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Progress Report for the Past One Year

1

(May 1, 1990 - April 30, 1991)

Introduction

This is a progress report for the past one year of research (year 4 of 5-year project) under the project titled "CSDP: Seismology of Continental Thermal Regime", in which we proposed to develop seismological interpretation theory and methods applicable to complex structures encountered in continental geothermal areas and apply them to several candidate sites for the Continental Scientific Drilling Project.

The past year continued to be extremely productive following up two major breakthroughs made in the preceding year. One of the breakthroughs was the derivation of an integral equation for time-dependent power spectra, which unified all the existing theories on seismic scattering including the radiative transfer theory for total energy and single-multiple scattering theories based on the ray approach. This unified theory finally lead us to a powerful method for uniquely separating the scattering attenuation from the intrinsic absorption attenuation from the field seismic data. We successfully applied the method to the data from the USGS regional seismic arrays in central California, Long Valley and Island of Hawaii, and obtained convincing results on the scattering Q^{-1} and intrinsic Q^{-1} in these areas for the frequency range from 1 Hz to 20 Hz. The frequency dependence of scattering Q^{-1} is, then, interpreted in terms of random medium with continuous or discrete scatterers.

The other breakthrough was the application of T-matrix formulation to the seismic scattering problem. This approach has been extended to the problem of imaging the geometry of subsurface body using various seismic source-receiver configurations. We solved the above inverse problem for the cases of 2-dimensional cavity, and are currently working on 2-dimensional inclusions with high and low velocity contrast with the surrounding medium.

In addition to the above two main lines of research, we were able to use so-called "T-phase" observed on the Island of Hawaii to map the Q value with a good spatial resolution. The T-phase is seismic waves converted from the acoustic waves propagated through the sofah channel of the ocean. We found that we can eliminate remarkably well the frequency dependent recording site effect from the T-phase amplitude using the amplification factor for coda waves, further confirming the fundamental separability of source, path and site effects for coda waves, and proving the effectiveness of stochastic modeling of high-frequency seismic waves.

Adding the publications and presentations made in the past year to the list for the preceding years, the following table lists 49 papers published, submitted, or presented in the past four years of the present project.

List of Publications and Presentations in the Past Four Years

1. Koyanagi, R., Chouet, B., and K. Aki, The origin of volcanic tremors, Part I: Data from Hawaiian Volcano Observatory 1969-1985, U.S.G.S. Professional paper 1350, ed. R. W. Decker, T. L. Wright and P.H. Stauffer, 1221-1259, 1987.
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Summary of Results

Results contained in the 12 new paper published, submitted or presented in the past year (papers 38 through 49) may be summarized under the following four headings:

- I. Imaging the shapes of buried cavities using waveform data obtained by various seismic-source-receiver configurations.
- II. Separation of scattering Q^{-1} and intrinsic Q^{-1} for crustal S waves in the frequency range from 1 Hz to 20 Hz.
- III. Interpretation of scattering Q^{-1} observed in Long Valley, Island of Hawaii and central California in terms of random media.
- IV. Mapping S-wave Q for the Island of Hawaii using the T-phase from circum Pacific earthquakes.

Let us now summarize major results for each of the above four areas. Details of the results are given in the preprint form if they are referred to in the list of publications shown above as "submitted" or "in press". They are available from the P.I. upon request.

Otherwise, the manuscript is currently in preparation.

I. Imaging the shapes of buried cavities using waveform data collected by various seismic source-receiver configurations.

Starting with the T-matrix method of Waterman (1969, 1976), we have studied the forward problem of seismic scattering from 2-D cavities for incident SH plane waves in papers 28 using a perturbation method. The unknown shape of cavity is expressed by Fourier coefficients of deviation from a circle, and the Fourier coefficients are linearly related to the complex amplitude of scattered wave field. Then, the linear inverse problem was solved for the single-frequency data available for a rather restricted seismic source-receiver configuration.

In order to test the validity of the perturbation method, we first compared the forward solution with that obtained by the well-tested boundary integral method. Figure 1 shows the comparison of scattered wave amplitude for SH waves incident on an ellipsoidal cavity with the ellipticity ($f = \frac{a-b}{a}$, where a and b are major and minor semi-axes, respectively) ranging from 0.1 to 0.4. The incident wave approaches the ellipse (major axis being horizontal) from the lower left side. The wave length is 2π times a , or wave number is $1/a$. The solid line (boundary integral method) agrees well with the dashed one (perturbation method) for f less than about 0.3.

The applicability of perturbation method is also limited in the range of wave number k . Figure 2 shows the comparison between the boundary integral and perturbation method for the cases in which ka varies from 1 to 10. In these cases, the incident waves are vertically from below. The agreement in calculated scattered wave amplitude is always excellent for the forward scattering direction, but good for the back-scattering direction only for ka less than 5.

With these limitations in mind, we studied the inverse problem from the following cases of restricted available data.

Figure 3 shows the case in which the scattered wave data are available only over a part of the circle surrounding the cavity (at distance $R=10a$). The true shape of the cavity is an ellipse as shown by the solid line. A single incident wave with $ka=1$ coming from below is the only seismic source. In this case, if the data are available over the upper right quarter of the cavity, the inversion result gives the image shown by the dashed line in the upper left corner. The image is gradually improved as the data coverage is increased as shown in the inset, and we get a nearly perfect result when the data are available from the whole circle.

A similar result is obtained when the true shape of cavity is more complex than a simple ellipse as shown in Figure 4.

Figure 5, on the other hand, shows the case in which the receiver is distributed only over the angular range from -30° to $+30^\circ$ from the top, but the seismic source increases from one plane wave sources from below, to four of them from below, above and two sides. The image of the cavity shape improves with the increase of seismic source, and is perfect when there are four sources.

A similar result is obtained when the true shape of cavity is more complex than a simple ellipse as shown in Figure 6.

Finally, Figure 7 shows the case in which the receiver is restricted to the angular range from -30° to 30° , and the source is only from below, but different wavelengths are used for the source. The image improves as more sources with different wavelengths are used. Lower two figures show that even for restricted receiver locations, one can recover the cavity shape perfectly if the data are available for three wave numbers ($ka=1,2,3$).

The above examples indicate that idealized models such as a cavity in a homogeneous medium can be used to address some fundamental questions on the necessary data for seismic imaging using waveform. The tomographic inversion using the travel time data requires many ray paths crossing a region in order to produce the image of the region. Such constraints may be relaxed if steady-state-source data (or long-duration

wave-form data) are available. For example, one may not need seismic stations in the Pacific ocean to determine the shape of the core-mantle boundary under there.

In the above examples, we studied the scattering of SH waves by cavities. It is easy to extend the study to the scattering by inclusions. As described in papers 39 and 45, we found that seismic scattering from a soft inclusion and that from a hard inclusion are different in the spectral domain. The soft inclusion shows more sharply peaked spectra corresponding to trapped modes inside the inclusion. It will be interesting to see how these differences are reflected in the inverse problem, for a variety of seismic source-receiver configurations.

Fig. 1

Fig. 2

Fig. 5

True vs. estimated model. ($k_0=1$, $R=10$)

Fig. 3

Fig. 4

Fig. 5

Fig. 6

Fig. 4

Fig. 6

II. Separation of scattering Q^{-1} and intrinsic Q^{-1} for crustal S waves in the frequency range from 1 to 20 Hz.

The starting point for separating scattering Q^{-1} from intrinsic Q^{-1} is the following integral equation derived by Zeng et al. (1990). It gives the seismic wave energy density $E(r,t)$ due to a point source applied at the origin and $t=0$ in a homogeneous unbounded medium characterized by the scattering coefficient η_s and the intrinsic absorption coefficient η_i .

$$E(r,t) = E_{in}(t - \frac{|r-r_0|}{v}) \frac{e^{-\eta|r-r_0|}}{4\pi|r-r_0|^2} + \int_V \eta_s E(r_1,t - \frac{|r_1-r|}{v}) \frac{e^{-\eta|r_1-r|}}{4\pi|r_1-r|^2} dV_1 \quad (1)$$

where v is the wave velocity, $\eta = \eta_s + \eta_i$, and $E_{in}(t)$ describes the incident wave energy density at the source. Zeng et al (1990) showed analytically that the radiative transfer equation used by Wu (1985) and the single and multiple scattering ray approaches used by Aki and Chouet (1975) and Chouet and Gato (1985) among others, are equivalent and can be derived from the above integral equation. Hoshiba (1990) demonstrated the same equivalence by a Monte-Carlo simulation. He further devised a very powerful method for separating the scattering Q^{-1} ($=v\eta_s/\omega$) and the intrinsic Q^{-1} ($=v\eta_i/\omega$). Hoshiba's method is based on plotting the seismic energy integrated over a time interval fixed relative to the arrival of incident wave as a function of the hypocentral distance, as shown in Fig. 8, where the time window is chosen to be 0 to 15 sec from the arrival time of incident wave. Different curves in Figure 8 correspond to different values of albedo B_0 which represents the fraction of scattering attenuation in the total attenuation, namely, η_s/η . The parameter g in the figure is the same as η_s , and Q_i is the intrinsic Q . Figure 9 shows the similar plot as Figure 8, but for a different choice of window, namely, 30 to 100 sec. from the arrival time of incident waves. The curves corresponding to different albedos are much more separated in Figure 9 than in Figure 8, suggesting that the combined use of different time windows may give unique determination of scattering Q^{-1} and intrinsic Q^{-1} .

Figure 10 shows the energy-distance curves for 3 different windows (0 to 15, 15 to 30, and 30 to 45 sec from the arrival of incident waves), which is used to determine the scattering Q and the intrinsic Q in Island of Hawaii, central California and Long Valley (paper 48).

Observed energy density spectra obtained from the U.S. Geological Survey regional networks are normalized to a common source and recording site by the correction method, and plotted in Figure 11 for the frequency band centered at 6 Hz. The different size of symbols used in Figure 11 correspond to different time windows, the largest corresponds to 0 to 15 sec, the intermediate 15 to 30, and the smallest 30 to 45 sec. It is clear that there is a regional variation in the observed energy-distance curve.

In addition to the regional variation, the energy-distance curve depends on frequency. Figure 12 compares the results from Long Valley and Island of Hawaii for 1.5, 3.0 and 6.0 Hz bands. These differences are attributed to the difference in frequency dependent scattering Q^{-1} and intrinsic Q^{-1} as shown in Figure 13.

Figure 13 shows a very consistent picture which emerges from all three regions. First, both scattering Q^{-1} and intrinsic Q^{-1} decrease with increasing frequency, but the former is more strongly frequency dependent. As a result, the scattering dominates over the intrinsic attenuation for lower frequencies and the opposite for higher frequencies. We find also a significant difference in the frequency dependence between the seismic tectonic area (central California) and volcanic area (Long Valley and Hawaii). The scattering Q^{-1} appears to decay more sharply with increasing frequency in the volcanic area than in the seismic tectonic area.

Fig. 8

Fig. 10

Fig. 11

Fig. 12

Fig. 13

III. Interpretation of scattering Q^{-1} observed in Long Valley, Island of Hawaii and central California in terms of random media

In order to explain the frequency dependent scattering Q^{-1} observed for the above three areas, we shall briefly summarize some of theoretical results on the scattering in random media. Benites (1990) (paper 39) and Benites et al. (1990, paper 38) applied the boundary integral method to the scattering in a medium containing numerous randomly distributed cavities. Figure 14 shows a schematic diagram of the scatterer-type waveform with wavelength λ and the radius of cavity a (see below), and calculated seismic waves along the line below the region shown in the lower part of Figure 14. Figure 15 shows the energy loss due to scattering of longer waves and shorter waves as shown in Figure 15. From these figures, the energy loss due to scattering is calculated, and the scattering Q^{-1} is shown as a function of frequency.

Figure 14 shows a schematic diagram of the scatterer-type waveform with wavelength λ and the radius of cavity a (see below), and calculated seismic waves along the line below the region shown in the lower part of Figure 14. Figure 15 shows the energy loss due to scattering of longer waves and shorter waves as shown in Figure 15. From these figures, the energy loss due to scattering is calculated, and the scattering Q^{-1} is shown as a function of frequency. Figure 16 shows the scattering Q^{-1} as a function of frequency. It is peaked when the wavelength is comparable to the scatterer size (more precisely when the wavelength is about π times the cavity diameter). For longer wavelength (lower k), Q^{-1} drops sharply because Q^{-1} is expected to be proportional to k^3 in the region of Rayleigh scattering. On the other hand, for higher k , we expect that the ray theory will apply and the attenuation η_s would be independent of k ($=\omega/v$) or frequency. In that case, the scattering Q^{-1} is expected to be proportional to ω^{-1} . Our result of numerical experiment appears to be in good agreement with this expectation. Since the cavity boundary is a sharp reflector for all frequencies, if the impedance varies gradually from the cavity surface toward the matrix interior, the reflection, and therefore the scattering loss, will be weaker for higher frequencies. Thus, we may expect the decay of Q^{-1} with increasing frequency faster than ω^{-1} for such a medium.

Another approach to interpreting the observed frequency dependence of scattering Q^{-1} is based on the random media characterized by the autocorrelation function. Figure 17,

reproduced from Frankel and Clayton (1986), shows the map view of fluctuation of velocity following two types of autocorrelation function, namely, exponential type and Gaussian (error-function) type. The frequency dependence of scattering Q^{-1} for these two types has been studied by Wu (1982) and Sato (1982) analytically, and by Frankel and Clayton (1986) numerically. Although there are some discrepancies, which we believe are due to the error in the finite difference calculation for large ka , it is clear that the decay of Q^{-1} with increasing frequency is steeper for the Gaussian type media, which shows greater smoothness in the velocity map than the exponential type. It appears that the seismic tectonic area is characterized by the exponential type autocorrelation function corresponding to the Q^{-1} decay proportional to ω^{-1} , while the volcanic area is characterized by the Gaussian type with steeper decay. This may be attributed to the effect of melting or hydrothermal activities which may smooth out the fractal nature of the crustal heterogeneities.

Fig. 14

a

b

Fig. 15

Fig. 16

(from Frankel and Clayton, 1986)

Fig. 17

IV. Mapping S-wave Q for the Island of Hawaii using the T-phase from circum-Pacific earthquakes.

T-phase (Tertiary phase) is the seismic waves recorded on islands or on land near a coast which are propagated through the sofar channel of ocean as acoustic waves and converted to seismic waves when the sofar channel touches the land.

Figure 18 shows examples of T-phase seismograms observed on the Island of Hawaii due to earthquakes in North America. From the arrival time of wave packet measured along the great circle connecting the epicenter and receiver, the wave velocity within the Island of Hawaii is estimated to be 1.8 to 2.0 km/s. This velocity value is similar to the S wave velocity in the shallower part of the Hawaiian crust.

We calculated the amplitude spectra at around 3 Hz from the records at many stations and plotted on a map as shown on the right hand side of Figure 19, separately for the Loma Prieta earthquake, northern California event and the south of Alaska event. The arrow in each figure shows the great circle direction from the epicenter. The amplitude pattern is extremely complex and irregular although there is a trend that the amplitude decrease from northeast to southwest.

This complex irregular pattern of spectral amplitude becomes a remarkably smooth one as shown on the left hand side of Figure 19, when we divided the T-phase amplitude by the site amplification factor at the same frequency band obtained for coda waves of local earthquakes.

The stations (triangles) and earthquake epicenters (solid circles) used for the determination of coda site amplification factor are shown in Figure 20. Typical records of local earthquakes with the indication of time window used for the coda measurement are shown in Figure 21.

The coda amplification factor A_c normalized to the station OTL is shown in Figure 22, together with the location map of station. We found that closer stations in

general show similar frequency dependence, although the absolute level of amplification factor may be quite variable.

The successful elimination of irregular complex pattern from the observed amplitude distribution of T-phase implies that both coda waves and T-phase share the common amplification factor, probably because they are both composed primarily of S-waves. We can then estimate the Q of S-waves from the map of iso-amplitude line shown on the left side of Figure 19. The interval between the neighboring lines is the distance over which wave amplitude is reduced by one half. This interval x is related to Q by

$$\ln 2 = \frac{\pi f x}{Q \beta}$$

where f is frequency in Hz and β is the S-wave velocity. For $\beta = 2$ km/s and $f = 3$ Hz, Q is simply $6.8 x$ (in km), as indicated in each figure. It is interesting to note that all three cases show distinctly low Q region near the Kilauea summit, where Q is about 30. Q is slightly higher along the eastern rift zone (~ 50), but markedly higher in the Mouna Loa area.

Fig. 18

Site Effect-Corrected T-Phase Amplitudes (3 Hz Band)
Loma Prieta Event

Observed T-Phase Amplitudes (3 Hz Band)
Loma Prieta Event

Northern California Event

Northern California Event

South of Alaska Event

South of Alaska Event

Fig. 19

Island of Hawaii

Stations and Earthquakes used for Coda Site Study

Fig. 20

Fig. 21

Fig. 22

V. Progress on VSP and Rock Fracture Studies (Section Prepared by Peter C. Leary)

One of the components of the DOE funded research of K. Aki centers on shallow borehole VSP work to investigate the nature of *in situ* fractures in crystalline rock along the lines of [1; 2; 3]. A program of VSP field studies to be conducted by P. C. Leary was submitted with the original proposal. In a parallel research effort Leary was co-P.I. for seismic studies at the NSF funded Cajon Pass deep drill hole near the San Andreas fault in southern California. It became evident that the Cajon Pass deep drill hole offered an unusually rich opportunity to study crustal fractures at significant depths. In addition to VSP seismic data in the drill hole [3; 4; 5], the Cajon Pass project permitted acquisition of a pair of short surface reflection seismic lines at the drill site [6], provided a complete suite of borehole geophysical logging data to a depth of 3.5 km [7; 8], and allowed examination of fracture fabric preserved in thin sections [9]. These data offered a unique opportunity to study fractures at a wide range of scale lengths over a wide range of fracture intensities. Finally, again in a parallel funding effort, Leary is funded to instrument the Cajon Pass borehole with a three-component seismic sensor to be placed at 3 to 3.5 km to record seismic activity in the immediate vicinity of the borehole and local San Andreas and San Jacinto fault microearthquakes. Data from this sensor allows direct comparison of seismic scattering with measured fracture heterogeneity in the crust.

From the combined Cajon Pass VSP and reflection seismic and logging data sets, we have gained a perspective on crustal fractures that we feel far surpasses in strength and utility the results that we might have gained from conducting the shallow borehole VSP program as originally proposed. The fundamental perspective gained from the Cajon Pass combined data sets is the *scale independence* or *fractal nature* of crustal fractures. This essential feature of crustal fractures was demonstrated using borehole geophysical logging data (sonic velocity and electrical resistivity) that spanned four orders of magnitude in fracture sensitive scale length. The logging data sampled the crust at 15 cm intervals for an uninterrupted interval of 1500 m. Treating the logging data with methods developed for long series of fluctuating physical data, it

was found that fracture fluctuations are scale independent over scale ranges of 1.5 m to 1500 and possess a fractal dimension $D \sim 2.3$ [7; 8]. The wide range of scale lengths on which fractures occur may extend down to the mm scale [8; 9]. With this view of crustal fractures it becomes imperative for seismic studies of the crust to greatly broaden the range of seismic frequencies used to explore the fracture character of a reservoir rock mass. While the traditional VSP seismic passband will return average information about small stress oriented fractures, we now can posit that a large component of the crustal fracture population is due to the clustering or coalescing of microfractures into large scale brittle shear fractures [8]. These larger fracture structures may not contribute to the average fracture fabric of the rock mass as detected by shear wave splitting [1; 2; 3;], but they form the bulk of the connected fluid pathways and must be imaged with transmitted [5] or reflected [4; 6] seismic waves of the highest possible frequencies. These perspectives support the scattering and Q^{-1} investigations of Aki and his students conducted under DOE auspices.

The fractal perspective on the fracture nature of the crystalline crust have been the subject of hour-long talks and subsequent discussions with scientists in industry (Amoco, Schlumberger), in universities (University of Uppsala (Sweden), Harvard University, University of Paris), in the United States (Andy Michael and Tom Hanks) and the British Geological Survey (Stuart Crampin), and with the Hot Dry Rock Project at Los Alamos (Mike Fehler, Bob Potter, Jim Albright).

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Proposed Research for the Next Year

One of the most exciting developments in the past year's project is the separation of scattering Q^{-1} from absorption Q^{-1} (which, according to the word of Dr. Spetzler at the recent workshop on Seismic Imaging at Berkeley, increased the value of attenuation measurement) and the structural implications of the resultant frequency dependent Q^{-1} 's determined for various volcanic and tectonic regions. There are several directions to proceed from here. First, we shall look back and reexamine the method used for the separation. This is important because our interpretation theory is based on the simple assumption that the background medium is homogeneous and unbounded, as expressed in equation (1) of the progress report. We believe that the apparently successful application of our simplified theory to the actual data is due to our particular method in eliminating the recording site effect from the observed seismic energy.

The method used for eliminating the recording site effect is based on the observationally well established fact that the source, path and local site effects on coda waves are easily separable as first recognized by Aki (1969). What is not clear is the

physical region where the local site effect should be attributed to. Our conjecture is that the region of local site effect is the earth overlying the laterally homogeneous part of the earth's crust. By eliminating the effect of this part, then, we may be dealing with the phenomena occurring in the homogeneous part of earth's crust. With regard to the neglect of surface waves, we argue that the attenuation is so strong for surface waves in the frequency range above 1 Hz, that they may not survive in the coda.

We shall address systematically the above fundamental questions regarding the applicability of our method for separating scattering Q^{-1} and intrinsic Q^{-1} . This problem will be attacked by the P.I. with the help of Y. H. Zeng who discovered equation (1) on which our method is based.

Another target of our research in the next year will be the application of our new method to other volcanic and geothermal areas to find if the results on scattering Q^{-1} and intrinsic Q^{-1} obtained from Hawaii and Long Valley are universal. We propose to study the Salton Sea geothermal area in comparison to non-geothermal seismic areas in the neighborhood. This part of the work will be done by S. Koyanagi and K. Mayeda under the supervision of P.I.

The third problem is to improve the interpretation of observed frequency dependent scattering Q^{-1} and intrinsic Q^{-1} , in order to find the physical and structural properties of the earth that are responsible for the observed results. For this purpose, a visiting scientist, L. S. Gao, has brought from Professor Poirier's laboratory, University of Paris, extremely interesting data on the Q^{-1} measured for aluminum samples which are invaded by gallium liquid along grain boundaries.

A preliminary analysis of the data on frequency dependence of Q^{-1} shows that liquid invasion dramatically changes the frequency dependence of Q^{-1} . Figure 23 shows the value of Q^{-1} measured for the Al-Ga sample when both are solid (14°C) as a function of frequency in the range from 100 to 2000 kHz. In this case, Q^{-1} decreases monotonically

with the increasing frequency, approximately inversely proportional to the square root of frequency.

Figure 24 shows the result when the temperature is raised to 31°C at which the liquid gallium spread over aluminum grain boundaries. Q^{-1} shows a peak at 500-600 Hz, and decays at higher frequencies proportional to $f^{-1.6}$.

Dr. Gao has a large amount of data brought from University of Paris, and will continue to analyze them to establish the frequency dependent change in Q^{-1} and interpret it in terms of known structure and material properties of Al-Ga samples.

In addition to the above three areas of research regarding the frequency dependent scattering Q^{-1} and intrinsic Q^{-1} , we shall also proceed with the fundamental study of seismic imaging based on the T-matrix method applied to the waveform data. We shall extend the inversion study for the geometry of buried cavities to buried inclusions. We shall study both hard and soft inclusions, and find out how the data from additional seismic sources and receivers improve the seismic image of inclusions in a manner similar to that used for cavities as explained in the progress report. This work will be done by M. Al-fares and X. Chen under the supervision of P.I.

For the next budget year, we also plan to continue our studies at the Cajon Pass drillhole using the new borehole instrument mentioned above. (We are also awaiting a response on a proposal submitted to DOE, Idaho Falls, to build a major seismic facility at Cajon Pass which would include 4 3-component geophones placed at various intervals in the well.) We will conduct several experiments around the well with our shallow sounding system to research various techniques. We are perfecting our techniques for our work in the new deep hole planned for the island of Hawaii and for the Katmai experiment, as well as future cooperation with the Los Alamos Hot Dry Rock Project.

Fig. 23

Fig 1a

Fig. 24

Fig 2

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