JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 100, NO. B10, PAGES 20,311–20,319, OCTOBER 10, 1995

Baylike and continuous variations of the relative level of the late coda during 24 years of observation on Kamchatka

Alexander A. Gusev[1]

Instituto de Geofisica, Unversidad Nacional Autónoma de México, Mexico City

Abstract. The relative amplitude levels of backscattered shear waves (coda) of local earthquakes as expressed in coda magnitude residuals were studied for nine Kamchatka seismograph stations. Coda magnitudes were determined from coda amplitudes measured on three-component photograph records of 1.2-s period displacement-recording (between 1 and 10 Hz) instruments. A coda magnitude scale was specially designed for this study. It is rather precise, with a standard deviation of about 0.15 for log amplitudes measured at any single station. At each station, for each event we compute the station magnitude residual, i.e., station magnitude (averaged over three components) minus network average magnitude. Station corrections for each station and component were determined and included in the procedure. Time series of coda magnitude residuals including 500-1000 events for each station for the 24-year period of observations (1967-1990) show moderate but statistically significant oscillations around a constant level. Superposed on this background, two prominent anomalies are revealed. One, at station KBG, of 3 years duration, preceded two $M \approx 8$ shallow earthquakes within 100 km from the station. Another, at station APH, of 1.5 years duration, preceded a major (volume of 2.5 km^3) fissure volcanic eruption within a 70-km distance from the station. No other comparable shallow earthquakes or eruptions took place on Kamchatka during this period. The sign and the amplitude of the anomalies indicate a possible 30% increase of S -wave attenuation in the lithosphere under the stations.

Introduction

A remarkable property of the local earthquake seismic coda is its nearly identical envelope of amplitude decay over various earthquakes recorded at closely spaced seismic stations. This phenomenon is believed to be a consequence of coda formation resulting from random (back)scattered waves [*Aki*, 1969; *Aki and Chouet*, 1975]. Relative stability of coda shape provides a basis for identification of temporal variations of attenuating properties of the lithosphere. This can be done by employing the coda decay rate [*Chouet*, 1979; *Gusev and Lemzikov*, 1980, 1984, 1985; *Aki*, 1985], the record (essentially, coda) duration [*Malamud*, 1974; *Jin and Aki*, 1986; *Sato*, 1986], or coda amplitude level as proposed here. Though coda amplitude level and coda duration are inherently closely related parameters, we prefer direct measurements of amplitude to record duration for the following reasons. Primarily, in terms of corresponding coda magnitude, the variance for duration magnitude is much larger. For example, compare rms (over network) residual of log coda amplitude of 0.14 found in this study with rms residual of duration magnitude of 0.25 to 0.5 for similar instrumentation type and magnitude range [*Tsimura*, 1967, Figure 8 a,b]. Second, for relatively small and large values of magnitude, the measurement procedure which determines duration magnitude employs coda segments with different lapse times (earlier segment for lower magnitude, later segment for larger magnitude). Coda waves constituting these two segments are formed by scattering from medium volumes of different size, whereas we would prefer the medium volume responsible for formation of analyzed coda segment to be as definite as possible. Finally, specifically for places like Kamchatka, duration magnitude depends on the level of ocean-wave-generated microseisms and will be apparently lower in bad weather and, generally, in the season of cyclones, when the level of background microseisms is high.

In order to monitor changes of relative coda level at a particular station, some reference is needed. For this, we employ the average coda level over the whole seismic network; in other words, we use coda magnitude residual. The study consists of three stages. The first stage is to design a coda level magnitude scale, for which we followed the approach of *Rautian et al.* [1981]. This is described in full detail elsewhere [*Lemzikov and Gusev*, 1989]. Then a coda magnitude database was created, and, finally, the present analysis was carried out.

Coda Shapes and Stability of Their Slope

As initial data, we use seismograms of the regional Kamchatka network (Figure 1). Seismographs in this network employ three-component 1.2-s pendulums of VEGIK or SM-3 type, together with GB-IV type galvanometers of 0.07-s period, with photographic recording. The system response is

[1]Institute of Volcanic Geology and Geochemistry, Russian Academy of Sciences, Petropavlovsk-Kamchatsky, Russia.

Paper number 95JB01571.
0148-0227/95/95JB-01571$05.00

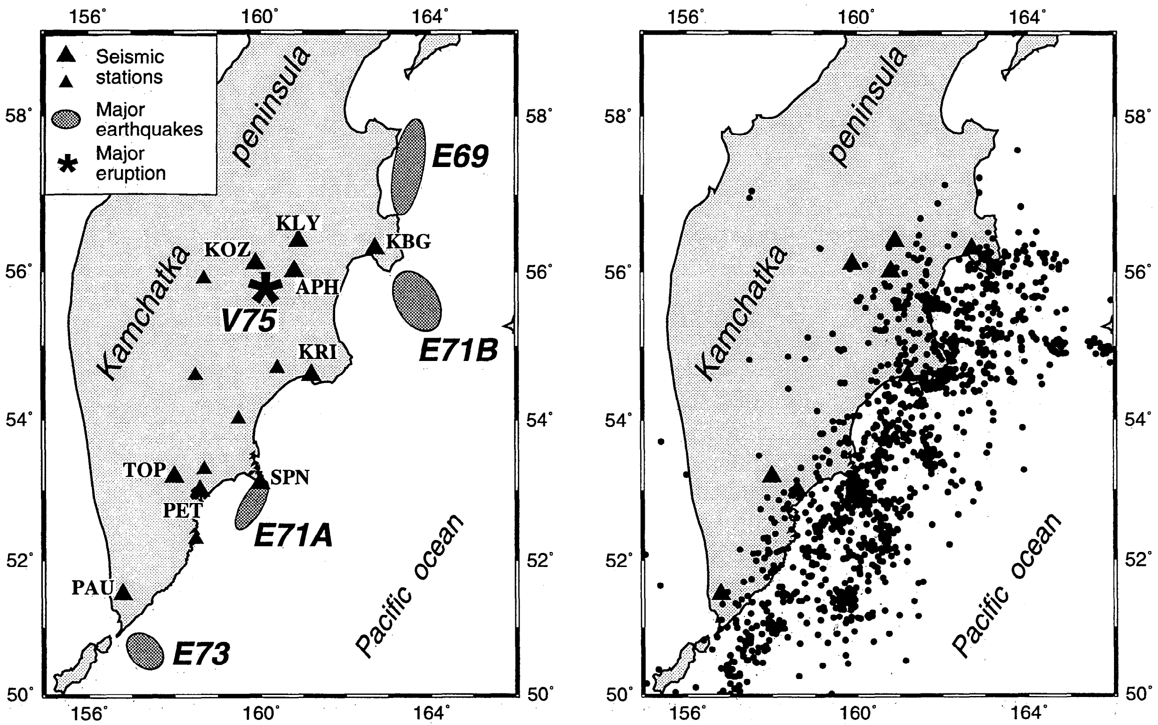

Figure 1. The region of the study. (left) Seismic stations (a larger triangle with the lettering denotes a station whose data are discussed here in detail) and major geophysical events marked by event codes. Refer to the text for detailed description of events. (right) Epicenters of ≈1600 earthquakes for 1967-1990 used in the study.

identical(±10-15%) for all stations and components and is practically flat for displacement between 1 and 10 Hz. The magnification is in the range 3-10 K. The typical visual frequency (measured as one-half the number of peaks and troughs over a 50-s window) of coda waves recorded by these instruments varies only weakly over the network. For example, it is 1.2-1.6 Hz at the lapse time t=75±25 s, 0.9-1.2 Hz at t=150±25 s and 0.75-1.1 Hz at t=250±25 s. (Lapse time here and throughout the paper is always measured from the origin time.) These frequency estimates are stable within the magnitude range studied in this paper. Such a narrow-band coda record is assumed to be due to a combination of intrinsic absorption in the propagation medium and the abrupt (f^3) low-frequency cutoff of the instrument response. A fast paper speed of 2 mm/s is used which results in good quality readable coda records. The complete calibration routine for these instruments was repeated once a year. In addition, at 8-hour intervals, a test signal from a stable-amplitude sweep-frequency generator was recorded on each component to check the stability of instrument response.

A segment of the tail part of a record of a near earthquake, from the time starting at (S arrival time + 1.5(S-P) time) until trace amplitude becomes twice as large as the microseisms (if such a segment could be found) was considered as the "coda window" within which the amplitude measurement was made. As a part of the data processing, we reduce each coda amplitude to the reference lapse time of 100 s. To minimize reduction errors, the amplitude is measured at a lapse time which is as near to 100 s as possible. We use the value of double amplitude 2A in μm of some clearly prominent excursion. This value is then reduced to the 100-s lapse time with the aid of a reference coda shape function. The standard coda shape function plays the same role as a calibration curve in the determination of a body-wave magnitude. To apply this reduction safely, it is necessary to check that coda shape does not depend on many possible distorting factors, such as station location, source depth etc.

Gusev and Lemzikov [1980] determined an empirical average coda shape function $a(t)$ for Kamchatka stations. Assuming it to describe real coda shapes accurately, one can reduce the value of 2A measured at some lapse time t to the reduced quantity $2A_{100}$ merely by

$$\log 2A_{100} = \log 2A - \log a(t) \tag{1}$$

where a(100 s)=1 by definition.

To check for the stability of the coda shape and to find out whether the empirical $a(t)$ function can be safely employed for magnitude determination purposes, let us consider the difference between some particular coda shape and the reference shape ($a(t)$) as a function of time, and expand this difference into Taylor's series around t=100 s. In this expansion, the constant term determines the absolute amplitude level, it is directly related to the magnitude value. It varies slightly for various stations and components. Using some particular station and component as the reference, one can determine corrections, to be used further for reduction of measurements to this reference station/component. (In our practice, the Z component of station PTR was chosen as the reference one.) To be consistent, we must assume that these station corrections does not depend on epicenter, depth, or magnitude, and this was verified successfully. Only temporal stability was lacking, thereby producing a basis for the present study. The linear term in the discussed expansion must vanish in theory, and in practice, it must be equal to zero on the average. To check this, we estimated α coefficient:

$$\alpha=\frac{d}{dt}(\log 2A(t)-\log a(t)) \quad (2)$$

through a regression [*Gusev and Lemzikov*, 1980]:

$$\log 2A(t)-\log a(t)=\alpha\, t+\beta. \quad (3)$$

Error components expressed by quadratic and higher terms are neglected. Thus we decide whether the employed $a(t)$ is an acceptable reference function based on how close to zero are average α values.

Lemzikov and Gusev [1989] examined α estimates for many earthquake records grouped by component, source depth, magnitude, station location and (calendar) time. Data for 1968-1984 were used from 12 stations and averaged separately for each station, for two periods, 1968-1973 and 1978-1984, and for four depth intervals. Almost for all data groups, and for all in the depth ranges of 0-60 and 60-120 km, average α values are in the range $\pm0.6\times10^{-3}$ s^{-1}, with estimated rms errors of these averages of the same order: $0.5\text{-}1\times10^{-3}$ s^{-1}. Hence average α in any subgroup was insignificantly different from zero. Based on these results, one can assume the true α values to be limited by the bounds $\pm0.5\times10^{-3}$ s^{-1}. To determine the related error in $2A_{100}$ estimate by (1), note that our $2A$ measurements were made within the lapse time window of 70-230 s. Therefore the reduction by (1) was never done over a time interval larger than 130 s, so that the bounds for the error in $\log 2A_{100}$ caused by station or depth dependence of $a(t)$ are $130(\pm0.5\times10^{-3})=\pm0.07$.

Systematic temporal variations of α have been revealed for several Kamchatka stations during some anomalous periods [*Gusev and Lemzikov*, 1980, 1984, 1985]. These can produce systematic deviations of $2A_{100}$ values. The actual average amplitude measurement time was about t_{av}=135 s. During a typical anomaly, α changed by $\delta\alpha=-3\times10^{-3}$ s^{-1}. However, the detailed study of temporal α anomalies has shown that they are localized mainly within the lapse time interval of 50-120 s. Thus the real deviation of α in the interval 100-135 s can be assumed to be equal to one-half of this value. This gives an estimate of $0.5(-3\times10^{-3})(135\text{-}100)\approx-0.05$ for this type of systematic error in $\log 2A_{100}$.

One can expect α to be magnitude dependent. The smaller the magnitude is, the higher the source corner frequency and the higher the average coda frequency. As wave attenuation usually increases with frequency, one can expect that the coda decay rate, measured by α value, will increase with decreasing magnitude. This effect was found to be real and well expressed, but it was observed only for magnitudes below a well-defined critical value, equivalent to $m_b\approx3.7$. The value of α equals to 0 for $m_b\approx4\text{-}6$, about -1×10^{-3} s^{-1} at $m_b\approx3.3$ and about -2.5×10^{-3} s^{-1} at $m_b\approx2.9$. The empirically determined critical magnitude value was later used to set a lower magnitude threshold in data selection.

Thus, the distorting effects of systematic variations of coda decay rate were quantitatively examined and found to be rather weak. As for the constant level differences, they were easy to compensate by appropriate constant corrections described below.

Lemzikov and Gusev [1989] reduced $\log 2A_{100}$ to the standard regional "energy class" magnitude scale K_S [*Fedotov*, 1972] which is based on S wave amplitude. (K_S is defined as $2\log(A_S/T)$+const at a given distance; it is closely correlated with m_b; $K_S\approx2m_b+2.1$ at $K_S<13$.). The linear regression of $\log 2A_{100}$ on Z component of station PET versus K_S gives the relation

$$K_S=1.6\log 2A_{100}+11.0\pm0.4\,. \quad (4)$$

The new coda level magnitude K_C was defined as the best approximation to K_S, by the similar relation:

$$K_C=1.6\log 2A_{100}+11.0+C_s+C_c \quad (5)$$

where C_s is the station correction ($C_s\equiv0$ for PET) and C_c is the component correction ($C_c\equiv0$ for the vertical component).

All the above do not apply to BKI station situated on Bering island. Because of T phase contamination [*Gusev and Lemzikov*, 1980; *Rautian et al.*, 1981], coda shape of earthquakes recorded at this station is not standard. Furthermore, both relative coda level and K_C-K_S difference at this station have unusually large variance. For this reason, BKI was excluded from this study.

Using a simplified version of analysis of variance, we estimated the contributions of various factors to the variance of an individual corrected K_C value for a particular station/component. Express this value as a sum:

$$K_C = K_C^* + \delta_p + \delta_c + \delta_s \quad (6)$$

where K_C^* is the "true" magnitude value and δ_p, δ_c, and δ_s are independent random errors with zero means and variances σ^2_p, σ^2_c and σ^2_s respectively. These errors are related to the following sources: δ_p ("interpeak") accounts for the choice of individual prominent peak in coda window, $\sigma^2_p\approx0.2^2$; δ_c ("intercomponent") accounts for the choice of a particular component, $\sigma^2_c\approx0.1^2$; δ_s ("interstation") accounts for the choice of a particular station, $\sigma^2_s\approx0.18^2$. The last parameter measures, in effect, the quality of station corrections: without corrections, $\sigma^2_s\approx0.4$. These estimates enable us to determine the accuracy of single-station (three-component average) K_C value:

$$\sigma_1^2=\sigma_s^2+\tfrac{1}{3}(\sigma_p^2+\sigma_c^2)\approx 0.22^2 \quad (7)$$

and of a typical five-station network average

$$\sigma_5^2=\tfrac{1}{5}\sigma_1^2\approx 0.10^2\,. \quad (8)$$

Note that differences between K_C and K_S are much larger and are specified by the rms error of 0.4. Therefore the S wave magnitude is markedly inferior to the coda level magnitude as a means of reference. All the listed σ values can be converted to errors in $\log 2A_{100}$ if divided by 1.6. In particular, an one-station $\log 2A_{100}$ value has the rms error of 0.14, to be compared to the typical rms magnitude residual of 0.3.

Measurement, Processing and Data Selection

Based on the technique described above, routine data processing was carried out, in the following manner. To determine station K_C value, at first the coda window is determined and a prominent coda excursion is found according to the procedure described above (if possible). Then the measurement of $2A$ is made by a ruler, $2A$ is converted into μm using nominal magnification of the instrument, and the lapse time t of the excursion is determined. The $(2A, t)$ pair is an input to the specially designed nomograph giving the component K_C value. After measurements has been done on Z, N-S and E-W records, component corrections are added and the three K_C estimates are averaged. This procedure is applied to all stations with a nonempty coda window, i.e., when the expected coda was not contaminated by microseisms. Then for each station, the K_C value is combined with the station correction giving the station K_C estimate, and all results are averaged to produce the final network K_C value.

Data selection was based mainly on magnitude (down to K_S=10.5, corresponding to $m_b \approx 4.2$ or $M_L \approx 4.4$) for reasons explained earlier. Depth was limited to the range 0-50 km. Events in dense swarms were intentionally decimated so as not to give excessive weight to data from a localized source volume. We used data from nine permanent stations (namely, PAU, PET, TOP, SPN, KRI, APH, KLY, KOZ, and KBG) out of approximately 17 which were operational in Kamchatka during various subperiods in the 24-year study period. Data from other stations were used for network determination of K_C only. Typically, there were five to eight stations per event for calculating K_C values (range between 3 and 12). For the study period of 1967-1990, about 1600 events were processed, and for each permanent station we could determine 550 to 1000 individual K_C values for the 0-50 depth interval.

For each event and station, we compute the magnitude residual

$$\Delta K_C = K_C(\text{station}) - K_C(\text{network}) . \tag{9}$$

We treat ΔK_C values as a time series, dividing them into successive 12-event nonoverlapping groups for statistical analysis. Data within a group were sorted, and the median and the interquartile width values were used to estimate the mean and the standard deviation.

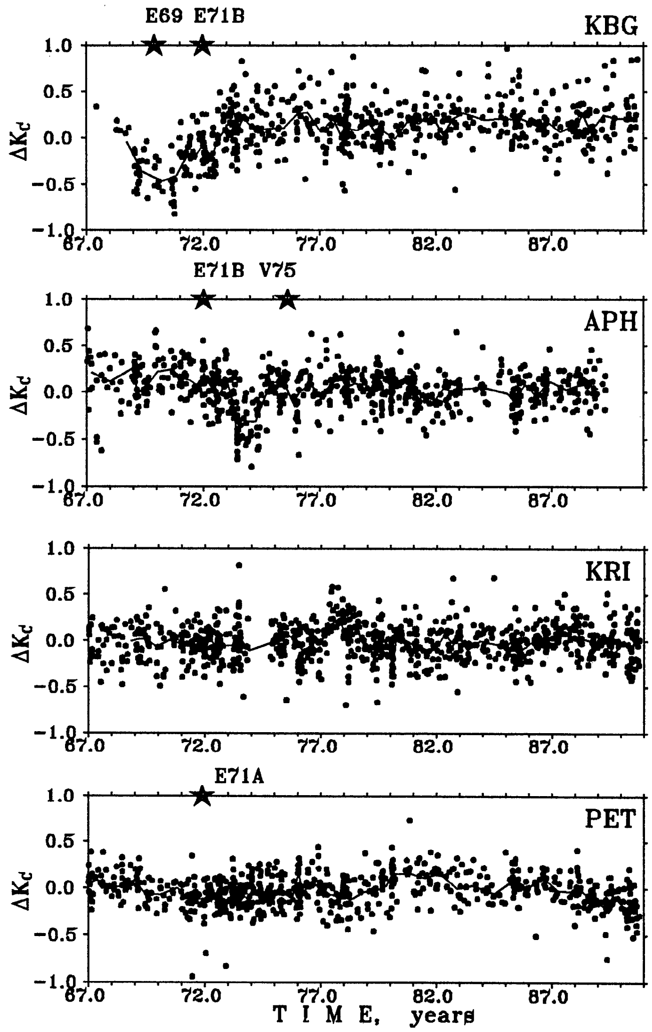

Figure 2. Individual ΔK_C values for four stations for 1967--1990 and their twelve-point average with no overlapping. Stars here and below denote major geophysical events in the vicinity of each station (see text and Figure 1).

Results

Figure 2 shows ΔK_C data for four representative stations. Note that in spite of the scatter, discernible temporal variations in average ΔK_C can be seen. However, except for APH during 1973-1975 and KBG during 1967-1973, these variations are weak. They are quite prominent within the mentioned periods at these stations. For the other five stations, plots are similar to those for PET and KRI, with no prominent features. Figure 3 shows the grouped data for all of the nine stations. The two prominent negative anomalies at APH and KBG show more than 2σ difference relative to the background, suggesting further investigation.

There are two different questions which necessitate different checks: first, whether any temporal variations are present at all and, second, whether the individual anomalies are real. It is clear a priori that the character of data precludes the possibility of identifying short-term anomalies because they cannot be distinguished from noise. Hence we confine our study to variations that are of long enough duration. To demonstrate the presence of this kind of variation, one can apply the standard analysis of variance and compare intergroup and within-group variances. The ratios of these variances (F ratios) were calculated for each station. They can be compared to the upper critical value $F(m\text{-}1, m(n\text{-}1), Q)$ where m is the number of groups, n is the group size and Q is the significance level. For n=12 and mn=550-1000, all $F(m\text{-}1, m(n\text{-}1), Q)$ are below $F(50, 500, Q)$, which is equal to 2.22 for Q=0.05% and to 1.88 for Q=0.5%. Out of nine calculated F ratios, eight are above 2.22 and one (for PAU) is above 1.88. Hence the answer to the first question can be considered to be positive.

The second question is whether the two particular baylike anomalies at APH and KBG are real. To check this, we separated 12-event groups into two subsets, "normal" and "anomalous", by eye. For APH, the "anomalous" subset consisted of the four adjacent groups forming the anomaly, all other groups were included into the "normal" subset. For KBG, the procedure was similar, only the number of "anomalous" groups was 7. Initial grouping was not revised in any way. To compare individual data of both subsets, the Kolmogorov-Smirnov test was applied. For each of the two stations, all ΔK_C values from "anomalous" groups were combined, to form one sample, and the values from "normal" groups were combined to form the second sample. The zero hypothesis of identity of distribution laws for these two

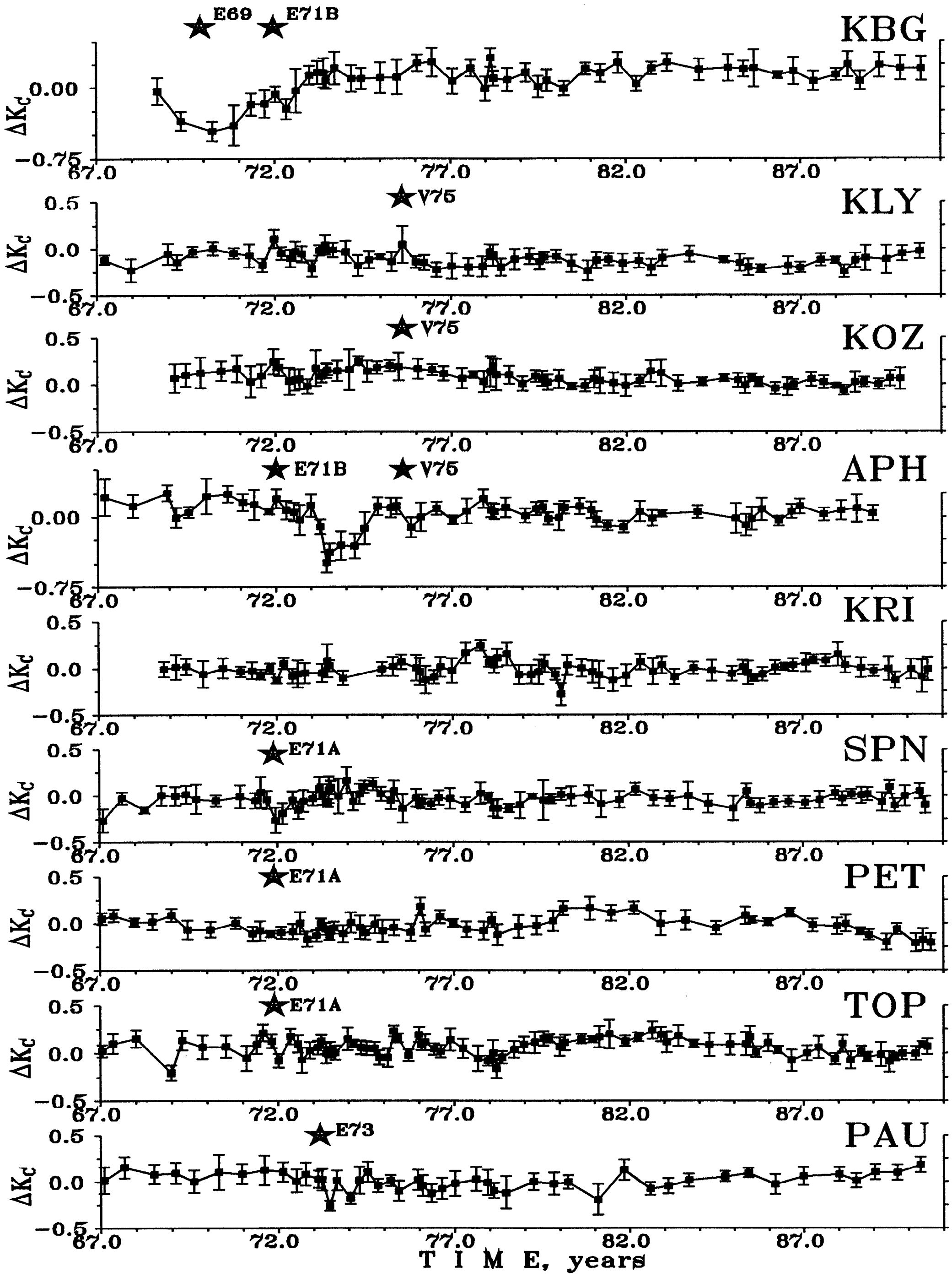

Figure 3. Successive twelve-point averages of individual ΔK_C data (with no data overlapping) for nine Kamchatka stations, with error bars denoting $\pm 1\sigma$ for each average. Continuous moderate oscillations at all stations as well as prominent anomalies at KBG and APH are seen.

samples is rejected at significance levels below 10^{-12} in each case.

However, this significance value could be accepted only if the compared subsets were chosen in a manner independent from the researcher (e.g., by some random mechanism). In fact, opposite is true: the "anomalous" subsets were intentionally chosen so as to maximize significance, thus introducing strong bias into the determined significance level. To account for this, we should add a factor to the significance value, representing the number of possible choices. To determine this number, consider the case of the zero hypothesis. Then we can assume that a fictitious "anomaly" can begin and end at any point, giving, for the grouped data of Figure 3, about N^2_{group} variants of such an "anomaly". For $N_{\text{group}}=49$, the factor of ≈ 2500 reduces significance level to less than 10^{-8}. Thus the significance is really high. Hence the answer to the second question is again positive.

Thus we have confirmed the presence of temporal changes in general, for all stations, as well as the reality of the two particular anomalies at APH and KBG. A few smaller anomalies can be detected visually on the graphs for KRI, SPN, and PAU. Their significance was not studied. As an additional informal check we applied the same procedure to data from earthquakes in the 50-150 depth range (Figure 4). Observe that the anomaly at KBG is practically identical to the previous case studied. However, for the APH data (Figure 5), no clear anomaly is seen, perhaps due to the lack of data in the critical time window. Hence no definite judgement is possible in this case.

Are the Anomalies of Natural Origin?

No statistical check of temporal anomalies can guarantee against time-varying systematic errors. Among those errors, variations of ΔK_C due to an instrument calibration error are prominent and, if present, would immediately appear as a fictitious anomaly. However, the same anomaly will appear in the value of K_S of this particular station. Network-average K_S magnitudes are determined routinely, so that the outlier for a particular station will not go unnoticed if it is large enough. Practically, one can safely assume that errors in excess of 0.3-0.4 units of K will be detected. Also, any pronounced change in the instrumental parameters is easily checked by visual inspection of the control signal recorded on each component of each seismogram. Because of poor quality of some seismic vaults and severe weather conditions, transfer function changes do sometimes take place. However, they are identified and corrected on a routine basis. Also, man-made calibration errors are possible. Any such error should produce an abrupt and steplike anomaly. This is not observed. Finally, the possibility of identical calibration errors on all three components of a station is a low probability. Figures 4 and 5 demonstrate individual data for the three components of KBG and APH showing similar anomalies on each component. Such check was done for all stations. In one case, for the station KII (not included into the group studied) we found a single step of 0.25 on only one component, probably reflecting a real calibration error. Several other contributing factors were analyzed and checked, such as systematic spectral variations of small earthquakes, the drift of the eigenperiod of the pendulum, variations of average earthquake depth or magnitude. All were found to be negligible.

An important source of systematic variation is the anomalous coda decay, but this cannot be considered as error

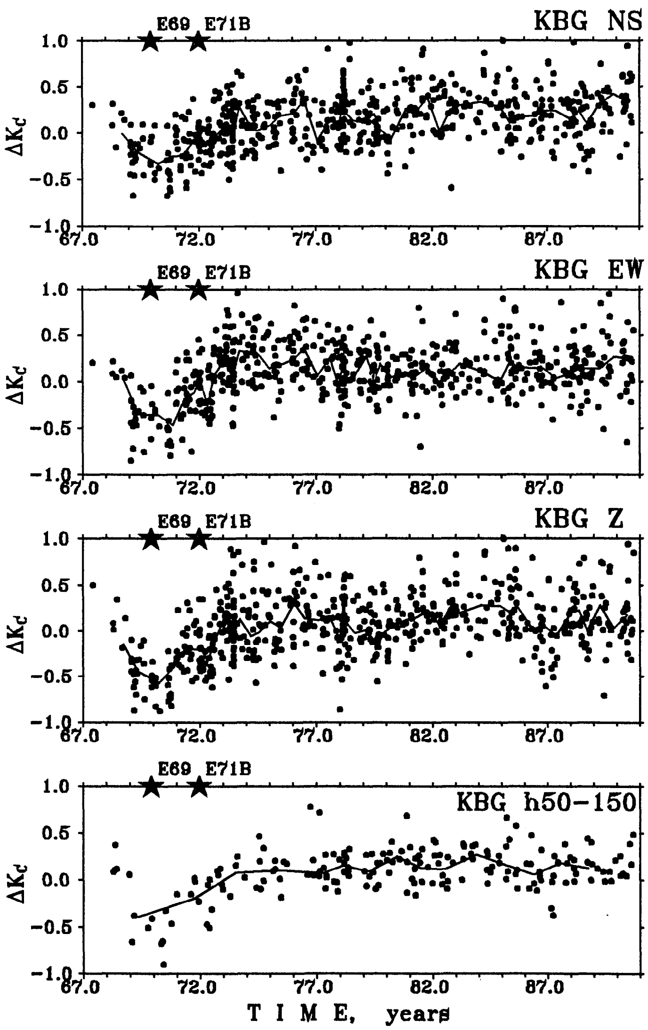

Figure 4. Single-component ΔK_C values of KBG station for the same events as shown in Figure 2 (three top graphs), and three-component-average ΔK_C values for data of the 50-150 km depth interval (bottom graph). Twelve-point averages are also given.

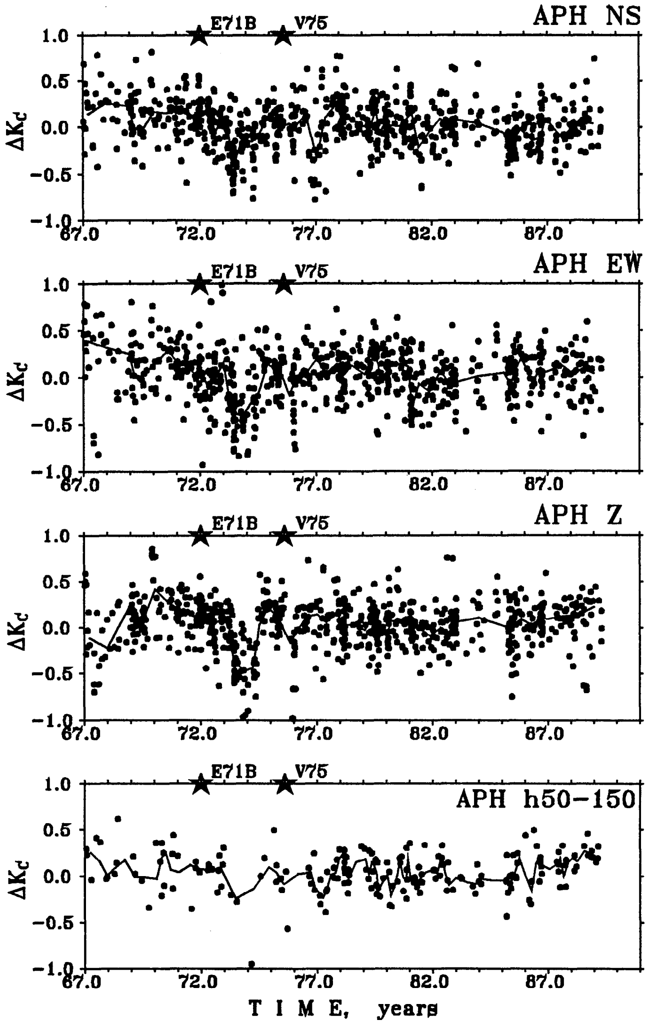

Figure 5. Same as Figure 4, for station APH.

proper in the present context: both coda decay and coda level variations can be thought of as different manifestations of the same phenomenon: temporal change of scattering and/or attenuating properties of the medium. If only this phenomenon is real one and not the artifact, the inability of our present technique to distinguish accurately between these manifestations is of secondary importance.

To exclude conclusively the possibility of calibration-related systematic error as a possible source of anomalies, one more check was done for the KBG and APH data. Instead of using network-average K_C as a background against which the temporal variations of the relative coda level of a particular station are determined, we used the *S* wave amplitudes of the same station. More precisely, we used the individual *S* wave station magnitude K_S as a reference for the individual K_C value of the same station. In this case, possible errors in station calibration would, in general, cancel. Note, however, that (1) K_S magnitude is defined as log *A/T*, and not log *A*; (2) it is related to a different spectral band (typically, 1.5-4 Hz instead of 0.7-1.4 Hz for K_C); and (3) it is measured on one of the horizontals, instead of averaging over three components. Hence some differences could be expected, due merely to procedural differences. Also, any systematic variations of *S* wave *A/T*, such as, for example, caused by systematic nodal plane variations of small earthquakes, or by their spectral variations (definitely known to take place in this case, see e.g. *Gusev and Lemzikov* [1984]) will additionally contribute to K_C-K_S difference. Apart from these general considerations, we know from observations that *S* wave magnitude is much less stable than the coda magnitude. Thus it would be too optimistic to expect any strong correlation between ΔK_C and K_C-K_S. However, some correlation can be seen (Figure 6), in particular, a positive ramp in 1969-1973 for KBG and a negative step in 1973-1974 for APH. We consider this partial correlation as an important argument confirming reality of temporal anomalies of the coda level.

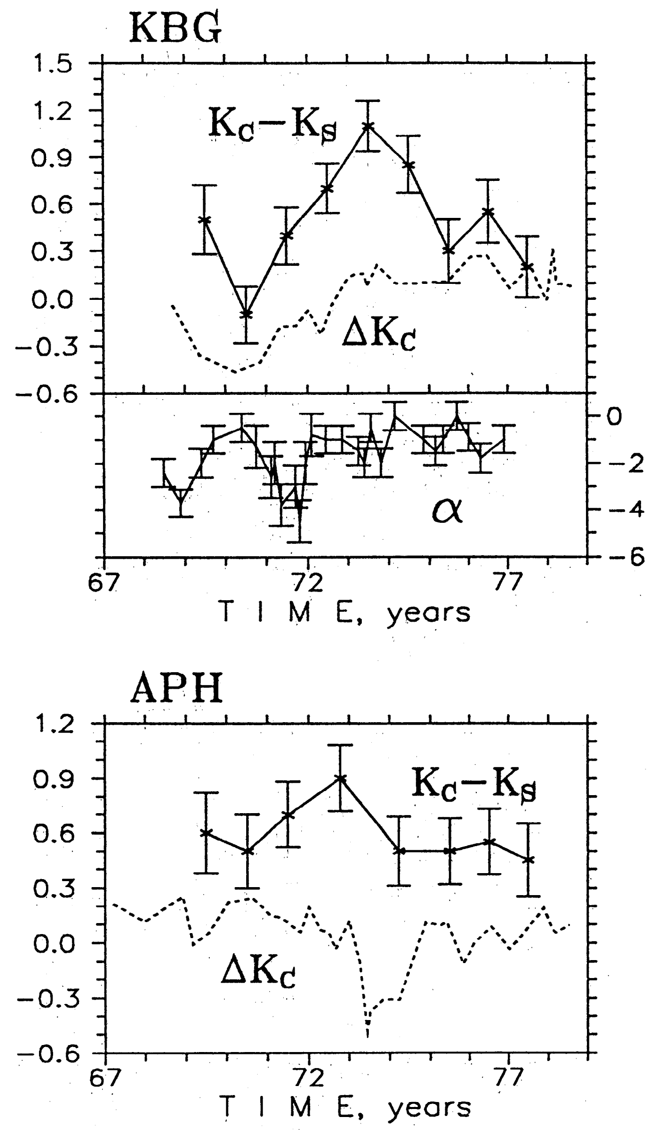

Figure 6. The trends of the difference K_C-K_S of single-station coda and *S* wave magnitudes (believed to be free from calibration errors but noisy) for the stations KBG and APH, for 1969-1978, compared with the ΔK_C trends reproduced from Figure 3. Presented are 1-1.5-year averages of K_C-K_S, over groups containing about 30 events each ($\pm 1\sigma$ error bars for the average are given). For KBG station, average α values (right scale, in units of 10^{-3} s^{-1}) measured on its vertical component are also plotted [after *Gusev and Lemzikov*, 1984].

Probable Precursory Character of the Baylike Anomalies

In 1967-1992, the most remarkable geophysical events on Kamchatka were the following (alphanumeric code after the number is an event identification on the figures):

1. (E69). The earthquake and tsunami ($M_w \approx M_t \approx 7.75$, M_s=7.3, surface -focus) of November 22, 1969. The nearest station is KBG, about 80 km to the southern (closest) part of the source.
2. (E71A). The earthquake (M_w=7.65, h=100 km) of November 24, 1971. The nearest stations are SPN and PET, at epicentral distance of about 60 km.
3. (E71B). The earthquake (M_w=7.7, surface focus) of December 15, 1971. The nearest station is KBG, at about 60 km to the NW (closest) part of the source.
4. (E73).The earthquake ($M_w \approx 7.3$, h=70 km) of February 28, 1973, at North Kurile islands. The nearest station is PAU at the epicentral distance of 120 km.
5. (V75). The volcanic eruption (fissure type) near the Tolbachik volcano which began in July 1975, and continued for 2 years, with the volume of its products $V \approx 2.5$ km^3. The nearest stations are APH and KOZ both at a distance of about 70 km.

All other earthquakes during 1967-1991 on Kamchatka had M_w below 7.2, and all other eruptions had V below 0.3 km^3. Therefore the five listed Kamchatka events constitute a naturally selected set. This greatly simplifies the task of analysis of possible precursory meaning of the ΔK_C anomalies. All the five events are shown on Fig.1 and are marked by stars on Figures 2, 3 and 4.

The events and the anomalies can be associated in a straightforward way. An anomaly at KBG during 1969-1972 preceded two earthquakes of 1969 (E69) and 1971 (E71B) near this station. The anomaly at APH in 1973-1974 preceded the eruption of 1975 (V75). The intermediate-depth earthquakes, the one of November 24, 1971 (E71A), as well as the weaker and more distant event in North Kuriles (E73) have no associated precursory anomalies.

In addition to the preseismic anomalies, postseismic anomalies of moderate amplitude can be seen after three earthquakes: November 24, 1971 (E71A), on SPN; February 28, 1973 (E73), on PAU; and also after December 15, 1971 (E71B) on KBG (in which case the anomaly is merged with the preseismic one). All three postseismic anomalies are of the same (negative) sign and of comparable amplitude and duration.

Physical Nature of Anomalies

Gusev and Lemzikov [1980, 1984, 1985] provide evidence for the temporal change of the coda decay parameter α. As described above in detail, it measures the deviation of the slope of a particular coda record from the slope of a reference coda shape. In those earlier studies, smaller earthquakes were selected out of a limited zone around the future epicenter of the December 15, 1971, event (E71B). Whereas in the present study the earthquakes had their epicenters all over the region. Thus the typical coda window was around 80 s in that study, against 135 s in the present study. Of the stations studied earlier, KBG, KRI, and KLY are common with those used in the present study. The earlier study showed two clear negative α anomalies at KBG in 1969 and 1971-1972; taken together, they approximately match the ΔK_C anomaly of 1969-1972 found at the same station in the present study (Figure 6). The signs of both anomalies agree in their physical meaning: they indicate the increase of shear wave attenuation. As for KRI, it too showed a negative anomaly of α in 1971, but ΔK_C values in the present study were practically constant. KLY showed no anomalies, in either α or ΔK_C.

Gusev and Lemzikov [1980, 1984, 1985] assumed the α change to arise due to the local Q_S decrease and estimated its value to be roughly 20%. Now we make a new estimate based on the ΔK_C data, which correspond to somewhat larger volume than one probed by *Gusev and Lemzikov* [1980, 1984, 1985]. Assume that the Q_S change takes place around the station so that during the last 100 s of energy propagation from a hypocenter to the station, scattered S waves move through the changed medium. At f=1.1 Hz, assuming Q_S=150, the expected attenuation of amplitude in "normal" period is $\delta \log A = 0.43 \delta \log_e A = 0.43\pi f t / Q_S \approx 1.0$. On this background, we observe additional "anomalous" $\delta \log A \approx 0.3$ (ΔK_C anomaly of about -0.5, divided by 1.6). Therefore, we can crudely estimate the attenuation increase as 30%, somewhat larger than in the earlier study. This difference should not be considered as meaningful: the order of magnitude is the same, and the difference seemingly results merely from greater sensitivity of the present technique.

Discussion

The amplitude level of the late coda at a station has been shown to be intrinsically a very stable parameter, enabling one to use it to develop an accurate magnitude scale. In terms of the logarithm of the amplitude, the standard deviation is 0.14 for one station and 0.065 for the average over a five-station network. We believe that this stability is due to the effective suppression of amplitude variations caused by differences in earthquake nodal plane orientation. The theory of single scattering [*Sato*, 1984] predicts prominent variations of coda shape due to this factor. However as lapse time increases and the multiple-scattering model [*Gusev and Abubakirov*, 1987] replaces the single-scattering model, coda shape must become independent of source radiation pattern. The threshold lapse time (when average multiplicity of scattering is unity) is equal to the mean free time MFT=$1/cg$, where c is wave velocity and g is turbidity. Ideally, the measurement time t should be large compared to MFT. Actually, t is around 130 s, and MFT is about 40 s [Abubakirov and Gusev, 1990], so that the average value of multiplicity is well above unity and, apparently, large enough to significantly reduce the errors of the type discussed above.

Our results can be compared to those of *Jin and Aki* [1989, 1993], who revealed clear temporal variations of logarithmic slope of decay of the late coda in California, thought to reflect temporal variations of Q_S. In their data, continuous oscillating variations as well as large individual excursions can be seen, closely matching the mode of variations found here.

I am well aware that temporal variations of the properties of the lithosphere, as revealed in coda parameters, nowadays provoke skepticism quite comparable to the optimism of the 1970s. However, I believe that results based on ample data, and of high formal significance, deserve certain attention.

Conclusion

Using high-accuracy coda magnitude data, we could monitor relative variations of the coda level as expressed in magnitude residuals. Based on 500-1000 small earthquake records for each of nine Kamchatka stations for 24 years (1967-1990) of observation, two types of temporal variation of magnitude residuals were found: first, small but significant background variations with a period of one to several years at all nine stations; second, superposed on this background, two prominent (-0.3 in terms of log amplitude) and at the same time significant ($p<10^{-8}$) anomalies of 1.5- and 3.5-year duration, each anomaly on a plot of one station. They preceded two out of two $M \approx 8$ shallow - focus earthquakes and one out of one of large ($V>1$ km^3) volcanic eruption that occurred on Kamchatka during the period of study. The lead times are 0.8 and 2.7 year for the earthquakes and 2.4 year for the eruption. The apparent cause of anomalies is a decrease of lithospheric Q for shear waves of the order of 30%.

Acknowledgments. The author is indebted to the administration of the Institute of Volcanology and its seismology team for providing access to the archive of seismograms. The technical help of Raisa Pyrina and Valentina Zubova is greatly appreciated. Data processing was partly conducted during the author's visit to the University of Southern California supported by Joint American--Soviet Cooperation Program in the Environmental Sciences. Financial support by the Russian Foundation for Fundamental Research, (RFFI grant 93-05-8514) is gratefully acknowledged. Useful comments of anonymous reviewers and H. M. Iyer helped to improve the manuscript. I am especially grateful to H. M. Iyer, who, as a Volunteer Editor of JGR-Solid Earth, also greatly improved my English.

References

Abubakirov, I. R., and A. A. Gusev, Estimation of scattering properties of lithosphere of Kamchatka based on Monte-Carlo simulation of record envelope of a near earthquake, *Phys. Earth. Planet Inter., 64*, 52-67, 1990.

Aki, K., Analysis of the seismic coda of local earthquakes as scattered waves, *J. Geophys. Res., 74*, 147-156, 1969.

Aki, K., Theory of earthquake prediction with special references to monitoring of the quality factor of lithosphere by coda method, *Earthquake Pred. Res., 3*, 219-230, 1985.

Aki, K., and B. Chouet, Origin of coda waves: source, attenuation and scattering effects, *J. Geophys. Res., 80*, 3322-3342, 1975.

Chouet, B., Temporal variation in the attenuation of earthquake coda near Stone Canyon, California, *Geophys. Res. Lett., 6,* 143-146, 1979.

Fedotov, S. A., *Energy Classification of Kurile-Kamchatka Earthquakes and the Problem of Magnitudes*(in Russian), 116 pp., Nauka, Moscow, 1972.

Gusev, A.A., and I.R. Abubakirov, Monte-Carlo simulation of record envelope of a near earthquake. *Phys. Earth. Planet Inter., 49*, 30-36, 1987.

Gusev, A. A., and V. K. Lemzikov, Preliminary results of the study of coda envelope shape variations of near earthquakes before the 1971 Ust-Kamchatsk earthquake (in Russian), *Vulkanol. Seismol., 6,* 82-93, 1980.

Gusev, A. A., and V. K. Lemzikov, Anomalies of coda wave parameters of small earthquakes before three large earthquakes of Kurile-Kamchatka zone (in Russian), *Vulkanol. Seismol., 4,* 76-90, 1984.

Gusev, A. A., and V. K. Lemzikov, Properties of scattered elastic waves in the lithosphere of Kamchatka: parameters and temporal variations, *Tectonophysics, 112,* 137-153, 1985.

Jin, A., and K. Aki, Temporal change in coda Q before the Tangshan earthquake of 1976 and the Haicheng earthquake of 1975, *J. Geophys. Res., 91,* 665-673, 1986.

Jin, A., and K. Aki, Spatial and temporal correlation between coda Q^{-1} and seismicity and its physical mechanism, *J. Geophys. Res., 94,* 14,041-14,059, 1989.

Jin, A., and K. Aki, Temporal correlation between coda Q^{-1} and seismicity- Evidence for a structural unit in the brittle-ductile transition zone, *J. Geodyn., 17,* 95-120, 1993.

Lemzikov, V. K., and A. A. Gusev, Energy classification of near Kamchatka earthquakes using coda wave level (in Russian), *Vulkanol. Seismol., 4,* 83-97, 1989.

Malamud, A. S., On a possible precursor of a strong earthquake (in Russian), *Dokl. Acad. Nauk Tajik. SSR, 17,* 31-34, 1974.

Rautian, T. G., V. I. Khalturin, M. S. Zakirov, A. G. Zemtsova, A. P. Proscurin, B. G. Pustovitenko, A. N. Pustovitenko, L. G. Sinelnikova, A. G. Filina, and I. S. Shengelia, *Experimental Studies of Seismic Coda,* 142 pp., Nauka, Moscow, 1981.

Sato, H., Attenuation and envelope formation of three-component seismograms of small local earthquakes on randomly inhomogeneous lithosphere, *J. Geophys. Res., 89,* 1221-1241, 1984.

Sato, H., Temporal change in attenuation intensity before and after eastern Yamanashi earthquake of 1983 in Central Japan, *J. Geophys. Res., 91,* 2049-2061, 1986.

Tsimura, T., Determination of earthquake magnitude from total duration of oscillation, *Bull. Earthquake Res. Inst., 15,* 7-18, 1967.

A. A. Gusev, Institute of Volcanic Geology and Geochemistry, 9 Piip Boulevard, Petropavlovsk-Kamchatsky, 683006, Russia. (e-mail: seis@volgeo.kamchatka.su)

(Received May 26, 1994; revised May 9, 1995; accepted May 17, 1995.)