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SEISMOLOGICAL INVESTIGATION OF CRACK FORMATION
IN HYDRAULIC ROCK FRACTURING EXPERIMENTS
AND IN NATURAL GEOTHERMAL ENVIRONMENTS

MASTER

Progress Report
for period September 1, 1979 - August 31, 1980

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Table of Contents

	Page
Disclaimer	iii
Progress Report COO-2534-6	1
Abstract	2
Scope of Investigation	4
A synthesis of seismic experiment at the Fenton Hill Hot Dry Rock system	7
Figures	15
Attenuation of high-frequency shear waves in the lithosphere	21
Figures	25
A new kinematic source model for deep volcanic tremors	27
Figures	30
Ground motion in the near-field of a fluid-driven crack and its interpretation in the study of shallow volcanic tremor	33
Table 1	48
List of Figures	49
Figures	54
Low velocity bodies under geothermal areas	76
Coso, California	76
Roosevelt Hot Spring, Utah	77
Geysers-Clear Lake, California	78
Yellowstone, Wyoming	78
Figure	81
Operation of event-recorders in Mt. St. Helens and Newberry Peak with preliminary results from them.	82
Figure Captions	85
Figures	86
References	89

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Abstract

This is a progress report for period September 1, 1979 to Aug. 31, 1980 on the project entitled "Seismological Investigation of Crack Formation in Hydraulic Rock Fracturing Experiments and in Natural Geothermal Environments." Under this title, we have been working on the following three major problems in geothermal seismology.

The first is to define the geometry and mechanical construction of a geothermal system such as the Hot Dry Rock geothermal energy source at the Fenton Hill site. The second is to find the transport rate of geothermal fluids such as magma from observations of seismic vibrations caused by their movement. The third is to map 3-dimensional distribution of seismic velocity and attenuation for defining the extent of geothermal energy sources.

In the past year, we concentrated our major efforts on the following specific problems.

(1) Synthesis of various seismic experiments carried out at the Fenton Hill Hot Dry Rock site in order to define the fracture system.

(2) Theoretical and experimental studies on attenuation of high-frequency shear waves in the lithosphere, in order to find the attenuation mechanism.

(3) A new kinematic source model for deep volcanic tremors, in order to relate the magma transport rate with the tremor amplitude.

(4) Theoretical studies on the ground motion in the near-field of a fluid-driven crack and its interpretation in

the study of shallow volcanic tremors.

(5) Review of seismic 3-D structures under geothermal areas.

(6) Operation of event-recorders in Mt. St. Helens and Newberry Peak for collecting data on volcanic tremors.

We published a paper on attenuation in lithosphere. Three papers are in press (3-D velocity anomaly, velocity change at the Hot Dry Rock site and summary of Kilauea Iki experiment. We contributed to the U.S.G.S. workshop in the Geysers - Clear Lake geothermal area, by reviewing our work on geothermal seismology. We presented three papers (3-D velocity anomalies, attenuation, and our Kilauea Iki experiments) at the XVII General Assembly of the International Union of Geodesy and Geophysics held at Canberra, Australia in December, 1979.

Scope of Investigation

The present research program was originated from our involvement in two major geothermal projects, namely the Hot Dry Rock project of Los Alamos Scientific Laboratories and the Magma Tap project of SANDIA. In the Hot Dry Rock project, we have been engaged in the interpretation of various seismic data obtained by the LASL scientists for defining the crack generated by hydraulic fracturing at Fenton Hill. In the Magma Tap project, we have carried out a variety of active and passive seismic experiments on the magma body in Kilauea Iki, a partially frozen lava lake.

From these studies, we learned that the traditional seismology based on layered earth models and earthquake or explosion sources was not always useful in interpreting the seismograms obtained from geothermal areas. The structure in a geothermal area shows 3-dimensional complexity, and the seismic sources there involve the motion of liquid and gas through cracks. In order to deal with seismic data from such complex sources and medium, we have been developing a new interpretation method based on seismic wave generation, transmission, scattering and attenuation in a medium which contains fluid-filled cracks.

In the present reporting period, the principal investigator has attempted to synthesize available experimental results from the Fenton Hill to come up with a model of fracture system which explains various observations. Our model will not be a simple one. It will consist of a system of large-scale cracks which are imbedded in a volume containing small-scale

cracks. The large-scale cracks are modelled deterministically, but a random medium is used to characterize the small-scale cracks.

The area to which the co-investigator put a major effort in the present reporting period was a theoretical study of volcanic tremors at near and far-field of the fluid-filled crack model (Aki et al., 1977a) buried in a realistic layered medium and its application to the observations at Kilauea. The result from this study will be useful for interpreting the volcanic tremors observed recently at St. Helens.

Last summer, we initiated a field experiment in collaboration with the Oregon State University at the Newberry Peak volcano using four digital event-recorders developed as a part of the present project. This year, M. Fehler of Oregon State installed one of them at the southern foot of Mt. St. Helens which was operating until the major eruption in May. We recovered the instrument successfully with numerous records of tremors, but unfortunately, the tape was used up one day before the eruption. Currently, we are operating eight event-recorders at St. Helens and Newberry Peak for further data collection.

In addition to the above works carried out by the people supported by the present project, the principal investigator has been collaborating with many other scientists on the broad area of geothermal seismology. He and his colleagues developed methods for studying (1) 3-D structure under geothermal areas by the use of teleseismic P time data (Aki et al, (1977b)), (2) Attenuation (Q^{-1}) of short-period waves by a combined

use of S and coda waves (Aki (1980)), and (3) Simultaneous determination of source parameters of earthquakes and 3-D velocity structure using local earthquake P time data (Aki and Lee, (1976)). These methods have been applied to many areas around the world, and the results are being collected, compared, and summarized by the principal investigator.

During the reporting period, the principal investigator published a paper on the attenuation of short-period shear waves in Physics of the Earth and Planetary Interiors, contributed to the U.S.G.S. The Geysers - Clear Lake workshop, presented two papers on 3-D inhomogeneities (which will be published in Tectonophysics) and attenuation at the XVII General Assembly of the International Union of Geodesy and Geophysics held in Canberra, Australia, lectured on volcanic tremors at Stanford Univ., Penn State, and U. of South Carolina, and organized a geothermal seminar at MIT. The co-investigator presented a summary paper on the Kilauea-Iki experiment at the Canberra IUGG meeting. The summary will be published in the Journal of Volcanology and Geothermal Research. M. Fehler is publishing a paper on the velocity change by cooling observed at the Hot Dry Rock site in the Journal of Geophysical Research.

In the following, we shall summarize the work in progress.

A synthesis of seismic experiments at the Fenton Hill Hot
Dry Rock system.

Various passive and active seismic experiments have been carried out at the Fenton Hill Hot Dry Rock site by the LASL scientists using sensors and sources placed in the boreholes. Their results have been described in LASL HDR Staff (1978) and Fehler (1979). Reviewing their results and re-examining original data, we came up with the following picture of the fracture system at the Fenton Hill site.

(1) A major vertical crack exists between GT-2 and EE-1 as evidenced in the converted waves observed in the Dresser Atlas dual well experiment. It is oriented in the NW-SE direction as indicated by the orientation of zone of seismic events generated by pressurization.

(2) The gneiss and granodiorite contact (dip angle 63° from horizontal, dip direction $N52^\circ E$) is opened by pressurization as evidenced by reflections observed in the shear-shadow experiment. This contact and others subparallel to it are partially open (self-popped) even without pressurization as evidenced in reflections observed in the Dresser-Atlas dual well experiment.

(3) In addition to the above system of large-scale cracks, small-scale cracks are generated in a volume surrounding the major crack system. The evidence for this comes from the fluctuation of P wave arrival time with the source depth and attenuation of coda waves of various events measured by Fehler. The linear dimension of small-scale cracks is around 2 meters.

In order to arrive at the model proposed above, we defined the fracture system as consisting of large-scale cracks imbedded in a volume containing small scale cracks, and applied a deterministic model to the former and a random medium model to the latter. A preprint of a paper by B. Ja. Gelchinsky and Karaev on seismic reflection from a complex geological body was helpful to understanding the deterministic part, and a book by Flatte et al (1979) on sound transmission through a fluctuating ocean was helpful to understanding the stochastic part.

When we consider waves reflected from a discontinuity (or crack) imbedded in a random medium, the following two factors play important roles. One is the complexity of the incident wave field. According to Flatte et al, the product of ϕ^2 (mean square phase fluctuation) and Λ ($= \frac{1}{2\pi} \left(\frac{R_F}{L}\right)^2$, diffraction parameter; where R_F is the Fresnel Zone radius, and L is the inhomogeneity scale length) is equal to the number of distinct ray paths arriving at a given sensor. For an isotropic random medium,

$$\phi^2 \sim \left(\frac{2\pi}{\lambda}\right)^2 \langle \mu^2 \rangle R L \quad (1)$$

and

$$R_F \sim \sqrt{\frac{R\lambda}{2}} \quad (2)$$

So, we have

$$\phi^2 \Lambda \sim 2\pi \langle \mu^2 \rangle \frac{R^2}{\lambda L}, \quad (3)$$

where R is the travel distance from the source to the reflector, $\langle \mu^2 \rangle$ is the mean square of fractional velocity fluctuation, λ is the wave length. If the above quantity is less than unity, we can safely assume that the incident wave is composed of a single ray. The incident wave is simpler, when the travel

distance R is shorter, wavelength λ is longer, or the scale-length L of inhomogeneity is longer.

The other important factor is the relative size of the reflector to the Fresnel zone radius. If the reflector radius a is much smaller than R_F , the reflection will be weak. If a is equal to or greater than R_F , and the receiver is in the lighted zone, the reflection will be strong.

Thus, for observing strong reflection, the following two conditions must be satisfied.

$$\Phi^2 \lambda \lesssim 1 \quad (4)$$

and

$$a \gtrsim R_F \quad (5)$$

Let us now apply the above guide to the interpretation of seismic signals obtained at the Fenton Hill site.

(1) In the so called shear-shadow experiment at the 2415-m shot line, a strong reflection was observed at about 40 m sec after the P arrival when the fracture is pressurized as shown in Fig. 1b. Such a strong reflection requires both conditions (4) and (5). Assuming that they are S to S reflections, their wave lengths are 5 to 10 meters. Since R is about 50 meters, R_F is 10 to 15 meters. Thus, the radius of reflector must be at least 10 to 15 meters.

Earlier (Aki, memo to G3, LASL, 1977), we proposed two alternative models for this reflector. One is the vertical crack oriented in NE-SW with the upper tip at about 2500 m depth. The other is the contact between gneiss and granodiorite as shown in Fig. 2 reproduced from LASL HDR staff (1978). We now prefer the latter, because of the higher strength of

reflection expected for the reflector plane normal to the incident wave direction than for the reflector p parallel to it. The strength may be proportional to the reflecting area vibrated with the same sense of motion. Then, the strength ratio of the latter to the former will be $\pi R_F^2 / R_F (\lambda/2) \sim 10$. Another reason for preferring the latter is that if we choose the former, the orientation of the crack become inconsistent with the trend of seismic events determined from passive experiments.

(2) The above choice of reflector is supported by the discovery of S to S reflection from the same reflector during the Dresser-Atlas dual well experiment. Fig. 3 shows the prominent reflection SS observed when the receiver was fixed at 2586 m in GT-2B and the source was moving in EE-1. Fig. 3 was obtained when the fracture was unpressurized. The neighboring seismograms correspond to the source spacing of 4 cm. The dominant frequency is about 10 KHz. The SS phase does not appear continuously, but shows up sporadically over a vertical distance of 1 m or less. If we draw an ellipsoid for each appearance of SS, in such a way that the total distance from the source to a point on the surface of ellipsoid and to the receiver is equal to the observed travel time for SS, we find that all the ellipsoids make contact with a common plane which is very close to the gneiss-granodiorite contact, strongly suggesting that SS is the reflection from the contact.

The Fresnel zone radius R_F in this case is about 1.5 m, implying that the contact is opened, even when the fracture is unpressurized, at patches with radius at least 1.5 m. Such

a small patch may only give weak reflection for large R_F as in the case of shear-shadow experiment, in which the reflection was not observed when the fracture was unpressurized.

(3) When the fracture is pressurized in the dual well experiment (at receiver depth 2586 m), SS disappears primarily because S disappears. They are attenuated by scattering or absorption. However, P waves are well observed, and seismogram shows a strong arrival which may correspond to the P to S conversion. Both scattering and absorption effect must be weaker for P waves than for S waves.

(4) For the section one step deeper than the above, we do not observe coherent S reflections even when the fracture is unpressurized. However, in this case, we observed P to P reflection which corresponds to one of the joints subparallel to the gneiss-granodiorite contact as shown in Fig. 2. We also observed very coherent regular waves which may be converted waves at the main crack as shown in Fig. 4. Detailed studies of these waves are currently undertaken by LASL seismologists.

(5) Fehler (1979) obtained the root mean square deviation of P arrival time from the parabola expected for a homogeneous body in the dual-well experiment. His result showed that the root mean square residual σ ranges from .01 to .02 m sec. For P waves with frequency 10 KHz, this corresponds to the value of ϕ^2 in the range 0.4 to 1.4.

Using the correlation distance $L \approx 2$ m also estimated by Fehler, we find from Equation (1)

$$\langle \mu^2 \rangle \sim 1 \sim 3 \cdot 10^{-4} \quad (6).$$

This is probably a minimum estimate for the velocity fluctuation, because we assumed that the fluctuation is uniformly distributed from the source to receiver.

Another way of estimating $\langle \mu^2 \rangle$ is to consider the complexity of incident S waves in the pressurized case which made coherent SS to disappear. From equations (3) and (4), we get

$$\Phi^2 \Lambda \sim 2\pi \langle \mu^2 \rangle \frac{R^2}{\lambda L} > 1$$

for S waves in the pressurized case. For $R = 12$ meters, $\lambda = 0.3$ meters, and $L = 2$ meters, we have

$$\langle \mu^2 \rangle > 7 \times 10^{-4} \quad (7)$$

which is higher than the estimates in (6).

Apparently, $\langle \mu^2 \rangle$ for S waves is greater than $\langle \mu^2 \rangle$ for P waves.

(6) We have an interesting evidence supporting a contention that the attenuation may be due to scattering. Fehler (1979) made measurements on the root mean square residual σ for both pressurized and unpressurized cases, and his result shows that σ^2 (and therefore Φ^2) is increased by 40% by pressurization. Fehler also studied the Q of coda part of seismograms (which is probably composed of back-scattered S waves) of the dual well experiment, and found that Q decreased by 34% by pressurization. (The coda waves are probably sampling the volume 10 to 20 meters from the main crack.) The percentage change in Q and Φ^2 agrees within the error of experiment. Since the loss of energy by scattering is proportional to Φ^2 , the above agreement suggests that the

attenuation of coda waves may be accounted for entirely by scattering due to inhomogeneity. However, there is a possibility that the attenuation is primarily caused by absorption, and the agreement between the increase in ϕ^2 and Q^{-1} is mere coincidence.

(7) Examining records of seismic events observed during a passive experiment (No. 203), we found that there are two distinct groups of earthquakes with different coda decay curves. Fig. 5 and 6 show the decay curve in db as a function of time from the S arrival. If we fit the theoretical curve of the form $t^{-1} \exp(\frac{-\pi f t}{Q})$ (t is time from the origin time) to the observed, one group gives Q of 35, and the other 106. Such an observation is very rare. Usually, the coda decay is common to all local earthquakes for a given station. Only exception is the station at Garm, Tadjikistan, which lies on the border of two distinctly different seismic regions. One is composed of crystalline rock, and the other is composed of deformed sedimentary rock. In this case, earthquakes in one area showed different coda decay (up to a certain time) from that for earthquakes in the other area. Thus, the observed difference in coda decay means distinctly different locations between the two groups; one in the low Q region including the fracture zone, and the other outside of it. Thus, the coda envelope will help to remove ambiguity in the single-station location method developed by LASL scientists.

(8) Fehler's results on Q of coda waves as well as additional data from passive experiments show that Q increases with frequency in the range 0.2 to 15 KHz. It appears that a peak of Q^{-1} lies around a fraction of 1 KHz. This again

supports the scattering hypothesis for attenuation, which predicts the peak at around $ka \sim 1$. In our case, $a = 2\text{m}$, $\lambda \sim 12\text{ m}$, and therefore the corresponding frequency is about 0.3 kHz for S waves in agreement with observation.

FIG 1

Triaxial geophone records from the (a) 2644-m and (b) 2415-m shot lines--full coda.

Figure 2

128-185

Figure 4

Figure 5

Figure 5

Figure 6

Attenuation of high-frequency shear waves in the lithosphere

As reported in last year's progress report (COO-2534-5) and published in Aki (1980), we developed a new reliable method for measuring attenuation of high frequency shear waves using micro-earthquake data and applied it successfully to the lithosphere under central Japan. The result was rather surprising: the quality factor Q increases with frequency f , for $f > 1$ Hz, in the form of $Q \propto f^m$, where $0.6 < m < 0.8$. Combining data from other regions and at other frequencies, we found an interesting frequency dependence of Q^{-1} as shown in Fig. 1. For stable areas such as central and eastern U.S., Q^{-1} is very low, or Q is 1000 or more for all frequencies in the range from 0.02 to 25 Hz. On the other hand, for tectonically active areas such as Japan, California and Iran, there is a peak in Q^{-1} around 1 Hz, and Q^{-1} decreases toward both ends. The curves for central Asia (Garm, Tajikistan) lies between the active and stable areas, with peak for Q^{-1} at a frequency lower than 1 Hz.

In order to explain these observations, we proposed two alternative hypotheses; one is the loss by scattering due to heterogeneity, the other is the grain boundary relaxation.

One of our graduate students, Scott Phillips, examined the latter hypothesis by using the parameters of grain boundary relaxation such as activation energy and volume determined by Jackson (1969, 1971). Temperature profiles for central Japan estimated by Horai and Uyeda (1964) were used to calculate the total Q^{-1} for various seismic ray paths. The result was then compared with observations in Fig. 1.

The agreement between the observed and predicted Q was rather unsatisfactory. First, none of the observed curves for Q^{-1} for the crust can be explained by the model because of too low temperature. Secondly, some permissible temperature gradients gave a good fit to observed Q^{-1} for mantle path for the frequency range $1 < f < 25$ Hz, but failed to fit observed high Q at low frequencies. Thirdly, the theory predicts that area B + C show lower Q than area A because of higher temperature at depth, but there was very little difference in observed Q between the two areas.

The scattering hypothesis, on the other hand, seems to be more promising. A visiting scientist from P.R.C., Wu Rushan, has been working with the principal investigator on the mathematical problems of wave scattering in random media characterized by an autocorrelation function for velocity fluctuation. He has shown, for an autocorrelation of the form $e^{-r/a}$, Q^{-1} due to scattering is proportional to f^3 (Rayleigh scattering) for low frequency, proportional to f^{-1} for high frequency, and has a peak between the above two regions at around $ka \approx 1$ where k is the wave number. The model which fits the observations for the central Japan gives the correlation distance a of $0.5 \sim 1$ km, and the mean square fractional velocity fluctuation $\langle u^2 \rangle$ of 0.02 to 0.04. These values correspond to the root mean square velocity fluctuation of 15 to 20%.

These fluctuations may appear unreasonably large. For example, from the travel time fluctuation of teleseismic P waves observed at LASA, in eastern Montana, Aki (1973) obtained

$\langle \mu^2 \rangle$ of 0.0016. However, since the Eastern Montana is in a stable area, we might expect higher Q and lower $\langle \mu^2 \rangle$ than Japan. Considering these and also the possibility that $\langle \mu^2 \rangle$ for S waves may be greater than that for P waves, the high $\langle \mu^2 \rangle$ value obtained by Wu Rushan may not be altogether unreasonable for tectonically active areas. Although the theory developed by Wu Rushan, is quite preliminary (neglecting, for example, S to P or P to S conversions), the scattering approach seems to be promising and encourages further studies.

An interesting observation shown in Fig. 1 is that Q^{-1} of coda waves is very close to Q^{-1} of S waves for Japan and central Asia. If this is universally true, we have a very simple method of measuring Q^{-1} of S waves. Fig. 2 shows Q^{-1} of coda waves for various regions. They all show a similar frequency dependence as seen in Fig. 1, except for station OTL located near the summit of an active volcano Kilauea, Hawaii. The Q for OTL is less frequency dependent than others which all show a familiar $Q \propto f^m$ dependence. This different frequency dependence suggests a different attenuation mechanism for Hawaii. Here, we can easily accept any temperature-sensitive activation process dominating the attenuation mechanism because we know that a large amount of magma is ascending through the lithosphere. Then, we can argue that the attenuation for the rest of the areas may be due to scattering rather than a thermally activated process.

The seismology group at CICESE (Centro de Investigacion Cientifica y de Educacion Superior de Ensenada, B.C.) has applied the same method to determine Q of S and coda waves

at Oaxaca and Cerro Prieto. At Oaxaca, they found a frequency dependence of Q similar to Japan and California, but at Cerro Prieto, another geothermal area, their preliminary result shows frequency dependence of Q similar to Hawaii.

In view of the usefulness of Q of coda also in the Hot Dry Rock characterization as discussed in the preceding section, coda waves will be considered as one of the major subjects of future study in the present project.

Figure 1

Figure 2

A new kinematic source model for deep volcanic tremors

As reported in last year's progress report (COO-2534-5), the principal investigator visited Hawaii Volcano Observatory in July, 1979 and examined 200 records of deep tremors. Preliminary analysis of the data was also reported in the progress report using a model of tremor earlier proposed by Aki et al. (1977a) for shallow tremors.

Since the high ambient pressure (~ 10 kbars) existing in the tremor source region will probably prohibit the formation of cavity during any mode of vibration, we now propose a kinematic model for dynamic transport of magma during which the total magma volume stays constant. Fig. 1 shows a schematic view of the model, which consists of two cracks connected by a channel; a version of the stationary crack model proposed by Aki et al (1977a). With the excess pressure building up in the crack on the left hand, magma starts to leave the crack and moves to the one on the right hand. Fig. 1 shows the change of crack shape as the magma movement spread over the whole crack area.

The above process can be simulated by the uniform displacement model shown in the lower half of Fig. 1, where the displacement of crack surface normal to the crack plane is shown as a function of time t and distance ξ measured from the channel center along the direction of magma movement. The magma front moves with a given velocity v , and displacement $\pm \Delta u$ is assumed to be constant. This model is as simple, and hopefully, as useful as the Haskell's model for an earthquake.

The above process will generate far-field P waves of the form as shown in Fig. 2, where W is the width of crack, L is the length of crack, α is the P wave velocity, ρ is the density, r is the distance to station, γ is the angle between the direction of magma movement and the direction to station, and θ is the angle between the normal to the crack plane and the direction to station.

We assume further that the process is repeated with the period approximately equal to the time that the pressure waves take to travel both ways through the length of crack. As shown in Fig. 2 of the next section, the motion of empty crack shows a very long-period oscillation. This period is comparable to the two way time of acoustic wave propagation in magma. There may be a coupling between liquid and solid vibration with this long period, which can be sustained for a long time. The validity of this conjecture will be tested in a future study using methods such as the one developed by Aki and Larner (1970) for studying seismic waves in a medium with irregular interfaces.

In the meantime, we accept the kinematic model as a simple scheme for recovering the physical parameters of the source from seismograms.

Since the flow q of magma is given by

$$q = WL\Delta u/T \quad (1)$$

$$\sim 2W\Delta uv \quad (2)$$

the flow can be expressed in terms of root mean square amplitude of P waves using the formula in Fig. 2.

$$q \sim 75\alpha r \text{ RMS}(u^P) \quad (3)$$

where r is the distance, we put $\lambda = \mu$, $\theta = 45^\circ$, and $\alpha/v = 4$.

The quantity on the right hand side of Equation (3) has been evaluated by the principal investigator for the deep tremors of Hawaii as shown in Fig. 3. The value depends on the tremor frequency, and increases as the square of tremor period, which is expected if large and small sources are similar. Since major tremors usually show the period of about 0.5 sec, we adopt the value of $r \times \text{RMS}(u^P)$ to be 30 cm^2 .

Then, we find that

$$q = 1.8 \times 10^3 \text{ m}^3 \text{ sec}^{-1}.$$

On the other hand, Swanson (1972) estimates the magma supply rate at Kilauea to be 0.1 km^3 in a year, and Koyanagi estimates the total duration of magma transport during the deep tremor period is

$$q = 1.5 \times 10^3 \text{ m}^3 \text{ sec}^{-1},$$

which is very close to the estimate by our formula.

For a more accurate comparison of tremor and magma transport it is necessary to find the tremor duration for different amplitudes. This work is currently being done at Hawaii Volcano Observatory by Dr. R. Koyanagi.

Figure 1

$$u^P = \frac{\lambda + 2\mu \cos^2 \theta}{4\pi\rho a^3 r} f(t)$$

$$f_1 = W\Delta u / \left(\frac{1}{v} - \frac{\cos \gamma}{a} \right)$$

$$f_2 = W\Delta u / \left(\frac{1}{v} + \frac{\cos \gamma}{a} \right)$$

$$t_1 = L \left(\frac{1}{v} - \frac{\cos \gamma}{a} \right)$$

$$t_2 = L \left(\frac{1}{v} + \frac{\cos \gamma}{a} \right)$$

$$T \sim 2L/v$$

Figure 2

FIG 3

Ground motion in the near-field of a fluid-driven crack and its interpretation in the study of shallow volcanic tremor

Volcanic activity, particularly at basaltic volcanoes and also at a few andesitic volcanoes, is often accompanied by sustained rhythmic motions of the ground which can last a number of hours. This type of motion is known as volcanic or harmonic tremor and it is generally accepted that its origin is related to the mechanism by which magma is transported at depth. The complexity of the observed seismic records, however, has so far prevented a rigorous analysis by the usual methods used in earthquake seismology.

The most obvious feature of the volcanic tremor is the peaked nature of its spectrum. The frequency of this peak sometimes changes in the course of eruptive activity and appears to be variable among stations closest to the source. Several attempts have been made to explain the nature of the tremor spectrum with models involving source or path effects (e.g., Omer, 1950; Shimosuru, 1961; Steinberg, 1965; Shima, 1958; Kubotera, 1974). Unfortunately, none of these models offers a natural explanation for the actual driving force of the tremor and neither do they offer a geometry compatible with the concept of magma transport. A more realistic approach was recently taken by Aki et al. (1977a) who proposed three possible models of a fluid-filled tensile crack driven by excess pressure to fluid to explain the tremor in terms of physical parameters of the source while at the same time offering a geometry adequate for transport. In the first two of these models, magma is transported through a succession

of jerky extensions at one or both ends of the crack, and in the third, it is the jerky opening of a narrow channel connecting two cracks which controls the migration of magma between neighboring cracks. In these models, it is the sudden extension of the crack or opening of the channel which acts as the source of seismic waves. Sustained tremor is then obtained by the superposition of many such events occurring randomly in time and depending on their spatial position on the actual path of magma. Calculating the far-field radiation from these sources and testing their results against the observations made during the October 1963 east rift eruption of Kilauea volcano, Aki et al. determined the model most compatible with the available data to be that of a chain of cracks connected by narrow channels which open and close in response to pressure fluctuations in the magma. They found that the length of individual cracks of 1 km, area of channel connecting adjacent cracks of 10^3 m^2 , magmatic excess pressure responsible for the channel opening of 20 bar, and frequency of opening of each channel of about 1 per 15 sec to be in reasonable quantitative agreement with the observation. Using this model of the source and the parameters that they determined, we have now succeeded in obtaining a complete representation of the motion of the ground in the near-field of a buried fluid-driven crack. Our approach uses Bouchon's discrete wave number representation of elastic wave fields for an arbitrary three-dimensional seismic source (Bouchon, 1979). The general configuration of the problem is depicted in Fig. 1. The source is the tensile opening of a vertical crack with rectangular dimensions of 2 by 1 km. The opening space-time function, specified in

Fig. 2, is assumed to depend on the horizontal dimension of the crack only, that is we take it to be constant in the vertical direction. In other words, the situation which we consider is that of the jerky opening of a vertical channel bisecting the crack in half. The data of Fig. 2 represent the finite difference solutions obtained by Aki et al. in response to a step in pressure applied in the channel; the results show the normal components of particle velocities and displacements calculated at five positions on the surface of the crack and given as dimensionless functions of the excess pressure ΔP , lengths L and Δl of crack and channel, as well as the compressional velocity α and rigidity μ of the rock. Advantage of the symmetry of the problem was taken by applying the pressure at two grids near the center of the crack and calculating the response for one half of the crack only. The grid labelled 5 is the location where the step in pressure was applied on this half of the crack. Note that the static opening at grid 1 is not zero because this position is offset by half a grid from the edge of the crack due to the discretization. The calculations were made assuming a Poisson ratio of 0.25 in the rock matrix and for two different values of the parameter C shown at the top of the figure. This parameter is called the crack stiffness and characterizes the effect of fluid in the model. The value $C = 0$ represents a dry (or empty) crack and the value $C = 5$ represents a minimum which may be compatible with the situation of a shallow crack filled with molten basalt. The implication of the latter choice was discussed in Aki et al. (1977a).

As defined in Fig. 1, the crack problem has a double symmetry, one with respect to the crack plane, and one with respect to the vertical plane normal to the crack and intersecting it along the channel axis. Denoting the three components of the ground displacement by (u,v,w) , where u and v are the horizontal longitudinal (parallel to the crack) and transverse (normal to the crack) components, respectively, and w is the vertical component, we find, specifically, that u is symmetric and v is antisymmetric with respect to the crack plane, while u is antisymmetric and v is symmetric with respect to the orthogonal plane containing the channel axis. The vertical component, w , itself has a symmetry with respect to both planes.

The calculation of the ground motion is carried out at 441 locations distributed over a square area of 3 by 3 km with two sides of it chosen to coincide with the traces of the two planes of symmetry. To improve the visual representation of the ground response, the data from this quadrant is then duplicated onto a second quadrant which corresponds to the other half of the crack. As indicated in Fig. 1 we thus consider a total of 861 locations in an area of 6 by 3 km adjacent to the fault trace. The separation between individual locations is 0.15 km in both longitudinal and transverse directions. Additional calculations are also carried out along 7 profiles normal to the crack and spaced 0.5 km apart. These profiles are marked by the heavy lines in Fig. 1; note that profile 1 corresponds to the plane of symmetry which is perpendicular to the crack and that profile 3 is aligned with the crack lip.

In our analysis we shall first compare the ground responses obtained for wet and dry cracks in a homogeneous half-space assuming a fixed source position. Restricting our attention to the fluid-filled crack we shall then investigate the effect of source depth and finally we shall analyze the ground response in the presence of a single surficial layer of unconsolidated volcanic deposits. The source-medium parameters for the various models are summarized in Table 1.

Snapshots of the ground motion for the first model are shown in Fig. 3. This is the case of the fluid-filled crack (see Fig. 2) buried at a depth of 0.5 km in a half-space with compressional and shear velocities of 3 and $\sqrt{3}$ km/s, respectively, and a density of 2.7 g/cm^3 . The snapshots are taken every 0.25 sec, starting 0.25 sec after the opening of the channel and covering the first 3 sec of the event. The static ground deformation resulting from the opening of the crack and calculated using Maruyama's (1964) formulation is shown at the bottom of the figure for comparison. By convention, the displacements outward from the source are positive. The snapshot sequence clearly shows the onset of the disturbance and the propagation of the various wavefronts over the area. The transition from the dynamic to the static solution is also easy to follow. It is interesting to note that the static vertical and longitudinal displacements are pointing down and inward directly above the source. Further details of the ground displacements associated with this event are depicted in the seismograms of Fig. 4 obtained at 8 stations along the 7 profiles shown in Fig. 1. The profiles are listed at the top

of the figure and the stations are set at distances of 0, 0.3, 0.6, 0.9, 1.2, 1.8, 2.4 and 3 km from the fault trace, shown from top to bottom for each individual component of displacement. The records start with the arrival of the P wave and last 3 sec from this arrival, including the entire ground response for the event at each location. All the displacements are normalized by the product $\Delta P \Delta l$ of the overpressure applied in the channel times the length of the channel. There are several outstanding features to these seismograms, namely, the first motion which is outward everywhere, a marked oscillatory character of the signal due to the combination of stopping phases from the edges of the crack and the vibratory motions of the source proper, a rapid damping of the vertical component of motion with range, and the presence of a large SH wave clearly observed on the two horizontal components at the most distant stations. The first motion is very small, indeed barely visible on the transverse component, close to the fault trace. The displacement, however, grows rapidly as the distance from the fault increases, reaching its peak at a range of about 1 km. The directions along the first profile and along the fault trace are nodes for the SH motion because of the antisymmetry of the problem for horizontal displacement about these two lines. A better view of the P-SH and P-SV motions of the ground is offered in Fig. 5, which shows polar plots of the particle trajectories in the three orthogonal planes defined by u , v and w . The configuration of the profiles and stations is the same as that of Fig. 4 with the exception that purely

vertical, transverse, and longitudinal motions are not represented. The Figure reveals a complex pattern of motion in the immediate vicinity of the source, but developing into a marked SH character as we move away along the azimuth which bisects the two nodal planes for this type of motion (see data at the range of 3 km on profile 7). This is, of course, the direction of maximum radiation for the far-field SH waves. The two nodal planes are planes where the motion is entirely P-SV. The trace of one of these planes - the direction of profile 1 - is the axis of the main lobe of radiation for the Rayleigh wave. While Rayleigh motion is not yet important in the range studied, it begins to be visible at the most distant station on profile 1 where a small loop with counter-clockwise motion, labelled by the letter R, can be seen at the end of the particle trajectory for that station. Our calculation shows that this loop develops into the well-known elliptical motion as the contribution of the Rayleigh poles become dominant with increasing range. The Rayleigh wave, in fact, dominates the record at distances beyond 9 km in the present situation. The motion in the perpendicular direction given by the fault trace, is predominantly compressional with a small amount of SV coupling. Our computation again indicates that this evolves to exhibit the typical Rayleigh motion at the distance of roughly 9 km. The Rayleigh radiation is at a minimum in that direction, however. This is naturally understandable because the component of the seismic moment tensor of the source which affects this particular lobe is smaller by a factor 2μ than

the component acting in the direction of profile 1.

So far we have only considered the ground response in the time domain, analyzing the conspicuous features of the displacement seismograms and investigating the type of motions induced by the source. Another description of the ground response which is perhaps even more useful in the analysis of the tremor situation is the representation in the frequency domain. This is done in Figs. 6 and 7. Both figures exhibit space-frequency spectra of ground velocities over the band from 0 to 5 Hz obtained from the entire record of motion along individual profiles; the first figure gives a three-dimensional representation emphasizing the prominent characteristics of the spectra at a glance, and the second shows contour maps of the spectral amplitudes for a detailed interpretation. The seven profiles are depicted from top to bottom in Fig. 6 and the first four of these appear from left to right in Fig. 7. Contour intervals and spectral amplitude scales for the normalized components of ground velocities are given at the bottom and right in the latter figure. The feature which immediately comes to attention in those figures is the peaked structure of the spectra; moreover not a single but several peaks are observed for some profiles (see transverse component). That the spectra have a dominant peak is not astonishing in itself since the source clearly has a peaked spectrum of its own as can be gathered from the marked overshoot and ringing associated with the wall motion depicted in Fig. 1. It is thus quite plain under the present circumstances that it is the source, not the medium, which is

responsible for this peaked characteristic of the ground velocity spectra. More surprising, however, is the clear evidence in the first figure that the position of the peak is not fixed in frequency. We see, for example, that the strong peak is observed at low frequency on the transverse component eventually vanishes completely giving place to another peak at a higher frequency as we move from profile 1 to profile 7. The other two components of velocity also present some frequency shift but the trend is less marked in those cases. It is remarkable that the frequency of these dominant peaks can vary by as much as one octave within a single component of motion. In the present situation, this variation appears to reflect the difference between the characteristics of the first motion and those of the prevailing SH displacement observed in the records.

An improved understanding of the effect of the fluid on ground excitation in the model may be gained through the comparison with a different source situation. An interesting approach is to perform this comparison for the limiting case of a dry crack. This is the option taken in our second model. For this study, the setting of the source and stations and the physical properties of the medium remain identical to those used earlier (see Table 1) but the source space-time function is now that corresponding to the case $C = 0$ in Fig. 2. Plots of the surface displacements, seismograms and contours of spectral amplitudes are given in Figs. 8, 9 and 10, respectively. Note that the scales of displacement and time are changed in Fig. 9 and so are the frequency scales and the

normalized spectral amplitudes of ground velocities in Fig. 10. Several differences with respect to the previous results are readily discernible in those plots: the total duration of motion is longer and lasts about 6 s, the dynamic and static displacements are larger than those induced by the fluid-driven crack, the motion is dominated by the transverse component over the first four profiles, the SH displacement is comparatively small, and Rayleigh motion is not observed within the area depicted. Both transverse and vertical displacements show a long period signal which distinctly stands out, especially on the transverse component in profiles 1 and 2, and appears to reflect the fundamental mode of vibration of the source (compare with Fig. 2). This characteristic of the signal is more clear yet in the spectral plots of Fig. 10 where we observe a dominant peak at the frequency of 0.33 Hz for the transverse velocity near the source. Again more than one spectral peak is associated with each velocity component. Of particular interest is the presence of four marked peaks in the bandwidth of 0 to 1.5 Hz and within 1.5 km of the fault trace for the vertical ground velocity. Also worth noticing is the variation in the positions of the spectral peaks between the two horizontal components.

Another effect which can be studied in the homogeneous half-space is that of source depth. To do this study we go back to the model of the fluid-driven crack (case C = 5 in Fig. 2) and , keeping all other parameters the same as before, move the source depth from 0.5 km to 0.05 km (see Table 1).

The results of our calculation are shown in Figs. 11, 12 and 13. Apart from the amplification effect and obvious differences in the transverse and vertical components of motion in the immediate vicinity of the source, the overall features of the seismograms are not greatly affected by the shallower depth of burial of the crack. Strong ground motion is confined to a zone, roughly 1.5 km wide, adjacent to the fault trace. The displacement in that region is largely transverse and the static solution for this component clearly outlines the static opening on the crack (compare with Fig. 2). Peak displacements are reached within 0.3 km of the epicenter and, similarly, the spectral peaks shown in Fig. 13 are positioned closer to the fault trace, becoming mostly coincident with it in the case of the vertical component of velocity. A marginal increase is observed in the dominant frequency associated with the transverse velocity, otherwise the frequencies are not too different from those depicted for case 1.

Having considered the effects of source fluid and source depth we shall now investigate the ground response in the presence of a surficial layer of unconsolidated volcanic materials. Apart from this new layer, the configuration and parameters of the source, medium, and receivers are identical to those used in our study of case 1. We shall assume the density, compressional and shear velocities of the layer to be fixed to the respective values $\rho = 2.3 \text{ g/cm}^3$, $\alpha = 1.7 \text{ km/s}$, $\beta = \alpha/\sqrt{3} \text{ km/s}$, but allow the thickness to vary and study three cases for which it takes the values 0.25, 0.5, and 0.75 km. The relevant parameters for these models are listed as cases 4, 5 and 6 in Table 1 and the results of the calculation

are presented in Figs. 14, 15 and 16.

The pattern of radiation is the same as in the previous cases studied. The direction along the fault trace and the direction perpendicular to the fault and intersecting the epicenter are nodes for SH motion; the maximum radiation for motion of this type is in the direction bisecting these two nodal lines. These lines also correspond to the axes of the lobes of Rayleigh radiation with maximum Rayleigh motion found in the direction normal to the fault. The first motion, which is always directed outward from the source, is very small in the vicinity of the fault and reaches a maximum within a range of 1.5 km from the epicenter. The presence of the layer results in an amplification of the ground displacements by a factor of 2 to 3 and in a longer signal duration which increases with range as the motion evolves into a wave train consisting of Love and Rayleigh waves as well as multiply reflected SH and SV waves. The vertical component of motion attenuates somewhat faster than the two horizontal components with the result that beyond a few km from the fault SH motion dominates the record. In the presence of a thin layer (case 4), well-developed coherent Rayleigh waves are clearly observed in the later part of the signal on the vertical component at distances beyond 2 km along the first profile (Fig. 15a). Compared with horizontal motion, the period of these Rayleigh waves is distinctly shorter, as can be seen by matching the records obtained on the last station of profile 1 with the records of the last few stations on profile 7. The displacement waveforms increase in complexity as the layer gets thicker and as the source becomes coupled with it. We note, in

particular, that the period of Rayleigh motion increases while Love and SH waves now consist of short period wavelets superposed onto a longer period signal. An outstanding feature of the horizontal records is the large pulse-like direct shear phase from the source which is observed along profiles 3 and 4. This phase is associated with the conspicuous peak visible in the transverse velocity spectra depicted in Figs. 16a,b,c and is, in fact, the largest feature seen in the latter two figures. As in the case of the half-space model, the spectral structure of the velocity records is always peaked and, depending on station position, may show a double peak or a dominant peak with smaller subsidiaries.

One of the major results from our study is thus the discovery of dominant peaks in the ground velocity spectra, a well-known feature observed in shallow tremor and which can be related to the dimensions of the fault plane in our model. The relationship between the dominant frequency and actual source parameters is not straightforward, however, as the spectrum varies considerably throughout the epicentral region, depending not only on the receiver location but also on the particular component of motion being considered. This is due to the fact that the spectral characteristics in the near-field reflect the rapid evolution of complex and as yet undecoupled compressional and shear motions generated by the source toward the more easily recognizable phases observed at larger distances. Depending on receiver position, the observed spectral peak may represent purely compressional motion, or compression coupled with shear or unpolarized shear or, as

the epicentral range increases, SH or SV waves, or finally in the far-field, Love or Rayleigh waves. The presence of even a single layer in the medium further adds to this complexity by modulating the spectral peaks to the extent that their frequency, bandwidth, and amplitude may change. Aside from the expected amplification of the ground displacements due to the layer, we found that a slight change in the source-medium configuration may affect the dominant frequency by up to two octaves or more on any component of motion. The importance of medium and path effect in the modulation of the tremor spectrum is well illustrated in Figs. 7 and 16a,c,b. Thus it appears imperative to take the effect of medium structure into consideration in the determination of the actual source parameters from available observation.

Using the statistical method of Aki et al (1977a), we find the following relationship between the RMS tremor amplitude and the spectral amplitude calculated for a single event in our model:

$$\langle g^2(t) \rangle^{1/2} = \sqrt{2n\Delta f} |\dot{G}_0(\omega_0)|/\omega_0$$

where $g(t)$ represents the amplitude of total seismic motion due to the various cracks, n is the frequency of channel opening, Δf is the bandwidth observed for the tremor peak, and $|\dot{G}_0(\omega_0)|$ represents the spectral amplitude of the appropriate component of ground velocity measured at the dominant tremor frequency ω_0 .

We have applied the above formula to the data from the October 1963 flank eruption of Kilauea volcano using the medium structure of our model number 4 appropriate for the summit region around Kilauea caldera. Measurements of $|\dot{G}_0(\omega_0)|$ were made directly from the data of Fig. 16a at the observed tremor frequency ω_0 . For the dominant frequency of 1.1 Hz, bandwidth of 0.1 Hz and RMS amplitude of 0.3 μm observed for the tremor at the station nearest to the source, and for the range of $0.4 \cdot 10^{-6} \Delta P \Delta l$ to $1.2 \cdot 10^{-6} \Delta P \Delta l$ measured for $|\dot{G}_0(\omega_0)|$, we estimated the product $n^{1/2} \Delta P \Delta l \approx 4\text{-}13 \text{ bar} \cdot \text{m/s}^{1/2}$. Thus, for a channel length of 1 m and total rate of channel openings of 1 per sec, the excess pressure of magma responsible for the jerky opening of individual channels was found in the range of 4 to 13 bars, in good agreement with the far-field estimate obtained from Rayleigh waves by Aki et al. (1977a).

A detailed version of the above study will be submitted to the Journal of Geophysical Research for publication, and a copy of this paper will be sent to D.O.E. under separate mail.

TABLE 1. MODEL PARAMETERS

Case	Crack Stiffness, C	z_1 (KM)	z_2 (KM)	h (KM)	α_1 (KM/S)	β_1 (KM/S)	ρ_1 (G/CM ³)	α_2 (KM/S)	β_2 (KM/S)	ρ_2 (G/CM ³)
1	5	0.5	1.5	-	3	$\sqrt{3}$	2.7	-	-	-
2	0	0.5	1.5	-	3	$\sqrt{3}$	2.7	-	-	-
3	5	0.05	1.05	-	3	$\sqrt{3}$	2.7	-	-	-
4	5	0.5	1.5	0.25	1.7	0.98	2.3	3	$\sqrt{3}$	2.7
5	5	0.5	1.5	0.50	1.7	0.98	2.3	3	$\sqrt{3}$	2.7
6	5	0.5	1.5	0.75	1.7	0.98	2.3	3	$\sqrt{3}$	2.7

LIST OF FIGURES

Fig. 1. Configuration of the source, medium, and receivers used in the computation of the ground motion produced by the sudden opening of the wet and dry cracks. The source is represented by the vertical rectangular plane with the dimensions of 2 by 1 km. The representation of the ground displacement is carried out at 861 locations over an area of 3 by 6 km. Each of these locations is depicted by one grid point in the figure. The separation between stations is 0.15 km in both transverse and longitudinal directions. The heavy lines give the positions of the 7 transverse profiles, spaced 0.5 km apart, which are discussed in the text. The plane at depth h represents the interface of the surficial layer used in our study of cases 4, 5, and 6.

Fig. 2. Normal component of the particle velocity and displacement plotted as a function of time and position on the crack surface for both cases of wet and dry cracks opening in the middle. The crack is 10 grid intervals long but, because of the symmetry of the problem, only 5 positions are depicted on one half of the source. Position 5 corresponds to the point of application of the overpressure ΔP and position 1 is near the crack edge. Velocity, displacement and time are given as dimensionless functions of the crack and medium parameters (see text for details).

Fig. 3. Snapshots of the surface displacements produced by the jerky opening of the wet crack in a homogeneous half-space.

The source is buried at a depth of 0.5 km (see Case 1, Table 1). The three components of displacement u, v, w , shown at the top of the figure are, respectively, longitudinal, transverse and vertical. The snapshots are taken every 0.25 s, starting 0.25 s after the event origin time and ending 2.75 s later, thus covering the first 3 s of the event history. The static displacement resulting from this event is shown at the bottom of the Figure.

Fig. 4. Seismograms of the three components of the ground displacement obtained at 8 stations along the 7 profiles depicted in figure 1 for the case of the wet crack opening in a homogeneous half-space (see Case 1, Table 1). For each component marked at the left and for each profile shown at the top of the figure the stations shown from top to bottom are at the distances of 0., 0.3, 0.6, 0.9, 1.2, 1.8, 2.4 and 3 km from the fault trace. Each individual record starts with the arrival of the P wave from the hypocenter and is approximately 3 s in duration. The time scale is indicated at the bottom. Note that the displacement amplitudes have been normalized by the product $\Delta P \Delta l$ of overpressure times the channel length.

Fig. 5. Polar plots of the ground displacement trajectories in the three orthogonal planes defined by u, v, w . The data shown is that of Figure 4 and the configuration of the plots is the same with the exception that purely vertical, transverse and longitudinal motions are not represented. The scale of the displacements indicated at the bottom is given for the particular value $\Delta P \Delta l = 1 \text{ bar} \cdot \text{m}$.

Fig. 6. Space-frequency spectra of the three components of the ground velocity obtained at 21 stations along the 7 profiles of Figure 2 for the case of the wet crack opening in a homogeneous half-space (see Case 1, Table 1). The profiles are numbered at the left and the components $\ddot{u}, \ddot{v}, \ddot{w}$ are indicated at the top of the figure. The disposition of the plots is shown at the upper left, where Δ stands for the distance along the profile, measured in km from the fault trace, and f is the frequency in Hz. The spectra are representative of the entire record of motion of 3 s duration studied in that case.

Fig. 7. Contour maps of the ground velocity spectra given as a function of distance along the first four profiles of figure 1 for the data depicted in Figure 6. The normalized components of velocity are indicated at the right and individual profiles are shown at the top of the figure.

Fig. 8. The same as Figure 3, for the case of a dry crack suddenly opening in a homogeneous half-space. The source is buried at a depth of 0.5 km (see Case 2, Table 1).

Fig. 9. The same as Figure 4, for the case of a dry crack opening at a depth of 0.5 km in a homogeneous half-space (see Case 2, Table 1). The individual records start with the P-arrival and last approximately 6 s.

Fig. 10. Contour maps of the ground velocity spectra given as a function of distance along the first four profiles of Figure 1 for the case of the dry crack opening in a homogeneous half-space (see Case 2, Table 1). The spectra are representative of the entire record of motion of 6 s

duration in the bandwidth from 0 to 3 Hz.

Fig. 11. The same as Figure 3, for a source depth of 0.05 km (see Case 3, Table 1). Note the steep wavefront of the disturbance in the epicentral area.

Fig. 12. The same as Figure 4, for a source depth of 0.05 km.

Fig. 13. Contour maps of the ground velocity spectra given as a function of distance along the first four profiles of Figure 1 for the case of the wet crack opening at a depth of 0.05 km in a homogeneous half-space (see Case 3, Table 1). The spectra are representative of the entire record of motion of 3 s duration in the bandwidth from 0 to 5 Hz.

Fig. 14a. Snapshots of the surface displacements produced by the jerky opening of the wet crack in a single-layered half-space. The source is buried at a depth of 0.5 km and the layer thickness is 0.25 km (see Case 4, Table 1). The configuration of the displacements and sequence of snapshots is identical to that of Figure 3, with the exception that the static solution is now shown.

Fig. 14b. The same as Figure 14a, for a layer thickness of 0.5 km (see Case 5, Table 1).

Fig. 14c. The same as Figure 14a, for a layer thickness of 0.75 km (see Case 6, Table 1).

Fig. 15a. Seismograms of the three components of the ground displacement obtained at 8 stations along the 7 profiles of Figure 1 for the case of the wet crack opening in a single-layered half-space. The source depth is 0.5 km and the layer thickness is 0.25 km (see Case 4, Table 1). The disposition of the records is the same as that of Figure 4.

Each individual seismogram starts with the arrival of the P wave from the hypocenter and is about 4 s in duration. The displacement amplitudes are normalized by the factor $\Delta P \Delta l$.

Fig. 15b. The same as Figure 15a, for a layer thickness of 0.5 km (see Case 5, Table 1).

Fig. 15c. The same as Figure 15a, for a layer thickness of 0.75 km (see Case 6, Table 1).

Fig. 16a. Contour maps of the ground velocity spectra given as a function of distance along the first four profiles of Figure 1 for the case of the wet crack opening in a single-layered half space. The source depth is 0.5 km and the layer thickness is 0.25 km (see Case 4, Table 1). The disposition of the plots is the same as that of Figure 7. The spectra are representative of the entire record of motion of 4 s duration in the bandwidth of 0 to 5 Hz.

Fig. 16b. The same as Figure 16a, for a layer thickness of 0.5 km (see Case 5, Table 1).

Fig. 16c. The same as Figure 16a, for a layer thickness of 0.75 km (see Case 6, Table 1).

Figure 1

Figure 2

Figure 3

NORMALIZED DISPLACEMENT
AMPLITUDE (BAR⁻¹)
-4 10⁻⁷ -2 10⁻⁷ 0 2 10⁻⁷ 4 10⁻⁷

Figure 4

MOTION IN THE VERTICAL PLANE PERPENDICULAR TO THE FAULT

MOTION IN THE HORIZONTAL PLANE

0.3
0.6
0.9
1.2
1.8
2.4
3.0

DISTANCE ALONG PROFILE (KM)

SCALE
0.54 μ

$\Delta P \Delta l = 1 \text{ BAR} \cdot \text{M}$

Figure 5

Figure 6

1 2 3 4

INTERVAL BETWEEN CONTOURS: $2 \cdot 10^{-8} \text{ BAR}^{-1}$

Figure 7

Figure 8

NORMALIZED DISPLACEMENT
AMPLITUDE (BAR⁻¹)
-2 · 10⁻⁶ -10⁻⁶ 0 10⁻⁶ 2 · 10⁻⁶

Figure 9

DISTANCE ALONG PROFILE (KM)

Figure 10

Figure 11

NORMALIZED DISPLACEMENT
AMPLITUDE (BAR⁻¹)

-2 10⁻⁶ -10⁻⁶ 0 10⁻⁶ 2 10⁻⁶

Figure 12

Figure 13

Figure 14a

Figure 14b

Figure 14c

NORMALIZED DISPLACEMENT
AMPLITUDE (BAR⁻¹)
-10⁻⁶ -0.5 10⁻⁶ 0 0.5 10⁻⁶ 10⁻⁶

0 1 2 3 4
TIME (SEC)

Figure 15a

NORMALIZED DISPLACEMENT
AMPLITUDE (BAR⁻¹)
-10⁻⁴ -0.5⁻⁴ 0 0.5⁻⁴ 10⁻⁴

0 1 2 3 4
TIME (SEC)

Figure 15b

NORMALIZED DISPLACEMENT
AMPLITUDE (BAR⁻¹)
-10⁻⁶ -0.5·10⁻⁶ 0 0.5·10⁻⁶ 10⁻⁶

Figure 15c

DISTANCE ALONG PROFILE (KM)

Figure 16a

Figure 16b

Figure 16c

Low velocity bodies under geothermal areas

At shallow depths, bodies associated with geothermal energy sources may have higher seismic velocity than the surrounding medium because of compositional or petrologic contrasts between magma and the surrounding medium or healing effect of high temperature on cracks which may be responsible for velocity reduction in the surrounding medium. In the lower crust and upper mantle, however, we expect lower velocity for the geothermal body than the surrounding colder region because the velocity-reducing defects, like cracks containing partial melts, may be more abundant in the former.

We shall summarize here the 3-D velocity structure determined by the method of Aki et al. (1977b) using the teleseismic P time data obtained at array stations covering several geothermal areas. Let us start with the smallest one, namely, Coso, California.

(1) Coso, California

Reasenberget al. (1980) determined a 3-D velocity structure under the Coso geothermal area using the teleseismic P delay data from an array of seismographs about 25 km in diameter. They found an intense low-velocity body at the depth range from 5 to 20 km with the maximum velocity contrast of over 8% between 10 and 17.5 km depth beneath the surface expression (domes) of recent rhyolitic volcanism. The body has a width of approximately 5 km and becomes elongated in approximately N-S with increasing depth. The direction of elongation is roughly perpendicular to the axis of least compression in this area obtained from fault plane solutions

of local earthquakes (Weaver and Hill, 1979), and they suggest that the thermal system which resulted in the deposit of rhyolite at the surface originated with the intrusion of basaltic magma into extensional fractures in the lower crust.

(2) Roosevelt Hot Spring, Utah

A similar low-velocity body was found beneath the Roosevelt Hot Spring, Utah by Robinson and Iyer (1979) using the teleseismic P time data from an array with a diameter of about 30 km. At shallow depths (0~5 km), the velocity distribution is dominated by the geology of Basin and Range, with high velocity under the Mineral Mountains and low velocity under the neighboring valleys filled with sediments. The results for depths below 5 km are in sharp contrast to those above. A pipe-like body with a diameter about 15 km and velocity 5~7% lower than the surrounding extends from about 5 km depth down at least as far as the uppermost mantle. The horizontal section of low-velocity body is roughly circular at shallower depths centered near the Roosevelt Hot Springs geothermal area, but become somewhat elongated in the N-S direction at depth. Robinson and Iyer discussed possible causes for the low-velocity body and preferred the abnormally high temperature and presence of a small fraction of molten rock, with acidic composition in the shallower part of the crust and basic composition in the deeper part.

(3) Geysers-Clear Lake, California

Large teleseismic P-time delays were also discovered in the Clear Lake volcanic area, California by Iyer et al (1979). 3-D velocity anomalies have been obtained using the data from an array with a diameter of about 70 km (Oppenheimer, personal communication). An intense low-velocity body extending to the depth of 45 km was found. The velocity contrast is more than 20% for depths from 0 to 15 km, 10% for 15 to 30 km, and 5% for 30 to 45 km. Interestingly, the anomaly has an elongated shape with the direction gradually rotating from NNE-SWW to N-S with increasing depth. The latter direction is consistent with the idea that the regional tectonic stress (with maximum compression in the N-S direction in this area) may be determining the orientation of fractures in the lithosphere through which magma ascends.

(4) Yellowstone, Wyoming

The existence of an enormous low-velocity body under the Yellowstone National Park penetrating deep into the mantle has been known by the work of Iyer and his colleagues at the U.S. Geological Survey (Iyer, 1975). Zandt (1978) determined the 3-D seismic image using their teleseismic P time data obtained from an array of seismographs with diameter about 100 km. The low-velocity body extends from the surface to at least a depth of 100 km. At shallow depths (0-20 km), the horizontal section of the body closely follows the Caldera boundary and has an elliptical shape elongated in the N60°E direction, with length 100 km and width 40 km. The velocity contrast is as much as 20%. The

velocity contrast decreases to about 10% for depths below 20 km. The elongated shape in the N60°E direction persists to the depth of 40 km, below which it becomes roughly circular for the depth range 40 to 70 km, and again elongated in the NEE direction for the depth range 70 to 100 km. There is some indication of low velocity anomaly extending deeper than 100 km down to perhaps 200 km. The velocity contrast, however, is only a few percent and the shape is not clearly defined for depths greater than 100 km.

As shown in Fig. 1, the N60°E direction of the elongation of the low-velocity body agrees with the trend of volcanic zone in the eastern Snake River Plain, where the initiation of volcanic activity is progressively younger toward northeast, the youngest being the Yellowstone Caldera.

By a remarkable coincidence, a large aperture seismic array (LASA) was operated for many years in eastern Montana located precisely on the extension of the Snake River Plain in the N60°E direction about 300 km beyond the northeastern edge of the Yellowstone Caldera. Aki et al (1976) determined a 3-D seismic image from the teleseismic P-time data. They found a zone of low-velocity sandwiched between high-velocity regions also trending in the N60°E direction. The low-velocity zone penetrates from the surface to the depth of about 100 km, and was interpreted as a zone of weakness or shear zone associated with a system of wrench faults present in the Rocky Mountains. One set of these fault trends N60°E, including the Fromberg, Unadilla, Weldon and Brockton-Froid faults which form a 500 km long

discontinuous fracture from a position less than 100 km northeast of Yellowstone, through LASA, nearly to the Canadian border. Thus, there is good evidence that the Snake River Plain is aligned along a zone of weakness trending N60°E that cuts deeply (~100 km) into the lithosphere, suggesting that magma found its way through fractures in the lithosphere.

From the above four cases of geothermal areas, we may say that the shape and extent of low-velocity body determined by the 3-D inversion method gives some constraints on the tectonic framework in which a physical model of each geothermal system may be constructed.

Figure 1

Operation of Event-Recorders in Mt. St. Helens and Newberry Peak with Preliminary results from them.

During the first six months of the year, most of our effort has been devoted to data collection and very little progress has yet been made on data analysis. Data analysis has progressed slowly because of the time expenditure required for servicing the instruments in the field which leaves little time left for analyzing data. Data analysis will progress more rapidly when winter weather forces reduction in field operations.

Data collection has been undertaken in two areas. Our initial objective was to study the crustal structure and seismicity of Newberry Volcano, Oregon. The eruption of Mt. St. Helens in nearby Washington State brought the opportunity to study seismic activity generated by an active Cascade volcano. On April 4, 1980 we placed a digital event recorder on the west side of St. Helens volcano. Our objective was to record earthquakes and harmonic tremor being generated by the newly active mountain. The seismic activity at St. Helens was at such a high level that one three hour digital tape was completely filled with events within twelve hours of operation. We decided to change tapes in the instrument on a weekly basis, thus enabling us to obtain a sample of the seismic activity as the eruption progressed.

We serviced our event recorder at St. Helens on a weekly basis until May 16, 1980. On May 18, the first and largest eruption of the volcano occurred. Our instrument was inaccessible by road after this eruption. We recently

retrieved this instrument by helicopter with the assistance of the U.S. Geological Survey. The tape that we recovered with this instrument contains a record of seismic activity up to within one day of the initial blast. We recently installed another digital event recorder southwest of St. Helens. By maintaining an instrument near the mountain we hope to obtain high quality three component digital recordings of seismic activity generated by Mt. St. Helens.

Figure 1 shows records of two distinct classes of earthquakes we recorded prior to the initial eruption of Mt. St. Helens. Figure 1a is a record of a typical deep (6-15 km) earthquake. The waveform contains frequencies in the range of 1-30 Hz and has clearly discernible P and S arrivals. Figure 1b is the waveform of a magnitude 3.8 earthquake that occurred on April 12, 1980. The waveform of this event contains very little high frequency content. It is difficult to locate an S arrival on this record. Earthquakes of character similar to that in Figure 1b have been found to occur mainly at shallow depths under Mt. St. Helens.

As an initial attempt to learn something about the elastic properties of material composing Mt. St. Helens, we can study the Q of coda using the technique of Aki and Chouet (1975). From their paper, we find that the amplitude, A, of singly scattered waves as a function of time t is of the form

$$A(t) \sim \frac{1}{t} e^{-\frac{\pi f t}{Q}}$$

where f is frequency and Q is quality factor which is inversely proportional to the amount of attenuation of waves. Using the above equation we can find the value of Q for a given frequency f by finding the slope of the $tA(t)$ plotted vs. time. Preliminary results for deep events yield Q of 250 at 13 Hz and 1200 at 22 Hz. These values are lower than found by Aki and Chouet (1975) in Japan and California.

Data collection at Newberry volcano was resumed in late June, 1980. With the fairly regular occurrence of earthquakes at Mt. St. Helens, it was decided to study arrival times of refracted waves from earthquakes originating at Mt. St. Helens in order to determine crustal properties under Newberry volcano. To perform this study, we set one instrument up for the duration of the experiment at a site north of Newberry Caldera. The remaining event recorders are set up in a region containing the caldera and south of the caldera. Each time a Mt. St. Helens earthquake is recorded, the event recorders are relocated to different sites. With this approach we hope to establish a grid in the vicinity of Newberry volcano and at each point in the grid we hope to have an arrival time for at least one St. Helens earthquake. By measuring arrival times across a grid, we hope to be able to locate velocity anomalies that may be associated with the occurrence of thermal anomalies or magma. The occurrence rate of earthquakes of large enough magnitude to be recorded by the event recorders at Newberry volcano has fallen off since our work at Newberry began. We are still optimistic

that there will be enough earthquakes of sufficient magnitude at St. Helens by the end of fall, 1980 to allow us to complete collection of data this year.

As an example of the type of data we hope to collect, Figure 2 shows data collected for one event. Six stations recorded the event. Approximate locations of each station recording the event are shown in Figure 3. We plan to use aerial photos to pinpoint exact locations of our stations and, until we obtain precise locations of stations, we cannot draw any firm conclusions from the records shown in Figure 2.

Figure Captions

1. Typical seismograms recorded at Mt. St. Helens - a) lower seismogram: record of a deep earthquake; b) upper: record of a shallow earthquake.
2. Seismograms recorded from Mt. Hood earthquake of July 7, 1980. Records recorded by event recorders operating at Newberry volcano plotted as a function of distance from the epicenter. Locations for each seismogram shown in Figure 3. Record from furthest station shown at top of figure, closest station on bottom of figure.
3. Location of event recorders recording signals shown in Figure 2. Locations shown by number; 1 is closest station, 6 is furthest from epicenter.

Figure 1

10 SEC.

FILE DAY HOUR MIN. SEC. OFF. GAIN NORM

OFF IS TIME FROM MARK TO BEGIN
SEC. IS FIRST MARK

3. KM.

10 SEC.

Figure 2

Figure 3

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