fluid penetration. The chemistry of the circulating fluids also exerts a strong influence over reservoir permeability: meteoric recharge is largely dependent upon soluble elements in the rock matrix to provide it with reactive species, whereas seawater intruding into a high-temperature hydrothermal system is extremely reactive, resulting in rapid alteration of the reservoir rocks and deposition of secondary minerals (Mottl, 1983; Seyfried and Bischoff, 1981; Seyfried and Mottl, 1982; Thomas, 1987).

The interplay of geologic and geochemical phenomena that control hydrothermal activity in Hawaii is best demonstrated by the hydrothermal systems present on Kilauea volcano, where detailed surface investigations and deep drilling have provided a broad array of data with which to develop models of the active hydrothermal systems present there. Other volcanic systems in the Hawaiian chain have been much less intensively investigated; however, the data that are available do provide insight into their current conditions as well as the aging processes of island-based hydrothermal systems.

KILAUEA VOLCANO

The shallow magma reservoir at Kilauea lies at a depth of approximately 3 to 6 km beneath the summit caldera. Magma is recharged to this reservoir at a rate estimated to be approximately $9 \pm 3 \times 10^6$ m^3/mo (Dzurisin and others, 1984), where it may reside for a period of weeks to years before it is discharged as a summit eruption or as an intrusion into the east or southwest rift zone. Conductive and convective heat loss from the summit reservoir drives an extensive hydrothermal system within the confines of the summit caldera. This system is evidenced by numerous persistent fumaroles and solfataras on the caldera floor and at its boundary faults (Macdonald and others, 1983). The extremely permeable nature of the surface basalts has not permitted the formation of hot springs or surface geysers such as those associated with other caldera systems. The temperature of the discharges from these summit features is typically 96°C or lower, but, in areas of intense outgassing, temperatures of 150°C have been measured (Casadevall and Hazlett, 1983).

The subsurface temperature regime within this hydrothermal system is not presently well defined. Measurements made in the only deep research well drilled in the caldera (Keller, 1974; Zablocki and others, 1974) showed nearly isothermal conditions, at 20°C to 30°C, to a depth of 488 m, where a steep increase occurred, to a temperature of more than 80°C, when a perched water body was encountered. Below the perched aquifer, at about 700 m, the temperature declined for an interval of about 300 m and then began to rise sharply to a maximum temperature of 137°C at a bottom hole depth of 1,250 m. This temperature profile is probably not representative of the entire caldera, but it does suggest that convective processes, through isolated surface vapor discharges and deeper groundwater circulation, are more important than conductive processes in the loss of heat from the summit region.

The typically low temperature of the surface discharges suggests that meteoric recharge buffers the temperature of the shallow hydrothermal system. The buffering action is a function of both the high rate of rainfall at the summit (200 cm/yr; Armstrong, 1973) as well as the high permeability of fractured subaerial basalts in the caldera region (up to 10^4 millidarcies; Keller, 1974). Because the surface basalts cannot sustain a significant pressure differential, the temperature of the fumarolic discharges can exceed the local boiling point only when the magmatic, gas-driven heat flow exceeds that consumed by down-flowing meteoric recharge. Hence, only the most gas-rich discharges exceed 96°C, and even these show temperature variations in response to periods of exceptionally heavy rainfall (Casadevall and Hazlett, 1983).

Because the fluid recharge to the summit hydrothermal system is predominantly meteoric water, the chemistry of the hydrothermal system is strongly influenced by volatiles discharged from the magma chamber. The fumarolic discharges contain substantial concentrations of carbon dioxide and sulfur dioxide and lesser amounts of hydrochloric, hydrofluoric, and sulfuric acids, and in some of the lower temperature fumaroles, hydrogen sulfide (Naughton and others, 1963; Greenland and others, 1985). Mass discharge rates of carbon dioxide and sulfur dioxide from the summit chamber have been estimated using the magma recharge rate of Dzurisin and others (1984) and volatile partitioning data (Gerlach and Graeber, 1985), to be approximately 3.7×10^9 g/day and 7.2×10^8 g/day respectively; discharge of carbon dioxide and sulfur dioxide from the thermal features on the caldera floor have been measured at 1.6×10^9 g/day (Greenland and others, 1985) and 1.5 to 3.0×10^8 g/day (Casadevall and Hazlett, 1983; Greenland and others, 1985) respectively. The gas discharge rates estimated from the magma recharge rate are in remarkably good agreement with the measured values for carbon dioxide. The apparent discrepancy in the measured and predicted sulfur emission rates suggests that loss of the more reactive sulfur species from the fumarolic discharges is quite extensive. Candidate mechanisms for removal of sulfur gases include disproportionation, air-oxidation, or auto-oxidation (Claus reaction) of sulfur dioxide and hydrogen sulfide (Mizutani and Sugiura, 1966; Giggenbach, 1980).

The inferred depth of Kilauea's summit magma chamber places the top of the intrusive complex at least 1 km below sea level. Thus the shallow, vapor-dominated hydrothermal system seen at the summit is probably underlain by a liquid-dominated zone at or near the boiling point. The sulfur dioxide disproportionation reaction noted above and the inferred mass emissions of hydrochloric and hydrofluoric acids from the magma chamber (Gerlach and Graeber, 1985) suggest that this zone will be strongly acid and contain moderate to high concentrations of acid-leached salts from the basalt matrix.

The hydrothermal system associated with the Kilauea summit caldera can be inferred to be a gas-rich, steam-dominated system at shallow levels; and at depth, a highly acid, liquid-dominated system driven by volatile discharge and conductive heat loss from the roof of the summit magma chamber (Fig. 52).

KILAUEA EAST RIFT ZONE

Geologic mapping of older, eroded rift systems (Macdonald and others, 1983; Ryan and others, 1983) indicates that the shallow (<3 km depth) intrusive complex associated with the Kilauea east rift zone (ERZ) consists of near-vertical, tabular bodies ranging in thickness from a few centimeters to several meters, interspersed with screens of extruded lavas. The core of the rift zone, where intrusives make up the bulk of the rock mass, is believed to consist of a nearly continuous zone of plastic, near-molten rock (Hardee, 1982; Dzurisin and others, 1984; Swanson and others, 1976b). Magma recharge into the ERZ has been estimated by Dzurisin and others (1984) to be approximately 5×10^6 m^3/mo, resulting in a heat influx of about 2,800 megawatts of thermal energy (Thomas, 1987). The width of the dike complex is not well known; aeromagnetic data (Godson and others, 1981) and the distribution of surface vents and ground cracks (Holcomb, 1980) suggest a width of 2 to 4 km, whereas interpretation of gravity and ground-based magnetic data suggest that the width of the earlier, now buried, rift zone may extend to 10 to 25 km at depth (Furumoto, 1978).

The ERZ dike complex has a profound influence on the shallow groundwater hydrology of the eastern flank of Kilauea; every groundwater well drilled within the surface expression of the rift zone (Fig. 53) produces brackish water having temperatures ranging from 35°C to near boiling (Thomas, 1987); wells south and downgradient of the rift show geochemical and sometimes thermal evidence of being influenced by hydrothermal outflow from the rift. Geophysical surveys along virtually the entire length of the rift zone (Keller and others, 1977; Kauahikaua and Mattice, 1981) show low-resistivity anomalies that are attributed to outflow of thermally altered or saline groundwater and indicate that an active hydrothermal system is associated with the entire length of the ERZ.

THE HYDROTHERMAL SYSTEM

Our understanding of conditions within the hydrothermal system on the ERZ has been greatly assisted by eight deep geothermal exploration and research wells that have been drilled into the rift during the last decade. These wells have all been drilled into the lower rift area (Fig. 53) and have penetrated to depths ranging from 1,920 to 2,557 m. Although complete geology and engineering data sets are not available from the privately funded

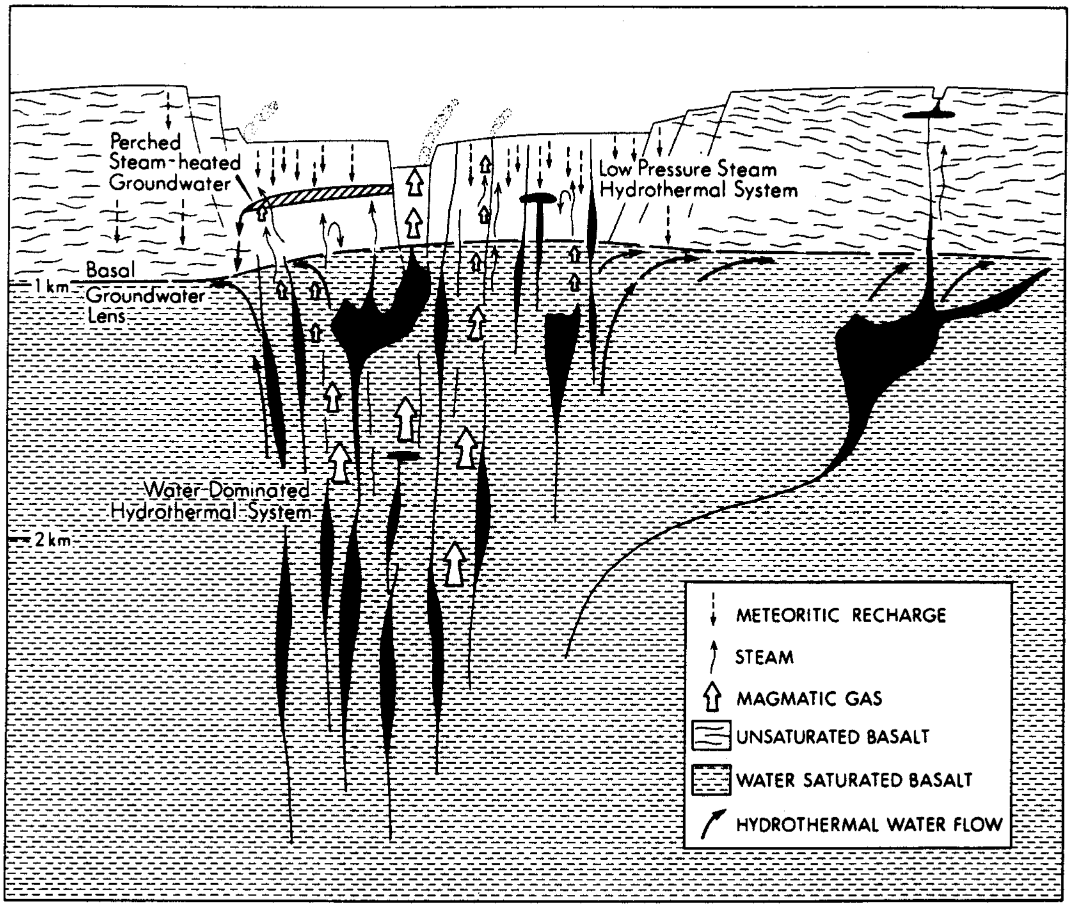

Figure 52. Conceptual model of hydrothermal system associated with the Kilauea summit caldera and magma chamber.

exploration wells, we do have sufficient data to draw a generalized model of the active hydrothermal system associated with the ERZ.

The deep wells (Fig. 54) encountered an interval of warm (30°C to 100°C), nearly isothermal conditions where large volumes of meteoric recharge (2.5×10^6 $m^3/yr/km^2$; Thomas, 1987) mix with deep hydrothermal fluids discharged from depth. In most wells this mixing interval extended to depths of nearly 1,000 m, where temperature gradients began to steepen. The rate of temperature increase varied widely from well to well, however. Wells that encountered productive aquifers showed a steep temperature gradient overlying a nearly isothermal region where permeabilities were high enough to allow hydrothermal circulation and mixing, whereas those that encountered low permeability formations showed a conductive temperature gradient to total depth. Bottom-hole temperatures of six of the deep wells exceeded 300°C, and the highest temperature encountered exceeded 370°C. Two wells drilled on the southern edge of the surface manifestation of the rift zone showed temperature reversals at depth (e.g., 2883-04A, Fig. 54) that were interpreted to indicate that the wells had passed through a zone of hydrothermal outflow and had entered cold country rock below. It is believed that the latter wells mark the southern extent of active hydrothermal circulation on the lower ERZ.

Subsurface permeabilities encountered during drilling varied substantially with depth. At shallow depths, extremely high permeabilities were found where interflow a'a rubble or clinker layers and lava tubes permit virtually unhindered fluid flow. At deeper levels, where dikes are interspersed with dense pillow lavas, the predominant flow channels consist of tectonically and volcanically induced fractures. The only well for which deep permeability data are available is the HGP-A well (2883-01, Fig. 54) that was found to have a reservoir permeability thickness of approximately 1,000 millidarcy feet over the lower 1,300 m of hole and was able to sustain a production rate of about 50 tonnes/hr of a mixed fluid consisting of steam and liquid water. Engineering data suggest that the majority of this production was derived from two or three discreet intervals in the well; hence, the conclusion that fluid flow is fracture controlled. Production rate data from the other high-temperature wells on the rift (Iovanitti and D'Olier, 1985; Thomas, 1987) indicate similar or slightly lower permeabilities toward the interior of the rift, and much lower ones on the southern boundary of the rift. It is of note that the characteristics of the fluids produced from these wells has

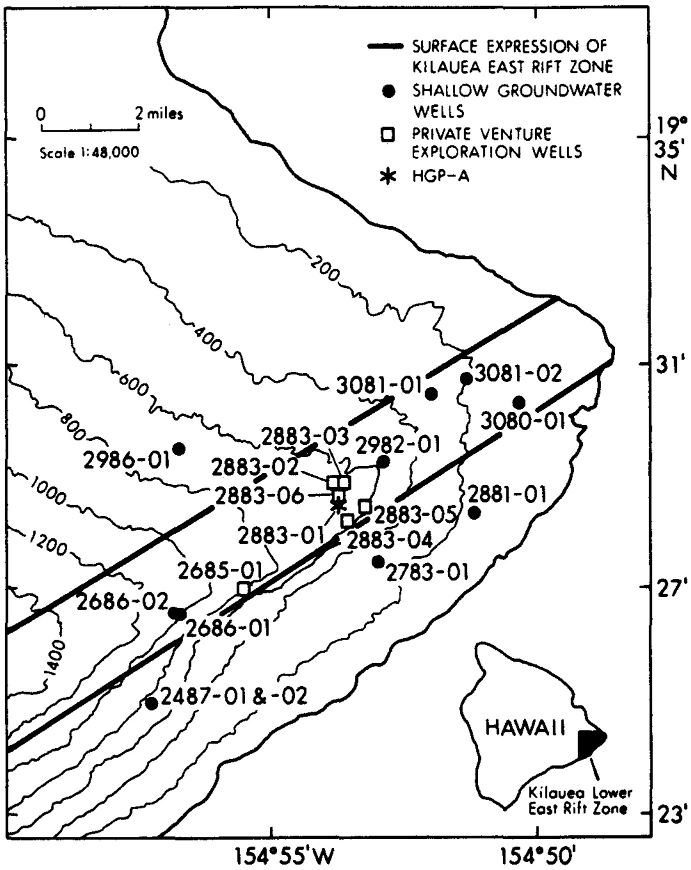

Figure 53. Map of Kilauea east rift zone showing groundwater wells and deep geothermal exploration wells. The numbers adjacent to each well represent the U.S. Geological Survey well designation.

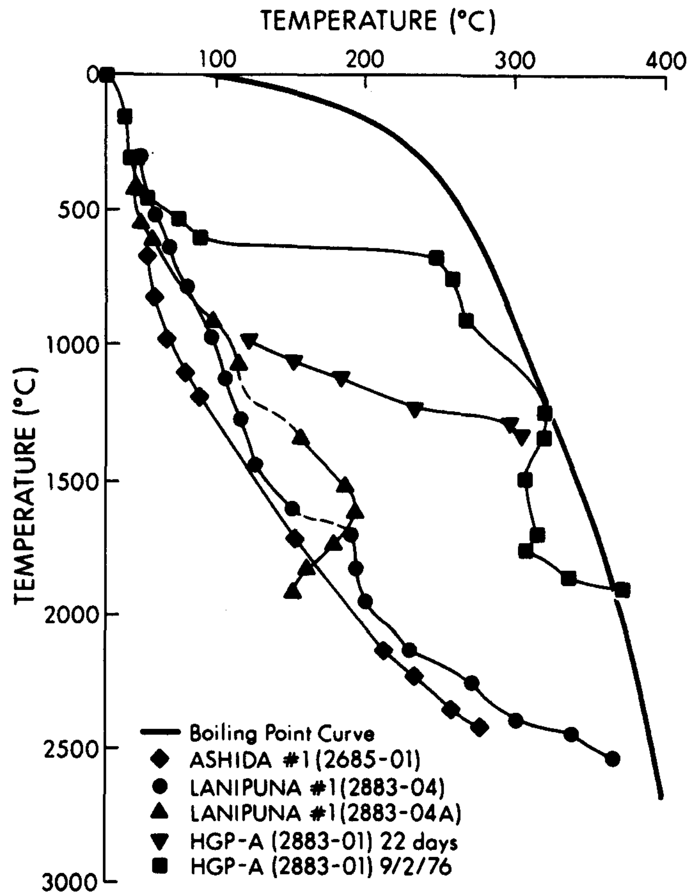

Figure 54. Plot of available temperature gradient data for deep geothermal wells in the east rift zone. Plots are synthesized from several data sets to provide the current best estimate of subsurface temperatures in the rift. Two plots are shown for the HGP-A well showing the temperature profile when fluid circulation is occurring (■) and when circulation is impeded (▼).

varied widely: some (such as HGP-A) have produced mixed fluids of steam and water, whereas others have generated dry steam only (Iovanitti and D'Olier, 1985). Down-hole temperature and pressure data indicate that the unperturbed reservoir consists of a single phase (liquid), and that the occurrence of dry steam in some wells is the result of the low permeability found in the rift zone (Thomas, 1987).

DOWNHOLE MORPHOLOGY AND PETROLOGY

The morphology of the cuttings and cores from the wells in the rift indicates that porous, highly fractured rocks are present throughout the first 300–m section. A transition zone is then encountered in some wells, consisting of oxidized hyaloclastites, below which lie pillow basalts interspersed with intrusive bodies (Thomas, 1987). In general, the density of the pillow basalts increases, and the porosity and permeability decrease with depth of emplacement, reflecting the greater hydrostatic pressures under which they were intruded or extruded (Stone, 1977).

Little alteration beyond normal weathering processes was found in the porous subaerial basalts, but as the depth and temperature increased, intermittent hydrothermal alteration of the reservoir rocks was apparent. The grade of metamorphism was found to generally increase with depth, but in every case, intermittent alteration was found in bands of variable thickness interspersed with bands of virtually unaltered pristine basalt. The alteration assemblage included zeolites, montmorillonite, illite, pyrite, and hematite at shallower levels, and calcite, pyrite, anhydrite, chlorite, epidote, and quartz at depth. The similarity of this assemblage to those found in studies of other seawater-dominated hydrothermal systems (e.g., Reykjanes, Iceland [Kristmannsdottir, 1975, 1983]; the Mid-Atlantic ridge [Humphris and Thompson, 1978]; and laboratory studies of seawater-basalt reactions [Mottl, 1983]) clearly indicates that seawater circulation through this system has had a substantial influence on the hydrothermal alteration found. The discontinuity of the alteration further suggests that seawater entry into the system is not pervasive but probably occurs through fractures generated by volcanic or tectonic activity within the rift. The intense seawater alteration found in some of the wells, although indicative of earlier seawater circulation, does not necessarily correspond to current permeability conditions found in the wells. In spite of pervasive alteration, some wells showed extremely low permeabilities, suggesting that an important aspect of seawater hydrothermal alteration of the basaltic reservoir rocks is the eventual elimination of the fracture permeability by deposition of secondary minerals.

RESERVOIR FLUID CHEMISTRY

The most extensive data base available for the chemistry of the deep reservoir fluids on the ERZ comes from the HGP-A well, which has undergone testing for more than a decade. Preliminary testing of HGP-A showed that the reservoir fluids contained much lower concentrations of dissolved solids than were anticipated on the basis of accepted models of island hydrology (Macdonald and others, 1983; Thomas, 1980). Whereas the latter model predicted a transition from fresh to saline water at depths of less than 400 m, the presence of fresh water at nearly 2,000 m suggests that hydrothermal circulation has permitted cold, fresh water to displace hotter, less dense saline water at depth (Thomas, 1987). The dike complex apparently limits the infiltration of saline fluids by inhibiting seawater intrusion across the rift from the south because of the low permeability of the dike rocks, but the complex permits rapid vertical and east-west movement of fresh water through fractures paralleling the strike of the rift. It is believed that the permeability contrast between fresh and saline water entry is further enhanced by the deposition of secondary minerals from the entering seawater (Thomas, 1987). Thus, access of seawater to the rift zone is controlled by thermally driven convection, the rift zone structure, and chemical alteration associated with high-temperature seawater-basalt reactions. In spite of the limitations imposed by the dike complex on circulation, the residence time of fluids in the rift is indicated by ^{14}C activities to be less than about 12,000 years (Thomas, 1980).

Withdrawal of fluids from the HGP-A well for a continuous period of five years resulted in a 500 percent increase in the salinity of the fluids produced. The major element composition of the increasingly saline fluids shows strong similarities to that found at the Reykjanes geothermal system in Iceland (Ragnarsdottir and others, 1984) and in fluids produced by sea-floor hydrothermal vents (Mottl, 1983) and hence is clearly of seawater origin. However, the relative cation concentrations in the fluids have shown that the intruding fluids have been heavily altered by seawater-basalt reactions; magnesium has been nearly quantitatively removed from the seawater, whereas lithium, potassium, and calcium have been enriched by as much as 2,000 percent (Thomas, 1987). Changes in the relative cation concentration with time also indicate that the degree of alteration is sensitive to the increasing effective seawater:rock ratio in the reservoir. The apparent fluid equilibration temperatures calculated on the basis of the cation concentrations (Fournier, 1981) have also shown a decrease from approximately 300°C to a temperature of about 250°C. Whether this apparent change reflects a real trend in the reservoir temperature or is simply an artifact of the seawater:rock reactions remains in doubt at present (Thomas, 1987). The change in cation chemistry of the fluids was also accompanied by a decline in pH from about 7.4 to about 6.8. This decline is consistent with laboratory results that have shown that, at increasing seawater:rock ratios, the loss of magnesium ion as $Mg(OH)_2$ from seawater results in an increase in hydrogen ion concentration that cannot be buffered by the exchange of other cations—such as calcium—from the reservoir rock; under conditions where seawater greatly exceeded the reactive reservoir basalt available, the pH fell to values as low as 2. Continued production of saline fluids from the HGP-A well may demonstrate whether such conditions occur in a natural hydrothermal system.

The evolution of the fluid chemistry provides insight into the changes that occur as seawater circulation begins in a newly formed fracture system, as well as into the differences between the conditions that might be anticipated in the seawater—and meteoric—water dominated portions of the ERZ. The data suggest that hydrothermal alteration is extremely rapid initially; magnesium in seawater is exchanged for lithium, potassium, and calcium in the fracture wallrock until the latter alkalis are depleted; when this reaction becomes more sluggish the hydrogen ion concentration of the fluids begins to rise, thus generating more aggressive acid, saline hydrothermal fluids. The occurrence of intensely altered, high-temperature, low-permeability aquifers, as well as low-temperature, highly permeable aquifers on the southern flank of the rift, suggests that the ultimate fate of the seawater-dominated system may depend on the extent of the fracture system, as well as its proximity to a heat source. The character of a hydrothermal system dominated by seawater, such as would be found on the southern flank of the ERZ, is indicated to have a highly aggressive fluid chemistry and to rapidly evolve as seawater circulation occurs. This contrasts sharply with the more benign, and possibly more stable, hydrothermal system on the interior of the rift that is recharged predominantly by meteoric water.

The composition of the gases dissolved in the hydrothermal fluids on the lower rift zone is distinctly different from those produced by the fumaroles at Kilauea's summit. The most important differences found are that the predominant sulfur species on the lower rift is hydrogen sulfide, as opposed to sulfur dioxide at the summit, and that the carbon to sulfur ratios of the geothermal gases are nearly an order of magnitude lower for the ERZ than are found at the summit. Although the former difference is easily explainable on the basis of thermodynamic equilibria of sulfur species (Gerlach and Nordlie, 1975; Helgeson and others, 1981), the latter is not as well understood. Two mechanisms may contribute to the carbon-to-sulfur ratio observed in the HGP-A fluids: (1) enrichment of sulfide due to reduction of incoming seawater sulfate by the reservoir rock (McDuff and Edmond, 1982); or (2) the preferential loss of CO_2 from magmas at the summit chamber that are subsequently intruded into the rift zone. Whereas the available iron in the reservoir rocks may limit the sulfide contribution from seawater, recent studies by Gerlach and

Graeber (1985) and Greenland and others (1985) suggest that magma loses a substantial portion of its CO_2 as it ascends to, and resides in, Kilauea's summit reservoir. Hence, the low carbon-to-sulfur ratio found on the ERZ probably reflects the elemental ratios in the reservoir basalts.

The overall picture of the hydrothermal system associated with the ERZ can be inferred to have the general characteristics described in the next two paragraphs.

1. The shallow subsurface environment, owing to the extremely high permeability and moderate to high rainfall, will have low temperatures except in areas of active discharge of deeper hydrothermal fluids; even in these areas, temperatures are unlikely to exceed the boiling point of water at the combined atmospheric and hydrostatic pressures at a given depth. Active hydrothermal circulation within the shallow, permeable aquifers is likely to maintain nearly uniform, or only slowly increasing, temperatures to depths of 500 m or more, where permeabilities begin to decline. The temperature within the deeper portion of the rift zone increases substantially, and although the highest recorded reservoir temperature to date in the rift is slightly more than 368°C (M. Gardener, personal communication, 1986), geophysical data of Godson and others (1981) and Flanigan and Long (1987) suggest that temperatures in excess of the Curie temperature of basalt are present at depth. The temperature profile across the rift zone is somewhat more problematic; at the southern boundary of the rift, temperatures drop off drastically and show steep gradients in the horizontal and vertical dimensions (Fig. 55). Temperatures on the northern boundary of the rift are not as well characterized but are expected to show a more gradual decline across the older, now buried, northern extent of the rift zone's dike complex.

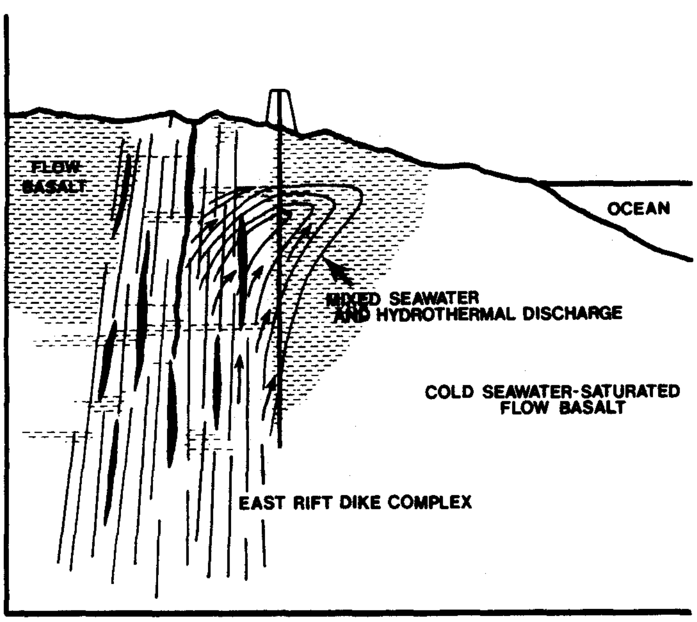

Figure 55. Conceptual model of rift zone hydrothermal system in cross section normal to the strike of the rift.

2. The fluid chemistry data suggest that both meteoric and seawater recharge play significant roles in the chemistry of the east rift hydrothermal system. The dike complex exerts a structural control over permeability that allows meteoric recharge to displace seawater from the interior of the rift zone. However, on the southern boundary, seawater intrusion into the rift may occur along fracture-induced permeability. The hydrothermal chemistry in the seawater-dominated portion of the rift is expected to have a high dissolved solids concentration and, at high seawater:rock ratios, a low pH. Fluids from the freshwater system have a nearly neutral pH and low dissolved solids concentrations. The aggressive nature of the saline fluids may allow rapid alteration of the native reservoir rock and deposition of alteration minerals, resulting in a loss of the fracture-induced permeability in the reservoir. The freshwater-dominated portion of the hydrothermal system may be able to sustain an active high-temperature hydrothermal circulation system, whereas the latter may be able to maintain circulation only at lower temperatures where alteration processes may occur more slowly, allowing fracture-induced permeability to persist for a longer period of time.

OTHER HYDROTHERMAL SYSTEMS IN HAWAII

The extent of the data available for hydrothermal activity associated with other volcanic systems in Hawaii is far more limited than that for Kilauea. Recent geothermal exploration studies (Thomas and others, 1979; Thomas, 1985) have found strong evidence for such activity on the older, more dormant, volcanic systems in the Hawaiian chain; the data upon which these hydrothermal systems have been inferred are presented in Table 11.

The characteristics of these systems are, in many respects, similar to those of Kilauea; however, some important differences exist. One of the more interesting of these differences is found in the hydrothermal system associated with Mauna Loa's summit magma chamber, both of which are entirely above sea level and receive only limited meteoric recharge. As a result, the temperatures of the fluids within this system are not significantly buffered by boiling processes, and hence the temperature of the surface discharges in several locations are in excess of 350°C; this system is believed to be a low-pressure vapor-dominated system. Another important aspect of Mauna Loa's hydrothermal activity is its limited extent: despite the high-temperature summit discharges, there is very little surface evidence of hydrothermal activity associated with the lower elevations of either rift zone. The lack of such evidence may be the result of high rates of meteoric recharge masking deep hydrothermal discharges; however, the short fluid residence times and high rates of heat loss calculated for the Kilauea ERZ (Thomas, 1987) suggest that the rift zone hydrothermal systems may be ephemeral features persisting only as long as frequent intrusive activity occurs on the rift zone.

Hydrothermal discharges from the caldera complexes of the older, now dormant, volcanic systems on Maui and Oahu islands have been tentatively identified on the basis of thermal and geo-

TABLE 11. OTHER HYDROTHERMAL SYSTEMS IN HAWAII

Geologic Structure	Date/Age of Most Recent Activity	Data Upon Which a Hydrothermal System Is Inferred and Probable Inferred Characteristics of System
Kilauea Southwest Rift	Eruption: 1974 Intrusion: 1981-1982	Recent magmatic intrusions; steaming ground and fumaroles; resistivity anomalies. High temperature, possibly seawater-dominated system. More limited in extent than ERZ.
Mauna Loa Summit	Eruption: 1975, 1984 Intrusions: 1980-1984	High temperature fumarolic discharge; aeromagnetic anomaly. High temperature, vapor dominated; driven by convective/conductive heat and gas loss from magma chamber.
Mauna Loa Rift Systems	Eruption: 1975, 1984 Intrusion: 1975	Intermediate temperature fumarole discharges at upper elevations (>2,700 m); aeromagnetic and self-potential anomalies on upper rift. Limited high temperature vapor-dominated system on high-elevation portion of rift; intermediate temperature, "blind" system on lower flanks is possible.
Hualalai	Eruption: 1801 Intrusion: 1929	Hydrothermal alteration at summit; self-potential and resistivity anomalies at summit and on upper west rift system; intense aeromagnetic anomaly at summit. Low to moderate temperature, low pressure water-dominated system driven by summit magma chamber relict heat.
Puu Loa Cinder Cone; Kohala Volcano	Eruption: ~80 ka	Ground-water temperature (33°C) and chemistry anomalies; soil chemistry, and resistivity anomalies. Low-temperature, limited discharge associated with relict heat from post-erosional intrusive event.
Haleakala East and Southwest Rift System	Eruption: 1790; several other isolated vents est. at <1,000 yr old	Resistivity and soil geochemical anomalies. Low to intermediate temperature system; possibly seawater-dominated at depth.
West Maui	Eruption: 20 ka; Shield building terminated at 1.2 Ma	Warm (33°C) ground-water wells; ground-water chemistry and resistivity anomalies. Low to moderate temperature system; driven by relict heat from West Maui caldera complex or from late-stage intrusive activity in southern flank; possibly seawater dominated at depth.
Waianae Caldera	Shield building terminated at 2.4 Ma	Ground-water temperature (27°C) and chemistry anomalies and resistivity anomalies. Low-temperature hydrothermal system driven by relict heat from Waianae caldera complex; probably seawater dominated at depth.

chemical anomalies found in shallow ground-water supplies. These discharges are typified by chemical alterations that are distinctly different from those found in the high-temperature systems on Kilauea. Whereas magnesium and sulfate are typically depleted in the young active systems, low-temperature discharges show substantial enrichments of these species and much stronger enrichments of calcium than are found in the young systems (Cox and others, 1979; Kennedy, 1985). These chemical differences suggest that, as temperatures decline in an aging hydrothermal system, the mineral assemblages formed in its more active phase (e.g., chlorite, smectite, and anhydrite) are "weathered" to low-temperature clays, allowing the remobilization of a significant fraction of the major and trace elements sequestered during high-temperature hydrothermal activity. Thus the life cycle of a Hawaiian hydrothermal system consists of a period of formation of high-grade metamorphic mineral assemblages followed by eventual decay of these species to low-temperature clays as the heat source driving hydrothermal circulation is exhausted.

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MANUSCRIPT ACCEPTED BY THE SOCIETY JUNE 21, 1987

ACKNOWLEDGMENTS

Clague and Dalrymple thank Tom Wright, Jim Natland, Tracy Vallier, and Bob Decker for their review, Kay McDaniel for her patience in typing it all, and Brigitta Fulop for drafting the figures. The section by Wright and Clague has benefited from reviews by P. E. Baker and E. W. Wolfe. Klein and Koyanagi thank Jenny Nakata, Wil Tanigawa, and George Kojima, without whose help the data on Hawaiian earthquakes would not exist; and also Pete Ward, Jim Moore, Dave Harwood, and Pete Palmer for their helpful reviews. Barbara Decker, Don Hussong, Fred Klein, Noye Johnson, and Pete Palmer have thoughtfully reviewed the section by Decker. Thomas thanks Roger Henneberger, Al Truesdell, Chuck Helsley, and Jon-Steinar Gudmundsson for their helpful reviews of his section on hydrothermal systems. Donald M. Thomas: HIG contribution no. 1953.