Temporal Change in Scattering and Attenuation Associated with the Earthquake Occurrence—A Review of Recent Studies on Coda Waves

HARUO SATO

Abstract—The study of coda waves has recently attracted increasing attention from seismologists. This is due to the fact that it is viewed as a new means by which the stress accumulation stage preceding a large earthquake can be measured, since the scattering paths nearly uniformly cover a fairly large region around the focus and observation stations, compared with the direct ray paths. To date, we have had many reports on the temporal variation of the relation between coda duration and amplitude magnitude, and that of the coda attenuation $Q_C^1$ which is estimated from coda amplitude decay. Some of these have shown a precursor-like behavior; however, others seem to have shown a coseismic change. We have critically reviewed these reports, and discussed what these observational facts tell us about the change in the heterogeneous crust. We found significant temporal variations, not only in the mean but also in the scatter of $Q_C^1$, associated with the mainshock occurrence. The formation of new cracks, the reopening and growing of existing cracks, the interaction of these cracks, and the pore water movement through these cracks might correspond to such variations. In addition, we may expect an inhomogeneous distribution of crack clusters in a fairly large region, compared with the aftershock region. The gradual appearance of such crack clusters seems to be the most plausible mechanism by which coda decay gradients are caused to largely scatter in the stress accumulation stage.

Key words: Attenuation, scattering, coda, cracks, inhomogeneities, heterogeneities, seismograms, earthquake prediction.

1. Introduction

The decay rate of coda amplitude is in general very stable, and such a trait is common to all local earthquakes located near a given station (AKI, 1969). Coda waves have been interpreted as scattered S waves due to heterogeneities in the lithosphere. Measurements of scattering coefficient $g[\text{km}^{-1}]$ (SATO, 1978) and coda attenuation $Q_C^1$ (RAUTIAN and KHALTURIN, 1978; TSUJURA, 1978), which characterize coda excitation strength and coda decay gradient, respectively, have been made based on the single scattering assumption (AKI and CHOUET, 1975; KOPNICHEV, 1975; SATO, 1977) or the diffusion model (WESLEY, 1965). At first, we show you a schematic illustration of coda decay measurements in Figure 1(a). Coda duration $t - \tau$ time is usually measured from a visual seismogram, but $Q_C^1$ from a band-pass filtered

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(a) Schematical illustration of the coda analysis. The coda duration \( f - o \) time is measured from the earthquake origin time \( o \) to the time \( f \) when peak-to-peak amplitude decreases to a certain level being larger than noise level usually on a visual seismogram. \( Q_c^1 \) is usually measured from a band-bass filtered seismogram. The standard decay curve is predicted by the single scattering model, the diffusion model or an empirical model. A steeper coda decay corresponds to a larger \( Q_c^1 \).

(b) Reference space-time distributions of the single isotropic scattering model for coda energy density without coda attenuation \( (Q_c^1 = 0) \): \( W_0 \), S wave source energy; \( g_0 \), total scattering coefficient; \( f \), frequency; \( r_0 \), hypocentral distance; \( t \), lapse time; \( \beta_0 \), average S wave velocity (after SATO, 1977).
seismogram. Assuming single isotropic scattering and uniform distribution of scatterers (Sato, 1977) and introducing attenuation $Q_C^1$, we get the space-time distribution of coda energy density as

$$E_{\text{SSS}}(f; r_0, t) = \left[ \frac{W_0 g_0}{(4\pi r_0^2)} \right] K(\beta_0 t / r_0) \exp (-Q_C^1 2\pi ft) \quad \text{for} \quad \beta_0 t > r_0,$$

where $K(x) = (1/x) \ln [(x + 1)/(x - 1)]$; $f$, frequency; $r_0$, hypocentral distance; $t$, lapse time measured from the origin time; $W_0$, S wave source energy; $g_0$, total scattering coefficient; $\beta_0$, average S wave velocity (see Figure 1(b)). The asymptote of this function is

$$E_{\text{SSS}}(f; r_0, t) \sim \left[ \frac{W_0 g_0}{(2\pi \beta_0^2 t^2)} \right] \exp (-Q_C^1 2\pi ft) \quad \text{for} \quad \beta_0 t \gg r_0.$$

This coincides with Aki and Chouet's single back scattering model (1975). The equality $Q_C^1 = Q_S^1$ has been accepted in many reports; however, let us treat $Q_C^1$ as a phenomenological parameter characterizing coda decay in the following. These functions have been commonly used as the geometrical reference functions for the estimation of $Q_C^1$. Regional differences in $Q_C^1$ have been studied in various areas in the world in relation to tectonic activity (Singh and Herrmann, 1983; Sato, 1986b). The spectral structure of the inhomogeneity has been inferred from the frequency dependent measurements of the scattering coefficient and the $P$ and $S$ wave attenuation (Aki, 1980; Wu, 1982; Sato, 1984a,b).

Chouet (1979) first paid attention to the temporal change in $Q_C^1$ for small earthquakes in California. Recently, the study of coda waves has attracted increasing attention from seismologists, since it is viewed as a new means of measuring the stress accumulation for the purpose of earthquake prediction (Aki, 1985). This new way of measuring stress accumulation has been thought effective, since the scattering paths do not one-dimensionally, but three-dimensionally, cover a fairly wide region around the focus and observation stations nearly uniformly. To date, we have had many reports on the temporal variation of scattering and attenuation associated with the earthquake occurrence, based on coda wave analysis (Gusev and Lemziko, 1980, 1984, 1985; Jin, 1981; Tsukuda, 1985; Novelo-Casanova et al., 1985; Jin and Aki, 1986; Sato, 1986a, 1987; Sato et al., 1987; Lee et al., 1986; Peng et al., 1987). Some showed a precursor-like behavior; however, others seem to have shown a coseismic change. Reported studies were principally based on two kinds of analyses: the comparison between coda duration and amplitude magnitude, and the spectral measurements of $Q_C^1$.

Here, we have critically reviewed these reports, and discussed what these observations tell us about the change in the earth medium.

2. Recent Studies on Coda Waves

2.1. Relation between coda duration and single station magnitude

Phenomenological studies of the empirical relation between coda duration and
Table 1

Temporal change in the mean of $Q_c^2$

<table>
<thead>
<tr>
<th>* Coseismic changes</th>
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<tbody>
<tr>
<td>Haicheng E.</td>
<td>$(M_s = 7.3, \text{Feb. 4, 1975})$</td>
<td>$Q_c^2$ (Bef.) $&gt;$ $Q_c^2$ (Aft.) 4 \sim 12 \text{ Hz}</td>
</tr>
<tr>
<td>E. Yamanashi E.</td>
<td>$(M_s = 6.0, \text{Aug. 8, 1983})$</td>
<td>$Q_c^2$ (Bef.) $&gt;$ $Q_c^2$ (Aft.) 2 \sim 20 \text{ Hz}</td>
</tr>
<tr>
<td>Petatlan E.</td>
<td>$(M_s = 7.6, \text{Mar. 14, 1979})$</td>
<td>$Q_c^2$ (Bef.) $&gt; Q_c^2$ (Aft.) 6 \text{ Hz}</td>
</tr>
<tr>
<td>Misasa E.</td>
<td>$(M_s = 6.2, \text{Oct. 31, 1983})$</td>
<td>$Q_c^2$ (Bef.) $&lt; Q_c^2$ (Aft.) 5 \text{ Hz}</td>
</tr>
<tr>
<td>Round Valley E.</td>
<td>$(M_s = 5.7, \text{Nov. 23, 1984})$</td>
<td>$Q_c^2$ (Bef.) $&gt; Q_c^2$ (Aft.) \text{ Near the aftershock region}</td>
</tr>
<tr>
<td></td>
<td>$Q_c^2$ (Bef.) $&lt; Q_c^2$ (Aft.) 3 \sim 12 \text{ Hz}</td>
<td></td>
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</table>

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<tr>
<th>* Precursor-like changes</th>
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<tbody>
<tr>
<td>Urup E.</td>
<td>$(M_s = 8.1, \text{Oct. 13, 1963})$</td>
<td>$Q_c^2$ (1959–1962) $&lt; Q_c^2$ (1963–1964) \sim 0.8 \text{ Hz}</td>
</tr>
</tbody>
</table>

Amplitude magnitude have been intensively done.

A temporal magnitude change in the relation was reported for the Tangshan earthquake $(M_s = 7.8, \text{July 28, 1976})$ by Jin and Aki (1981), and Jin and Aki (1986). Coda durations for earthquakes of the same magnitude were abnormally shorter in a three-year period preceding the mainshock, compared with those before and after that period. On the other hand, from studying the Jianchuan earthquake $(M_L = 5.3, \text{July 3, 1982})$ in Yunnan, Yan and Mo (1984) found that coda durations were longer for the foreshocks than those for the aftershocks for the same maximum amplitude of S waves. Gao (1985) briefly reviewed recent studies of this kind made in China.

Hasegawa and Hori (1985) analyzed vertical component coda at a station just above the focal area of the southeastern Akita earthquake $(M_s = 6.2, \text{Oct. 16, 1970})$ in Japan. The mainshock focus was 6 km in depth. They found that the maximum amplitudes for the P and S waves of the foreshocks were greater than those for the aftershocks of the same coda duration for 10 \sim 15 \text{ Hz}. By assuming a constant scattering coefficient, they concluded that the attenuation above the mainshock focus increased after the mainshock occurrence.

In these studies, magnitudes were calculated from the maximum amplitudes of direct waves at a well calibrated single station. Therefore, there remains a possibility
that the reported change might be caused by the change in focal mechanisms, not by the real change of attenuation in the earth medium.

2.2. Relation between coda duration and average magnitude

In order to avoid the effect of radiation pattern changes, it is necessary to investigate the relation between coda duration and average magnitude, which is the arithmetic average of station magnitudes at the stations surrounding the epicenter.

The eastern Yamanashi earthquake ($M_S = 6.0$, Aug. 8, 1983) in Japan

SATO (1986a) found that coda durations ($f - o$ time) at two stations, TRU and HHR, in the close vicinity of the aftershock region were shorter before than those after the mainshock occurrence for the same average magnitude. For the analysis, he
used earthquakes which took place in a small volume of dimensions $10 \times 10 \times 5$ km$^3$ (hatched region in Figure 2) near the mainshock hypocenter. Station magnitudes were calculated from the maximum amplitudes of vertical seismograms, irrespective of the P or S waves. Especially, TRU is located just above the aftershock region and equipped with a seismometer at the bottom of a well 160 m in depth. Such differences were significant for coda durations shorter than about 30 s (see Figure 3), which correspond to a travel radius of about 50 km. The residuals of station magnitude from average magnitude were lower before the mainshock than after at five stations, ENZ, TRU, HHR, AKW and ASG, in the vicinity of the aftershock area; however, changes in the residuals of station magnitude at two farther stations, SMB and JIZ, took the opposite sign. It is difficult to explain these changes by the change in focal mechanisms.
TEMPORAL VARIATIONS

Temporal variations of the ratio $P_1$ for the number of earthquakes of $2 \leq M_L < 2.5$ to that of $1.5 \leq M_L < 2$ within a radius of 40 km and 60 km from the mainshock hypocenter (after IMOTO and ISHIKURO, 1986). High $P_1$ values correspond to low $b$ values.

From the two independent analyses, Sato concluded that the S wave attenuation intensity was higher before the mainshock than after, within a 26 km radius above the mainshock focus, which was 18 km in depth, in the upper crust, as illustrated by the speckled hemisphere in Figure 2: the decrease in $Q_s^{-1}$ was $1.4 \times 10^{-2} f^{-1}$ for $f = 2 \sim 3$ Hz from the coda analysis and for $f = 10 \sim 20$ Hz from the direct wave analysis, where he did not assume $Q_s^{-1} \propto f^{-1}$.

There were other reports of precursor-like changes associated with this earthquake. Figure 4 shows the temporal variation of $P_1$, which is defined by the ratio of the number of earthquakes of $2 \leq M_L < 2.5$ to that of $1.5 \leq M_L < 2$ around the mainshock focus (IMOTO and ISHIKURO, 1986). The $P_1$ value was higher, that is, the $b$ value was lower, during about a one year period before the mainshock than that before and after this period. $Q_s^{-1}$ was negatively correlated with the $b$ value. Tilt anomaly was observed 18 days preceding the mainshock at ENZ, which is located in an active fault at a 31 km distance from the mainshock epicenter (SATO et al., 1984).

The western Nagano earthquake ($M_s = 6.8$, Sep. 14, 1984) in Japan

SATO (1987) studied the relation between coda duration at GER close to the mainshock epicenter ($\Delta = 24$ km) and average magnitude measured from peak motion (see Figure 5). The vertical seismograms of earthquakes, which took place in a small restricted volume of $6 \times 6 \times 5$ km$^3$ (surrounded by a broken line in Figure 5(a)) in the middle of the aftershock region, were analyzed for a period from Feb. 1982 to Dec. 1984. Coda durations for earthquakes of the same average magnitude were abnormally longer in period B, which began about 16 months before and continued until 7 days after the mainshock, than those in the preceding time period A and in the following period C (see Figure 6). Figure 7 shows the temporal
Western Nagano Earthquake of Sep.14, 1984

Mt. Ontake

GER

Mainshock
(M₃ = 6.8)

Core Volume
(6x6x5 km³)

35.807°N

137.554°E

(a)

(b)

Figure 5
(a) Epicenter distribution before and after the western Nagano earthquake, Japan. Earthquakes which took place in the core volume are used for the coda analysis at GER. (b) Distribution of seismic stations and the mainshock epicenter (star) (after SATO, 1987).

Plots of the residual of the logarithm of coda duration from the linear regression. According to the one-sided statistical t-test, the significance level is less than 0.1% where the null hypothesis is that the means of duration residuals for two samples are equal. A gradual decrease is found even in period C. The predominant frequencies were about 5 Hz at the end of coda. We note that the temporal plots of the residual of station magnitude at GER from average magnitude scattered largely, however, did not show any systematic variations. Sato concluded that the most plausible mechanism for the elongation of coda durations was the increase of scattering intensity in the crust, within a radius of about 50 km from the mainshock focus, since the anomaly was significant for coda durations shorter than about 30 s, as shown in Figure 6.

YOKOYAMA (1986) reported that earthquake swarms began to concentrate close
Figure 6
Plots of the logarithm of coda duration (s) at GER against average magnitude: crosses, period A (Feb. 18, 1982–May 7, 1983); open circles, period B (May 7, 1983–Sep. 21, 1984); closed circles, period C (Sep. 22–Dec. 30, 1984); a solid line, linear regression (after Sato, 1987).

Western Nagano Earthquake (M_s = 6.8)

Figure 7
Temporal variations of the residual of the logarithm of coda duration from the linear regression before and after the western Nagano earthquake. Here, a bold line indicates the mean in each period.
to the mainshock epicenter after 1978. Sugisaki and Sugiuira (1986) reported an anomalous appearance of \( H_2 \) in bubble gases at a spring (\( \Delta = 52 \text{ km} \)), beginning one month before the mainshock. Katoh et al. (1986) reported that an anomalous increase of radon concentration in soil gas occurred from 1982 to the summer of 1983 at three sites around the mainshock epicenter, as illustrated in Figure 8. The chronological coincidence between the radon anomaly and the coda duration anomaly in the earthquake preparation stage suggests that the formation of new cracks and the reopening and growing of existing cracks progressed not only in the close vicinity of the focal area, but in a fairly large area. Scattering is strong when the wavelength is comparable to the dimension of a scatterer in general. Even though the length of each crack is much smaller than 1 meter, if we imagine a cluster of cracks with a dimension of the order of several hundred meters, the resulting contrast in elastic parameters functions as an effective scatterer for S waves of frequencies around 5 Hz. An appearance of such clusters in the stress accumulation stage might elongate coda durations.

Studying not the maximum amplitude but the relation between the pulse width of initial P waves observed at GER against travel time, Ohtake (1987) found that
Figure 9

$Q_p^1$ in the focal region was more than twice that of the surrounding region, not only for the post-seismic stage of 1.3 years, but also for a 2.5 year period prior to the mainshock, as shown in Figure 9. Ohtake did not find any temporal increase of $Q_p^1$ in the 16-month period before the mainshock, which was the anomalous period in the analysis done by Sato (1987). The highest value appeared for from one day to two months following the mainshock, period A-II, following which the $Q_p^1$ monotonously decreased, approaching the value of the surrounding area. The latter fact coincides well with Shibuya's (1979) investigation for the temporal change of $Q_p^1$ in the focal area of the off-Izu earthquake ($M_s = 6.9$, May 9, 1974) in Japan for the aftershock stage.

2.3. Coda decay gradient

The Tangshan earthquake ($M_s = 7.8$, July 28, 1976) in China

Jin and Akı (1986) analyzed vertical seismograms at station PG, which is 120 km distant from the mainshock epicenter, based on the single isotropic scattering model (Sato, 1977). They estimated that $Q_c^{-1} = 5 \times 10^{-3}$ around 2 Hz for Feb. 1969–Oct. 1972, $Q_C^{-1} = 1.4 \times 10^{-2}$ for coda just after the S wave, but $3.2 \times 10^{-3}$ for a latter part of coda for Apr. 1973–Feb. 1976, a three year period before the mainshock, and $Q_C^{-1} = 4 \times 10^{-3}$ for Sep. 1976–Jan. 1978 as listed in Table 1. They concluded that $Q_c^{-1}$ was anomalously high in an elliptic area 100 km wide, that is, the aftershock area plus the area extending to the NW direction in the three year period, but that the $Q_c^{-1}$ of a broad region outside that area became lower than normal. In this
Table 2

<table>
<thead>
<tr>
<th>Mainshock</th>
<th>Station</th>
<th>Inequality (Alternative hypothesis)</th>
<th>Frequency (Hz)</th>
<th>Significance level (One-sided F-test)</th>
</tr>
</thead>
<tbody>
<tr>
<td>* K = 13.3 E. in the Peter-the-1st Range, Soviet Central Asia (Mₛ = 5.2, Feb. 26, 1983)</td>
<td>Garm</td>
<td>σ (Bef.) &gt; σ (Aft.)</td>
<td>0.62</td>
<td>1.8%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>σ (Bef.) &gt; σ (Aft.)</td>
<td>1.25</td>
<td>0.003</td>
</tr>
<tr>
<td></td>
<td></td>
<td>σ (Bef.) &gt; σ (Aft.)</td>
<td>2.5</td>
<td>1.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td>σ (Bef.) &gt; σ (Aft.)</td>
<td>5</td>
<td>0.3</td>
</tr>
<tr>
<td></td>
<td></td>
<td>σ (Bef.) &gt; σ (Aft.)</td>
<td>10</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td></td>
<td>σ (Bef.) &gt; σ (Aft.)</td>
<td>18</td>
<td>6.1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>σ (Bef.) &gt; σ (Aft.)</td>
<td>27</td>
<td>30.4</td>
</tr>
<tr>
<td>* Tangshan E. (Mₛ = 7.8, Jul. 28, 1976)</td>
<td>PG</td>
<td>σ (2.1969–10.1972) &lt; σ</td>
<td>1 ~ 4 Hz</td>
<td>41.0%</td>
</tr>
<tr>
<td></td>
<td></td>
<td>σ (4.1973–2.1976) &gt; σ</td>
<td>16.4</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>(9.1976–1.1978)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>* Haicheng E. (Mₛ = 7.3, Feb. 4, 1975)</td>
<td>SHE</td>
<td>σ (Bef.) &gt; σ (Aft.)</td>
<td>5 ~ 12 Hz</td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>0.7</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>σ (Bef.) &gt; σ (Aft.)</td>
<td>4 ~ 12 Hz</td>
<td></td>
</tr>
<tr>
<td>* Round Valley E. (Mₛ = 5.7, Nov. 23, 1984)</td>
<td>The outer region</td>
<td>σ (Bef.) &gt; σ (Aft.)</td>
<td>1.5 ~ 24 Hz</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>(Long Valley Caldera)</td>
<td>t: 20 ~ 45 s</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Near the aftershock region (Regions S &amp; T)</td>
<td>σ (Bef.) &lt; σ (Aft.)</td>
<td>1.5 ~ 24 Hz</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>t: 20 ~ 45 s</td>
<td>except 3 Hz</td>
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<tr>
<td></td>
<td></td>
<td>σ (Bef.) &gt; σ (Aft.)</td>
<td>1.5 ~ 24 Hz</td>
<td></td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>t: 30 ~ 60 s</td>
<td>except 3 Hz</td>
</tr>
</tbody>
</table>

Here calculated from Tables 1 and 2 in Jin and Aki (1986), where σ for Q₋₀

Here evaluated from Fig. 8 in Gusev and Lemziko (1984), where σ for Q₋₀

abnormal period, the b value took a lower value in a 250 × 250 km² area extending to the SW of the aftershock area.

Analyzing the P wave spectra at two stations to the WNW of the mainshock epicenter, Zhu et al. (1977) found that the stress drops of small earthquakes were higher before than after the mainshock. Supposing frequency independent Q₋₀ for 1 ~ 10 Hz, they also found the increase of Q₋₀: Q₋₀ ~ 10⁻³ before and 1.5 × 10⁻³ after the mainshock. This brings to mind the report of a high Q₋₀, around 20 Hz in the earthquake swarm region, compared with that in the surrounding region for the Matsushiro earthquake swarm in Japan (Suzuki, 1971). What then does the different sign between the changes of Q₋₀ and Q₋₀ mean?

A big difference in Q₋₀ appeared in coda just after the direct S wave. We are
afraid of the effect of focal mechanism changes for such a time window. Precisely examining Table 1 of Jin and Akı (1986), we can test whether the standard deviation of $Q^{-1}$ in the three year period was larger than that in the surrounding time periods or not based on the one-sided statistical $F$-test. The significance levels for the rejection of the null hypothesis are 41.0\% and 16.4\% for the former pair and the latter pair, respectively (see Table 2).

*The Haicheng earthquake ($M_s = 7.3$, Feb. 4, 1975) in China*

Analyzing coda decays at SHE station in the vicinity of the mainshock epicenter, Jin and Akı (1986) estimated $Q^{-1} = (2.0 \pm 0.3) \times 10^{-3}$ before and $(1.4 \pm 0.15) \times 10^{-3}$ after the mainshock around 7 Hz, as shown in Figure 10. In this case, the difference appeared even in the latter part of coda. The travel radius of the coda was estimated to be about 60 km, much larger than the size of the aftershock area. They found a decrease of $Q^{-1}$ at another station, HEN, too. The negative correlation between the $b$ value and $Q^{-1}$ held good also in this case.

Analyzing the P wave spectra, Zhu et al. (1977) reported that $Q^{-1} \sim 2 \times 10^{-3}$ almost uniformly around the focal region before and after the mainshock, but that $Q^{-1}$ increased to $6 \times 10^{-3}$ only in the SW area of the mainshock epicenter, as shown in Figure 11. Finding the increase in the pulse width of initial P waves at a station in the close vicinity of the mainshock epicenter, Zhao et al. (1985) reported an increase of $Q^{-1}$ after the mainshock. Here again a discrepancy between $Q^{-1}$ and $Q_{C}^{-1}$ appeared.
Examining Table 2 of Jin and Aki (1986), we found that the standard deviation of $Q_C^{-1}$ was larger before than that after the mainshock. The significance levels for the rejection of the null hypothesis are 5% and 0.7% at SHE and HEN, respectively, based on the one-sided $F$-test (see Table 2).

The Ust-Kamchatsk earthquake ($M_S = 7.8$, Dec. 15, 1971) in Kamchatka

Analyzing three-component short period seismograms at three stations surrounding the focal area, Gusev and Lemziko (1985) found an anomalous change in the gradient of the coda envelope decay. The earthquakes used for this analysis were distributed in a 300 $\times$ 200 km$^2$ area covering the aftershock area, where the minimum coda duration used was 70 s. They measured the $a$ value, which represents the deviation of logarithmic coda decay gradient: a negative $a$ indicates a steeper decay corresponding to a larger $Q_C^{-1}$. They did not refer to any theoretical model but took the empirically averaged coda decay curve as the reference time function. They plotted the running means of 8 successive data, where each datum was an average of all the components at three stations (see Figure 12(a)). Compared with the relatively stable background level during 1966–1970, a pronounced anomalous decrease and recovery of $a$ was observed during the one year period preceding the mainshock. They reported such a decrease of $a$, corresponding to an increase of $Q_C^{-1}$, had been found for two other large earthquakes in the Kuril-Kamchatka zone, but the recovery was not found (Figures 12(b) and (c)). They concluded that the decrease in the mean of $a$ was a precursor and that the most probable cause was the increase of $Q_S^{-1}$. They
interpreted that a micro-crack formation would start abruptly about one year preceding the mainshock in a large area and seismic wave energy was spent on crack growth and compression of gas or fluid filled in cracks.

They also reported an appearance of a large amplitude phase at lapse time around 100 s in the anomalous period. Recently, Gusev (1986) radically proposed a seismic emission mechanism (i.e., Rykunov et al., 1979) for the explanation of this phase in coda.

However, we were unable to find an anomalous decrease in plots of $\alpha$ for vertical and horizontal components at four stations surrounding the Ust-Kamchatsk earthquake epicenter, as shown in Figure 13 (Figure 4 in Gusev and Lemzikov (1980)). An increase of the standard deviation of $\alpha$ during the period from 1971 to 1973 looks more anomalous to us (see Table 2). Gusev and Lemzikov (1984) reported that the significance level was less than 5% according to the statistical F test, where the null hypothesis was that the standard deviations of $\alpha$ for 1966–1970 and 1971 are equal.
Earthquakes of the maximum magnitude 5 or 6 have taken place with a several-year interval in the Peter-the-first range, Sauristan region. EW-component band-pass filtered seismograms (RAUUL and KANTUN) have been in operation since 1978 at station Garm, which is close to the mainshock epicenter (A = 20 km). Sato et al. (1987) measured log (C) for 7 frequency bands from coda decays for lapse times greater than twice the S wave travel time and smaller.
\[ A(f,t) \propto \frac{\exp(-Q_c^{-1}fnt)}{t} \]

for \( t > 2t_s \)

Figure 14
Temporal variations of \( \log(Q_c^{-1}f) \) for 7 frequency bands before and after a \( K = 13.3 \) earthquake (star) near Garm, in the Peter-the-first range, Soviet Central Asia, where \( f \) is the central frequency in Hz, open and closed circles correspond to the foreshocks and aftershocks, respectively (after Sato et al., 1987).

than 100 s (see Figure 14). Earthquakes that occurred in the aftershock region of dimensions \( 25 \times 11 \times 20 \) km\(^3\) during 1979–1984 were used for the analysis. At 5 Hz band, \( \log(Q_c^{-1}f) \) took a significantly lower value in the three-year period preceding the mainshock, where a gradual increase began in early 1981 (see Figure 15). They could not find a clear difference between the means of \( \log(Q_c^{-1}f) \) before and after the mainshock for the other 6 frequency bands. They, however, preferably pointed
Figure 15
(a) Temporal variations of log ($Q^{-1}$) at 5 Hz band, where a bold line and sand area show the mean and the standard deviation, respectively, in each period: crosses, Feb. 22, 1979–Sep. 1, 1979; open boxes, Oct. 20, 1979–Oct. 15, 1982; closed boxes, Feb. 12, 1983–Sep. 5, 1984. (b) Monthly number of earthquakes of $K \geq 7$ which occurred in the aftershock region. (After Sato et al., 1987).

Figure 16
Temporal variations of $Q^{-1}$ at 12 Hz band for hypocenters in two regions before and after the Round Valley earthquake (arrow): A (aftershock region) and B (next to the aftershock region) (after Lee et al., 1986).
out a significant difference between the standard deviations of \( \log(Q_{C1}^1 f) \) before and after the mainshock. Standard deviations were significantly larger before the mainshock than after, for 0.62, 1.25, 2.5, 5 and 18 Hz bands, based on the one-sided statistical \( F \)-test. The significance level for each frequency band is given in Table 2.

From the instability of coda decay, they interpreted the larger standard deviation of \( \log(Q_{C1}^1 f) \) to be due to the appearance of clusters of cracks in a fairly large region in a stress accumulation stage prior to the mainshock failure.

*The Round Valley earthquake (\( M_S = 5.7 \), Nov. 23, 1984) in California*

Since Oct. 1978, four earthquakes of magnitude around 6 and preceding several swarms have occurred in the Mammoth Lakes area. Peng et al. (1987) analyzed vertical seismograms of earthquakes which occurred in a period from Apr. 1984 to Jan. 1985. Applying the single scattering model to coda for lapse times of 20–45 s and 30–60 s, they reported significant temporal changes in the mean of \( Q_{C1}^1 \), in relation to the location of the mid point of hypocenter-station pair. Those changes were as follows: (1) \( Q_{C1}^1 \) was smaller before the mainshock than after, in regions near the mainshock epicenter, mostly for frequencies 3 ~ 12 Hz; (2) \( Q_{C1}^1 \) was larger before the mainshock than after, in regions farther away from the mainshock epicenter; (3) the high \( Q_{C1}^1 \) in the Long Valley Caldera, which was comparably higher than that for the outside area, disappeared after the mainshock. They introduced the doughnut pattern model (Mogi, 1977) for the interpretation, expecting a low \( Q_{C1}^1 \) in the area of quiescence and a high \( Q_{C1}^1 \) in the active zone. But, the finding of a high \( Q_{C1}^1 \) in the outer region runs counter to the case of the Tangshan earthquake.

We found that the standard deviation of \( Q_{C1}^1 \) seems to have systematically changed for wide frequency bands, from the calculation based on Tables 1–6 of Peng et al. (1987), as shown in Table 2. The standard deviation of \( Q_{C1}^1 \) was larger before the mainshock than after in the Long Valley Caldera, region I. From the study of \( Q_{C1}^1 \) in different time windows, the standard deviation of \( Q_{C1}^1 \) before the mainshock was larger than those after the mainshock for 30–60 s, but the change was opposite for 20–45 s, in the regions S and T close to the aftershock region. Coda for the shorter lapse time samples the closer region. The larger standard deviations before the mainshock are seen in Figures 4 and 5 in Lee et al. (1986) as they suggested (see Figure 16). We may say that the coda decay gradients were more unstable before the mainshock than after, in a fairly large region outside of the aftershock region; however, they became more unstable after the mainshock than before, in the vicinity of the aftershock region.

*Other coda decay studies on natural earthquakes*

Novelo-Casanova et al. (1985) reported high \( Q_{C1}^1 \) before and low \( Q_{C1}^1 \) after the Petatlan earthquake (\( M_S = 7.6 \), Mar. 14, 1979) in Mexico at 6 Hz band. We are
apprehensive of the strong affection of the radiation pattern changes, since they applied the single isotropic scattering model to the early part of coda for the estimation of $Q_c^{-1}$.

TSUKUDA (1985) analyzed vertical seismograms of micro-earthquakes before and after the Misasa earthquake ($M_s = 6.2$, Oct. 31, 1983) in Tottori-ken, Japan. He reported that the mean of $Q_c^{-1}$ at 5 Hz band was significantly higher after the mainshock than that during the preceding seven years as shown in Figure 17, where the significance level was less than 2%, according to the statistical $t$-test.

We have summarized the reported changes in the means and the standard deviations of $Q_c^{-1}$ in Tables 1 and 2.

2.4. Ratio of coda duration of horizontal component to that of vertical component

Coda durations strongly depend on the magnification scale of seismograms. Thus, MALAMUD (1974) began to monitor the nondimensional quantity: the ratio of coda duration of the horizontal component to that of the vertical component. Analyzing earthquakes within 100 km in epicentral distance, and where the durations were mostly from 100 to 300 s, he found that the ratio had a lower value in the few months preceding the major earthquakes of $K \geq 13$ ($M_L > 5$) in Soviet Central Asia, as shown in Figure 18.

YAN and Mo (1984) studied the Jianchuanch earthquake ($M_L = 5.3$, July. 3, 1982) in Yunnan, China. They found a clear decrease of the ratio in the few-day period preceding the mainshock, as plotted in Figure 19, when coda durations ($f - p$ time) were mostly from 10 to 30 s.

These facts suggest the possibility of different values for $Q_s^{-1}$ between vertical and
Figure 18
Temporal variation in the ratio of coda duration of horizontal component ($t_h$) to that of vertical component ($t_v$), where an arrow indicates an earthquake of $M_L > 5$ in Soviet Central Asia (after MALAMUD, 1974).

Figure 19
Temporal variation of the ratio of coda duration of horizontal component to that of vertical component ($t_h/t_v$) before and after the Jianchuan earthquake in Yunnan, China (after YAN and MO, 1984).

horizontal polarizations. GUSEV (1986) introduced crack-induced anisotropy of $Q_s^{-1}$ for the explanation. In addition, the possibility is also suggested that coda amplitudes might be different for different components, according to the change in focal mechanism.
2.5. Coda of explosion seismograms

Nikolaev's group at the Institute of the Physics of the Earth, USSR intensively repeated explosions in two lakes, Iskanderkuli and Chashmaisangok, in the western part of the Gissar-Kokshal zone, in Soviet Central Asia from 1978 to 1984. They recorded seismic waves at epicentral distances of 50 ~ 130 km. Analyzing the power spectrum of the radial component coda waves, Gamburtseva et al. (1983) found an anomalous decrease in the spectrum around 2.5 Hz for lapse time of 21–40 s prior to the two nearby \( K \sim 12 \) (\( M_L \sim 4.5 \)) earthquakes of 1979 and 1980, as shown in Figure 20. From the same experiments, S. P. Starodubrovskaya (personal communication, 1985) found the appearance of a large amplitude phase in coda waves at lapse time around 40 s in a period just before these earthquakes. Such a phase was observed at stations for which the direct ray paths crossed active faults; however, it did not appear in a seismogram at a station for which the direct ray path did not cross the active fault.

In China, Gu et al. (1982) studied the temporal change in coda of explosion seismograms before and after the Liyang earthquake (\( M_S = 6.0 \), July 9, 1979). They found anomalies in the relation between the predominant frequency of coda and lapse time, as shown in Figure 21(a). They found that the exponential decay factor \( \alpha \) for the coda envelope was abnormally small (see Figure 21(b)), that is, there was a small \( Q_c^{-1} \), and that the coda durations were anomalously long in the one year period prior to the mainshock.

3. Discussion

3.1. Overview of the reported changes in coda waves

We can summarize the above reports as follows:

1. Changes in the relation between coda duration and amplitude magnitude were found to be associated with the earthquake occurrence, where predominant frequencies were relatively low, such as 2 ~ 6 Hz.

2. In most cases, the mean value of \( Q_c^{-1} \) was higher before the mainshock than after, or higher in the few-year period around the mainshock than the surrounding time periods for frequencies mostly lower than or equal to 12 Hz, and this change has been interpreted, based on the change in \( Q_S^{-1} \). In some cases, however, the change was opposite. Many reports showed a negative correlation between the \( b \) value and \( Q_c^{-1} \).

3. Spatial difference of the change in the mean of \( Q_c^{-1} \) was found to be based on the coda decay analysis for different time windows. The doughnut pattern model was introduced for the explanation.

4. Temporal change in the scatter of \( Q_c^{-1} \), in relation to the earthquake occurr-
Temporal variation in the coda power spectrum of horizontal component explosion seismograms in the radial direction for lapse time of 21 ~ 40 s observed at Shaambari. The shot point was the Iskanderkuli Lake in Soviet Central Asia (after Gamburgseva et al., 1983).

...reference, was systematic for wide frequencies. In most cases, the standard deviation was larger before the mainshock than after. The increase in the scatter of $Q_C^{-1}$ began a few years before and/or continued until a little after the mainshock in some cases. The study of the Round Valley earthquake suggests that the standard deviations of $Q_C^{-1}$ were larger before than after the mainshock, especially in the outer region of the aftershock region.

5. Abnormal increases in radon concentration and an appearance of $H_2$ gas were observed at large epicentral distances in a period of long coda durations.

6. These observational facts suggest that physical property changed in a fairly large region, compared with the aftershock region prior to the mainshock.

However, we are diffident that the above summary might be too optimistic. Consequently, we would like to add the following comments:

1. We know that different time windows give different values for $Q_C^{-1}$; however, some papers did not clearly mention the onset and the length of the time window used.

2. Changes in the relation between coda duration and single station magnitude might be caused by the change in radiation pattern. The change in the mean of $Q_C^{-1}$ also might be caused by the change in focal mechanism in some cases, since the single isotropic scattering model was applied to the early part of coda close to the direct S wave, mostly in the vertical component. Anomaly appeared mostly in small magnitude earthquakes, for which the focal mechanism determination was difficult. Sato (1984a) recently showed that the decay gradient of the coda envelope is different for different components, and strongly depends on the focal mechanism in such a time window, based on the single scattering assumption for vector waves. Figures 22(a), (b) and (c) show synthesized envelopes for a three-component seismogram of an $M_L = 3$ earthquake at 30 km in distance for three different focal mechanisms,
Figure 21
Temporal variations in coda parameter of explosion seismograms before and after the Linyang earthquake in China. (a) Temporal change in $k$ (s$^{-2}$), where $k$ is the negative gradient of the predominant frequency $f$ of coda against lapse time $t$. $f = a - kt$. (b) Temporal change in the exponential decay parameter $\alpha$ (s$^{-1}$) of coda envelopes. (After Gu et al., 1982).

where the randomness of elastic moduli is characterized by the exponential autocorrelation function with 10% in RMS fractional fluctuation and 2 km in correlation distance.

3. In some reports, the single scattering model was applied to a latter part of
Synthesized envelopes of three component seismograms of an $M_L = 3$ earthquake at a distance of 30 km in the inhomogeneous lithosphere based on the single scattering approximation for vector waves, where $\kappa_{\text{in}} = 10^{-2}$ m/s$^2$. The randomness of elastic moduli is characterized by the exponential auto-correlation function with $10^6$ in RMS fractional fluctuation and a correlation distance of 2 km. The normal vector $\vec{n}$ and the slip vector $\vec{s}$ characterize the point shear dislocation. (a), (b) and (c) show different focal mechanisms. The third axis is taken to the radial direction. (After SATO, 1984a).
coda of lapse time longer than the mean free time of S waves, which was roughly estimated to be about 30 s (SATO, 1978, 1984a). For such a long lapse time, coda amplitudes are theoretically different in different components, under the single scattering assumption, as illustrated in Figure 22. Multiple scattering should be considered in such a lapse time. GAO et al. (1983) constructed a multiple scattering model up to the 7th order. A numerical simulation approach was taken by FRANKEL and CLAYTON (1986) for the case of short mean free path. They demonstrated the discrepancy between $Q^1_C$, evaluated from coda decay gradients, and the scattering attenuation of direct wave amplitude. GUSEV and ABUBAKIROV (1986) proposed a reference time function for acoustic energy density, interpolating the single scattering model and the diffusion model, based on the Monte-Carlo simulation for non-isotropic multiple scattering. Will multiple scattering decrease the difference between the coda amplitudes in different components for long lapse time?

4. Contrary to our expectation, there are very few reports on the temporal change in the mean of $Q^1_C$ for high frequencies.

5. There are many reports on the increase of $Q^1_P$ in the close vicinity of the aftershock region after the mainshock occurrence, compared with that before. Does the contradiction between the temporal changes of $Q^1_C$ and $Q^1_P$ indicate the physical difference or the spatial difference between the sampled regions? We should pay attention to this discrepancy.

6. We should take more note of the change in scattering coefficient $g$, which controls the absolute level of coda excitation. If the scattering coefficient largely increases in a restricted region, coda durations become longer and coda decay gradients of earthquakes in this region become steeper than normal at a nearby station. It may be interpreted as an increase of $Q^1_C$. In the opposite case, if the scattering coefficient increases larger in the outer region, coda durations become longer, but $Q^1_C$ apparently becomes smaller than normal. Like this, $Q^1_C$ reflects the spatial inhomogeneity of the scattering coefficient distribution. We should remember that $Q^1_C = Q^1_S$ holds good only when the spatial distributions of $Q^1_S$ and $g$ are uniform and isotropic, and when the scattering is weak.

7. We must note that there was a possibility for misinterpretation of the spatial differences as the temporal differences in $Q^1_C$, since some papers reviewed here did not restrict the spatial distribution of hypocenters in a small volume.

8. Coda excitation strength may be due to local site effects as much as scatterers in large volumes (PHILLIPS and AKI, 1986). Different sites may observe different coda decay rates for the same earthquake. We should have devoted more attention to the local site effects, even in the study of temporal changes.

9. We must note that there is a report of little temporal change in $Q^1_C$ before and after the Adak earthquake ($m_s = 5.8$, May 6, 1984) in the Aleutian islands (SCHERBAUM and KISSLINGER, 1985). Poupinet et al. (1984) found earthquakes of very similar wave form before and after the Coyote Lake earthquake ($M_s = 5.9$, Aug. 6, 1979) in California.
Figure 23
Temporal plots of $V_p/V_s$ (a), and the RMS of the residuals of the Wadati-diagram data from the regression line at (b), before and after a $K = 13$ ($M_L \sim 5$) earthquake in Soviet Central Asia (after Kulagina et al., 1982).

Figure 24
Temporal changes of $V_p$ along various ray traces against uniaxial loading stress for a large granite block. The scatter of $V_p$ increased before the fracture. (After Sobolev, 1984).
3.2. Cluster formation hypothesis

We can easily imagine a formation of new cracks, reopening and growing of existing cracks, and interaction and connection of nearby cracks in the stress accumulation stage (NUR, 1972; SCHOLZ et al., 1973; MJACHKIN et al., 1975). CRAMPIN et al. (1984) proposed the extensive-dilatancy anisotropy of micro-cracks throughout a fairly wide region around an impending earthquake. It is expected that $Q_C^1$ will detect the change in heterogeneous structure in a broad region, including an earthquake source region as suggested from distant precursors observed (Aki, 1985). There is, however, a contradiction between both the temporal and spatial changes in the mean value of $Q_C^1$ in different reports. Recently, GUSEV (1986) proposed different time histories for the dilatancy region around the impending earthquake, and the outer region in the compressional stress field, to explain the doughnut pattern. Presently, we are not sure whether these discrepancies will be settled by the doughnut pattern model or not.

Here, let us direct more close attention to the change in the standard deviation of $Q_C^1$, not as a credibility parameter for the mean of $Q_C^1$, but rather as a physical parameter. Many reports have suggested an abnormal increase in the scatter of $Q_C^1$ for the period just before or just around the mainshock occurrence. In addition, there are reports on the anomalous appearance of a large amplitude phase in coda. These observational facts suggest the instability of the coda decay gradient in the stress accumulation stage.

KULAGINA et al. (1982) showed an anomalous increase in the scatter of the Wadati-diagram plots in the earthquake preparation stage in Soviet Central Asia, as shown in Figure 23. SOBOLEV (1984) experimentally reported the appearance of high and low velocity spots and clusters of acoustic emission hypocenters, not only in the close vicinity of the fault preparation zone, but also in a wide region of a big granite block under the uniaxial loading, just before the fracture. He mentioned that the most prominent precursor was the increase of the scatter of seismic velocity, as shown in Figure 24.

The elastic moduli are inhomogeneous, as inferred from well log data (SATO, 1984b), and the in situ stress distribution is also nonuniform, as measured precisely by the hydraulic fracturing experiment in a drilled borehole, as illustrated in Figure 25 (IKEDA and TSUKAHARA, 1987).

It is very natural for us to imagine a nonuniform cluster formation of cracks in a fairly large region, compared with the focal region. The decrease of the $b$ value in this stage is in harmony with the repression of the independent appearance of small cracks. The resulting contrast in elastic parameters functions as a scatterer for S waves. We can expect a wide distribution for the size of clusters. Clusters of a dimension of the order of several hundred meters are able to change the relation between coda duration and magnitude in relatively low frequencies less than 10 Hz.
Horizontal in situ stress in a borehole at Ashigawa-mura, Japan measured by the hydraulic fracturing experiment: open and closed circles correspond to the maximum and the minimum, respectively. A stress concentration is found between 160 m and 190 m in depth. (After Ikeda and Tsukahara, 1987.)

We can expect the movement of pore water through cracks in such clusters. Then, a cluster of cracks containing water as viscous fluid or as adsorbed on the mineral surfaces (Clark et al., 1980) works as a strong intrinsic attenuation spot for S waves. We may expect that the increase of elastic scattering will dominate over the increase of intrinsic and scattering attenuation for the coda wave formation in some cases, and that the opposite situation will appear in the other case.

The gradual appearance of the inhomogeneously distributed scatterers and attenuation spots will make the coda decay gradients largely scatter in the stress accumulation stage. We may interpret these phenomena by analogy with the increase of the fluctuation in the critical state in physics. Just before or a little after the mainshock occurrence, we may expect that the growth of clusters of cracks will stop in the outer region, and then coda decay gradients will become stable. On the other hand, coda decay gradients will be more unstable after the mainshock in the aftershock region, because of the mainshock and aftershock fractures and the pore water movement through them.

3.3. Suggestions for further research

The following coda researches are recommended for a better understanding of the temporal change in inhomogeneities in the crust, in relation to earthquake occurrences.

1. Simultaneous measurements of $Q^{-1}$, $Q_{S}^{-1}$ and $Q_{T}^{-1}$, and examining the con-
sistency or the inconsistency between their temporal changes. We have to clarify the
constraints on which component of ground motion is measured, how time window
is defined, and how magnitude and focal mechanism is factored into the analysis.

2. Monitoring not only the mean but also the scatter of $Q_C^{-1}$ itself as a physical
parameter. The latter is valuable for examining the cluster formation hypothesis.

3. Precise examination of individual phases in coda waves.

4. Comparison of coda measurements, not only with the $b$ value and stress drops,
but also with geochemical observation (Rn, $H_2$, ...), groundwater-level monitoring,
and electric resistivity measurement for examining the crack formation and the pore
water contribution.

5. Re-examination of the coda portion of explosion seismograms. It is effective
for the quantitative evaluation of the scattering coefficient $g$.

6. Observation of coda by a three-component seismometer, in order to clarify
the energy partition into different components.

7. Comparison between $Q_S^{-1}$ measured from direct S waves, and $Q_C^{-1}$ and $g$
measured from coda waves. $Q_C^{-1}$ is probably a complicated function of the spatial
distribution of $Q_S^{-1}$ and $g$.

8. Theoretical construction of the inversion scheme for the spatial distribution of
$Q_S^{-1}$ and $g$ from coda envelopes, since different lapse time window samples different
region in space. It will be useful to examine whether or not the doughnut pattern
model holds good. At first, we should try to make such a scheme for coda of short
lapse time based on the single scattering approximation.

9. Theoretical construction of the multiple scattering model for the envelopes of
three-component seismograms for long lapse time.

10. Precise measurements of in situ stress in deep boreholes by using the
hydraulic fracturing technique to clarify the inhomogeneous scale in the crust.

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