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THE SEISMOLOGY OF GEOTHERMAL REGIMES

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Introduction

This is the final technical report on the research project titled "The Seismology of Geothermal Regimes", in which we have been developing seismological interpretation theory and methods applicable to complex structures encountered in geothermal areas for a better understanding of the earth's geothermal regimes.

The questions we have addressed in our research may be summarized as "what is going on in the earth's crust under tectonically active regions? What are the structures and processes responsible for such activities as earthquakes and volcanic eruptions? How can we capture their essence effectively by means of seismological studies?"

In the past year, the P.I. divided his time between southern California, where his research continues at USC with several Ph.D. students and post-doctoral fellows, and the island of Réunion in the Indian Ocean, where he had access to a wealth of seismological data on an active volcano gathered by a modern monitoring network maintained by the Institute of Physics of Globe in Paris.

First, we found clear evidence for localization of scattered seismic energy in the deep magmatic system of the volcano. The seismic coda of local earthquakes show concentrated energy in the intrusive zones as late as 30 to 40 seconds after the origin time. This offers a very effective method for defining a zone of strong heterogeneity on a regional scale, complementary to the high resolution study using trapped modes as pursued in our past project.

Secondly, we identified about 700 long-period events with various frequencies and durations from the data collected during the past 5 years which included three episodes of eruption. We are applying a finite-element method to the simplest event with the longest period and the shortest duration in order to find the location, geometry and physical properties of their source deterministically. The preliminary result described here suggests that their sources may be a horizontally lying magma-filled crack at a shallow depth under the summit area.

In addition to the above work on the Reunion data, we have continued the theoretical and observational studies of attenuation and scattering of seismic waves. In our theoretical works, we studied seismic scattering in layered media using both radiative transfer and wave-theoretical approaches. In our observational works, we analyzed extensive seismic data now available from the data center of the Southern California Earthquake Center in order to find spatial and temporal variation of coda Q as a geophysical parameter in southern California. The products from these works are included in the following list of publications for the past year.

List of Publications and Presentations

- (1) Koyanagi, S., K. Aki, N. Biswas and K. Mayeda, Inferred attenuation from site effect corrected T-phases recorded on the Island of Hawaii, *PAGEOPH*, **149**, No. 1, 1995.
- (2) Roberts, P. M., K. Aki and M. C. Fehler, A shallow attenuation anomaly inside ring fracture of the Valles caldera, New Mexico, *J. Volcan. Geoth. Res.*, **67**, 79-99, 1995.
- (3) Gao, L. S. and K. Aki, Effect of finite thickness of scattering layer on coda Q of local earthquakes, *J. Geodynamics*, **21**, 191-203, 1996.
- (4) Aki, K., Seismic coda waves: a stochastic process in earth's lithosphere, in *Stochastic Models in Geosystems*, ed. S. A. Molchanov and W. A. Wyczynski, IMA volumes in Mathematics and its Applications, **85**, 1-24, 1996.
- (5) Aki, K., Scale dependence in earthquake phenomena and its relevance to earthquake prediction, *Proc. Natl. Acad. Sci., USA*, **93**, in press, 1996.
- (6) Yomogida, K., K. Aki, and R. Benites, Coda Q for double-layered random media, *Geophys. J. Int.*, in press, 1997.

- (7) Ouyang, H. P., Spatial and temporal characteristics of coda Q as a geophysical parameters in southern California, Ph.D. thesis, University of Southern California, 1996.
- (8) Chen, X., Seismogram synthesis for multi-layered media with irregular interfaces by global generalized reflection/transmission matrices method, III. Theory of 2D P-SV case, *Bull. Seis. Soc. Am.*, **86**, 389-405, 1996.
- (9) Yomogida, K., and R. Benites, Coda Q as a combination of scattering and intrinsic attenuation: Numerical simulations with the boundary integral method, *PAGEOPH*, vol. 148, nos 1/2, 1996.
- (10) Aki, K. and V. Ferrazzini, A snap-shot mapping of localized heterogeneities in the lithosphere using the coda wave data from a sparse seismic array, *Geophys. Res. Letters*, in press, 1997.
- (11) Chen, J., and K. Aki, A finite element simulation of volcanic long-period events observed at Piton de la Fournaise, to be presented at the annual meeting of the Seismological Society of America, Hawaii, April, 1997.
- (12) Aki, K., the use of long-period events for defining the magma system under the Piton de la Fournaise, to be presented at the annual meeting of the Seismological Society of America, Hawaii, April, 1997.

Summary of Results

(1) Inferred attenuation from site-effect corrected T-phase recorded on the Island of Hawaii

The amplitude of T-phase from the circum Pacific earthquakes as recorded at the seismic network of the Island of Hawaii shows very irregular spatial distribution except that it decreases in general away from the epicenter. We found that the application of the coda amplification factor to the T-phase amplitude remarkably smooth out the irregularity throughout the island, and the resultant contours of equal amplitude enables a reliable local estimation of Q . The estimated Q is low at the Kilauea summit area ($Q=30$), and also along the rift zone ($Q=60$), but increases to 150 outside these areas.

A similar remarkable reduction of irregular amplitude distribution using the coda amplification factor was also found for T-phases observed at the Réunion island from mid-ocean ridge earthquakes as described in the 1995 Annual Report of the Observatoire du Piton de la Fournaise.

(2) A shallow attenuating anomaly inside the ring fracture of the Valles caldera, New Mexico

Spectral ratios of teleseismic direct and scattered P waves observed in the Valles Caldera, New Mexico, show a systematic pattern of low amplitudes at sites inside the caldera relative to sites on or outside the ring fracture. Waveforms recorded at caldera stations are considerably more complex than those recorded outside the caldera. The data used in this study were collected during a passive seismic monitoring experiment conducted in 1987. Twenty-four teleseismic events were recorded on two linear arrays spanning the caldera. To first order, the pattern of low amplitudes did not vary with source incidence angle or azimuth of approach, and could not be explained by anomalous amplification at the ring fracture. This observation suggests the presence of a shallow, attenuating zone associated with the caldera fill material inside of the ring fracture. We estimated the general features of the calderas near-surface structure for the two-dimensional vertical cross section beneath the array, using a modification of the Aki-Larner discrete-wavenumber method to forward model the observed amplitude variations. Our results indicate that the caldera fill material must be subdivided into at least two distinct zones: a strongly attenuating lower zone, extending to depths in excess of 4 km, and a mildly attenuating surface layer. To fit the data we had to assign an unrealistically low value to seismic Q in the deeper attenuating anomaly. We attribute this to the inability of the Aki-Larner method to account for strong re-direction of energy away from the caldera due to local heterogeneity that we could not include within the low- Q anomaly. This interpretation is consistent with the pervasive, fractured hydrothermal system that is known to exist in the caldera fill material.

(3) Effect of finite thickness of scattering layer on coda Q of local earthquakes

Theories based on the assumption of an infinite medium show that the coda attenuation factor, coda Q^{-1} , of local earthquakes should be very close to the intrinsic Q^{-1} , whereas many observations show the coda Q^{-1} is close to the total Q^{-1} . We show by numerical modeling that coda Q^{-1} depends on the thickness of the scattering layer. For source-receiver coincidence, if the thickness is greater than a half of the mean free path, coda Q^{-1} lies between intrinsic Q^{-1} and total Q^{-1} . However, when the thickness is less, coda Q^{-1} may be higher than total Q^{-1} . In addition, coda Q^{-1} depends on the epicentral distance. The thinner the layer, the faster the change with distance in coda Q^{-1} . For an unbounded scattering medium coda Q^{-1} is even lower than intrinsic Q^{-1} at long epicentral distances.

(4) Seismic Coda Waves: A stochastic Process in Earth's Lithosphere

Reviewing observational and theoretical studies of coda waves of local earthquakes made in the last quarter century, we reached the following conclusions.

The coda waves are a powerful tool for seismologists because of the simple separability of the effects of source, path and recording site, and have been used for a variety of practical applications such as the mapping of frequency dependent site amplification factor, discrimination of quarry blasts from earthquakes, single station method for determining frequency-dependent attenuation, and normalizing the regional seismic network data to a common source and receiver site condition.

The definition of coda Q^{-1} in terms of the single-scattering model proved to be useful because of its simplicity. It has become, however, a necessary routine to specify the time window for each measured value of coda Q^{-1} to avoid the effect of multiple scattering.

In order to clarify the physical meaning of coda Q^{-1} , attempts have been made to separately determine the scattering loss and the intrinsic absorption by the use of theories including multiple scattering. Although there is a need for further improving the interpretation theories, we found empirically that coda Q^{-1} are rather narrowly bounded between the intrinsic Q^{-1} and the total Q^{-1} .

We found that the coda Q^{-1} measured from a time window represents the seismic attenuation property of the earth's crust averaged over the volume traversed by the singly back-scattered S-waves. If the time window is longer and later, the resultant coda Q^{-1} map will have poorer spatial resolution.

A map of coda Q^{-1} in the mainland China made with spatial resolution of about 200 km shows a strong correlation with the location of large historic earthquakes ($M > 7$). The low Q regions are full of them, and the high Q regions are devoid of them. There is some evidence for the migration of a low Q zone associated with the high seismicity in North China during the past 30 years.

A characteristic pattern of coda Q^{-1} change in time sometimes occurs before a major earthquake, but we found that it is not a reliable precursor. Instead, the highest correlation (more than 0.8) with seismicity was found between coda Q^{-1} and the fraction of seismicity in a narrow magnitude range $M_c < M < M_c + 0.5$, where M_c is characteristic to a given seismic region, for example, 3.0 for southern California and 4.0 for central California. This correlation was predicted by the creep model proposed by Jin and Aki (1989) for explaining the erratic (sometimes positive and sometimes negative) correlation between coda Q^{-1} and b-value. The creep model hypothesizes that the creep in the ductile part of lithosphere occurs over fractures with a unique size corresponding to M_c . High creep activity may increase coda Q^{-1} and enhance seismicity with magnitude around M_c because of stress concentration in the brittle part. This model is consistent with the observation that coda Q^{-1} shows a minimum value during the period of high aftershock activity, and sometimes shows a peak during the quiescence. Another support for the deeper (ductile rather than brittle part of lithosphere) source of the coda Q^{-1} in southern California during 1986-87 and the simultaneous increase in electrical conductivity in the same region, which is attributed to the lower crust.

In summary, observations suggest that the temporal change in coda Q^{-1} may reflect the degree of creep fractures in the ductile part of lithosphere.

(5) Scale-dependence in Earthquake Phenomena and its Relevance to earthquake Prediction.

The recent discovery of a low-velocity, low-Q zone with a width of 50-200 m reaching to the top of the ductile part of the crust, by observations on seismic guided waves trapped in the fault zone of the Landers earthquake of 1992, and its identification with the shear zone inferred from the distribution of tension cracks observed on the surface support the existence of a characteristic scale length of the order of 100 m affecting various earthquake phenomena in southern California, as evidenced earlier by the kink in the magnitude-frequency relation at about M3, the constant corner frequency for earthquakes with M below about 3, and the source-controlled $f_{\max}$ of 5-10 Hz for major earthquakes. The temporal correlation between coda Q^{-1} and the fractional rate of occurrence of earthquakes in the magnitude range 3-3.5, the geographical similarity of Q^{-1} and seismic velocity at a depth of 20 km, and the simultaneous change of coda Q^{-1} and conductivity at the lower crust support the hypotheses that coda Q^{-1} may represent the activity of creep fracture in the ductile part of the lithosphere occurring over cracks with a characteristic size of the order of 100 m. The existence of such a characteristic scale length cannot be consistent with the overall self-similarity of earthquakes unless we postulate a discrete hierarchy of such characteristic scale lengths. The discrete hierarchy of characteristic scale lengths is consistent with recently observed logarithmic periodicity in precursory seismicity.

(6) Coda Q for Double-Layered Random Media.

This is our first attempt to use wave-theoretical approach for understanding the effect of stratified random media on coda waves.

Using the indirect boundary integral scheme for multiple scattering of seismic waves developed by Benites et al. (1992), we compute SH-wave seismograms and measure frequency-dependent characteristic of coda Q in 2-D random media with a flat layer over a half space. Many circular cavities are randomly distributed in both the upper layer and the half-space, down to a certain depth (called the lower layer), simulating the upper and lower crusts, as shown in Figure 1. The scattering strength of cavities and intrinsic attenuation factor Q_i are different for each layer, and the S-wave velocity is prescribed constant throughout the media so that the computation of Green's functions for boundary integral is simple. Considering two basic parameters of our random media, scattering strength of cavities and intrinsic attenuation Q_i , we represent the actual shallow earth structure by an upper crust with large intrinsic attenuation and a lower crust with effective scatterers.

Figure 2 shows an example of theoretical seismograms for the case in which the radius of cavity $\alpha = 0.5$ km, the numbers of cavities are 16 and 48, respectively, in the upper and lower layer, and the intrinsic Q is 100 and 500, respectively in the upper and lower layer. The seismic source is located at a depth of 9 km as shown by the cross in Figure 1. The Ricker wavelet of a central frequency of 1.8 Hz is used as the source time function.

Computations of coda Q for several values of those parameters show that when the scattering is relatively strong, coda Q^{-1} is grossly independent of frequency. This result differs from the case of a uniformly random model, in which the value of coda Q^{-1} peaks around $kd=2$, as a function of kd , where k is the wavenumber and d is the cavity diameter. If the scattering strength in the lower layer is large enough to produce significantly more multiply scattered waves than singly scattered waves, the value of coda Q^{-1} strongly depends on the intrinsic attenuation in the lower layer, Q_{i2}^{-1} strongly depends on the intrinsic attenuation in the lower layer Q_{i2}^{-1} , and these two values (coda Q^{-1} and Q_{i2}^{-1}) become similar. Looking at the entire envelope and comparing the results for seismograms composed of only singly scattered waves, this result can

$a=0.5$ upper: 16 cavities, $Q_i=100$
 lower: 48 cavities, $Q_i=500$

Fig. 2

Fig. 1

be explained as follows: the waves scattered in the upper layer attenuate very quickly due to the large intrinsic attenuation and do not contribute to the coda envelope in a wide time-window started at twice the travel-time of the direct wave. The multiply scattered waves in the lower layer eventually arrive at the surface, forming the coda envelope which decays at a rate determined by the intrinsic attenuation factor in the lower layer, Q_{12}^{-1} . The hypothesis that the temporal decay of coda is controlled not by the scattering but by the energy leaking into a "transparent" underlying mantle is ruled out in general by our numerical simulations, except for very low frequencies. Although our model may be too simple to simulate the overall details of the observed coda Q accurately, the value of coda Q is quite likely to reflect the intrinsic attenuation in the Earth's lower crust.

(7) Spatial and Temporal Characteristics of coda Q^{-1} as a Geophysical Parameter in Southern California.

We have completed a systematic measurement of coda Q^{-1} using the enormous amount of digital data stored at the data center of the Southern California Earthquake Center for the period from 1987 to 1994. This period is particularly interesting for the coda Q^{-1} study because of four major earthquakes, namely, the 1992 Joshua Tree - Landers - Big Bear sequence and the 1994 Northridge earthquake occurred within the regional seismic network.

The seismic stations and the earthquake epicenters used for the coda Q^{-1} measurement are shown in Figures 3 and 4, respectively.

The measurement of coda Q^{-1} is made for five octave bands centered at 1.5, 3.0, 6.0-, 12.0 and 24.0 Hz, and for four lapse time windows, 30-30, 20-35, 20-40, and 20-45 sec. To qualify for the coda waves, the record must have a good signal to noise ratio for the time window, and the hypocentral distance is short enough so that the beginning of window must be later than twice the S wave arrival time. For the S velocity of 3.5 km/s, our windows beginning at the 20 sec lapse time require that the distance is shorter than 35 km.

The measured coda Q^{-1} for each frequency and time window is assigned to the mid-point between the epicenter and the station. We then group all the coda Q^{-1} with midpoints with 24' by 24' grid areas, as shown, for example, in Figure 5 and Figure 7, where the histogram of coda Q^{-1} is plotted for each grid area for the case of 1.5 Hz and 20-45s window and for the call of 6 Hz and 20-30 s window, respectively. We then calculate for each histogram the following statistics:

$$\begin{aligned} \text{Mean}(x_1 \dots x_n) &= \bar{x} = \frac{1}{N} \sum_{j=1}^N x_j \\ \text{Variance}(x_1 \dots x_N) &= \sigma^2 = \frac{1}{N-1} \sum_{j=1}^N [x_j - \bar{x}]^2, \\ \text{Skewness}(x_1 \dots x_n) &= \frac{1}{N} \sum_{j=1}^N \left[\frac{x_j - \bar{x}}{\sigma} \right]^3 \\ \text{Kurtosis}(x_1 \dots x_N) &= \left\{ \frac{1}{N-1} \sum_{j=1}^N \left[\frac{x_j - \bar{x}}{\sigma} \right]^4 \right\} - 3 \end{aligned}$$

The Chi-square test is also applied to find if each histogram obeys the Gaussian distribution. Pearson's test statistic q is calculated as

Figure 4 Map shows epicentral locations (dots) of earthquakes used to obtain coda Q^{-1} . Stars are epicenters of the Joshua Tree, Landers, Big Bear, and Northridge earthquakes.

Figure 5

dit20-45 X from -14.72 to 20

			65 10.633 2.523 0.066 -0.364 13.535	139 11.637 3.274 -0.011 -0.415 24.956					77 9.426 2.962 0.399 -0.383 19.356					
			100 9.779 3.128 -1.023 3.737 15.861	181 10.127 3.665 -0.725 4.360 15.426	135 9.434 2.636 0.284 -0.328 11.657				76 8.528 2.688 0.221 -0.630 27.386	85 9.939 2.914 -0.056 -0.772 19.756				
		50 7.509 2.120 0.190 -0.587 27.207	85 8.220 2.680 1.334 2.556 48.523	125 8.379 2.633 0.671 0.604 18.503	644 8.275 2.779 0.458 -0.258 34.310	324 8.904 2.880 0.185 -0.494 27.852	974 8.643 2.747 -0.314 2.622 28.007	276 8.522 3.063 -0.428 4.987 24.602	573 9.883 3.019 0.472 0.083 26.036	599 9.085 2.609 0.277 -0.413 28.474	81 8.989 2.473 0.489 -0.454 19.141			
				88 7.045 2.611 -1.085 6.099 20.976	201 6.786 2.785 -0.484 2.969 46.039	104 7.084 3.621 -0.471 2.087 25.973	311 8.487 2.641 0.193 0.693 24.239	602 8.025 2.495 0.228 0.842 25.895	915 9.705 3.065 0.183 -0.287 27.611	720 8.697 2.832 0.372 -0.280 23.923	1000 8.602 2.892 0.338 0.045 47.720			
									236 9.587 3.093 0.285 -0.363 26.517	642 8.543 3.457 -0.689 3.348 71.468	184 5.304 4.765 -1.963 4.091 194.701			
										135 6.674 2.575 -0.385 3.571 25.376	290 6.505 3.142 -1.052 5.200 72.164	87 1.487 6.506 -0.473 -1.071 74.245		
											248 7.569 2.761 0.159 1.298 33.768	180 4.948 6.266 -1.324 1.073 162.092		

Figure 6 (Continued)

Figure 7

d3120-30 X from -5.45 to 8

$$q = \sum_{i=1}^m \frac{(k_i - np_{oi})^2}{np_{oi}}$$

where n is the total number of data, m is the number of partitions, k_i is the number of data within the i -th partition, and np_{oi} is the number of successes within i -th partition according to the hypothetical distribution. The hypothesis is acceptable if and only if q is less than tabulated value $\chi^2_{1-\alpha}(m-1)$, where $1-\alpha$ is the confidence level.

Figure 6 shows the number of coda Q^{-1} measurement, the mean of coda Q^{-1} in 10^{-3} , the standard error of the mean, skewness, and kurtosis, and the test statistic q for each grid area corresponding to Figure 5. Figure 8 shows the same corresponding to Figure 7. In both figures, the value of q meeting the test for Gaussian distribution is shown in bold letters.

We found that longer the time window and lower the frequency, more grid areas meet the test for Gaussian distribution. Figure 6 shows, for example, 24 out of 34 (70%) grid areas meet the test for the 1.5 Hz and 20-45 s window, while Figure 8 shows only 1 out of 22 (9%) meet the test for the 6 Hz and 20-30 window.

Furthermore, it is remarkable that the departure from the Gaussian distribution occurs in the form of bi-modal distribution as shown clearly in Figure 7. One of the peak occurs at a positive value of coda Q^{-1} , and the other at a negative value. In order to find the cause of the bimodal distribution, we examined the mid-point location, focal depth, epicentral distance, time of measurement, magnitude for coda Q^{-1} belonging to separate modes. An example of results is shown in Figures 8 and 10 for the coda Q^{-1} at 6.0 Hz and 20-30 sec window within selected region.

They show that no obvious separation of the two groups depending on these parameters. The positive Q^{-1} and negative Q^{-1} can be found close to each other in these diagrams. We see, however, very subtle differences. The mid-points corresponding to the negative Q^{-1} group appears to lie in a zone trending NW to SE, subparallel to the San Andreas and other strike-slip faults in this area. We also see that there is only one negative Q^{-1} observed for focal depths deeper than 12 km. Since the negative Q^{-1} implies that strongly scattered waves arrive in the later part of time window, some erratic (specular?) reflections from the fault zone may be involved. The observed bimodal distribution might be related to the discrete hierarchy of structure as discussed earlier in items (4) and (5). In any case, it will require some regularity (bimodal) and irregularity (erratic behavior) on the scattered distribution. We shall propose modeling this by non-uniform scattering media.

In mapping coda Q^{-1} geographically, we eliminated the data belonging to the negative Q^{-1} mode, and measured the Q^{-1} value corresponding to the peak of the positive Q^{-1} mode. A typical example of the resultant map is shown in Figure 11 for the case of 12 Hz and 20-30 sec window. Remarkable features of these maps which are generally common to different frequencies and windows are (1) high Q (low Q^{-1}) region corresponding to the Peninsular Range where seismic velocity at mid-crust depth is high, gravity anomaly is high, and heat flow is low, and (2) low Q (high Q^{-1}) regions corresponding to the epicentral areas of 1952 Kern County, 1992 Joshua Tree - Landers - Big Bear, and 1994 Northridge earthquakes.

With regard to the temporal change in coda Q^{-1} we found several stations which show an increase in coda Q^{-1} about a year before the earthquake for Joshua-Tree and Landers, but other stations close to them do not show any change. We must conclude that the coda Q^{-1} precursor may be real but not reliable. This unreliability may be related to the cause of the bimodal distribution discussed above.

				58 3.166 1.515 -1.597 5.497 34.554									
					193 2.089 2.508 -1.020 -0.128 149.749	145 2.308 2.139 -1.317 1.317 118.891	259 0.950 2.573 -0.499 -1.161 239.756	76 1.383 2.688 -1.021 -0.185 70.710	146 2.449 1.987 -1.450 2.538 105.483	196 2.765 1.559 -1.237 3.192 126.605			
					146 2.946 1.634 -1.278 3.420 79.130	58 3.096 1.216 -0.247 1.761 21.216	130 1.264 2.745 -0.570 -0.718 95.781	179 1.599 2.444 -0.932 -0.143 154.122	329 2.783 1.814 -1.400 3.497 170.027	302 3.018 1.447 -1.066 3.789 137.912	406 3.034 1.582 -1.243 4.087 162.413		
									70 2.795 2.254 -1.077 2.155 53.566	202 1.918 2.189 -1.147 0.605 197.896	81 2.443 2.353 -1.253 0.682 81.090		
										50 3.054 1.689 -1.956 5.307 42.550	101 2.690 1.592 -1.403 4.352 44.281	52 1.804 2.574 -0.846 -0.743 64.461	
											90 3.011 1.176 -0.653 5.115 18.232	64 1.546 2.658 -0.902 -0.431 61.717	

Figure 8 (Continued)

Figure 9 Mid-point locations of coda 1/Q measurements at 6.0 Hz for 20 to 30 seconds within the regions bounded between latitude 33.7 degree and 34.5 degree, and between longitude 117.1 degree and 117.9 degree. It is shown that the two groups of coda 1/Q do not differ in spatial locations. Open circles represent larger 1/Q values, and solid diamonds smaller 1/Q values.

Figure 10 Coda 1/Q measurements at 6.0 Hz for 20 to 30 seconds, within a region bounded between latitude 34.1 degree and 34.5 degree, and between longitude 117.5 degree and 117.9 degree.

A: indicates no dependence on time; B: indicates no dependence on earthquake magnitude; C: indicates no dependence on focal depth; D: indicates no dependence on epicentral distance.

Figure 11 Map of coda Q^{-1} at 12.0 Hz and 20-30 seconds. See Figure 6.1 for detailed explanations.

(8) Seismogram Synthesis for Multi-Layered Media With Irregular Interfaces by Global Generalized Reflection/Transmission Matrix Method.

In this third article of a series study attempting to present a new method of seismogram synthesis for irregular multi-layered media problem, we present the theoretical development for the 2D P-SV case. Following the similar procedures developed for solving the SH problem presented in an earlier article (Chen, 1990), we first set up the basic matrix equations starting from the basic boundary integral equations, then introduce and determine the global generalized reflection/transmission (R/T) matrices, and finally solve the basic matrix equations by using the global generalized R/T matrices. It can be demonstrated that for the case of horizontally layered media (a special case of the general model), the general solution derived in this study naturally becomes identical to that of the generalized reflectivity method, which is directly developed for horizontally layered media. Therefore, this new method can be viewed as an extension of the generalized reflectivity method to the case of irregular topography, we derive the simplified formulas. This new method can provide an efficient, accurate, and unified algorithm to calculate the synthetic seismograms for a variety of the laterally heterogeneous media.

(9) Coda Q as a Combination of Scattering and Intrinsic Attenuation: Numerical Simulations With the Boundary Integral Method.

Numerical modeling of SH wave seismograms in media whose material properties are prescribed by a random distribution of many perfectly elastic cavities and by intrinsic absorption of seismic energy (anelasticity) demonstrates that the main characteristics of the coda waves, namely amplitude decay and duration, are well described by singly scattered waves in an elastic media rather than by multiply scattered waves in either elastic or an elastic media. We use the Boundary Integral scheme developed by BENITES et al. (1992) to compute the complete wave field and measure the values of the direct wave Q and coda waves Q in a wide range of frequencies, determining the spatial decay of the direct wave log-amplitude relation and the temporal decay of the coda envelope, respectively. The effects of both intrinsic absorption and pure scattering on the overall attenuation can be quantified separately by computing the Q values for corresponding models with (an elastic) and without (elastic) absorption. For the models considered in this study, the values of coda Q^{-1} in an elastic media are in good agreement with the sum of the corresponding scattering Q^{-1} and intrinsic Q^{-1} values, as established by the single-scattering model of Aki and Chouet (1975). Also, for the same random model with intrinsic absorption it appears that the singly scattered waves propagate without significant loss of energy as compared with the multiply scattered waves, which are strongly affected by absorption, suggesting its dominant role in the attenuation of coda waves.

(10) A Snap-Shot Mapping of Localized Heterogeneities in the Lithosphere Using the Coda Wave Decay from a Sparse Seismic Array.

The coda waves from local earthquakes have been analyzed by use of a parameter called coda Q^{-1} , which describes the rate of decay of coda power spectrum with lapse time after correction for geometrical spreading assuming single scattering (Aki and Chouet, 1975). The parameter served well on a regional and global scale revealing a systematic variation of its absolute value as well as its frequency dependence with the degree of tectonic activity.

The mapping of coda Q^{-1} on a local scale within a given tectonically active region, however, encountered a difficulty because of the large random variability of individual measurement. Attempts have been made to overcome this difficulty either by the use of the so called "doublets" data (Got et al., 1990; Beroza et al., 1995) or by averaging numerous measurements. A most comprehensive and systematic study of coda Q^{-1} using the enormous amount of high quality data from the data center of the Southern California Earthquake Center by Ouyang (1996) revealed that the statistical distribution of measured coda Q^{-1} often deviates from

the Gaussian distribution, and in some areas, forms a bimodal distribution with two peaks, one at a positive Q^{-1} and the other at a negative Q^{-1} .

Traditionally, coda Q^{-1} was measured for a single station, assuming that a common decay curve characterized by a constant Q^{-1} exists in the study area. This assumption may be justified if the scattering and absorption properties vary smoothly over the area, but may not be adequate if they vary sharply and strongly.

Some interesting and intriguing results have been obtained recently by Nishigami (1991) and Revenaugh (1995a, b and c) who inverted the coda wave data from a seismic array to determine the spatial distribution of scatterer intensities. Because of many factors involved in the inversion analysis, the link between the direct observation and the final result obtained by these sophisticated methods tends to be obscured. It is, therefore, desirable to present the interpretation of observations with the least manipulation of data.

The present paper describes a simple method of reducing the coda wave data from a seismic array into the form that may be related to a localized heterogeneity in the lithosphere. The method will be illustrated using the data from the seismic network at La Reunion in the Indian Ocean operated by the Observatoire Volcanologique du Piton de la Fournaise, Institut de Physique du Globe de Paris.

The Method

Figure 12 shows a map of seismic stations at La Reunion. The first step in our method is to find the amplification factor at each station by the coda method (see e.g., Kato et al., 1995 for demonstration of the validity of the coda method). Since the coda waves are not explained in any standard textbook of seismology, we give a brief description here. Following an earthquake, P, S, and surface waves propagate away from the epicenter. After these primary waves are gone, the epicentral area is still vibrating with a uniform amplitude except for the local site effect with larger amplitude at softer ground. The coda Q^{-1} measures the rate of decay of this vibration, which is roughly common to all stations and earthquake locations in a given area, while the amplitude at a particular site gives the station's amplification factor. The closest thing to this coda of local earthquakes is the residual sound in room acoustics discovered by Prof. Sabine of Harvard at the turn of the century. The decay rate of the residual sound was roughly independent of the locations of the sound source or receiver. The residual sound is due to ergodic multiple reflections at the room boundaries. Since it is absurd to propose a room-like structure in the lithosphere, coda waves are attributed to randomly distributed scatterers in the lithosphere. From an extensive study of coda waves in the central Asia, Rautian and Khalturin (1978) proposed a widely accepted rule of thumb that waves arriving at a given station after twice the S wave arrival time are qualified as coda waves with a common decay rate.

In order to obtain the station amplification factor for the Reunion network, we measured the amplitude of coda waves from a $M=3$ event near the northern end of the island (one of the largest recorded events in this low-seismicity area; its epicenter marked by a star in the inset of Fig. 12) for the lapse time window of 60 to 70 sec which meets the Rautian-Khalturin rule at all stations for this earthquake. The dominant frequency of coda waves was 2 to 3 Hz at all stations. The peak-to-peak amplitude averaged within the time interval at each station is normalized to the summit station BOR, and is shown beside each station in Fig. 1. For these measurements, we used the data processing software designed by J-L. Veinante and A. Nercissian of the Institut de Physique du Globe de Paris for a HP1000 at the observatory. The software includes a Fourier analysis package, from which we found that the amplification factor measured for the peak amplitude applies well to the frequency range from 1 to 3 Hz. The normalized amplitude, which we call the coda amplitude factor, varies from station to station by more than a factor of 3 with no systematic geographic distribution. We found that these

Fig. 12. Map of seismic stations on the island of Reunion, Indian Ocean. The size of the island is 50km x 70km, as shown in the inset. The number attached to each station is the coda amplification factor applicable to the frequency range 1 to 3 Hz. They are obtained from coda waves from an earthquake off-shore north of the island as marked by a star in the inset. The coda amplitude is measured for the lapse time window of 60 to 70 sec and is normalized to a summit station BOR.

amplification factors work very well for eliminating the site effect from various seismic signals recorded by the Reunion network such as T phases from mid-ocean ridge earthquakes, seismic motions caused by rock falls, and the long period events originating from the volcano as described in the Observatory's Seismology Annual Report for 1995. In most cases, we can draw contours of equal amplitude with a factor of 2 spacing smoothly interpolating values at observation points after this site correction.

According to the Rautian-Khalturin rule, the coda amplification factor at a given station obtained from different earthquakes should be the same as long as the coda window is chosen later than twice the S arrival time. We found a striking departure from what we expected at La Reunion. We measured the coda amplification factor for a $M=2.5$ earthquake that occurred at a depth of 17 km below the summit area. The epicenter of this earthquake is marked by a star in Fig 13. Coda wave in the time window 30 to 40 sec was band-pass filtered between 1 and 3 Hz, and the measured amplitude at each station is again normalized to station BOR. Fig 13 also shows two examples of seismograms on which the origin time (O.T.) and the 30 and 40 sec window are marked. The ratio of the normalized amplitude to the coda amplification factor given in Fig 12 is, then, taken at each station and is plotted in Fig 2. We expected this ratio to be close to 1.0 at all stations because the 30 and 40 sec. window meets the Rautian-Khalturin criterion for the $M=2.5$ earthquake. The ratio shown for station BOR is 1.0 by definition, but it decreases smoothly to a small value away from the summit area.

To check if this decrease is due to the source-dependent scattering processes, we made the same measurement for the two other events occurring outside the summit area, one in the southeast (Fig 14) and the other in the northwest (Fig 15). The time window of 30 to 40 sec again meets the Rautian-Khalturin criterion. As shown in Figs 13 and 15, the general pattern of contours of equal ratio remains the same independent of the location of seismic source, although the pattern is more strongly concentrated in the summit area when the seismic source is close to it. Preliminary applications of the present method to other areas such as Southern California, however, seem to show stronger source dependence of the coda ratio pattern than in the case of La Reunion.

The location and trend of peaks of the coda ratio contours in Figs 13 through 15 coincide with those of the presumed magmatic paths under the volcano marks as "NW intrusive axis, and NE and SE intrusive zones" in Fig 16 reproduced from Lenat and Bachelery (1988). This coincidence strongly suggests that an unusual localization of seismic waves of a few km/sec, the observed concentration of scattered energy within several km distance at a lapse time of 30 to 40 seconds is extraordinary. This trapping of seismic energy cannot be due to a low-velocity wave guide, because A. Hirn (personal communication) and his coworkers at the Institut de Physique du Globe de Paris demonstrated by extensive tomography and refraction surveys that these zones of presumed magmatic ascension at La Reunion are definitely of higher velocity than the surroundings.

In the following discussion of the above result, we shall assume that the distribution of the scattered seismic energy is nearly uniform in space at the lapse time window of 60 to 70 seconds, and the ratio of the coda for the 30 to 40 sec window to that for the 60 to 70 sec will be close to the site-effect corrected coda for the 30 to 40 sec window.

Discussions on the results from La Reunion

What impressed us most in Fig 13 through 15 is the smoothness of variation of the coda amplitude ratio from station to station. It is so smooth that we were able to draw meaningful contours of the ratio at an interval of 0.1, demonstrating the precision of the measured ratio within a few percent. Furthermore, we believe that the smooth spatial variation of the coda ratio is physically expected. Granted that taking the ratio eliminated the effect of local conditions of

Fig. 13. The coda amplitude of an earthquake located below the summit (epicenter marked by an arrow) for the time window of 30 to 40 sec and the frequency band between 1 and 3 Hz is normalized to the station BOR, divided by the coda amplification factor shown in Fig. 1, and the resultant ratio (called coda ratio) is shown beside each station. The seismograms at BOR and ICR are also shown to indicate the appearance of coda in the 30 to 40 sec time window. It is remarkable that we can draw smooth contours of the coda ratio at intervals of .1 through measured values. The peak of the contour coincides with the intrusive axis and zones shown in Fig. 16

Fig. 14. The same as Fig. 13, except for the earthquake located southwest of the summit as marked by a star. In spite of the difference in source location, the contour map of the coda ratio is similar to the one in Fig. 13.

Fig. 15. The same as fig. 13, except for the earthquake located northwest of the summit as marked by a star. The contour map of the coda ratio is again similar to those shown in Figs. 13 and 14.

Fig. 16. Generalized structural map of Piton de la Fournaise, reproduced from Lenat and Bachelery (1988).

recording, the ratios shown in Figs 13 and 15 represent snap shots of the scattered seismic amplitude at about 35 sec after the origin time. By this time, primary P waves and S waves would be at distances of 150 km and 8 km, respectively, away from the source. These figures show a concentration of scattered wave energy within a region of 10 km or so. The scattered energy must be following a diffusion-type equation like temperature, and therefore its spatial distribution must be smooth. For this reason a relatively sparse network like the one at La Reunion can be used to draw precise contour maps of scattered energy distribution.

In order to show how extraordinary the seismic scattering at La Reunion is, we shall put it in a perspective of wave scattering in general physics. For this purpose, we shall refer to a recent workshop where physicists of various branches working on wave diffusion in random media exchanged their observations and ideas. The workshop was called "Ecole interdisciplinaire sur la diffusion des ondes en milieu aleatoire", and was held at Cargese, Corsica on 21-27 May, 1996. We learned at this workshop that, in the terminology of physicists, our usual seismic coda waves belong to the so called "opaque" region, because the mean free path (10 to 100 km) for typical seismic coda falls between the wavelength (about 1 km) and the travel distance (100 to 1000 km). We also found that our observation at La Reunion may be a very exceptional case of wave scattering. Using elementary formulas for the scalar wave diffusion, one can make a very rough estimate of the mean free path from Figs 13 to 15. These figures suggest a mean spread of scattered energy of around 5 km for a lapse time of 35 seconds. From the distance x traveled by a particle walking randomly $\langle x^2 \rangle = 3Dt$, we estimate that diffusion coefficient D as $D = 25/(3 \times 35) = 0.24 \text{ km}^2/\text{s}$. In the framework of energy transport theory, D can be related to the transport mean free path l^* by $D = vl^*/3$ where v is the wave velocity. This give $l^* = 360 \text{ m}$ assuming $v = 2 \text{ km/s}$. Since the scattering mean free path is shorter than the transport mean free path, the wavelength is about 1 km, we find that the scattering mean free path may be much shorter than the wavelength in the zone of magmatic path at La Reunion. This is the region of scattering called "Anderson localization" by physicists (Anderson, 1958) which has not been produced in the laboratory for acoustic waves or elastic waves in 3-D.

Thus, the observed localization of scattered waves at La Reunion is not only exceptional for seismic coda waves, it is also quite unusual from the point of wave scattering in general physics. What is causing such an exceptional phenomenon on this island? A hint for answering this question appears to be in the vertical-component seismogram recorded at a summit station BOR shown in Fig 13. We find that the record had no clear first arriving P wave. P waves are very strong at other stations as exemplified by the ICR record also shown in the same figure. We found that all the events recorded at BOR share the same absence of clear first arriving P wave except for shallow events closer than 2 or 3 km to the station.

This suggests that the P wave may be trapped in the magmatic ascent path. A pseudo-honeycomb structure (not as regular as a real honeycomb), made of hard solid containing fluid magma, may behave like a grossly high velocity body, and still trap acoustic energy within the low-velocity magma becoming a source of extraordinary localization of seismic scattering. Unfortunately, because of the low seismicity of this region, we do not have enough data at present for definitively separating the effects of source radiation pattern, heterogeneous path, and possible temporal change in the structure in the absence of P waves.

The above speculation on the cause of the localization of seismic scattering, however, is in harmony with the peculiar behavior of this volcano before an eruption. As compared with other oceanic basaltic volcanoes, Piton de la Fournaise shows no classic cycles of inflation and deflation as observed for Kilauea, Hawaii and Krafla, Iceland. In this respect, it resembles Mt. Etna, where, for example, a major eruption starting on December 14, 1991 was preceded by negligible deformation and seismicity, but an increase of gravity in the summit area. This increase in gravity was interpreted in terms of passive intrusion of magma into a pre-stressed fracture that had remained open since the end of the last eruption in 1989 (Rymer et al, 1993).

One can imagine, under Piton de la Fournaise, a similar chain of cracks or cavities in a pseudo-honeycomb structure of hard rock which is filled with magma in an active period, and remains open when magma is drained back to a deeper reservoir in a quiet period.

If the above interpretation is valid, our proposed new method may offer a tool as useful as the microgravity monitoring for volcanoes covered by a seismic network with density comparable to that at La Reunion.

(11) A Finite Element Simulation of Volcanic Long-Period Events Observed at the Piton de la Fournaise, La Reunion.

Numerous long period (LP) events with the predominant frequency in the range of 1-3 Hz have been observed during the active period of the volcano Piton de la Fournaise in the past several years with the following characteristic: their arrivals at the stations located along the rim of the summit caldera are the earliest and nearly simultaneous; their amplitudes (vertical component) are the largest at the summit area, decay to 0.4 to 0.5 of the maximum at a distance of about 5 km from the summit using station correction based on the coda method. We started with the simplest model of a pressure-lined source buried in a homogenous half-space, and found that the observed arrival times of the first cycle of the event can be explained very well with the source depth of 2 km using seismic velocities appropriate for the vicinity of the summit area. This depth is close to the depths of volcano-tectonic earthquakes preceding an eruption which occur as a clustered swarm beneath the summit station. This simple model, however, underestimates the amplitude at distant stations relative to the summit station. In order to eliminate the above discrepancy, we introduce a low-velocity body to house the pressure source. We found that a vertically elongated reservoir would further decrease the amplitude ratio. On the other hand, the horizontally elongated body tends to reduce the discrepancy. We thus conclude that the magma reservoir responsible for the generation of LP events under the Piton de Fournaise is horizontally elongated.

(12) The use of long-period events for defining the magma system under Piton de la Fournaise, La Reunion.

A modern seismic network has been operated by the Institut de Physique du Globe de Paris since 1980 to monitor the volcanic activity of the Piton de la Fournaise in the Indian Ocean. The volcano erupted 28 times during the first 15 years of monitoring and has been completely quiet since the last eruption in August, 1982. We found that long-period (LP) events are originated beneath the central cone (2500 m elevation) at about the sea level, and their predominant frequency ranges from 1 to 3 Hz. LP events were abundant during the active period, and showed up immediately after the last eruption of August, 1982. Then, those with the predominant frequency around 1 Hz disappeared after November, 1993, and those with frequencies around 1 Hz disappeared in April, 1995. Since then, for more than one and a half years, we have not seen any LP events. This is consistent with the idea that the sources of the LP event involve magma, and we have no LP events if the reservoir is empty. It has been well known that the eruption of this volcano is preceded by a swarm of volcano-tectonic earthquakes located beneath the central cone. We found the precursor time correlates strongly with the elevation of the eruption site. It is less than 1 hour for a summit eruption, and as long as 9 hours when the eruption site is near the rim of the caldera along the presumed rift zone. There were no LP events as precursors of a summit eruption, while there were many of them as a precursor to rift zone eruptions. This suggests that reservoirs for the rift zone eruption are the source of LP events. There is a distinct difference between the volcano and Kilauea, Hawaii in the development of rift zone and central cone and the difference can be attributed to the difference in plate speed relative to the mantle.

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Figure Captions

Fig. 12. Map of seismic stations on the island of Reunion, Indian Ocean. The size of the island is 50km x 70km, as shown in the inset. The number attached to each station is the coda amplification factor applicable to the frequency range 1 to 3 Hz. They are obtained from coda waves from an earthquake off-shore north of the island as marked by a star in the inset. The coda amplitude is measured for the lapse time window of 60 to 70 sec and is normalized to a summit station BOR.

Fig. 13. The coda amplitude of an earthquake located below the summit (epicenter marked by an arrow) for the time window of 30 to 40 sec and the frequency band between 1 and 3 Hz is normalized to the station BOR, divided by the coda amplification factor shown in Fig. 1, and the resultant ratio (called coda ratio) is shown beside each station. The seismograms at BOR and ICR are also shown to indicate the appearance of coda in the 30 to 40 sec time window. It is remarkable that we can draw smooth contours of the coda ratio at intervals of .1 through measured values. The peak of the contour coincides with the intrusive axis and zones shown in fig. 16.

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Fig. 16. Generalized structural map of Piton de la Fournaise, reproduced from Lenat and Bachelery (1988).