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SEARCH FOR SHALLOW MAGMA ACCUMULATIONS
AT AUGUSTINE VOLCANO

MASTER

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OBJECTIVES AND RESULTS

Scope of Investigation

The main objective of the study was a search for shallow magma accumulations beneath Augustine Volcano using primarily three geophysical techniques: (1) temperature and heat flow measurements, (2) seismic refraction, and (3) three-dimensional modeling of aeromagnetic data. With these studies we hoped to gain insight into the interval structure of Augustine Volcano, to delineate, if possible, the size and shape of near surface magma bodies and to assess the potential of the volcano as a natural laboratory for hot rock and magma geothermal energy research. Augustine was chosen because it is a very young and very active volcano with several historic eruptions in 1812, 1883, 1935, 1963/64. One of the main targets for our geophysical studies was a summit lava dome of about 0.05 km^3 volume, extruded in 1963/64 and suspected to still contain considerable residual heat, perhaps be still partially molten years after its intrusion. Five months after the field work in 1975 this dome was exploded in January 1976. One month later, a hot (about $650\text{--}800^\circ\text{C}$) viscous dome was intruded into the January summit crater.

Significant Results

Thermal Studies:

- (1) Outlined surface temperature anomalies on the surface of the 1964 summit lava dome intrusion, using infrared (IR) remote sensing techniques for 2 successive years, 1974 and 1975, just prior to the January 1976 eruptions. The surface temperature increased monotonically from the base to the top of the dome.

- (2) From the IR data and micrometeorologic parameters calculated a minimum heat flux for the hottest region on the top surface of the 1964 dome of $2669 \text{ HFU} \pm 1061$ for July, 1974 and of $4523 \text{ HFU} \pm 3045$ for August, 1975. There was no significant change in heat flow from one year to the next signaling the impending eruption, however, the stability of the heat flux could only be determined within an order of magnitude.
- (3) A minimum heat flux for the same region on the top surface of the 1964 dome was also determined via direct temperature measurements and the effective conductivity of the ashy soil, and was $2730 \text{ HFU} \pm 1861$ in August, 1975. This number is indistinguishable from the IR based heat flux estimate.
- (4) Investigated a fumarole field at the base of the 1964 dome through direct heat flow measurements and a shallow drill hole:
- (a) Found no fumaroles above the boiling point.
 - (b) Fumarole temperatures decrease radially outward from the base of the 1964 dome, from 95°C (boiling point) to 43°C , 66 m distant.
 - (c) Minimum heat flow, based on effective conductivity and direct temperature measurements, ranged from 4000 HFU near the base of the '64 dome to 500 HFU at the edge of the field.
 - (d) Drilled 24.7 m deep hole 51 m from the base of the '64 dome. Penetrated about a 9 m thick water vapor table almost immediately. The temperature of the vapor was 46.8°C on Aug. 4, 1975, and 48.9°C on Sept. 26, 1975. Beneath the vapor table the temperature dropped along a linear gradient.

- (e) Fumarole field is result of steam travelling radially outward from probable ring fracture near the base of the 1964 lava dome through about 9 m thick cavernous dome crumble breccia, capped by ash layer of varying thickness. Steam quickly cools and mixes with air as it diffuses and convects outward. Steam has a meteoric origin. A relatively impermeable layer with no heat sources underlies the fumarole field.
- (5) Found IR thermal methods very imprecise at least with the simple equipment we had available, allowing only order of magnitude estimates of heat flux. Several important parameters such as (1) surface emissivity, (2) surface wetness, and (3) the windfield could not be determined and would have required (1) a multiband IR radiometer, (2) reflectometer to obtain surface albedo and relate that to surface wetness, and (3) a sophisticated way of determining the wind field, e.g., large array of wind measuring stations and time dependent numerical modeling.
- (6) Direct heat flow measurements suggest that source of heat was indeed the central '64 lava dome, which still retained a considerable amount of heat 11 years after its intrusion. The gases emitted were principally steam from reheated meteoric water, no significant amounts of magmatic gases were detected.

Seismic Refraction:

- (1) Defined velocity structure of Augustine Volcano above 1 km depth. The refractor near that depth severely limited our penetration.
- (2) High velocity (5.1 km/sec) zeolitized Mesozoic sandstones underlie the volcano at very shallow (900 m) depth, faulted up on the south side. Overall velocities of relatively small volcanic pile above

sea level are surprising low ranging from 2.56 km/sec for the more solid core to velocities as low as 1.2 km/sec for the pyroclastics apron. There is evidence for a higher velocity (4.4 km/sec) central pipe.

- (3) No magma bodies were located above 1 km depth.
- (4) Usefulness of refraction methods: If low velocity magma bodies with a characteristic dimension $> 1/2$ km existed at the levels sampled prior to the '76 eruptions, we would have probably located them. However, because of the nature of the refracted wave paths, often guided up the central higher velocity material, we could not sample all regions of the volcano uniformly. In the central intrusion region magma filled dikes could have existed but they would have needed to be smaller than the diffraction limit for seismic waves (order $1/2$ km).
- (5) Even though we did not locate magma at very shallow depth (shallower than 1 km) beneath or within Augustine the velocity structure resulting from the refraction experiment allowed highly accurate hypocenter location of volcanogenic earthquakes.

Earthquake Seismology:

- (1) All 1976 eruption precursor events occurred at depths shallower than 8 km and outline a central plumbing system. Attenuation patterns tentatively indicate a very shallow (few km deep) small magma chamber (Lalla, 1979).
- (2) The majority of the precursor events occur above 1 km depth.
- (3) Benioff zone lies at 110 km depth beneath Augustine, dipping landward at an angle of 50° and reaching a maximum depth of 200 km (Pulpan and Kienle, 1979). The deeper feeder system of Augustine Volcano from 8 to 110 km appears to be aseismic as observed elsewhere in island arcs.

- (4) Usefulness to locate magma: We rank passive seismic methods highest amongst the geophysical methods tried, followed by refraction seismic and then thermal and magnetic techniques.

Magnetics:

- (1) Acquired pre- and post-1976 eruption aeromagnetic surveys.
- (2) The overall magnetizations for Augustine volcanics are 2 to 5×10^{-4} emu/cc for pyroclastic debris, which makes up the bulk of Augustine Island, and about 5×10^{-3} emu/cc for the more consolidated flows and central domes. The magnetization is in the direction of the current earth's magnetic field. There is no evidence for reversely magnetized material.
- (3) A three-dimensional magnetic structure model was developed for Augustine Volcano that matches the aeromagnetic data. Integration of the magnetic fields of known geologic features matches the observed field fairly well except that there is strong evidence for an additional magnetized sill at shallow (0-1 km) depth. The magnetic model is consistent with the seismic refraction data.
- (4) We found no evidence for large non-magnetic (above the Curie point) regions indicative of magma bodies above depth of 1 km.

Significant Departures

- (1) We did not locate a shallow magma body beneath Augustine at the levels sampled and therefore could not delineate its shape and size and estimate its viscosity.

Potential of Augustine as a Natural Geothermal Laboratory
for Hot Rock and Magma Geothermal Energy Research

We rate the potential high: In 1976 a viscous, hot (650-800°C) dome intrusion of magma occurred at Augustine's summit. The intrusion is now 3 years old and very hot fumarole temperatures (order 700°C, Johnston, personal communication) on the top surface of the dome indicate that melt lies very near the surface. We have demonstrated that the summit region of Augustine is accessible to drilling and a variety of geophysical experiments. Seismicity studies by Lalla (1979) indicate that a small magma chamber (less than 1 km typical horizontal dimension) may lie at a depth of a few km beneath the center of the volcano.

The following 4 chapters give a detailed account of the 4 geophysical techniques used at Augustine Volcano to try to locate shallow magma accumulations: (1) Thermal, (2) active seismic, (3) passive seismic and (4) magnetics.

1. HEAT FLOW ON THE SUMMIT REGION OF AUGUSTINE VOLCANO 1974-1975

INTRODUCTION

Areas of steaming ground and fumarole activity at Augustine are associated with lava domes extruded on the summit in 1935 and 1964. The most active thermal regions are visually apparent in Figure T1. Fumaroles are concentrated in upper and lower moats MM concentric to the summit spine S.S. and along radial fractures F-F on the endogenous 1964 dome. A large fumarole lies at the 1935/1964 dome interface just below the wing tank in Figure 1 and a small fumarole field F.F. lies at the base of the 1964 dome. An investigation into the source, distribution, magnitude and stability of summit thermal areas was begun in July 1974 using ground based infrared (IR) mapping of the southwest face of the 1964 lava dome and fumarolic temperature measurements. We returned to the volcano in late July 1975, repeated the IR survey, directly measured ground and fumarole temperatures, conducted air flow calorimetry and drilled a 24.7 m hole (D.M.) in fumarole field F.F. at the base of 1964 lava dome. On September 26, 1975 the drill hole was relogged and the fumaroles in field F.F. were remeasured.

The stability of thermal features are of special interest from the point of view of eruption prediction since the violent vent clearing eruptions of January 23, 1976 were preceded by shallow seismic activity (< 8 km) beginning in May 1975 (Lalla, 1979), although this was not recognized at the time that the second set of thermal measurements were taken.

The thermal analysis begins with the estimation of the relative heat flux on the 1964 dome from the IR data and then proceeds to the analysis of the directly obtained temperature and heat flow measurements.

Figure T1: Summit of Augustine Volcano prior to January, 1976 eruption. F-F radial fumarole lines controlled by fractures, M-M moats of endogenous dome, S.S. summit spine, L.D. 1935 lava dome, F.F. fumarole field, D.H. drill hole site (photo by J. F. Luthy, from Alaska Geographic, Alaska's Volcanoes, Vol. 4(1), 1976).

A. REMOTE RELATIVE HEAT FLUX ESTIMATION

Ground based IR radiometry provides a rapid means of identifying small scale surface temperature differences on inaccessible volcanic edifices, as previously shown by Birnie (1973) and Lange and Avent (1973a,b). The absolute magnitude of the geothermal heat flux can be deduced from the surface temperature only by considering the complete time-dependent mass and heat flow problem. This necessitates considering both micro-meteorologic and subsurface heat and mass transfer processes (Rosema, 1975; LeShack and DelGrande, 1976; Kahle, 1977). An order of magnitude estimate of the relative heat flux within a geothermal area may be obtained without time-dependent numerical modeling via the surface heat balance method of Sekioka and Yuhara (1974).

In this study we use the surface heat balance method to estimate the relative geothermal flux over part of the surface of the 1964 lava dome, look for differences between the July 1974 and July 1975 surveys, and provide an analysis as to the significance of those differences.

Theory of Estimating Relative Geothermal Heat Flux by IR Radiometry

To estimate the relative heat flux of a region from an apparent radiometer temperature the following procedure must be followed:

- 1.) Convert apparent temperature to true surface temperature.
- 2.) Convert surface temperature to heat flux via a heat balance model.

Conversion of apparent temperature to true surface temperature.

The apparent temperature indicated by the radiometer is equal to the true surface temperature only if there is no absorption or emission by the intervening air column, the surface acts as a black-body radiator,

and there is no incoming long-wave sky radiation. To obtain the true surface temperature the total radiation balance must be written.

$$\int_0^{\infty} B(T_m, \lambda) D(\lambda) d\lambda = \int_0^{\infty} \epsilon_{\lambda} B(T_s, \lambda) D(\lambda) d\lambda \quad (1)$$

$$+ \int_0^{\infty} (1 - \tau_{\lambda}(x_0)) B(T_A, \lambda) D(\lambda) d\lambda + \int_0^{\infty} (1 - \epsilon_{\lambda}) B(T_{sky}, \lambda) \tau_{\lambda}(x_0) D(\lambda) d\lambda,$$

where:

$B(T_m, \lambda)$ = Black body distribution functions evaluated at radiometer indicated temperature.

$B(T_s, \lambda)$ = Black body distribution function evaluated at true surface temperature.

λ = Wavelength of radiation

τ_{λ} = Atmospheric transmission over path x_0 long at wavelength λ .

ϵ_{λ} = Emissivity of surface at wavelength λ .

$D(\lambda)$ = Normalized radiometer response function.

$B(T_{sky}, \lambda)$ = Black Body distribution function evaluated at sky temperature.

$B(T_A, \lambda)$ = Black Body distribution function evaluated at air temperature.

Equation (1) is the same as that given by Lorenz (1968) except that an atmospheric transmission term is also included. The atmospheric transmission values used in equation (1) were calculated in 25 cm^{-1} intervals using the absorption model of Davis and Vezee (1964). Their model consists of a series of analytical expressions which gives the transmission coefficient as a function of pressure, temperature and amount of absorber. The true surface temperature is found by numerically solving equation (1).

In this study we have assumed that:

1. The surface being measured acts as a grey Lambertian body.
2. The sky radiates as a black body at the average temperature measured by the radiometer.
3. The temperature and composition of the air between the observation point and surface being measured is uniform.

Accuracy of Calculated Surface Temperature

The effect of any relevant variable on the calculated surface temperature T_s may be derived from equation (1). However, the expression is rather cumbersome for evaluating the error in T_s due to the uncertainty in the measured variables. Figure T2 illustrates the sensitivity of the calculated surface temperature to air temperature for various surface temperatures, assuming a path length of 300 meters with a water vapor content of $7.4 \times 10^{-6} / \text{cm}^3$. The magnitude of the correction is a simple linear function of the air temperature. Figure T3 is a plot of the magnitude of the correction versus the difference between the surface temperature and the sky temperature. For temperature contrasts of practical interest ($<60^\circ\text{C}$) a linear function may be used to represent the correction with a precision of 0.2°C . The nearly linear functional relationships is a result of sampling the radiated energy over a narrow band of wavelengths for a small range of temperatures (Bartolucci et al., 1976). For the purpose of error analysis the radiant energy balance may be represented by a two-term Taylor expansion and therefore the temperature correction may be written as the following simple linear relationship (Tein, 1974):

$$T_s = T_m + \frac{A}{(1-A)} (T_m - T_a) + \frac{(1-\epsilon)}{\epsilon} (T_m - T_{\text{sky}}), \quad (2)$$

Figure T2: Effect of intervening humid air on apparent IR surface temperature measurements, including (1) radiance of the air mass (2) absorption of humid air, calculated for 300 m long path and a humidity of 7.4×10^{-6} g H_2O per cm^3 of air.

Figure T3: Effect of sky radiation on apparent IR surface temperature for materials with different emissivities ϵ , ash $\epsilon = 0.95$, andesite $\epsilon = 0.90$.

where

T_s = Surface temperature

T_m = Apparent temperature measured

A = Effective absorption of column between observation point and surface.

ϵ = Emittance of surface

T_a = Air temperature

T_{sky} = Effective sky temperature

We will use equation (2) with the root-sum-square method (Hewitt, 1973) to estimate the uncertainty in the calculated surface temperature for a single point and in the calculated surface temperature difference between two points during a given temperature survey. The general expression for the uncertainty of a function:

$$T = f(x_1, x_2, \dots, x_n)$$

is

$$\Delta T = \left[\left(\frac{df}{dx_1} \Delta x_1 \right)^2 + \left(\frac{df}{dx_2} \Delta x_2 \right)^2 + \dots + \left(\frac{df}{dx_n} \Delta x_n \right)^2 \right]^{1/2} \quad (3)$$

for random uncorrelated errors.

Applying the method to equation (2), the following expression results for the uncertainty of a single temperature measurement:

$$\begin{aligned} \Delta T = & \left[\left(-\frac{1}{\epsilon} \left(\frac{A}{1-A} \right) \Delta T_A \right)^2 + \left(\left(\frac{1-\epsilon}{\epsilon} \right) \Delta T_{sky} \right)^2 \right. \\ & + \left(\left(-\frac{1}{\epsilon^2} \left(\left(\frac{A}{1-A} \right) (T_m - T_A) + (T_m - T_{sky}) \right) \Delta \epsilon \right)^2 \right. \\ & + \left(\left(1 + \frac{A}{(1-A)} + \frac{1-\epsilon}{\epsilon} \right) \Delta T_m \right)^2 \\ & \left. \left. + \left(\frac{(T_m - T_a)}{\epsilon} \left(\frac{1-2A}{(1-A)^2} \right) \Delta A \right)^2 \right]^{1/2} \quad (4) \end{aligned}$$

Since we are looking for temperature anomalies it is important to be able to estimate how small a temperature difference is significant. For the temperature difference at two points 1 and 2 the resulting expression for uncertainty is:

$$\Delta T_{2-1} = \left[\left(-\left(\frac{1}{\epsilon_2} \left(\frac{A_2}{1-A_2} \right) - \frac{1}{\epsilon_1} \left(\frac{A_1}{1-A_1} \right) \right) \Delta T_A \right)^2 + \left(-\frac{1}{\epsilon_2} \left(\frac{A_2}{1-A_2} \right) T_{m2} + \frac{1}{\epsilon_2} \left(\frac{A_2}{1-A_1} \right) T_A - \frac{T_{m2}}{\epsilon_2} + \frac{T_{sky}}{\epsilon_2} \right) \Delta \epsilon_2 \right)^2 + \left(\frac{1}{\epsilon_1} \left(\frac{A_1}{1-A_1} \right) T_{m1} - \frac{1}{\epsilon_1} \left(\frac{A_1}{1-A_1} \right) T_A + \frac{T_{m1}}{\epsilon_1} - \frac{T_{sky}}{\epsilon_1} \right) \Delta \epsilon_1 \right)^2 + \left(\left(1 + \frac{A_2}{\epsilon_2(1-A_2)} + \frac{1-\epsilon_2}{\epsilon_2} \right) \Delta T_{m2} \right)^2 + \left(-\left(1 + \frac{A_1}{\epsilon_1(1-A_1)} - \frac{(1-\epsilon_1)}{\epsilon_1} \right) \Delta T_{m1} \right)^2 + \left(\left(\frac{T_{m2} - T_a}{\epsilon_2} \right) \left(\frac{1-2A_2}{(1-A_2)^2} \right) \Delta A_2 \right)^2 + \left(-\left(\frac{T_{m1} - T_a}{\epsilon_1} \right) \left(\frac{1-2A_1}{(1-A_1)^2} \right) \Delta A_1 \right)^2 \right]^{1/2} \quad (5)$$

Relationship of Surface Temperature to Geothermal Heat Flux

In this study we use the net heat balance technique of Sekioka and Yuhara (1974) but use the formulation used by Burke (1945), Malkus (1962), Kahle (1977) to estimate sensible and latent heat loss.

The principle assumptions of the technique are:

1. All micrometeorologic parameters are uniform over the surface.
2. The surface has uniform slope, emissivity, reflectivity and thermal inertia.
3. The instantaneous temperature difference between any two points on the surface is equal to the differences in the mean daily temperature between the two points.

4. The heat transport process in the soil may be considered a diffusive process.
5. Complications due to atmospheric instability and varying surface roughness may be neglected.

Even though there are many assumptions Sekioka and Yuhara (1974) found that the geothermal flux estimated by this method for the Owakadari geothermal area, Hakone volcano, agreed with heat flow values calculated from snow fall calorimetry within a factor of two or three. The resulting formula for the difference in geothermal flux between any two points is:

$$G_{2-1} = \epsilon \sigma (T_{s2}^4 - T_{s1}^4) + P_a C_d c_p W (T_{s2} - T_{s1}) + m P_a C_D W l (q_2 - q_1) \quad (6)$$

where

G_{2-1} = Difference in geothermal flux between any two points

T_s = Surface temperature

W = Wind speed + 2m/sec

C_D = Drag coefficient = $.002 + .006 (z/5000)$

z = Altitude

P_a = Density of air at the surface

c_p = Specific heat of dry air at constant pressure

m = Surface wetness

q = Mixing ratio of air near the ground

l = latent heat of evaporation of water

ϵ = Emissivity of the surface

σ = Stefan-Boltzman constant

If we calculate the uncertainty of the relative heat estimate via the root-sum-square method (Hewitt, 1973), we find:

$$\Delta G_{2-1} = \left[\left(\frac{\epsilon \sigma y}{2} (9x^2 - y^2) \Delta x + P_a C_D c_p \Delta x \right)^2 + \left(\frac{3\sigma \epsilon x}{2} (x^2 - y^2) \Delta y \right)^2 + \left(\left(P_a C_D c_p x + P_a C_D \ell m (q_2 - q_1) \right) \Delta W \right)^2 + \left(P_a W C_b \ell (q_2 - q_1) \Delta m \right)^2 + (P_a C_D \ell m W \Delta q_2)^2 + (P_a C_D \ell m \Delta q_1)^2 + (\sigma (T_{S2}^4 - T_{S1}^4) \Delta \epsilon)^2 \right]^{1/2} \quad (7)$$

where: $x = T_{S1} - T_{S2}$ and $y = T_{S2} + T_{S1}$.

Data Analysis

Instrumentation and Procedure

All apparent temperatures were measured with a Barnes PRT-5 precision radiation thermometer. The unit has a band pass of 8-14 microns and a field of view of 2°. The survey distance varied between 60-350 meters; therefore the measurements are integrated values over near circular areas with major axes between 1.8 and 10.5 meters. The targets to be measured were aligned with the radiometer using a six power rifle scope. The experiment was documented by plotting apparent surface temperatures on Polaroid photographs.

Wet bulb and dry bulb temperatures at the site were measured with an Assaman psychrometer. In 1974 we monitored atmospheric pressure with a recording drum barometer. In 1975 the instrument failed during the IR survey and we had to use the Homer WSO pressure corrected for altitude.

No wind recording device was used in 1974; however, the steam from the fumaroles rose nearly vertically, indicating a wind velocity of less than 3 km/h. In 1975 we used a hot wire anemometer to estimate the wind velocity at the measuring site.

Calibration

The Barnes PRT-5 is factory specified to an accuracy of $\pm .5^{\circ}\text{C}$. To provide a check on this calibration, surface temperature measurements of melting snow fields located on or near the 1964 lava dome were interspersed with lava dome temperature measurements. The results are:

Mean Snow Field Temperatures, Corrected for Atmospheric Effects

1974 Survey:	2.2°C	$\pm .4^{\circ}$
1975 Survey 1:	0.4°C	$\pm .1^{\circ}$
1975 Survey 2:	0.9°C	$\pm .1^{\circ}$

Melting snow has an emissivity of .995, thus one would expect a surface temperature of 0°C . The extremely large departure in 1974 was found to be the result of a weak battery which powered the heating unit for the reference cavity. Some of the difference in 1975 is probably the result of the inability of the cavity reference temperature to maintain itself when exposed to the forced convective cooling due to 30 km/h winds. For this study we simply reduced all temperature measurements by the mean snow field temperature values, thus the absolute accuracy of an apparent temperature is dependent on the snow temperature truly being 0°C . The standard deviation of the snowfield temperature is assumed to be totally due to instrument drift thus providing a maximum estimate of error induced by radiometer instability.

IR Temperature Surveys

All apparent temperatures were converted to surface temperatures via equation (1). In using equation (1) we have assumed a constant emissivity of 0.95, the mean value measured by Sekioka and Yuhara (1974) for a geothermal area on Hakone Volcano, an andesitic stratovolcano in Japan.

The 1964 lava dome is covered with dacitic-andesite ash, composed chiefly of glass with small amounts of feldspar and pyroxene crystal fragments. The major source of variations in emissivity over the surface is varying surface wetness, the most active areas of steaming ground being the wettest. A variation in wetness from saturated to dry could cause a change of the emissivity of order 0.03 (Buettner and Kern, 1965).

The air temperature over the measured path has been assumed to be equal to that at the observation point. Since the elevation varied only by 100 m the maximum temperature variation assuming a dry adiabatic lapse rate of $1^{\circ}/100$ m would be equal to 1° .

The accuracy of the apparent temperature measurements was estimated from the standard deviation of the snowfield calibration. The accuracy of the surface temperature or surface temperature difference has been estimated with equation 4 and equation 5, respectively. A complete listing of calibration, apparent and corrected temperatures, for the 1974, 1975 surveys are given in Tables T1-5.

Results

Surface temperature maps of the southwest face of the 1964 lava dome for 1974 and 1975 are shown in Figure T4b and Figure T5b, respectively.

TABLE T1

1974 CALIBRATION USING SNOW FIELDS

July 14, 1975

T_m °C	Distance (meters)	T_c °C	$(T_m - 2.23)$ °C	T_s	Point No.
1.9	60	1.6	-0.33	-0.7	1
2.5	60	2.3	0.27	-0.1	2
3.1	150	2.6	0.87	0.3	3
3.0	230	2.4	0.77	0.0	4
Mean 2.62 ± .55		Mean 2.23 ± .43		Mean -0.13 ± .42	

TABLE T2

1975 CALIBRATION USING SNOW FIELDS

Survey No. 1 July 28, 1975

T_m °C	Distance (meters)	T_c °C	$(T_m - 0.42)$ °C	T_s	Point No.
.6	60	0.4	0.18	0.0	1
.6	60	0.4	0.18	0.0	2
.7	60	0.6	0.28	0.1	3
.8	90	0.4	0.38	0.0	4
.6	90	0.3	0.18	-0.2	12
.8	90	0.4	0.38	0.0	13
.9	110	0.5	0.48	0.1	22
.6	110	0.3	0.18	-0.2	26
.9	150	0.4	0.48	-0.2	17
1.0	150	0.5	0.58	0.0	18
1.0	230	0.4	0.58	-0.2	19
Mean 0.77 $\pm$.16		Mean 0.42 $\pm$.09		Mean -0.05 $\pm$.12	

Survey No. 2 August 4, 1975

1.0	60	0.8	0.2	0.0	1
1.1	60	0.9	0.3	0.1	2
Mean 1.05 $\pm$.07		Mean 0.85 $\pm$.07		Mean 0.05 $\pm$.07	

TABLE T3

1974 IR DATA

Point No.	T_m	$(T_m - 2.23)$	T_s	Surface	Distance (meters)
1	1.9	-0.3	-0.7	S	60
2	2.5	0.3	-0.3	S	60
3	11.2	9.0	9.3	G	90
4	10.8	8.6	8.9	G	90
5	10.5	8.3	8.6	G	90
6	10.7	8.5	8.8	G	90
7	4.3	2.1	-	SG	90
8	7.0	4.8	-	SG	90
9	10.9	8.7	9.0	G	90
10	11.3	9.1	9.4	G	90
11	13.8	11.6	12.1	G	90
12	5.2	3.0	-	SG	130
13	3.1	0.9	0.3	S	150
14	7.1	4.9	-	SG	250
15	3.0	0.8	0.0	S	230
16	5.6	3.4	-	SG	230
17	9.3	7.1	-	SG	90
18	10.2	8.0	8.2	G	60
19	9.5	7.3	7.4	G	130
20	12.1	9.9	10.3	G	130
21	12.5	10.3	10.7	G	110
22	12.5	10.3	10.7	G	110
23	12.5	10.3	10.7	G	110
24	12.5	10.3	10.7	G	110
25	12.4	10.2	10.6	G	110
26	13.0	10.8	11.2	G	110
27	12.5	10.3	10.7	G	150
28	12.3	10.1	10.5	G	150
29	10.4	8.2	8.4	G	150
30	3.3	1.1	-	SG	300
31	12.5	10.3	10.7	G	110
32	12.5	10.3	10.7	G	110
33	13.2	11.0	11.4	G	110
34	7.1	4.9	-	SG	110
35	12.4	10.2	10.6	G	110
36	13.1	10.9	11.3	G	110
37	13.9	11.7	12.3	G	150

S = Snow field

G = Bare ash

SG = Both snow and ash

(TABLE T3, Cont.)

Point No.	T_m	$(T_m - 2.23)$	T_s	Surface	Distance (meters)
38	13.8	11.6	12.1	G	230
39	10.6	8.4	8.8	G	300
40	11.2	9.0	9.4	G	300
41	10.8	8.6	-	SG	300
42	12.8	10.6	11.0	SG	110
43	12.6	10.4	10.8	SG	110
44	13.0	10.8	11.2	SG	110
45	13.0	9.8	10.2	SG	110
46	13.0	10.8	11.5	SG	300
47	15.1	12.9	13.5	SG	140
48	14.3	12.1	12.7	SG	140
49	13.1	10.9	11.4	SG	140
50	13.1	10.9	11.4	SG	200
51	12.4	10.2	10.6	SG	200
52	15.8	13.6	14.5	SG	250
53	14.4	12.2	13.0	SG	290
54	13.6	11.4	12.0	SG	290
55	15.0	12.8	13.6	SG	250
56	15.0	12.8	13.6	SG	250
57	14.9	12.7	13.5	SG	250
58	13.3	11.1	11.5	SG	150
59	13.2	11.0	11.4	SG	150

TABLE T4

1975 IR DATA

Survey No. 1

Point No.	T _m	T _m -.4	T _s	Surface	Distance (meters)
1	0.6	0.2	0.0	S	60
2	0.6	0.2	0.0	S	60
3	0.7	0.3	0.1	S	60
4	0.8	0.4	0.0	S	90
5	7.1	6.7	8.8	G	90
6	6.9	6.5	8.6	G	90
7	5.4	5.0	-	SG	90
8	6.6	6.2	8.2	G	90
9	6.4	6.0	8.0	G	90
10	6.4	6.0	8.0	G	90
11	6.6	6.2	8.2	G	90
12	0.6	0.2	-0.2	S	90
13	0.8	0.4	0.0	S	90
14	6.5	6.1	8.1	G	150
15	5.0	4.6	6.6	G	150
16	4.8	4.4	6.4	G	150
17	0.9	0.5	-0.2	S	150
18	1.0	0.6	0.0	S	150
19	1.0	0.6	-0.2	S	230
20	5.9	5.5	7.5	G	125
21	4.8	4.4	6.0	SG	90
22	0.9	0.5	0.1	S	110
23	6.4	6.0	8.0	G	110
24	6.4	6.0	8.0	G	110
25	6.6	6.2	8.2	G	110
26	0.6	0.2	-0.2	S	110
27	6.9	6.5	8.3	G	120
28	7.6	7.4	9.3	G	150
29	7.4	7.0	9.1	G	150
30	5.9	5.5	7.5	G	220
31	6.4	6.0	8.0	G	220
32	6.4	6.0	8.0	G	220
33	7.4	7.0	9.0	G	200
34	7.4	7.0	9.0	G	230
35	6.7	6.3	8.3	G	230
36	6.9	6.5	8.5	G	230
37	7.9	7.5	9.5	G	220
38	8.9	8.5	10.7	G	220
39	6.6	6.2	8.6	G	230
40	7.4	7.0	9.0	G	230
41	8.4	8.0	10.1	G	300
42	6.9	6.5	8.7	G	300
43	7.4	7.0	9.0	G	300
44	6.4	6.0	8.0	G	300
45	6.5	6.1	8.3	G	300
46	6.4	6.0	8.2	G	300

TABLE T5

1975 IR DATA

Survey No. 2

Point No.	T_m	$T_m - 0.4$	T_s	Surface	Distance (meters)
1	1.0	0.1	0.0	S	60
2	1.1	0.2	0.1	S	60
3	7.5	6.6	8.1	G	90
4	6.5	5.6	7.1	G	90
5	6.0	5.1	6.6	G	90
6	6.0	5.1	6.5	G	90
7	5.8	4.9	6.3	G	90
8	6.8	5.9	6.3	G	90
9	6.0	5.0	6.6	G	90
10	6.8	5.9	6.4	G	130
11	6.6	5.7	6.2	G	110
12	7.4	6.5	8.0	G	110
13	7.0	6.1	7.6	G	110
14	7.0	6.1	7.6	G	110
15	-	-	-	-	-
16	6.5	5.6	7.1	G	150
17	7.6	6.7	7.3	G	150
18	9.0	8.1	9.4	G	150
19	9.0	8.7	10.5	G	220
20	9.0	8.0	10.0	G	300
21	7.5	6.5	8.2	G	300
22	7.5	6.5	8.2	G	300

Figure T4:

- a. Location map of apparent surface temperature measurement points on the SW face of the 1964 lava dome taken on July 15, 1974. Circles are measurement points, numbers are referenced to temperatures given in Table T3.
- b. Contoured surface temperatures in °C, calculated from July 15, 1974 I.R. data, corrected for sky radiation, radiance and absorption of air, and assuming an emissivity of $e = .95$. All data normalized, assuming a melting snow field temperature of 0°C. Dotted areas are snow fields, diagonally ruled area is the reference zone for relative heat flux calculations.

Figure T5: a. July 28 and August 4, 1975, location map of apparent surface temperature measurement points. Squares July 28 measurements, Triangles August 4 measurements. Numbers are referenced to temperatures in Tables T4 and T5.
 b. Contoured surface temperatures in °C calculated from July 28 and August 4, 1975, IR data, corrected for sky radiation, radiance and absorption of air and assuming an emissivity of $e = .95$. All data normalized assuming a melting snow field temperature of 0°C. Dotted areas are snow fields on August 4, 1975. Diagonally ruled area is the reference zone for relative heat flux calculations.

Isotherms above the snow fields have been drawn at a 1° contour interval. It should be noted that the high sky temperature that prevailed during the 1974 survey minimizes the effect of variable emissivity on the uncertainty in the surface temperatures. The error analysis indicates that surface temperatures are significantly different if the difference is greater than:

July 14, 1974 : 0.8°C

July 28, 1975 : 1.6°C

August 4, 1975: 0.7°C

A complete listing of the effect of each relevant variable on the minimum temperature difference which is significant is given in Table T6.

The highlights of the temperature maps for 1974 and 1975 are:

1. In both years we find a zone of significantly higher surface temperatures near the summit. This region coincides with an active fumarole area in lower moat MM (Figure T1). A temperature profile from a pit dug in this moat is shown in Figure T6.
2. The 1974 temperature map shows a finger of high temperature extending down the left side of the map. This zone coincides with a radial fracture in the 1964 dome (Figure T1, F-F in front of wing). The feature is not evident in the 1975 survey.
3. The snow fields melt pattern is strikingly similar in both years (Figures T4b and T5b). A quantitative interpretation of this pattern would require accurate estimates of the snow pack and the long term meteorologic data. However, although the total heat flux may have changed, the relative flux must have remained constant in order to preserve the snow melt pattern.

A comparison of the relative magnitude of the temperature difference between the two surveys is not very meaningful since the temperature difference is dependent on the micrometeorologic conditions. We will therefore make an order of magnitude estimate of the relative geothermal flux over the dome which takes into account the micrometeorologic conditions.

Relative Geothermal Flux

The diagonally-rated region in Figure T4b and Figure T5b was chosen as a reference area for both years. The relative heat flux is calculated via equation 6 between this area and the highest temperature isotherm on each survey. The total, radiant, sensible and latent heat loss are listed below:

Survey 1974

Maximum surface temperature isotherm	=	13.5°C ± .6
Reference area average surface temperature	=	8.8°C ± .6
Wind speed	=	0.9 m/sec ± .4
Mixing ratio of water vapor to air at surface		
q_1	=	$8.0 \times 10^{-3} \text{ g/g} \pm .4 \times 10^{-3}$
q_2	=	$11.2 \times 10^{-3} \text{ g/g} \pm .6 \times 10^{-3}$
Surface wetness m	=	0.5 ± .5
ϵ	=	0.95 ± .015
T_A	=	8°C ± 1°
Total heat flux	=	2669 HFU ± 1061
Relative radiant heat loss	=	555 HFU ± 189
Relative sensible heat loss	=	1151 HFU ± 273
Relative latent heat loss	=	963 HFU ± 1008

1975 Surveys

Maximum surface temperature isotherm	=	$10^{\circ}\text{C} \pm 1.2^{\circ}$
Reference area average surface temperature	=	$7.4^{\circ}\text{C} \pm 1.2^{\circ}$
Wind speed	=	$9 \text{ m/sec} \pm 4.5 \text{ m/sec}$
q_2	=	$8.8 \times 10^{-3} \text{ g/g} \pm .6 \times 10^{-3}$
q_1	=	$7.3 \times 10^{-3} \text{ g/g} \pm .6 \times 10^{-3}$
m	=	$.5 \pm .5$
ϵ	=	$.95 \pm .015$
T_A	=	$9.0^{\circ}\text{C} \pm 1$
Total heat flux	=	$4523 \text{ HFU} \pm 3045$
Relative radiant heat loss	=	$311 \text{ HFU} \pm 188$
Relative sensible heat loss	=	$2509 \text{ HFU} \pm 1861$
Relative latent heat loss	=	$1703 \text{ HFU} \pm 2404$

The highest temperature isotherm for both 1974 and 1975 follows the trace of lower fumarole moat (Figure T1, MM). Since the reference area has a heat flux equal to or greater than zero, the relative heat flux calculated for both years is a minimum estimate for the lower moat. Later in this report the lower moat is found by direct temperature measurements and the effective thermal conductivity to have an average conductive flux of 2730 HFU in 1979 with a standard deviation of 1861 HFU over the length of the moat. This is also a minimum estimate of the total flux. The direct and indirect heat flux determinations in 1975, $4523 \text{ HFU} \pm 3045$ and $2730 \text{ HFU} \pm 1861$, respectively, agree reasonably well.

The important question of thermal stability between the 1974 and 1975 can only be constrained within an order of magnitude. The minimum estimates of heat flux for moat MM are indistinguishable at the one sigma level. A major source of error is the unknown variation of the surface wetness. In the error analysis we have conservatively allowed the wetness parameter to have a standard deviation from completely dry to completely

saturated. The saturated condition would be approached as the boiling point isotherm neared the surface, while the dry condition might exist in a region of very low heat flux. This difficulty could be eliminated by indirectly measuring the surface albedo and finding a relationships between the wetness and albedo. The other major problem is the estimation of the wind velocity over a region of large topographic relief. No easy solution exists for this problem without having numerous monitoring sites or complex wind field modeling. The problem of emissivity variations could be eliminated if a multiband radiometer were available so that the ratio technique of LeShack and Del Grande (1976) could be used.

Conclusions: The southwest face of the 1964 lava dome surface temperature increased with altitude, the highest temperature isotherm being coincident with the approximate trace of lower moat fumarole area MM. A minimum heat flux estimate for this lower moat is $2669 \text{ HFU} \pm 1060$ in 1974 and $4523 \text{ HFU} \pm 3045$ in 1975. The relative heat flux pattern observed in 1974 is in agreement with the 1975 data. However, the error limits only allow constraint of the stability to an order of magnitude.

B. DIRECT TEMPERATURE AND HEAT FLOW MEASUREMENTS

The thermal areas on Augustine consists of regions of steaming and hot ground with only weak fumarole activity; no temperatures were measured above 95°C, which is the boiling point of water at the summit elevation. Although not measured, the concentration of SO₂ and H₂S must be low since only a faint odor was detectable.

Heat Transport in Regions of Steaming Ground

The transport of energy in steaming ground is a highly complex problem of coupled heat and mass transport. Extensive field studies were made by Dawson (1964). White et al. (1971) recognized that the latent heat of vaporization is a primary energy source in near surface geothermal processes. Recently, there has been a new wave of papers in the literature dealing with the modeling of two-phase geothermal systems, e.g., Mercer and Faust (1975), Martin et al. (1976), Moench (1976); however, none of these authors treat the physics of the important near surface boundary layer in which shallow heat flow measurements were made in this study. A typical temperature vs. depth profile for this boundary layer is shown in Figure T6, top; a moisture content vs. depth profile is shown in Figure T6, bottom. The temperature curve can be divided into three regions on the basis of the temperature gradient:

Region 1 is characterized by an extremely small temperature gradient. It is the top of the two-phase zone defined by White (1969) as a vapor dominated system and the subject of many recent modeling efforts. A one-

Figure T6: Top: Temperature vs. depth profile at location H6 (Figure 7) in fumarole field F.F. (Figure T1). Curve is divided into three regions R-1, R-2, R-3 on the basis of temperature gradient.

Bottom: Moisture content vs. depth profile, also taken at location H6.

dimensional steady state counter-flow model has been presented for this region by Shcubert and Strauss (1979).

Region 2 is characterized by rapidly increasing temperature gradients and large amounts of vapor condensation. The increasing moisture content causes a rapid decrease in empty pore volume and thus a rapid decrease in vapor conductivity; in turn, the hydraulic conductivity increases, making it easier for liquid to drain away, thus acting as a negative feedback mechanism that limits the size of the zone of near saturation.

Region 3 is characterized by a near linear temperature gradient and small amounts of condensation. Heat transport by Fourier conduction becomes important; however, a significant amount of heat is transported by the diffusion of water vapor. The transport of liquid and vapor is coupled to the heat transport equation and may be described in terms of temperature and moisture dependent matrix potentials (DeVries, and Peck, 1958a, b). The relationship of the matrix potential and temperature for a given moisture content must be determined for each soil.

Estimation of Surface Heat Flow

Over most of the geothermal regions on Augustine shallow temperature measurements were made in a zone of near linear temperature gradients; thus, if the effective thermal conductivity could be estimated, the heat flux could be deduced. Two methods were used to estimate the geothermal flux:

1. A total estimate of heat flux via air flow calorimetry.
2. A minimum estimate of heat flux is using the temperature gradient and the effective conductivity.

Air Flow Calorimetry

An air flow calorimeter similar to that of Benseman (1959) and Yuhura (1970) was utilized in areas of hot ground, useful to approximately 10,000 HFU as a lower limit, to measure both the vapor and total heat transport.

The device functions by drawing air at a measured volume, temperature and humidity through an insulated box which cover an area of steaming ground. By measurement of the outgoing absolute humidity, temperature and flow rate and then finding the difference in energy and water content of incoming to outgoing flow, the heat and mass transport can be estimated. The tacit assumption is made that the artificial boundary conditions simulate the existing natural state.

The pertinent equations are:

1. $Q = P_A V C_p (t_{o_d} - t_{i_d}) + V(H_2 E_2 - H_1 E_1)$, where

P_A = density of air

V = flow rate cm^3/sec

C_p = heat capacity of air at constant pressure

t_{o_d} = outgoing dry bulb temperature

t_{i_d} = ingoing dry bulb temperature

H_2 = outgoing absolute humidity

H_1 = ingoing absolute humidity

E_2 = outgoing enthalpy

E_1 = ingoing enthalpy

2. $J_v = \frac{V(H_2 - H_1)}{A}$, where

A = area cm^2

J_v = water vapor flux g/cm^2

3. $Q_A = Q/A$, where $Q_A = \text{Flux} \frac{\text{cal}}{\text{cm}^2 \text{sec}}$.

The actual device consisted of a box insulated with 2" of urethane foam with a collection area of 0.27 m^2 . The input and output temperatures were measured with YSI thermister probes and the humidity was estimated from the dry and wet bulb temperatures. The air velocity was estimated with a hot wire anemometer.

Initially it was planned that the device would be used to calibrate the ground temperature in terms of heat flow; however, the fan failed after only a few measurements were made in fumarole field F.F. (Figure 1). The measurement did however provide a maximum estimate of heat flow for the field F.F.

Heat Flux Estimation Via Effective Thermal Conductivity

The effective thermal conductivity of a soil is dependent on:

1. The thermal conductivity and shape of the minerals composing the soil matrix.
2. The soil's moisture content and porosity.
3. The soil's temperature.

A moist soil's thermal conductivity may vary by a factor of 2-3 over a temperature range of $0 - 100^\circ\text{C}$. The major reason for such a large variation is the heat transported by the latent heat of vaporization of water. The partial pressure of water vapor in air is an exponential function of temperature and thus small temperature gradients may induce a large diffusion rate of water vapor through soil pores. The first unified approach in estimation of thermal conductivity and its dependence on moisture content and temperature was presented by DeVries (1952). He

borrowed a technique from electromagnetic theory which had been applied to estimate the dielectric constant of a material. For the case of spherical pores and mineral grains, the formulation reduces to:

$$\lambda = \frac{\sum_{i=0}^N K_i X_i \lambda_i}{\sum_{i=0}^N K_i X_i} \quad (8)$$

$$K_i = [1 + (\frac{\lambda_i}{\lambda_0} - 1) 0.333]^{-1}, \text{ where} \quad (9)$$

λ = effective thermal conductivity

λ_0 = thermal conductivity of the continuous medium (water)

λ_i = thermal conductivity of dispersed particles of medium i

X_i = volume fraction occupied by medium i.

For water vapor saturated air-filled pores the thermal conductivity was estimated to be:

$$\lambda_p = \lambda_a + \lambda_v, \text{ where} \quad (10)$$

λ_p = thermal conductivity of the gas filled pore

λ_a = thermal conductivity of air

λ_v = thermal conductivity of water vapor

$$\lambda_v = \frac{HDP}{RT(P-P_w)} \frac{dP_w}{dT} \quad (11)$$

$$D = \frac{17.6}{P} \left(\frac{T}{273} \right)^{2.3}, \text{ where} \quad (12)$$

H = latent heat of vaporization of water

D = diffusion coefficient of water vapor through air

P = total atmospheric pressure

Pw = partial pressure of water vapor

R = gas constant of water vapor

T = absolute temperature.

DeVries (1952) found agreement to within 5 - 10% of his theoretical estimates of thermal conductivity and those obtained via transient heat probe experimental determinations of thermal conductivity. However, in experiments designed to measure water vapor diffusion in the steady state case Cary and Taylor (1962), Hadas (1968) and Week et al. (1968) found that equation 12 underestimated the water vapor flux by factors of up to 5. It was suggested that thermal gradients are higher across gas filled pores than measured across the bulk medium, which induces larger water vapor migration than expected. For the purpose of this study we have used DeVries equations 8 to 12 for estimation of the temperature dependent conductivity. The method provides a minimum estimate of the total heat flux. If the vapor flux were underestimated by a factor of 5 for Augustine soil, where 40% of the available pore space is filled with water vapor, the heat flux would be underestimated by 30%. To correctly estimate vapor flux, equation 12 should be multiplied by a correction factor B (Cary and Taylor, 1962).

In order to use the DeVries' thermal conductivity estimation technique, the thermal conductivity of the matrix minerals must be determined; additionally, the porosity of the bulk sample must be known. Porosity was determined directly from a 3-inch diameter Shelby tube sample taken at location HC in Figure T7 and also from samples lab packed by repeated drying and wetting. The total porosity was found to be $50\% \pm 2\%$.

The average thermal conductivity of the matrix minerals was determined by first measuring the effective conductivity of water saturated samples using the transient needle probe technique. The average mineral conductivity was then found by solving equations 8 and 9. The value was then checked by measuring the samples with a saturation of 60%.

Needle Probe Thermal Conductivity Measurement

The probe technique was first used on soils by Hooper and Lepper (1950). The assumptions and pitfalls of the technique have been exhaustively discussed by DeVries and Peck (1958a, b) and Woodside and Messmer (1961). The technique is dependent on the needle probe approximating an infinite line heat source while the medium must behave as though it were infinite in extent. In the experimental situation the probe has finite length, finite conductivity, measurable heat capacity, finite contact resistance and is implaced in a moist finite medium. In order for line heat source theory to be valid, conflicting criteria must be met:

1. A long enough time must be allowed for the effect of contact resistance and the heat capacity of the probe to be negligible.
2. The experiment must be run for a short enough time, such that the finite sample size and axial heat flow do not effect the results.

The minimum sample size may be estimated by the relation:

$$\exp(-R_1^2/4at_1) \ll 1 \text{ (DeVries, 1958a), where}$$

a = diffusivity

R_1 = sample size

t_1 = experiment run time.

The maximum error due to axial heat loss may be estimated from:

$$\text{Error} = \pi^{1/2} \left[\tau^{1/2} + \frac{2(1-aR)}{a \tau_1^{1/2}} (\ln \tau - \gamma - 2n) \right] \exp\left[-\frac{1}{\tau_1}\right] \quad (\text{Blackwell, 1956})$$

where

$$\tau_1 = 4at/\ell^2 \quad \ell = 1/2 \text{ probe length}$$

$$\tau = at/R^2$$

R = radius of probe in cm

$$a = \lambda/\lambda_i, \text{ where } \lambda = \text{thermal conductivity of probe}$$

$$\lambda_i = \text{thermal conductivity of medium}$$

$\gamma = .5772$ Euler's constant

$n = \lambda/RH = \text{dimensionless contact resistance}$

H = heat transfer constant.

If these criteria are met, then

$$\lambda_i = \frac{Qs}{4\pi}$$

Q = power input to needle probe/unit length

s = slope of temperature vs. $\ln(\text{time})$ graph

From an empirical point of view, the line source approximation is valid over the region in the graph where the temperature vs. $\ln(\text{time})$ relation becomes linear. For moist soils a plot of temperature vs. $\ln\left(\frac{t}{t-t_1}\right)$ may be used to check for moisture migration where t_1 is the time at which the power was cut off.

Experimental Arrangement

Thermal conductivity measurements were conducted with a Pittsburgh Corning Thermal Conductivity Probe, Model CS-148. The unit is four inches long and 0.02 inch in diameter. The internal heater is a 0.001 inch Constantan wire, bifilar wound. The temperature sensing element was a Chromel P thermocouple. The probe's small size allowed use of small samples.

It also required that very small amounts of power be put into the probe, resulting in very small temperature rises during the course of the experiments, on the order of 0.1°C . The thermocouple's change in output over the course of an experimental run was thus of order 0.006 mv. This put severe requirements on the stability of the amplification system if the output was to be monitored by a chart recorder. The problem was solved by first nulling the output of the thermocouple with a Leeds and Northrup potentiometer. The difference signal was then amplified using a chopper stabilized Medistor microvolt meter and recorded on a Bausch and Lomb chart recorder. The following procedure was followed for the measurements.

1. The sealed sample with probe in place was placed in a temperature controlled oven for 24 hours prior to measurement.
2. The temperature was monitored for drift for at least 20 minutes prior to an experimental run which typically lasted for about three minutes.
3. The data corrected for drift was plotted on semi-log paper and the linear segment was used to calculate the sample's thermal conductivity.
4. The cooling data was plotted to check for errors due to moisture migration.

The following experimental results were obtained:

Sample	% Saturation	Temperature $^{\circ}\text{C}$	System Drift $^{\circ}\text{C}/\text{sec}$	Input Power (watts)	Thermal Conductivity ($\text{mcal}/\text{cm sec}$)
Core HC	60	22.5	$+1.42 \times 10^{-4}$	4.62×10^{-2}	1.64
Core HC	60	24.8	$+1.52 \times 10^{-4}$	4.65×10^{-2}	1.70
Packed Sample	60	22.2	-1.00×10^{-4}	4.62×10^{-2}	1.81
Packed Sample	60	26.0	-1.67×10^{-4}	4.63×10^{-2}	1.77
Packed Sample	100%	20.0	$+5.56 \times 10^{-5}$	4.62×10^{-2}	2.19

Using equation 8 and 9 we may calculate the average thermal conductivity of the matrix minerals from the experimental thermal conductivity. Of the saturated sample, the result being 3.25 mcal/cm sec, which is in the range of silica glass. Using this value for the 60% saturated case, the bulk thermal conductivity at 24°C is 1.78 mcal/cm sec and compares to a mean value from the experimental measurements of 1.73 mcal/cm sec $\pm$.08. DeVries method is thus used as a means of estimating the thermal conductivity over the range of water contents and temperatures found in the field. Some typical values are listed below.

% Saturation		Temperature (°C)	λ B=1 (mcal/cm sec)	λ B=5 (mcal/cm sec)
typical	60	10°	1.70	1.89
typical	60	80°	2.62	3.45
	100	10°	2.14	2.14
	100	80°	2.36	2.36

Fumarole Field at Base of 1964 Dome

Structure

A prominent fumarole field (F.F., Figure T1) is located at the base of the 1935 and 1964 lava domes. It is a small region of irregular topography, formed by rock slides from the 1964 dome. Figure T8 shows an idealized cross-section of the field; it is composed of three distinct layers:

1. A relatively impermeable basement of andesite.
2. An extremely permeable layer of dome crumble breccia which is the major passage way for hot water vapor, water and air. It is composed of angular clasts of widely varying size, with large open cavities between the clasts.

Figure T7: Temperature measurement points, drill hole location, fumaroles, and effective conductive heat flow map for fumarole field F.F. at the base of the 1964 lava dome.

FUMAROLE FIELD CROSS SECTION

Figure T8: Idealized cross-section of fumarole field F.F. at the base of the 1964 lava dome.

3. An ash layer of andesitic-dacitic composition caps the irregular surface of layer 2; however, some of the larger angular blocks of layer 2 penetrate to the surface. Typically, fumaroles occur where the ash cover is penetrated by the breccia.

Fumarolic activity is effusive with no well-defined jets, no vent temperatures above the boiling of water, and barely detectable (odor) amounts of SO_2 . The fumarole temperatures Table T6 decrease from 95°C (boiling point at this elevation) for fumarole 10 at the base of the 1964 lava dome to 43°C for fumarole 1 at the edge of the fumarole field (Figure T7). The majority of the fumaroles are located along a nearly linear rubble flow between fumaroles 10 and 1.

Temperature Measurements and Heat Flow

Shallow temperatures were measured at 142 points in fumarole field F.F., at 7.6 cm, 20.3 cm, and 30.5 cm depth and in 7 shallow pits. The data is listed in Tables T7-9 and the locations are given in Figure T7. In addition, air flow calorimetry was used at pits H6 and H7 and on fumaroles 1 and D1. Initially, it was hoped that an empirical relation between ground temperature and heat flow could be obtained using air flow calorimetry; however, failure of the fan on the air flow calorimeter after only a few measurements had been made precluded getting sufficient data to derive such a relationship.

A minimum estimate of heat flux was obtained using Fourier's law and the effective conductivity of the Augustine ash. We have assumed that the water vapor is being transported by simple diffusion ($B=1$). The temperature gradient was obtained from the 7.6 cm, 20.3 cm, and 30.5 cm

TABLE T6

FUMAROLE TEMPERATURES IN FIELD F.F. (FIGURE T1)

1975

Fumarole No.	25 July 1975 (°C)	26 July 1975 (°C)	27 July 1975 (°C)	28 July 1975 (°C)	26 Sept. 1975 (°C)
1	43.0	----	----		
2	38.9	----	38.5		
3	65.0	----	63.0		
4	58.5	----	50.5		
5	74.0	----	74.9		
6	68.5	----	----		
7	82.5	----	82.6		
8	82.4	----	82.5		
9	94.9	----	95.0		95.0
10	60.0	----	22.0		
11	82.8	----	84.8		
12	71.5	----	72.1		
13	87.6	----	87.5		
14	86.5	----	87.3		
15	93.0	----	93.5		
D1	----	69.9	70.0	70.0	70.0

TABLE T7
GROUND TEMPERATURES IN FIELD F.F. (FIGURE T1)
August 3, 1975

Line No.	(FT)	30.5 cm (°C)	20.3 cm (°C)	7.6 cm (°C)
1 - 5	5	27.5	21.5	12.0
	10	20.0	16.5	10.1
	15	21.0	17.5	10.0
	20	--	--	13.8
	25	24.5	21.5	13.0
	30	22.1	19.0	11.0
	35	23.5	20.7	14.5
	40	18.6	15.0	9.8
	45	17.2	15.0	9.0
	50	13.5	10.3	7.1
	55	11.0	9.6	7.0
	60	17.9	15.0	10.0
	65	14.9	11.8	7.8
	70	16.0	12.4	8.5
	75	10.0	9.1	7.0
	80	12.2	8.6	7.2
	85	12.0	10.0	7.2
	90	14.0	10.9	8.1
	95	16.5	13.5	8.6
	100	16.9	13.0	8.6
5 - 6	5	32.0	22.6	13.0
	10	31.2	21.6	13.1
	15	34.0	25.4	14.9
	20	--	23.4	15.1
	25	39.8	29.0	18.8
	30	--	28.5	14.6
5 - 7	5	39.1	33.8	18.0
	10	38.8	31.4	19.0
	15	34.9	26.9	15.2
	20	--	--	19.6
	25	--	--	--
	30	--	25.1	15.1
	35	--	--	30.5
5 - 8	5	41.5	29.6	15.1
	10	56.8	38.1	19.9
	15	--	--	23.9
	20	--	25.0	14.8
	25	--	34.9	19.0

Table T7 (Cont'd)

H1 - (3-13)	0		16.7	15.1
	5		17.9	16.0
	10		26.5	20.0
	15		19.3	16.9
	20		15.0	14.7
	25		13.0	13.0
	30		10.0	12.1
	35	10.2	13.3	13.5
	40		11.2	12.1
	45		11.3	12.8
	50		18.8	16.0
	55		53.5	32.0
5 - 15	5		26.4	18.7
	10		30.0	20.3
	15		30.1	20.9
	20		30.0	21.1
	25		57.9	37.9
	30		--	--
	35		78.5	49.9
	40		--	26.9
	45		60.0	41.0
	50		55.9	31.0
	55		--	55.0
	60		--	51.9
	65		--	54.1
	70		41.1	27.4
	75		47.4	28.2

TABLE T8
GROUND TEMPERATURES IN FIELD F.F. (FIGURE T1)

Line No.	(FT)	20.3 cm (°C)	7.6 cm (°C)
5 - 4	5	23.3	14.4
	10	17.0	11.4
	15	16.4	9.7
	20	25.5	14.9
5 - 8	5	37.1	20.1
	10	32.3	17.8
	15	29.6	17.9
	20	--	21.1
	25	35.6	19.1
	30	--	25.0
	35	48.4	25.9
	40	35.5	20.9
	45	38.4	21.9
	50	--	--
5 - D1	5	29.6	18.5
	10	28.6	16.0
	15	32.0	21.2
	20	24.2	16.0
	25	18.4	13.5
	30	24.6	17.1
15 - D1	5	90.2	82.9
	10	90.4	91.0
	15	90.4	82.7
	20	--	41.3
	25	71.6	54.7
	30	34.0	22.5
	35	53.3	35.4
	40	36.0	22.0
	45	26.3	18.8
	50	29.2	18.0
	55	25.4	17.0
	60	23.6	15.1
	65	27.5	12.4
3 - 13	5	--	22.1
	10	23.4	15.9
	15	17.0	14.0
	20	15.3	12.5
	25	25.7	18.3
	30	18.9	15.0
	35	17.6	14.5
	40	12.0	10.9
	45	--	--
	50	12.8	10.9
	55	10.9	10.6
	60	10.6	11.0
	65	16.9	13.8
	70	31.4	20.0
	75	39.3	24.7
10 - 12	5	--	30.5
	10	--	--
	15	--	--
	20	--	18.6
	25	--	--
	30	--	87.0
	35	--	52.8
	40	61.6	35.2
	45	27.0	19.5
	50	33.0	19.2
	55	--	25.4
	60	48.9	32.1
	65	18.7	15.0
	70	12.0	11.5
	75	11.2	11.1
	80	8.8	9.5
	85	13.3	12.0
	90	15.1	13.0
	95	21.1	13.5

TABLE 9

FUMAROLE FIELD - PIT TEMPERATURES

Pit #	Date	Depth (in)	Temperature (°C)
H1	July 27, 1975	1"	21.0
		2	22.0
		4	23.5
		6	26.0
		8	28.1
		10	30.0
		12	33.2
		14	34.9
		16	37.0
		18	39.0
H2	July 26, 1975	1"	16.0
		2	17.0
		4	19.0
		6	21.0
		8	23.0
		10	25.0
		12	26.5
		14	28.5
		16	30.3
		18	33.1
HC	August 5, 1975)	1"	22.0
		3	23.0
		5	25.0
		7	27.1
		9	29.2
		11	32.1
		13	34.0
		15	37.8
		17	40.1
		19	42.8
		21	45.0
H3	July 26, 1975	1"	18.0
		2	18.5
		4	20.1
		6	22.5
		8	24.9
		10	28.2
		12	29.9
		14	31.8
		16	34.5
		18	37.2

Table 9 (Cont'd)

Pit #	Date	Depth (in)	Temperature (°C)
H4	July 26, 1975	1"	23.1
		2	22.8
		4	23.0
		6	25.0
		8	27.8
		10	30.0
		12	33.3
		14	36.9
		16	39.2
H5	July 26, 1975	18	42.3
		3"	33.2
		5	40.9
		7	48.3
		9	57.0
		11	62.2
		13	68.0
		15	73.9
		17	79.0
		19	82.1

measurements and the effective thermal conductivity is calculated at the mean temperature of the measurements. The heat flow values range from about 4000 HFU near the 1964 dome to 500 HFU at the site of the drill hole, 46 m from the base of the dome. In detail, the heat flux may vary by a factor of up to 10 over a five foot distance. Shallow pits dug into the field revealed that this was the result of cavities in layer 2 being covered with a very thin layer of ash. A minimum heat flow map for the field is shown in Figure T9.

An estimate of the maximum heat flux conducted and advected through the ground was obtained via air flow calorimetry at position H6 (Figure T9). In this region fumarole temperatures were at the boiling point, the ground temperature at 7.6 cm was 51°C, and there was a minimum conductive flux of 2640 HFU. With the air flow calorimeter a heat flux of 19,500 HFU was derived. Assuming that heat flux is a unique function of ground temperature as Dawson (1964) suggested, only five points have a heat flow greater than 19,500 HFU.

The maximum heat flux advected through the ground, assuming that the flux is uniform over the entire field at 19,500 HFU, is 3.6×10^5 cal/sec. The average effective conductive flux over the entire field is 1306 HFU $\pm$ 1007 based on the mean of all 142 temperature measurements. The minimum total heat output through the ground is thus 2.5×10^4 cal/sec. The heat output of fumarole 1 was found to be 8.7×10^2 cal/sec while that of D1 was 1×10^4 cal/sec; i.e., the fumaroles emit a significant amount of energy.

Isotope Ratios

D. Johnston (1978) condensed steam from fumarole 1 in August 1975 and found an isotope value of $\delta^{18} = -19.50/00$, $\delta D = -1480/00$. These values

fall on the meteoric water line, indicating a meteoric source for the steam with no significant rock-water interaction. Johnston inferred from this that the circulation must be shallow and/or rapid.

Drill Hole

We originally planned to probe the 1964 lava dome, but logistics and weather forced locating the drill hole in fumarole field F.F. (Figure T1). The drill hole was located 51 m from the base of the '64 dome and penetrated 24.7 meters into the fumarole field. The down hole temperature was logged with a YSI-Sostman Platinum Probe 539X and a Leeds and Northrup Wheatstone Bridge. The probe was a four-terminal device; thus, lead resistance could be compensated for by taking two sets of measurements. The accuracy of the temperature and depth measurements are $\pm 0.25^{\circ}\text{C}$, ± 0.05 meters respectively. Figure T9 shows the results of two temperature loggings, one on August 4, 1975, and a second on September 26, 1975. The August 4, 1975 data shows a maximum temperature of 46.8°C at 6.4 meters, while the maximum on September 26, 1975 was 48.9°C at 4.4 meters. The 2.1°C change in temperature and 2 meter change in the position of the temperature maximum are probably the result of small changes in the circulation of hot vapor around the hole after drilling had stopped. Both maximum values are within the range that would be expected, considering the drill hole's position in the fumarole field.

If we look at the entire temperature profile, from 0-3.5 meters the data is influenced by convection inside the casing and incomplete sealing outside the hole due to the use of an initially larger diameter drill. Below nine meters the temperature begins to decrease linearly indicating a conductive regime with no additional heat sources down to a depth of 24.7 m. The nine meter point is interpreted as the bottom of the vapor table. The drill hole data thus gives further substance to the idea that heat and water vapor are flowing laterally from the 1964 lava dome.

Figure T9: Temperature vs. depth log for drill hole in fumarole field F.F. located 51 m from the base of the 1964 lava dome. Photograph shows long year drill rig at 4000 ft level of Augustine Volcano, agitation of vapor table from drilling vibrations causes increased fumarole activity; snow covered '64 dome in the background. The first 3 m were drilled with N-rod and a 3 1/2 inch tri-cone bit, the remaining 21.7 m were drilled with A-rod (1 1/2 inch o.d.) and a 2 15/16 inch tri-cone bit.

Summary of Fumarole Field Data (F.F., Figure T1)

We found that:

1. No fumaroles are above the boiling point, and those that are near the boiling point are at the boundary of the 1964 dome and the field.
2. The fumarole vapor temperature decreases in a near monotonic fashion as one moves away from the 1964 dome.
3. The minimum average heat flux for the field is $1306 \text{ HFU} \pm 1007$.
4. The vapor table is no more than nine meters thick at the drill hole site.
5. The source of steam is meteoric in origin with shallow and/or rapid circulation.

The fumarole field is the result of steam traveling outward from the 1964 dome through the cavernous dome breccia which is capped with an ash layer of varying thickness. The steam quickly cools and mixes with air as it diffuses and convects outward. The fumaroles are channels within the breccia uncovered by ash; they serve both as exit ports for the hot water vapor and entry points for ambient temperature and pressure air.

The Summit Moat Fumarole Fields

The position of the lower summit moat fumarole field MM is shown in Figure T1. It is located in a moat which partially rings the summit of the 1964 lava dome. Endogenous domes often consist of a series of inverted nested cones (e.g., MacDonald, 1972, p. 116). Concentric moats and arches are the surface expression of this structure and can clearly be distinguished in Figure T1. No temperatures were measured above 95°C and there were no well defined jets; however, there was a large amount

of crystallized sulfur evident in the regions with near boiling point temperatures. A list of the shallow temperatures measured is given in Table T10. The mean minimum average heat flux was $2730 \text{ HFU} \pm 1891$ which is statistically indistinguishable from the measurement at fumarole field F.F.; however, we found that 17% of the measurements at 7 cm are above 51°C as compared to only 3.5% at fumarole field F.F.. Thus it is probable that the total heat flux is significantly higher on the summit of the 1964 dome. It should be noted that the minimum heat flux calculated from the IR data is in agreement with that found by direct measurements.

TABLE T10

GROUND TEMPERATURES IN 1964 DOME MOAT M.M. (FIGURE T1)

Line No.	(FT)	20.3 cm (°C)	7.6 cm (°C)
1	5	19.0	18.1
	10	16.3	16.4
	15	14.8	14.1
	20	16.3	15.0
	25	20.0	16.1
	30	30.1	19.9
	35	30.5	21.9
	40	30.0	20.9
	45	26.1	19.5
	50	30.3	21.3
	55	43.6	29.0
	60	66.6	41.1
	65	61.8	40.5
	70	44.5	28.2
	75	33.5	23.1
	80	33.0	20.5
	85	44.2	28.6
	90	52.5	29.8
	95	61.0	35.0
	100	46.0	26.0
	105	67.0	49.5
	110	86.0	62.6
	115	--	32.0
	120	--	--
	125	--	28.0
	130	44.0	25.8
	135	--	29.0
	140	57.0	33.9
	145	89.9	51.9
	150	94.1	76.0
	155	79.5	42.3
	160	77.5	51.1
	165	94.5	45.5
	170	88.9	62.0
	175	66.0	42.5
	180	33.0	23.8
	185	32.0	21.8
	190	64.0	42.3
	195	59.8	35.7
	200	58.6	33.2
2	5	39.0	24.1
	10	42.0	27.1
	15	--	29.2
	20	50.0	32.8
	25	64.0	43.0
	30	63.9	40.8

Table T10 (Cont'd)

Line No.	(FT)	20.3 cm (°C)	7.6 cm (°C)
	35	67.7	39.1
	40	--	45.5
	45	--	47.5
	50	40.9	25.9
	55	37.0	23.5
	60	84.0	48.5
	65	87.7	51.3
	70	69.0	39.7
	75	79.5	47.2
	80	63.0	37.9
	85	62.6	38.1
	90	58.5	36.9
	95	--	44.0
	100	--	57.3
	105	--	52.0
	110	--	26.8
	115	50.0	47.0
	120	--	66.0
	125	93.6	93.2
	130	93.0	79.2
	135	71.0	46.1
	140	59.7	34.3
	145	88.5	65.1
	150	51.0	31.9
3	0	18.0	10.5
	5	17.0	12.0
	10	--	--
	15	11.5	9.0
	20	10.5	9.0
	25	--	9.0
	30	--	8.0
	35	14.5	10.0
	40	19.0	12.0
	45	23.5	14.5
	50	--	13.1
	55	--	14.2
4	5	15.8	11.5
	10	17.0	11.5
	15	--	12.5
	20	24.5	9.5
	25	33.5	20.0
	30	31.5	17.0
	35	21.8	11.0
	40	16.5	10.5
	45	43.0	25.0
	50	--	27.5

2. ACTIVE SEISMIC EXPERIMENT

Objectives

The objective of the active seismic experiment on Augustine was to determine the shallow structure of the volcano and to try to locate molten or partially molten rock.

Seismic Properties of Magma

In order to try to locate magma by seismic means one must first know the velocity of seismic waves in molten or partially molten rock. This problem is more complex than it might first appear. Aki et al. (1978) conducted passive and active seismic experiments on the partially frozen lava pond of Kilauea Iki and made two important discoveries: (1) Explosion P-wave data showed that the magma lens beneath the frozen upper crust had extremely low P-wave velocities in the range of 0.2 to 0.6 km/sec and (2) surprisingly, the magma lens transmitted SV-waves from distant earthquakes with periods of about 0.1 seconds.

Love and S-wave data indicated a less than 10 m thick magma lens which in response to the weak explosion signals behaved like a viscous liquid with apparent viscosity of about 10^7 poise and apparent shear velocity of about 0.2 km/sec. Shaw et al. (1968) made in situ viscosity measurements in the nearby Makaopuhi lava lake and found an apparent high viscosity at low stress levels, which they attributed to the presence of vesicles. A few volume percent of vesicles in the melt apparently causes it to behave like a Bingham body, i.e. as a solid below threshold stress, and as a liquid above it. The P-wave velocity in such a viscous liquid is $\sqrt{\frac{k + \frac{8}{3} n \omega}{\rho}}$ and the S-wave velocity $\sqrt{\frac{2n\omega}{\rho}}$, where n is the

viscosity of the liquid, k is the bulk modulus, ρ is the density and ω is the angular frequency of the propagating wave. Aki et al. feel that the extremely low P-wave velocity of the Kilauea Iki magma lens can be explained if the bulk modulus k is small compared to the $\frac{8}{3} \eta \omega$ term. A few volume percent of vesicles in the melt can reduce the apparent bulk modulus to as low as the apparent rigidity, inferred from Love and S-wave data, resulting in P-velocities as low as 0.3 km/sec for a melt containing about 5 percent vesicles. Such a low P-wave velocity is in disagreement by a factor of about 7 with laboratory measurements of P-wave velocities, about 2 km/sec in bubble-free basaltic melts, which behave like a Newtonian fluid (Murase and McBirney, 1973).

According to Aki et al., the apparent bulk modulus k of a mixture consisting of liquid and gas with bulk moduli of k_l and k_v respectively is $\frac{1}{(1-\epsilon) k_l^{-1} + \epsilon k_v^{-1}}$, where ϵ is the volume fraction of gas. For an adiabatic process, k_v is equal to $\gamma \times P$ where γ is the ratio of the heat capacity at constant pressure to the heat capacity at constant volume and P is the pressure (Officer, 1958). For example, the bulk modulus of water vapor contained in melt near the surface at say 15 bar and 1150°C would be 1.8×10^7 dyne cm^{-2} . The bulk modulus of vesicle-free liquid andesitic melt can be calculated from the P-wave velocity, density and viscosity. With values given by Murase and McBirney (1973) for andesites, a bulk modulus of 1.5×10^{11} dyne cm^{-2} is obtained. As little as two volume percent of vapor filled vesicles in the melt will decrease the bulk modulus by two orders of magnitude to 1×10^9 dyne cm^{-2} .

The other parameter which critically controls the P-wave velocity of rock melt is viscosity, which is strongly controlled by the solid and

gas phases contained in the liquid. Murase and McBirney (1973) state that the viscosity of naturally occurring magmas cannot be reliably predicted from their chemical composition. For lack of better data, we assume 10^7 poise, the viscosity found by Aki et al. for Kilauea Iki, for near surface magma at Augustine.

Assuming a bulk modulus of 10^9 dyne cm^{-2} , a viscosity of 10^7 poise, and a density of 2.45 g cm^{-3} (Murase and McBirney, 1973) for vesicular, shallow melt at Augustine, and a frequency of 10 Hz for seismic P-waves, we estimate that the P-wave velocity of the melt could be as low as 0.33 km/sec. Allowing viscosities of 10^8 poise only increases the P-wave velocity to 0.8 km/sec. Thus, if the magma contains as little as two volume percent of gas bubbles it will be characterized by an extremely low P-wave velocity. Most of this gas is likely to be water.

The solubility of water in a silicate melt is a function of pressure and therefore depends on the depth of the magma. Measurements of the solubility of water in andesitic melts suggest that if a magma body contains as little as 3 percent water by weight, the melt will become saturated at a pressure of 500 bars (Burnham, 1972), which is equivalent to a depth of 2 km, assuming the stresses are lithostatic and the average density of the crust is 2.55 gm/cc . Burnham (1972) suggests that 3 percent by weight of water is a minimum value for andesitic melts based on petrographic evidence and energy considerations. Consider an andesitic magma with 3 percent water content by weight that is rising from depth: When the magma reaches a depth of 2 km below the surface it will become saturated and degassing begins. Most likely not all the gas will escape.

from the melt as evidenced by the Kilauea Iki lava pond which still contains considerable quantities of vesicles, in spite of the fact that (a) the lava at Kilauea Iki was extruded onto the surface more than 17 years ago in 1959, (b) basaltic magma has, at least for large applied shear stresses, a much lower viscosity than andesitic magma, and (c) basaltic magma has a much lower volatile content than andesitic magma. Consequently, andesitic magma above 2 km depth at Augustine is likely to be vesicular and characterized by very low seismic P-wave velocities approximately equal to the S-wave velocity and magma deeper than 2 km is probably characterized by velocities of order 2.5 km/sec, the measured ultrasonic P-wave velocity of molten andesite without vesicles (Murase and McBirney, 1973).

Field Experiment

From August 4-6, 1975, we conducted a seismic refraction experiment in cooperation with Sandia Laboratories, Albuquerque, N.M., to locate low velocity regions indicative of magma at shallow depths beneath Augustine Volcano. Sandia had an interest in the experiment as the group that is presently spearheading research on direct heat extraction from rock melts (e.g., Colp et al., 1976; Hardee and Larson, 1976).

Layout:

Since our prime target for finding molten or partially molten rock was the 1964 summit lava dome and associated magma conduit, we had originally planned to use a simple fanned array of geophones. A few explosions would have been set off on Augustine Island and recorded across two lines of seismometers, one placed in front of the central conduit and dome, the other behind it. By shooting in different

quadrants of the volcano and at different elevations, the position of regions of extreme low P-wave velocity could then be inferred by comparing pairs of stations. This plan would have required that the stations be moved after every shot in order to obtain optimum array configurations. In practice, this procedure proved to be impossible because of the difficult terrane and the limited battery life of the twelve portable seismic stations that Sandia Laboratories provided. Instead, we decided in the field to combine the permanent University of Alaska seismic array and the portable Sandia stations into one fixed array and move only the shotpoints. This arrangement allowed us to increase the number of explosions to 10 and to complete the shooting in two days with only one battery change in between. The price paid for the logistic convenience was much greater difficulty in interpreting the results.

The position of the ten shotpoints, the five-station Augustine permanent array (AU1-5) and the twelve-unit portable Sandia array are shown in Figure AS1. One of the permanent Augustine stations, AU4, is not shown since it was not operating during the fan shooting experiment. In order to synchronize the permanent array with the portable Sandia array, two Sandia stations units 10 and 12, were placed next to stations AU2 and AU3 of the permanent Augustine array. Sandia unit 1 had a horizontal seismometer and was placed next to unit 6.

By using so many shots and stations we could afford to lose some data due to poor arrivals of the signals, poor telemetry or other unforeseen electronic or logistic problems that we anticipated to develop in a terrane as difficult as that found on the very steep slopes and cliffs of Augustine Volcano.

SEISMIC FAN SHOOTING

August, 1975

Figure AS1: Layout of Augustine seismic refraction experiment, shot points-hexagons, portable Sandia array (12 stations)-circles, permanent Augustine array (5 stations)-squares.

Shaped charges were used in shots 1 - 6 (Figure AS1). Conventional explosives were used for 4 additional shots (7 - 10, Figure AS1): Shot 7 was a fifty-pound petrojel water shot in an embayment at Burr Point at the northern end of Augustine Island. Shot 10 consisted of 100 pounds of petrojel which we detonated in an ice cave near the summit of Augustine Volcano. For shots 8 and 9, we detonated a pile of 200 pounds of excess petrojel on the beach near the water's edge.

Pearson (1977) compared in detail the coupling of energy of the three different types of charges into the ground. The shaped charges (1 - 6) were clearly more efficient than the ordinary surface charges (8, 9) but both types were less effective than the one shot in the water (7).

Detectors, Telemetry and the Recording System

Sandia Array of Detectors

The twelve portable short period seismic stations provided by Sandia Laboratories consisted of a 4.5 Hz Geospace seismometer, a preamplifier, a voltage controlled oscillator (VCO), a transmitter, and a power package. Nine of the preamplifiers had both a high and low gain output. For these nine stations, the high and low gain outputs were modulated by different VCO's, with center frequencies of 2.3 KHz and 1.7 KHz respectively. The two FM signals were mixed and then fed into the transmitter. For the other three stations, only the high gain output was available. The shot break was amplified and broadcast using similar equipment. The only difference was that the VCO frequency for the shot-break was 1.3 KHz. Figure AS2a shows the overall frequency response curve of the system.

Of the total of 117 potential rays, (9 shots times 13 sites; shot 2 was a repeat of 1) 66 were used in the final analysis, the remainder having poor arrivals or not being recorded (e.g., Sandia unit 4 never functioned).

Energy Source

Of the ten charges, six were shaped charges, one was a water shot, and three were conventional surface charges. Shaped charges are explosive charges with a molded cavity at one end and a cap and primer at the other. These charges concentrate energy axially. If placed vertically, most of the energy can be directed down into the ground rather than being radiated omnidirectionally as in conventional charges, thus eliminating the need for drilling shotholes in order to insure good coupling between the explosion and the ground. The use of shaped charges in exploration geophysics is fairly new. The technique was pioneered by Petróleo Brasileiro in 1969 and has been used routinely by that company since 1973 (Bahia and Silvia, 1973).

The six shaped charges were designed by MSgt. Roy Fehrenbach of the U.S. Army, Fort Wainwright, Fairbanks, and fabricated at the Geophysical Institute. We welded two five-gallon oil drums with tops and bottoms removed together, and used an inverted cone made from sheet copper as the bottom cavity of the charge. The shaped charge was loaded at Augustine Island with 135 pounds of C-4, a commercially available explosive. The explosive was later detonated from the top end of the charge by an electric blasting cap and primer.

Log of the gain plotted vs. log of the frequency, for the Sandia stations.

Frequency response curve for the Augustine stations.

Figure AS2: System responses. Left: Portable Sandia stations, log gain versus log frequency. Right: Permanent Augustine stations, frequency versus magnification.

Permanent Augustine Seismic Array

The permanent Augustine short period seismic array consisted of five stations. Each station included a 2 Hz vertical seismometer, an Emtel preamp VCO unit and a VHF transmitter. Station AU2H had an additional horizontal unit. Figure AS2b shows the overall frequency response curve of the system.

Telemetry

The data from the portable Sandia stations and the shotbreak were transmitted in the VHF-FM mode through simple, nondirectional dipole antennas to an airplane (Beaver) circling over Augustine Island at an elevation of about 7,000 feet. By transmitting the data to an airplane, Sandia was able to solve the severe line of sight problems that almost inevitably result when one is telemetering data in mountainous terrane. The data from the permanent stations of the Augustine array were transmitted in the VHF-FM mode to Homer, 100 km east of Augustine Island, through five element directional Yagi antennas.

Recording

The signals from the portable Sandia array were recorded on a 14 channel magnetic tape in the airplane at a speed of 60 i.p.s. Of the 14 channels on the Sandia magnetic tape, 12 contained seismic data, one was a voice channel and one contained a time signal. The voice channel was used by Sandia technicians to identify the shots. The signals from the Sandia stations were recorded without being discriminated. Since the high and low gain signals were modulated by VCO's with different center frequencies, it was possible to record both the high and low gain signals

from each station on one data channel. The shot break was recorded on a data channel along with the signal from one of the single gain seismic stations (unit 1). The signals from the permanent Augustine array were recorded on 16 mm film and magnetic tape in Homer.

Playback and Scaling of the Seismic Signals

The Sandia magnetic tape was played back on a 14 channel tape recorder at the Geophysical Institute. The seismic signal and the shot break were discriminated, and then recorded on a 2 channel "Brush" recorder. Because of excessive electronic noise, it was often necessary to filter the discriminated data through a band pass filter, set to pass signals between 0.2 and 20 Hz. Since first arrivals generally showed frequencies of 10 ± 2 Hz the seismic signals were not distorted by the filters. The time delay between the first arrivals and the shot break were scaled manually from the "Brush" records. Repetition showed that this measurement had a precision of better than 0.01 seconds. The accuracy of the measurement varied from shot to shot depending on the quality of the first arrival.

The first arrivals from the permanent Augustine seismic array were measured directly from the screen of a 16 mm "Geotech" film viewer. The precision of this measurement was about 0.02 seconds.

Data Analysis

Introduction

The interpretation of the Augustine seismic refraction data was complicated by the severe topography and heterogeneity of the volcano. As a first step in the analysis we divided the 66 usable seismic rays into several subsets that sampled known morphologic or geologic subdivisions.

Outer Rays: A major subset of rays crossed the lower flank of the volcano outside the 1000 foot contour line, which marks a fairly sharp break in the slope of the volcano. Most of these rays, being fairly long, were critically refracted from the underlying basement of Augustine Volcano which, at least on the south side of the volcano, is known to be sedimentary. These sediments crop out on the southern flank of the volcano up to the 1000 foot contour line and consist of shallow marine sandstones and shales of the Upper Jurassic Naknek formation in fault contact with Upper Cretaceous Kaguyak formation (Detterman, 1973; Buffler, 1976). The sediments probably also underlie the northern half of the Island. Pyroclastic flows and mudflows form the bulk of the overlying debris found on the outer flanks of the volcano. The outer rays were analysed using plane layered models.

Inner Rays: Inside the 1,000 foot contour lies the more vertical cylindrical core structure of the volcano. A small group of rays traversed this structure above the 1,000 foot contour line more or less horizontally and makes up a second subset, henceforth called Inner Rays. This data set was analysed using a cylindrical ray tracing routine.

Mixed Rays: A third subset consisted of rays that crossed from the outer to the inner regions, or vice-versa. Most of these rays were critically refracted from the basement and sampled the lower inner structure and conduit system of the volcano (2,840 ft. to basement).

Each of these three principal subsets of rays and some other small subsets that sample specific regions of the volcano will be discussed separately in the following paragraphs. Tables AS1-7 are listings of all the subsets of rays used in the analysis.

Figure AS3: Plan view of ray paths of a: inner rays, traversing the Augustine cone above the 1000 ft. contour line
 b: outer rays, starting and ending outside the 1000 ft. contour line
 c: mixed rays, passing from the inner to the outer parts of the volcano or vice versa, crossing the 1000 ft. contour line
 Contour Interval 500 ft. up to 3000 ft.

Outer Rays

The outer rays (Figure AS3a and Table AS1) have shot and receiver locations that lie outside the main volcanic core (outside the 1,000 ft. contour).

A short unreversed refraction profile, shot in 1974 near Burr Point at the northern end of Augustine Island, revealed a simple two-layer horizontal structure as shown in Figure AS4. The upper pyroclastic layer has a velocity of 0.73 km/sec and extends to 52 meters below sea level and overlays a second layer with a velocity of 2.6 km/sec. A plot of the raw travel time data for the outer rays, uncorrected for elevation, revealed a third layer with a velocity of about 5 km/sec. We therefore adopted a plane-layered structure of three layers to model the outer structure of Augustine Volcano.

The velocity of the pyroclastic apron making up the lower flanks of most of the island was measured at Burr Point (0.73 km/sec). Seismic ray 10-3 (1.27 km/sec) and ray 1-7 (1.08 km/sec) sampled the pyroclastics higher up on the volcano. We felt that the Burr Point velocity was somewhat low because the refraction profile was shot in finer material between prominent ridges of coarser material within the pyroclastic flow that makes up the Point. We chose an average value of $V_1 = 1.2$ km/sec from rays 10-3 and 1-7 for the layer 1 pyroclastics.

Unfortunately, only the Burr Point data allows an estimate of the depth to layer 2, which we interpret to be most likely sedimentary. At that location it lies at a depth of 52 meters below sea level. Since bathymetric and geologic evidence indicates that the volcano is built on the sedimentary platform that underlies Kamishak Bay at very shallow water depths (less than 30 m deep) south and west of the volcano, layer 2 was

Figure AS4: Seismic refraction profile at the northernmost extension of Augustine Island (Burr Point), shot in 1974 by J. Kienle (unpublished data). A 2.6 km/sec refractor underlies a very low velocity, 0.73 km/sec, surface layer at a depth of 52 m.

TABLE AS1. Outer Rays
Critically Refracted By 5.1 km/sec Layer For Split Layer 2 Structure

	$T_o^{(1)}$	$D_h^{(2)}$	$T_o - T_c^{(3)}$
7-AU2	1.76	5.41	-.06
7-AU5	1.53	3.87	+.06
7-7	2.14	7.54	+.05
7-AU3	2.37	8.81	+.02
8-AU2V	1.95	6.04	+.01
8-AU3	2.03	7.34	-.03
8-7	1.48	4.37	+.01
6-AU3	1.42	4.04	-.04
6-AU5	1.90	5.64	+.04
6-7	2.06	7.06	+.02
9-AU5	1.71	5.40	-.06
9-AU2V	2.51	8.96	-.01
9-AU3	1.96	6.70	+.02

Mean = $0.00 \pm .04$

Critical Refracted By 2.6 km/sec Layer

8-AU5	0.95	1.52	+.25
9-7	1.00	2.2	+.04
6-AU2	0.82	1.31	+.05

(1) T_o : observed travel time in seconds

(2) D_h : horizontal distance between shotpoint and receiving station in km

(3) $T_o - T_c$: time residual between observed and calculated arrival times in seconds

chosen to begin at sea level for our models. The estimated velocity of layer 2 is $V_2 = 2.6$ km/sec, based on the Burr Point data. Rays 6-AU2, 8-AU5, 9-7 (Table AS1, bottom) suggest a somewhat lower value of 2.1 km/sec, obtained by reducing these rays to a sea level datum using $V_1 = 1.2$ km/sec for the surface layer and $V_2 = 2.6$ km/sec.

A time-distance plot of the elevation corrected data is shown in Figure AS5. This data defines layer 3. All shot points and stations were reduced to sea level using a surface layer velocity of $V_1 = 1.2$ km/sec, except for station AU3 which lies on sediments of velocity 2.1 km/sec (ray 5-AU3). The data from shot 5 were deleted from the determination of the final least square best fit line and will be discussed in a special sub-section. The best least-square fit line indicates a velocity of 5.1 km/sec for layer 3. Assuming that the velocity for layer 2 is 2.6 km/sec the depth to the 5.1 km/sec layer is 0.9 km (with a velocity of 2.1 km/sec for layer 2 the depth becomes 0.6 km).

In detail, Figure AS5 reveals a systematic scatter of data points: The diamonds, scattering above the line, are travel times for rays where both shot point and receiver are located on the north side of Augustine Island (residual 0.052 ± 0.049); the x's are for rays that crossed only the south side and plot systematically below the least square fit (residual -0.135 ± 0.035); and the pluses (+) which fall fairly well on the least square line (residual -0.02 ± 0.03) are for rays crossing from south to north and vice versa. The systematic scatter can be explained if layer 2 has a higher velocity on the south side of Augustine Island. This possibility will be discussed in the sub-section on mixed rays.

Figure AS5: Outer rays (excluding rays from shot 5) travel times, elevation corrected to sea level using 1.2 km/sec, except for arrivals at AU3 which were corrected using 2.1 km/sec. Diamonds for northside rays, X's for southside rays, pluses for rays crossing from N to S or S to N. Least square fit shows 5.1 km/sec refractor at a depth of 600 or 900 m, depending on the velocity used for the first layer, 2.1 or 2.6 km/sec.

We like to stress the point that in calculating the depth to the 5.1 km/sec we have implied that the dip of the interface is zero. Conventionally, the dip of a refractor is calculated by using two perpendicular reversed refraction profiles, a method that we could not use for our layout. Pearson (1977) developed a method to calculate an upper bound on the dip of the 5.1 km/sec refractor and found it to be 2.2 degrees.

Inner Rays

This subset of 11 rays traversed the core of the volcano above the 1,000 ft. contour line (Figure AS3b and Table AS2). Because of the axial symmetry and conical shape of the volcano we had to use ray-tracing methods to interpret the travel times. The structure of the volcano was approximated as a stack of cylindrical structures, each one consisting of a series of nested coaxial cylindrical shells, assumed to be homogeneous isotropic and of constant velocity.

Pearson (1977) wrote a computer program to trace a ray through this type of velocity structure allowing for refraction and reflection at each boundary according to Snell's law and to calculate theoretical travel times. Furthermore, an inversion algorithm was developed which allowed to systematically vary the velocity and radii of the shells in order to minimize the mean square difference between the calculated and observed travel times.

Assume a structure consisting of N vertical coaxial cylinder shells traversed by seismic rays: If T_{oi} is the observed travel time for the ith ray, and T_{ci} is the calculated travel time through the theoretical structure, the mean squared error E between the calculated and observed travel times is:

$$E = \frac{1}{N} \sum_{i=1}^N (T_{ci} - T_{oi})^2 \quad (1)$$

TABLE AS2. Inner Rays

Lower Cylinder

	T_o	D_h	$T_o - T_c$
4-6	0.69	1.43	.00
4-5	1.02	2.33	+.01
4-AUT	1.21	2.93	-.01
3-8	1.29	2.74	.00
Mean = 0.00 ± 0.01			

Middle Cylinder

4-2	1.14	2.57	.00
1-8	1.04	2.03	-.06
1-6	1.09	2.17	+.07
Mean = 0.00 ± 0.07			

Upper Cylinder

10-9	0.50	0.73	+.03
10-11	0.36	0.74	+.01
2-9	0.78	1.01	-.03
2-11	0.71	1.01	-.01
Mean = $0.00 \pm .03$			

If V_j and P_j are the velocity and outer radius of the j th cylindrical shell, T_{ci} can be thought of as a function of V_j and P_j , i.e.,

$$T_{ci} = T_{ci}(V_1, P_1, V_2, P_2, \dots; V_m, P_m)$$

Note, that although there are only m layers, T_{ci} is a function of $2m$ variables. The velocities and radii of the cylindrical layers in the model can be varied to minimize the mean squared error between the calculated and observed travel times. As a first step, the calculated travel times are approximated by the first order terms in the Taylor series, expanded about the point (V_{jo}, P_{jo}) :

$$T_{ci}(V_j, P_j) = T_{ci}(V_{jo}, P_{jo}) + \sum_{j=1}^N \left[\frac{\partial T_{ci}}{\partial V_j} \bigg|_{V_{jo}, P_{jo}} (V_j - V_{jo}) + \frac{\partial T_{ci}}{\partial P_j} \bigg|_{V_{jo}, P_{jo}} (P_j - P_{jo}) \right] \quad (2)$$

Substituting equation (2) back into equation (1):

$$E = \frac{1}{N} \sum_{i=1}^N \left[T_{ci}(V_{jo}, P_{jo}) + \sum_{j=1}^N \frac{\partial T_{ci}}{\partial V_j} \bigg|_{V_{jo}, P_{jo}} (V_j - V_{jo}) + \frac{\partial T_{ci}}{\partial P_j} \bigg|_{V_{jo}, P_{jo}} (P_j - P_{jo}) - T_{oi} \right]^2$$

In order to test for the minimum value of E , the first partials of E with respect to V and P are set equal to zero:

$$\frac{\partial E}{\partial V_k} = \frac{2}{N} \sum_{i=1}^N \left[T_{ci} + \sum_{j=1}^N \frac{\partial T_{ci}}{\partial V_j} \bigg|_{V_{jo}, P_{jo}} (V_j - V_{jo}) + \frac{\partial T_{ci}}{\partial P_j} \bigg|_{V_{jo}, P_{jo}} (P_j - P_{jo}) - T_{oi} \right] \frac{\partial T_{ci}}{\partial V_k} \bigg|_{V_{jo}, P_{jo}} = 0 \quad (3a)$$

$$\frac{\partial E}{\partial P_k} = \frac{2}{N} \sum_{i=1}^N \left[T_{ci} + \sum_{j=1}^N \frac{\partial T_{ci}}{\partial V_j} \bigg|_{V_{jo}, P_{jo}} (V_j - V_{jo}) + \frac{\partial T_{ci}}{\partial P_j} \bigg|_{V_{jo}, P_{jo}} (P_j - P_{jo}) - T_{oi} \right] \frac{\partial T_{ci}}{\partial P_k} \bigg|_{V_{jo}, P_{jo}} = 0 \quad (3b)$$

The terms $\frac{\partial T_{ci}}{\partial V_j}$ and $\frac{\partial T_{ci}}{\partial P_j}$ can then be calculated from numerical approximations of the derivative. Equations (3a), (3b) represent a linear set of $2m$ equalities and $2m$ unknowns. They may be solved for the V_j and P_j using a variety of standard techniques, such as Gauss Jordan Elimination and Gauss-Siedel Iteration (Hornbeck, 1975).

Since we had only a limited amount of data, we modelled the structure of the mountain using only three stacked cylindrical structures. The first structure extends from 1,000 feet to 2,000 feet, the second from 2,000 feet to 2,840 feet, and the third from 2,840 feet to the summit volcano. The division between the second and third cylindrical structure was placed at 2,840 feet because this choice simplified the inversion process; 2,840 feet is the elevation of shots 1 and 2. Therefore, rays travelling up from shots 1 and 2 will have their entire ray path contained within cylinder 3, and rays travelling down from shots 1 and 2 will have their entire ray path contained by the lower two cylinders.

In order to cut down the number of variables in the inversion process we first attempted to obtain an estimate of the thickness and velocity of the outer rubble layer in the upper parts of the volcano.

Given the general axial symmetry and conical shape of Augustine Volcano, the structure can also be approximated by a series of coaxial conical shells each with a different seismic velocity. It is quite difficult to write a computer program to trace a ray through a structure of this kind. However, if the shotpoint, the receiver and the axis of symmetry of the cones are coplanar, that is if the shot is directly up or down the flanks of the cone, the ray will be refracted in a similar way as shots the same distance apart would be refracted by a horizontally layered structure.

TABLE AS3. Flank Rays

	T_o	D_h	$T_o - T_c$
1-AU1	0.93	1.54	-.04
3-11	0.84	1.26	-.01
1-9	0.79	1.01	+.04
10-2	0.75	1.09	-.02
3-9	0.80	0.98	+.08
1-2	0.65	0.89	+.01
1-3	0.51	0.71	-.09
4-AU3	0.81	1.4	-.03

Mean = $-.01 \pm .05$

There were seven rays that fit this criterion (Table AS3). A travel time plot for these rays shows a clear linear trend (Figure AS6) which we interpreted as representing refracted arrivals from an underlying higher velocity layer dipping down the flank of the mountain. The best fit least square line through the data gave a velocity of 2.25 ± 0.5 km/sec. Assuming that the refracting layer lies parallel to the outer flank of the volcano, and that the rubble velocity is 1.18 km/sec (rays 1-7, 10-3) we estimated the rubble layer to be about 200 m thick.

Cylindrical Structure of the 1,000 - 3,000 Foot Region

Four rays passed nearly horizontally through the mountain at the 1,500 foot level (Table AS2 top). Since these rays never left the lower cylinder, they were used to determine its structure which we modeled as three concentric cylindrical shells. The two outer radii and velocities of the outer two layers were known from the results of the flank shots and the general topography of the mountain. With the conical model we determined the thickness of the outermost skin layer at 0.2 km, having a velocity of 1.18 km/sec.

Suppose the average radius of the mountain at some specified elevation is ρ , the average slope is θ and the thickness of the cone-mantling layer is d . At any elevation the horizontal cross-section will be the same for the conical and cylindrical models. If the outer radius of the outermost cylindrical shell is ρ , the inner radius becomes $\rho - d/\sin \theta$. Augustine's upper slopes are about 30° . Since the average radius of the 1,500 foot contour line, the level at which the 4 rays passed the mountain, is 1.88 km, the radius of the middle cylindrical shell becomes 1.48 km. We are now

Figure AS6: Refraction profile along upper flank of Augustine volcano. Least square fit shows 2.25 km/sec refractor, the more consolidated inner core of a conical volcano structure, overlain by 200 m of low velocity (1.2 km/sec assumed) mantling material.

left with only two parameters that can be varied in order to minimize the difference between the theoretical and the observed travel times. These are the velocity and radius of the innermost layer.

For the case where only the velocity and radius of one layer is being varied, equation (2) becomes:

$$0 = \sum_{i=1}^4 (T_{ci} - T_{oi}) \left. \frac{\partial T_{ci}}{\partial V_3} \right|_{\substack{V_3=V_{30} \\ P_3=P_{30}}} + \sum_{i=1}^4 \left(\left. \frac{\partial T_{ci}}{\partial V_3} \right|_{\substack{V_3=V_{30} \\ P_3=P_{30}}} \right)^2 (V_3 - V_{30}) + \sum_{i=1}^4 \left. \frac{\partial T_{ci}}{\partial P_3} \right|_{\substack{P_3=P_{30} \\ V_3=V_{30}}} \left. \frac{\partial T_{ci}}{\partial V_3} \right|_{\substack{V_3=V_{30} \\ P_3=P_{30}}} (P_3 - P_{30})$$

$$0 = \sum_{i=1}^4 (T_{ci} - T_{oi}) \left. \frac{\partial T_{ci}}{\partial P_3} \right|_{\substack{V_3=V_{30} \\ P_3=P_{30}}} + \sum_{i=1}^4 \left(\left. \frac{\partial T_{ci}}{\partial P_3} \right|_{\substack{V_3=V_{30} \\ P_3=P_{30}}} \right)^2 (P_3 - P_{30}) + \sum_{i=1}^4 \left. \frac{\partial T_{ci}}{\partial P_3} \right|_{\substack{P_3=P_{30} \\ V_3=V_{30}}} \left. \frac{\partial T_{ci}}{\partial V_3} \right|_{\substack{V_3=V_{30} \\ P_3=P_{30}}} (V_3 - V_{30})$$

The partial derivative of T_{ci} with respect to V_3 and P_3 were calculated using the standard two point derivative formula:

$$\left. \frac{\partial T_{ci}(V, P)}{\partial V} \right|_{V_{30}, P_{30}} = \frac{T_{ci}(V_{30} + \Delta V, P_{30}) - T_{ci}(V_{30}, P_{30})}{\Delta V} = a_i$$

$$\left. \frac{\partial T_{ci}(V, P)}{\partial P} \right|_{V_{30}, P_{30}} = \frac{T_{ci}(V_{30}, P_{30} + \Delta P) - T_{ci}(V_{30}, P_{30})}{\Delta P} = b_i$$

$$\text{Let } T_i = T_{ci} - a_i V_{30} - b_i P_{30}$$

Substituting a_i and b_i into equation (2):

$$0 = \sum_{i=1}^N (T_i - T_{oi}) a_i + \sum_{i=1}^N a_i^2 V_3 + \sum_{i=1}^N b_i a_i P_3$$

$$0 = \sum_{i=1}^N (T_i - T_{oi}) b_i + \sum_{i=1}^N a_i b_i V_3 + \sum_{i=1}^N b_i^2 P_3$$

There are two equations with two unknowns. They may be solved for V_3 and P_3 using Cramers rule:

$$V_3 = \frac{\begin{vmatrix} -\sum_{i=1}^4 (T_i - T_{oi})a_i & \sum_{i=1}^4 a_i b_i \\ -\sum_{i=1}^4 (T_i - T_{oi})b_i & \sum_{i=1}^4 b_i^2 \end{vmatrix}}{\begin{vmatrix} \sum_{i=1}^4 a_i^2 & \sum_{i=1}^4 a_i b_i \\ \sum_{i=1}^4 a_i b_i & \sum_{i=1}^4 b_i^2 \end{vmatrix}} \quad (4a)$$

$$P_3 = \frac{\begin{vmatrix} \sum_{i=1}^4 a_i^2 & -\sum_{i=1}^4 (T_i - T_{oi})a_i \\ \sum_{i=1}^4 a_i b_i & -\sum_{i=1}^4 (T_i - T_{oi})b_i \end{vmatrix}}{\begin{vmatrix} \sum_{i=1}^4 a_i^2 & \sum_{i=1}^4 a_i b_i \\ \sum_{i=1}^4 a_i b_i & \sum_{i=1}^4 b_i^2 \end{vmatrix}} \quad (4b)$$

The cylindrical structure was then determined using the following iterative scheme. First, we made an initial guess of the structure and then calculated theoretical travel times for each ray using the ray trace program. Next, improved values of V_3 and P_3 , representing more refined guesses at the velocity and radius of the innermost cylinder, were calculated and substituted into the model. More theoretical travel times were calculated which gave again new values for V_3 and P_3 , etc. The process stopped when the root mean square difference between the calculated and observed travel times decreased by less than 0.0015 sec in two successive iterations.

From an initial guess of 2.8 km/sec and 1.3 km for V_3 and P_3 the algorithm converged to 2.56 km/sec and 1.35 km in three iterations. The root mean square difference between the calculated and observed travel times is 0.01 sec (Table AS2 top).

Cylindrical Structure of the Region Above 2,840 Feet in Elevation

The uppermost cylinder starts at the elevation of shot 1 and includes the summit of the mountain. This region of the volcano consists mainly of lava domes, including the 1964 dome, which we suspected to still be hot, perhaps partially molten, the 1935 dome, and minor breccias, rubble and pyroclastic material. Since the average radius of the 3,000 foot contour line is only about 0.7 km, we modeled this region of the mountain with only two concentric cylindrical shells. The outermost shell was assigned a velocity of 1.2 km/sec. This is the average of the velocities of rays 10-3 and 1-7. The inner shell was assigned a velocity of 2.25 km/sec, determined by simple refraction along a cone as discussed above, a value which seems reasonable for the vesicular, fractured andesite of the central domes.

Since the velocities of both shells have already been determined, the only parameter that can be varied in order to minimize the difference between the calculated and observed travel times is the radius of the inner layer. If T_{ci} (the calculated travel time through the theoretical structure) is a function of only one variable (the radius of the inner layer) then equation (2) reduces to:

$$0 = \sum_{i=1}^4 (T_i - T_{ci}) a_i + \sum_{i=1}^4 a_i P_2 \quad (5)$$

where a_i is the two point numerical approximation to the derivative:

$$\left. \frac{\partial T_{ci}}{\partial P_2} \right|_{P_2 = P_{20}} = a_i = \frac{T_{ci}(P_{20} + \delta P_2) - T_{ci}(P_{20})}{\delta P_2}$$

Equation (5) can be easily solved for P_2 :

$$P_2 = \frac{\sum_{i=1}^4 a_i (T_i - T_{oi})}{\sum_{i=1}^4 a_i^2} \quad (6)$$

The iterative process that was used to determine the best value of P_2 was the same as the one used in the last section, except that equation (6) rather than equation (4b) was used to determine the new improved values of P_2 .

Four shots started and ended within cylinder three. The structure was determined from them. From an initial guess of $P_2 = 0.6$ km, the iterative process described above converged to a final value of 0.48 km in three iterations. The root mean square difference between the calculated and observed travel times is 0.02 sec (Table AS2 bottom).

Structure of the Region Between 2,000 and 2,840 Feet in Elevation

There are no shots that are confined to the middle cylinder only. However, there were three rays that passed through both the middle and bottom cylinders. Since the structure of the bottom cylinder is known, the travel times of these rays can be thought of as functions of the structure of the middle cylinder.

Having only three rays to work with we modeled the cylinder as a simple two shell structure. It contains the same rock types as the cylinder above it, principally dome material, two short lava flows and a mantle of pyroclastic debris. Analogous to the upper cylinders we specified an inner and outer velocity of 2.25 and 1.2 km/sec, respectively, and varied the radius of the inner shell to minimize the difference between the calculated and observed travel times. Since only one parameter

was varied the same least squares inversion algorithm as for the upper cylinder was used. From an initial guess of 0.75 km for the radius of the inner layers, the algorithm converged to 0.56 km in three iterations. The root mean square difference between the calculated and observed travel times is 0.05 sec (Table AS2 middle).

Figure AS7 summarizes the inner cylindrical velocity structure of the volcano above the 1,000 foot contour line.

The travel times of the 11 inner rays above the 1,000 foot contour line are consistent (Table AS2) with a rather simple model of increasing velocity toward the center of the volcano. The outer skin of rubble, dome breccias and pyroclastic material has an average velocity of 1.2 km/sec. The more solid inner core of the lava domes has a velocity of 2.56 km/sec. The more fractured outer skin of the domes has a velocity of 2.25 km/sec.

The rather elaborate method of analysis was designed to detect a low velocity core within the 1964 lava dome, but the results suggest the reverse, high velocity material in the central conduit system. In the sub-section on core rays we will present evidence for a very high velocity pipe-like central conduit system of the volcano that existed prior to the 1976 eruption.

Mixed Rays

The subset of rays that passed from the inner to the outer parts of the volcano or vice versa crossing the 1,000 ft. contour line and with either station or receiver below 2840 ft are henceforth called mixed rays (Figure AS3c and Table AS4). We used this group of rays to extend the inner velocity structure of the volcano from 1,000 ft. down to the 5.1 km/sec refractor and also to check the velocity of layer 2 on the south side of the island.

Figure AS7: Cylindrical velocity structure of Augustine Volcano above 1,000 ft. contour, derived by ray tracing (inner rays), showing upward and outward velocity decrease.

In the previous section we showed that the upper (above 1,000 ft.) part of the volcano is composed of dome material with core velocities of 2.56 km/sec, decreasing upward and outward to values of 2.25 km/sec with the outermost skin as low as 1.2 km/sec. In order to avoid having to trace rays from the cylindrical upper structure of the volcano to the horizontally layered lower structure, we approximated the upper part of the volcano as a layered structure consisting of a 2.6 km/sec layer from 1,000 to 2,000 ft. elevation and a 2.3 km/sec layer from 2,000 ft. to the summit. Table AS5 gives elevation corrections to sea level in seconds for all shots and receivers, based on this generalized velocity structure.

Assuming that the 2.6 km/sec core extends to sea level we calculated the travel times for those rays that were critically refracted by the 5.1 km/sec layer. Most of calculated travel times were too long, suggesting that the velocity of layer 2, defined by the outer rays to be 2.6 km/sec, was too low. We had already indications from the analysis of the outer rays that on the south side of the island the velocity of layer 2 was higher. We therefore divided layer 2 into two parts. The northern half was set equal to 2.6 km/sec and the velocity of the southern half of layer 2 was varied until the mean residual between the observed and calculated times equaled 0.00. The final set of 12 mixed rays which were critically refracted by the 5.1 km/sec layer are listed in Table AS4: Adjusting the group mean to zero results in a standard deviation of 0.05 sec (Table AS4) for the 12 rays and yields a velocity of 3.4 km/sec for the southern half of layer 2 between sea level and 0.9 km depth.

When this same model for a split layer 2 was applied to the 13 outer rays (Table AS1) which were critically refracted by the 5.1 km/sec

TABLE AS4. Mixed Rays

Critically Refracted By 5.1 km/sec Layer

	T_o	D_h	$T_o - T_c$
4-AU2	1.59	4.00	-.03
4-AU5	1.59	4.50	-.09
7-2	1.69	5.04	-.03
7-5	1.90	6.02	+.07
7-6	2.10	6.96	+.08
7-8	2.22	7.69	+.04
8-5	1.75	5.61	-.01
8-6	1.83	6.11	-.02
9-5	2.18	7.54	+.05
9-6	2.01	7.16	-.05
9-8	1.88	5.94	+.04
8-2	1.39	3.62	-.05

Mean = $0.00 \pm .05$

Critically Refracted By 3.4

4-7	1.35	3.90	-.04
6-8	1.47	3.79	-.09

Critically Refracted By 3.4 and 2.6

3-AU2	1.33	2.91	-.04
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Flank Rays

6-6	1.11	2.26	
3-AU5	1.23	2.08	

layer, the mean residual equalled 0.00 with a standard deviation of .04 sec. The combined mean for the outer and mixed critically refracted rays is the same, 0.00 sec $\pm$.04. Figure AS8 is a replot of Figure AS5 correcting the outer ray travel times for a split velocity structure of layer 2, 2.6 km/sec for the northern half of Augustine Island and 3.4 km/sec for the southern half. The scatter is now much reduced.

There are 2 additional mixed rays which were probably refracted by the 3.4 km/sec layer, rays 4-7 and 6-8. Both rays were too fast by -.04 and -.09 sec respectively. The rays may have had part of their travel path through a high-velocity dome-like intrusion on the southern flank of Augustine, the Kamishak vent plug (Detterman, 1973).

Core Rays

The subset of rays which passed from the inner to the outer parts of the volcano or vice versa crossing the 1,000 ft. contour and with either the shot or receiver above 2,840 ft. are designated core rays (Table AS6). For ease of calculation we will only quantitatively consider those rays which were critically refracted by the 3.4 km/sec layer or the 5.1 km/sec layer. There are 5 rays: 10-AU3, 10-7, 1-AU3, 7-3, 7-11. These rays have a group mean of -.1 sec $\pm$.03 for the model developed thus far which differs by 2.5 standard deviations from the group mean of the combined outer and mixed rays. By using the F and t statistical tests we find that we can reject that the two populations means are the same at the 99% confidence level.

We are restricted by the small data base and the method of analysis to considering simple models to account for this travel time anomaly. The 1976 eruption revealed columnar jointing in the walls of the crater

Figure AS8: Outer rays (excluding rays from shot 5), elevation corrected and adding a correction for a higher (3.4 instead of 2.6 km/sec) velocity of layer 1 on the southern half of Augustine Island (replot of Figure AS5).

TABLE AS5. Shot and Station Elevation Corrections to Sea Level in Seconds for Generalized Volcano Structure as Follows: 0-2000 ft.-2.6 km/sec, and 2000 to 4304 ft.-2.3 km/sec, with 1.2 km/sec Pyroclastic Apron and 2.1 km/sec Sediments on the South Side

Shot	Elevation (km)	Refracted 5.1 (sec)	Refracted 3.4 (sec)	Refracted 2.6 (sec)
1-2	.869	.30	.24	
3	.610	.20	.15	
4	.528	.18	.13	
5	.004	.00	.00	
6	.172	.14	.13	
7	.002	.00	.00	
8	.003	.00	.00	
9	.006	.00	.00	
10	1.12	.40	.33	
Station				
AU1	.494	.16	.13	
AU2	.195	.16	.15	.14
AU3	.293	.13	.11	
AU4	1.021	.36	.31	
AU5	.152	.12	.12	.11
2	.678	.24	.18	
3	1.21	.44	.36	
4	.229	.08	.06	
5	.497	.16	.13	
6,1	.497	.16	.13	
7' (shot 4-10)	.146	.12	.11	.11
8	.528	.18	.14	
9	.913	.32	.26	
AU3=10	.610	.13	.11	
11	1.03	.37	.30	
AU2=12	.195	.16	.14	

TABLE AS6. Core Rays

Critically Refracted By 5.1 km/sec Layer and Core

	T_o	D_h	$T_o - T_c$
7-3	1.94	5.84	+.02
7-11	1.90	5.75	+.05

Critically Refracted By 3.4 km/sec Layer and High Velocity Core

10-AU3	1.31	3.26	-.01
10-7	1.40	3.58	-.01
1-AU3	1.25	3.30	-.03

Mean = $0.00 \pm .03$ Critically Refracted By 3.4 km/sec and 2.6 km/sec Layers
(High Velocity Core Not Considered)

			$T_o - T_c$ (3.4 km/sec)	$T_o - T_c$ (2.6 km/sec)
1-AU2	1.46	3.58	+.02	-.30
1-AU5	1.40	2.60	+.29	-.04
10-AU2	1.35	3.26	-.09	-.39
10-AU5	1.47	3.00	+.14	-.17

Figure AS9: Seismograms of selected outer rays, numbers indicate shot points and stations. Vertical lines indicate calculated arrival times for split layer 2 structure model. Left side shows anomalously early arrivals for shot 5.

implying that a small high velocity pipe may have existed prior to the eruption. If this pipe is assumed to be restricted to lie between the summit and sea level and if in addition we assume that the rays travel vertically along the pipe, adjustment of the residual of the 5 core rays to $0.00 \text{ sec} \pm 0.03$ (Table AS6) yields a pipe velocity of $4.4 \text{ km/sec} \pm 0.6$. The thermal studies discussed in the previous chapter indicate that the inner parts of the pipe filling could potentially still be hot, i.e., in reality the 2.56 km/sec core derived in the sub-section on inner rays could consist of a high velocity (4.4 km/sec) outer wall that surrounds a low velocity core. Unfortunately, there were not enough rays to distinguish between this more complex structure and the average structure presented in the sub-section on inner rays.

South Side Rays

Rays radiating from shot 5 are called south side rays (Table AS7). The mean residual for these rays calculated on the basis of the model developed thus far, including the vertical pipe, is $-0.14 \text{ sec} \pm .04$. Figure AS9 demonstrates this relative early arrival of signals from shot 5. We know that shot 5 lies in a region of uplifted sediments (Buffler, 1976). If we attribute the velocity anomaly to higher velocity sediment between sea level and the 5.1 km/sec refractor and again adjust the mean residual of this group of 4 rays to $0.00 \text{ sec} \pm .04$ (Table AS7), the velocity of the sediments would be 4.85 km/sec .

Summary

A generalized three-dimensional velocity structure of Augustine Volcano is shown in Figure AS10. The little more than 1,000 m high cone

TABLE AS7. South Side Rays
Critically Refracted By 5.1 km/sec Layer

	T_o	D_h	$T_o - T_c$
5-AU5	1.75	6.71	-.03
5-5	1.23	4.07	+.02
5-7	1.39	5.36	-.04
5-11	1.30	4.06	+.05
Mean = $0.00 \pm .04$			

Refracted By 3.4 km/sec Layer

5-8	0.78	2.26	-.02
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Figure AS10: Generalized velocity structure model of Augustine Volcano. Top: Sea level plan view. Left: E-W cross section. Right: N-S cross section.

above sea level was modeled as a cylindrical structure and consists of a central complex of andesitic-dacitic lava dome extrusions with an average core velocity of 2.6 km/sec. The velocity decreases outward and upward to 2.3 km/sec, probably reflecting increased fracturing. The central igneous structure is mantled on the lower slopes by very low velocity (1.2 km/sec) pyroclastics, consisting of debris avalanche, pumice flow and mudflow material, except for the southern flank, where uplifted sediments abut against the igneous complex. Even though we cannot constrain this average structure further with the limited number of rays that traversed it, there are several other lines of evidence that support a more complex innermost structure. For example, a still hot, low velocity, non-magnetic central feeder of the 1964 lava dome surrounded by a solidified, magnetized, high velocity pipe is not at variance with the thermal data discussed in the preceeding chapter and the magnetic data presented in the following chapter. In fact, the evidence for a high velocity pipe based on the core ray data almost requires a compensating low velocity core in order to not create a conflict with the inner ray data. For example, we could not distinguish between a single cylinder structure of 2.7 km diameter with an average velocity of 2.56 km/sec as shown in Figure AS7 and a more complex structure of the same outer diameter but containing an inner low velocity core surrounded by a high velocity outer cylindrical shell. We feel that such a model is entirely possible as the 1976 dome intrusion, which we suspect to be similar to the 1964 dome intrusion, was viscous with a temperature near or in excess of 700°C (Johnston, personal communication).

The basement of Augustine was modeled as a layered structure. At a depth of about 900 m beneath the volcano lies a rather high velocity layer of 5.1 km/sec. A layer with a velocity of 4.9 km/sec has been traced from a well, located 45 km northeast of Augustine Volcano, to the Island by the oil industry using seismic profiler systems. The layer smoothly deepens from Augustine to the well, where it was encountered at a depth of 6850 ft. (2.1 km) (Figure AS11). Wills et al. (1978) report that the high velocity is due to zeolitization of Lower Cretaceous sediments, filling the pore spaces and destroying most porosity and permeability originally in the rock.

The intermediate layer between the 900 m deep zeolitized zone and sea level is laterally heterogeneous, in general increasing in velocity from north to south across the island (Figure AS10 top). The zeolitized rocks have been uplifted to very near sea level on the southern flank perhaps by the igneous activity of Augustine Volcano. The northern part of the volcano is underlain by 2.6 km/sec material which we tentatively interpret as not-zeolitized sediments. Beneath the central part of the volcano we find a transitional velocity of 3.4 km/sec, which could be an average value for non-zeolitized sediments intercalated with volcanic sills for which there is magnetic evidence as discussed in the next chapter.

At the levels we probed seismically we did not locate any large very low velocity bodies indicative of melt or partially molten rock.

Figure AS11: Velocity derived at 10 millisecond intervals from the integrated, borehole compensated sonic log for the Lower Cook Inlet COST well No. 1 (from Wills et al., 1978).

3. PASSIVE SEISMOLOGY

Objectives

The Augustine Island based short period seismic network with its limited aperture of about 10 km, supplemented by a few stations in Lower Cook Inlet (RED, CDA, SHU, SLV, Figure AS1) did, up to 1975, not allow accurate hypocenter locations of deeper events beneath the volcano. In order to study the deeper seismicity in Lower Cook Inlet we expanded the Augustine array to span the Inlet. The main objectives of this new array were: (1) to locate the Benioff zone beneath the Lower Cook Inlet volcanoes, and (2) to study the seismicity of Augustine's conduit system.

Expanded Array

Three new short-period seismic stations were added between September 15 and 22, 1975, to the existing rudimentary network in Lower Cook Inlet. Two of these, Oil Point (OPT), 35 km NNE of Augustine, and Bruin Bay, 32 km W of the volcano, were sponsored by this contract. The array was slightly modified in summer 1976, when we added a station in Homer (HOM), and in mid-1977, when we moved the Bruin Bay station to the summit of Augustine Volcano (AUH). Figure AS1 shows the configuration of the current array. All the University of Alaska data is radio-telemetered to Homer on the east side of the Inlet.

Results

1. Regional Seismicity

We are now continuously monitoring the seismicity of Lower Cook Inlet. At present, this program is entirely supported through NOAA's Outer

Figure PS1: Current configuration of short-period seismic network in Cook Inlet, on Kodiak and on the Alaska Peninsula.

Continental Shelf Assessment Program (OCSEAP). Figures PS2 and PS3, taken from Pulpan and Kienle (1979), are current epicenter and hypocenter plots of the seismicity in Cook Inlet. Details on the routine location procedures, crustal structure, etc., have been discussed by Estes (1978) and in the quarterly series of seismological bulletins issued by the Alaska Earthquake Analysis Center since November, 1977.

Figure PS2 demonstrates the unusually high level of seismicity of the Lower Cook Inlet as compared to the adjacent sections of the Aleutian island arc structure. In this area we are now routinely locating an average of about 150 events per month. Most of the seismicity is located in the deep Benioff zone. The Benioff zone is particularly active beneath Iliamna volcano but smaller deep clusters occur beneath Augustine and Douglas volcanoes (Figure PS2).

Figure PS3 shows best located Lower Cook Inlet hypocenters in a 170 km wide arc section projected onto a vertical plane perpendicular to the volcano line (Pulpan and Kienle, 1979). The Benioff zone is quite thick, about 30 km, which is not an artifact of the projection direction (e.g., Estes 1978). It extends to a depth of about 200 km and plunges more steeply below 100 km. This knee-like bend beneath the volcano line has been seen elsewhere in the Aleutian Arc, e.g. in the central Aleutians (Engdahl, 1976) and in the Shumagin Islands (Jacob et al., 1977). The cross section also shows a cluster of shallow (less than 10 km) seismicity beneath Augustine Volcano but no seismicity at greater depths where magma is presumably generated and transported upwards in near vertical feeder dyke systems. This aseismic

Figure PS2: Epicenter map for Cook Inlet, best class (RMS < 1 sec, horizontal and vertical location error < 10 km) events, February 1976-September 1978, showing the dominating deep seismic cluster near Iliamna Volcano and less intense deep clusters near Augustine Volcano (153.4°W, 59.4°N) and Mt. Douglas (153.6°W, 58.9°N). The letter code indicates the depths of the events: A 0-25 km, B 26-50 km, etc. The event is located at the lower left hand corner of the letter. Letter size is proportional to magnitude; letter size of geographic coordinates corresponds to magnitude 2. Dots indicate events for which the magnitude is not known.

Figure PS3: Benioff Zone in Lower Cook Inlet, best class events, February 1976-September 1978. Projection plane perpendicular to volcano line, width of arc segment plotted is 170 km.

wedge beneath the volcanic arc is typical of island arc seismicity and has been observed elsewhere in the Aleutians by Engdahl (1977). In the wedge the presence of hot melt and partial melt may account for the non-brittle behavior of the lithosphere.

2. Seismicity of Augustine Volcano

As it turned out, the most important contribution of the active seismic experiment -- since we did not locate any substantial region of molten rock -- was a detailed three-dimensional velocity structure of the volcano (Figure AS10): As we discussed in the previous chapter, the generalized final structure to a depth of 1 km below sea level fits the bulk of the explosion travel time data with a standard deviation of about 0.05 sec. The velocity structure below 1 km was inferred from a well drilled 45 km northeast of the volcano (Wills et al., 1978) and crustal structures developed for the area and consists of a 5.1 km/sec layer from 1 to 6 km underlain by a 5.4 km/sec layer to a depth of 11.6 km.

Using this model Lalla and Kienle (1978) analysed the precursor seismicity to the January 1976 eruptions. The primary data used in this analysis were from the five stations of the Island based array. It operated reliably until January 2, 1976, about three weeks before the onset of the eruptions. The failure of the Island network just prior to the eruptions was indeed an unlucky break. Since all but one station were deeply buried by the ejecta of the eruptions, we could never determine whether the failure was due to preliminary volcanic activity or due to winter storms, severe antenna icing and electronic failure at cold temperatures. Most likely, both factors brought the system down; for

example, when we examined the sole surviving station on the south side of the Island, we found the electronics intact, but the antenna severely damaged by falling debris and the co-ax cable ripped out of its socket, apparently by a strong blast or gust of wind. On the other hand, we also experienced station failures of the Kamishak Bay stations which were typical winter storm problems. One unusual failure occurred at the Bruin Bay station. When we visited it after the initial vent clearing eruptions of late January 1976 the Augustine facing hill side where the station was located was thickly glazed with ice from the water vapor discharged by Augustine plumes which were carried predominantly towards Bruin Bay by the prevailing winds. The antenna mast had broken under the weight of the iced-up antenna.

In spite of these problems we were able to continue to monitor the volcano throughout the eruptions with the surviving Kamishak Bay and the U.S. Geological Survey stations further inland to the west and southwest, albeit with a higher detection threshold magnitude. The detection threshold for Augustine events with the Island based network is about $M_L = 0$, while the larger aperture (60 km) Kamishak Bay array allowed us only to locate Augustine events down to magnitude 1.5.

Knowing the velocity structure of the volcano fairly well allowed us to resolve the depths of the precursor seismic events to less than about 1 km. The precursor seismicity began in May 1975 and is clearly restricted to depths less than 8 km. Figure PS4 from Lalla (1979) is a summary plot of the January through December 1975 precursor seismicity to the 1976 eruptions. Most events occur within an apparent central conduit system of the volcano which seems to extend to about

Figure PS4: Epicentral (top), E-W (bottom left) and N-S (bottom right) cross sectional views of Augustine 1976 eruption precursor seismicity, January-December 1975. Letter code indicates depths: A 0-1 km, B 1-2 km, etc. In the bottom 2 plots the symbols denote different time periods: squares-January to April, triangles-May, circles-June/July, diamonds-August, crossed out circle-September, star-October, pluses-November, x's-December. Only best events are used: RMS < 0.15 sec, horizontal error < 2 km, vertical error < 1.5 km, at least 5 arrivals. Note, intense clustering of earthquakes above 1 km depth (from Lalla, 1979).

8 km. These events are now being interpreted in terms of a temporal and spatial migration of seismic energy release probably reflecting actively changing magmatic and hydrothermal processes (Lalla, 1979).

To date, we have not observed any eruption related precursor seismicity from depths greater than 8 km suggesting that the rhyolitic to andesitic magma erupted in 1976 comes from a shallow magma chamber located less than 8 km beneath Augustine Volcano. Independent confirmation of this conjecture comes from geochemical data. Johnston (1978) presented a convincing petrochemical model, based on the geochemistry of the ejecta and gas inclusions, suggesting that several months prior to the eruptions basaltic melt was injected into or mixed with rhyolitic melt contained in a shallow magma chamber at 2 to 10 km depth. The mixing could have coincided with the onset of the precursor seismicity in May 1975.

4. THREE-DIMENSIONAL MAGNETIC MODELING

Objectives

A non-magnetic body, such as molten or hot rock above the Curie temperature, is expected to give rise to a negative magnetic anomaly, if it is in contrast with magnetized country rock. The objectives of the three-dimensional modeling of aeromagnetic surveys flown over Augustine Volcano were to determine the magnetic structure of the volcano and, if possible, to locate the Curie isotherm, the position of which could be indicative of regions of hot or molten rock.

Data

1. Aeromagnetic Data

The primary aeromagnetic survey used in this study was flown by Lockwood, Kessler and Bartlett (LKB), Inc., Airborne Geophysics Division of Pasadena, California, in August of 1972, prior to this contract period. The LKB survey consisted of 17 east-west flight lines, approximately 11 miles long, centered over Augustine Island, and flown approximately 1 km apart, extending from just north of Burr Point to approximately 1.5 km south of the Island. These lines and their associated tie lines were all flown at an altitude of 4,000 feet above sea level. In addition, three lines were flown, 500 feet above mean terrain level, around the 100-foot, 500-foot and 1500-foot contour lines.

The instrument used for the LKB survey was a fluxgate magnetometer measuring only the vertical component of the magnetic field. Therefore, the absolute magnetic field value is not known. Figure M1 shows the results of the survey. A regional magnetic gradient of 6.8 gammas per

Figure M1: Aeromagnetic contour map of the 1972 Lockwood, Kessler and Bartlett survey. Magnetic field values are in gammas and are relative to some arbitrary baseline. The flight lines are indicated by the dots P1-P17. Hatchured contours indicate magnetic lows. Survey flown at 4000 ft ASL.

mile (4.2 gammas/km) in the direction of N 35° E has been removed. The main features are the strong magnetic high just east of the summit and an opposing, very weak low to the northeast of the summit. At the location of Augustine Island (153° 25' W, 59° 22' N) the earth's field has a total intensity of approximately 54,500 gammas, with an inclination of 72.6° and a declination of N 23° E. The gradient from high to low in Figure M1 is fairly closely aligned with the direction of the declination of the earth's field, N 23° E, indicating that the volcano is magnetized in the direction of the earth's present field.

If the background field is arbitrarily taken to be 1980 gammas, then the value of the magnetic high is 2751 gammas and the low is 31 gammas. The field resembles that of a monopole. A monopole field is observed when the measurements are made much closer to one end of a dipole than to the other. In the case of Augustine, the upper end of the dipole lies closest to the magnetometer, within 100 feet at the closest approach, while the other end of the dipole probably lies near the base of the volcano, 4,000 feet lower.

Typically, aeromagnetic surveys give field values relative to some arbitrary datum, not absolute values, but fortunately, the most important parameter indicative of depth and shape of a magnetized body is the vertical gradient. For example, Muroi (1973) carried out multilevel aeromagnetic surveys of several volcanoes and calderas in Japan pointing out the value of measuring the vertical gradient to determine subsurface structure and magma reservoirs. We, therefore, and also to test the temporal stability of the magnetic features seen in the LKB survey, flew additional surveys in 1975 under this contract. On 28 August 1975, Alaska Aeromagnetics (AA) of Fairbanks, Alaska, with participation from

J. Kienle, flew a series of aeromagnetic surveys of Augustine at 2,000, 4,000, and 5,000 foot altitudes. We used a proton precession total field magnetometer which was attached to the wing of a Cessna 180. A video tape camera was used for ground control.

The 2,000 foot survey consisted of seven flight lines, three in a north-south direction and four in an east-west direction. Figure M2 shows the flight lines and the resultant magnetic field with the background field of 54,000 and the regional magnetic gradient removed. Since the survey was flown at 2,000 feet summit features were not measured. The map shows a large number of relatively small scale high and low anomalies on the flanks of the volcano.

The 4,000 foot survey repeated the LKB survey and consisted of six east-west trending flight lines. Figure M3 shows the flight lines and the resultant measured magnetic field with the background field and regional magnetic gradient removed. None of the flight lines passed over the summit and we therefore have no maximum value of the positive anomaly at the 4,000 foot altitude. The flight lines were rather widely spread to determine the details of the magnetic field. Nevertheless, one can clearly see that the shape of the field is quite similar to that of the 1972 LKB survey.

The 5,000 foot survey consisted of three east-west flight lines and one north-south flight line. Figure M4 shows the flight lines and the resultant magnetic field with the background field and regional gradient removed. One east-west flight line and the north-south flight line passed close enough to the summit that the maximum field value at the 5,000 feet altitude is known. The maximum field anomaly was 2001 gammas, 1381 gammas above the lowest value of 620 gammas. The steepest gradient

Figure M2: Aeromagnetic contour map of the 1975 2000-foot survey flown by Alaska Aeromagnetics. Flight lines are shown as dashed lines. The magnetic field values are in gammas. The earth's total field of 54,000 gammas and the regional gradient of 6.8 gammas/mile at N35°E have been removed.

Figure M3: Aeromagnetic contour map of the 1975 4000-foot survey flown by Alaska Aeromagnetics. Flight lines are shown as dashed lines. The magnetic field values are in gammas. The earth's total field of 54,000 gammas and the regional gradient of 6.8 gammas/mile at N35°E have been removed.

Figure M4: Aeromagnetic contour map of the 1975 5000-foot survey flown by Alaska Aeromagnetics. Flight lines are shown as dashed lines. The magnetic field values are in gammas. The earth's total field of 54,000 gammas and the regional gradient of 6.8 gammas/mile at N35°E have been removed.

appears to be in the same direction as that seen in the LKB 1972 survey, approximately N 23° E.

A third survey was flown over Augustine following the vent clearing eruptions of January, 1976, with emergency funds from the State of Alaska. On 1 February 1976, following the vent-clearing eruptions of January 22-25, Alaska Aeromagnetics resurveyed the volcano using the same instrument as in 1975. A total of sixteen flight lines were acquired at an altitude of 4,000 feet, 14 of which were flown in a northwest-southeast direction and two crossings of these in a northeast-southwest direction. Several of the lines passed very near or over the summit, which at that time consisted of a steep-walled crater, about 250 m deep. Figure M5 shows the resultant magnetic field with the background field and regional gradient removed. The shape of the field is very similar to that of the LKB survey and the 1975 AA survey. The gradient between the strong high anomaly and the weak low anomaly to the northeast of the high trends approximately N 30° E. The field difference between these two points is 1404 gammas, with the strong high and weak low anomalies again indicative of a monopole field. The field shows two characteristic departures from the 1972 survey. First, there is a low directly north of the large high caused by the new 250 m deep crater formed during the January 1976 eruptions. Second, a small high of unknown origin appears to the northwest of the peak.

2. Rock Magnetizations and Susceptibilities

In order to calculate the magnetic field of Augustine Volcano, the magnetizations and susceptibilities of the material composing the volcano need to be known. Three sets of rock samples were available to determine

Figure M5: Aeromagnetic contour map of the 1976 4000-foot survey flown by Alaska Aeromagnetics. Flight lines are not shown. The magnetic field values are in gammas. The earth's field of 54,000 gammas and the regional gradient of 6.8 gammas/mile at N35°E have been removed.

their remanent magnetizations and susceptibilities. During the summer of 1972, R. Metzner of the Geophysical Institute of the University of Alaska collected a number of unoriented samples (R1-R10, Table M1) and D. Stone, also of the Geophysical Institute, collected oriented samples (C1-C21, Table M2). During the summer of 1975, 27 oriented hand samples (SB1-SB27, Table M3) were collected by S. Barrett (this contract). Locations for samples C1-C21 and SB1-SB27 are shown in Figure M6. Accurate locations of samples R1-R10 were not recorded.

Samples R1-R10. Since these samples, collected in July 1972 and analyzed in October 1972, were not oriented, only the intensities and not the directions of magnetization were measurable. Table M1 gives the values as measured on a Schoensted Model SSM-1A spinner magnetometer. All of the values are natural remanent magnetizations and no susceptibilities were measured for this group.

Table M1: Samples R1-R10

Sample No.	Total Intensity of Magnetization $\times 10^{-6}$ emu/cc	Sample No.	Total Intensity of Magnetization $\times 10^{-6}$ emu/cc
R1	2855	R6	4172
R2	4022	R7	3636
R3	2428	R8	1651
R4	2276	R9	1628
R5	1425	R10	4116

Table M2: Samples C1-C21

Sample Number	Remanence 10^{-6} emu/cc	Susceptibility 10^{-6} emu/cc	Geologic Feature and/or Location
C1	2253	2500	boulders in mudflow
C2	1784	3100	boulders in mudflow
C3	1833	2800	boulders in mudflow
C4	2101	2200	boulders in mudflow
C5	11950	1200	boulders in mudflow
C6	6907	2900	rock and breccia
C7	2761	124	rock and breccia
C8	4122	144	rock and breccia
C9	24940	1050	lava flow front
C10	15080	640	lava flow front
C11	15130	770	lava flow front
C12	13930	785	lava flow front
C13	15540	850	lava flow front
C14a/b*	0.918/0.669	9.0/9.0	southside sediments
C15a/b*	5.040/8.558	71/84	southside sediments
C17a/b	12220/4506	1950/2350	summit volcanics
C18a/b	17170/14630	2000/2100	summit volcanics
C19a/b	7843/6876	1500/1800	summit volcanics
C20a/b	2532/2439	2650/3150	summit volcanics
C21a/b	1825/----	2500/2550	lava flow top

* a and b indicate the top and bottom cores, respectively, of the same sample.

Table M3: Samples SB1-SB27

Sample Number	Remanence 10^{-6} emu/cc	R_{ave} 10^{-6} emu/cc	Susceptibility 10^{-6} emu/cc	X_{ave} 10^{-6} emu/cc	Koenigsberger Ratio R_{ave}/X_{ave}	Location and/or Geologic Setting
SB1a/b*	2551/255	-----	1700/1700	-----	-----	Burr Point Rubble
SB2a/b	8132/9058	8595	1570/1420	1495	5.75	Kamishak Dome
SB3a/b	-----	-----	1220/600	-----	-----	Lagoon Beach Rubble
SB4	1865	1865	2550	2550	0.731	Mud flow, N.W. Lagoon
SB5a/b	3423/3442	3432	2050/2100	2075	1.65	Mud flow, N.W. Lagoon
SB6a/b	2644/2929	2786	2200/2300	2250	1.24	Mud flow, N.W. Lagoon
SB7a/b	555/101	328	1160/1300	1230	0.267	Mud flow, N.W. Lagoon
SB8	13710	13710	2200	2200	6.23	Dome 64, summit spine
SB9a/b	4352/4996	4674	330/500	415	11.3	Dome 35
SB10a/b	7962/7408	7685	940/920	930	8.26	Dome 35
SB11	-----	-----	520	520	-----	Dome 35(?)
SB12a/b	2492/2206	2349	1120/1000	1060	2.22	Dome I, prehistoric, high west flank
SB13a/b	1671/1988	1829	820/920	870	2.10	Dome I
SB14a/b	11850/11650	11750	1880/1860	1870	6.28	Dome H, prehistoric, middle west flank
SB15a/b	5310/3554	4432	2400/2250	2325	1.91	Dome H
SB16a/b	4969/5876	5424	2200/2500	2350	2.31	Dome H
SB17a/b	17220/20880	19050	820/1020	920	20.7	Pinnacles
SB18a/b	4785/4337	4561	1660/1420	1540	2.96	Pinnacles
SB19a/b	870/761	815	118/142	130	6.27	Near Pinnacles
SB20	-----	-----	1220	1220	-----	Rubble near pinnacles
SB21a/b	17910/17830	17870	720/640	680	26.3	Pinnacles
SB22	-----	-----	1180	1180	-----	Rubble near pinnacles
SB23a/b	5437/5682	5804	2900/2950	2925	1.98	Spine below pinnacles
SB24a/b	3369/3659	3514	1510/1400	1455	2.42	Top of large lava flow
SB25a/b	3446/3369	3407	3100/3200	3150	1.08	Face, small lava flow
SB26a/b	18900/16940	17920	730/800	765	23.4	Center, large lava flow
SB27a/b	2154/2359	2256	470/420	445	5.07	North end, large flow

* a and b indicate two cores taken from the same hand sample.

Figure M6: Map of Augustine Island showing the location of rock samples C1-C21 (circles) and SB1-SB27 (triangles).

Samples C1-C21. Samples C1 through C21 were collected in July of 1972 using a gasoline driven coring device with a diamond cutting edge. The cores were marked and oriented in situ. They were analyzed for both remanent magnetizations and susceptibilities in April 1973.

Table M2 lists the magnetizations in units of 10^6 emu/cc, as well as the type of geologic feature they were obtained from, e.g., mudflow, dome, etc. There is some correlation between the remanence of each sample and the type of geologic feature from which the sample was obtained. Samples C1 through C8 were obtained from boulders and other breccia from the Burr Point pyroclastic flow. With the exception of sample C5, which has an anomalously high remanence, all of these samples have remanent magnetizations in the range of approximately 2 to 7×10^{-3} emu/cc. The samples obtained from the front of a lava flow on the north side of the island have an order of magnitude larger magnetization ranging from 1.3 to 2.5×10^{-2} emu/cc. These values are similar to those of two summit lave dome samples (C17 and C18), while the other two summit samples (C19 and C20), as well as sample C21 from the top of the lava flow, all have remanences similar to those of the boulders and breccia. The sediments (samples C14 and C15) have several order of magnitude lower remanences.

Sample C9 is anomalous and has by far the largest remanence, nearly twice that of the other samples in its group. Sample C5, a boulder sample, is also anomalous, having a remanence more than five times that of the other boulder samples. These two samples may have been struck by lightning, which can alter the magnetization directions and magnitudes significantly.

The north end of the measured inclinations and declinations of the remanent field of the samples are plotted on a Wolff stereonet in Figure M7. Samples C1 through C8, which were obtained from pyroclastic flow and mudflow debris and loose breccia, scatter widely, indicating that they had cooled below the Curie temperature when the flow came to rest. As will be seen later when we discuss the 1975 samples, the low magnetic viscosity of the samples is the apparent reason for the great scatter of the data points.

Table M2 also lists the susceptibilities of samples C1 through C21 as measured on a susceptibility bridge. The values (actually magnetization $M = XH$ where $H \approx 1$ emu) range from less than 1×10^{-5} emu/cc for the sediments, to more than 3×10^{-3} emu/cc for the summit dome material. Generally, the higher susceptibilities are associated with the lower remanences and vice versa. Some samples have susceptibilities that are greater than their respective remanences (e.g. C1-C4).

Figure M8 is a plot of susceptibility, k , versus remanent magnetization, M , for samples C1 through C21. If all of the samples have the same Koenigsberger ratio (k/M) then such a plot should give a straight line with a positive slope. (The Koenigsberger ratio is the ratio of natural remanence to induced magnetization and can be used as a measure of the stability of the magnetization of rock samples.) The slope is the value of the Koenigsberger ratio. With the exception of two boulder samples (C7 and C8), the values roughly fit a curve that shows susceptibility increasing as remanence decreases. This indicates that the samples may have a low magnetic viscosity, i.e., changes in the ambient field would cause the magnetizations of the samples to change. The samples were probably affected by many different magnetic fields between the time

Figure M7: Stereonet plot of the inclination and declination of samples C1-C21. Closed circles (open circles) indicate that the inclination of the magnetic field of the samples is down (up) with respect to the earth's surface.

Figure M8: Plot of measured remanence versus measured susceptibility for samples C1-C21.

they were collected and the time their magnetizations were measured. After collection, the samples would have been oriented differently with respect to each other during transport and storage and therefore each sample would have been affected differently. However, since the susceptibility is a bulk property of the rocks, it would not have altered. Also, those samples containing magnetically harder minerals would have a high spontaneous magnetization and be more difficult to alter external magnetic fields, whereas those containing magnetically softer minerals would have a lower remanent value but would be easier to realign in an external field (Stacey and Banerjee, 1974).

Samples SB1-SB27. These samples were obtained during the summer field season of 1975. They were collected as oriented hand specimen. The hand samples were reoriented later in the laboratory and cored.

This method is not nearly as accurate as the field-coring method used in obtaining samples C1-C21. The Brunton compass may not give accurate readings of declination because of the effects of nearby magnetic material, including the samples themselves. Also, hand samples are of necessity near-surface material, which may be affected by lightning and weathering. Thus, their remanences may be significantly altered. Finally, errors in orienting the samples while drilling the core in the laboratory may occur.

Table M3 gives the remanences and susceptibilities along with the locations of samples SB1 through SB27. The remanences ranged from 1×10^{-4} emu/cc to 2×10^{-2} emu/cc. Only three of the samples, SB7a/b and SB1b have values below 10^{-3} emu/cc. Samples SB7a/b were obtained from the same pyroclastic flow as samples SB4-SB6 and were of similar

composition, yet had a much lower remanence. Samples SB1a and SB1b were adjacent cores of the same sample collected from a large boulder of andesite of the Burr Point pyroclastic flow. They should therefore have nearly the same remanence values. Since sample SB1a has a remanence value similar to that of other debris material, sample SB1b is probably anomalous. These anomalies cannot be readily explained.

There does not appear to be significant correlation between the remanence values and their geologic location. For example, the remanences of the three samples obtained on a prehistoric dome on the lower W-flank of Augustine, SB14-SB16, vary by as much as a factor of three, from 3.6×10^{-3} to 11.7×10^{-3} emu/cc. Samples obtained from the pinnacles near the summit vary even more. Similarly, the samples from the lagoon mud flow on the northwest side of the island, excepting sample SB7, vary from 2.6×10^{-3} to 9×10^{-3} emu/cc.

The remanent field directions of the samples were plotted on different stereonetts according to sampling sites. Figure M9 shows clockwise the field directions of pyroclastic flow samples, lava flow samples, dome samples, and pinnacle samples. The present direction of the earth's magnetic field is also indicated on the stereonetts. The dome sample plot also has samples C17-C20 included for comparison.

As can be seen, none of samples SB1-SB27 have remanence directions close to the present field direction. The closest ones are two of the pinnacles samples and sample SB12. Generally, there is a wide scatter of points. For those cases where two cores were obtained from the same sample, the field directions of the two cores were similar, in some cases very close, for example, samples SB17, SB19, and SB21 of the

Stereonet plot of mudflow samples SB2, SB4, SB5, and SB6. Open circles-inclination up; closed circles-inclination down.

Stereonet plot of lava flow samples SB25 and SB27.

Stereonet plot of pinnacles samples SB17, SB18, SB19, SB21 and SB23. Open circles-inclination up; closed circles-inclination down.

Stereonet plot of dome samples SB8, SB9, SB10, SB12, SB15 and, for comparison, samples C18-C20. All of the inclinations point downwards.

Figure M9: Magnetic field directions of Augustine samples collected in 1975, grouped according to sampling site, clockwise: pyroclastic- and mud flows, lava flow, central lava domes, pinnacles.

pinnacles, and sample SB5 of the mud flow. This implies that the laboratory measurements were fairly accurate.

The randomness of the orientations is probably the result of some magnetic property inherent in the samples. In an attempt to determine the cause of the randomness, the samples were demagnetized and measured in a cryogenic magnetometer. It soon became apparent (D. B. Stone, personal communication) that the randomness was due to a large component of viscous magnetization. The samples would acquire a new magnetization from stray magnetic fields immediately after being demagnetized. Therefore, we conclude that the direction of the remanent magnetization of the rocks forming Augustine Island is likely to be parallel to the earth's present field.

The viscous magnetization effect is further evidenced in Figure M10, which is a plot of remanent magnetization versus susceptibility for samples SB1-SB27, excepting the samples for which no remanences were calculated and sample SB1, which has an order of magnitude intensity change between adjacent cores a and b. Both the remanent magnetizations and the susceptibilities show a wide scatter, with values varying by nearly an order of magnitude. Within this scatter of points, individual groups of data points form straight lines on a log-log plot. The data points for the lagoon pyroclastic flow (SB2-SB7) form a near-perfect straight line, with the exception of point SB7, which is anomalously low in remanence. However, the slope of the line is negative, which is itself anomalous. If all of the samples were similar, then the line should have a positive slope, because the Koenigsberger ratio should be the same. A large component of viscous magnetization seems to be the only explanation for these anomalous results.

Figure M10: Plot of measured remanence versus measured susceptibility for samples SB1-SB27.

Summarizing the results of the three sets of rock magnetization data, it can be concluded that (1) a large component of viscous magnetization is evident which has altered the original direction and intensities of magnetization of the samples in situ, (2) the values of the natural remanences are probably in the range of 2×10^{-3} to 1×10^{-2} emu/cc in the direction of the earth's magnetic field, and (3) the susceptibility values are in the range of 1×10^{-4} to 3×10^{-3} emu/cc. The remanences and susceptibilities are thus directly additive, both being in the earth's field direction. The remanences are dominant for in situ rock masses, but the susceptibilities dominate the magnetic field values for the pyroclastics.

Since a large part of Augustine Island may be composed of unconsolidated materials, the susceptibility of a sample of lightly packed dry ash and pyroclastics was measured giving a value of 4.4×10^{-5} emu/cc. This places a lower limit on the susceptibility value for Augustine Island as a whole.

Data Analysis

The 1972 LKB survey was the primary survey used in the analysis since the flight lines were more closely spaced giving a higher resolution of the anomalies than the pre-eruption 1975 AA surveys. The AA surveys, however, complemented the data, giving us approximate absolute values of the anomalies and vertical gradients. The 1976 survey presented the opportunity of comparing a post-eruption magnetic field with a pre-eruption field to try to determine if thermal changes of the volcano and hence a change in the Curie isotherm could be observed.

Barrett (1978) developed a computer program using the method of Talwani (1965) to three-dimensionally model Augustine Volcano. The program was written in FORTRAN IV language and is split into two parts to simplify calculations. The first part calculates the volume integrals, and the second part uses these volume integrals to calculate the magnetic field. The detailed theory and computer programs are described by Barrett (1978).

Modeling the magnetic structure of Augustine involved matching the calculated magnetic field of some postulated structure to the known field from aeromagnetic data. The datum plane was 4000 feet ASL. This was done by changing the geometry of the magnetic components, which involved recalculating the volume integrals, and by changing the magnetization of the rocks until the calculated field matched the observed. The model was constrained, as discussed by Barrett (1978), with the known geology, measured magnetizations of rock samples and the results of the active seismic experiment (Pearson, 1977). By trial and error the best model was found after about 100 different models had been tried.

As a first step the volcano's topography was approximated by a stack of slabs, which consisted of straight line approximations around each 500 foot contour interval. The 4000 foot high volcano was thus approximated by eight 500-foot thick slabs. Other slabs representing for example known lava flows and domes and cores of various magnetizations were added, as required by the mismatch of the anomalies.

The following is a sampling of simple models discussed by Barrett (1978) that eventually led to the more complex final model.

Model 1 is a simple homogeneous cone with a uniform magnetization J . Figure M11 shows the calculated field for a model consisting of slabs 1-8A (Table M4) and having a uniformly low magnetization of $J = 6 \times 10^{-4}$ emu/cc with an inclination of 72.6° in a direction $N 23^\circ E$, i.e., the direction of the present geomagnetic field. All of the following models assume the same orientation of the magnetizations. The value of 6×10^{-4} emu/cc is within the range of susceptibilities (1×10^{-4} emu/cc to 3×10^{-3} emu/cc) measured for the loose debris. Since this model assumes that the volcano is composed entirely of unconsolidated pyroclastics, there would be no remanent magnetization and the total magnetic field anomaly of the volcano would be due to susceptibility alone. Support for this idea is actually seen in the aeromagnetic survey, where the direction of steepest gradient from the positive high to the negative low of the anomaly is approximately $N 30^\circ E$; the declination of the earth's field is $N 23^\circ E$. The field of model 1 is strongly dipolar in the direction $N 25^\circ E$, which is approximately the same as the magnetic declination. The magnitudes of the highs and lows are amazingly close to the observed field, as are the field gradients. The peak of the high, however, lies southwest of that actually observed, and the shape of the field is too dipolar compared to the monopole character of the aeromagnetic survey (Figure M1). This implies that there must be other magnetically significant structures affecting the total field.

A major part of the modeling consisted of determining the position of the Curie isotherm, which would have indicated non-magnetized hot rock. For Model 2 (Figure M12), a large core of non-magnetized material (Table M4, slab 17) was tried along with a cap of highly magnetized rock (slabs 1, 2) and a sedimentary wedge (slabs 10, 11, 12) that has been

MODEL 1

$$J = 6 \times 10^{-4} \text{ emu/cc}$$

Figure M11: Calculated magnetic field of uniformly magnetized volcano (slabs 1-8A, basic cone).

Table M4 : Augustine slabs used for 3-D magnetic modeling (from Barrett, 1978, partial listing)

Slab No.	Description.	Distance in feet (from flight level at 4000 ft.)		Characteristic horizontal dimensions in meters
		Top	Bottom	
1	basic volcano slab	100	600	300
2	basic volcano slab	600	1100	800
3	basic volcano slab	1100	1600	1400
4	basic volcano slab	1600	2100	1800
5	basic volcano slab	2100	2600	2500
6	basic volcano slab	2600	3100	3500
7	basic volcano slab	3100	3600	5000
8	basic volcano slab	3600	4100	6500
8A	basic volcano slab	3600	4000	6500
9	basic volcano slab	4000	4340	11000
10	sediment block	3100	3600	1500 x 3000
11	sediment block	3600	4000	2000 x 4000
12	sediment block	4000	4400	3000 x 6000
16	small central core	600	4100	300
17	large core	1600	4100	1750
22	pinnacle top	1100	1600	250 x 400
26	sill	5050	7050	6500
30	lower pinnacle block	1600	3100	450 x 850
31	Kamishak plug	2600	4000	450
33	lava flow slabs	3250	3500	200 x 600
34	lava flow slabs	3000	3300	250 x 600
35	lava flow slabs	2750	3100	350 x 650
36	lava flow slabs	2500	2850	350 x 650
37	lava flow slabs	2200	2550	250 x 700
38	lava flow slabs	2000	2350	250 x 700
43	South Peak slab	700	2100	250 x 700

MODEL 2

$$J_0 = 0$$

$$J_1 = 2 \times 10^{-3}$$

$$J_2 = 1.5 \times 10^{-2}$$

Figure M12: Calculated magnetic field of basic cone (slabs 1-8A) plus large hot core (slab 17), a magnetized cap (slabs 1 and 2), and a sediment wedge (slabs 10, 11, 12).

documented by geologic mapping on the south flank of the volcano. This produced a modified dipole field with the calculated peak of the positive anomaly well southwest of that observed. The magnitude of the positive anomaly is comparable to the observed field. However, the magnitude of the calculated negative anomaly is much too high by nearly an order of magnitude, and the gradients are dissimilar. It appears from this that there is no large demagnetized region above sea level within Augustine Volcano.

Model 3 (Figure M13) shows a structure opposite to the hot core Model 2. A highly magnetized cold core (slab 17) with a very thin hot pipe (slab 16) down the center makes up the center of the volcano. The large solid core of highly magnetized rock increased the positive anomaly to almost 9000 gammas, more than a factor of three greater than that observed. Obviously, Augustine Volcano does not possess such a core.

Models 2 and 3 indicated that the central conduit, hot or cold, had to be fairly small. It also became apparent that the Pinnacles just NNE of the summit are magnetically important. The effect of including a magnetized cylindrical structure beneath the Pinnacles was to move the peak of the field northeasterly to its correct position east of the summit. Figure M14 (top) shows the calculated field for the stack of basic slabs with a cylindrical magnetic structure added for the Pinnacles and a hot small (less than 100 m radius) central core.

Magma bodies at depths below-sea level beneath the volcano would be undetectable magnetically because of the low magnetization contrast between the non-magnetic magma and the sedimentary units (10^{-6} emu/cc) which, based on the explosion seismic data, underlie the Island. There is, however, strong evidence in the aeromagnetic data for a magnetized

MODEL 3

$$J_0 = 0$$

$$J_1 = 1 \times 10^{-4}$$

$$J_2 = 1.5 \times 10^{-2}$$

Figure M13: Calculated magnetic field of basic cone (slabs 1-8A) plus large cold core (slab 17), and a central hot core (slab 16).

Calculated field of basic slabs 1-8a at $J=6 \times 10^{-4}$ emu/cc, pinnacle slabs 22 and 30 at $J=5 \times 10^{-3}$ emu/cc and hot core slab 16 at $J=0$.

Calculated field of magma sill slab 26 at $J=1 \times 10^{-3}$ emu/cc.

Figure M14: Top: Calculated field of basic slabs 1-8a at $J=6 \times 10^{-4}$ emu/cc, pinnacle slabs 22 and 30 at $J=5 \times 10^{-3}$ emu/cc and hot core slab 16 at $J=0$. Bottom: Calculated field of magma sill slab 26 at $J=1 \times 10^{-3}$ emu/cc.

sill at shallow depth beneath Augustine. Introducing a slab of magnetized material at shallow depth solves a major problem inherent in all models discussed so far, in that the ratio of the magnitudes of the high and low of the calculated field was at least a factor of 2 smaller than for the observed field. Introducing the sill adjusts the calculated ratio to more clearly fit the observed one. Figure M14 (bottom) shows the field of a circular sill, 300 m thick, with its base at a depth of 600 m below sea level and a magnetization of 1×10^{-3} emu/cc. The effect of adding the field from the sill to the total field is to widen the anomaly and reduce the amplitude of the low. Adding other known geologic features such as the South Peak, a magnetized cap representing the 1964 summit lava dome, a lava flow on the north side of the Island and the Kamishak intrusion produces a fairly reasonable match of calculated and observed anomalies.

Results: Best Model

The best model of Augustine Volcano is shown in Figure M15 (see also Table M4). The bulk of the volcano consists of pyroclastics (slabs 1-9) with a magnetization of 2 to 6×10^{-4} emu/cc; we used 4×10^{-4} emu/cc. The uplifted sediments on the southern flank of Augustine, having a similar susceptibility as the pyroclastics, were included in the basic stack of slabs 1-9 (Table M4) that approximate the overall shape of the volcano. There is no large interior magma chamber which is consistent with the refraction data, but we included in our model a narrow non-magnetic ($J = 0$ emu/cc) feeder pipe 300 meters in diameter down the center of the volcano (slab 16). The feeder pipe is encased with a magnetic solid shell ($J = 10^{-3}$ emu/cc) in our model extending to

Figure M15: Top: Calculated magnetic field of best magnetic structure model of Augustine Volcano consisting of (1) a basic cone extending to 340 ft. below sea level (slabs 1-9 at $J = 4 \times 10^{-4}$ emu/cc), (2) a central hot core, 300 m in diameter (slab 16 at $J = 0$), (3) a magnetic sill (slab 26'), 6.5 km in diameter at a depth of 1200 to 1505 m at $J = 5 \times 10^{-4}$ emu/cc and (4) slabs 22, 30 (pinnacles), slab 31 (Kamishak plug), slabs 33-38 (north side lava flow) and slab 43 (south peak) at $J = 1 \times 10^{-3}$ emu/cc.

Bottom: 1972 LKB aeromagnetic survey, flown at 4000 ft., ASL.

the base of the volcano. Surface manifestations of this shell are the South Peak (slab 43) and the Pinnacles (slabs 22, 30). The feeder pipe is capped by the 1964 lava dome, modeled as a 100 to 200 m thick cap with the same magnetization as the casing ($J = 1 \times 10^{-3}$ emu/cc). The intrusion on the south side of the volcano near VABM Kamishak was interpreted as an old plug dome (Kamishak Dome) with a magnetization of approximately 1×10^{-3} emu/cc (slab 31). This intrusion may be responsible for the uplift of the basement sediments on that side of the volcano.

Below Augustine Island, at a depth of about 1 km, lies a large sill of solidified magma (slab 26). The extent of the sill is not known, but is probably at least 100 meters thick and several kilometers in diameter. It is as highly magnetized as the flows and domes on the surface.

A prominent lava flow at the northern flank of the volcano was modeled as slabs 33-38, with a magnetization of 1×10^{-3} emu/cc. There may be other solid lava flows as well as sills and dikes of solid magma buried beneath the pyroclastics and interspersed throughout the volcano, but apparently these have a minor effect on the total field anomaly.

As discussed by Barrett (1978) the cumulative error, due to the way the contouring is done, the spacing of flight lines which affects resolution, the accuracy of navigation and the geometrical approximation of magmatic bodies by straight line segments is insignificant when compared to the uncertainty in the actual determination of the size, shape and magnetization of the various slabs in the model. We do not claim that the best fit model is accurate in detail, but it is consistent with the geologic data and the explosion seismic data. We therefore believe that the basic components of the model are fairly realistic, even though their shape and dimensions may depart somewhat from the true dimensions inside the volcano.

The main new piece of information from the magnetic data is (1) the probable existence of a solidified magnetized sill at shallow depth beneath Augustine which is confirmed by the seismic refraction data, and (2) that there are no large volumes of material below Curie temperatures within the volcano proper, a fact again confirmed by the seismic data.

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APPENDIX

Photogrammetric Mapping of the Summit Region of Augustine Volcano

The U.S. Geological Survey Iliamna Quadrangle Map B-2 of Augustine Island is based on aerial photography flown in 1957 and field annotated in 1958. The 1963/64 eruption altered the volcano, particularly the summit region, significantly. For example, after the 1964 dome intrusion the summit was 279 ft. higher.

With funding from this contract a new post 1964-pre 1976 topographic map was prepared of the Augustine summit region above the 2,500 ft. contour line, based on photography flown by North Pacific Aerial Surveys, Inc., Anchorage, on June 16, 1973. The original map has a scale of 1" = 400' and a contour interval of 10'. Figure A1 shows a photographic reduction of this map.

The 1976 eruption again altered the summit region dramatically. With funds from the State of Alaska, Department of Natural Resources, Division of Geological and Geophysical Surveys, we again acquired new vertical photography through North Pacific Aerial Surveys and contracted them to prepare a new topographic map of Augustines summit above the 2,500 ft. contour line at the same scale as the previous map. Figure AS2, based on August 21, 1976 photography, shows the dramatic changes that took place during the 1976 eruptions, which resulted in the removal of the 1964 lava dome, the creation of a large crater which was subsequently filled partially by a new dome intrusion which took place between February and April, 1976.

Figure A1: Topographic map of Augustine Volcano's summit region based on aerial photography taken on June 16, 1973, i.e., post 1963/64 eruption but pre-1976 eruption. The contour interval is 10 ft. and the distance between tickmarks is 2,000 ft. (prepared by North Pacific Aerial Surveys, Inc., Anchorage, Alaska).

Figure A2: Topographic map of Augustine Volcano's summit region based on aerial photography taken on August 21, 1976, i.e., post-1976 eruption. The contour interval is 10 ft. and the distance between tickmarks is 2,000 ft. (prepared by North Pacific Aerial Surveys, Inc., Anchorage,