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

Progress Report
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Abstract

Through several years' involvement in geothermal projects such as the Hot Dry Rock project of LASL and the Magma Tap project of SANDIA, we learned that a multiple approach is essential for exploring a complex geological body such as geothermal system. To put this multiple approach on a sound scientific basis, we have been developing an interpretation theory based on the "fluid-filled crack" model of geothermal system and testing it against a variety of data from various geothermal areas.

During the period covered by the present progress report, we worked on the following subjects: (1) analysis and interpretation of seismic data obtained from the bore holes at the LASL Hot Dry Rock geothermal site. (2) Collection, analysis and interpretation of data on deep volcanic tremors at the USGS Hawaii Volcano Observatory. (3) Development of a new method for determining seismic attenuation at frequencies higher than 1HZ, and initiation of a cooperative work with Mexican seismologists on the attenuation measurements at various geothermal areas in Baja, California using the method. (4) Installation of 3-component digital event-recorders at four stations in the Newberry Peak volcano, Oregon. (5) Developing a computer program for calculating seismic motion generated by the vibration of fluid-filled crack in a layered medium.

We published two papers on seismic attenuation in the lithosphere and on source mechanisms of seismic events in a partially frozen lava lake. A Ph.D thesis was completed on a seismological study of Hot Dry Rock geothermal system. We contributed to the USGS workshop at Taos, New Mexico in October, 1978, by reviewing our work on the Hot Dry Rock and Magma Tap projects. We presented a paper on volcanic tremor and magma transport mechanism at the IAVCEI symposium on Intraplate volcanism held at Hilo, Hawaii in July, 1979.

Scope Of Investigations

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 formed 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. In both studies, our interpretation theory has been built upon the seismic wave generation, transmission, scattering and attenuation in a medium which contains a fluid-filled crack.

Our main conclusion from these studies is that a multiple approach is essential for exploring a complex geological body like a geothermal system. We need both conventional and unconventional techniques, based on both active experiments (using artificial signal sources such as seismic waves generated by buried explosions) and passive experiments (using signals generated by the geothermal system itself or by the natural process outside the system). For example, the lateral extent of magma lens in Kilauea Iki was best defined by the distribution of seismic events occurring in the crust above the lens by stress due to cooling, and by the sudden change in Love

wave phase velocity (Aki et al 1978). The strongest constraint on viscosity and thickness of magma lens, on the other hand, was obtained by measurements of SH-waves from distant earthquakes on the floor of Kilauea Iki.

To put this multiple approach on a sound scientific basis, it is necessary to develop a comprehensive interpretation theory based on a quantitative model of geothermal system and test it against a variety of data from different geothermal areas.

Our "liquid-filled crack" model of geothermal system has proved to be quite versatile and useful for various problems, static or dynamic. It was used not only in the study of the fracture in Fenton Hill and the magma lens in Kilauea Iki, but also enabled us to relate the volcanic tremor data from Kilauea with the rate of magma transport during an eruption (Aki et al, 1977).

In the present reporting period, we continued to work along the same line of approach, with our attention focused on the following subject areas.

- (1) Continued analysis and interpretation of seismic data obtained from the bore holes at the LASL Hot Dry Rock geothermal site. This work was done by M. Fehler as his Ph.D thesis work supervised by the principal investigator.
- (2) Collection, analysis and interpretation of data on deep volcanic tremors at the USGS Hawaii Volcano

Observatory. This work is being done by the principal investigator and C. Jaupart.

- (3) Development of a new method for determining seismic attenuation at frequencies higher than 1HZ, and initiation of a cooperative work (at no cost to our project) with Mexican seismologists on the attenuation measurements at various geothermal areas in Baja California using the method. This work was also done by the principal investigator.
- (4) Installation of 3-component digital event-recorders at four stations in the Newberry Peak volcano, Oregon. This work was done by the principal investigator, co-investigator and M. Fehler who now works at the Oregon State University. H. Berman was responsible for the instrument operation.
- (5) Developing a computer program for calculating seismic motion generated by the vibration of fluid-filled crack in a layered medium. This work was done by the co-investigator.

During the reporting period, M. Fehler completed his Ph.D thesis on the seismological study of Hot Dry Rock geothermal system. The principal investigator published a paper on seismic attenuation, and the co-investigator published a paper on source mechanisms of seismic events in Kilauea Iki. The above three attended a USGS workshop on seismic methods for geothermal energy exploration at Taos, New Mexico in October,

1978 and reported on the Hot Dry Rock and Kilauea Iki experiments. The principal investigator presented a paper at a symposium on seismic attenuation held at the Stanford University in June, 1979. He also presented a paper on volcanic tremor and magma transport mechanism at the IAVCEI symposium on Intraplate Volcanism held at Hilo, Hawaii, in July, 1979.

In the following, we shall briefly describe the work in progress.

Interpretation of seismic data from the Fenton Hill Hot Dry Rock System

The results on these topics are described in detail in M. Fehler's Ph.D thesis entitled "Seismological investigation of the mechanical properties of a Hot Dry Rock Geothermal System", a copy of which has been sent to D.O.E. Here, we shall summarize the highlights of the results.

- (1) A Hot Dry Rock geothermal system provides an excellent field laboratory for study of seismic properties of rock. Because of the abilities to control fluid pressure, obtain core for laboratory studies and introduce new microcracks through heat extraction we have a unique opportunity to compare seismic field measurements with laboratory measurements and theoretical formulations of rock properties.
- (2) A total of 16,800 seismograms in the frequency range 8-15 KHz obtained from two dual-well experiments have been successfully used to probe the detailed structure of the reservoir.
- (3) Velocity decrease due to introduction of microcracks during heat extraction has been observed as shown in Fig. 1. Velocity variations with fluid pressure changes have also been observed. We find that velocity decreases almost linearly with change in temperature from the virgin in situ temperature. Rate of change of velocity with temperature is $1.07 \text{ (km/sec)/}100^{\circ}\text{C}$.

Thus velocity may be used to effectively monitor changes in temperature in the reservoir.

- (4) Using methods developed by Cheng (1978) to model laboratory measurements of velocity, the in situ microcrack spectra have been estimated from field velocity measurements. Microcrack porosity is estimated to be as much as four times the virgin microcrack porosity. Changes in porosity with pressure were computed and found to be approximately one order of magnitude larger than were found from flow experiments indicating that some flow occurs along a path whose channel capacity is little affected by pressure. Those paths are called self-propped fractures since they are open even when outside fluid pressure is not applied.
- (5) Spectral analysis of windowed seismograms revealed that logarithm of spectral amplitude times the time t measured from source origin time decreased nearly linearly with t , when the amplitudes were averaged over many seismograms. Spectral amplitudes of individual seismograms showed no consistent pattern.
- (6) Envelopes of spectral amplitude in unpressurized step 2 were modeled using Sato's (1977) single isotropic scattering model and Q of approximately 465 at 12 KHz was found. At 15 KHz, Q was found to be about 315. Diffusion model of energy propagation adequately explains energy propagation in the highly

fractured region of steps 11 and 16. Values of Q were also estimated for these regions. Diffusion coefficients were poorly constrained.

- (7) Average spectral amplitudes of pressurized steps 2 seismograms decreased as time increased from source firing except for two jumps. The first jump in amplitude is due to the arrival of the S wave. The second jump is probably due to some reflection off the top of the region from which heat is being extracted. The arrival of scattered waves is not apparent in individual seismograms and was not successfully amplified by phase shifting and stacking of seismograms.
- (8) Differences in travel times between pressurized and unpressurized cases were found to be smooth functions of seismic source location at steps 2 and 11 in 1978 and steps 2 and 7 in 1977. Differences in travel times in step 30 in 1977 and step 16 both years showed variation with source depth with a characteristic length of approximately 3 meters. This indicates the existence of heterogeneity that has scale of 3 meters in these regions. Velocity variation in the heterogeneity due to pressurizing the reservoir is approximately 3%.
- (9) Diffraction of P and S waves by finite empty and fluid-filled (zero viscosity) cracks has been studied. Only waves with wavelength on the order of

the crack length were considered.

- (10) Transmission and reflection of P and S waves incident on a fluid layer imbedded between two elastic half spaces have been studied as a function of elastic properties of fluid and solid, fluid viscosity as well as fluid layer thickness.
- (11) Frequency dependence of the attenuation of direct P waves was compared to that of later arriving scattered waves and interpreted in terms of the number and thickness of large fractures present in the geothermal reservoir. Attenuation of scattered waves is due to the anelastic properties of the medium while direct P waves lose energy to anelasticity as well as reflection by the fractures. It was found that at the time of the 1978 experiment two to four fractures, each with thickness of 4 mm in the pressurized case and 2 mm in the unpressurized case, are required to explain the observed low amplitude of high frequency waves. Higher frequencies observed in 1977 mean that fractures were less than 1 mm thick at that time. Vertical extent of the fractures may be as much as 200 m. Orientation of the fracture is either NW-SE or NE-SW as determined from seismic noise generated during hydro-fracturing as well as microseismic events. S wave scattered off the top of the fracture implies a NE-SW fracture

direction while a televiwer placed in the borehole indicated a NW-SE direction.

In addition to these conclusions, we feel that the mechanical properties of the system could be further clarified, if a future study includes the following experiments:

- (1) Repeat of the dual well seismic experiment using lower frequency source. Lower frequencies are less affected by fine scale structure of the medium and may provide a clearer image of the gross structure of the reservoir.
- (2) Collection of three component seismic data at two or more receiver locations so that type of waves may be determined and propagating wavefronts reconstructed.
- (3) Detailed laboratory studies on core samples removed from various depths in the reservoir. Measurement of static strain and velocity as a function of effective pressure could be made to check the in situ measurements as well as find pore aspect ratio spectra of cracks.
- (4) Careful and extensive dual well seismic measurements made in a new reservoir location before hydrofracturing. This is important so that the effects of fracturing and heat extraction on seismograms can be easily and accurately determined.

- (5) Use of two or more stations to record micro-seismic signals generated by hydrofracturing. This would eliminate the present uncertainty in location of seismic sources.
- (6) Perform a diffraction experiment using surface sources and receivers as well as borehole receivers. Use of long period (.1 sec) waves may provide information on the total reservoir in three dimensions, as the long period seismology provides information on the gross features of earthquake fault motion.

Fig. 1

Volcanic tremors and magma transport mechanism

Our "fluid-filled crack" model of volcanic tremors developed in the present project enables a quantitative comparison of the sources of shallow and deep tremors, giving some insight into the magma ascent mechanism under the Kilauea volcano.

Volcanic tremors are continuous vibrations observed at an active volcano. They must be related to the magmatic process in a volcano, but their seismograms are rather complex and the conventional analysis technique used for earthquakes are not applicable to them.

Several attempts have been made to interpret observations on the tremor. For example, the predominant period of tremor was interpreted as an eigen-period of vibrating magma body in a spherical chamber (Kubotera, 1974) or in a cylindrical chamber (Shimozuru, 1961). Those models, however, were not very specific as to what excites the vibration, and never capable of explaining observed amplitude and duration of the tremor.

Our model, on the other hand, will not only explain observed period, amplitude and duration of the tremor, but also open up a new possibility for relating these seismological observations with the amount and rate of magma transport and other geological data observable at the surface [Aki et al. 1977].

The basic idea is to consider magma contained in a crack. An important parameter of our model is the excess pressure

in magma, which causes a jerky extension of crack, setting up vibrations to be observed as tremor. At the same time, magma is transported by filling the extended portion of crack.

Several versions of this basic model have been considered. If the magma supply comes from a large reservoir, the crack size may be expanded as the crack tip advances. This model may explain the period change of seismic noise observed at the Fenton Hill Hot Dry Rock system by Potter and Dennis (1974). On the other hand, the tremor observed during the East Rift zone eruption of Kilauea, such as the 1963 October eruption, seems to be adequately modeled by a chain of cracks, which vibrate as magma is pushed through the channel between consecutive cracks (Aki et al, 1977).

The third version is the model proposed by Weertman (1971). In this model, the difference in pressure gradient between liquid and solid causes excess pressure at one end and deficient pressure at the other. If the liquid has lower density, the upper tip extends and the lower tip closes, resulting in upward movement of magma. In this case, the magnitude of excess pressure is given by the density contrast and the difference in height between the upper and lower tips.

In any case, the period of tremor is determined by the crack size. If the crack size gets bigger, the period becomes longer.

The amplitude of tremor is determined by several factors. Aki et al (1977) showed that the amplitude is proportional to the excess pressure of VP of magma, crack length L , incremental

area ΔS of crack by one jerky extension, and the square root of frequency η of such extensions, assuming statistically independent occurrence of jerky extensions. The product $L\Delta P\Delta S\sqrt{\eta\Delta f}$ (where Δf is the band width of tremor) may be called seismic moment rate.

The rate of magma transport, in our model, is proportional to the width d of crack, incremental areas ΔS and frequency η of jerky extensions.

In the case of 1963 October eruption, we know both the period (1 sec), amplitude and duration (10 hours) of tremor. We also know the amount of magma (25 million cubic meters) transported from the summit reservoir to the eruption site. They give the following constraints on the model parameters

$$L \sim 1-2 \text{ Km} \quad (1)$$

$$L\Delta P\Delta S\sqrt{\eta\Delta f} \sim 10^{19} \text{ dyne cm/s} \quad (2)$$

$$d \Delta S\eta \sim 3 \times 10^8 \text{ cm}^3/\text{s} \quad (3)$$

The crack width d may be estimated from width of dike measured on the surface. According to Swanson et al (1976), a typical width for many observed Hawaiian dikes is 50 cm. Considering that the width in our transport formula may refer to the peripheral part of the crack, we assumed here

$$d \sim 30 \text{ cm} \quad (4)$$

Final constraint comes from the fact that the tremor was a continuous vibration and not a sequence of discrete events.

Then, frequency η must be equal to or greater than the predominant frequency of tremor.

$$\eta \gtrsim 1 \text{ s}^{-1} \quad (5)$$

There are five unknowns (L , ΔP , ΔS , η , d), and five constraints. But, the last one is inequality. If we put $\eta = 1$, we get the following result:

$$\begin{aligned} \Delta P &\sim 20 \text{ bars} \\ \Delta S &\sim 10^3 \text{ m}^2 \end{aligned} \quad (6)$$

If $\eta > 1$, ΔS must be less than the above value, and ΔP slightly larger. The above result, however, is obtained by fitting a single crack model (Aki et al, 1977). For a chain of crack, which was found more appropriate for this eruption, the number of extensions contributing to tremor amplitude become greater than that contributing to magma transport by a factor equal to the number of cracks in a chain. If we take that effect into account, ΔP must be reduced by several times. Thus, our best guess on ΔP may be around 10 bar.

Let us now apply the same model to the deep tremors of Kilauea. Results are preliminary but very exciting.

For many years, the USGS Hawaii Volcano Observatory (HVO) has been routinely reporting data on the volcanic tremors. They are classified into shallow, intermediate and deep tremors under the supervision of R. Koyanagi. Very few people are aware of these observations, and almost nobody outside HVO believe the existence of deep tremor.

The principal investigator visited HVO in July, 1979 and examined 200 records of deep tremors. He was convinced that Koyanagi's identification was correct. These deep tremors are tremors and not a sequence of earthquakes. Earthquakes show usually a change of period with time and distance. Their coda vanishes rather quickly in Hawaii because of high attenuation. These tremors, on the other hand, persists for a long time with the same period and show identical period at all the stations. Fig. 2 shows an example of records at 19 stations of the USGS Hawaii network for 20 seconds from 19:39 local time Dec. 10, 1973.

The main evidence for deep focus is the nearly simultaneous arrival of first motion at widely distributed stations, and roughly uniform amplitude at these stations as shown in Fig. II, there focus can be located when the tremor has a sharp beginning. Most of the deep tremors were found to come from the depth of about 40 km, in an area of 20 x 40 km centered directly at the coast directly south of the Kilauea summit.

A preliminary study of amplitude was made on selected events with known locations. The frequency of tremor varied from 1.5 to 8 Hz. We measured the amplitude at the peak of an event, and evaluated the quantity in equation (2), namely the seismic moment rate, $L\Delta P\Delta S\sqrt{\eta\Delta f}$. We found that the seismic moment rate is approximately proportional to the square of tremor period T .

1973 Dec. 10 1939

If we assume dynamic similarity between large and small crack, we expect that L is proportional to T , ΔS to T^2 , and $\sqrt{\eta \Delta f}$ to T^{-1} . Thus, the seismic moment rate $L \Delta P \Delta S \sqrt{\eta \Delta f}$ should be proportional to T^2 (as observed), if dynamic similarity holds. This dynamic similarity means that the excess pressure ΔP is independent of the crack size.

The value of seismic moment rate $L \Delta P \Delta S \sqrt{\eta \Delta f}$ for the 1963 October eruption was 10^{19} dyne cm/s (Aki et al, 1977) as shown in equation (2). If we choose the deep tremor for the similar crack size, we find that the corresponding value of seismic moment rate is about 6×10^{20} dyne cm/s; 60 times the value for 1963 October events.

The rate of magma transport can be estimated, if we assume that all the magma transported to Kilauea (about 0.1 km^3 per year) are causing deep tremors. The duration of deep tremors, on the other hand, is a little more than 1000 minutes per year according to the listing of tremors prepared by the HVO staff. Thus, the rate of transport during a tremor episode is roughly $15 \times 10^8 \text{ cm}^3/\text{s}$, which is 5 times the rate for the 1963 October eruption. This higher rate of transport is consistent with the higher value of seismic moment rate. If L , ΔS , d , and Δf are common between the shallow and deep tremors, the observed difference in magma transport would mean that η is 5 times greater for the deep tremors. Then, the observed difference in seismic moment rate suggests that the excess pressure may be about 30 times greater for the

deep tremors, or several hundred bars. This gives much higher excess pressure than implied in the Weertman model mentioned earlier. Thus, effects other than gravity may be necessary for the magma transport in the source region of deep tremors.

A remarkable thing about the deep tremor is that their activity has been stationary over many years, unaffected by major eruptions and large earthquakes. This implies that the magma transport at depths around 40 km is very steady.

Attenuation of shear waves in the lithosphere

As reported in last year's progress report (COO-2534-4), a new method was developed by the principal investigator to study the seismic attenuation in the earth's crust and upper mantle. The method was applied to the Kanto region in central Japan, using about 900 local earthquakes occurring at depths from 0 to 150 km, and the result will be published shortly in the Journal of Physics of the Earth and Planetary Interior. A preprint of the paper has been sent to DOE.

Here, we shall briefly summarize the method and result.

The measurement of attenuation high frequency seismic waves is difficult because of complex effects of source function, radiation pattern, heterogeneity along the wave path and local site condition at the observation point which must be removed from the observed record before measuring the attenuation effect.

We eliminate the local site effect by the use of a single station. In the past, several single station methods have been used, but they had to make assumptions on the source spectra of seismic events and on the frequency dependence of attenuation. We make no assumptions, either on the source spectra or on the frequency dependence of attenuation.

The source spectra of individual earthquakes are estimated by the coda method of Aki and Chouet (1975) and Rautian and Khalturin (1978). Because the coda waves are

the sum of scattered waves from heterogeneities distributed in a large region, the source spectra estimated from coda waves represent an average over all radiation directions. Thus, it is necessary to eliminate further the radiation pattern effect for the path connecting a particular source and a particular receiver. This was done by averaging the normalized spectral amplitude of S waves (normalized with respect to the coda source spectra) over many paths sharing the same distance. The averaged normalized spectral amplitude is then plotted as a function of distance. The decay constant in this plot gives the attenuation for each frequency.

Our method can be tested by comparing results from two independent stations. The results for the Kanto region obtained from two stations: Tsukuba and Dodaira (to which most of events studied had distinctly different radiation directions) were essentially the same, confirming the validity of our result. The result was extremely interesting. The attenuation depends strongly on frequency, with Q proportional to f^m in the frequency range from 1 to 25 Hz where m varies from 0.6 to 0.8 depending on the area. A similar frequency dependence was observed for central Asia by Rautian and Khalturin (1978) and for the Kuril region by Fedotov and Boldyrev (1969).

Fig. III and IV show the value of Q^{-1} as a function of frequency for area A and area B+C in central Japan. Area A is the northern part of the Kanto region, where the cold

Pacific plate is subducting, and temperature is considered to be lower than under area B + C, where volcanoes exist and tectonic is more complex because of interaction among three plates; Eurasia, Pacific and Philippine sea. The value of Q^{-1} is, however, quite similar between the two areas.

We show in these figures the value of Q^{-1} for lithosphere estimated from long-period Rayleigh and Love waves, which is much smaller than the extrapolated value from high-frequency. Thus, the Q^{-1} vs frequency curve must have a peak somewhere around 0.5Hz. The resultant curve is quite similar to what is expected for a simple linear relaxation model (Zener, 1949). The relaxation frequency of 0.5Hz may be attributed to the grain-boundary relaxation and/or to the loss by scattering. A quantitative interpretation of observed Q^{-1} vs frequency relation is now contemplated in terms of these two mechanisms.

In order to apply our new powerful method to various geothermal areas, we have initiated a cooperative work with Mexican seismologists at no cost to our project. The principal investigator was invited to the CICESE (Centro de Investigacion Cientifica y de Educacion Superior de Ensenada, B.C.), where he supervised the implementation of our new method to their computer processing routine. The seismology group at CICESE is well equipped with digital data acquisition and processing systems, and is planning to acquire seismic data from two major geothermal areas; namely Cerro Prieto and Los Azufres. The following processing scheme is

designed by the principal investigator, and is currently tested at CICESE using digital data from aftershocks of the Oaxdcd earthquake of 1978.

Fig. 3

Fig. 4

Procedure for analysing seismograms of local earthquakes for attenuation measurements

(1) A few definitions: A given area is divided into several regions designated by "p", where

$$p = 1, 2, \dots, P$$

Station is designated by "i", where

$$i = 1, 2, \dots, I,$$

and Event is designated by "j", where

$$j = 1, 2, \dots, J.$$

(2) All the seismograms are filtered by a band-path filter with band width Δf and center frequency f . The RMS amplitude of the output of filter for station i is written as $A_i^j(f/t)$ at t measured from the origin time of the earthquake j . In particular, the maximum of $A_i^j(f/t)$ for the P-wave group is written as $A_i^j(f/t_p)$ and that for the S-wave group as $A_i^j(f/t_s)$. The response of filter, that is, the ratio of RMS amplitude (at f) to the ground displacement (including the seismograph magnification) is $\text{Mag}(f)$ for the i 'th station.

(3) The simplest subject is the coda waves, because they can be written as

$$A_i^j(f/t) = \text{Site}(\text{coda}, i, f) \cdot \text{Source}(\text{coda}, j, f) p(f/t) \dots (1)$$

where,

-Site (coda, i, f) is the local site effect of station i on coda waves,

-Source (coda, j, f) is the source effect of source j on coda waves, and

-P(f/t) is the coda decay curve.

Usually in the form of $t^{-m} \exp(-\frac{\pi f t}{Q_{\text{coda}}(f)})$ where $Q_{\text{coda}}(f)$ is the Q-factor for coda waves.

P(f/t) may be a more complex function of t.

The important thing here, is that P(f/t) is independent of station and source.

Equation (1) can be written as

$$\log A_i^j(f/t) = \log \text{Site}(\text{coda}, i, f) + \log \text{Source}(\text{coda}, j, f) + \log P(f/t) \dots (2)$$

The assumption of independence of P(f/t) on i or j can be tested by determining log P(f/t) for, say, i = 1 and i = 2 separately, and see if they agree.

From a set of data (i = 1, ..., I and j = 1, ..., J) measured at $t = t_1, t_2, \dots, t_T$, we have I X J X T equations of the form (2). The unknowns, P(f/t), Site (coda, i, f). Source(coda, j, f) are only T + I + J. Therefore, a least squares method can be used if

$$I + J + T < I \times J \times T \quad (3)$$

(4) The above analysis gives the following results:

(a) The calibration curve of coda: P(f/t).

$Q_{\text{coda}}(f)$ is known to be roughly the same as $Q_s(f)$, and

give the general average attenuation property of the area.

- (b) Site (coda,i,f) is known (see Tsujiura, 1978) to be the of site(S,θ,i,f) over various direction θ. Site (S,θ,i,f) is the Site effect on S waves at station i, frequency f when S waves arrive from the direction θ. We write this:

$$\text{Site(coda,i,f)} \sim \langle \text{Site(S,}\theta \text{ i,f)} \rangle_{\theta} \quad (4)$$

- (c) Source(coda,i,f) is known to be the average of Source (S,θ,i,f) over various direction θ. Source(S,θ,i,f) is the Source effect on S waves at station i at frequency f when S waves radiate to the direction θ.

The absolute values of the functions $P(f/t)$, Site (coda,i,f) and Source(coda,j,f) are undetermined because of unknown effects of scattering. However, the dependence of $P(f/t)$ on t, or $Q_{\text{coda}}(f)$ can be uniquely determined: the relative site effects $\text{Site(coda,i,f)}/\text{Site(coda, o,f)}$ to a standard station 0: the relative source effects Source(coda,j,f) to a standard event 0 can be also uniquely determined. These quantities are useful for earthquake engineering micro-zonation, attenuation, and stress drop estimation.

- (5) Next step is to determine $Q_s(f)$ for S-waves averaged for the entire region. We assume that the amplitude of S-waves corrected for filter-seismograph-response has the following form:

$$\frac{A_i(f/t_s)}{\text{Mag}(f)} = \text{Site(S,}\theta_{ij},i,f) \text{ Source(S,}\theta_{ij},j,f) \frac{\exp\left(-\frac{\pi f x_{ij}}{Q_s(f) \cdot \beta}\right)}{x_{ij}} \quad (5)$$

where θ_{ij} indicates the direction between source j and receiver i , x_{ij} the distance between them, $Q_s(f)$ is the Q-factor of S-waves, and β is the velocity S-waves.

Let us divide (5) by

$$\text{Site}(\text{coda}, i, f) \quad \text{Source}(\text{coda}, j, f)$$

which had been obtained from equation (2). Then, we have

$$\begin{aligned} & \frac{A_i^j(f/t_s)}{\text{Mag}(f) \cdot \text{Site}(\text{coda}, i, f) \cdot \text{Source}(\text{coda}, j, f)} \\ &= \frac{\text{Site}(S, \theta_{ij}, i, f)}{\text{Site}(\text{coda}, i, f)} \frac{\text{Source}(S, \theta_{ij}, j, f)}{\text{Source}(\text{coda}, j, f)} \frac{\exp(-\frac{\pi f x_{ij}}{Q_s(f) \beta})}{x_{ij}} \end{aligned} \quad (6)$$

Multiply equation (6) by x_{ij} and taking logarithm, we have

$$\begin{aligned} \log \frac{A_i^j(f/t_s) \cdot x_{ij}}{\text{Mag}(f) \text{Site}(\text{Coda}, i, f) \cdot \text{Source}(\text{coda}, j, f)} &= \\ \log \Omega_{ij} - \frac{\pi f x_{ij}}{Q_s(f) \beta} \end{aligned} \quad (7)$$

We shall assume here that average of $\log \Omega_{ij}$ over many i and j , for which $|x_{ij} - x| < \Delta x$, will smooth out its directional dependence. In other words, after the average, the left-hand side depends on x only through the factor $\frac{\pi f x}{Q_s(f) \beta}$.

With this assumption, we have

$$\begin{aligned} & \langle \log \frac{A_i^j(f/t_s) x_{ij}}{\text{Mag}(f) \cdot \text{Site}(\text{coda}, i, f) \cdot \text{Source}(\text{coda}, j, f)} \rangle_{|x_{ij} - x| < \Delta x} \\ &= \text{const.} - \frac{\pi f x}{Q_s(f) \cdot \beta} \end{aligned} \quad (8)$$

Thus, plotting the left-hand of equation (8) against x , the slope gives $1/Q_s(f)$

(6) Once the $Q_s(f)$ is found for various frequencies, we can obtain the absolute source spectrum for S-waves.

For an unbounded, homogeneous, isotropic medium [see Brune (1970) for correction free surface effect],

$$A_{ij}^j(f/t_s) \exp\left(\frac{\pi x_{ij} f}{Q_s(f) \beta}\right) = \frac{c |\dot{M}_0(f)|}{4\pi \rho \beta^3 x_{ij}} \quad (9)$$

where ρ is the density, $0 < c < 1$ is a geometrical factor, and $|\dot{M}_0(f)|$ is the moment rate spectral density. $|\dot{M}_0(0)|$ gives seismic moment.

It is recommended to collect the smallest, closest events which show the simplest wave form that justify the application of equation (9), and average them to eliminate the directional effect over events sharing a limited range of magnitudes. The plot of $\log |\dot{M}_0(f)|$ vs $\log f$ will give the corner frequency and stree-drop for the average of these simplest earthquakes.

(7) We now combine the above result with the relative source effect $S(\text{coda}, j, f)$ obtained in (4) (c), to derive the scaling law of S-wave source spectra [that is, how the spectra grow with earthquake magnitude, see Aki (1967)]. Choose the event group of previous section as the reference event 0, and the source spectra for event j may be given as

$$\frac{\text{Source}(\text{coda}, j, f)}{\text{Source}(\text{coda}, 0, f)} |\dot{M}_0(f)| \quad (10)$$

Plot the above ratio (10) against f for various regions separately, to find any local, regional variations in the scaling law. The scaling law would give the estimate of stress drop and the scale length of fault zone heterogeneities (Chouet et. al., 1978, Rautian & Khalturin, 1978).

(8) Next, divide the whole area into regions. Assign unknown $Q_s(f)$ value for each region, such as $Q_s(f,p)$ for region p .

Instead of equation (7), formulate the following equation;

$$\log \frac{A_i^j(f/t_s) \cdot x_{ij}}{\text{Mag}(f) \cdot \text{Site}(\text{coda}, i, f) \cdot \text{Source}(\text{coda}, j, f)} =$$

$$\log \Omega_{ij} - \frac{\pi f}{\beta} \sum_p \frac{x_{ij}(p)}{Q_s(f,p)}$$

where, $x_{ij}(p)$ is the length of portion of wave path (between source j and station i) lying in region p . Equation (10) may be solved for $Q_s(f,p)$, by putting

$$\log \Omega_{ij} = + \epsilon_{ij}$$

and minimizing the sum of the squares of random errors ϵ_{ij} in equation (10).

This gives the attenuation of S-waves for various regions separately.

(9) The steps (5) through (8) can be applied to P waves by changing

$$A_i^j(f/t_s) \rightarrow A_i^j(f/t_p)$$

$$\text{Site}(S, \theta_{ij}, i, f) \rightarrow \text{Site}(P, \theta_{ij}, i, f)$$

$$\text{Source}(S, \theta_{ij}, j, f) \rightarrow (P, \theta_{ij}, j, f)$$

$$Q_s(f) \rightarrow Q_p(f)$$

$$\text{and } \beta \rightarrow \alpha.$$

Installation of digital event-recorders at Newberry Peak,
Oregon

The extreme youth and wide extent of volcanism in the Cascade range suggest that important geothermal resources are present. However, surface expressions of hydrothermal activity are weak and sparse. Seismic studies made so far by University of Washington, U.S. geological survey and others have not succeeded yet in delineating the geothermal source volumes.

As an initial step toward application of our multiple seismic methods, we installed 3-components digital event-recorders at four sites in the Newberry Peak volcano as shown in Fig. 5. The Newberry Peak was chosen because of the possibility of a shallow magma chamber suggested from a petrological point (J. Eichelberger, personal communication). The installation was done during the week of July 23, 1979.

The operation has been quite successful. Fig. 6a through 6d show an example of seismograms of a distant earthquake ($M = 4.7 \sim 5.2$, off coast of N. Calif.) recorded at all the stations on Aug. 1, 1979. The high frequency noises in the records were found due to tape mechanism, and are being eliminated. The seismic waves from distant earthquakes sample the deep root of volcano, and should give information about the deep interior of volcano if sufficient data are accumulated covering various incident

Fig. 5

Fig. 6 a

Fig. 6 b

Fig. 6 c

DPO

Vertical

North-South

East-West

Fig. 6 d

directions.

In addition to the seismograms from distant earthquakes, we are interested in volcanic tremors which may be generated by the magmatic process. As a matter of fact, we found a very curious continuous vibration with 1 HZ period repeatedly recorded at the station located inside the Newberry Caldera. We are currently examining the origin of the vibration.

The stations have been maintained by M. Fehler who now works at the Oregon State University.

Synthetic seismograms for arbitrary three-dimensional seismic sources in layered media

We use the discrete wavenumber representation of Bouchon (1979) and Dunkin's (1965) propagator technique to obtain the three components of ground motion near an arbitrary three-dimensional seismic source embedded in a layered medium. We have computed the elastic radiation from both shear and tensile dislocations propagating over rectangular or circular fault planes; more complex source functions, appropriate for vibrating cracks containing vacuum or fluid, have also been used to study the radiation from geothermal sources including volcanic tremors. The following are examples of our numerical results obtained for a variety of sources and structures.

We first tested our computer program with Bouchon's program which was developed for earthquake-type shear dislocation sources. Then, we compared seismic waves from the shear-dislocation source with those from tensile-faults which may be used to model thermal cracking such as the seismic events in Kilauea Iki studied by Aki et al (1978) and Chouet (1979).

We first studied the case of a rectangular fault as considered by Bouchon where the rupture originates along a segment of a line of length W and propagates over a distance L with uniform velocity c and constant slip D

parallel to the direction of propagation. Fig. 7 shows the example of a right-lateral strike-slip fault with length $L = 7$ km and width $W = 9.5$ km; the fault lies in the plane $y = 0$ and the rupture propagates at the velocity $c = 2.2$ km/s in the positive x - direction, starting at $x = 0$. The ground displacements u, v, w generated by this source are computed at a point located 100 m away from the fault trace and 5 km away from the epicenter ($x = 5$; $y = 0.1$) for the two cases where the medium is a simple half-space (Fig. 8a) and where the fault breaks through a surficial layer of sediments with low seismic velocities (Fig. 8b). The parameters of the layer and half-space are referred by the subscripts "1" and "2", respectively, in Fig. 7. The starting and stopping P and S phases from the onset and end of the rupture are indicated by P_0 , S_0 and P_1 , S_1 in Fig. 8; a second subscript "r" indicates a refracted wave in Fig. 8b. The largest amplitude of the ground motion is found in the v -component transverse to the fault; the peak ground motion on this component is amplified about eight times by the presence of sediments. This large SH pulse is the bow wave generated by the passage of the rupture front through the sediments having shear velocity less than the rupture velocity. The above results are in a complete agreement with the results obtained by Bouchon using the same method but different programming techniques.

In the second example (Fig. 9) we consider the case of

$$\alpha_1 = 2.8 \text{ km/s}$$

$$\alpha_2 = 6.0 \text{ km/s}$$

$$\beta_1 = 1.6 \text{ km/s}$$

$$\beta_2 = 3.5 \text{ km/s}$$

$$\rho_1 = 2.2 \text{ gr/cm}^3$$

$$\rho_2 = 2.7 \text{ gr/cm}^3$$

Fig. 7

Fig. 8a,b

$$L = 18.67 \text{ km}$$

$$\alpha = 5.6 \text{ km/s}$$

$$\beta = 3.2 \text{ km/s}$$

Fig. 9

a fault with a dip of 40° , length 18.67 km and width 10 km, buried in a half-space with velocities $\alpha = 5.6$ km/s and $\beta = 3.2$ km/s. The rupture originates at a depth of 13 km, propagates upward with a velocity of 2 km/s and stops 1 km below the surface. A comparison is made in Fig. 10 of the ground displacements obtained for a tensile dislocation, with opening D perpendicular to the fault plane (Fig. 10a) with those derived for a thrust fault (Fig. 10b). The calculation is made for a location above the fault plane at equal distance between the two sides of the fault and 12 km away from the epicenter. Of particular interest is the prominence of the SP wave (the wave radiated at the source as a shear wave and critically reflected at the surface as a P wave) on the horizontal component of displacement due to the thrust fault (Fig. 10b) and the absence of this phase for the tensile fault (Fig. 10a). Fig. 11 shows the case of a tensile fault of length 7 km and width 8.5 km located in the plane $y = 0$ and buried 1.5 km below the free surface of a half-space with wave velocities $\alpha = 6$ km/s and $\beta = 3.5$ km/s. The rupture originates at $x = 0$ and propagates in the positive x -direction with a velocity of 2.2 km/s. The ground displacements for this source (Fig. 12) are calculated at five locations A,B,C,D,E, set along a line transverse to the fault at distance of 0 to 4 km from the fault trace and 5 km away from the epicenter. We note the downward static vertical displacement at the stations A and B closest to the fault.

$\alpha = 6 \text{ km/s}$

$$\beta = 3.5 \text{ km/s}$$

Fig. 11

Fig. 12

In Fig. 13 we consider a circular tensile dislocation of constant amplitude which nucleates at a point and propagates at a constant velocity of 2 km/s until it stops at the final fault radius of 2 km. The fault is set in the plane $x = 0$ and the nucleation point is on the z -axis, 5 km below the surface. The ground displacements at the locations O, A, B, C, D, (Fig. 13) are calculated for the two cases of a fault in a half-space (Fig. 14) and in a layer over a half-space (Fig. 15) with parameters as indicated in Fig. 13. The subscripts "o" (Fig. 14) denote the phases for the nucleation point, and the subscripts "1" and "2" denote stopping phases from the closest and farthest edges of the fault. As in the previous case (Fig. 12) we note the downward static vertical displacement at the stations closest to the source. The small impedance contrast between the layer and half-space (Fig. 13) accounts for the weak amplification of the ground displacements shown in Fig. 9 as compared to those depicted in Fig. 14. The above examples are useful as a basic guide for interpreting the seismograms due to formation of thermal cracks such as observed in Kilauea Iki. However, they are not appropriate to model seismic sources involving fluid as encountered in the geothermal system. As examples of these types of sources we consider the solutions obtained by Aki et al. (1977) for the motion of dry or fluid-filled tensile cracks driven by an excess pressure ΔP . Using the finite-

$$\alpha_1 = 4.7 \text{ km/s}$$

$$\alpha_2 = 5.6 \text{ km/s}$$

$$\beta_1 = 2.7 \text{ km/s}$$

$$\beta_2 = 3.2 \text{ km/s}$$

$$\rho_1 = 2.2 \text{ gr/cm}^3$$

$$\rho_2 = 2.7 \text{ gr/cm}^3$$

Fig. 13

Fig. 14

Fig. 15

difference method they calculated the vibration of a crack generated by the jerky extension of one or both ends of the crack or by the sudden opening of a narrow channel connecting two cracks. A critical parameter which characterizes the effect of the fluid in their model is the crack stiffness factor $b/\mu \cdot L/d$ where b is the bulk modulus of the fluid filling the crack, μ is the rigidity of the rock matrix, and L and d are the crack length and opening, respectively. Examples of the particle velocities and displacements across the crack plane obtained from their finite difference solution to the three crack problems considered are shown in Figs. 16 through 18 for a stiffness factor of 5 and in Figs 19 through 21 for a stiffness factor of 0 (dry crack). Solutions are shown at 10 grid points for a crack extending at one end (Figs. 17 and 20) and at 5 grid points or a half of the crack when the crack extends at both ends (Figs 18 and 21) or for the sudden opening of a channel connecting adjacent cracks (Figs 16 and 19).

The ground displacements generated by a rectangular crack with the source-time function as given in Fig. 16 are shown for 8 stations in Figs. 22 and 23. The double crack has a length of 2 km and width of 1 km and is buried 1.5 km below the free surface of a half-space with velocities $\alpha = 3$ km/s, $\beta = \sqrt{3}$ km/s; the orientation of the fault plane is identical to that previously shown in Fig. 11. The stations are set on the line $x = 1.5$ and at distance of 0, .5,

Fig. 16

Fig. 17

Fig. 18

Fig. 19

Fig. 20

1, 2, 4, 5, 6, 7, and 8 km from the fault trace. The ground displacements calculated at the four closest stations (0 to 4 km) are compared with those derived from a bilateral tensile dislocation nucleating at $x = 1$ (Fig. 24) with the rupture velocity 1.57 km/sec, roughly stimulating the source function depicted in Fig. 16.

The fluid-filled double crack vibrating by a jerking opening of connecting channel shows distinctly different seismic signature from the tensile crack open from the center to the ends by a bilateral rupture propagation. The former (Fig. 22) shows the lack of sharp corner, lack of DC component, and predominance of oscillatory intermediate frequency waves as compared to the latter (Fig. 24). These differences clearly correspond to the observed contrast between volcanic tremors and earthquake-type waves (e.g. Aki et al 1977).

Fig. 21

Fig. 22

Fig. 23

Fig. 24

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