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Magma Energy Research Project State-Of-The-Project Report – October 1, 1978

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John L. Colp, Richard K. Traeger

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MAGMA ENERGY RESEARCH PROJECT
STATE-OF-THE-PROJECT REPORT - OCTOBER 1, 1978

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ABSTRACT

The Magma Energy Research Project is investigating the scientific feasibility of extracting energy from magma bodies. This report summarizes work done in FY 76, 77, and 78 in the four tasks of the Project:

1. Resource Location and Definition: To evaluate geosensing techniques, a series of electric, electromagnetic, and seismic studies were combined with drilling results to define the horizontal margins and the vertical structure of Kilauea Iki lava lake. The hydrothermal-hot dry rock regions were clearly identified by electrical techniques, but the depth and nature of the liquid-lava lens could not be determined because physical properties of the in situ lens materials are unknown.

2. Source Tapping. To determine if deep drill holes would be structurally stable, strengths of andesite, granodiorite, obsidian, and basalt were measured to 1000°C with confining pressures of 0 and 50 MPa. Preliminary stress analyses indicated 10-km boreholes would be stable if the horizontal stress exceeded the vertical stress. Conventional drilling penetrated the Kilauea Iki crust into molten rock at >1050°C, but a rigid structure in the liquid lens prevented penetration through the lens.

3. Magma Characterization and Materials Compatibility: To define magma environments, thermodynamic codes have been developed to define at depth equilibrium volatile compositions in magma bodies. Volcanic gas analyses have been obtained and chemical reaction effects have been determined. The existing data have then been used to define gas fugacities in deep magma systems for material and compatibility studies. Materials screening tests have been completed in fumaroles and in simulated magma systems. Ni- or Co-rich superalloys show minimum deterioration in magmatic environments.

4. Energy Extraction: Magmatic thermal energy is being considered for the generation of steam or the reaction of water and biomass to generate synthesis gas (CO , H_2 , CH_4). The iron in basalts may also enhance fuel production by reduction of water to hydrogen. Major uncertainties are the mechanisms and rates of heat and mass transport. Calculations of fluid dynamics and heat transfer in magma bodies suggest that energy extraction rates of 1 to 5 kW_T/m^2 are possible. Measurements in the laboratory and in the 1977 Kilauea lava flow have given rates of 1 to 200 kW/m^2 .

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I. INTRODUCTION

The nation's geothermal energy resources exist in a variety of forms: dry steam, hot water, geopressed water, hot/dry rock, thermal gradients in the earth's crust, and magma. The objective of the Magma Energy Research Project now underway at Sandia Laboratories is to investigate the scientific feasibility of extracting energy directly from deeply buried circulating magma sources.

With temperatures on the order of 1000°C, buried magma sources represent great amounts of clean, high-quality energy. The United States Geological Survey (USGS) in Circular 726 (White and Williams, 1975) estimates the energy contained in molten and partially molten magma within 10 km of the surface of the US at 5×10^4 quads. Total energy consumption in the US in 1976 equaled 75 quads. Thus, the estimated energy in the US magma resource is equal to approximately 700 times the nation's annual total energy consumption.

Any program devoted to tapping the energy of a buried magma source is considered long term and high risk because of the technological problems to solve and the many unknowns to resolve. The program at Sandia, under funding by the Division of Basic Energy Sciences of the Department of Energy, emphasizes the investigation of basic scientific and engineering questions of energy extraction from magma; it is not structured toward early on-line power production from these geothermal resources.

The initial proposal by Sandia Laboratories to investigate the extraction of thermal energy from buried molten rock deposits was put forward in mid-1973. From that proposal, investigative studies on a small scale were begun under discretionary funding. This level of effort was continued until mid-1974, when the initial funding for the project (at that time called Magma Tap) was received from AEC (now DOE) Division of Physical Research (now Division of Basic Energy Sciences).

It became apparent during that year that a logical approach would be to divide the project into four tasks.

TABLE I-1

<u>Major Tasks of Magma Energy Research Project</u>	
<u>Task</u>	<u>Current Efforts</u>
1. Source Location of Definition	Cooperate with USGS and universities on magma source surveys and sensing methods development. Lava lake sensing experiment and drilling program.
2. Source Tapping	Texas A&M University borehole stability study.
3. Magma Characterization and Materials Compatibility	Magma geochemistry studies. Material compatibility studies at high pressure and high temperature. Field studies with University of Hawaii. Design/fabrication of magma simulation facility.
4. Energy Extraction	Molten lava/single tube boiler experiment. Long tube heat exchanger experiment. Design of lava lake energy extraction experiment. Magma fluid mechanics, heat transfer, and geochemistry calculations. Flowing lava-heat transfer measurements. Fuels from magma laboratory experiments.

Description of the efforts expended during the initial year and during FY 1975 can be found in Colp et al, 1976.

To provide additional scientific guidance to the project, a Magma Energy Research Advisory Panel was formed in 1974 and has functioned continuously since with few personnel changes and additions. The panel consists of persons from institutions throughout the US, carefully selected for their recognized expertise in relevant fields. The panel's purpose is to review the progress of the project and to provide advice on directions in which it should proceed. Each member also provides advice and assistance in his specialty, as required, to those persons at Sandia responsible for that part of the project. The panel meets as a body about twice yearly. Individual consultations between panel members and project investigators occur more frequently as need arises.

The members of the panel over the years, their affiliations, and their fields of expertise are

Prof. Robert Decker Dartmouth College Hanover, NH	Volcanology/Geophysics
Dr. Gordon Eaton (since 1976) USGS National Center	Volcanology/Geophysics
Dr. James Moore (since 1975) USGS Menlo Park, CA	Magma Petrology
Prof. Melvin Friedman Texas A&M University College Station, TX	Tectonophysics/Rock Deformation
Dr. Peter Ward USGS Menlo Park, CA	Geophysics/Seismic
Prof. John Hermance Brown University Providence, RI	Geophysics/Electrical

Prof. Roger Staehle
Ohio State University,
Columbus, OH

Material Compatibility

Dr. Dallas Peck (until 1975)
USGS
National Center

Magma Petrology

Prof. William Brigham (until 1976) Heat Transfer
Stanford University
Stanford, CA

Other scientists frequently attend the panel meetings and have contributed to the development of this program.

II. TASK 1 -- SOURCE LOCATION AND DEFINITION

Introduction

A major requirement in the development of viable processes to use magma sources for energy applications is the capability to define the existence of a magma body, its depth, its areal extent, and its general form (i.e., whether it is in a finite homogenous pool or in a honeycomb of solid crevices filled with molten material).

A few studies have been conducted to locate and identify a buried magma source, with most of the existing remote sensing methods--seismic, microseismic, resistivity, gravity, magnetic, magnetotelluric, infrared, thermal--having been used (Eaton et al, 1975). The USGS has examined areas under Yellowstone National Park; Long Valley, CA; Kilauea Volcano, HI; Geysers/Clear Lake, CA; and at other locations in the Western US. While these studies have provided some evidence of magma sources, there are sufficient uncertainties associated with each of the remote sensing methods that conclusive proof of magma location and other characteristics has not yet been obtained.

Recognizing this state of uncertainty, the Magma Workshop (Colp et al, 1975) recommended that various geophysical techniques be tested and confirmed on a known buried molten rock body so that the effectiveness of the different methods could be determined. A parallel recommendation was to use geologic information from eroded and exhumed fossil magma systems to aid in characterizing active magma systems.

These recommendations have resulted in a comprehensive investigation of the configuration and physical characteristics of a known, small-scale

molten lava system in Kilauea Iki Crater (Richter & Moore, 1966) in Hawaii. The Magma Energy Research Project supported, in collaboration with the USGS, a series of geophysical-sensing experiments during 1976 on Kilauea Iki. Also in 1976, the first attempt to obtain confirmatory information by drilling into and through the lava lake was attempted. While the primary objective of this drilling experiment was not achieved, it established the basis for the second drilling experiment to be conducted during FY 79. In addition to the geophysical and drilling experiments, the studies of Moore and Evans (1967) on the exposed prehistoric lava lake at the Makaopuhi pit crater have also been used to hypothesize the characteristics of Kilauea Iki.

With respect to establishing the configurational characteristics of the Kilauea Iki lava lake via geophysical techniques, the following constraints were known:

- The total thickness of the lava fill in the crater, both chilled and molten, was estimated at 115 to 120 m (Holcum, 1976) or 111 m (Richter and Moore, 1966)
- 1976 contemporary drilling indicated that the melt zone was reached at 45 m
- Thermal history calculations and observations on the prehistoric Makaopuhi lava lake indicated that the slowest rate of cooling and maximum temperatures were at a depth of approximately 55% to 60% of its total thickness (Evans & Moore, 1968)
- Temperature logs in drill-holes revealed isothermal temperatures at 100°C from 10 to 33 m, at which point a rapid increase in temperature occurred until melt (probably 1070° +5°C) was encountered at 45 m.

Lava Lake Geophysical Sensing Experiments

In early 1976, a series of geophysical sensing experiments was initiated on Kilauea Iki lava lake, with the following objectives in mind:

- Monitor the geophysical properties of lava in-situ.
- Map the thickness and edge of the lava lens as a prelude to an extensive drilling program
- Assess the feasibility of using geophysical methods for detecting and delineating major magma structures in the earth's crust (2-10 km).

Seismic and electromagnetic experiments (Table II-1) were carried out by personnel from the USGS, MIT, the University of Texas at Austin, and Sandia Laboratories. Much of the following description of the results from the Lava Lake Geophysical Sensing Experiments was extracted from the summary report by Hermance, Forsyth, and Colp (1979).

Active Electromagnetic Experiments

The interpretation of the active electromagnetic experiments has led to quantitative constraints on the depth, thickness, and lateral dimensions of the lava lens.

Defining the Vertical Structure--Two classes of electrical experiments were carried out to study the vertical structure of the lava lens. Zablocki of the USGS conducted Schlumberger (galvanic four-electrode electrical) soundings along two perpendicular lines in Kilauea Iki, North-South and East-West. The USGS also performed two-loop electromagnetic soundings.

1. The Schlumberger (Galvanic) Soundings. The Schlumberger data shown in Figure II-1 has contributed significantly to defining the vertical structure of the chilled resistive layers above the molten lava lens. This data indicated that the surface layers have the thicknesses and electrical resistivities indicated in Table II-2.

TABLE II-1

Geophysical Experiments Performed on
Kilauea Iki Lava Lake

<u>Experiment</u>	<u>Performed By</u>
Dipole-Dipole Electromagnetic Induction Experiments	Frischknecht, Zablocki, B. Smith, Flanigan--USGS
Very Low Frequency Radio Wave Methods	Zablocki, Anderson--USGS
Galvanic Four-Electrode Resistivity Soundings	Zablocki, B. Smith--USGS
Audio-Frequency Magnetotelluric Profiling and Sounding	Bostick, S. Smith, Boehl-- UT at Austin
Passive Monitoring of Seismic Noise	Aki, Chouet, Colp-- MIT/Sandia
Teleseismic Measurements	Aki, Chouet--MIT
Small-Scale Seismic Refraction Experiments Using Artificial Charges	Aki, Chouet--MIT

TABLE II-2

Interpretation of Schlumberger Data
(Zablocki, 1976)

<u>Layer Number</u>	<u>Depth to Top (metres)</u>	<u>Resistivity (Ω/m)</u>
1	0	25K
2	1.7	5K
3	5.4	600
4	33	(40K?)
5	44?	(3?)

Figure II-1. Kilauea Iki VES. North-south (NS) and east-west (EW) reduced VES data (dots) and theoretically computed apparent resistivities (solid line). (After Smith et al, 1977, Figure 5.)

The high resistivity of layer 1 is undoubtedly due to a low concentration of pore fluids (meteoric water) in a very porous and heavily fractured material. The somewhat lower resistivity of layers 2 and 3 may reflect a higher pore-fluid concentration, along with an increase in the amount of dissolved salts in the fluid due to its recycling in a fairly active convection zone.

The value of resistivity for layer 4 is undoubtedly large, and 40 $\text{k}\Omega/\text{m}$ could easily be a minimum value. Zablocki (1976) has stressed that the thickness of layer 4 and the resistivity of layer 5 as indicated in Table II-2 are only possible interpretations; these parameters are not well constrained by the Schlumberger data alone. On the other hand, the two-loop electromagnetic induction measurements as described in the next section significantly constrain the interpretation of these deeper layers.

2. The Electromagnetic (Two-Loop) Soundings. Two-loop electromagnetic soundings over the approximate center of the lava lens at two fixed transmitter-receiver spacings, 108 and 200 m, were conducted, and the results are summarized in Table II-3 (Smith et al, 1977). Several layer parameters were constrained in the least-squares inversion of the electromagnetic data: the resistivity and thickness of the first layer (reasonably well known from the Schlumberger measurements), the thickness of the second layer (assumed to be a warm wet layer), and the resistivity of the third layer.

TABLE II-3

Combined Interpretation of Geoelectric Section
Using Schlumberger and Loop-Loop Electromagnetic Sounding
(Smith et al, 1977)

<u>Layer Number</u>	<u>Depth to Top (metres)</u>	<u>Resistivity (Ω/m)</u>
1	0	25K
2	4.5	300-750
3	28	>10,000
4	32.3-44.4	2.3
5	49.8-60.1	>10,000

The resistivity of the melt layer (layer 4) was determined from a previous loop-loop electromagnetic sounding performed in 1962, when the crust of the lake was only 13 m thick. Since it is not clear whether the melt resistivity changed appreciably over the ensuing 14 yr, 2.3 Ω /m may represent a lower limit on the present-day resistivity.

The electromagnetic soundings cannot resolve the resistivity and the thickness of the lava lens separately; however, the conductivity-thickness product (the conductance) of the melt layer is well-determined, and has a

value of 7.64 mhos from the 108-m sounding. The value of 6.86 mhos for the conductance of the lava lens from the 200-m sounding may be biased toward somewhat lower values by either the finite geometry of the lava lens or by lateral changes in the lava lens itself (Smith et al, 1977).

Although the USGS interpretation advocates a thickness of 17.5 m for the lava lens, this is true only if the resistivity of the melt has not changed since it was determined originally in 1962.

A schematic of a conceptual model proposed by Smith et al to account for the essential elements of the Schlumberger and two-loop electromagnetic sounding experiments is presented in Figure II-2. The Schlumberger data have quite adequately defined the presence of a relatively dry surface layer; a warm, wet conductive layer (300 to 700 Ω/m); and a dry, resistive layer (on the order of 40 k Ω/m) in which temperatures are sufficiently high to drive off significant amounts of water vapor. The two-loop electromagnetic measurements have also defined conductive melt, 17.5 m thick, with a resistivity of 2.3 Ω/m and a dry resistive layer below.

Figure II-2. Conceptual Model of Kilauea Iki Lava Lake (after Smith et al, 1977, Fig. 4)

In order to reconcile the electromagnetic interpretation with the drilling results and thermal modeling estimates, Smith et al presented the geological model shown in Figure II-3. The top of the principal molten-lava lens is pegged to the depth that drilling encountered "melt." The fact that the Sandia Laboratory hole KI 76-I penetrated only 0.5 m before encountering an obstacle is accounted for by floating, more refractory regions in an otherwise liquid lava lens 17 to 18 m thick.

Figure II-3. Complex Geological Model for Kilauea Iki Lava Lake Showing Stratification in the Melt Zone. The lower shaded zone is less conductive than the upper (not shaded) zone. (After Smith et al, 1977 Figure 14.)

Defining the Lateral Boundaries--Flanigan and Zablocki (1977) have used data from a Turam experiment on the lava lake to define its lateral boundaries. A large circular loop of wire placed on the surface of the lake approximately over the center served as the transmitter source; the complex horizontal gradient of the vertical magnetic field component was

measured as a function of distance from its center. The field data was then compared with a small analog scale model in the laboratory. Melt conductivity and thickness were inferred from the electromagnetic soundings just described. The depth to the melt was constrained on the basis of the 1976 drilling results. Although the possibility exists that any inference regarding the location of the edge of the melt zone from the Turam interpretation is inbred with inferences regarding its depth and conductance as determined by other observations, these uncertainties contribute second order effects, at most, for the aspect ratio of depth to radius of the structure considered here.

Three frequencies (200, 400, 800 Hz) were used in the actual field experiment, although results from only 400 Hz were quantitatively interpreted. The variation between the results from each profile when contoured on a plan view of the lava lake suggest that the target is somewhat more complicated than a homogeneous lens implanted in a homogeneous half-space. An intercomparison of the normalized field ratios among the various profiles shows a radial asymmetry of 10%. This asymmetry could be caused by difference in the conductivity thickness product (layer conductance), depth to the molten basalt, or the asymmetric position of the transmitter coil relative to the edges of the molten lava lens.

Although a relatively broad spread is evident when data from all profiles are compared (Figure II-4), an interpretational device to minimize the effects of such a large spread between curves was used. The data along a single radial direction from the transmitter were used to determine the radius of the best fitting disk for that single profile (Figure II-5). Although data from the next radial give a slightly different radius for the disk, this difference is treated as an actual perturbation of the edge of the conducting lava lens, rather than an intrinsic error in the data. In this way an outline of the lava lens (Figure II-6) which is somewhat distorted from a simple disk can be constructed.

Figure II-4. Real Component of the Electromagnetic Field, Kilauea Iki Lava Lake. The ordinate is in percent of field from an arbitrary point. The abscissa is the ratio of A/R where A is the distance from the center of the transmitting loop to the center of the two receiver loops and R is the radius of the transmitting loop. Shaded region shows the range in the field curves obtained along all 9 traverses. (After Flanigan and Zablocki, 1977, Figure 3.)

Figure II-5. Real Component of a Model Curve (circles) and the Real Component of the East Field Traverse. The edge of the conductive model represents a field distance of 320 m from the center of the transmitting loop. (After Flanigan and Zablocki, 1977, Figure 7.)

Figure II-6. Proposed Lateral Boundary of Molten Lens in Kilauea Iki Lava Lake

The Very Low Frequency (VLF) Experiment

Preliminary magnetotelluric profiling at VLF, 23.4 kHz (USN Sta. NLK, azimuth N34°E) and 18.6 kHz (USM Sta. NPM, azimuth N53°W), presented by Flanigan and Zablocki (1977), shows dramatic increases in apparent resistivity near the edges of the melt zone (Figure II-7). Apparent resistivities are quite low over the lens itself but, as one gets close to the edge, a high-resistivity zone is encountered first, then a more conductive zone, and finally another high-resistivity zone. These variations are most prominent for the apparent resistivity determined from the electric field polarized perpendicular to the edge, the so-called "perpendicular resistivity."

Figure II-7. Apparent Resistivity Profiles From Two VLF Stations Obtained Along Two Lava Lake Traverses. Circled-numbers show locations of conductor edge (data from L. A. Anderson, written communication, 1976). (After Flanigan and Zablocki, 1977, Figure 8.)

The variations are less pronounced for the "parallel resistivity," which is the apparent resistivity determined from the electric field parallel to the signal direction, but nevertheless, the results are consistent. At the frequencies used here and their associated short attenuation lengths, the parallel apparent resistivity is probably more representative of the true value of resistivity at depth than is the perpendicular apparent resistivity.

The smooth flat behavior of apparent resistivities over the interior of the lava lake indicates that one-dimensional models fitting the VLF apparent-resistivity data over the center of the lake require the presence of a low resistivity layer ($120\Omega/\text{m}$) above the molten lens, underlain by a zone of much lower resistivity, presumably the melt zone. Apparently at these frequencies the electromagnetic fields are absorbed in a more resistive halo above the molten lens and are responding only indirectly to the presence of the molten lens itself.

An interesting feature of the VLF apparent-resistivity profiles is the fine structure associated with the edge of the lens. A persuasive argument has been advanced that this fine structure is associated with the presence of a shallow convective hydrothermal cell driven by the thermal energy released by the cooling lava (Figure II-8). Moreover, Flanigan and Zablocki point out that the flat trend of apparent resistivities over the lake continues beyond what was previously thought to be the approximate edge of the lake.

Figure II-8. Conceptual Model of the North Edge of the Lava Lake. Shows the possible configuration of the melt lens, the hydrothermal system, and temperature distribution within the lake that might account for the generalized resistivity profile shown on top. (After Flanigan and Zablocki, 1977, Figure 9.)

Audio-Magnetotelluric Measurements

The Electrical Geophysics Laboratory of the University of Texas, Austin, performed audio-magnetotelluric measurements at 20 sites in Kilauea Iki Crater, as well as other sites over the summit of Kilauea Volcano (Bostick et al, 1977).

Bostick et al calculated, as part of a prefield planning exercise, the plane-layered (one-dimensional) magnetotelluric response for the layer parameters given in Table II-4. These estimated values were based on electromagnetic work as well as on thermal-modeling calculations performed previously by USGS workers.

TABLE II-4

Plane-Layered Model to Simulate Lava Lake
(Bostick, Smith, and Boehl, 1977)

<u>Layer Number</u>	<u>Depth to Top (metres)</u>	<u>True Resistivity (Ω/m)</u>
1	0	1000
2	46	2
3	69	1000

The observed data, shown as the points in Figure II-9, looked very little like the expected plane-layered response calculated prior to the field program. The most striking difference is that the parabolic-shaped minimum in the theoretical curve by the lava lens is clearly absent in the field data. This apparently is due to the effects of the finite size of the lava lens. Based on magnetotelluric data only from sites over the middle of the lava lake, one would not have been led to assert, based on experience with plane-layered interpretations, that a lens-like conductor was present. However, since both 1976 drilling and the active

electromagnetic-sounding experiments indicated molten material at a depth of only 40 to 45 m, the lava lens was believed to exist, even though not evident in a simple plane-layered interpretation.

Figure II-9. Observed Apparent Resistivities Compared With the Theoretical Response of a Plane-Layered Model With the Characteristics Given in Table II-4.

In order to model the possible effect of the finite geometry of the lava lens, it is assumed that the lowest frequency (17 Hz) is of a sufficiently long period that induction effects can be neglected. The response of the lava lens can then be modeled as simply the distortion of a uniform dc-telluric field by an imbedded conducting inhomogeneity. A series of straightforward, though approximate, corrections can be made to account for the fact that the lava lens is beneath the floor of a crater-type structure. The final model evolved is shown in Figure II-10.

Figure II-10. Assumed Form of the Melt Zone in Kilauea Iki Lava Lake. (After Bostick et al, 1977, Figure 10.)

The smooth curves in Figure II-11 represent theoretical magnetotelluric apparent resistivities "normalized" to values away from the crater disturbance, for various thicknesses of the lava lens in metres. The apparent resistivities are calculated for the electric field perpendicular to the edge of the melt and parallel to the edge of the melt. The curves for a given melt thickness show the greatest separation near the edge of the melt zone, whereas over the center of the lens, the theoretical curves from the two polarizations should come together for each model. In fact, this does not seem to happen in the data shown as the points in Figure II-11, where the true distance of each site from 17N2 is normalized by the assumed radius of the lava lens (275 m). Clearly certain qualities of the theoretical model data are reflected in the experimental data; for example, the increased separation between the parallel and perpendicular resistivities with increasing distance. Nevertheless, there can be little doubt, from inspecting Figure II-11, that the lava lake is more complicated than the simple model used in the interpretation.

Figure II-11. Experimental Data Compared With Calculated Apparent Resistivity Curves for Different Assumed Values of Melt Thickness. (After Bostick et al, 1977, Figure 15.)

Bostick et al noted that the observed minimum in apparent resistivity over the center of the lava lake (site 17N2) is compatible with a lens thickness of approximately 4.6 m, which gives a conductance for the lens of 2 mhos; this is much less than the 6.9 to 7.6 mhos determined by Smith et al (1977) using the two-loop electromagnetic-sounding method. To explain this discrepancy, Bostick et al have invoked the possible effect of a resistive "steam" sheath surrounding the lava lens. Zablocki (1976) apparently detected such a zone directly above the lava lens by using Schlumberger soundings. If such a zone completely surrounds the lens, it undoubtedly represents a very high resistance in series with the conducting lens. The net effect would be to decrease the apparent conductance

that one would infer from surface telluric measurements, perhaps enough to reduce the apparent conductance from 7 mhos to 2 mhos (Bostick et al, 1977).

This in itself is an important observation; it appears the telluric fields are conveying a great deal of information about the edges of the lava lens as well as about the interface between the molten lens and the surrounding medium. Hence, the telluric field measurements may have provided a significant insight into aspects of the problem that have a minimal effect on loop-loop E-M soundings. On the other hand, the loop-loop measurements contributed an estimate of the lens conductance that appears far less contaminated by "edge effects." In this sense, the telluric measurements and the loop-loop measurements complemented each other, each providing information that presents a clearer picture of the physical processes involved.

Seismic Measurements

Introduction--A preliminary experiment on the passive seismic activity of Kilauea Iki lava lake was made by B. Chouet (MIT) in August 1975. This experiment revealed a high activity of seismic events occurring at this location. A seismograph located on the surface of the lake near its center recorded about 100 events/h.

In March of 1976, Aki and others from MIT, together with equipment and personnel from Sandia, conducted both active and passive experiments using 13 seismic channels with various combinations of vertical and horizontal seismographs. In the active experiment, in addition to the usual P wave refraction survey, the propagation of SH and Love waves was studied. In the passive experiments, not only were the events occurring within the Kilauea Iki crater studied, but also seismic signals propagated through the molten rock lens from earthquakes outside the location (Aki et al, 1978).

In September 1976, Colp and Moritz of Sandia made a more detailed study of the passive seismic events occurring in Kilauea Iki. These

observations clearly reinforced the previous data from Chouet and Aki et al, but with an appreciably finer mesh of stations than was used in the earlier studies.

Active Experiments (MIT)--As soon as the crust of Kilauea Iki was solidified following the 1959 eruption, nails were planted at regular intervals for a precise leveling survey. These nails were used as the geophone sites. The nails planted along the major axis of the crater at intervals of 30.5 m were numbered 1 through 40. For refraction experiments, geophones were placed at nails 3 through 39 (as is shown in Figure II-12). The leveling stations 17N7-17N12 at intervals of 30.5 m along the minor axis of the crater were also occupied.

Figure II-12. Locations of Geophone Sites for Active Experiments. Shot points are marked by crosses. (After Aki et al, 1978)

The geophones (HS-1, 4.5 Hz) were coupled to the ground with construction plaster. The signals from the geophones were preamplified and used to modulate constant bandwidth voltage-controlled oscillators for transmission, via cable, to a recording station set up in a building near the rim of the crater. Two channels of FM multiplexed data representing 13 seismic signals were directly recorded on magnetic tape. Following demodulation, the seismic data were also displayed on a high-speed chart recorder.

Twenty shot holes of 3.2-cm diameter were drilled to a depth of 2 m at the locations marked by crosses in Figure II-12; 0.23 to 1.6 kg (1/2 to 3-1/2 lbs) of dynamite was used in each hole, depending on the distance between the shot point and the geophone array. For each shot, a geophone was placed near the shot hole and its signal was used to define the origin time.

Interpretation of P Wave Data. The seismograms obtained for shots E2 and E3 are arranged according to the epicentral distance in Figure II-13; typical features of refraction records obtained on the floor of Kilauea Iki are shown. The first P arrival attenuates very strongly with distance. The velocity of this arrival is low, ranging from 0.2 to 2.2 km/s. The attenuation is stronger along the path with lower velocity. At distances greater than about 400 m, prominent later phases are observed that show the apparent velocity close to the velocity of the first P arrival. The particle motion observed by the two-component array along the minor axis indicates that they are P waves. The first arrival shall be referred to as P_0 , and later arrivals as P_1 , P_2 , etc.

Later-arriving P waves with apparent velocities nearly equal to surface velocity may be explained by the presence of a low-velocity layer. For simplicity, one can assume a doubly-layered half space with the velocity of the top layer and half space v_0 , and the velocity of the intermediate low-velocity layer v_1 ($< v_0$). Then one can consider the direct path P_0 in the layer and the path P_1 as reflected from the top of the half space (inset of Figure II-14). For a large horizontal distance, it can be shown that the travel time difference $\Delta\tau$ between P_0 and P_1 does not depend on the thickness of the top layer, and can be approximated very well by

$$\Delta\tau = 2 \left(d v_1^{-2} - v_0^{-2} \right)^{1/2} \quad (1)$$

where d is the thickness of the intermediate layer.

Figure II-13. Records of Vertical Geophones Along the Major Axis of the Lake From Shots at the Eastern End. (After Aki et al, 1978)

Figure II-14. Theoretical Seismograms of Vertical Displacements Observed at A and B Due to a Line Explosive Source at O. The structure and the ray paths associated with P_0 and P_1 arrivals marked on theoretical seismograms are shown at the top. (After Aki et al, 1978)

Equation (1) alone cannot uniquely define the velocity and thickness of the low-velocity layer. However, there are some constraints on the thickness from other sources. The thickness of lava in Kilauea Iki at the end of the eruption in 1959 was 111 m at the deepest point. Since then the floor has subsided by 2 m. Drilling in March 1975 revealed that the thickness of the upper crust was 44 m. This gives the maximum possible thickness of melt as about 65 m. Crust is also being formed from the bottom, as is demonstrated in an exposed section of the prehistoric Makaopuhi lava lake (Moore and Evans, 1967). Thermal history calculations, based upon the assumption of conductive heat transfer at the bottom, predict that the lower crust is thinner than the upper crust (Shaw et al, 1977; Peck et al, 1977). The minimum thickness of melt may be obtained as 21 m by assuming the same thickness for upper and lower crust. From Eq (1) we find the range of velocities for the low-velocity layer corresponding to the above range of thickness estimated as 0.2 to 0.6 km/s.

Interpretation of Love Wave Data. In both Figures II-15 and -16, the Love wave-phase velocities change abruptly when the waves pass the neighborhood of nail 22. The velocity is lower on the east side of nail 22 than on the west side, independent of the direction of wave approach.

The P velocity of the upper crust did not show any change across nail 22 (Figure II-17); therefore, the source of change in Love wave velocity must be deeper. As will be shown later, the number of passive seismic events drastically changes across nail 22. The activity is 10 times higher on the east side of nail 22 than on the west side. Nail 22 is also located close to a zone of long, sublimate-encrusted joint cracks (as described by Peck and Minakami, 1968). These data seem to suggest that nail 22 marks the edge of the magma lens. The Love wave-phase velocity is low on the east side because of the magma lens. The thermal cracks are generated there because of continued cooling and contraction of crust above the magma. The zone of joint cracks might have accelerated cooling to freeze the part of the lava lake on the west side of nail 22.

A trial and error method was used to determine model structures that would give Love wave-phase velocities agreeing with those observed. For the presumably frozen part of the lava pond, a single layer over a half space, as depicted in the upper left of Figure II-18, gives a satisfactory fit with observations. The model consists of a 20-m-thick layer of very low velocity (0.55 km/s), overlying a half space with a velocity of 0.95 km/s. The low velocity at the top may be due to the abundance of cracks which at greater depths will be healed by higher pressure and temperature. Models for the part underlain by the presumed molten lava lens are determined by inserting low-velocity channels (to correspond to the melt) of various cross sections into the model obtained for the frozen part.

If the low-velocity channel (melt lens) is thick, Model B (depicted in the upper right of Figure II-18) matches the observations. However, for a thin, low-velocity channel, Model C best fits the data. Thus, it is not possible to determine the unique velocity and thickness of the low-velocity layer. Its shear velocity may be as low as 0.2 km/s if its thickness is 10 m, and as high as 0.4 km/s if the thickness is 50 m.

Figure II-15. Records of Love Waves by Transverse Horizontal Geophones Along the Major Axis of the Lake From a Shot at the Eastern End. Notice that the wave velocity becomes higher after crossing nail 22. (After Aki et al, 1978)

Figure II-16. Records of Love Waves by Transverse Horizontal Geophones Along the Major Axis of the Lake From a Shot at the Western End. The wave velocity becomes lower after crossing nail 22. (After Aki et al, 1978)

Figure II-17. Velocity of First Arriving Waves (P_0) at Short Distances Plotted on the Measured Section (After Aki et al, 1978)

Figure II-18. Observed and Theoretical Phase Velocities of Love Waves. The structures corresponding to theoretical curves are shown in the top marked with the same letters. The phase velocities for structures A and C are the fundamental modes, but those for structure B are the first higher mode. (After Aki et al, 1978)

Passive Experiments (MIT)--To determine the depth and location of foci of seismic events occurring within the Kilauea Iki Crater, two different methods were used. First, a seismic array with various configurations and combinations of sensor orientations (Figure II-19) was deployed. Second, a visible drum-recording (Sprengnether Model MEQ800) seismograph was placed at 17 locations marked in Figure II-20 for a few hours at each location. The number of events per half hour, with peak-to-peak amplitude greater than 2 mm on the record, is noted at each site on Figure II-20.

Figure II-19. Locations of Seismograph Sites for Passive Experiments (After Aki et al, 1978)

Figure II-20. The Number of Seismic Events With Peak-to-Peak Amplitude Greater Than 2 mm on the Record of a Portable Seismograph (maximum magnification 1.6×10^4 at 50 Hz) is Counted at Different Sites, and the Number Per Hour is plotted at Each Site. The frequency drops abruptly as nails 17N10 and 22 are crossed toward the edge of the lake. (After Aki et al, 1978)

The microseismic events shown in Figure II-20 may originate from thermal cracks formed during the cooling and solidification of the crust around the molten lens. The sharp drops in frequency of events shown in Figure II-20 as nails 17N10 and 22 are crossed imply the edge of the molten lens.

Interpretation of S Waves. Array 1 (Figure II-19) was operated specifically to find the diffraction effect of the edge of the magma lens on P waves arriving from earthquakes outside Kilauea Iki. The array location spread on both sides of the leveling station 17N10, beneath which the

magma edge was suspected from the spatial distribution of frequency of local seismic events. Other arrays also recorded earthquakes outside Kilauea Iki. Epicenter locations of five selected earthquakes were determined by using the USGS seismic network on the island of Hawaii. These are either volcanic earthquakes associated with Kilauea, or after-shocks of the Kalapana earthquake. The former arrived at Kilauea Iki from SSW, and the latter from ESE (Figure II-19).

The records of an earthquake ($M \approx 1.7$) were adequately recorded at four stations along the minor axis and are shown in Figure II-21. In agreement with the array result, the observed small ratio of north-south to vertical-component P wave amplitudes indicates that the wave path approached transverse to the minor axis. The large arrival on the north-south component observed at about 2.4 s after the P waves must be SH waves. The S-P time of 2.4 s is appropriate for the epicentral distance of 16 km. Nearly simultaneous arrivals at all the stations located along the minor axis were observed. Therefore, since they are SH waves, they cannot be attributed to converted waves that went through the molten rock lens as P waves. Thus, it must be concluded that S waves are propagated through the presumed molten rock lens.

Figure II-21. Records of SH Waves by Vertical and Horizontal Geophones Along Array 4 in the Middle of the Lake Due to an Earthquake Outside Iki. Large-amplitude SH waves with a period of 0.11 s are clearly recorded. (After Aki et al, 1978)

A study of the transmission of S waves through a viscous liquid was made to help explain this surprising result. Figure II-22 shows a simple contour plot of transmission coefficient as a function of viscosity and liquid layer thickness. It also shows that the transmission through a liquid layer with thicknesses as small as a few metres requires a viscosity of around 10^7 P.

Figure II-22. Relation Between Viscosity (η) and Thickness of Layer H for Selected Values of $|T|$. (After Aki et al, 1978)

As mentioned earlier, the thickness of upper crust was determined by drilling, and the thermal history calculation (based on the assumption of conduction at the bottom) suggests a magma lens thicker than 10 m. On the other hand, the value of 10^7 P for viscosity is too high for basaltic magma, for which most of the reported values lie in the range 10^3 - 10^5 P (MacDonald 1972). Here the Bingham body model of basaltic magma proposed by Shaw et al (1968) aids in understanding the dilemma. Shaw et al made an in-situ measurement of viscosity of the magma in Makaopuhi Lava Lake, Hawaii, using a stainless-steel spindle inserted through the crust. They found that magma behaves as a liquid, with Newtonian viscosity of 10^3 - 10^4 P, only under stresses above a certain yield stress. The measured yield

stress was 700 dyn cm^{-2} at one depth and 1200 at another. The stress level of the seismic signal is of that order or less.

From observations of Love wave dispersion, it was found that if the low-velocity channel is as thick as 50 m, the shear velocity is about 0.4 km/s, from which an apparent viscosity of $4 \times 10^7 \text{ P}$ can be calculated. Then, from Figure II-22, it can be seen that the thickness of the liquid layer probably does not exceed 10 m.

Passive Experiments (Sandia)--In the late summer of 1976, following the passive seismic surveys described above, Colp and Moritz made a more extended investigation of the passive seismic activities in Kilauea Iki Crater. An easily portable array of six geophones, spaced 15.25 m apart, was assembled. The array was wired directly to signal-conditioning equipment, and the data were recorded on both a magnetic tape and a paper chart. The geophones were attached to the lake surface by construction plaster at 17 array locations in the northeast quadrant of Kilauea Iki Crater. Recordings of the seismic events at a total of 102 geophone locations were made for a minimum of 30 min each.

Preliminary analysis of the data gathered is shown in Figure II-23. As can be seen, the demarcation between high- and low-seismic activity can be clearly seen between geophone stations located only 15.25 m apart. Assuming that the high rates of seismic activity are associated with thermal-fracturing events in the crust overlying the molten lens, while low rates are associated with locations not over the lens, we can draw an indicated edge of the molten lens (Figure II-23).

Figure II-23. Preliminary Analysis of the Results of a Passive Seismic Survey of the Northeast Quadrant of Kilauea Iki Crater. (Numbers are microearthquakes per 30 min.)

A Proposed Model for the Lava Lake

The Vertical Structure of the Lava Lake

A model representing the present configuration and structure of Kilauea Iki Lava Lake must embody the four constraints mentioned in the Introduction. The active electromagnetic experiments provide added details to the vertical structure:

- Schlumberger electrical-sounding data define a very resistive surface layer, 5 m thick; a wet, less resistive, warm layer at 5 m extending to approximately 30 m; and underlain by a very resistive layer representing the top of the transition zone.
- Two-loop electromagnetic-sounding data reveal a layer at a depth of 38 to 44 m with a total conductance of 6.68 mhos.

Drilling results in 1976 defined the minimum upper crust thickness as 45.5 m.

A proposed model meeting these constraints is shown in Figure II-24. The gravitational settling of olivine crystals and the upward migration of residual fluid eventually forming glass are useful indicators of rheological and temperature conditions in the melt at intermediate times in the cooling history of the lava layer. Smoothed versions of the concentrations of olivine phenocrysts and residual glasses as observed on the prehistoric Makaopuhi Lava Lake (Moore and Evans, 1967) are also shown in Figure II-24. Since the prehistoric lava lake had a depth of 67 m and the present Kilauea Iki lava lake has a depth of 115 to 120 m, the depth scale of Moore and Evans has been simply expanded by a linear factor of 115/67.

The model in Figure II-24 has seven layers:

1. A dry, resistive surface layer, from the surface to approximately 5 m.
2. A warm, wet layer at 100°C, containing appreciable water and/or steam, from 5 m to a depth of 33 m.

3. A transition layer, from 33 to 45 m, over which temperatures increase from 100° to 1070°C. The top of this layer will appear resistive as water is flashed as superheated steam; the bottom will appear conductive as temperatures become sufficiently high to activate appreciable conduction in the rock itself.
4. A thin zone, up to 5 m thick, representing a plexus of molten sills in solid layers.
5. A layer with the highest temperature, lowest viscosity, and lowest density of olivine phenocrysts, from 49 $\pm$ 2 m to approximately 70 m (modified from Smith et al, 1977).
6. A smooth transition zone, from a low viscosity melt at about 70 m to a crystalline mush, and then into a cooler melt with a lower density of olivine phenocrysts. The uncertainty in the depth to the bottom of this layer (87 $\pm$ 7 m) is shown by the error bar.
7. Solid basalt with a temperature which, analogous to the calculation of Shaw et al (1977), gradually decreases from 1070°C at the top of the layer (80 to 95 m) to 700° or 750°C at the bottom of the layer (115 to 120 m).

Figure II-24. Proposed Model of Vertical Section Through Kilauea Iki Lava Lake.

The Lateral Boundaries of the Lava Lens

Both active electromagnetic experiments and the passive monitoring of local seismic events have served to define the lateral boundaries of the lava lens. In addition to those experiments previously discussed, preliminary VLF data furnished by O. Anderson (1976) has proven useful in this phase of the study.

A map (Figure II-6) showing the boundary of the lava lens as determined by all of these workers shows excellent agreement between the various methods and indicates that uncertainty regarding the precise location of the edge of the lens lies within a band from 25 to 50 m wide. The minimum diameter of the lava lens is 500 m north-south and the maximum diameter is 750 m east-west.

Speculation on the Thickest Portion of the Lava Lens

The geophysical sensing experiments to date have only obliquely suggested lateral variations in either the material characteristics or the thickness of the lava lens. At present it is felt that the best information on where the lava lens is thickest is provided by (a) the maximum total thickness of the entire lava lake, as derived from subtracting preflow topography from present topography, and (b) the area of maximum subsidence of the surface due to thermal contraction and crystallization of the lava upon cooling.

Figure II-25 shows a composite contour map in which the total lake thickness and the surface subsidence data of Holcum are compared. From this comparison, one would conjecture that the melt would be a relatively uniform maximum thickness over an area 300 m wide and 500 m long, centered approximately on nail 17N2.

Figure II-25. Composite Contour Map of Kilauea Iki Lava Lake Showing Total Thickness and Subsidence. (After Holcum, 1976.)

III. TASK 2 -- SOURCE TAPPING

Introduction

A problem whose solution is vital to the use of magma energy is that of reaching the source and extracting its energy. Magma sources may be >10 km below the surface of the earth in many locations. Conventional drilling technology at this time is limited to depths of about 10 km (30 000 ft). It has been difficult to do deep drilling in areas suspected of being near magma sources because of the increasing temperature (Keller, 1974). This project was not originally intended to be directly involved in developing deep-drilling methods in the high-temperature magma environment.

Task 2 of the original project plan included only a study of the deformation of the heated rocks overlying a magma source and an assessment of the stability of the borehole. In 1976, the performance of the Lava Lake Sensing experiment led to the drilling of two holes in Kilauea Iki to gain information on the existence and form of the molten lens. This initial drilling work was a part of Task 1; however, drilling technology developed for the 1000°C molten rock body will provide a base for future engineering drilling programs. Therefore, the Lava Lake drilling effort is included as part of Task 2.

Other organizations in this country are considering deep-hole drilling. The drilling subcommittee of the US Geodynamics Committee has long recognized the desirability of developing deep-drilling techniques. Under their sponsorship, a workshop on Continental Drilling was convened in June 1974 (Shoemaker, 1975) and summarized objectives of a continental drilling program for scientific purposes. A panel on continental drilling (D. L. Peck, Chairman) issued a report (Peck et al, 1977) recommending that a 5 to 10 yr Federal Continental Drilling Program be implemented.

In July 1978, the US Geodynamics Committee of the National Academy of Science sponsored a Workshop on Continental Drilling for Scientific

Purposes at Los Alamos, NM. One of the panel groups in this workshop, under the direction of Robert Decker (Dartmouth College), addressed the subject of "Thermal Regimes of the Crust, Particularly Those Related to Hydrothermal and Magmatic Systems" (National Academy of Science, 1979).

Sandia Laboratories has had a DOE-funded program in drilling research and development for several years that is separate from the Magma Energy Research Project. The objective of this program is to devise methods and equipment for improving the speed and depth of drilling, particularly in geothermal environments. Although this program is not concerned at present with drilling development at near-magma temperatures, methods being investigated may be applicable to this project in the future. Close liaison is maintained between the drilling program and this research project.

In preparation for the USGC/NAS workshop held in July 1978, Sandia Laboratories convened a workshop on Magma/Hydrothermal Drilling and Instrumentation in June 1977. The report of this workshop was issued in July 1978 (Varnado and Colp, 1978).

Borehole Stability

As drilling approaches the magma source, the temperature of the formation is going to increase, causing the rocks to approach a plastic state. A basic question must be answered: Will a hole through such a hot plastic material stay open (or are there methods for keeping it open) so that the energy-extraction equipment can be inserted into the magma source? Little is known about the physical properties of rocks at these elevated geostatic pressures and temperatures. To answer these questions, Sandia Laboratories has a continuing research program with the Center for Tectonophysics at Texas A&M University. This work is being performed under the direction of John Handin and Melvin Friedman from that Center. The following discussion is largely from the Center's progress reports to Sandia Laboratories.

The objectives of this experimental study, Mechanical Properties of Rock at High Temperature and Pressure, are to determine in static triaxial experiments the mechanical properties and behavior of appropriate igneous rocks under confining pressures to at least 3 kb, temperatures to at least 1000°C, and at strain rates from 10^{-4} to 10^{-6} s⁻¹. The four igneous rocks selected for this study and references describing them are

1. Mount Hood Andesite (Handin and Friedman, 1977)
2. Charcoal Granodiorite (Handlin and Friedman, 1977)
3. Newberry Rhyolite Obsidian (Murase and McBirney, 1973)
4. Cuerbio Basalt (Disbrow and Stoll, 1957)

Experimental Results

Crystalline Rocks--The instantaneous failure strength and strain of room-dry Mt. Hood Andesite, Charcoal Granodiorite, and Cuerbio Basalt, shortened at confining pressures of 0 and 50-MPa temperatures to 1000°C and a strain rate of 10^{-4} /s, is shown on the stress-strain curves in Figure III-1. These curves reveal the following significant facts:

1. The reproducibility of data on strength and shortening at failure is reasonably good for specimens deformed under similar conditions.
2. The failure strengths of all three rocks decrease with increasing temperature at constant-confining pressure and increase with increasing-confining pressure at constant temperature (Figures III-2, 3, and 4). The average strength-reduction of the andesite is essentially linear from 125 MPa at 25°C to 80 MPa at 1000°C unconfined, and from 280 MPa at 25°C to 195 MPa at 900°C, at 50-MPa confining pressure (1.2 and 2.4 MPa/25°C). Strength reductions for the granodiorite are much steeper. At 50 MPa, the reaction is essentially linear from 625 MPa at 25°C to an average of 190 MPa at 1000°C (11 MPa/25°C). Unconfined, the reaction is nonlinear from 250 MPa at 25°C to 35 MPa at 1000°C. The strength reduction for the basalt is also nearly linear to about 900°C; gradients are 8 MPa/25°C at 50-MPa confining pressure, and 3 MPa/25°C unconfined. Thus at 1000°C, the unconfined strength of all three rocks is less than 100

MPa; only the granodiorite at 50-MPa confining pressure remains fairly strong (~200 MPa).

3. The strengths of the rocks vanish when even partial melting occurs. This takes place at a nominal temperature of about 1030°C at the hotter end of a basalt specimen.
4. All three rocks are brittle throughout the temperature range explored until melting occurs. Shortenings at failure are less than 3%.
5. Most of the stress-shortening curves exhibit pronounced work-softening after the onset of macroscopic failure. This reflects displacement along faults that have not lost total resistance to differential stress.
6. A few tests on the andesite and granodiorite were made at strain rates of 10^{-7} /s, at temperatures of 25° and 400°C, and pressures of unconfined and 50 MPa. Except for a 30% weakening of the andesite at room temperature and 50 MPa, the 1000-fold decrease in strain rate from 10^{-4} /s has negligible effects on strength or ductility.
7. Two types of tests were made to investigate the causes(s) of the temperature-weakening. All three rocks were first heated at 50 MPa confining pressure to temperatures between 750° and 1000°C, cooled to 25°C, and then deformed at 25°C and 50 MPa confining pressure. The rocks are as strong after thermal cycling as before (Figures III-2, -3, and -4), so that the thermal cracking caused by the relatively rapid heating and cooling (1.5°C/min or 10°C/min) is not the cause of strength reductions. Indeed, specimens of the andesite are an average of 25% stronger after thermal treatment (Figure III-2), for reasons not yet understood. On the other hand, similar thermal treatment greatly reduces the strength of the unconfined rocks, presumably because the effect of thermal cracking has not been suppressed by confining pressure.

Figure III-1. Stress-Shortening Curves for Three Crystalline Rocks Deformed at Confining Pressure (P_c) and Temperatures Shown

Figure III-2. Ultimate Strength vs Temperature for Mt. Hood Andesite. Triangles (▲) represent specimens heated to 930°-1000°C, cooled, and deformed at 25°C, 50 MPa. The unconfined strength at 25°C is virtually the same whether or not a specimen has been preheated and cooled.

Figure III-3. Ultimate Strength vs Temperature for Charcoal Granodiorite. Triangle (▲) represents a specimen heated to 1000°C, cooled, and deformed at 25°C, 50 MPa. Open square (□) represents three unconfined specimens heated to 900°-1000°C, cooled, and then deformed at 25°C.

Figure III-4. Ultimate Strength vs Temperature for Cuerbio Basalt. Triangle (Δ) represents specimen heated to 750°C, cooled, and deformed at 25°C, 50 MPa. Open squares ($\square$) represent unconfined specimens heated to 905°-1000°C, cooled, and then deformed at 25°C.

Obsidian--The strength and ductility of dry Newberry-Rhyolite obsidian has been determined at 0 and 50-MPa pressure, at temperatures to 800°C, and at a strain rate of 2×10^{-4} /s. Ultimate strength decreases dramatically from 665 MPa at 600°C to 15 MPa at 800°C (Figure III-5); the brittle-ductile transition occurs between 600° and 700°C; and work-hardening ceases and steady-state flow is obtained between 700° and 750°C.

Data at 50 MPa are not appreciably different. Between 700° and 750°C the obsidian is somewhat stronger at 50 MPa than when unconfined, but the opposite is true between 750° and 800°C. In short, the increment of 50 MPa in confining pressure has insignificant effects on short-time strength and ductility.

Figure III-5. Ultimate Strength vs Temperature for Newberry Rhyolite Obsidian

Stability Analysis

Ideally borehole stresses depend on the regional state of stress in the undisturbed rock; this state of stress is not generally known a priori. Few measurements of the in-situ stress have been made in crystalline igneous rocks at depths of more than a few meters (Table III-1). Using the hydraulic-fracturing technique, the staff of the Los Alamos Scientific Laboratory has determined the stress at a depth of 2926 m in the Precambrian crystalline basement of the Jemez Mountains. Taken as the weight per unit area of the superincumbent rock, the vertical stress, σ_v , is 72 MPa and the minimum horizontal stress, σ_h , is 34 MPa. Since available data are too few for meaningful generalizations, the ratios of vertical to horizontal stresses elsewhere are not predictable. For calculations, therefore, it is assumed that horizontal stresses are equal in all directions and that σ_h/σ_v is 0.5, 1, or 2.

The principal stresses ($\sigma_1 > \sigma_2 > \sigma_3$), together with differential stresses ($\sigma_1 - \sigma_3$) and mean stresses ($\sigma_m = (\sigma_1 + \sigma_2 + \sigma_3)/3$), are given

in Table III-2 for the different ratios (σ_h/σ_v). Note that for a water-filled hole in water-saturated rock, the tangential stress can become tensile when the far-field horizontal stress is low relative to the overburden stress. This state of stress would favor unintentional hydraulic fracturing and lost circulation of the drilling fluid.

These estimates of the borehole stresses should be regarded as minimal because the stresses are higher where $\sigma_x \neq \sigma_y$, as is commonly observed in nature, and because the walls of a real drillhole are not smooth, so that stress concentrations may occur.

Figure III-6 shows one correlation of strength with borehole failure. Using the quantities listed in Table III-2 for dry rock ($p_p = 0$) and the ratio $\sigma_h/\sigma_3 = 1$, we plot differential stress ($\sigma_1 - \sigma_3$) as a function of mean pressure (σ_m) at the borehole wall for an open hole ($p_b = 0$) and a water-filled hole ($p_b = 10$ MPa/km) in impermeable rock. Choosing the proper stress gradients, $\sigma_1 - \sigma_3 = 54$ and 34 MPa/km, respectively, we relate the stresses to depth below the surface. At any given depth, borehole pressure reduces the differential stress and tends to stabilize the hole.

On this diagram of the borehole stresses as functions of depth, we superimpose the plots of ultimate compressive strength versus mean pressure, using the data of Figure III-2 for dry andesite at temperatures of 25° , 400° , 700° , and 900°C and confining pressures of 0 and 50 MPa. At a given mean pressure, say 50 MPa, the strengths exceed the borehole differential-stress so that even the open hole is stable at all temperatures. The intersections of the two sets of curves mark the boundaries between the fields of stability (containing all points to the left of the borehole-stresses loci) and of instability (containing all points to the right). Since the strength curves are subparallel to each other and to the borehole-stress curves, although the stability-instability boundaries are not sharp, they can be related approximately to depth: about 6 , 4 , 3 , and 2 km at 25° , 400° , 700° , and 900°C , respectively, for the open hole. The water-filled hole is stable at all temperatures to a depth of at least 10 km.

TABLE III-1

Some In-Situ Stress Measurements in Igneous Rocks

Place	Rock	Depth (m)	Stresses (MPa)			References
			Vertical (Overburden)	Maximum Horizontal	Minimum Horizontal	
California	Granite	300	8.2	9.5	5.4	Haimson (1976)
N Carolina	Granite	230	6.2	22.8	15.9	Haimson (1976)
Wisconsin	Granite	136	3.5	16.0	7.0	Haimson (1976)
New Mexico	Granite	2926	71.7	?	24.1	Brown (1976)
Iceland	Basalt	203	5.4	6.3	4.1	Haimson (1977)
Iceland	Basalt	375	10.0	7.0	6.4	Haimson (1977)
Africa*	Dunite	350	11.5	15.5 (average)		Gay (1975)

*Strain-relief measurement; all others done by hydraulic-fracturing technique.

TABLE III-2

Borehole Stresses (MPa/km)

Condition	σ_h/σ_v	σ_1	σ_2	σ_3	$\sigma_1 - \sigma_3$	σ_m
<u>Dry Rock</u>						
Open Hole	2	$\sigma_\theta = 108$	$\sigma_z = 27$	$\sigma_r = 0$	108	45
	1	$\sigma_\theta = 54$	$\sigma_z = 27$	$\sigma_r = 0$	54	27
	0.5	$\sigma_\theta = 27$	$\sigma_z = 27$	$\sigma_r = 0$	27	18
Filled Hole	2	$\sigma_\theta = 98$	$\sigma_z = 27$	$\sigma_r = 10$	88	45
	1	$\sigma_\theta = 44$	$\sigma_z = 27$	$\sigma_r = 10$	34	27
	0.5	$\sigma_z = 27$	$\sigma_\theta = 17$	$\sigma_r = 10$	17	18
<u>Saturated Rock</u>						
Open Hole	2	$\bar{\sigma}_\theta = 88$	$\bar{\sigma}_z = 17$	$\bar{\sigma}_r = 0$	88	35
	1	$\bar{\sigma}_\theta = 34$	$\bar{\sigma}_z = 17$	$\bar{\sigma}_r = 0$	34	17
	0.5	$\bar{\sigma}_z = 17$	$\bar{\sigma}_\theta = 7$	$\bar{\sigma}_r = 0$	17	8
Filled Hole	2	$\bar{\sigma}_\theta = 78$	$\bar{\sigma}_z = 17$	$\bar{\sigma}_r = 10$	68	35
	1	$\bar{\sigma}_\theta = 24$	$\bar{\sigma}_z = 17$	$\bar{\sigma}_r = 10$	14	17
	0.5	$\bar{\sigma}_z = 17$	$\bar{\sigma}_r = 10$	$\bar{\sigma}_\theta = -3$	20	8

$$\sigma_z = \sigma_v = \rho_r g z = 27 \text{ MPa/km for } \rho_r = 2.7 \times 10^3 \text{ kg/m}^3.$$

$$P_b = P_p = \rho_w g z = 10 \text{ MPa/km for } \rho_w = 10^3 \text{ kg/m}^3.$$

Figure III-6. Differential Stress vs Mean Pressure at Wall of an Open and Water-Filled Borehole in Impermeable Rock (Table III-2). Superposed are plots of ultimate compressive strength vs mean pressure.

Table III-3 gives the maximum depths for borehole stability, determined by the same procedure for all conditions treated in Table III-2 and for both andesite and granodiorite. It appears that boreholes in these rocks would be stable even at 900°C unless the in-situ horizontal stresses are very high. The data on basalt are similar. The obsidian is even stronger than the crystalline rocks up to about 600°C, above which, however, it is much weaker, its strength approaching zero at 800°C.

TABLE III-3

Maximum Depth (km) of a Stable Borehole

Conditions		σ_h/σ_v	25 °C	400 °C	700 °C	900 °C	
Andesite	Dry	Open Hole	2.0	2	2	1	1
			1.0	6	4	3	2
			0.5	>10	>10	>10	>10
		Filled Hole	2.0	4	3	2	1
			1.0	>10	>10	>10	>10
			0.5	>10	>10	>10	>10
	Saturated	Open Hole	2.0	2	2	1	1
			1.0	9	7	5	4
			0.5	>10	>10	>10	>10
		Filled Hole	2.0	5	4	3	2
			1.0	>10	>10	>10	>10
			0.5	*	*	*	*
Granodiorite	Dry	Open Hole	2.0	6	3	1	>1
			1.0	>10	>10	>10	2
			0.5	>10	>10	>10	>10
		Filled Hole	2.0	>10	>10	6	1
			1.0	>10	>10	>10	>10
			0.5	>10	>10	>10	>10
	Saturated	Open Hole	2.0	6	3	>1	1
			1.0	>10	>10	>10	>2
			0.5	>10	>10	>10	>10
		Filled Hole	2.0	>10	>10	>10	2
			1.0	>10	>10	>10	>10
			0.5	*	*	*	*

*Tangential Stress is Tensile

*Tangential Stress is Tensile

These predictions are optimistic, since the borehole stresses would be higher than supposed if the far-field horizontal stresses were not equal and if the stress concentrations in a real hole were greater than those idealized for the calculations. Furthermore, the effects of pore water and time have not yet been fully accounted for.

Conclusions

The major conclusions to be drawn from the experimental results now available are:

1. The failure strengths of the crystalline rocks decrease gradually with increasing temperature to about 900°C, and then more rapidly as partial melting is approached between 1000° and 1100°C. The obsidian is even stronger than are the crystalline rocks up to 600°C, above which, however, the glass softens and strength vanishes at 800°C.
2. The crystalline rocks are essentially brittle throughout the temperature range explored until partial melting occurs. They deform primarily by fracture. Ductile flow in the groundmass where it is present, some gliding flow (kinking) in the biotite, and thermal alteration of the mafic minerals contribute to the strength reduction by an unknown, but probably minor, amount. Shortenings at failure are less than 3%. Virtually all the permanent shortening is ascribable to macroscopic faulting.
3. Several facts strongly suggest that the temperature-weakening of the crystalline rocks under confining pressure is due primarily to intrinsic effects of temperature on the process resulting in macroscopic shear fracturing (faulting), not merely to degradation of strength by thermal cracking, at least when the rocks are subjected to confining pressure:
 - brittle fracture accounts for virtually all the permanent strain;

- the degree of microfracturing associated with the mechanical loading, as distinguished from the thermoelastic stressing, is about the same at all temperatures from 25° to 1000°C; and
- specimens heated to 1000°C, cooled, and then deformed at 25°C are at least as strong as those that have not been preheated.

On the other hand, thermal cycling degrades the strengths of the unconfined rocks, presumably because the effect of the thermal cracks on subsequent failure is not suppressed by confining pressure; that is, heating significantly modifies the mechanical state of the material. Thus, data from unconfined tests alone cannot be used for valid predictions of rock properties under in situ conditions, particularly for geological-geophysical applications. However, such tests may be useful to estimate strengths during drilling where unconfined conditions and rapid cooling may occur.

4. Because the deformations of the rocks are essentially cataclastic at all temperatures, it is supposed that the borehole will fail in shear when the differential stress reaches the ultimate compressive strength. A hole as deep as 10 km in the dry rocks under short-time loading is not likely to fail even at 1000°C, unless the in-situ horizontal stress is equal to or greater than the vertical stress and the hole is open (Table III-3).
5. Predictions of borehole stability are optimistic, because the effects of loading duration (creep) and water-weakening which should significantly lower the strengths of the rocks at the higher temperatures have not yet been explained. Assessment of these effects is the next objective of the research. Predictions must remain qualitative until sophisticated numerical methods can be applied to the problem. This cannot be done until adequate constitutive relationships are known over the entire range of significant variables--at least of confining pressures from 0 to 200 MPa, temperatures from 25° to 1000°C, and periods of a few seconds to several years--let alone the chemical effects of interstitial fluids.

Source Definition Drilling Experiments

The validation of the geophysical sensing experiments must be achieved by drilling into and through the lava lake to provide physical confirmation of its structure and properties.

Since 1960, fifty-five holes have been drilled through the crusts of three Hawaiian lava lakes. Forty-two of these were small holes ($\approx 1\text{--}3/16$ in. dia), drilled into Alae and Makaopuhi lava lakes along the Kilauea east rift by the USGS Hawaiian Volcano Observatory (HVO). The remaining 13 holes were drilled into Kilauea Iki lava lake (Richter and Moore, 1966; Rawson and Bennett, 1961; Okamura, 1977).

The last holes in Kilauea Iki were drilled by Sandia Laboratories in 1976, in cooperation with HVO. These were the first attempts to drill through the lava lens to determine the thickness of the molten rock and to verify the geophysical sensing experiments. Subsequently, the Lava Lake Drilling Program, an effort to design and develop molten rock drilling systems specifically to explore Kilauea Iki, was initiated in FY77. Field drilling experiments using these systems will be conducted in FY79.

The 1976 Drilling Experiments: Kilauea Iki 76-1 and 76-2

The drilling plan was formulated, using two methods of penetration through the melt. Although the previous drilling efforts at the lava lakes had not attempted to penetrate the melt, the methods were based on extensive experience gained by the USGS/HVO and were limited by conventional drilling technology. The first method called for drilling through the upper crust to the solid/melt interface, using a diamond coring bit and water as the drilling fluid. Upon penetrating the interface the driller would let the drill string fall or push it to the bottom crust of the molten lens. The second method was to drill through the upper crust to the solid/melt interface; then, with maximum flow of cooling water, chill a solid zone below the bit, drill to the melt, chill a solid zone ahead, and repeat until the bottom crust was reached.

Sandia negotiated a contract with Continental Drilling Hawaii, Inc. to furnish a drill rig, equipment, and drilling crew. The HVO provided the services of R. Okamura, who had done all of their previous lava lake drillings, to act as drilling consultant. R. Helz, USGS, supervised the coring activities and will be responsible for core analysis.

The drilling of hole KI 76-1 was started on August 18, 1976. On August 25 the solid/melt interface was reached at a depth of 45.5 m (149.3 ft). Core recovery was 99% of rock drilled. The core barrel dropped 0.7 ft into the melt before it was retrieved.

On the following day, drilling was started at 44.2 m (145 ft); the solid material at the 44.2-m (145-ft) level resulted from the molten rock oozing back up the hole overnight and cooling. Molten rock was encountered at about 45.5 m (149+ ft), and the drill string was pushed to a depth of 46 m (151+ ft), where progress stopped. Simultaneously, the drill string was frozen in the melt so that no movement (rotational or vertical) was possible.

Drilling activities were resumed on August 26. With the aid of temperature equilibration, the drill string could be rotated and raised but could not be pushed further into the melt. It was realized that the core bit that was downhole could not have survived at the temperature to which it was subjected. An unsuccessful attempt was made to hammer the drill string deeper into the melt with a 300-lb weight.

Figure III-7 shows a photo of the retrieved bit and an identical, unused one. Subsequent attempts to use borehole 76-1 were unsuccessful because of an obstruction wedged at 42-m (138-ft) depth, apparently associated with the retrieved bit.

Borehole 76-2 was to be drilled with the second method, "chill and drill," from the melt interface downward. To be successful the driller must know when he encounters the molten rock so he can withdraw his drill string and pump quantities of water downhole to chill the melt.

Figure III-7. Special Carbide Insert Core Bits
Used in Kilauea Iki Lava Lake Drilling in
August 1976

The drilling of hole KI 76-2 started on August 30, 1976. Core recovery was 98% on this hole. The solid/melt interface was reached at 45.7 m (149.9 ft) on September 2. Two attempts to penetrate the melt using the "chill and drill" method were unsuccessful because the drilling core bit was blocked by quenched glass from the molten rock. The inability to prevent blockage of the core bit resulted from the combination of two effects: inadequate means to detect the onset of the melt zone, and an inadequate water supply to effectively overcome the pressure of the molten rock and chill the melt an adequate distance in front of the bit. Figures III-8 and -9 show photos of the plugged bit.

Figure III-8. Standard NX Carbide Core Bit Used in 1976 Kilauea Iki Lava Lake Drilling Program, Showing Solidified Glass Plug

Figure III-9. Section Cut Through Glass-Plugged NX Core Bit Used in 1976 Kilauea Iki Drilling Program

From this initial drilling effort, considerable knowledge was gained:

1. The liquid pressure in the molten lens was greater than 80-100 psi (≈ 0.69 MPa).
2. Molten rock was encountered at 45.5 m (149.3 ft).
3. The molten lens contains at least one rigid obstruction to penetration.
4. A molten rock drilling system for use in this location must have
 - (a) The ability to drill through a rigid obstruction (rock) encountered in the melt zone at a temperature of about 1050°C .
 - (b) A drill stem capable of operating while immersed in molten rock at a temperature of $\approx 1070^{\circ}\text{C}$ or having a method of positively preventing the molten rock from contacting the drill stem.
 - (c) Any holes in the drilling system open to the molten rock filled with a material (liquid or gas) at a pressure greater than the maximum molten rock liquid pressure that could be encountered ($P_{\text{max}} = 288$ psi or 2 MPa)
 - (d) A sealing method that will prevent molten rock from being forced up and freezing between the drill stem and the drill hole in the solid upper crust, causing a seizure.

The Lava Lake Drilling Program

Because of the need to obtain physical information on Kilauea Iki to evaluate the geophysical sensing experiments, Sandia decided early in FY77 to concentrate the majority of the Magma Energy Research Project on the development of drilling systems that could penetrate a lava lake. The experience gained from the 1976 drilling experiments indicated that conventional drilling technology would not be adequate and the complexity of the lava lake negated the concept of pushing the drill string through the melt; consequently, more attention had to be given to high temperature materials and design practices. The efforts concentrated on the development of a system for drilling very hot ($\approx 1050^{\circ}\text{C}$) solid rock while the drill string, immersed in molten ($>1075^{\circ}\text{C}$) rock rotates.

Numerous concepts of methods and equipment for performing the Lava Lake Drilling Program were suggested. These were outlined in Table III-4, along with their status. A summary of development activities on the uncooled drag bit concepts and jet-augmented drill are presented below.

Uncooled Drag Bit--A three-wing drag bit, made of a superalloy material with adequate strength properties at 1100°C, was selected because it is simple, strong, and has no moving parts. Chip clearance is effected by the use of low-volume, high-velocity jets of air/water mist through small diameter holes in the bit. The bit is not designed to be cooled, but some cooling will take place due to the flow of the chip disposal fluid. Figure III-10 shows a photo of the cast drag bit.

Figure III-10. Superalloy MAR-M 509 Drag Bit, Used for Drilling Molten Rock

TABLE III-4

Summary and Status of Proposed Drilling Concepts

<u>Proposed Concept</u>	<u>Bit Type</u>	<u>Drill Stem Type</u>	<u>Lava Seal Req'd?</u>	<u>Fixed Casing Req'd?</u>	<u>Coolant Type</u>	<u>Surface Equipment Req'd</u>	<u>Status</u>
Conventional Core Bit w/Hi-Pressure Coolant	Core bit	Insulated	Yes	Yes	Water	Rotary drill	No further work under way or contemplated
Explosive Tip	Open Penetrator	Insulated	Yes	Yes	Water?	Axial thruster	No further work under way or contemplated
Exothermic Tip	Closed Penetrator	Insulated	Yes	Yes	Water?	Axial thruster	No further work under way or contemplated
Electrical Subterrene	Closed Penetrator	Insulated	Yes	Yes	Inert	Axial thruster	No further work under way or contemplated
Bubble-Blowing/ Freezing Bit	Special bit	Standard	Yes?	Yes	Air and Water	Rotary drill	No further work under way or contemplated
Uncooled Drag Bit	Special drag bit	Insulated	Yes	Yes	Air/Water Mist	Rotary drill	Design, analyses, and fabrication underway
Jet-augmented Drill	Special	Standard	Yes?	Yes	Air and Water	Rotary drill	Design, analyses, and fabrication underway

Mar-M 509, a cobalt-based superalloy, is the material used for the drag bit. Titanium-carbide inserts were selected for the cutting surfaces of the bit. The final bit design has an outside diameter of 4.75 in.

A double-pipe, insulated drill stem is proposed. This is a feature common to all of the proposed concepts (see Table III-4). The inner pipe will be cooled to a temperature $\approx 200^{\circ}\text{C}$ during operation. It will transmit the major portion of the required torque to the drill bit. The outer pipe will be insulated from the cooled inner pipe so that it can attain a temperature close to that of the molten rock ($\approx 1075^{\circ}\text{C}$) which it contacts directly. This is necessary to prevent a buildup of solidified rock on the outer surface of the outer pipe, with the resultant high increase of viscous drag during rotation in the molten rock.

A lava seal is proposed to prevent the upward flow of molten rock into the annulus between the rotating drill stem and the cased bore hole in the solid upper crust. Air pressure (190 psi) greater than the assumed maximum molten rock liquid pressure encountered at the top of the molten lens will be maintained on top of the downhole lava seal.

A hot (uncooled) experiment section can be placed between the uncooled drag bit and the lower end of the insulated drill stem. The experiment section would contain the transducers required to perform active experiments such as temperature gradients of the molten lens, heat transfer coefficients, etc. Figure III-11 depicts a preliminary layout of this concept, showing the insulated drill stem, lava seal, and other details.

This concept was selected as the primary system to be used in the Lava Lake Drilling Program. It has been designed, laboratory tested, and fabricated for that use.

Figure III-11. Insulated Drill String Schematic

The Jet-augmented Drilling System--The jet-augmented bit is similar to conventional bits, except that coolant is forced through the bit and into the formation ahead of the bit. When the coolant encounters molten rock, the liquid vaporizes and simultaneously freezes the liquid rock. The result of this, ideally, is the formation of a chilled porous-rock crust (foam) ahead of the bit. The basic principle is to chill and foam the medium ahead of the bit and then use conventional drilling technology to drill through that solid medium. Simple model tests and preliminary calculations indicate that this concept might work in molten rock.

Model drilling experiments were first run on this concept using a 1/1-scale drill rig which was designed to drill through a crust of solidified wax and into a molten wax region below. The drill was designed to carry Freon-11 as a coolant. High-speed movies were taken, and it was demonstrated that a porous or foam structure of solidified wax was formed around and ahead of the drill bit. Later experiments were run in molten glass and molten basalt; it was shown that a porous foam structure could be formed by passing coolant into these molten materials. Figure III-12 shows such a test being run in molten glass.

Two drilling techniques are proposed for entering molten rock: a water-jet-augmented drag bit--a modified drag bit with a central nozzle used to jet water forward into the magma to cool and solidify it into a porous matrix; and a water-cooled core barrel and jet-augmented core bit to jet the cooling water forward into magma--the resultant solidification will allow a 10-ft-long core of solidified molten rock to be obtained.

1. Jet-Augmented Drag Bit. The jet-augmented drilling concept is shown schematically in Figure III-13, and the modified drag bit is shown in Figure III-14. The bottom seal is created by drilling into liquid lava and allowing it to flow up the drill hole and solidify. The top of the casing is sealed with a rotary seal, and a steam outlet and air-flow inlet connections are provided. The entire inner casing is pressurized with air to a pressure above the lake geostatic pressure. With water providing the cooling, a porous-solid matrix should be formed. Should a crack develop in the solid structure, steam or water would be forced into the opening, thereby cooling it and closing it off. Nozzle calculations for water-jet-augmented drilling indicate a flow rate of 8.4 gpm with nozzle in molten rock, and a jet velocity of 86 ft/s in molten rock.

The maximum drilling rate is calculated at 2 ft/h, which should allow sufficient cooling to solidify a 5-ft-dia cylinder. When the original melt-solid lake interface is penetrated, the feed rate will be stopped for about 30 min to establish a solid base at the interface of about 4.5-ft diameter, from which the porous matrix can be grown downward.

2. Jet-Augmented Core Barrel and Bit Assembly. The jet-augmented core system will be used as shown in Figure III-13, where the core barrel and the bit will replace the drag bit. The modified bit is shown in Figure III-15. The core barrel is a standard Longyear NV series.

The core dimensions are 1-7/8 in. dia by 10 ft long. The bit has been modified to contain six nozzles of 0.081-in. dia, spaced 60° apart, and located so that water is jetted ahead of the bit. Calculated flow parameters are a water-flow rate of 8.4 gpm to 18.8 gpm and a jet velocity of 86 ft/s. A solid base will be established at the lake interface in the same manner as described above for water-jet-augmented drag-bit drilling.

Insulated Drill Stem--The insulated drill stem concept shown in Figure III-11 consists of an inner tube that is cooled to ~200°C by an internal flow of coolant; a layer of high-temperature, high-efficiency insulation; and an outer tube that remains at the temperature of the molten rock.

The material selected for the inner tube was RA 310, a high nickel, high-temperature-resistant alloy stainless steel--in case the design temperature of 300°C was exceeded. The insulation between the inner and outer tubes was an assembly composed of a nickel-foil radiation shield next to the inner pipe, approximately 1/4-in.-thick kaowool blanket, another nickel-foil radiation shield, another 1/4 in. of kaowool blanket, and a layer of stainless-steel shim stock 0.002 in. thick which was spirally wound and held in place by nichrome wire. The outer tube was fabricated from RA 330, another high nickel, high-temperature-resistant stainless-steel alloy.

Figure III-12. Molten Glass Laboratory Experiments

Figure III-13. Pressurized Drilling Schematic

Figure III-14. End View of Jet-Augmented Drag Bit

Figure III-15. Jet-Augmented Core Bit

A specially designed instrumentation section joined the lower ends of the insulated section to the drill bit. The outer and inner tubes were welded to this instrumentation section, which was a high-nickel, heat-resistant stainless-steel alloy, RA 333. The Mar-M 509 drag bit was threaded on its lower end. The instrumentation section also contained the terminations of several Inconel-sheathed Chromel-Alumel thermocouples for temperature and heat-flow experiments measurements to be made while the instrumentation section is immersed in the molten rock.

A splined section to allow differential linear thermal expansion joins the upper ends of the tubes forming an ~60-ft-long insulated section.

Lava Seal--Although the insulated drill string appears to offer solutions to the problems of transmitting adequate torque to the bit and to inhibiting the build-up of solidified lava on the outside, one other problem remains. This is the flow of molten rock up the space between the drill string and the drilled hole in the rock of the solid upper crust. The migration (under geostatic pressure) of this molten ooze upward to a cooler ($\sim 100^{\circ}\text{C}$) environment will result in its freezing the drill string against the bore hole in the solid upper crust.

To eliminate or minimize this problem, the concept of a lava seal was developed. The seal was designed to confine a flow of air along the outside of the drill string, thus maintaining an air pressure greater than the expected liquid pressure in the molten lava. This pressure differential was designed to prevent the upward migration of the molten lava.

Confinement of the airflow is achieved by installation of a full-length casing inside the borehole. The lower end of the casing has a lip that will allow the drill bit and drill string to pass but will retain a series of rings designed to minimize the airflow but maintain the required air pressure. The final design consists of a series of four seal rings. Adjacent pairs of rings either are solid and have a thermally induced interference fit against the inner diameter of the lower section of the fixed casing, or are split and have a spring fit against the outer diameter of the rotating outer insulated tube.

The rings are attached to the instrumentation section at the lower end of the insulated drill stem during insertion by frangible pins. At the start of drilling, these pins are sheared off by the drill-rig thrust, allowing the rotating-drill stem to pass downward through the seal. The rings are made of RA 333, a high-nickel stainless steel. The lip and the lower 2 ft of the casing are made of Inconel 625; the next 40 ft of casing is made of 316 stainless steel. The remainder of the casing is conventional carbon steel. Figure III-16 is a drawing of the seal rings in cross-section.

Program Status--Equipment for each of these concepts is being designed, fabricated, and assembled. Various parts of the systems are being tested to the extent possible in the laboratory. The completed systems will be demonstrated in the FY79 Lava Lake Drilling Program.

Figure III-16. Insulated Drill String Schematic

IV. TASK 3 -- MAGMA CHARACTERIZATION AND MATERIAL COMPATIBILITY

Magma Characterization

Introduction

The Magma Workshop held in March 1975 (Colp et al, 1975) included a working group on Chemical and Physical Properties of Magma. From the discussions of that working group has come, in a large part, the direction for the Sandia studies in magma characterization.

Knowledge of many chemical and physical properties of molten rock (magma) and of solid rock at high temperatures is essential for evaluating the feasibility of extracting energy directly from magma bodies. In order to evaluate which properties are most critical, the panel on chemical and physical properties first reviewed the three major stages in developing energy directly from magma: (1) finding a magma body; (2) developing the body (drilling into it and emplacing heat-transfer devices sufficiently strong and corrosion-resistant to last an appreciable period); and (3) extracting energy from the body. Various properties were then reviewed in terms of their relationships to these three stages and our present knowledge about them. Possible approaches toward increasing our knowledge of the critical properties were discussed, and recommendations for future research were prepared.

The critical factors in evaluating the feasibility of extracting energy from a given magma body are the heat-transport mechanisms and rates. Will magma in the body convect heat at a rate sufficient to

increase heat extraction significantly above that of a nonconvecting magma? If not, can convection be induced? The magma very probably will be below liquidus temperatures and will consist of a mixture of melt and crystals, with the possible addition of a gas phase. Relevant critical properties that need to be determined are the distribution of liquid, crystals, gas, and temperature in the magma; the viscosity and density of crystal-liquid-gas mixtures; and the heat-transfer coefficients over the relevant range of temperature for magmas of different compositions. Other properties that are critical to the exploration and development of magma bodies are the content of H, C, O, S, F, and Cl in the magma and overlying rock; the strength, elasticity, and ductility of the margin; and in-situ physical properties of partially and completely molten magma and hot rock relevant to exploration, such as seismic, electrical, and magnetic properties, density, and permeability.

The panel recommended a two-pronged attack on the problem:

- The selection of a site overlying a probable magma chamber for exhaustive geologic, geochemical, and geophysical studies, including research drilling to the magma chamber.
- Measurements in the laboratory of the critical physical properties and related computed modeling.

Magmatic Gas Studies

Magmas contain volatile elements which can form a separate gas phase at pressures appreciably above atmospheric pressure. Magma characterization with respect to volatiles is of concern in the experimental determination of physical properties of magmas (density, viscosity, electrical resistivity, etc) which are sensitive to volatile composition; the selection of engineering materials for magmatic environments; and the design of heat exchangers exposed to mass transfer effects from circulating volatiles. Accordingly, we have followed the recommendation of the 1975 Magma

Energy Workshop (Colp et al, 1975) that characterization studies be made of complex magmatic volatiles.

Computer Code Development--A set of computer codes has been developed to calculate equilibrium distributions in multicomponent gas mixtures over a wide range of specifiable conditions. The calculations can be carried out by treating the gas phase as an ideal or nonideal mixture. Nonideal properties are calculated from measured P-V-T data (if available) or from a corresponding state version of the Redlich-Kwong equation. The equilibrium distributions can be calculated at a specified temperature and pressure for a full set of mass balance constraints, or one or more mass balance constraints can be replaced by fixed fugacities or activities. The latter feature is useful in many geochemical problems where typically some activities and some, but rarely all, mass balance constraints are known. For example, these codes can be used to calculate the atomic compositions of all gas mixtures in a complex system that would be compatible with a fixed chemical potential determined by the presence of a particular mineral assemblage.

The codes are not reaction-specific, in that species distributions can be determined for a system of any arbitrary number of components provided that standard-state free energy data are specified for all species of interest. Therefore, it is not necessary to rewrite the codes each time new chemical systems are investigated. Data can be processed at fixed pressure over a range of temperatures (up to 1700°C), or for any arbitrary combinations of temperatures and pressures. The codes have been equipped with input features that allow random access to a thermodynamic data file prepared for 275 gas species. Output results are listed and can also be displayed with flexible graphics subroutines designed to produce two- and three-dimensional projections (examples given in Figure IV-1 and -2).

Figure IV-1. Surfaces Define the Elemental Compositions in the C-O-H-S System Compatible With Nickel-Nickel Oxide (NNO), Quartz-Fayalite-Magnetite (QFM), Magnetite-Wustite (MW), and Graphite (G) at 1200°C and 1-kbar Total Pressure. Element compositions have been expressed as logarithms of oxygen, hydrogen, and sulfur normalized to carbon. O_2 fugacity isobars are nearly coincident for this system at the stated conditions except for gases with elemental S/C greater than 1.

Figure IV-2. The Mole Fractions (XI) of the Principal Species in C-O-H-S Gases With Elemental S/C = 0.2 at 1200°C and 1 kbar with O_2 Fugacity Equivalent to That for QFM are Shown as a Function of Elemental H/C.

A computer code developed to compute stability diagrams for hydrothermal mineral assemblages (Helgeson et al, 1967) has been adapted to compute mineral stability diagrams for high-temperature magmatic systems (e.g., Figure IV-3). This code is also being used to calculate the stability fields of metal in terms of volatile fugacities. The fugacities of magmatic gases can be plotted on these diagrams for any desired set of temperatures and pressures; variable activities for melt and solid-phase components are permitted. These diagrams provide a useful tool for characterizing magmatic volatiles, and for predicting the stability of engineering materials in magmatic environments.

Figure IV-3. The Stability Fields for Solid Oxides and Sulfides in the System Fe-SiO₂-O-S Are Expressed in Terms of the Fugacities of O₂ and S₂ at 1400°C, 1-bar Total Pressure, and Unit Activity for All Components in Solid Phases

Evaluation of Volcanic Gas-Collection Data--The computer codes described above have been employed in detailed studies of approximately 100 volcanic gas analyses for collections made in the past 20 yr from Nyiragongo Lava Lake (1959), Surtsey Volcano (1964, 1965, 1967), Mount Etna (1970), and Erta Ale Lava Lake (1971, 1973, 1974). Investigations have been restricted to these collections because they were made from well-described sites in close proximity to active outgassing lava vents. All collections were made at temperatures above 950°C and most were made above 1050°C. Nevertheless, our studies of these collections show they have been severely to moderately affected by secondary alterations, even after correction for atmospheric contamination. Secondary modifications are related mainly to problems in the collection and analytical procedures; they include oxidation-reduction and sulfidation reactions between erupted gases and metal collection devices, water condensation and reevaporation in lead-in tubes, analytical errors in H₂O determinations, and other relatively minor modifications. These secondary modifications are responsible for the erratic and chaotic compositions reported for several of the above gas collections.

It is possible to remove the major imposed modifications from the reported gas analyses and to restore them to compositions more representative of the actual erupted gases. The details of the restoration procedures have been described elsewhere (Gerlach, 1978, 1978a, 1978b, 1978c). Information obtained in these investigations has also been used to make several improvements in gas-collection devices and techniques; an example is described elsewhere in this report.

The restored analyses allow, for the first time, a meaningful comparison of data from different volcanoes and of data taken at different times from the same volcano. The restored analyses show relatively regular and uniform compositions for each series of collections. They show little evidence of short-term (minutes to days) variations. Hence, it appears that significant data can indeed be obtained over short periods of time in the proximity of active outgassing lava sources.

Volcanic gases from tholeiitic lavas (Surtsey and Erta Ale) tend to be relatively H₂O-rich (70% to 90%). They have compositions which are consistent with observed outgassing patterns of submarine tholeiites from progressively shallower depths along zones of seafloor spreading. They also show composition changes over long periods of time (months to years) which are dominated by a decreasing CO₂ content. A trend of this sort is compatible with the relative solubilities of CO₂, water, and sulfur in silicate melts at low to moderate pressures. The volcanic gases associated with more alkalic lavas (hawaiite at Etna and nephelinite at Nyiragongo) are considerably more CO₂-rich (20% to 50%) than those associated with tholeiitic lavas. This result is consistent with recent findings in experimental petrology indicating that alkaline partial melts are favored with increasing CO₂ fugacity, while increasing H₂O fugacity favors tholeiitic partial melts. Because of their higher CO₂ content, alkaline magmas outgas appreciably at greater depths than do tholeiitic magmas, and they tend to erupt more explosively. The restored analyses also indicate that lavas with high O₂ fugacities (e.g., Etna) tend to release gases with a larger total sulfur content (SO₂ + H₂S + S₂) than do lavas with relatively low O₂ fugacities (e.g., Nyiragongo). These relationships can be accounted for by the mechanism of sulfur solubility in mafic lavas, and they help explain the heretofore puzzling sulfur and sulphur/carbon variations observed in volcanic gases.

The results of our investigations of volcanic gas collections are summarized below for each volcano.

Nyiragongo. Restoration methods have been employed to estimate the gas-phase composition at Nyiragongo lava lake, based on samples obtained in 1959. H₂O data were not reported in 11 of the 13 original analyses. Correction procedures have been used to estimate the H₂O contents of the sample and to adjust the analyses for atmospheric contamination, loss of sulfur, and pre- and post-collection oxidation of H₂S, S₂, and H₂. The estimated gas compositions (Table IV-1) are low in total sulfur and relatively reduced, containing approximately 35% to 50% CO₂, 45% to 55% H₂O, 1% to 2% SO₂, 1% to 2% H₂, 2% to 3% CO, 1.5% to 2.5% H₂S, 0.5% S₂,

and 0.1% COS over the collection temperature range 1020° to 960°C. The O_2 fugacities of the gases are consistently about half an order of magnitude below QMF. The total sulfur content of the gases is low, but the estimated S_2 fugacities are relatively high. These characteristics and the high H_2S/SO_2 of the estimated compositions are related to the relatively low f_{O_2} of the lava lake. At temperatures above 800°C and pressures of 1 to 1.5 kbar, the estimated gas compositions resemble those observed in primary fluid inclusions believed to have formed at similar temperatures and pressures in nephelines of intrusive alkaline rocks; cooling to 300°C results in gas compositions very rich in CH_4 (50% to 70%) and resembling secondary fluid inclusions formed at 200° to 500°C in alkaline igneous rocks. Below 600°C, the gases become supersaturated in carbon as graphite. These characteristics are corroborated in several reports of hydrocarbon gases and bituminous compounds in intrusive alkaline rocks.

TABLE IV-1

Estimated 1959 Nyiragongo Lava Lake Gas Compositions (mole %)

Sample No.	CO_2	H_2O	CO	H_2	H_2S	SO_2	S_2	COS	T (°C)	Log f_{O_2}
2	48.55	43.50	2.20	1.29	1.72	2.02	0.62	0.09	970	-12.0
12	45.44	45.90	2.72	1.59	1.41	2.30	0.55	0.08	1020	-11.3
13	36.35	55.62	2.13	2.18	2.45	0.81	0.38	0.08	960	-12.4

Surtsey. Evaluations have been made of volcanic gas collections from Surtsey volcano eruptions between 1964 and 1967. The reported analyses for eight high-temperature (1125°C) samples taken at or very near the crater closely approach a state of chemical equilibrium and are clearly representative of the erupted gases. They have been only slightly effected by atmosphere contamination and by reduction processes resulting from interactions with stainless-steel collection tubes. After corrections for these alterations, all resulting compositions are relatively water-rich (Table IV-2); they contain on the order of 80% to 90% H_2O , 1% to 10% CO_2 , 2% to 4% SO_2 , 1.5% to 3% H_2 , 0.1% to 0.7% CO, 0.1% to 0.9% H_2S , and 0.01% to 0.25% S_2 . O_2 fugacities are very near or slightly below the QMF buffer at the collection temperature. The gas compositions show evidence of a

decrease in CO_2 over the period 1964 to 1967, suggesting a progressive decrease in the CO_2 content of the erupting magmas. The water-rich gas compositions are consistent with the observed degassing patterns of submarine basalts from progressively shallower depths along the Reykjanes Ridge.

TABLE IV-2

Improved Compositions for the Surtsey Volcanic Gas
Samples at 1125°C (mole %)

Sample No.	CO	CO ₂	HCl	H ₂	H ₂ O	H ₂ S	SO ₂	S ₂	Log f _{O₂} bar
(15 Oct 1964)									
12	0.69	9.29	0.81	2.80	81.13	0.89	4.12	0.25	-9.80
13	0.71	9.79	0.81	2.77	81.98	0.64	3.15	0.13	-9.78
(21 Feb 1965)									
17	0.38	5.01	0.40	3.12	87.88	0.63	2.46	0.10	-9.82
22	0.36	6.43	0.40	2.27	87.83	0.24	2.43	0.03	-9.55
24	0.39	5.54	0.43	2.86	87.40	0.54	2.72	0.09	-9.75
(31 March 1967)									
29	0.11	3.31	1.16	1.42	91.11	0.06	2.81	0.01	-9.11
30	0.05	1.21	1.04	1.79	92.47	0.14	3.27	0.02	-9.30
31	0.05	1.31	0.52	1.74	92.46	0.15	3.74	0.02	-9.27

Mount Etna. The 1970 Mount Etna volcanic-gas collections are among the most reduced volcanic-gas samples ever obtained, containing 20% to 40% H₂, 2% to 3.5% CO, and 2% to 5% H₂S. Calculated oxygen fugacities for most of the analyses are well below QFM, several are more reduced than MW, and all are many orders of magnitude less than those measured in the gas-streams of the collection sites at the time the samples were taken. The analyses show no similarity to calculated equilibrium compositions at any temperature. Deviations between analytical and equilibrium composition indicate the gases have undergone extensive reduction mainly involving loss of oxygen. The evidence indicates that the reduced analyses are not the products of unusually reduced lavas, but originated from reactions of the erupted gases with the metal sampling device used in the collection procedure. There also is limited evidence for sulfur-loss from sulfidation reactions between the erupted gases and the copper collection tube. The

oxygen deficiencies of the analyses have been restored using the analytical atomic hydrogen, carbon, and sulfur data and the oxygen fugacity field data. The restored analyses (Table IV-3) are much more representative of the erupted gases, which were remarkably rich in CO_2 (15% to 35%) and SO_2 (15% to 35%). Unlike the original analyses, the restored analyses show relatively uniform and steady compositions at each collection site over periods of observation ranging from hours to days.

Erta Ale. Analyses of 18 volcanic-gas samples collected at 1075°C over a 2-h period from Erta Ale lava lake in December 1971 have highly erratic compositions caused principally by H_2O analytical errors and atmospheric contamination and oxidation. These imposed modifications are responsible for the gross chemical disequilibrium in the analyses. The restored analyses for the 1971 collections are given in Table IV-4. Correction procedures have also been employed to restore the anhydrous mean compositions reported for 46 collections made from the lava lake at temperatures of 1100° and 1210°C in January 1973. The results indicate the 1973 erupted gases had compositions very similar to those of the 1971 gases (69.7% to 71.3% H_2O , 1.6% to 2.4% H_2 , 17.8% to 19.4% CO_2 , 0.8% to 1.6% CO , 4.9% to 8.8% SO_2 , 0.2% to 0.5% S_2 , and 0.4% to 1.0% H_2S).

Analyses of 18 volcanic-gas samples collected over a 0.5-h interval from the active lava lake at Erta Ale volcano in 1974 have moderately erratic compositions caused by secondary modifications imposed on the erupted gases during collection. Most variability in the analyses resulted from condensation and reevaporation of water entrained in the stainless-steel collection tube. Alterations of smaller magnitude included atmospheric contamination, HCl fluctuations, and irregularities in sulfur species resulting from reactions with the collection tube and possibly with atmospheric O_2 . Minor irregularities in the H_2 apparently were caused by HCl and/or H_2O reactions with the collection tube. In spite of the high collection temperature (1125° to 1135°C), all analyses are grossly out of chemical equilibrium even after removal of atmospheric contamination. The restored 1974 analyses obtained after removal of the above modifications are given in Table IV-5.

TABLE IV-3

Factored Analyses for the 1970 Mount Etna Volcanic Gases at 1075°C, $f_{O_2} = 10^{-9.475}$ atm

Hornito No.	Date (July)	Time	Sample No.	CO	CO ₂	H ₂	H ₂ O	H ₂ S	SO ₂	S ₂	SO	
1	12	11:32	11	0.48	22.76	0.30	27.71	0.22	47.70	0.76	0.06	
		13:20	16	0.31	14.76	0.71	66.22	0.20	17.68	0.10	0.02	
	13	11:30	8	0.48	22.87	0.50	46.91	0.22	28.70	0.28	0.04	
	14	6:13	13	0.46	22.23	0.48	44.59	0.24	31.61	0.33	0.04	
2	12	12:40	10	0.54	26.06	0.51	47.26	0.20	25.18	0.21	0.03	
		12:50	17	0.51	24.55	0.50	46.80	0.21	27.14	0.25	0.03	
		15:05	14	0.52	24.78	0.61	57.09	0.16	16.72	0.09	0.02	
		15:10	12	0.46	22.12	0.64	59.45	0.17	17.04	0.10	0.02	
	13	10:40	5	0.51	24.32	0.53	49.33	0.20	24.86	0.21	0.03	
		11:45	7	0.50	24.00	0.46	43.09	0.23	31.34	0.33	0.04	
		13:05	6	0.49	23.56	0.54	49.91	0.21	25.05	0.21	0.03	
		14:30	9	0.41	19.71	0.52	48.27	0.25	30.48	0.31	0.04	
	14	15:45	15	0.36	17.08	0.51	47.82	0.27	33.54	0.38	0.04	
	3	13	11:00	1	0.70	33.58	0.53	49.50	0.13	15.45	0.08	0.02
			12:00	4	0.71	33.93	0.54	49.91	0.12	14.69	0.07	0.02
<hr/>												
			*16A	0.49	23.42	0.51	47.10	0.22	27.96	0.26	0.04	
			14A	0.45	21.69	0.54	49.88	0.22	26.93	0.24	0.03	
			12A	0.42	20.00	0.57	53.69	0.22	24.85	0.21	0.03	

*16A, 14A, 12A are preferred alternative restorations for Samples 16, 14, 12.

TABLE IV-4

Restored 1971 Erta Ale Analyses (mole %)

Sample	H ₂ O	H ₂	CO ₂	CO	SO ₂	S ₂	H ₂ S	COS
1032	69.41	1.57	17.16	0.75	9.46	0.59	1.02	0.02
1130	69.41	1.57	17.12	0.75	9.49	0.59	1.03	0.02
1131	69.45	1.57	17.92	0.79	8.77	0.50	0.85	0.02
1132	68.59	1.57	16.13	0.71	10.17	0.68	1.10	0.02
1133	68.60	1.57	17.58	0.77	8.83	0.52	0.97	0.02
1150	70.16	1.59	17.11	0.75	8.86	0.52	0.97	0.02
1151	68.36	1.55	18.88	0.83	8.87	0.52	0.94	0.02
1152	68.83	1.56	18.22	0.80	9.04	0.54	0.97	0.02
1153	70.88	1.60	17.47	0.77	7.95	0.41	0.88	0.01
1154	69.43	1.57	18.06	0.79	8.67	0.49	0.94	0.02
1155	68.72	1.55	18.81	0.83	8.63	0.49	0.92	0.02
1157	71.85	1.62	16.61	0.73	8.03	0.42	0.90	0.01
1214	69.55	1.57	18.10	0.80	8.53	0.48	0.92	0.01
1216	69.27	1.56	18.38	0.81	8.53	0.48	0.92	0.02
1224	69.99	1.58	17.93	0.79	8.31	0.45	0.91	0.02
1225	70.21	1.59	18.30	0.81	7.80	0.40	0.85	0.02
1226	66.88	1.51	18.57	0.82	10.29	0.70	1.07	0.02
1227	70.40	1.59	18.00	0.79	7.89	0.41	0.87	0.01

TABLE IV-5

Restored 1974 Erta Ale Analyses (mole %)

Sample	H ₂ O	H ₂	CO ₂	CO	SO ₂	S ₂	H ₂ S	HCL
910	77.24	1.39	11.26	0.44	8.34	0.21	0.68	0.42
911	76.05	1.57	12.52	0.56	7.59	0.31	0.93	0.42
912	79.09	1.57	10.60	0.46	6.86	0.21	0.76	0.42
914	79.01	1.56	10.65	0.46	6.90	0.21	0.77	0.42
916	75.06	1.59	13.08	0.60	7.84	0.36	1.01	0.42
920	78.28	1.48	10.92	0.45	7.51	0.21	0.72	0.42
921	79.83	1.73	10.09	0.47	6.27	0.25	0.92	0.42
922	78.66	1.69	10.47	0.48	6.18	0.22	0.85	0.42
929	79.16	1.71	10.47	0.49	6.50	0.27	0.95	0.42
930	81.06	1.57	9.49	0.40	6.22	0.16	0.66	0.42
931	77.52	1.68	11.36	0.53	7.12	0.32	1.01	0.42
932	76.52	1.66	11.94	0.56	7.46	0.35	1.05	0.42
933	77.87	1.61	11.38	0.51	7.02	0.26	0.88	0.42
934	77.47	1.53	11.51	0.49	7.47	0.25	0.82	0.42
935	77.45	1.57	11.53	0.51	7.37	0.26	0.86	0.42
936	78.19	1.76	11.22	0.55	6.46	0.31	1.05	0.42
937	76.76	1.70	11.93	0.57	7.16	0.35	1.08	0.42
938	81.04	1.49	9.55	0.38	6.37	0.13	0.59	0.42

The investigations of the 1971, 1973, and 1974 gas collections from Erta Ale lava lake support the conclusion that the gases exhaled from the lake are essentially at chemical equilibrium. High-temperature reaction equilibria were best preserved in the 1974 samples, which were taken with a stainless-steel collection tube. There is no evidence in the sets of restored analyses of significant short-term (minutes, hours) chemical fluctuations in the gases released from the lava lake. The O_2 fugacities of all the restored analyses are within a range of 0.20 log units when compared at the same temperature, and they are close to those of the QMF buffer from 1200° to 1000°C. S_2 fugacities are also nearly constant for all the restored gas analyses at a given temperature. These relationships suggest that the O_2 and S_2 potentials of the lava lake did not change appreciably over the period 1971 to 1974.

There is a marked change in the compositions of the restored analyses between 1973 and 1974, involving mainly a decrease in CO_2 and an increase in H_2O . These changes are qualitatively compatible with magma degassing controlled by solubility differences. The solubilities of H_2O , CO_2 , and sulfur as a function of depth (pressure) also appear to account for the composition difference between the Erta Ale gases and the CO_2 -rich vesicles of submarine basalts similarly associated with crustal spreading. By analogy with submarine basalts, H_2O and sulfur in the Erta Ale gases are contributed mainly by magma degassing at shallow depths (~50 to 250 m) while significant CO_2 contributions come from larger volumes of magma at depths of at least 1000 m.

Evaluation of 1977 Kilauea Eruption-Gas Collections--An investigation, similar to those described above, has been made of the Sandia gas collections from the September 1977 Kilauea eruption. A discussion of the corrections and interpretations of this data is presented under "Field Studies" later.

Organohalogen in Hawaiian Volcanic Gases--High concentrations of CH_3Cl and CH_3I have recently been reported for the atmosphere near a cooling 1977 Kilauea lava flow and for fumarole gases on Mauna Loa. These observations have prompted concern about the concentration levels of toxic

methylhalogens in Hawaiian volcanic gases. Thermodynamic calculations have been employed to investigate the stability of these compounds in Hawaiian volcanic-gas systems over a range of temperatures and pressures (Gerlach, 1978d). All results indicate that equilibrium methylhalogen concentrations in Hawaiian volcanic gases would be many orders of magnitude below the reported values (Table IV-6). The high concentrations of methylhalogens in active volcanic environments appear to be related to pyrolysis of plant material.

TABLE IV-6

Maximum CH_3Cl in Hawaiian Volcanic Gases (ppb, molar)

Sample	1 bar	10 bars	100 bars
1 ^a	3×10^{-10} (1100°C)*	1×10^{-8} (1200°C)	2×10^{-7} (1200°C)
2 ^b	3×10^{-11} (800°C)	1×10^{-9} (900°C)	7×10^{-8} (1200°C)
3 ^c	3×10^{-8} (1200°C)	7×10^{-7} (1200°C)	1×10^{-5} (1200°C)

NOTES:

- a. Based on Nordlie (1971): 13.73% C, 56.12% O, 20.84% H, 9.29% S, 0.01% Cl
- b. Based on Naughton et al (1974): 1.34% C, 34.93% O, 63.38% H, 0.33% S, 0.009% Cl
- c. Based on Graeber et al (1978) Sample No. 5: 1.71% C, 36.14% O, 60.20% H, 1.84% S, 0.12% Cl

*Temperatures in parentheses are those at which maximum calculated CH_3Cl concentrations are found.

Outgassing of Basaltic Magmas--Studies are in progress to model the C-O-H-S-Cl volatiles outgassing from basaltic magmas at 1 kbar to surface pressures. The computer codes discussed above, solubility data, and field data for concentrations of H_2O , CO_2 , Cl, and S in submarine basalts are being used to follow the compositional evolution of a gas phase separated from magma as a function of depth (pressure). Currently, the work is focused on mid-ocean ridge tholeiites, Hawaiian basalts, and oceanic-island alkali basalts. The preliminary results indicate that basalts have

gas phases richer in CO_2 and CO at higher pressures compared to basaltic volcanic gases. The gas phase also has relatively small amounts of sulfur at high pressure, due to the pressure effect on the solubility of sulfur in mafic silicate melts. SO_2 declines in importance as pressure increases; S_2 and H_2S increase in importance with pressure. Some examples typical of the preliminary results are given in Table IV-7 for submarine basalt bases at 1200°C and 2500 m (250 bars) water depth.

TABLE IV-7

Submarine Basalt Gases (mole %) at 1200°C , 250 bars

<u>Species</u>	<u>Mid-ocean Ridge Tholeiite</u>	<u>Hawaiian Submarine Basalt</u>
CO	4.56	2.62
COS	0.66	0.38
CO_2	79.96	45.97
H_2	0.25	0.96
H_2O	11.45	43.90
H_2S	1.07	4.11
SO	0.005	0.005
SO_2	1.27	1.27
S_2	0.73	0.73
S_2O	0.04	0.01
HCl	0.004	0.006

Field Studies

Kilauea 1975 Eruption--

General Description: The collection and analysis of gases being emitted from fissures following the eruption of a volcano provides information about one part of the diverse magmatic-volcanic environment to which engineering materials might be exposed. Such studies may lead to insight into the processes of gas-rock interaction above igneous intrusions.

Gas samples were collected from several gas vents located on fissures in the caldera of Kilauea volcano in November and December of 1975 by K. M. Htun, J. Larsen-Basse, and E. Herlicksa of the University of Hawaii and by M. Davis, E. J. Graeber, and P. Modreski of Sandia. The sampling locations were from the November 29, 1975 eruptive fissure several days after the eruption took place, and from fumaroles on older fissures which formed and erupted lava in 1971 and 1974 at the south end of the caldera floor and which have continued to emit gas and form sulfur deposits. Such gases are believed to have complex origins that may involve magmatic volatiles, ground water, air, and the interaction of these fluids with hot basaltic rocks underground.

The examination of the fissure vent of the November 29, 1975 eruption at Kilauea was particularly interesting. First visited on December 3 (4 d after the eruption), the downslope portion of the fissure was emitting hot gases that appeared to be nearly all steam. Rainfall at Kilauea had been quite light during this period, about 0.08 in. total from November 29 through December 3 and a total of about 0.025 in. over the next 3 d. A small amount of solid sulfur was being deposited around small cracks in the pahoehoe lava adjacent to the fissure: a probe inserted into the main fissure brought up globs of liquid sulfur. The decline in temperature at several locations along the fissure during a 3-d period of observation is given in Table IV-8.

Gas Collection Techniques: Gases were collected in Naughton-type sample tubes, which contain an evacuated volume plus a silica gel adsorbent to separate the polar gases and prevent their reaction after collection. Gas samples were collected by two techniques: first, holding the long Naughton tube in the gas vent and breaking a seal to allow the gases to fill the evacuated tube; and second, pumping the gases out of the fumarole to the Naughton tube via a hand-operated peristaltic pump attached to a tygon or teflon tube inserted into the fumarole. The use of a pump permits the collection of samples which are less contaminated by air during the collection, but this technique is complicated by a condensation of water in the teflon tube. Such condensates have been collected and analyzed along with the gases.

TABLE IV-8.

Temperatures Measured at Site of November 1975 Kilauea Eruption
(Temperatures in °C; time of day in parens)

<u>Location</u>	<u>12/4/75</u>	<u>12/5/75</u>	<u>12/7/75</u>
1975 fissure			
Vent A, broken bubble	240° (1400)	222° (1215)	204° (1120)
1975 fissure			
Vent B, tortoise	237° (1430)	205° (1245)	185° (1055)
1975 fissure			
Vent C, east end of fissure	211.5° (1445)	189° (1300)	174.5° (1130)
1975 fissure			
Vent D, west end of fissure	210.5° (1500)	184° (1325)	165° (1145)

Gas Analyses: A procedure was developed for analysis of the gas samples; this is summarized in Figure IV-4. Those gases in the Naughton tube not absorbed on the silica gel are transferred into another evacuated glass bulb for gas chromatographic analysis; the remaining gases and water are removed from the silica gel by heating under vacuum and condensing the gases in a bulb at liquid-nitrogen temperature. The condensed water and a water rinse of the residual silica gel are also analysed for dissolved ionic species. Table IV-9 presents the completed analyses of seven gas samples: four from a single location on the August 1971 eruptive fissure, one from a fumarole on the September 1974 fissure, and two from the November 1975 fissure. Comparison of the oxygen and nitrogen contents of Sample 6-2 to those of 6-1, 6-3, and 6-4 shows the superiority of the pumping technique at preventing atmospheric contamination during sampling. The lower water content of this sample, compared to that of 6-3 and 6-4, reflects loss of water by condensation in the teflon tubing. These fumarolic samples are dominantly $H_2O-CO_2-SO_2$ gases, with varying proportions of air. The gas being emitted from the 1974 fissure contained very little sulfur. This same fissure was examined in September 1977; only a small amount of warm, moist gas (probably air) was being emitted from the

fissure, still with no noticeable deposition of sulfur. This suggests that this crack did not maintain the open communication to deeper levels of the rocks within the caldera that the 1971 and 1974 fissures have.

Analyses of condensate samples collected in the field are given in Table IV-10. The small volumes of liquid collected in these samples prohibited analyses for additional dissolved species. The exact influence of water condensation on the composition of a gas sample could be determined if the composition and amount of condensate, the solubilities of the various gas components in water, the temperature of condensation, and the duration of pumping during which condensate was accumulating were all known. However, the last two of these factors are not accurately known for the present series of samples.

Interpretation of Analyses: Evaluating the results of the gas analyses can be complex and subject to various interpretations. The presence of molecular oxygen and nitrogen in the samples indicates admixture of air into the gases. We would like to determine whether this was incorporated subsequent to sampling, during sampling, at or near the vent of the fumarole as air mixed with the gases flowing out of the fissure, or at greater depth due to circulation of air or of ground water containing dissolved air through the porous basaltic rocks of the caldera floor. The low N_2 and O_2 contents of Sample 6-2 suggest that most of the N and O_2 reported in analyses 6-1, 6-3, and 6-4 represent contamination during sampling.

The SO_3 reported in the analyses may represent H_2SO_4 truly present in the fumarolic gases, or it may represent sulfate formed by the postcollection reaction of O_2 with SO_2 ; $SO_2 + 1/2 O_2 + H_2O = H_2SO_4$. The free-oxygen content in six out of seven samples is lower than that expected from the amount of nitrogen present, assuming that all the nitrogen is atmospheric; this indicates that some of the atmospheric oxygen did indeed react with SO_2 at some stage. If one compares the amount of oxygen anticipated on the basis of the nitrogen content to the combined total of free O_2 (oxygen required to produce the observed SO_3 by oxidation of SO_2), one sees a

close balance (Table IV-11); this suggests that the bulk of the observed sulfate was not present in the gas prior to incorporation of the air. The slight excess of oxygen calculated in this way relative to the expected atmospheric oxygen may correspond to that portion of the sulfate originally present in the gas.

Figure IV-4. Gas Samples: Collection, Treatment, and Analysis

TABLE IV-9

1975 Kilauea Gas Analyses (mole %)

Sample No. Lava Flow	6-1 1971	6-2 1971	6-3 1971	No. 1 1971	No. 3 1974	No. 2 1975 "A", Broken Bubble	No. 4 1975 "B", Tortoise
Vent	No. 6	No. 6	No. 6	No. 6	No Name	No Name	No Name
Method	Direct, hand-held	Pump	Direct, tied to stick	Direct	Pump	Pump	Pump
Condensate Collected?	--	No	--	--	Yes	Yes	Yes
Date	11/7/75	11/7/75	11/7/75	12/3/75	12/5/75	12/5/75	12/4/75
Time	10:06 am	2:10 pm	2:20 pm	~noon	3:30 pm	12:25 pm	2:50 pm
T, °C	~121	~121	~121	122	94	222	237
N ₂	1.47	0.090	2.60	1.05	0.39	3.47	2.23
O ₂	0.13	<0.005	0.39	0.015	<0.002	0.99	0.60
Ar	NA	0.0017	0.020	~0.01	0.0019	0.016	0.018
H ₂	NA	0.0006	0.0003	<0.0001	0.0009	0.0007	NA
CO ₂	2.32	5.62	1.81	2.60	14.5	0.53	0.20
CO	0.009	<0.005	0.004	<0.001	0.002	<0.002	<0.002
SO	0.30	3.59	0.039	0.20	0.99	<0.002	<0.002
H ₂ O	Assume 95	90.6	94.6	95.6	83.9	95.0	96.8
HCl	0.094	0.010	0.013	0.007	0.003	0.012	0.006
HF	NA	<0.0002	<0.0003	<0.0003	<0.0004	0.0007	0.0008
SO ₃	0.65	0.11	0.51	≥0.54	0.23	0.13	0.12
NH ₃	A	tr(~0.004?)	tr(≥0.016?)	tr(≥0.017?)	<0.008	tr(~0.009?)	tr(~0.007?)
S	0.025	NA	NA	NA	NA	NA	NA
H ₂ S	NA	0.004	NA	NA	<0.004	<0.002	<0.001

NA means Not Analyzed. <0.00X means that the component was not detected in the analysis and its concentration is assumed below the estimated level of detectability, 0.00X. The estimates for NH₃ are based on a qualitative test only and the numerical values are only gross estimates. The value of S for Sample 6-1 represents solid sulfur remaining in the Naughton tube after removal of the gases; it may represent sulfur dust mechanically scraped up or sucked into the tube during sampling of the fumarole. The amount of water in Sample 6-1 was not determined; it is arbitrarily assumed to be 95.0 mole %, by comparison with 6-3 and 6-4, and the other components are calculated accordingly.

TABLE IV-10

Analysis of Water Condensates From Kilauea

Sample	Sample Volume, (cm ³)	pH	ppm SO ₂	ppm SO ₄ ⁼	ppm Cl ⁻	ppm F ⁻
1974 fissure hand-pumped 12/5/75	12	3-4	>3100	265	8.4	<0.5
1975 fissure tortoise hand-pumped 12/4/75	4.5	~7	420	<100	4.6	NA
1975 fissure tortoise hand-pumped 12/5/75	10	5-6	460	<100	2.9	6.9
1975 fissure broken bubble hand-pumped 12/5/75	9	5-6	310	<100	10.3	6.4
1975 pahoe-hoe small cracks in flow 12/5/75	32.5	5-6	NA	<100	12.8	0.9

NA - Not Analyzed

TABLE IV-11

Kilauea Gas Analyses:
Comparison of Free O₂ Plus Sulfate Oxygen
to Expected Atmospheric Oxygen

Sample No.	6-1	6-2	6-3	No. 1	No. 2	No. 3	No. 4
Measured N ₂	1.47	0.090	2.60	1.05	3.47	0.39	2.23
N ₂ x 0.26825	0.394	0.024	0.697	0.282	0.931	0.105	0.598
Measured Free O ₂	0.13	<0.005	0.39	0.015	0.88	<0.002	0.60
1/2 SO ₃	0.325	0.055	0.255	≥0.270	0.065	0.115	0.060
Free O ₂ + 1/2 SO ₃	0.455	0.055	0.645	0.285	0.945	0.115	0.66
Free O ₂ + 1/2 SO ₃ - 0.26825 N ₂	+0.061	+0.021	-0.052	≥0.003	+0.014	+0.010	+0.062

(0.26825 is the atmospheric O₂:N₂ ratio)

The ratio of argon to nitrogen in the analyses is, in five out of six samples, lower than the atmospheric ratio of 0.0120. Since other studies have generally indicated that both magmatic volatiles and the gases dissolved in ground water have a higher Ar:N₂ ratio than air, this may indicate that the analytical determinations of argon are low.

The observed concentrations of CO and H₂ in the gases indicate a very reducing environment, if one assumes equilibrium between CO and CO₂ and between H₂ and H₂O at the temperatures of collection. The calculated oxygen fugacities would be in the range of 10⁻⁶⁰ to 10⁻⁴⁷ bars at 400 K.

Such oxygen fugacities are, however, compatible with those that would be expected for gases in equilibrium with basaltic rock; for example, the f_{O_2} of the QFM assemblage, a good model for basalt, is about 10⁻⁵⁵ bars at 400 K. The observed CO and H₂ are therefore consistent with an origin by interaction between water and basalt, either by magmatic degassing or through lower temperature interactions with solid rock.

The observed CO and H₂ concentrations are, however, not compatible with the measured SO₂ content of the gases. The equilibria $3H_2 + SO_2 = H_2S + 2H_2O$ and $3CO + H_2O + SO_2 = H_2S + 3CO_2$ should lie far to the right, with H₂ and CO partial pressures of about 10⁻⁹ and 10⁻¹⁴ bars, respectively. The most reasonable interpretation of this is that the CO and H₂ contents reflect an earlier state of the gas, and that these species are not in equilibrium with the other sulfur and oxygen species in the gas in the vicinity of the vent. Solid or liquid sulfur was observed at all the fumaroles and calculation of f_{O_2} , assuming equilibrium in the reaction $S(5 \text{ or } 2) + O_2 = SO_2$, yields oxygen fugacities still quite low but somewhat more oxidizing (around 10⁻⁴¹ bars at 400 K) than those derived from the CO and H₂ data.

Hydrogen sulfide was not detected in any of the samples, but the accurate determination of this species in volcanic gas samples is poor, due to reaction of H₂S with metals, with oxygen, and particularly with SO₂. The deposition of sulfur around fumaroles has been attributed to

the reaction between H_2S and SO_2 ($2\text{H}_2\text{S} + \text{SO}_2 = 3\text{S} + 2\text{H}_2\text{O}$) with falling temperature. Hydrogen sulfide fugacities calculated from this reaction, assuming equilibrium between solid or liquid sulfur and the observed partial pressures of SO_2 and H_2O , are in the range of 10^{-14} to 10^{-2} bars at collection temperatures. Concentrations at the lower end of this range could have escaped detection, particularly if the collected gases did react to some extent.

Kilauea 1977 Eruption--

General Description: During the period September 20 - October 4, 1977, Sandia conducted field studies on the east zone of Kilauea volcano. The eruptive phases started on September 12 with deflation of the summit area, accompanied by harmonic tremor and shallow earthquakes in the upper east rift zone. On September 13, fountains occurred along the fissure system between Kalualua and Puu Kaua, increasing in height to 25 to 40 m the following day. Flows from this vent area moved 3 km east and the activity continued sporadically until September 19.

On September 18, a new vent began erupting 500 m west of Kalualua. Fountain heights were generally 30 m or less, causing viscous slab pahoehoe and aa flow, which moved slowly northeast. Eruptive activity ended at this vent by late morning on September 20. On the following day, a gas sample was obtained from a crack in the lava flow (955°C was measured in the crack). Three additional gas samples were taken from the interior of the spatter cone, which was still emitting red-hot gases on September 23.

The third and most voluminous phase of the eruption began on September 25. Fountaining up to 60-m high occurred along a 300-m fissure situated 1 km west of the September 13 vent; it rapidly built up a spatter cone. A flow to the east of the cone branched into a major channel, advancing at 300 m/h to the southeast towards Kalapana. Fountaining at this major cone (later named Puu Kia'i) continued until October 1. Temperature measurements of gas vents and flows, collection of gas samples, testing of material compatibility samples, and thermal measurements from a heat-

transfer probe were fielded during this vigorous eruptive phase. The flow moved rapidly toward Kalapana until September 30, at which time it spread laterally northwest of the village. The 10-km-long flow stopped within 700 m of the nearest house in its path. A total of $35 \times 10^6 \text{ m}^3$ of highly differentiated tholeiitic basalt were erupted, covering an area of 8 km^2 .

Gas Collection Sites: A description of the collection sites is given in Table IV-12. Three gas samples (Nos. 5, 6, and 7) were taken from a hot conduit, and three (Nos. 1, 10, and 11) from lava flows. It is note-worthy that in one sample (10) some gas products from pyrolysis of vegetation were suspected when the advancing pahoehoe overran an older 1963 flow. Temperatures ranged from 955° to 1062°C ; where gas temperature measurements could not be made directly, the temperature was assumed to be that measured by a calibrated Inconel-sheathed, Chromel-Alumel thermo-couple emplaced in the flow directly beneath the collection site. The highest lava temperature measured during the eruption was 1095°C , using a thermo-couple immersed in the margin of the flowing lava river a few hundred feet from the fountaining vent on September 30, 1977.

TABLE IV-12

Collection Sites

Sample No.	Site Description
1	Crack in lava flow west of Kalualua. 9/21/77, 12:45, 24 h after end of 9/20-phase of eruption; 955°C
5, 6, 7	Vertical conduit on NE side of spatter cone west of Kalualua; this spatter cone was the main vent of the 9/18-9/20 eruptive phase. 9/23/77, 14:36, 14:41, 14:49; 1000°C (estimated temperature within conduit)
10	Crack in old lava flow beneath advancing pahoehoe lobe. Pyrolysis of plants on old flow suspected. 9/27/77, 11:16; 1062°C (lava temperature)
11	Surface of advancing pahoehoe lobe after removal of surface crust to expose glowing lava. 9/27/77, 11:20; 1062°C (lava temperature)

Gas Collection Techniques: The gases were sampled with a collection device consisting of an evacuated metal tube and a Pyrex bottle with a volume of approximately 250 ml (Figure IV-5). One end of the bottle contained a break seal housed in glass tubing, which served as a connector to a gas chromatograph/mass spectrograph analytical instrument. The other end was fabricated with 1.27-cm-OD thin-walled (0.25-mm) platinum tube and break-off seal. A series of graded-glass seals were used to approximate the thermal expansivities of platinum and Pyrex, thereby minimizing cracking from thermal shock when the device was immersed in a high-temperature gas stream.

Figure IV-5. Configuration of Volcanic Gas Collection Bottle

In operation, the device is supported in a metal pole and is inserted into a gas vent or suspended above a lava flow. The end seal is broken by pulling on a coiled Nichrome wire attached to the break-off seal. After collection the platinum tube is cold-welded by a special pliers.* The entire procedure can be easily executed in less than 10 s. Simulated laboratory tests with this type of collector have demonstrated that the inside bottle temperature is still increasing at the time of welding;

* Carl Hermann and Associates, Inc. Model PCD-375.

hence, there is an inside bottle pressure to minimize additional air intake. This increasing temperature was also observed in collection devices by Finlayson et al (1968).

Gas Analyses: All of the gases sampled were analyzed by both gas chromatography and mass spectroscopy. The sample bottles were then rinsed with ethanol to remove and extract condensates. Equal aliquots of this rinse were used for wet chemical analysis of total sulfur, sulfate, chloride, and water. The volume of each collector was determined by weighing and measuring water displaced from the bottle/tube assembly. In Table IV-13 the analytical results (in mole %) are given for main constituents of the gases. All chloride is reported as HCl. No SO₂ was detected in the gas phase of the samples; sulfur measurements did not reveal the presence of any significant amounts of reduced or elemental sulfur.

TABLE IV-13

Analyses (mole %) of Gases Collected From
September 1977 Eruption of Kilauea Iki

Sample No.	N ₂	O ₂	Ar	H ₂ O	H ₂	CO ₂	CO	SO ₂	HCl ⁺	Ar/N ₂	O ₂ /N ₂
1	42.01	6.57	0.11	38.27	3.07	8.88	0.48	0.06	0.49	0.003	0.156
5	75.78	20.10	0.90	2.73	ND	0.16	ND	0.17	0.01	0.012	0.265
6	77.61	17.69	0.83	3.73	ND	0.08	ND	0.03	0.01	0.011	0.228
7	67.84	17.91	0.86	12.54	ND	0.25	ND	0.59	0.02	0.013	0.264
10*	64.67	14.94	0.79	12.59	1.39	2.79	1.75	0.09	0.05	0.012	0.231
11**	74.63	17.98	0.91	4.10	0.33	2.01	ND	0.03	0.01	0.012	0.241

⁺ All chloride in condensate reported here as HCl

ND: Not detected

* Also contains 0.49% CH₄, 0.04% C₂H₆, 0.28% C₂H₄, 0.05% C₃H₈, 0.03% C₃H₆, 0.03% C₄H₁₀, 0.02% C₆H₆, 0.01% C₆H₅CH₃

** Also contains 0.012% CH₄

Interpretation of Analyses: The analyses generally have Ar/N₂ and O₂/N₂ in excellent agreement with atmospheric ratios (0.012 and 0.270, respectively), suggesting that N₂, O₂, and Ar are mainly the result of air contamination. The low Ar/N₂ of Sample 1 and the low O₂/N₂ in Samples 1,

oxidation of plant material as discussed below. A recalculated set of analyses without N₂, O₂, and Ar is given in Table IV-14.

TABLE IV-14

Analyses (mole %) of Gases Collected From
September 1977 Eruption of Kilauea Iki After Removal of N₂, O₂, and Ar

Sample No.	H ₂ O	H ₂	CO ₂	CO	SO ₂	HCl ⁺	S/C
1	74.55	5.98	17.30	0.94	0.11	0.96	0.006
5	89.10	ND	5.06	ND	5.45	0.35	1.08
6	96.86	ND	1.99	ND	0.84	0.20	0.42
7	93.58	ND	1.87	ND	4.42	0.12	2.36
10*	64.20	7.09	14.23	8.92	0.46	0.25	0.014
11**	63.19	5.09	30.93	ND	0.46	0.21	0.015

⁺All chloride in condensate reported here as HCl

*Also contains 2.50% CH₄, 0.20% C₂H₆, 1.43% C₂H₄, 0.25% C₃H₈, 0.15% C₃H₆, 0.15% C₄H₁₀, 0.10% C₆H₆, 0.05% C₆H₅CH₃

**Also contains 0.18% CH₄

ND: Not detected

Because of the high air content (48.7 to 96.8 mole %) of the analyzed samples, the content of atmospheric water and even of CO₂ becomes significant in evaluating the composition of the volcanic component of the samples. Assuming that the ambient atmosphere contained 50% relative humidity at 25°C, the observed quantities of included air would have introduced between 0.9 and 1.7 mole % H₂O to the samples. This is equivalent to a significant fraction of the water reported in samples 5, 6, and 11; therefore, the apparent water-rich composition of these gases must be accepted with some caution. The CO₂ content of the air would account for approximately 0.02 to 0.03 mole %, which is an appreciable fraction of the amount of CO₂ reported in Sample 5.

The high proportion of atmospheric gases was not unusual in the samples collected from the surface of the lava flows, but the high air content of the samples from the conduit was unexpected. The conduit samples were collected from a point about 2 m below the lip of a cylindrical, vertical lava conduit within the spatter cone west of Kalualua. At the time of sampling, the interior of the conduit was glowing orange-red and there was evident emission of hot gas from the vent, though not under great pressure. Pale-red flames flickered around the mouth of the vent, but the appearance and behavior of the flames in the conduit suggested an open communication with degassing magma at depth. Additional eruptions took place within about 2000 m of this site during the following week. It is evident from the analyses that convective circulation of air into the open mouth of the conduit, or through the fracture system at greater depth, was quite extensive. Samples less contaminated by air could probably be recovered if a long tube were inserted into a degassing vent to conduct the gases to the sampling point at the surface.

The recalculated analyses point up marked differences between samples taken from the conduit (5, 6, and 7) and those taken from lava flows (1, 10, and 11). The latter flows tend to be poorer in H_2O , richer in CO_2 , deficient in SO_2 , and have very low atomic S/C. The flow samples are rich in H_2 and CO, and Sample 10 contains significant amounts of various hydrocarbons. The analyses for the flow samples are very different from their equilibrium distributions at all temperatures. Their equilibrium distributions contain almost no hydrocarbons and have oxygen fugacities one to three orders of magnitude less than QMF at temperatures near the collection temperature. They are much too reduced for the atomic composition limits proposed for basaltic volcanic gases by Gerlach and Nordlie (1975).

The above characteristics and the tendency of the flow samples to have O_2/N_2 and Ar/N_2 less than atmospheric suggests contamination by plant material. This possibility has been investigated for the atomic compositions of the analyses after removal of SO_2 and HCl. The C-O-H atomic compositions for all analyses are shown in Figure IV-6, along with the

tie-line between the average composition of the conduit samples and common cellulose ($C_6H_{10}O_5$). The tie-line gives approximate compositions of mixtures resulting from plant contamination, assuming that Samples 5, 6, and 7 are reasonably representative of the atomic composition of the volcanic gases. Sample 10, which is the most reduced sample and contains the largest proportion of hydrocarbons, falls close to the tie-line. The interpretation favored here is that contamination by plant pyrolysis led to disequilibrium gas mixtures along the tie-line. Progressive oxidation of the pyrolyzed plant material produced gas mixtures with compositions like those of 1 and 11. We suggest that hydrocarbons, other than small quantities of CH_4 , observed in previous volcanic gas studies (e.g., Muenow, 1973) have a similar origin.

Figure IV-6. The C-O-H Atomic Compositions of Analyses of Samples From Flows (squares) and a Vent (circles) Are Shown After Removal of SO_2 and HCl . The triangle gives the bulk composition of common cellulose ($C_6H_{10}O_5$). The dashed line shows compositions formed from mixtures of cellulose and the average of the vent analyses.

The analyses for Samples 5, 6, and 7 show no evidence of plant contamination. The small amounts of H_2 and CO normally present in volcanic gases have apparently been oxidized by air in these samples, either before or after collection. These analyses represent the atomic compositions of the volcanic gases from the conduit, except for slight gains in oxygen. The preoxidized compositions of the volcanic gases can be estimated from the atomic C, H, and S of the analyses and the oxygen fugacity of Hawaiian basalts. The f_{O_2} of the basalt at the collection temperature (ca. 1000°C) was assumed similar to that measured for basalt in Makaopuhi lava lake (Sato and Wright, 1966). Sato and Wright data give a mean $\log f_{O_2}$ of -10.8 at 1000°C, slightly greater than that of the QMF buffer. The estimated volcanic gas compositions based on Sample 5, after removal of small amounts of oxygen sufficient to give $\log f_{O_2}$ values of -10.3, -10.8, and -11.3 at 1000°C, are shown in Table IV-15. Similar results were obtained for Samples 6 and 7 and for the mean f_{O_2} of Sato and Wright at 1100° and 900°C.

TABLE IV-15

Estimated Volcanic Gas Compositions at Spatter Cone
Based on Sample 5 and the Oxygen Fugacity Data of Sato and Wright
(1966) at 1000°C

$\log f_{O_2}$	H_2O	H_2	CO_2	CO	SO_2	H_2S	S_2	HCl
-10.3	88.47	0.66	5.00	0.06	5.35	0.09	0.01	0.35
-10.8	87.99	1.17	4.98	0.11	4.86	0.45	0.08	0.35
-11.3	87.20	2.06	4.97	0.19	3.20	1.66	0.34	0.36

The estimated volcanic-gas compositions shown in Table IV-14 are H_2O -rich, a feature suggesting the gases may have been contaminated by sub-surface water. The high humidity of the Hawaiian atmosphere may also have contributed to the H_2O -rich compositions. On the other hand, petrographic examination of the 1977 lava indicates it was differentiated (Eaton, 1977) and had presumably been emplaced under Kilauea from some time before eruption. It is therefore possible that the lava had degassed less

soluble constituents (CO_2 and sulfur) at depth, leading to H_2O -rich volcanic gases after eruption.

Kilauea Iki Drill Core Study--Drillhole Kilauea Iki 76-2 was part of the 1976 Drilling Experiments described earlier under Task 2. One of the drill bits plugged at 44 m (145 ft) with solidified melt was brought back for analysis (Figure III-9). Microscopic and microprobe examination of a sample of banded glass from the drill bit showed it as a mixture of rapidly quenched glass and devitrified glass, each having the same chemical composition. Most of the sample was composed of quenched glass, showing no evidence of crystallization whatsoever. The devitrified bands, constituting about 20 to 25% of the sample, contain phenocrysts of 50 to 120 μ olivine, 50 to 60 μ clinophyroxene (augite), and 100 to 150 μ plagioclase (bytownite) in a fine-grained cryptocrystalline groundmass.

Thermal expansion curves were determined for both the quenched and the devitrified glasses. As shown on Curve A (Figure IV-7), the expansion of the devitrified glass was fairly uniform up to 950°C. Curve B (Figure IV-7) shows that the rapidly quenched glass goes through a transition point at 575°C, a softening point at 670°C, and continues to melt to the terminal temperature of the experiment at 750°C. Both rocks have the same chemical composition; however, the glass represents a quenched nonequilibrium solid that, if allowed to cool under equilibrium conditions, would have resulted in a basalt of aphanitic texture.

Laboratory Studies

Introduction--In accordance with the recommendation of the Magma Workshop (Colp et al, 1975) and the advice of Sandia's Magma Advisory Panel, it has been decided to concentrate on experiments to measure the effects of pressurized volatile atmospheres on the physical properties of molten rocks and the selected aspects of the chemical equilibria between silicate melt and a gas phase. Most of the properties of molten rock that are of interest have at least approximately been determined at atmospheric pressure in the laboratory or in the field, but relatively little work has been done to date at the elevated pressures required to maintain dissolved volatiles in the melts.

Figure IV-7. Thermal Expansion of Kilauea Iki Lava Lake Basalt (145-ft depth, August 1976)

The experimentation at Sandia will be done at high pressures (to 4 kbar) and at atmospheric pressure. The atmospheric pressure work will be done mainly to test the experimental techniques used at elevated pressures and to provide baseline data for comparison with the results of the high pressure studies. A number of areas remain in which careful measurements at atmospheric pressure alone would be of real value; for example, studies of viscosity and other physical properties of multiphase magma systems (melt plus suspended crystals or gas bubbles).

Apparatus Description--The following types of apparatus have been set up at Sandia for experimentation on lavas at atmospheric and elevated pressures:

Atmospheric Pressure Gas-Mixing System. A horizontally mounted tube furnace through which gases circulate within a mullite tube is being used

primarily for materials compatibility experiments (Figure IV-8). Mixtures composed of CO_2 , CO , H_2O , and SO_2 are being used; other gases, such as H_2 , H_2S , and HCl , could be added as desired. The maximum temperature is 1200°C ; the length of the isothermal zone ($\pm 5^\circ\text{C}$) within the 5.08-cm-ID tube is about 13 cm. The gases are metered through calibrated flowmeters, with the exception of water; the flow of water is calculated from the vapor pressure of water in a constant temperature bath through which argon or one of the other gases is bubbled as a carrier. Lava samples for the compatibility tests are usually contained in high purity (99.8% Al_2O_3) alumina crucibles, which are relatively nonreactive to the lava, for runs of 10- to 100-h duration. Due to solution of Al_2O_3 in the lava, such crucibles are not suitable for careful-phase equilibrium studies on the lavas. This system has at present no provision for rapid quenching, and the samples are cooled fairly slowly (about 4°C min^{-1}) by shutting off the furnace power. The system is not equipped with an oxygen fugacity probe, and the concentrations and activities of gas species at temperature are calculated from the input mixture assuming equilibrium conditions.

Figure IV-8. Controlled-Atmosphere Test System

Atmospheric Pressure Rapid-Quench System. A 1250°C furnace has been set up for experiments on lava or other samples in a controlled gas atmosphere (Figure IV-9). CO_2/CO , CO_2/H_2 , or similar gas mixtures flow through a mullite tube within a vertically mounted tube furnace; lava samples are suspended in the gas stream by hanging a drop of melt from a loop of fine platinum wire (about 0.125-mm-dia) attached to a quartz rod. This technique facilitates rapid equilibration between the lava and the gas atmosphere, and minimum contamination of the sample by reaction with the container material; some loss of iron from the sample does occur by reduction of iron and diffusion of Fe^0 into the platinum wire. The samples are quenched rapidly in air by pushing the quartz rod carrying the samples out through the bottom of the furnace. An air reference ZrO_2 solid electrolyte oxygen fugacity electrochemical cell is incorporated within the furnace for direct measurement of oxygen fugacity adjacent to the sample, as a check on that calculated for a given gas mixture at equilibrium. It is intended that this system generally be kept "clean," using only $\text{CO}/\text{CO}_2/\text{H}_2/\text{H}_2\text{O}$ gas mixtures without any sulfur-containing gases.

Figure IV-9. Rapid-Quench System

The main use of this apparatus is for phase equilibrium studies of the lavas (melting range, crystalline phases present at equilibrium, etc); a controlled-gas atmosphere is essential to regulate oxygen fugacity in these experiments. These data will provide baseline information for comparison with phase relations in the presence of volatiles at higher pressures, and for planning and interpretation of the physical properties measurements which are sensitive to the degree of crystallization of the lava. This apparatus will also be useful for atmospheric pressure duplication of some of the physical-properties measurements to be made at high pressures, such as electrical conductivity and viscosity.

Cold-Seal (1-kbar) Pressure Vessels. Two TZM alloy (Mo-0.5% Ti-0.08% Zr) cold-seal pressure vessels are presently set up for phase equilibrium experiments on lavas plus volatiles under pressure, and for initial compatibility tests in pressurized gas atmospheres. The vessels of relatively small internal volume (6.4 mm dia x 286 mm total length) are externally heated in a resistance furnace and pressurized with argon. Maximum operating temperature is 1100°C; the maximum operating pressure in the present setup is 1000 bars, although the system could be replumbed to achieve the upper operating limit of the pressure vessels, 3000 bars at 1100°C. An Inconel sheath purged with N₂- 2% H₂ protects each vessel from oxidation. The system is currently pressurized by condensing argon into a pressure reservoir at liquid-nitrogen temperature and warming the reservoir to produce the desired gas pressure, but a diaphragm-type argon compressor is being added to the system. Temperature is measured with a thermocouple either inserted internally or attached to a well in the outside of the sheath; the external readings are related to the internal temperature by calibration runs. The length of the hot zone (+10°C) within the vertically oriented vessels is about 3.5 cm.

One of the vessels is presently equipped with a "Shaw membrane" to control oxygen fugacity within water-containing capsules via diffusion of hydrogen through a platinum tube connected to an external reservoir of hydrogen at known pressure (P_{H₂} typically about 3 to 15 bars) (Figure IV-10).

Figure IV-10. Shaw Membrane: H_2 Fugacity Control

Experiments are normally run with solid and gaseous reactants enclosed within thin-walled capsules of gold (below 1050°C) or $\text{Ag}_{50}\text{Pd}_{50}$ tubing. Volatiles can be added to the experiment capsules in the form of solid or liquid phases which vaporize or decompose upon heating (e.g., water, oxalic acid, silver oxalate, sulfur, silver sulfate) to give the desired atmosphere. The small volume available puts a limit on the amount of volatile material that can be sealed within a capsule to maintain a gas phase of constant composition during the experiment, particularly with regard to depletion of sulfur by reaction with metals in materials capability experiments. This problem can be overcome to some extent by using oxide-phase assemblages to buffer oxygen fugacity (e.g., $\text{Fe}_2\text{SiO}_4\text{-Fe}_3\text{O}_4\text{-SiO}_4$, Ni-NiO , $\text{FeO-Fe}_3\text{O}_4$, etc), and sulfide phases such as pyrrhotite, Fe_{1-x}S , to buffer sulfur fugacity.

Four-kbar Pressure System (Magma Simulation Facility). The Magma Simulation Facility (Figure IV-11) consists of a 4-kbar large volume

(18-cm-ID x 40-cm-length) internally-heated pressure vessel built by Autoclave Engineers. The vessel is pressurized with argon using a Harwood DA-4 two-stage compressor. An internal furnace is now being built for the vessel by Autoclave Engineers; this will be a 1600°C molybdenum single-zone resistance furnace employing a convective heat-transfer design to heat a 10-cm-dia x 10-cm-long hot zone ($\pm 10^\circ\text{C}$) located above the heating elements. The system is located in a blast cell (Bldg 9938) and will be operated from an adjacent control room (Bldg 9939) because of the blast hazard of the large volume of pressurized argon. An on-line computer will be used to collect and process data and to control the operating parameters and safety interlocks of the system. The computer is a Hewlett-Packard 2171A with 64-kbyte memory, a 15-mbyte disc, and RTE-III operating system software; it is interfaced with an HP-9648A CRT/graphics terminal, a Tektronix 4632 video copy unit, an HP-9866A thermal printer (240 lpm), and an HP-9603R remote measurement and control station with digital and analog input/output channels.

Figure IV-11. Magma Simulation Facility Schematic

The relatively unique features of this system are the large volume, permitting both large samples and the use of a large number of electrical measurement circuits or mechanical devices for simultaneous measurements within the pressure vessel; the availability of the computer for data handling; and the planned incorporation of a system for introducing (and sampling or replenishing) multicomponent gas mixtures to an interior chamber within the pressure vessel. The furnace will include ports for pressure lines which can be used for the introduction or removal of gases (Figure IV-12). Full operation with multicomponent gas mixtures will require

- A suitable inert chamber within the furnace to contain the "reactive" gas mixtures and to isolate them from the argon which fills the rest of the vessel. A molybdenum chamber with a liner of platinum or a similar noble-metal alloy might be suitable.
- A gas-delivery system to mix, compress, and supply complex gas mixtures to the pressure vessel. A magnetically driven pump suitable for circulating gases at high pressure through such a chamber has been designed.
- A pressure-balancing cell to maintain the pressure differential between the internal chamber and the rest of the pressure vessel at a tolerable value (e.g., up to a few tens of psi).

Once the internal chamber is operating, it can be used for

- A continuous supply of controlled gas atmospheres to buffer the activities of volatile species over simulated magma samples
- Compatibility experiments of long duration
- Experiments on the interactions between introduced gases and liquid rock (e.g., phenomena of bubble formation, fluid injection, magma fuels generation, etc).

Figure IV-12. Schematic of Containment Vessel for Reactive Gas Atmospheres Within 4-kbar Pressure Vessel

Prior to operation of the sealed chamber, experiments on volatile-containing melts will be made using samples within sealed noble-metal tubes, the conventional technique for hydrothermal experimentation. It should be possible to insert a Shaw membrane within the vessel to control hydrogen fugacity, and also operate a ZrO_2 electrochemical-oxygen potential cell with a solid buffer internal reference to measure oxygen fugacity within the vessel or within the internal chamber. The temperature capability of the system ($1600^\circ C$) will permit measurements on magma both within the crystallization interval and for several hundred degrees above the liquidus temperature. This will allow activation energies to be calculated from the temperature dependence of a property over a significant temperature interval within the completely liquid region.

Experiment Plans--The initial experiments will consist of work on phase-equilibria and gas-solubility studies, viscosity, and electrical conductivity. The experiments will be done on a series of igneous rock

compositions, which include basalt (1971 Kilauea tholeiite, 50 wt % SiO_2); andesite (Mount Hood andesite, 61 wt % SiO_2); and rhyolite (Newberry Caldera rhyolite obsidian, 73 wt % SiO_2). The proposed plan of experiments was discussed with C. W. Burnham (Penn State), I. S. E. Carmichael (Berkeley), J. Holloway (Arizona State), T. Shankland (LASL), H. Shaw (USGS), H. Waff (University of Oregon), P. J. Wyllie (University of Chicago), and H. S. Yoder (Carnegie Institution) among others.

Experiments on molten rocks will be carried out at atmospheric pressures of up to 1 kbar in cold-seal pressure vessels and up to 4 kbar in an internally-heated pressure system. Equilibrium-melting relations of a basalt, an andesite, and a rhyolite will be determined to provide necessary data at atmospheric pressure and at elevated pressures with water and other volatile components present. The solubilities of volatile components of interest (H_2O , CO_2 , $\text{SO}_2/\text{H}_2\text{S}/\text{S}_2$, HCl , HF) will be studied, along with the phase equilibria. The first physical properties measured will be viscosity (using the falling-sphere method) and electrical conductivity, in the presence of some of the above volatile components; the same properties will also be measured at atmospheric pressure for comparison. Some of the viscosity and conductivity measurements will begin as soon as the 4-kbar system is operation, and will proceed at the same time as the melting-equilibria, gas-solubility, and materials-compatibility experimentation. A number of additional important physical parameters exist which can be the subject of future experiments.

Magma/Material Compatibility

Introduction

A significant uncertainty in considering the feasibility of using magma energy is whether engineering materials used in exploration, development, and production can survive and function in the various environments that exist, from the molten rock source up the borehole to the surface. These environments can be divided roughly into three zones:

- MAGMA

Conditions - High temperatures (600° to 1250°C), reducing atmosphere, high sulfur content.

Problems - Sulfidation corrosion, thermal-strength degradation, metal dissolution.

- HYDROTHERMAL

Conditions - Moderate temperatures (150° to 500°C), oxydizing atmosphere, corrosive hydrothermal-brine content.

Problems - Oxidation, brine corrosion.

- NEAR SURFACE

Conditions - Low temperatures (<150°C), oxydizing atmosphere, water-rich gas content.

Problems - Oxidation corrosion.

To date material compatabilities have been studied at atmospheric pressure only under several conditions simulating parts of the expected environments. Studies have been divided into two groups:

- Field Exposures:

Fumarole gases (low temperature) at Kilauea Volcano

Hot gases at volcanic vents and lava flows

- Laboratory Exposures:

Degassed basalt, andesite, and rhyolite lava in an inert (argon) atmosphere

Basalt lava plus simulated magmatic-gas atmospheres.

Most of the experimentation has been made on metals (pure elements and iron-and-nickel-base alloys) but some carbides, ceramics, and other compounds have also been examined.

Experiment Descriptions

Field Experiments--Field exposures have been made in gases at fumaroles at Kilauea Volcano, Hawaii, for periods from a few days to several months. The gases at these low-temperature (near 100°C) vents have a

complex history involving degassing of magma; interactions with cool, solid rocks; and mixture with groundwater and possibly air. The gases are rich in H_2O , CO_2 , and SO_2 , with traces of H_2S ; many of the fumaroles are depositing solid or (below-ground level) liquid sulfur. Materials behavior in these gases may be relevant to the environment that would be encountered at the upper, cooler levels of a heat-exchange system tapping magmatic heat.

Materials have also been exposed to much hotter gases (800° to 1000°C) during the September 1977 east rift zone eruption of Kilauea. The gases at these sites of exposure, however, appear from our analyses to have contained a large amount of admixed air due to convective circulation within the volcanic vents, making the environment more oxidizing and not representative of the uncontaminated magmatic volatiles. Attempts to immerse and recover materials from molten lava at the 1977 eruption were mostly unsuccessful; only one very short (10 min) exposure was recovered from the lava.

Compatibility testing in the natural environments at the volcano and in laboratory experiments has also been carried out by M. K. Htun and J. Larsen Basse of the Department of Mechanical Engineering, University of Hawaii at Manoa, under Sandia support.

Laboratory Experiments--Laboratory exposures have been designed to include the most severe environments that might be encountered in extracting energy from magma--direct exposure to uncooled, gas-containing molten magma. Actual exposure conditions are likely to be less severe, involving either internally cooled metal heat exchangers, which may also be protected by a surrounding shell of solidified magma, or contact with hot steam and gas in pipes under conditions which are more similar to conventional environments in turbines and power-generating systems. The worst conditions might only occur in drilling operations and as transient exposures.

In the tests on metals plus degassed lavas, metal coupons were covered with granular lava in alumina crucibles and heated at 1200°C for

about 100 h. Basalt (1971 Kilauea tholeiite, 50 wt % SiO_2 , 12 wt % FeO_x), andesite (Mount Hood andesite, 61 wt % SiO_2 , 6 wt % FeO_x), and rhyolite (Newberry Caldera rhyolite obsidian, 73 wt % SiO_2 , 2 wt % FeO_x) lavas have been studied. The samples were contained in an alumina or mullite tubes, which were evacuated and then purged with argon.

A similar arrangement was used in the tests with simulated magmatic gas atmospheres, except that a CO_2 - CO - SO_2 - H_2O gas mixture flowed through the tubes continuously (Figure IV-13). The mixtures were regulated by flowmeters to give gas compositions calculated to have oxygen and sulfur fugacities typical of magmatic gases once the mixtures attained thermodynamic equilibrium at the temperature of the tests (usually about 1130°C). Two simulated magmatic gas atmospheres were used: one with an oxygen fugacity equal to that of the QFM phase assemblage and one with a slightly higher oxygen fugacity, equal to that of the NNO assemblage. Typical gas compositions and fugacities of the gas species for calculated equilibrium conditions are listed in Table IV-16.

Figure IV-13. Controlled Atmosphere Test System

TABLE IV-16

Controlled Atmosphere Compatibility Experiments:
Typical Gas Compositions

		Oxygen Fugacity QFM	Oxygen Fugacity NNO
Inlet Composition (Mole Fraction)	CO ₂	0.50	0.63
	CO	0.15	0.012
	H ₂ O	0.01	0.02
	SO ₂	0.34	0.33
Calculated Equilibrium State (Fugacity, bars)	CO ₂	0.54	0.54
	CO	0.024	0.008
	H ₂ O	0.01	0.02
	SO ₂	0.25	0.28
	H ₂	0.002	1×10^{-4}
	H ₂ S	0.003	2×10^{-5}
	S ₂	0.02	3×10^{-4}
	O ₂	2×10^{-10}	2×10^{-9}

T ~ 1130°C

P ≈ 0.85 bar

Total Gas Flow ≈ 10^{-2} mole min⁻¹

Experiment Results

Table IV-17 lists the materials that have been exposed in the various test environments to date, and Figures IV-14 through IV-23 illustrate some typical results obtained in the experiments. Table IV-18 summarizes the approximate overall rates of corrosion observed in the various laboratory tests. Full technical reports on results of the various tests conducted to date are in preparation.

The behavior of metal alloys can be predicted to a certain extent from thermochemical and phase equilibrium data on the stabilities of the

oxides and sulfides of the metals. Figure IV-24 is an example for the Fe-S-O system; the chemical potentials of oxygen and sulfur used in the metal plus lava plus gas laboratory experiments are plotted relative to the stability fields of the iron-containing phases. The test compositions are plotted near the boundary of the magnetite and iron-sulfide phase fields; although the fugacities of the N-NO test conditions fall within the magnetite field, iron sulfide has been the dominant product observed in the actual tests. This type of correlation is being continued for the other metallic elements under study.

TABLE IV-17

Materials Examined in Compatibility Tests
As of September 1978

	Degassed Lavas	Lava Plus Simulated Magmatic Gas	Direct Exposure to Simulated Magmatic Gas	Exposures at Kilauea
Titanium	X	X	-	-
Vanadium	-	X	-	-
Chromium	X	X	-	-
Iron	X	X	X	X
Cobalt	X	X	-	-
Nickel	X	X	X	X
Zirconium	-	X	-	-
Niobium	-	X	-	-
Tantalum	X	X	-	X
Molybdenum	X	X	X	X
Tungsten	X	X	-	X
Rhenium	-	X	-	-
Rhodium	-	X	-	-
Palladium	X	X	-	-
Platinum	-	X	-	-
1030 Carbon Steel	X	X	X	-
70Fe-30Ni	X	-	-	X
Nichrome (80Ni-20Cr)	X	-	-	X
Kovar (Fe-29Ni-17Co-2Mn)	X	-	-	X
302 SS (Fe-18Cr-9Ni-2Mn)	-	-	-	X
310 SS (Fe-25Cr-20Ni)	X	X	X	X
316 SS (Fe-17Cr-12Ni-2Mo)	X	X	X	X
330 SS (Fe-19Cr-36Ni)	-	X	X	X
446 SS (Fe-25Cr)	X	-	-	X
A-286 (Fe-15Cr-26Ni-2Ti-1.25Mo-0.3V)	-	X	X	X
Inconel 600 (Ni-15.5Cr-8Fe)	X	-	-	X
Inconel 718 (Ni-19Cr-18.5Fe-0.9Ti-0.5Al-3Mo-5.3(Nb+Ta))	X	X	X	X
Hastelloy C (Ni-16Cr-5Fe-2.5Co-17Mo-4W-0.4V)	-	-	-	X
Hastelloy X (Ni-22Cr-18.5Fe-1.5Co-9Mo-0.6W)	-	X	X	X
Udimet 700 (Ni-15Cr-18.5Co-3.5Ti-4.5Al-5Mo)	X	X	X	X
Rene 41 (Ni-19Cr-11Co-3.1Ti-1.5Al-10Mo)	-	-	-	X
MAR-M509 (Co-23Cr-10Ni-2Fe-0.2Ti-7W-3.5Ta-0.5Zr)	-	X	X	-
MP35N (35Ni-35Co-20Cr-10Mo)	X	-	-	X
WC 3015 (49Nb-30Hf-14W-4.5Ti-1.5Zr)	X	-	-	X
TZM (Mo-0.5Ti-0.08Zr)	X	-	-	-
WC C-6 (WC+11Co, 8-15(TiC, TaC))	-	-	X	-
TiC K151A (TiC+19(Ni, Mo))	-	X	X	-
BN	X	-	-	-
UCAR (BN/TiB ₂)	X	-	X	-
Al ₂ O ₃	X	X	X	X
ZrO ₂	X	-	-	-

Figure IV-14. Moderate Amount of Corrosion of Stainless Steel 316 Plus Degassed Basalt, 97 h, 1200°C. From left to right: basalt; Fe-rich metal grains; oxide layer composed of (Mg, Mn, Fe) (Cr, Al)₂O₄ spinel plus Cr₂O₃; Cr- and Mn- depleted alloy containing inclusions of MnCr₂O₄ spinel; unaffected alloy. Field-of-view 0.4 x 0.5 mm, reflected light.

Figure IV-15. Cobalt Plus Rhyolite, 98 h, 1200°C. Light region surrounding metal is zone in which ferric iron has been depleted by reaction with cobalt ($2\text{Fe}^{3+} + \text{Co}^0 = 2\text{Fe}^{2+} + \text{Co}^{2+}$). Inner, slightly dark half of this zone appears blue and represents extent of outward diffusion of Co^{2+} . 23 x 18 mm f.o.v., transmitted light.

Figure IV-16. Photomicrograph of Cobalt-Rhyolite Interface, Showing Absence of Any New Phases as Reaction Products. This reaction, $\text{Fe}^{2+} + \text{Co}^0 = \text{Fe}^0 + \text{Co}^{2+}$, produces only Fe^0 which dissolves in the cobalt metal, and Co^{2+} which dissolves in the lava. 0.21 x 0.27 mm f.o.v., reflected light.

Figure IV-17. Iron, 316 Stainless Steel, Tungsten Carbide, and Molybdenum (clockwise from right) Immersed in Basalt Which Was Exposed to a Simulated Magmatic Gas Atmosphere (1130°C, 21 h, CO_2 -CO- SO_2 - H_2O Gas with Oxygen Fugacity Equal to that of the QFM buffer). Moderate corrosion of the SS 316 and Mo but complete consumption of the iron. Inner diameter of crucible, 27 mm.

Figure IV-18. 1030 Carbon Steel (originally a rectangular coupon) Exposed to Simulated Magmatic Gas (1130°C , 8 h, $\text{CO}_2\text{-CO-SO}_2\text{-H}_2\text{O}$ gas with oxygen fugacity equal to that of the NNO buffer). The steel was completely converted to a molten or partially molten mass of iron sulfide. Inner diameter of crucible, 27 mm.

Figure IV-19. Extensive Sulfide/Oxide Corrosion Scale Developed on 316 Stainless Steel Exposed Directly to Simulated Magmatic Gas (1130°C , 9.5 h, oxygen fugacity = NNO). Scale broken off to show remaining metallic core. 28 x 35 mm f.o.v.

Figure IV-20. Photomicrograph of Remaining Core of 316 SS Specimen, Showing Intergranular Formation of Sulfides. Reflected light, 5.2 x 6.7 mm f.o.v.

Figure IV-21. Moderately Severe Corrosion of 316 Stainless Steel Immersed in Basalt and Exposed to Simulated Magmatic Gas (same specimen as in Figure IV-16). From top to bottom: basalt; Cr_2O_3 oxide layer (gray, irregular); (Fe,Cr)S sulfide (light gray) plus Fe-, Ni-rich metal grains (white); Cr-depleted alloy containing inclusions of chromium and chromium-iron sulfides; unaffected alloy. Reflected light, 1.5 x 2.0 mm f.o.v.

Figure IV-22. Moderate Corrosion of Udimet 700 Alloy (Ni-18.5Co-15Cr-5Mo-4.5Al-3.5Ti) Exposed Directly to Simulated Magmatic Gas (1130°C, 8 h, oxygen fugacity = NNO). 11 x 14 mm f.o.v.

Figure IV-23. Moderate Corrosion of Udimet 700 Immersed in Basalt Exposed to Simulated Magmatic Gas (1130°C, 8 h, oxygen fugacity = NNO). From top: basalt; thermal contraction gap now filled with mounting medium; thin outer scale of sulfides and oxides; alloy with sulfide inclusions. Reflected light, 0.45 x 0.57 mm f.o.v.

TABLE IV-18

Compatibility Tests: Approximate Observed
Corrosion Rates at 1130°C

	QFM Gas (mm/h)	QFM Gas + Basalt	NNO Gas (mm/h)	NNO Gas + Basalt (mm/h)	Basalt + Argon
1030 Steel	0.3	0.3	0.33	0.25	10^{-4}
316 SS	0.04	-	0.01-0.08	10^{-3}	10^{-4}
A-286	-	-	0.08	10^{-3}	-
310 SS	0.02	-	10^{-3}	10^{-3}	-
330 SS	0.05	-	0.001-0.005	10^{-3}	2×10^{-4}
Inconel 718	-	-	0.0015	10^{-3}	-
MAR-M509	0.1	-	0.005-0.01	10^{-3}	-

Figure IV-24. Gas Compositions Used in Laboratory Tests Plotted on an f_{O_2} - f_{S_2} Diagram for the Fe-S-O System. Phase fields are taken from published equilibrium diagrams for the Fe-O and Fe-S systems; the boundaries are dashed for the region involving ternary Fe-S-O liquid for which data do not exist.

Conclusions

In general, the following conclusions have been reached:

1. The worst environment for material compatibility is that simulating direct exposure to sulfur-bearing magmatic gases. Transport of the gaseous reactants to the metal surface is rapid; most alloys are rapidly attacked with formation of solid or liquid sulfides of elements such as Fe and Cr, and with oxide products often present as well. Corrosion is more severe in gases which have lower oxygen fugacity and therefore higher activity of H_2S and sulfur. The nickel- and cobalt-base "superalloys" and high-alloy stainless steels such as 310 (25Cr-20Ni) and 330 (19Cr-36Ni) are the most resistant to attack, but even these are corroded significantly in the gases of highest sulfur activity that have been studied. Ordinary carbon steel is unstable in this environment and is rapidly and completely destroyed.

2. Corrosion of alloys immersed in basaltic magma which is saturated with a sulfur-containing gas phase is much less rapid than corrosion of those in direct exposure to the gas, but the effects are similar. Sulfides and oxides are formed, but the presence of elements such as Mg and Al in the magma results in the formation of oxide phases such as $Mg(Cr,Al)_2O_4$. Transport of sulfur to the metal is slower since it is limited by diffusion through the liquid magma. Titanium carbide exposed to the lava and gas environments has proven considerably more resistant to attack than tungsten carbide.

3. For materials immersed in completely or partially degassed molten lava surrounded by an inert atmosphere, corrosion is less rapid still, involving mostly the formation of layers of oxides such as $MgCr_2O_4$. In many cases the oxide layers appear tightly adherent and act as a protective diffusion barrier. Some internal formation of sulfide phases is still observed from residual sulfur in the lavas. Metals such as Ni and Co, the

oxides of which are less stable than the iron oxide present in magma, show little corrosion, with only a small amount of this solution of the metal, accompanied by some absorption of iron by the metal. In the degassed lavas, nearly all alloys examined in the laboratory tests (even iron and carbon steel) hold up quite well to attack, aside from those which contain major proportions of mets that are very easily oxidized (e.g., Ti).

4. Corrosion is more rapid in basalt lava than in andesite, and is slowest in rhyolite. The lower viscosity of basalt (hence the more rapid transport of dissolved reactants and products) and its higher iron content both promote a greater reactivity. Iron is the major oxidizing agent in degassed lavas, and both ferrous and ferric iron in magma can oxidize metals that form more stable oxides (e.g., Cr, Mg, Al, Ti, Si, Mo, W, Ta):

The reduced iron (Fe^0) which is produced may appear either as discrete grains of relatively pure iron or it may dissolve in and diffuse into the alloy.

5. In the field exposures at fumaroles, nickel-base alloys such as Inconel 718, Hastelloy-C and Rene-41 were found most resistant to corrosion. Iron and some of the low alloy stainless steels (such as Type 302) are corroded fairly rapidly. A pure nickel sample was not corroded at 124°C in a SO_2 -rich fumarole but was severely attacked at another fumarole (Sulphur Bank) in which the gases are inferred to have a higher $\text{H}_2\text{S}:\text{SO}_2$ ratio.

V. TASK 4--ENERGY EXTRACTION

Introduction

Molten magma bodies offer a source of high quality, pollution-free thermal energy. Evaluation of the scientific feasibility of energy extraction is the objective of this task of the Magma Energy Research Program.

A number of schemes for energy extraction have been considered; these include thermionics, thermoelectrics, steam generation for conversion to electricity, hydrogen production, and generation of synthesis gas (H_2 , CO, CH_4). The last three extraction schemes (steam, hydrogen, and synthesis gas) were chosen for these scientific feasibility studies, since thermodynamics and surface processing technologies are well defined. The basic questions of energy extraction are questions of heat and mass transfer rates in and from magma bodies. Definition of the properties of molten magmas, in Task 3, is critical to understanding energy extraction limits. Efforts in the Energy Extraction Task have been aimed at limited studies on the dynamics of magma bodies, extraction rates from molten lavas, and preliminary studies on the internal dynamics of heat exchangers.

Although any economic analyses are premature, a "back-of-the-envelope" calculation was made to determine if energy extraction from magma was competitive within several orders of magnitude. Results (Figure V-1) suggest heat-transfer rates on the order of $1-10 \text{ kW}_T/\text{m}^2$ are desirable. Calculations (Hardee 1974) on a conduction-dominated, noncirculating magma indicate that heat-transfer rates will decrease from an initial 5 to $10 \text{ kW}_T/\text{m}^2$ to $\sim 1 \text{ kW}_T/\text{m}^2$ after 30 yr. Calculations (Hardee 1976) of heat

transfer from convecting chambers suggest potential extraction rates of $5 \text{ kW}_T/\text{m}^2$ using estimated magma properties from shallow magma bodies. Innovative concepts of increasing local magma convection or total heat-transfer area would also improve energy extraction efficiencies. More accurate economic studies are not warranted at this time because of lack of data, and no further economic studies are planned as part of this research program.

Figure V-1. Preliminary Cost Comparison Between Magma/Electric Plant and Nuclear/Electric Plant.

Results of energy extraction studies are summarized in the following discussion.

Dynamics of Magma Bodies

Understanding of the thermal and mechanical dynamics of magma bodies is critical to evaluation of energy extraction concepts. Approaches used in this program include theoretical calculations using assumed properties for magma bodies, measurements in a lava flow, and theoretical plus experimental studies in Kilauea Iki.

Heat Transfer From a Lava Flow

The September-October 1977 eruption of Kilauea Volcano, Hawaii, produced a river of basaltic lava which flowed for several days. A heat-transfer probe (Figure V-2) containing two heat-flux gauges measured the conduction-dominant heat-transfer rates from molten lava into a cold probe inserted in an eddy in the lava river. During a 5-min test period, the primary heat-flux gauge indicated transient-decaying conduction-dominant heat fluxes ranging from 200 kW/m^2 down to 50 kW/m^2 . A secondary heat-flux gauge verified that convective heat-flux rates, due to either natural or forced convection in the eddy, were on the order of 10 kW/m^2 or less. Theoretical calculations of the expected conduction-dominant heat-transfer rates between the lava and the probe agree with the experimental data from the primary gauge to within an error of 20% or less (Figure V-3).

Figure V-2. Heat Flux Probe

Figure V-3. Energy Extraction Rates in Flowing Lava During 1977 Kilauea Eruption

Magma Thermal Characterization Studies

An analytical study of heat-transfer mechanisms in magma bodies was undertaken in order to improve our estimates of potential heat-extraction rate. One aspect of the studies was evaluation of convective upmelting (Hardee, 1977). Only one of four possible heat-transfer regimes was found which would support the process of convective upmelting. The length-to-diameter ratio of an upmelting magma body was theoretically found to be limited to a maximum value of 5. An analytical heat-transfer model of forced-intrusion magma flow, coupled with available experimental evidence, suggests that magma chambers likely have low superheat.

Thermal calculations were made to study the effect of viscous dissipation heating which can cause thermal runaway in magma conduits. Our calculations confirmed the earlier calculation of Fujii and Uyeda (1974), which suggested that thermal runaway could occur in magma conduits of diameters greater than 1 m. This effect may present problems for magma heat-exchanger designs, which usually require a diameter of ~1 m to be economical.

Analytical solutions have been found for predicting the surface heat-flux distribution above rectangular (dikes) and spherical magma sources. The solution for the dike geometry predicts the steady-state heat-flux distribution at the surface of the earth for a conduction-dominant region above the magma source. The solution for the spherical magma source includes transient effects. These solutions have application in the location of magma bodies. The surface heat-flux distributions can be used to estimate characteristics of the magma body such as the extent of the body and depth. The dike solution has been compared to some recent measurements made on the Yellowstone Caldera (Morgan, Blackwell, and Spafford, 1977) and the spherical solutions have been compared with data from Coso and Avachinsky (Hardee 1979).

Hydrothermal Circulation and Solidification in Kilauea Iki Lava Lake

Kilauea Iki lava lake is a unique small-scale model of a magma heat source. A molten lava source is known to exist at a shallow depth, and downhole temperature measurements have been made extending to the melt. Downhole temperature measurements in Kilauea Iki suggest that the geothermal system above the molten zone is composed of two regions. The top region (currently down to 35-m depth) is a two-phase porous convective hydrothermal zone. The lower region (currently extending from 35- to 46-m depth), which extends to the melt, is a single-phase porous zone with no appreciable circulation; the controlling heat-transfer mechanism is conduction. Downhole temperature measurements in this single-phase region are shown in Figure V-4. These temperature profiles have a significant curvature in the single-phase zone (35 to 46 m) where the temperature exceeds the boiling point of water. Analytical studies show that the temperature profile curvature effects in the lower zone appear due to transient conduction effects of the moving solidification front rather than to effects such as porous advection or variable thermal properties (Hardee 1979). The temperature profile curvature factor can be shown to be fundamentally related to the rate of solidification and, in fact, the rate of solidification can be determined directly from the temperature profile data. Calculations indicate that the permeability of the crust is around $1/3$ Darcy; however, in the lower zone of the crust, the gas in the pores or fractures is expected to be relatively stagnant because of the low value of the single-phase Rayleigh number. The thickness of the lower conduction-dominant zone is expected to be relatively constant and a function of only the temperature and physical properties of the lava crust.

Figure V-4. Kilauea Iki Downhole Temperature Profiles

Steam Generation

One energy extraction concept is to insert a closed heat exchanger into a magma chamber and generate steam for subsequent generation of electricity. A single-tube boiler experiment, done in 1974, demonstrated heat-extraction rates of 180 to 300 kW_T/m² for an exchanger immersed in 1400°C molten basalt (Hardee 1975). Recent studies have evaluated potential problems in the internal dynamics of a long tube exchanger needed for energy extraction from deep (2 to 10 km) magma bodies.

Because of the large vertical distances involved and the resulting hydrostatic head, a heat exchanger would likely have large pressure variations, potentially ranging from atmospheric to above the critical pressure of water (3206 psia). After examining the literature, it became apparent that little or no work has been done characterizing heat exchangers with large internal-pressure variations. Therefore, a test facility was designed and constructed to study part of this problem (Figure V-5).

The basic long tube heat-exchanger design is a closed-loop counter-flow heat exchanger consisting of two concentric tubes. Water flows down the inner tube and then up the annulus between the outer and inner tube, while radiant (resistive) heaters provide thermal input to the external surface of the outer tube.

The areas of greatest uncertainty and concern for operating a heat exchanger at high (boiling) temperatures with a hydrostatic pressure variation include flow instabilities and "liquid-deficient" flow regions which could lead to tube-wall burnout (melting). The long tube heat exchanger (LTHE) facility was designed with the instrumentation and controls necessary to investigate these phenomena and to quantitatively characterize the flow and heat transfer. Preliminary tests indicate that the system can be operated in a stable mode, and that the long tube heat-exchanger concept is a technically feasible method of energy extraction

from high energy-density fields. However, high heat-transfer rates from the heated annulus to the cold central tube may be a problem in this counterflow system.

A complete description of the facility, preliminary test results, and recommended modifications can be found in the report by Lee et al (1978).

Figure V-5. Long Tube Heat Exchanger Prior to Insertion Into Underground Experimental Facility

Fuels Generation

Production of liquid and gaseous fuels is desirable to provide a storable, versatile energy source. Thermal energy from magma bodies could be used to convert water and biomass to synthesis gases, or the ferrous iron in basaltic systems could possibly be used to generate hydrogen by the reduction of water (Northrup, Gerlach, Modreski, and Galt, 1978). Effects of temperature, pressure, and fugacities on equilibrium conversions are well defined. The most significant scientific unknowns are the rates of heat and mass transport in magma bodies.

Hydrogen From Water-Basalt Interactions

The potential fuel-producing capacity of basaltic magma and hot basaltic rocks resulting from equilibration with water-rich fluids can be predicted from the O_2 fugacities of simulant solids. This approach has been followed in thermodynamic calculations for the range of oxygen fugacities characteristic of basaltic magmas.

Figure V-6 shows hydrogen concentrations resulting from the equilibration of water with the various solid assemblages for a total pressure of 100 MPa (1000 bars), a reasonable value for basalt at shallow depths (for example, under the ocean ridges). The results are presented in $\log_{10}$ mole % hydrogen for the temperature range 600° to 1300°C, which includes the subsolidus to liquidus region for most magmas at such pressure. The curve labeled H_2O refers to the percent of hydrogen generated by the thermal dissociation of water. Hydrogen concentrations are mainly determined by temperature and oxygen fugacity, although a pressure increase from 0.1 MPa (1 bar) to 100 MPa (1000 bars) causes a decrease of about 0.25% in the region of NNO and QFM for most temperature. Hydrogen concentrations for systems with oxygen fugacities near NNO and QFM range

from 0.5% to 2.5%; hydrogen fugacities in this region range from 2.1 MPa (21 bars) to 0.3 MPa (3 bars) at 100-MPa total pressure. Before eruption and posteruption oxidation, most basaltic magmas probably have oxygen fugacities equal to or somewhat less than QFM. Hence, hydrogen concentrations of 3% are possible from typical water-basalt interactions.

Figure V-6. Hydrogen Concentration as a Function of Temperature for Several Mineral Assemblages.

The calculated hydrogen percentages, based on the simulant solids, have been tested and verified experimentally for water vapor reacting with a crushed basalt. Water vapor at low pressures was passed through a bed of porous, vesicular, degassed basaltic lava at 950°C, about +100°C below its solidus temperature. Tholeiitic basalt from the September 1971 pahoehoe flow in Kilauea caldera, Hawaii, was used. This lava contains about 10.5 wt % FeO and 1.5 wt % FeO_{1.5}. Seventy-seven grams of the lava were placed in a fused-silica tube open at both ends, and enclosed in a larger fused-silica tube placed in a furnace and connected via stainless-steel flanges

and tubing to a mechanical vacuum pump. The lava was first degassed at 400°C for 7 h, then heated to 950°C; the experiment began when total pressure fell to 6 Pa (45 microns). Gases being evolved at the end of the degassing period contained approximately 0.6 mole % hydrogen at a total pressure of 6 Pa. Water vapor from a flask of deaerated water at room temperature was allowed to flow over the bed of lava. Gases were analyzed with a UTI quadrupole mass spectrometer; samples were first withdrawn into a liquid-nitrogen-cooled cold trap to condense water vapor, and were then admitted to the mass spectrometer. The analyses were calibrated both by the use of a calibrated hydrogen-leak cylinder and by the introduction of a measured pressure of hydrogen into the system.

The observed production of 1 to 3 mole % hydrogen is in good agreement with that expected for this rock, with an anticipated oxygen fugacity between that of the QFM and NNO buffers.

Enhancement of Fuel Production

Although the capacity for fuel production by water injection into magma is large, the reducing capacity of any magma body is finite. Addition of biomass to the injected water is a way of enhancing the geochemical fuel-production capability of the magma. Mixtures of water with small percentages of biomass material are themselves reducing in nature and therefore decrease the consumption of FeO and H₂O per unit of fuel generated. Thus, small percentage additions of these components to the working fluid may allow fuel production from less FeO-rich, lower temperature andesitic and rhyolitic magmas. The possibility also exists of producing fuels directly from mixtures of water and biomass matter, using magma or hot rock as a source only of thermal energy.

Biomass may generally be regarded as fuel itself; however, many forms of biomass are not being used today and are in such a form or location as to make them unattractive for most uses. A relevant example is the seaweed, sargassum, which occurs in sea fields close to attractive magma

sources along the mid-Atlantic ridge. There is a possibility, however, that new forms of biomass can be refined to higher quality fuel in magma and that other, more conventional forms can be used as well.

Equilibrium calculations have been made for the addition of cellulose ($C_6H_{10}O_5$) to the injection fluid, assuming no reducing action by the magma. Results similar to those presented below have been obtained for a variety of other common components of biomass (nitrogen-containing cellulose, lignin, etc). Figures V-7 and V-8 summarize the concentrations of principal fluid species in mole % for $H_2O-C_6H_{10}O_5$ mixtures at 100 MPa (1000 bars) and temperatures of 600° and 1300°C, which bracket the temperature range for the more common rhyolitic to basaltic magmas. At 600°C, hydrogen concentrations level out at 2% after the addition of 3 wt % cellulose. With the addition of large proportions of cellulose, methane generation predominates over hydrogen. At temperatures greater than 800°C, fluid equilibria shift toward larger hydrogen and carbon-monoxide concentrations. Thus, mixtures of water with a few weight percent cellulose in a closed system loop heated by magma or hot rock could produce fuels of hydrogen and methane at lower temperatures and of hydrogen and carbon monoxide at higher temperatures.

Status of Energy Extraction

The thermodynamics of generating steam or gaseous fuels is well-established and shows that production of high quality steam, hydrogen, and synthesis gas is feasible at temperatures and pressures of magma bodies. Laboratory and field measurements indicate that heat-transfer rates may be adequate for practical energy production. However, new concepts for generating increased in-situ heat-transfer areas (e.g., forming a frozen, structural foam) are desirable.

Figure V-7. Effect of Cellulose Feed on the Equilibrium Concentrations of Principal Species.

Figure V-8. Effect of Cellulose Feed on the Equilibrium Concentrations of Principal Species.

More information on the physical and chemical properties of potential magma energy sources is needed to better define the scientific feasibility of energy extraction. Specific properties include phase equilibria and crystallization kinetics to evaluate effects of cooling on transport behavior; viscosity, thermal conductivity, and density of magma with fugatives to estimate heat-transfer rates; and mass diffusivities of Fe in hot, frozen magma to evaluate the hydrogen-generation potential. Need for this data emphasizes the importance of the Magma Characterization task.

Future studies under the Energy Extraction Task will include analytical studies defining the magma body - margin - hot dry rock thermal behavior, possibly using Kilauea Iki as a model for a geothermal system. Studies of the dynamics of energy extraction will be made on the lava lake or lava flows as the opportunity arises. Further engineering and economic studies are not planned.

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