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

PROGRESS REPORT

For Period September 1, 1982 - August 31, 1983

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Abstract

This is a progress report, C00-2534-9, for the period September 1, 1982 to August 31, 1983 on the project entitled "Seismological Investigation of Crack Formation in Hydraulic Rock Fracturing Experiments and in Natural Geothermal Environments."

During the present reporting period, we have accomplished the following work, covering both experimental and theoretical aspects of the project.

(1) Publication of a paper on the operation of a digital seismic network on Mount St. Helens volcano and observations of long period seismic events that originate under the volcano.

(2) Publication of a paper on ground motion near an expanding preexisting crack.

(3) Analysis of spectral character of "tremor" in Long Valley, California, during 1980-1983 (joint work with CDMG).

(4) Interpretation of seismic data from the recent activities in Long Valley, California.

(5) Theoretical study on a source model for the long-period events and volcanic tremor observed at Mount St. Helens.

(6) Calculation of synthetic seismograms for the above tremor source model.

(7) Analysis of the Easter Island tremor records.

(8) Testing and application of the Gaussian beam synthetic method to observations at Mt. St. Helens.

(9) Field testing and development of a high-temperature borehole seismograph.

Scope of Investigation

This program was started as an involvement in two major geothermal projects, namely, the Hot Dry Rock Geothermal Energy Development Project of the Los Alamos National Laboratory and the Magma Tap Project of Sandia National Laboratories. The theory and methods developed for interpretation of various seismic experiments conducted at Fenton Hill, New Mexico and Kilauea Iki, Hawaii, however, found a variety of applications to other geothermal areas and volcanoes and our research has been evolving into what might be called Volcanic Seismology. The list of publications in refereed journals and Ph.D. theses from the work supported by the present contract is shown below and attest to this evolution

In this research program, we are interested in both passive and active seismology. We constructed a set of digital event recorders for collection of high-quality data as used in our study of Mt. St. Helens. Our interpretation methods for observed seismic signals from volcanoes and geothermal areas include the effects of seismic wave generation, transmission, scattering and attenuation in a medium containing fluid-filled cracks.

Our study areas include the hot dry rock fracture system at Fenton Hill, magma lens in Kilauea Iki, deep and shallow volcanic tremor sources under Kilauea, inside Mt. St. Helens, and Long Valley. We obtain data from our own event recorders as well as those collected by other universities and government agencies in the U.S. We are also collecting data from Mexico,

Japan, New Zealand and France through cooperation with universities and research institutions in these countries.

We are also developing a borehole seismograph which can be operated at high temperature.

List of Publications in Refereed Journals and Ph.D. Theses
by the Present Contract

- Chouet, B., Source, scattering and attenuation effects on high frequency seismic waves, PhD thesis, Massachusetts Institute of Technology, 183pp, 1976.
- Aki, K., M. Fehler, and S. Das, Source mechanism of volcanic tremors: Fluid driven crack models and their application to the 1963 Kilauea eruption, J. Volcan. Geotherm. Res., 2, 259-287, 1977.
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- Chouet, B., K. Aki, and M. Tsujiura, Regional variation of the scaling law of earthquake source spectra, Bull. Seis. Soc. Am., 68, 49-79, 1978.
- Fehler, M., and K. Aki, Numerical study of diffraction of plane elastic waves by a finite crack with application to location of a magma lens, Bull. Seis. Soc. Am., 68, 573-598, 1978.
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- Aki, K., Attenuation of shear-waves in the lithosphere for frequencies from 0.05 to 25 Hz, Phys. Earth Plan. Int., 21, 50-60, 1980.
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- Chouet, B., Ground motion in the near field of a fluid-driven crack and its interpretation in the study of shallow volcanic tremor, J. Geophys. Res., 86, 5985-6016, 1981.
- Chouet, B., and K. Aki, Seismic structure and seismicity of the cooling lava lake of Kilauea Iki, Hawaii, J. Volcanol. and Geotherm. Res., 9, 41-56, 1981.
- Fehler, M., Changes in P wave velocity during operation of a hot dry rock geothermal system, J. Geophys. Res., 86, 2925-2928, 1981.
- Zandt, G., Seismic images of the deep structure of the San Andreas fault system, central coast ranges, California, J. Geophys. Res., 86, 5039-5052, 1981.
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- Aki, K., M. Fehler, R.L. Aamodt, J.N. Albright, R.M. Potter, C.M. Pearson, and J.W. Tester, Interpretation of seismic data from hydraulic fracturing experiments at the Fenton Hill, New Mexico, hot dry rock geothermal site, J. Geophys. Res., 87, 936-944, 1982.
- Fehler, M., and B. Chouet, Operation of a digital seismic network on Mount St. Helens volcano and observations of long period seismic events that originate under the volcano, Geophys. Res. Letters, 9, 1017-1020, 1982.
- Fehler, M., Interaction of seismic waves with a viscous liquid layer, Bull. Seis. Soc. Am., 72, 55-72, 1982.
- Fehler, M., Using dual-well seismic measurements to infer the mechanical properties of a hot dry rock geothermal system, J. Geophys. Res., 87, 1982.
- Chouet, B., Free surface displacements in the near field of a tensile crack expanding in three dimensions, J. Geophys. Res., 87, 3868-3872, 1982.
- Wu, R.S., Attenuation of short period seismic waves due to scattering, Geophys. Res. Letters, 9, 9-12, 1982.
- Chouet, B., Ground motion near an expanding preexisting crack, J. Volcanol. and Geotherm. Res., in press, 1983.

Summary of Results in the Present Reporting Period

We shall summarize the results obtained in the present reporting period (Sept. 1, 1982 - Aug. 31, 1983) in the following. First, we shall give a list of titles of papers published in this period about which a summary was already given in our last year's progress report: C00-2534-8. We shall then describe the progress on various aspects of the project made during the past year.

(1) Operation of a digital seismic network on Mount St. Helens volcano and observations of long period seismic events that originate under the volcano (M. Fehler and B. Chouet), Geophysical Research Letters, 9 (1982), 1017-1020.

(2) Ground motion near an expanding preexisting crack (B. Chouet), to appear in the December issue of the Journal of Volcanology and Geothermal Research.

(3) Spectral character of "tremor" in Long Valley, California, during 1980-1983 (C.R. Real, C.H. Cramer, and B. Chouet). This work is in collaboration with the California Division of Mines and Geology (CDMG), and a paper was presented at the annual meeting of the Seismological Society of America, Salt Lake City, May 1983, with the following abstract published in Earthquake Notes, 54, p. 89, 1983.

Occurrences of "spasmodic tremor" at Long Valley, Calif. have been suggested by Ryall (1982). Whether these unusual swarms are actually of volcanic origin, or, instead, represent intense bursts of normal tectonic earthquakes is crucial for assessment of the hazard and remains an open question at this

time. Recent work by Fehler and Chouet (GRL, 1982) at Mount St. Helens, Washington, suggests a spectral discriminant between spasmodic-harmonic tremor (magma-fluid flow resonance) and intense earthquake swarms (stick-slip source). At Mount St. Helens low-frequency events precede eruptions, and these events have spectral peaks near 1-3 Hz due to a fluid resonance mechanism. Stick-slip source events have flat spectra out to corner frequencies of 20-40 Hz. Also, high-frequency events characteristic of brittle fracture are readily distinguishable by bandpass filtering.

Since 1979, the California Division of Mines and Geology has been operating digital event recorders (Sprengnether DR-100's) in the area. Brief high-quality digital samples of several suggested occurrences of spasmodic tremor have been recorded, including the most recent 7 January 1983 M 5.5 swarm, which occurred at the time of writing this abstract. This unique dataset is being analysed for spectral content and wave character, and the results and implications relevant to the hazard as well as insights into tectonic/volcanic processes behind continued earthquake activity at Long Valley will be presented.

(4) Interpretation of seismic data from the recent activities in Long Valley, California (K. Aki). This paper was presented at the 1983 Mammoth Lakes CSDP meeting held on April 26-28, 1983. We shall describe here the arguments and tentative conclusions expressed in this paper in some detail. This paper is not yet considered for publication because the

activities at Long Valley as well as further researches into them are still continuing.

Recently, two groups of seismologists (Julian and Cockerham, 1983; Ekström and Dziewonski, 1983), working with independent data sets, arrived at the same conclusion that three of the four recent largest earthquakes in the Mammoth Lakes region show a source mechanism called the Compensated Linear Vector Dipole (CLVD) which is distinctly different from the usual earthquake source characterized by a Double Couple and was attributed by Ishimoto (1932) to magma intrusion (see Haskell, 1964). Julian and Cockerham attribute the CLVD mechanism to a magma injection along vertical dikes striking N 25° W, which is the approximate strike of numerous faults in the area.

Seismic motion generated by a magma injection along a crack was studied extensively by Chouet (1981; see also C00-2534-6 and C00-2534-7) with a numerical simulation. Using the discrete-wave number method, Chouet calculated synthetic seismograms for stations on a layered halfspace in which magma is forced to flow jerkily through a channel connecting two magma filled cracks by an excess pressure in one of the cracks. The latter model was first considered by Aki et al. (1977; see also C00-2534-2 and C00-2534-3), as a source model of shallow volcanic tremors observed at Kilauea.

An important result from Chouet's numerical experiment is that the pattern of near-field displacements are similar to that of static displacement. For example, the horizontal displacement is directed away from the crack along the line perpendicular to the crack plane but it is directed toward the

crack center along the line above and parallel to the crack. This azimuthal sign reversal was most clearly observed in the nearfield (a few hundred meters from active volcanic centers) of Mt. Aso during the active period of 1930's by Sassa (1935). The horizontal displacement observed by Savage and Clark (1982) in Long Valley also shows a similar reversal of sign compatible with the orientation of hypothesized dikes.

One problem remains, however, that that the P wave first motion at far field should be compressive for any direction from a dike injection source and cannot correspond to CLVD alone but requires an additional explosive element. Chouet's numerical result also gives the compressive first motion to all directions. His result, however, suggests a possibility to get around this problem. He shows that the first motion is very weak and is followed by a sharp dilatational motion in certain directions. It is possible that the true first motions were missed in the noise in certain directions and the CLVD mechanism was obtained. If so, there should be a systematic delay of dilatational arrivals relative to compressional arrivals, which can be checked using the arrival time data.

In the magma injection model used by Chouet (1981), the seismic motion is generated by the opening of a channel by fluid movement, but the fluid in the crack responds only passively to the vibration of the solid part. An important parameter in this model is the crack stiffness factor, $BL/(d)$, where B is the bulk modulus of fluid, L is the crack length, k is the rigidity of solid, and d is the crack width (Fehler and

Aki, 1978 and Aki et al., 1977). If this parameter is much larger than 1, vibration is suppressed. For a thin crack to vibrate, it is necessary to have bubbles in magma to reduce its bulk modulus. In any case, as the synthetic seismograms of Chouet (1981) show, the vibrations do not last longer than a few cycles. Examples of synthetic seismograms are shown in Fig. 1 reproduced from Chouet (1981) for the case of the crack stiffness factor 5.

Originally, Aki et al. (1977) proposed the above model to explain the shallow volcanic (harmonic) tremor at Kilauea, Hawaii, considering the tremor as the excitation of crack vibration by jerky openings of channels by excess pressure in magma. We now realize that the spectral peaks of the tremor are much narrower than the theoretical peak expected for the above mode of crack vibration.

Recently, a consensus has been developing among seismologists working on seismic events in volcanoes that the harmonic tremors are caused by sustained occurrence of so-called "long-period events" or "low-frequency events." An example of such an event from Mount St. Helens (Fehler and Chouet, 1982) is shown in Fig. 2, where the long-period event (lower trace) is compared with the volcano-tectonic earthquake (upper trace). When these long-period events occur frequently, they become harmonic tremors.

Comparing Fig. 1 with Fig. 2, we find that our magma-injection model cannot explain long-lasting vibrations of the observed event.

As described in detail in a later section, we are now developing a new model in which fluid participates actively in vibration rather than passively. In the new model, the narrow spectral peak is attributed to the acoustic resonance of fluid in the magma container. The width of the peak is determined by the loss by viscous attenuation and by seismic radiation. Thus, from the point of this new model, the occurrence of long-period events and harmonic tremor implies the presence of a sufficient volume of liquid magma to allow a high-Q vibration.

From the work described above as item (3), our tentative conclusion about seismic events in Long Valley is that we have not seen long-period events or harmonic tremors there. According to our view, Allan Ryall's "spasmodic tremors" are volcano-tectonic events, or a swarm of microearthquakes occurring in the solid without the active participation of liquid magma. It appears that magma is being injected, as evidenced by earthquakes with the CLVD mechanism, but that it has not yet reached a condition in which magma actively vibrates. If we see long-period events in Long Valley, a prompt warning should be in order, because they appeared 2 days before the recent El Chichon eruption (Havskov et al., 1982) and 50 days before the main eruption of Mt. St. Helens.

Figure 1

0 1 2 3 4 5
TIME (SEC)

BCR Z

↑ 246 5 7 18

0.0103

BCR Z

↑ 246 7 24 45

0.0660

Figure 2

(5) A source model for the long-period events and volcanic tremor observed at Mount. St. Helens (B. Chouet).

We shall present here our first attempt to construct a model of volcanic tremor in which liquid magma participates actively in generating ground vibration. Because of the observed narrowness of the spectral peak(s), as mentioned in the preceding section, we need a resonator containing magma. For a quantitative interpretation of the seismogram, we must include the trigger of vibration as well as the radiation of seismic energy into the surrounding solid. Thus, our model must consist of trigger, resonator, and radiator. For simplicity, we shall assume a simple geometry for these three elements: a hemispherical trigger, a cylindrical resonator, and a circular radiator.

Let us consider a cylindrical coordinate system (r, θ, z) with the origin at the free surface, the z axis being taken positive downward, and assume a vertical pipe of length L extending between the depths z_1 and z_2 ($z_2 > z_1$), its axis coinciding with the z axis. The pipe is shut by a hemispherical cap at the top and by a circular disk at the bottom (see Fig. 3). We assume that the pipe is filled with a liquid characterized by a bulk modulus b lower than the rigidity μ of the surrounding medium. The hemispherical cavity contains a gas which, through the process of flashing, acts as an explosive trigger setting the oscillations of the fluid column. The source that we consider thus consists of three parts, (1) a trigger which applies a pressure transient on the

wall of the hemispherical cavity and into the pipe, (2) a pipe which deforms elastically in response to the oscillation in fluid pressure set up by the trigger, and (3) the disk surface shutting the downstream end of the pipe which yields to the force applied by the fluid column.

The Fourier transformed dilatational potential produced by a sudden excess pressure applied uniformly over the interior surface of a spherical cavity is well known (e.g., Kawasumi and Yosiyama, 1935; Sharpe, 1942; Fu, 1945; Blake, 1952). Using the form given by Harkrider (1964) we obtain for a source located at the depth z_1 the radial horizontal, u , and vertical, w , displacements at the free surface as

$$u(r, \theta, 0; \omega) = -4ik_{\beta}^2 S(\omega) \int_0^{\infty} \frac{k^2 \gamma}{F(k)} e^{-i\gamma z_1} J_1(kr) dk \quad (1a)$$

$$w(r, \theta, 0; \omega) = 2k_{\beta}^2 S(\omega) \int_0^{\infty} \frac{k \ell}{F(k)} e^{-i\gamma z_1} J_0(kr) dk \quad (1b)$$

with

$$S(\omega) = \frac{P(R, \omega) R^3 e^{i(k_{\alpha} R - \delta)}}{4\mu \left[\left(1 - \frac{R^2 k_{\beta}^2}{4} \right)^2 + k_{\alpha}^2 R^2 \right]^{1/2}}$$

$$\delta = \tan^{-1} \frac{k_{\alpha} R}{1 - \frac{R^2 k_{\beta}^2}{4}} \quad (2)$$

$$F(k) = \ell^2 + 4k^2 \nu \gamma$$

$$\ell = 2k^2 - k_{\beta}^2$$

where ω is the frequency, $P(R, \omega)$ is the Fourier transform of

the pressure applied at the cavity wall, R is the cavity radius, r is the epicentral distance from the source, F is the Rayleigh function, and J_0 and J_1 are the Bessel functions of zeroth and first order, respectively. By definition k, ν and k, γ are the r, z components of the wave number vectors $\vec{k}_\alpha$ and $\vec{k}_\beta$ with the relations

$$\begin{aligned} \nu &= (k_\alpha^2 - k^2)^{1/2}, & \text{Im}(\nu) < 0 \\ \gamma &= (k_\beta^2 - k^2)^{1/2}, & \text{Im}(\gamma) < 0 \\ k_\alpha &= \omega/\alpha \\ k_\beta &= \omega/\beta \end{aligned} \tag{3}$$

where α and β represent the compressional and shear velocities of the medium.

Pressure waves in the liquid generated by the explosive trigger will push the wall of the pipe primarily in the horizontal direction. This volumetric source has a seismic moment density tensor (moment per unit area) $\vec{m}$ (Aki and Richards, 1980) of the form (Aki and Richards, 1980)

$$\vec{m} = U_R(z_0, \omega) \begin{pmatrix} \lambda + 2\mu \cos^2 \theta & 2\mu \sin \theta \cos \theta & 0 \\ 2\mu \sin \theta \cos \theta & \lambda + 2\mu \sin^2 \theta & 0 \\ 0 & 0 & \lambda \end{pmatrix} \tag{4}$$

where (R, θ, z_0) are the cylindrical coordinates of a point on the pipe surface, λ is the Lamé constant, and $U_R(z_0, \omega)$ is the

Fourier transform of the radial displacement of the pipe wall, which is assumed to be independent of θ . Using this tensor we obtain the free surface displacements u and w in the form

$$\begin{aligned}
 u(r, \theta, 0; \omega) = & -2iR \int_0^{\infty} \frac{k^2 \gamma}{F(k)} J_1(kr) \cdot \\
 & \cdot \int_{z_1}^{z_2} U_R(z_0, \omega) \left\{ \left[2\left(\frac{\lambda}{\mu} k_{\alpha}^2 + 2k^2\right) \right. \right. \\
 & \left. \left. - \ell [J_0(kR) - \frac{2}{kR} J_1(kR)] e^{-i\gamma z_0} \right\} dz_0 dk
 \end{aligned} \quad (5a)$$

$$\begin{aligned}
 w(r, \theta, 0; \omega) = & R \int_0^{\infty} \frac{k}{F(k)} J_0(kr) \cdot \\
 & \cdot \int_{z_1}^{z_2} U_R(z_0, \omega) \left\{ \ell \left[\left(\frac{\lambda}{\mu} k_{\alpha}^2 + 2k^2\right) J_0(kR) - \frac{4k}{R} J_1(kR) \right] e^{-i\gamma z_0} \right. \\
 & \left. + 4k^2 \nu \gamma [J_0(kR) - \frac{2}{kR} J_1(kR)] e^{-i\gamma z_0} \right\} dz_0 dk
 \end{aligned} \quad (5b)$$

Finally, let us consider the force applied by the liquid on the circular area at the downstream end of the pipe. We obtain in that case the following displacements at the free surface:

$$\begin{aligned}
 u(r, \theta, 0; \omega) = & \frac{RP(z_2, \omega)}{\mu} \int_0^{\infty} \frac{k}{F(k)} J_1(kR) J_1(kr) \cdot \\
 & \cdot (2\nu \gamma e^{-i\gamma z_2} + \ell e^{-i\gamma z_2}) dk
 \end{aligned} \quad (6a)$$

$$\begin{aligned}
 w(r, \theta, 0; \omega) = & \frac{iRP(z_2, \omega)}{\mu} \int_0^{\infty} \frac{\nu}{F(k)} J_1(kR) J_0(kr) \cdot \\
 & \cdot (2\nu \gamma e^{-i\gamma z_2} + \ell e^{-i\gamma z_2}) dk
 \end{aligned} \quad (6b)$$

where $P(z_2, \omega)$ is the Fourier transform of the fluid pressure at the bottom of the pipe.

The horizontal and vertical ground displacements produced by the composite source are the sum of the contributions given by equations (1a), (5a), (6a) and (1b), (5b) and (6b), respectively. The synthesis of the individual spectral components of the source is easily performed by using the discrete wave number method (Bouchon, 1981). The inversion to the time domain is then achieved by using the Fast Fourier transform.

To obtain the radial displacement of the pipe surface $U_R(z_0, \omega)$ and the pressure $P(z_2, \omega)$ at the downstream end of the pipe one must solve the equations of the fluid dynamics. Before proceeding with this problem we shall first compare the seismic and hydraulic energies. The energy radiated into the elastic medium as long-period seismic waves can be estimated by performing the integral of the energy of plane waves radiated from a point source over a spherical surface surrounding the source (Haskell, 1964). For a single force F with time dependence $e^{i\omega t}$ applied at the downstream end of the pipe one obtains for the total energy E radiated per unit time in P waves and S waves the expression

$$E = \frac{\omega^2 F^2}{24\pi\rho\alpha^3} \left(1 + \frac{2\alpha^3}{\beta^3}\right) \quad (7)$$

where ρ is the density of the medium. On the other hand the energy supplied per unit time by the driving hydraulic force can be expressed as

$$E = \frac{F^2}{2\pi R^2 |Z_S|} \cos \phi \quad (8)$$

where $Z_S = |Z_S|e^{i\phi}$ is the radiation impedance expressed by the ratio between the driving pressure and the particle velocity normal to the disk surface in the steady-state driven motion. Equating (7) and (8) we get

$$\cos \phi = \frac{\omega^2 R^2 |Z_S|}{12\rho\alpha^3} \left(1 + \frac{2\alpha^3}{\beta^3}\right) \quad (9)$$

which expresses the ratio between the energy radiated into seismic waves and the energy available in the fluid motion. For a rough approximation one can obtain an estimate of $|Z_S|$ at low frequency by considering the static solution for the deformation of a penny-shaped crack under pressure (Sneddon and Lowengrub, 1969), i.e.,

$$w(r) = \frac{\lambda+2\mu}{\mu(\lambda+\mu)} \frac{\Delta P \sqrt{R^2-r^2}}{\pi} \quad (10)$$

where $w(r)$ denotes the displacement normal to the crack face as a function of the radial distance r , R is the radius of the crack, and ΔP is the pressure applied within the crack.

Assuming $\lambda = \mu$ and averaging over the crack surface we obtain

$$\bar{w} = \frac{R}{\pi\mu} \Delta P \quad (11)$$

which yields

$$|Z_S| = \frac{\pi\mu}{\omega R} \quad (12)$$

Putting (12) into (9) we get

$$\cos \phi = 0.3 \frac{\omega \pi R}{\alpha} \quad (13)$$

Assuming $\alpha \sim 4\text{km/s}$, $R \sim 10\text{m}$ and $\omega \sim 4\pi\text{s}^{-1}$ as typical values representative of our problem we obtain $\cos \phi \sim 10^{-2}$ which shows that the energy lost through radiation at low frequencies is a trivial part of the energy available in the fluid motion.

Performing a similar calculation for the pipe one obtains the total seismic energy radiated per cycle by the equivalent point source as

$$E = \frac{\omega^4 (L \Delta s)^2}{8\pi \rho \alpha^5} \left(\lambda^2 + \frac{8\mu^2}{15} + \frac{4\mu\lambda}{3} + \frac{2\mu^2 \alpha^5}{15\beta^5} \right) \quad (14)$$

The energy supplied per cycle of steady state fluid motion is in that case

$$E = \frac{\omega^2 L (\Delta s)^2 |Z_S|}{4\pi R} \cos \phi \quad (15)$$

where Z_S represent the radiation impedance of the pipe source. Combining (14) and (15) we obtain

$$\cos \phi = \frac{\omega^2 L R}{2\rho \alpha^5 |Z_S|} \left(\lambda^2 + \frac{8\mu^2}{15} + \frac{4\mu\lambda}{3} + \frac{2\mu^2 \alpha^5}{15\beta^5} \right) \quad (16)$$

At low frequency the radial displacement of the pipe wall can be obtained with good accuracy from the static solution. This yields

$$|Z_S| = \frac{2\mu}{\omega R} \quad (17)$$

Putting (17) into (16) and assuming $\lambda = \mu$ we derive

$$\cos \phi = 0.08 L R^2 \left(\frac{\omega}{\beta} \right)^3 \quad (18)$$

Assuming $\beta \sim 2\text{km/s}$, $L \sim 1000\text{m}$, $R \sim 10\text{m}$, and $\omega \sim 4\pi\text{s}^{-1}$ as values representative of our model we obtain $\cos \phi \sim 10^{-3}$ which shows that the energy loss due to the radiation from the pipe itself is smaller than the energy loss due to the radiation from the dead end of the pipe. Since the energy radiated as low-frequency seismic waves is such a trivial part of the available energy in the fluid, we can neglect this loss when solving the hydraulic problem. In other words, the fluid dynamics can be treated separately from the problem of seismic radiation.

The usual approach (e.g., Wylie and Streeter, 1978) to solving the problem of a fluid transient in a conduit is to write the equations for the conservation of momentum and the conservation of mass. For a slender tube having a diameter much smaller than the wavelengths of interest, sound pressure and axial particle displacement are taken to be functions of the axial coordinate and time only. Although some radial motion is induced by the elastic yielding of the conduit wall, the pressure gradients associated with this motion are assumed to be too small to affect the pistonlike motion in the axial direction. With the assumption of slowly varying pressure transients the motion in the radial direction is considered to be in equilibrium with the pressure existing at a given time. The same assumption of low frequencies also ensures the quasi-uniformity of the pressure over distances of several pipe diameters in the axial direction. Within this low frequency limit the relation between pressure and radial wall expansion

can be adequately described by the equations of static elasticity. This approach was taken by Halliwell (1963) to study the problem of the water hammer in pipelines as well as tunnels embedded into a homogeneous, isotropic, elastic full space. Neglecting the effect of gravity, the equation of motion gives

$$\frac{1}{\rho} \frac{\partial p}{\partial x} + \frac{2}{\rho R} \tau + v \frac{\partial v}{\partial x} + \frac{\partial v}{\partial t} = 0 \quad (19)$$

where ρ is the density of the fluid, R is the pipe radius, τ is the wall shear stress, p is the pressure, v is the flow velocity, and x and t denote the axial coordinate and time, respectively. The equation of continuity yields

$$\frac{\partial v}{\partial x} + \frac{1}{\rho a^2} \left(v \frac{\partial p}{\partial x} + \frac{\partial p}{\partial t} \right) = 0 \quad (20)$$

where a is the acoustic velocity of the fluid given by

$$a = \left(\frac{\frac{b}{\rho}}{1 + \frac{b}{\mu}} \right)^{1/2} \quad (21)$$

The continuity and momentum equations form a pair of quasi-linear hyperbolic partial differential equations in terms of two dependent variables, velocity and pressure, and two independent variables, distance along pipe and time. To simplify the analysis we shall neglect the convective terms $v \partial v / \partial x$ and $v \partial p / \partial x$, those terms being typically small in the present application. We shall further assume that the shear

stress at the conduit wall can be roughly described by the wall stress for a Hagen-Poiseuille flow. This provides the relation

$$\tau = \frac{4\eta v}{R} \quad (22)$$

where η is the viscosity of the fluid. Putting (22) into (19) and taking partial derivatives with respect to x and t of both (19) and (20) we obtain the telegraph equations

$$\frac{\partial^2 p}{\partial x^2} = \frac{1}{a^2} \frac{\partial^2 p}{\partial t^2} + \frac{\partial \eta}{\rho a^2 R^2} \frac{\partial p}{\partial t} \quad (23)$$

$$\frac{\partial^2 v}{\partial x^2} = \frac{1}{a^2} \frac{\partial^2 v}{\partial t^2} + \frac{\partial \eta}{\rho a^2 R^2} \frac{\partial v}{\partial t} \quad (24)$$

with the solutions

$$p(x,t) = (A_1 e^{\xi x} + A_2 e^{-\xi x}) e^{st} = P(x) e^{st} \quad (25)$$

$$v(x,t) = -\frac{s}{\rho a^2 \xi} (A_1 e^{\xi x} - A_2 e^{-\xi x}) e^{st} = V(x) e^{st} \quad (26)$$

where A_1 and A_2 are constants and s represents the complex frequency with real and imaginary parts σ and ω , respectively

$$s = \sigma + i\omega \quad (27)$$

in which σ denotes the damping of the fluid oscillation and ω is the circular frequency. The parameter ξ is the complex wave number defined by

$$\xi = \frac{1}{a} \left(s^2 + \frac{8\eta}{\rho R^2} s \right)^{1/2} \quad (28)$$

Equations (25) and (26) can be expressed in terms of the

pressure P_0 and velocity V_0 at the inlet ($x = 0$) of the conduit, i.e.,

$$P(x) = P_0 \cosh \xi x - V_0 \frac{\rho a^2 \xi}{s} \sinh \xi x \quad (29)$$

$$V(x) = -P_0 \frac{s}{\rho a^2 \xi} \sinh \xi x + V_0 \cosh \xi x \quad (30)$$

In keeping with the definition of the radiation impedance introduced earlier we now define the hydraulic impedance $Z(x)$ as the ratio between the acoustic pressure and the flow velocity, that is,

$$Z(x) = \frac{P(x)}{V(x)} \quad (31)$$

This yields at $x = L$:

$$Z_L = \frac{P_L}{V_L} \quad (32)$$

The boundary condition at $x = L$ is given by equating the hydraulic impedance with the radiation impedance, i.e.,

$$Z_L = Z_S \quad (33)$$

Defining the characteristic impedance Z_C as

$$Z_C = \frac{\rho a^2 \xi}{s} \quad (34)$$

we obtain from (29), (30), (31), (32), (33) and (34) the equations

$$P_0 = V_L (Z_S \cosh \xi L + Z_C \sinh \xi L) \quad (35)$$

$$V_0 = V_L \left(\frac{Z_S}{Z_C} \sinh \xi L + \cosh \xi L \right) \quad (36)$$

Using (29), (35) and (36) then yields the expressions for the pressure and velocity of the fluid

$$P(x) = P_0 \left(\cosh \xi x - \frac{Z_S \sinh \xi L + Z_C \cosh \xi L}{Z_S \cosh \xi L + Z_C \sinh \xi L} \sinh \xi x \right)$$

$$= P_0 \left\{ e^{-\xi x} + \sum_{n=1}^{\infty} (-1)^n \Gamma^n \left[e^{-\xi(2nL+x)} - e^{-\xi(2nL-x)} \right] \right\} \quad (37)$$

$$V(x) = \frac{P_0}{Z_C} \left\{ e^{-\xi x} + \sum_{n=1}^{\infty} (-1)^n \Gamma^n \left[e^{-\xi(2nL+x)} + e^{-\xi(2nL-x)} \right] \right\} \quad (38)$$

where

$$\Gamma = \frac{Z_S - Z_C}{Z_S + Z_C} \quad (39)$$

Equations (37) and (38) are the solutions of the hydraulic problem expressed in the Laplace domain. Notice that these equations are completely analogous to a transmission line equation. Let us first consider the case where the fluid viscosity is zero and assume also that the downstream end of the pipe has a strong rigidity, i.e., $Z_S \gg Z_C$. This yields $\xi = s/a$ and $\Gamma = 1$. Assuming a step $E(t)$ in pressure at the pipe inlet we obtain the time domain response as

$$p(x,t) = E\left(t - \frac{x}{a}\right) + E\left(t - \frac{2L-x}{a}\right) - E\left(t - \frac{2L+x}{a}\right) - E\left(t - \frac{4L-x}{a}\right) +$$

$$+ E\left(t - \frac{4L+x}{a}\right) + E\left(t - \frac{6L-x}{a}\right) - \dots \quad (40)$$

where we see that the reflected wave is of equal magnitude and has the same phase as the incident wave. On the other hand the reflected wave has the phase opposite to that of the incident wave when $\Gamma = -1$ ($Z_S \ll Z_C$). Finally, we note that

when the radiation impedance equals the characteristic pipe impedance, $\Gamma = 0$, and the tube behaves as if it were infinitely long.

The modes of this hydraulic source are given by the solutions of the equation

$$Z_S + Z_C \tanh \xi L = 0 \quad (41)$$

Using (12) and (13) and taking $n = 0$ we obtain at low frequencies the approximate relations

$$\begin{aligned} f_n &\approx n \frac{a}{4L} \\ \lambda_n &\approx \frac{4L}{n} \end{aligned} \quad (42)$$

$$\sigma_n \approx \frac{a}{2L} \ln \left[\frac{1 - \left(\frac{2\rho a R f_n}{\mu} \right)^2}{1 + \left(\frac{2\rho a R f_n}{\mu} \right)^2} \right]$$

$$n = 1, 3, 5, \dots$$

where λ_n is the wavelength of the n th mode. The free vibration frequencies are the odd harmonics as expected for a singly open-ended pipe. The value of σ_n is a property of the entire conduit system and is independent of position and time. It is also independent of the amplitude of oscillation. It is, however, a function of frequency since the dissipative nature of the system is dependent upon the relative amplitude of oscillation at the various resistive elements in that system. For the typical values of the parameters $\rho \sim 3 \text{ g/cm}^3$,

$a \sim 2 \text{ km/s}$, $R \sim 10 \text{ m}$, $L \sim 1000 \text{ m}$, $f \sim 2 \text{ Hz}$ and $\mu \sim 10^{11} \text{ dyne/cm}^2$, we obtain $\sigma \sim -0.001$ which confirms that the damping of the fluid transient due to seismic radiation is trivial. Indeed, in the simple pipe model which we have considered the damping of the fluid motion is caused primarily by the energy loss due to viscous drag at the conduit wall.

From (37) we obtain P_L which we introduce into (6a) and (6b). At low frequencies the radial deformation of the pipe is described with good accuracy by the static solution

$$U(x,t) = \frac{R}{2\mu} P(x,t) \quad (43)$$

which allows the solution of equations (5a) and (5b). Space-time plots of the fluid pressure and flow velocity calculated with equations (37) and (38) are shown in Figure 4c for the first cycle of the flow oscillation. The triggering event (fig. 4a) is a pressure transient of the form $P_0 = \Delta P \exp[\ln t - (t/T)] / \exp(\ln T - 1)$ applied at the pipe inlet. The pressure-time history at the downstream dead end of the pipe is depicted in Figure 4b. The model parameters are summarized in Table 1.

(6) Calculation of synthetic seismograms for the tremor source model described in (5) (B. Chouet)

Using the discrete wave number method our object is to obtain a complete representation of the motion of the ground in the near field of the tremor source depicted in Figure 3. Examples of the two components of motion synthesized using equations (1a), (5a), (6a), and (1b), (5b), (6b) are shown in Figure 5. The four traces depicted for each station represent from left to right the contributions of the trigger source, the pipe source, the disk source and the composite source resulting from the superposition of these three elements. The two components of the displacement are indicated at the left and the amplitude and time scale are shown at the extreme left and bottom, respectively. The medium considered is a homogeneous half space with the compressional velocity $\alpha = 4$ km/s and shear velocity $\beta = \alpha/\sqrt{3}$ km/s. The depth z_1 to the pipe inlet is 50 m and the ground displacements are obtained at 13 stations ranging in epicentral distance between 0 and 5 km. The parameters of the source are given in Table 1 and the individual source-time functions are those of Figure 4. The last four rows of traces for both components have been magnified by a factor of 10 to improve their visibility. The records start with the arrival of the P wave from the trigger source and last 15.3 s covering most of the duration of the ground response at each location.

The first motion from the trigger and pipe sources is always outward from the source. Interestingly this

compressional motion is very small on the vertical displacement due to the pipe source close to the epicenter and in fact it becomes practically undetectable in comparison to the strong dilatational pulse which follows the onset of the signal at the epicenter. Since the force applied by the fluid on the dead end of the pipe is oriented downward, the first motion due to this source is dilatational everywhere. Note also that the associated signal is delayed with respect to the other source components by an amount which represents the travel time of the acoustic wave in the fluid over the length of the pipe. Because of this delay the first motion of the composite source is made of the contributions from the trigger and pipe components only and thus is always a compression.

The ground response is dominated by the pipe component in the near field. While the horizontal motion vanishes at the epicenter by symmetry, it grows rapidly with distance showing a pulse-like waveform with smaller superimposed oscillations which reaches a peak amplitude within 100 m of the epicenter. The horizontal displacement due to the disk source displays a slower rate of growth and lesser amplitudes owing to the deeper location and smaller size of this source. The comparatively small volume of the trigger also accounts for its limited contribution to the ground motion in the immediate source region. The vertical component of ground motion, on the other hand, is largest at the epicenter for all sources and decays smoothly with increasing range. The vertical orientation of the force applied on the disk accounts for the

strong signal on this component in the near field compared to the horizontal response. The signal from the disk source indeed dominates the vertical ground response at epicentral distances beyond 100 m. The resulting response consists of harmonic oscillations which mirror the repeated hammering of the fluid at the downstream end of the pipe. No outstanding shear arrival can be recognized in the ground response.

In the intermediate field beyond 2 km the disk component of the source dominates the signal on both components of the ground response. This is naturally understandable as the displacement produced by a single body force has a slower rate of decay with distance than that due to a combination of dipoles. The resulting signal consists of a wave train with the harmonic signature typical of the vertical component of motion at shorter distances. Beyond 3 km this wave train exhibits the retrograde particle motion characteristic of Rayleigh waves and at distances beyond 4 km these surface waves clearly dominate the record (see Fig. 6). No SH waves are observed because the equivalent body forces representing the source are isotropically distributed in the horizontal plane.

The spectral characteristics of the ground response are presented in Figure 7 over the epicentral range from 0 to 500 m. This figure shows contour maps of the amplitude spectral density of the two components of ground displacement plotted as a function of frequency and distance from the source. The contours are obtained by taking the logarithm in base 10 of

the displacements premultiplied by a factor 10^9 and they are marked every decade in amplitude. Both maps exhibit a dominant peak at 0 Hz within 100 m of the source. The peak has a similar amplitude on the two components of motion and occurs at the epicenter in the vertical displacement and 75 m from the epicenter on the horizontal displacement. Several subdominant spectral peaks are observed for each component. The vertical motion shows peaks at frequencies near 0.5, 1.5, 2.5, and 3.5 Hz which correspond to the fundamental mode of vibration of the pipe and subsequent harmonics, respectively. These peaks have maximum amplitude at the epicenter. While the first three appear as a stable feature of the vertical ground spectrum with a smoothly decaying amplitude as range increases, the fourth shows more variability arising from the interference pattern between the pipe and disk components of the source. Interestingly, the fundamental mode of the source does not distinctly stand out on the horizontal component at these short epicentral distances. Similarly the corresponding harmonics near 1.5, 2.5 and 3.5 Hz do not become apparent until the epicentral distance exceeds 100 m. All three harmonics are a stable feature of the horizontal displacement spectrum beyond that range.

For added details Figure 8 shows ground displacement spectra obtained at selected distances from the source. The variability between the spectral signatures of the two components is well demonstrated in the synthetics obtained at the distances of 200 and 5000 m. The dominant peak near

0.5 Hz which is well marked at 200 m on the vertical component has no obvious counterpart on the horizontal displacement. On the contrary the same dominant peak is stronger on the horizontal component of the Rayleigh waves observed at 5000 m than it is on the corresponding vertical displacement. In addition this peak occurs at a slightly different frequency on the two components depending on location. At 5000 m for example the dominant frequency is 0.46 Hz on the horizontal component and 0.52 Hz on the associated vertical motion. Such fluctuation of 10% in the respective positions of the dominant peaks appears to be typical of the result obtained in the range studied.

The effect of the depth of burial of the source is investigated in Figure 9 for a station located at the epicentral distance of 100 m. The pattern of the ground displacements is similar to that observed in Fig. 5. The first motion from the composite source is always a compression which, however, becomes increasingly difficult to distinguish in relation to the strong dilatation following it on the vertical component of motion for source depths exceeding 100 m. The disk component of the source always exceeds the pipe component in amplitude on the vertical displacement. Horizontal motion, on the other hand, is dominated by the pipe component for shallow sources. For source depths exceeding 300 m the signal due to the fluid hammering at the dead end of the pipe dominates the record on both components. With the exception of sources buried at depths shallower than 100 m for

which the horizontal displacement is dominating the ground response, the vertical motion is always larger in amplitude and consists of a well-developed harmonic wave train with a stable dominant period related to the fundamental mode of oscillation of the fluid column. The effect of the trigger is not apparent for sources deeper than 400 m.

Finally in Figure 10 we show the effect of fluid viscosity on the spectra of the two components of ground displacements obtained at a station located at the epicentral distance of 200 m from the source. The source parameters are those of Table 1 and the depth to the pipe inlet is 50 m in that case. Lowering the viscosity has the expected effect of sharpening and amplifying the dominant and subdominant spectral peaks on both components. Note that this sharpening effect is accompanied by a slight shift in the position of some of the peaks. The dominant spectral peak for the vertical displacement, for example, shifts from 0.52 Hz to 0.49 Hz as the viscosity is lowered by a factor of 2. On the other hand, the subdominant peak observed near 2.5 Hz for both components is shifted toward higher frequencies on both components as the viscosity is lowered. The same behavior is observed for the position of the subdominant peak near 0.5 Hz on the horizontal displacement.

We have presented a model of volcanic tremor in which the modes of oscillation of a pipe of fixed dimensions are excited by the motion of the fluid flowing within the pipe in response to a short-term perturbation in pressure. The model shows the

relative importance of the various parts constituting this composite source in the radiated elastic field at near and intermediate distances.

The ground response shows a dominant period which is well marked on the vertical component of motion at all distances from the source. Interestingly this dominant period does not become apparent on the horizontal component until distances exceed 1 km. The dominant period and its associated harmonics are strongly controlled by the length of the pipe considered. Close to the source the pipe component dominates the ground response. At ranges beyond 1 km, however, the amplitude of ground motion is controlled primarily by the magnitude of the force applied by the liquid at the downstream end of the pipe. The peak displacement occurs at the epicenter on the vertical component and within 100 m of the epicenter on the horizontal component. The ground response evolves into a well-developed harmonic train of Rayleigh waves with roughly equal horizontal and vertical amplitudes at ranges beyond 3 km. No clear SV arrival can be detected and no SH wave is present in the signal. The contribution of the trigger to the ground motion is limited to the immediate vicinity of the source. Fluctuations of up to 10% are typical between the dominant periods observed on the two components of displacements. In the simple hydraulic model considered the number of cycles and thus the duration of the signal depends essentially on the ratio of the viscosity to the radius of the pipe. Changing the viscosity does result in a slight shift of the dominant and subdominant frequencies present in the signal.

TABLE 1. SOURCE PARAMETERS

$$\eta = 5 \cdot 10^4 \text{ poise}$$

$$b = 1.217 \cdot 10^{11} \text{ dyne/cm}^2$$

$$\mu = 1.6 \cdot 10^{11} \text{ dyne/cm}^2$$

$$\rho = 2.7 \text{ g/cm}^3$$

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

$$R = 4 \text{ m}$$

$$L = 800 \text{ m}$$

$$\Delta P = 10^6 \text{ dyne/cm}^2$$

$$T = 0.5 \text{ s}$$

LIST OF FIGURES

- Fig. 3 Configuration of the source, medium and receiver used in the computation of the ground motion produced by the excitation of a fluid-filled pipe. The composite source consists of a vertical pipe of radius R and length L capped by a hemisphere and shut by a horizontal disk at the bottom. The pipe is filled with a liquid while the hemispherical cap contains a gas. The depth to the pipe inlet is z_1 and the receiver is located at the epicentral distance r .
- Fig. 4 Pressure-time functions for the trigger (a) and disk (b) sources. c) Space-time dependence of the fluid pressure and flow velocity for the pipe source.
- Fig. 5 Synthetic seismograms of the two components of ground displacement resulting from the excitation of a fluid-filled pipe buried in a homogeneous half space. The depth to the pipe inlet is 50 m and the records are obtained at 13 stations located at increasing epicentral distances as indicated at the right. Four sources are represented. Identified at the top of the figures these are: the hemispherical cap which triggers the motion of the fluid column, the pipe, the disk shutting the bottom of the pipe, and the composite source resulting from the superposition of these three sources. The records start with the arrival of the P wave from the trigger and last 15.3 s. For each component marked at the left the

trace amplitude of the bottom four rows has been magnified by a factor of 10 for improved visibility.

Fig. 6 Polar plots of the particle trajectories in the radial vertical plane defined by u and w . The data shown are taken from the composite source observed at the four most distant stations depicted in Figure 4. The scale of the displacements is indicated at the bottom.

Fig. 7 Contour maps of the spectral amplitudes of the two components of ground displacement obtained in the band from 0 to 4 Hz and in the epicentral range from 0 to 500 m for the source model of Table 1. The plots show the logarithm in base 10 of the displacements magnified by a factor of 10^9 . Contours are marked every decade in amplitude.

Fig. 8 Displacement spectra obtained in the band from 0 to 4 Hz at four stations at the epicentral ranges of 0, 200, 1000 and 5000 m for the source model of Table 1. The spectrum of the horizontal component at the range of 200 m and the spectra of the vertical component at the ranges of 0 and 1000 m have been shifted upward by one decade in amplitude for improved visibility.

Fig. 9 Synthetic seismograms of the two components of ground displacement resulting from the excitation of a fluid-filled pipe buried in a homogeneous half space. The records are obtained at the epicentral distance of 100 m and for source depths ranging between 50 and 1000 m as indicated at the right. The four sources

identified at the top are identical to those depicted in Figure 5. The records start with the arrival of the P wave from the trigger and last 15.3 s.

Fig. 10 Effect of the fluid viscosity on the displacement spectra observed in the band of 0 to 3 Hz at the epicentral range of 200 m. Two values of the viscosity are considered, i.e., $2.5 \cdot 10^4$ and $5 \cdot 10^4$ poise. The other source parameters are those listed in Table 1.

Figure 3

Figure 4

Figure 5

EPICENTRAL DISTANCE (M)

2000

3000

4000

5000

 $0.2 \cdot 10^{-4} \text{ CM}$

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

(7) Analysis of the Easter Island tremor records

(K. Aki)

We have examined the records of tremors obtained at Easter Island by Walker, Woollard, Sutton, La Tourneau and Kausel (HIG report 72-2, prepared for the National Science Foundation under Grant GA-12851) for the possibility that the observed 5-sec period tremors may be associated with the magma ascent process under the mid-ocean ridge. An example of records is shown in Fig. 11.

The result of our analysis, however, is negative. We first examined the possibility that the continuous tremor during June-July, 1970 was a swarm of local earthquakes originating under the East Pacific Rise. This possibility was rejected because:

(a) No dispersion characteristic to earthquake records was recognized in the tremor wave train.

(b) The tremor period (5 sec) is always significantly shorter than the period at which earthquakes show the maximum amplitude.

(c) The EW/NS component ratio was not 1 for some earthquakes, but it was on the average close to 1 for the tremor.

Next, we tested the hypothesis that the tremor may be caused by a deep-seated source such as those observed under Kilauea, Hawaii.

We recognized a very strange change in the period of the tremors. In mid-June, 1970, the predominant period was about

1.2 sec. This increased to 1.8 sec on June 21, changed to 5 sec at about noon on June 22, and remained at 5 sec for the next 10 days. Similar increases in period have been observed during several episodes of the Hawaiian deep volcanic tremor (Aki and Koyanagi, 1981). Since the amplitude of the tremor was so large (0.1 mm), if it originated from a deep source, it could have been observed at distant stations. We checked records from several South American WWSSN stations for the same period. We found 5 sec tremors at GIE (Galapagos), and 7~8 sec tremors at ARE and ANT. Their amplitudes were large, but showed no indication of change in period at noon, June 22 as clearly observed at Easter Island; another negative conclusion.

Recent observations on volcanic tremor at Hawaii, Mount St. Helens and El Chichon, the tremor is usually preceded by the so-called "long-period events" as discussed in the above items (4) and (5) which are discrete events with unclear S phase and period close to the period of tremor. Such long-period events were not recognized in the Easter Island records.

With all these negative results, we are inclined to attribute the tremor to the winds and waves over the ocean.

In view of this negative conclusion, we reexamined the literature which reported very long-period volcanic tremors. The so-called "type 2" tremor observed at Mt. Aso during the eruptions of the early 1930's by Sassa (1935) had periods in the range from 3.5 to 8 sec. By reexamining the manuscript

and contacting the director of the Aso Volcano Observatory (Prof. A. Kubotera), it was confirmed beyond a doubt that Sassa's type 2 tremors are not due to meteorological effects but definitely originated from magmatic processes.

Portions of EISS seismograms for event 18 of Table 1. Note the 5-sec noise.

Figure 11

(8) The Gaussian beam synthetic method: Testing by the reciprocal theorem and application to observations at Mt. St. Helens (R. Nowack and K. Aki)

In order to develop an effective method for interpreting high-frequency seismic waves in complex structures such as volcanoes and geothermal areas, we have pursued the testing and application of the Gaussian beam method recently introduced to seismology by Cerveny, Popov and Psencik (1972).

There are several advantages to using the Gaussian beam method. In relation to the asymptotic ray method, the Gaussian beam method is always finite at caustics. In addition, no prior knowledge of caustic locations is required. Since the Gaussian beam method relies on local parabolic approximations, no global direction of propagation must be assumed as in the standard parabolic approximation. Finally, the Gaussian beam method is comparable in cost to ray methods, and possibly even faster since no two-point ray tracing is done.

Since the Gaussian beam method involves a number of approximations, it is important to test the applicability of the method to each problem. For this purpose, we use the reciprocal theorem for Green's function which is valid for any arbitrary heterogeneous medium (Aki and Richards, 1980). For example, we compare the seismogram for the vertical component displacement at point A due to a vertical point force applied at point B with the seismogram for the vertical component displacement at point B due to a vertical point force applied

at point A. The ray paths involved in the Gaussian beam synthesis are quite different between the above two cases, but the reciprocal theorem requires that the two seismograms should be identical. The discrepancy between the synthetic seismograms for reciprocal cases may then be considered as a measure of error.

Figure 12 shows the 2-D model of the velocity structure which was tentatively adopted for Mt. St. Helens. This model is based on evidence obtained by various workers on Mt. St. Helens (Malone, 1982) as well as on other similar types of volcanoes such as St. Augustine in Alaska (Kienle et al., 1979) and Showa-shinzan in Japan (Hayakawa, 1957). In all these cases, the internal structure of a volcano is depicted as a high-velocity central body with a P velocity of about 4 km/sec covered by lower-velocity surficial layers.

In order to check the accuracy of the computed seismograms, reciprocal seismograms were compared in Figure 13. In case A, a 2-D vertical point force is applied at a point 3.5 km beneath the summit, and the receiver is on the flank of the volcano. In case B, the vertical point force is on the flank of the volcano and the receiver is located 3.5 km beneath the summit. The free surface is neglected here. The vertical component seismograms are shown, for both cases, for frequencies of 5, 10, and 20 Hz. There is a satisfactory agreement in waveform for all three frequencies. The amplitudes agree to within 1% for 20 Hz, 6% for 10 Hz, and 12% for 5 Hz, showing better accuracy for higher frequencies, as expected.

During the summer of 1981, a 9-station array was deployed on the flanks and within the crater of the Mt. St. Helens volcano. This was a joint effort undertaken by Oregon State University, Massachusetts Institute of Technology, and the U.S. Geological Survey (Fehler and Chouet, 1982). Two of the more significant observations of the recorded high-frequency volcanic events were: (1) the arrival times at the crater station were between 0.4 and 0.7 seconds earlier than those at the flank stations, and (2) the amplitudes at the crater station were about 10 times greater than those recorded at the flank stations.

It was first thought that the observed high amplitude and early arrival time could be explained by locating the earthquake foci at the bottom of the hypothesized high-velocity column. Figure 12 shows the ray diagram for a focus located 3.5 km below the summit. The high-velocity column tends to cause energy to diverge from the summit. Vertical component Gaussian beam seismograms for stations located across the volcano from 2.5 km to 7.5 km along the surface are shown in Figure 14 using frequencies of 5, 10, and 20 Hz. The amplitudes at the flank stations appear to be comparable to those at the crater, and the differences in arrival time are within 0.25 seconds.

If the focal depth is moved upward to 1 km below the summit, the seismic signal changes dramatically. The ray diagram for a focus 1 km below the summit is shown in Figure 12. In this case, the ray paths going directly upward are

dense and short, while those toward the flank of the volcano are sparse and take a detour. The resultant seismograms for 5, 10, and 20 Hz are shown in Figure 15. These showed an amplitude at the crater station an order of magnitude greater than at the flank stations, and about 0.5 sec earlier arrival, in agreement with observation.

Thus, the P-velocity structure depicted in Figure 12 and focal depths less than 1 km from the summit can explain the two significant observations.

A similar observation was made for St. Augustine volcano by Laila and Kienle (1982), who found a correlation between the amplitude ratio of summit to flank stations and the focal depth. For focal depths shallower than about 2 km, the summit amplitude was greater by an order of magnitude than the flank amplitude, and for deeper events, the opposite was observed.

In any case, the sensitivity of the seismic signal to the focal depth is strong, due to the heterogeneous structure under a volcano. This result also suggests a promising way of accurately locating volcanic earthquakes using variations in amplitude as well as in arrival time.

MT ST HELENS

STRUCTURE

deep source

3.5 km

beneath summit

shallow source

1.0 km

beneath summit

Figure 12

20 hz

10 hz

5 hz

0 sec 2

Figure 13

Figure 14

Figure 15

(9) Development and field testing of a high-temperature borehole seismograph.

Our borehole seismometer is a twin-horizontal element instrument capable of operating for prolonged periods at 260°C. The seismometer elements use a pendulum to detect horizontal movement. The base of the pendulum provides a vertical force against a fixed piezo-ceramic disc for output. The piezo-ceramic crystals have been greatly improved since the first prototype to provide greater output and directional preference. The crystals are arranged so that maximum output is achieved along a preferred axis. With the two seismometer elements arrayed orthogonal to each other, the direction of the source can be determined from the sign and amplitude of the two signals. The arrangement of the piezo-ceramics also cancels out any vertical movement. Problems such as ringing and poor electrical connections have been eliminated in the several prototypes. Progress is being made on improving the response signal by experimenting with various methods of damping the pendulum.

The clamping mechanism used to fix the instrument to the borehole walls is inertial, relying on the weight of the instrument to wedge itself against the wall. The clamping device consists of three vertical spring-loaded leaves arranged symmetrically around the instrument. Our present winch system allows for a maximum depth of deployment of 50 meters in a 2.25" to 2.50" hole. A hard anodized coating has been applied to the outer casing to solve the problem of galvanic effects

when the instrument is deployed in a water-filled hole.

Improved electrical connections make it easier to service elements while in the field and provide more reliability to the system.

The instrument has been deployed successfully in Long Valley, California in a 50-meter-deep borehole with a maximum temperature of 80°C. Signals were recorded on a strip chart recorder via a pre-amp of weight drops at a distance of 75 meters. No natural events were recorded. The borehole seismometer provided good directional information of the weight drop with maximum temperature of 80°C at 50 meters. Signals from weight drops of 50 lbs. from 8 feet were recorded at distances of 75 meters. A recorder designed to provide information on the direction of the source displayed good directional information during the test.

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