Separation of scattering and intrinsic attenuation for the Kanto–Tokai region, Japan, using measurements of S-wave energy versus hypocentral distance

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SUMMARY
Determining the relative amount of attenuation caused by scattering and intrinsic absorption is important to the understanding of wave propagation and attenuation in the heterogeneous lithosphere. A recently developed model based on radiative transfer theory provides a method for quantifying the ratio of scattering loss to total attenuation, which is called seismic albedo. The basic assumption of this model is that the medium is composed of a uniform distribution of isotropic scatterers. The method is based on a theoretical result showing that the variation with source–receiver distance in the seismic energy integrated over time is directly related to albedo and total attenuation, $Q_s^{-1}$. We introduce an improvement in the previously used method which results in more reliable estimates of seismic albedo and $Q_s^{-1}$, which can be used to calculate the losses due to scattering and intrinsic absorption. We call our new method multiple lapse-time window analysis. The improvement is based on the observation that the relationship between integrated energy and distance is strongly dependent on the time duration over which energy is integrated. We show that parameters describing media attenuation can be estimated from measurements of two ratios from the integrated energy versus distance relations compiled using two time windows for integration. One ratio is the energy integrated from 0 to 15 s after the S-wave arrival observed at 50 km source–receiver distance divided by the energy in the same time interval observed at 150 km distance. The second ratio is the energy integrated from 0 to 15 s observed at 150 km divided by that from 30 to 100 s observed at the same source–receiver distance.

Integrated energy calculated for many source–receiver pairs may be corrected for relative site amplification and relative source amplitude determined using the coda-wave method. These corrections allow us to use data from many source–receiver pairs to find a well-constrained energy versus distance relation. Site amplifications relative to a reference station are calculated for three frequency bands by determining the ratio of the spectral amplitude in each band at one station for a 10 s time window to that at the reference station in the same 10 s time window. Ratios are calculated for many 10 s time windows for each of 10 events. For each station, we found little scatter in the ratios among the windows and events used. The average of all the ratios obtained for each station was taken to be the site amplification relative to the reference station.

We applied the multiple lapse-time window analysis method and source-site correction procedures to data from the Kanto–Tokai region of Japan and found that intrinsic attenuation $Q_s^{-1}$ is larger than scattering attenuation $Q_s^{-1}$ over three frequency bands; 1–2, 2–4, and 4–8 Hz. We found that estimates of coda-wave attenuation $Q_s^{-1}$ made using the coda-wave decay method, are similar to the
intrinsic attenuation $Q_i^{-1}$ for the frequency range of 2–8 Hz. We were unable to fit data for a third integration time window for the lowest frequency band studied. Our inability to fit all the data might be due to the assumptions that scattering is isotropic, that scatterers are uniformly distributed, and that only body waves are observed. We found that the site amplifications and seismic albedos determined for vertical and north–south components of motion are similar.

**Key words:** attenuation, Japan, scattering.

**INTRODUCTION**

Attenuation is caused by two factors; scattering and intrinsic absorption. Quantifying the contribution of scattering to total attenuation has been a subject of considerable interest to seismologists (Aki, 1980a, b; Taylor, Bonner & Zandt 1986; Jacobson 1987; Frankel & Wennerberg 1987). Arguments have been made that scattering is responsible for a majority of the observed attenuation (Aki 1980a, b) or that intrinsic mechanisms that convert elastic energy into heat dominate the observed attenuation (Frankel & Wennerberg 1987). Before realistic models of the attenuation process can be developed and used to interpret data, there is need for definitive measurements of the relative contributions of scattering and intrinsic attenuation to total attenuation in various frequency bands.

Recently, Wu (1985) introduced a method based on the radiative transfer theory (Ishimaru 1978) for determining the relative contributions of scattering and intrinsic absorption to total attenuation. His method is based on a theoretical model of energy propagation in a randomly heterogeneous elastic media that has no spatial variation in the heterogeneity. Scattering is assumed to be isotropic and multiple scattering is included. Two parameters describe the media in Wu's model. These are the scattering coefficient $g$ and the intrinsic quality factor $Q_i$. The scattering coefficient is the inverse of the scattering mean free path and is related to scattering $Q_s$ by

$$Q_s = \frac{\omega}{gV}$$  \hspace{1cm} (1)

where $\omega$ is angular frequency and $V$ is seismic velocity. Wu's modelling resulted in a set of characteristic curves showing the distribution of seismic energy versus distance from the source in the medium. Wu's results can be considered to represent the distribution of energy with distance from a continuously radiating source or, alternatively, as the distance dependence of the energy integrated over infinite time due to an instantaneous source. Each of the curves shows the seismic energy versus scaled distance from the source for a single value of seismic albedo, which is the ratio of the scattering attenuation to total attenuation

$$B_o = \frac{g}{g + \omega(Q,V)^{-1}} = \frac{Q_i^{-1}}{Q_i^{-1} + Q_s^{-1}}$$  \hspace{1cm} (2)

where $B_o$ is seismic albedo, which ranges from 0 to 1. Distance is scaled by $g + \omega(Q,V)^{-1}$, which is called the attenuation extinction coefficient. The curves developed by Wu (1985) each have a characteristic distance dependence that depends on $g$ and $Q_i$. Thus, by integrating the seismic energy over time at various source–receiver distances from earthquakes in some region, the distance dependence of integrated seismic energy may be obtained and compared with Wu's curves to determine $g$ and $Q_i$ for the region.

Wu & Aki (1988) applied the approach of Wu (1985) to estimate the relative amounts of scattering and intrinsic attenuation in the Hindu Kush region in various frequency bands. Their approach was to select a 32 s window on each seismogram beginning with the S-wave onset and compute the Fourier transform. They averaged the square of the spectrum over the desired frequency bands. Since many earthquakes recorded at a single station were studied, the energy for each event was corrected for source magnitude using the coda-wave technique (Aki & Chouet 1975). The plots of energy versus distance for each frequency band were then compared with the curves developed by Wu (1985). Wu & Aki found that seismic albedo is $\leq 0.5$, which means that scattering is not the dominant mechanism of attenuation in the Hindu Kush. Toksöz et al. (1988) measured seismic albedo in Canada and the northeastern United States using a method similar to that employed by Wu & Aki (1988) and found that scattering is the dominant mechanism of attenuation in this region. The seismic albedo found by Toksöz et al. (1988) ranged between 0.8 and 0.9. More recent studies by Mayeda, Su & Aki (1991) and McSweeney et al. (1991) of data recorded in Southern California and Alaska, respectively, found that seismic albedo is $>0.9$ in both of these regions which means that scattering attenuation is dominant.

Hoshina (1991) used a Monte Carlo method to simulate the temporal shape of the coda envelope of regional seismograms and to estimate the relative importance of higher orders of multiple scattering on the envelope shape. In his method, he models the propagation of S-wave energy in terms of moving packets of energy. Each packet is radiated from the source and moves in one direction. The travel direction is changed at some location that is considered to be the location of a scatterer. The scatterer location and new travel direction are determined by a probability function appropriate for a mean free path of $1/g$ and isotropic scattering, respectively. A packet that changes direction three times, for example, is interpreted to have undergone triple scattering. A large number of packets and large order of scattering are used to estimate the space–time distribution of multiple scattered coda wave energy. Hoshina's (1991) modelling is based on the same assumptions for the scattering process as those of Wu (1985): isotropic scattering and scatterers randomly and uniformly distributed. While the physics modelled by both
Hoshiba (1991) and Wu (1985) are the same, Hoshiba's numerical calculation resulted in a determination of the distribution of energy in space and time for an impulsive source while Wu's analytical expressions have no time dependence. More recently, Zeng, Su & Aki (1991) confirmed Hoshiba's (1991) numerical simulation based on their time domain solution in an integral form for the multiple isotropic scattering problem. Hoshiba integrated the energy over time to obtain integrated energy $E(r)$ versus distance $r$, for comparison with the results of Wu (1985). There was good agreement between the theoretical results of Wu (1985) and Hoshiba (1991). Hoshiba (1991) also showed that the shape of the calculated integrated energy versus distance relation is significantly affected by the time window over which the energy is integrated. When $g$ is large, scattering distributes energy into the later portion of the recorded seismogram and it is likely that the total energy will be underestimated when the integration time window is small. Hoshiba (1991) pointed out that when the length of the integration window is similar to that used in previous studies of regional data, the theoretical relation between integrated energy and distance is independent of seismic albedo for albedos greater than 0.5 so that no inference about albedo could be made. High albedo values reported by previous investigators (Mayeda et al. 1991; McSweeney et al. 1991) might be artifacts of the short time windows used to integrate energy in their analysis.

Figure 1(a) shows integrated energy versus distance for media with selected properties where energy has been integrated between 0 and 15 s after the $S$-wave arrival at each distance. The energy has been corrected for geometric spreading by multiplying by $r^2$. The calculations have been carried out using the method of Hoshiba (1991). Each line represents the result for one value of albedo. Fig. 1(b) shows curves for the same media where the energy has been integrated between 30 and 100 s after the $S$-wave arrival. Comparing the curves in Figs 1(a) and (b) for similar albedos shows that the ratio of energy in the later time interval to that in the early time interval is substantially higher for the high seismic albedo media than for the low-albedo media. In fact, for high-albedo media, the integrated energy in the later time interval exceeds that in the earlier time interval at large distance. Thus, if integration is over short time intervals, the Monte Carlo method should be used to compute theoretical fits to data since it allows us to generate model curves appropriate for the time intervals used in the data analysis.

Here, we first propose a new method, called multiple lapse-time window analysis, for measuring attenuation due to intrinsic mechanisms and scattering from regional seismic data. This method takes advantage of our ability to use the Monte Carlo method of Hoshiba (1991) to model the spatial dependence of energy over finite lapse times. Various curves corresponding to different sets of parameters are compiled in Hoshiba, Sato & Fehler (1991). We apply the method to data collected by the Kanto–Tokai seismic network, which is operated by the National Research Institute for Earth Science and Disaster Prevention in Tsukuba, Japan. Since data from many different source–receiver combinations will be used, we correct the signal amplitudes for local site amplifications and for source effects. These corrections are made using the coda-wave method (Aki & Chouet 1975; Phillips & Aki 1986). The seismic energy integrated from the $S$-wave onset was calculated for each station and corrected for the site amplification and source size. We integrated the seismic energy over equal-length time intervals on each seismogram, since Hoshiba (1991) has shown the importance of the time window's length on the shape of the energy versus distance relationship. Three time windows were used: 0–15, 15–30, and 30–100 s after the $S$-wave onset. Comparing the shapes of the energy versus

![Figure 1. Geometric spreading corrected integrated energy, $r^2E(r)$, versus distance, $r$, calculated using the method of Hoshiba for media having velocity of 4 km s$^{-1}$ and $g + a/Q, V = 0.03$. The number of particles used in the Monte Carlo simulation was $5 \times 10^6$ for albedos smaller than 0.7 and $2 \times 10^7$ for the higher albedos. (a) Result when energy is integrated from 0–15 s after $S$-wave onset at each distance. (b) Result when energy is integrated from 30–100 s after $S$-wave onset.](image)
distance curves and the relative amplitudes of the curves obtained for the earlier and later integration time windows with the results obtained from Hoshiba's (1991) modelling, we obtain stable and reliable estimates of the seismic albedo and mean free path for scattering, thus allowing us to quantify the relative amounts of scattering and intrinsic attenuation.

**DATA**

The area of study is the Kanto–Tokai region of Honshu Island, Japan. This area is near the triple junction of the Asian, Philippine Sea, and Pacific Plates and is one of high seismic activity. Many of the earthquakes are associated with plate subduction and there are also a large number of crustal earthquakes. Ishida & Hasemi (1988) discuss the seismic structure of the Kanto—Tokai region as inferred from 3-D inversion of P-wave arrival times and from the distribution of hypocentres in the region.

Waveform data were collected by the seismic network operated by the Japan National Research Institute for Earth Science and Disaster Prevention (NIED, formerly NRCDP), which is located in Tsukuba. Most stations have three-component 1 Hz velocity transducers. A few stations have 4 Hz horizontal transducers. Many of the seismometers are located in boresholes, some as deep as 3 km. Fig. 2 shows the locations of stations used in this study. Hamada et al. (1985) describe the NIED network and routine processing of network data. Waveform data are digitized at each station at a rate of 80 samples per second using an 8 bit scheme with logarithmic scaling to increase the dynamic range for unclipped signals. Digital data are transmitted to Tsukuba for event detection and further processing.

Waveforms from earthquakes that trigger the event detection algorithm are stored and used to measure arrival times for finding event locations. We used these routinely determined locations for our analysis. The location scheme, which uses both P- and S-wave arrivals, was described by UKawa et al. (1984) and the reliability of the method has been examined by Papanastassiou & Matsumura (1987). Epicentral and focal depth errors for events at depths between 10 and 50 km are less than 1 and 3 km, respectively, in the study area.

Fig. 2 also shows the epicentres of events used in this study. A subset of the events used to calculate the integrated energy were also used to measure the site amplifications. This subset of events is also shown in Fig. 2. Focal depths of all events used were less than 50 km.

Figure 3 shows seismograms obtained at two stations that were used in the study. The three time windows over which energy was integrated are shown for each seismogram.

**METHOD OF ANALYSIS**

We measure the integrated seismic energy as a function of source—receiver distance and frequency for earthquakes in the Kanto–Tokai region. Since interpretation is based on the shape of the integrated energy versus distance relation, we need to measure energy ratios and not absolute energy. Thus, we make all measurements relative to a reference station. We will make measurements for many earthquakes and many stations. The amplitude of the waveform recorded at a seismic station is affected by the source size and the local site amplification in addition to the media between the source and receiver so we correct for the source and site amplification effects to isolate the path effect. The correction is made using the following formula:

\[
E_u = \frac{E_u^0}{S_i L_j}
\]  

where \(E_u\) is the corrected integrated energy for event \(i\) at receiver \(j\) relative to the reference station, \(E_u^0\) is the uncorrected integrated energy at receiver \(j\) for source \(i\), \(S_i\) is
the source correction for source $i$, and $L_j$ is the local site amplification for receiver $j$ relative to the reference site. All measured integrated energy will be in arbitrary units since only relative values are required for the analysis. Both $E_{ij}^n$ and $L_j$ will be measured without making corrections for the differences in the instrumental frequency responses among the stations since the ratio shown in equation (3) normalizes for these differences.

Prior to analysis, all seismograms were inspected visually and regions where clipping or digital glitches occurred were discarded. We studied the following bands: 1–2, 2–4, 4–8, 8–16, and 16–32 Hz. We will first discuss the methods for making site and source corrections to the data and then discuss the method for measuring the integrated seismic energy versus distance.

**Site Amplification**

Site amplification for each station used in the analysis was determined using the coda-wave method as outlined by Phillips & Aki (1986). This method was first used by Tsujiura (1978), who showed that site amplification factors obtained from coda waves are consistent with those obtained from direct $S$-wave arrival data and that the coda measurements show less scatter than those from direct waves. This suggests that coda waves are composed predominantly of scattered $S$-waves. The physical basis of the coda-wave method for determining site amplifications is that after an earthquake, elastic energy is distributed nearly uniformly within a spherical volume far behind the $S$-wavefront (Frankel & Wennerberg 1987). After we determine $g$ for the Kanto–Tokai region, we will use Hoshiba’s (1991) method to show that this assumption is valid for our data set. Since energy is uniformly distributed, the ratio of the signal amplitudes measured at two stations in the same time interval after an earthquake is equal to the ratio of the local site amplifications at those stations. We wish to determine the site amplification of each site relative to that at some reference station. To apply the coda method we measure the coda amplitude at a given station averaged over some frequency band for a data window centred at a pre-chosen, fixed lapse time after the earthquake origin time. We then calculate the ratio of this amplitude to that measured at the reference site for the same frequency band and at the same lapse time. Tsujiura (1978) and Phillips & Aki (1986) found that the ratio of the coda amplitudes so measured is independent of the lapse time as long as the lapse time used is greater than twice the traveltime of the direct $S$-wave from the source to each receiver and that the source–receiver distance is less than about 100 km. This ratio is then a measure of the relative site amplification for both direct $S$ and coda waves.

We used coda waves from 10 earthquakes (see Fig. 2) to determine site amplifications. We chose earthquakes with local magnitudes larger than 3.4 to ensure that signal amplitude was large for lapse times greater than two times the $S$-wave traveltime to each station. The events had focal depths of less than 50 km and locations throughout the study area (see Fig. 2). We used 10 s coda windows, tapered both ends of the data with a cosine taper whose length was 5 per cent of the length of the data. The windowed data were Fourier transformed and the spectra were averaged over each frequency band. We compared the amplitude in each frequency band with the amplitude calculated for a noise window taken before the $P$-wave arrival time. If the signal exceeded the noise by a factor of 3, the data were used. We used as many 10 s windows as possible within the coda of each seismogram for each source–receiver pair. The first 10 s window used began at 60 s after the earthquake origin time to eliminate the possibility of the late arriving phase observed by Obara & Sato (1988) from contaminating the results. Data for source–receiver pairs separated by less than 120 km were used. We averaged the amplitude ratios obtained for all 10 s windows for all events at each site to obtain the site amplifications $L_j$ for each site.

**Source amplitude corrections**

Source corrections were also obtained using the coda method. Aki (1969) demonstrated that the amplitude of the coda envelope in a given frequency band measured at one station at a fixed lapse time measured from the earthquake origin time is proportional to the seismic moment and independent of the source–receiver distance, provided the distance is less than about 100 km and the lapse time is greater than two times the direct $S$-wave traveltime. Thus, by measuring the amplitudes of coda waves from many events at one station but at the same lapse time, we can estimate the relative seismic moments of the corresponding earthquakes. We used a 10 s window beginning at 50 s lapse time. The time window beginning at 50 s lapse time was chosen because signals in this window were neither clipped for big events nor buried in the noise for the small events. The time domain data were tapered, Fourier transformed, and averaged over the frequency band of interest. Signal-to-noise ratios were estimated in the same manner as for the site corrections and results obtained from stations with poor signal-to-noise were not used. The average over each frequency band was first normalized by the corresponding site amplification $L_j$. Then, by averaging the site-corrected amplitudes over many stations, we obtained a stable estimate of the size of the earthquake, $S$, which is used in equation (3).

**Measurement of energy**

All previous investigators who used Wu’s (1985) theoretical results to obtain seismic albedo (Wu & Aki 1988; Toksöz et al. 1988; Mayeda et al. 1991) have estimated the energy $E_{ij}^n$ in equation (3) by windowing the seismic trace beginning with the $S$-wave, tapering, Fourier transforming, squaring the Fourier transform and averaging over selected frequency bands. While this method should yield reliable results when the $S$-wave arrivals are not impulsive, we were concerned that tapering a signal that begins with an impulsive arrival may lead to a significant underestimation of the energy. Such an underestimation will be most pronounced at short source–receiver distances where first arrivals are most likely to be impulsive. This underestimation may then lead to an erroneous conclusion that seismic albedo is high since Wu’s (1985) results show that the quantity $r^2E(r)$ increases with distance at short source–receiver distances for high-albedo media (see fig. 6 of Wu 1985).

To avoid possible side-effects from tapering, waveforms
were filtered over the frequency band of interest using a causal fourth-order Butterworth filter. Since this filter is causal, a high-amplitude portion of the signal that occurs at the beginning of the window is smeared into the portion of the waveform after the arrival but is not lost. The error caused by similar smearing of energy out of the end of the window used to estimate energy is small since the later portion of the seismogram is much smaller than the earlier portion. The filtered signals were squared and summed over the time window of interest to obtain an estimate of the relative integrated energy for the event.

Hoshiba (1991) has shown that the shape of the integrated energy versus distance curve is a function of the time window over which the energy is integrated. While Hoshiba’s result points to a significant problem in data analysis using Wu’s (1985) model, it is not practical to integrate energy over longer time periods to obtain a more reliable comparison with Wu’s results. Seismograms with long codas above noise level generally have clipped direct S-waves. Signals from small events, with unclipped S arrivals, decay into the noise level too quickly to allow the use of very long integration time windows. Rather than trying to integrate seismic energy over large time windows using one set of earthquakes, we used earthquakes of several sizes and integrated the energy over three distinct time intervals. We used seismograms from earthquakes with magnitudes \( M_L \) ranging from 2.0 to 4.2, which had unclipped S arrivals, for the time intervals from 0 to 15 s and 15 to 30 s after the S arrival. We then used larger events, with magnitudes ranging from 2.9 to 6.1, to calculate the integrated energy in the time interval from 30 to 100 s after the S arrival. The signal-to-noise level of most of the small events was too small to allow us to estimate the energy for time intervals significantly longer than 30 s after the S-wave onset. By using the coda-wave method to normalize all events to the same source amplitude, we obtained a measure of the relative energies in all three time windows over a wide range of source–receiver distances.

**RESULTS**

**Site amplifications**

Figure 4 shows site amplifications obtained using the coda-wave method for the vertical component of the motion relative to station TRU for frequency bands of 1–2, 2–4, and 4–8 Hz on station location maps. Station TRU was chosen as the reference station because it is located near the centre of the NIED network and visual examination of seismograms showed this to be a very quiet station. The sensor is located in a 160 m deep borehole at station TRU. Results were not obtained for some of the frequencies at some stations analysed due to poor signal-to-noise ratios. Scatter in the overall results was investigated by calculating the difference between individual measurements made for each station and the mean for that station. Fig. 5 shows the scatter in station corrections obtained for all of the time windows used, where measurements for individual 10 s windows are normalized by the mean \( L_j \) obtained for each station.

\[
D = 100 \frac{O_{i,j} - L_j}{L_j}
\]  

where \( O_{i,j} \) is the site amplification observed for the \( k \) th 10 s window of event \( i \) at site \( j \). Note that the number of individual measurements decreases with increasing frequency. Fig. 5 shows that the station corrections for the three lowest frequency bands have little scatter about the means obtained for each station. Results for two higher frequency bands, 8–16 and 16–32 Hz, have more scatter and are considered unreliable. From now on, we will discuss results for only the 1–2, 2–4 and 4–8 Hz bands. We found no systematic variation in calculated site amplifications when using events from different source regions.

We also calculated station corrections for the north–south horizontal component of motion. We found that results for the horizontal component of motion were similar to those obtained for the vertical component except in the lowest

![Figure 4](image-url) Vertical component site amplifications obtained for stations in the NIED seismic network using the coda-wave method. Amplifications are relative to station TRU. Station names are shown in Fig. 1. (a) Result for 1–2 Hz band. (b) Result for 2–4 Hz band. (c) Result for 4–8 Hz band.
frequency band. For the 1–2 Hz band, results differed because there is a 4 Hz geophone at station TRU and 1 Hz geophones at most other stations. While seismograms have been corrected for electronic gain, they have not been corrected for differences in the shapes of the frequency responses between the stations. If we choose station TR2, a station near TRU that has 1 Hz geophones for all three components, as a reference station, the relative site amplifications for vertical and horizontal components are similar for all frequency bands except at stations where instrument response is significantly different between the two components of motion. Fig. 6 shows the horizontal site amplifications plotted against vertical amplifications when station TR2 is chosen as the reference station. The correlation between amplifications for the two components of motion is quite good. A few sites, all having surface sensors, have exceptionally large horizontal amplifications.

Figure 7 shows the vertical component site amplifications calculated using the coda-wave method for the 4–8 Hz band plotted against magnitude residuals calculated for stations in the NIED network by Noguchi (1990). Magnitudes are determined from the maximum trace amplitude measured on the seismograms at each station and residuals are the differences between magnitudes determined at a given station and the average magnitude determined from all stations. The figure shows a good correlation between site amplifications determined by the coda method and the magnitude residuals. There is also a clear correlation between the 1–2 and 2–4 Hz site amplifications and magnitude residuals determined by Noguchi (1990) but the correlation is best for the 4–8 Hz site amplification results.

**Energy versus distance**

Figure 8 shows $r^2E_y$ versus $r$ for the vertical component data for the 2–4 Hz band in the three intervals, 0–15, 15–30 and 30–100 s. We obtained an average value of energy for each...
distance by averaging individual measurements over 15 km intervals. The result of this average, along with the 1 standard deviation range for the average are shown for each set of data. The figure shows that the scatter among individual measurements is less for the 15–30 and 30–100 s integration windows than for the 0–15 s window. The scatter is largest for the 0–15 s window since it includes direct wave energy that is most affected by the source radiation pattern, which has not been accounted for in the present analysis.

Figure 9 shows \( r^2E(r) \) versus \( r \) for vertical component data for the three frequency bands, 1–2, 2–4, and 4–8 Hz. Three curves derived from data and the theoretical fits to the data, discussed below, are shown for each frequency band corresponding to the three time intervals of integration. The results for the north–south component of motion are nearly identical to those for the vertical.

Figures 8 and 9 show that the energy in later time intervals after the direct S-wave arrival is significantly smaller than the energy in early time intervals. This dramatic reduction in energy in the medium with increasing time can only be caused by intrinsic absorption removing elastic energy from the medium. If scattering dominates the attenuation of the direct arrival, the energy would be redistributed to times after the direct arrival and the curves for integration times of 15–30 and 30–100 s would have larger amplitude than we found (see Fig. 1). Since the energy we observe in the 0–15 s time window is orders of magnitude larger than that in the later time windows, the 0–15 s window can be considered a good estimate of the energy integrated over infinite time. Wu's (1985) curve calculated for an infinite time window in a medium with high albedo, e.g. strong scattering, increases with distance out to some distance and then decreases. Our result for the 0–15 s time window shows no peak, contrary to the results obtained by Mayeda et al. (1991) for California and McSweeney et al. (1991) for Alaska. In fact, our results are more similar to the ones obtained by Wu & Aki (1988), who examined deep earthquakes in the Hindu Kush region. Thus, our preliminary conclusion is that scattering is not dominant in the Kanto–Tokai region of Japan.

To quantify the ratio of scattering-to-total attenuation, we used the Monte Carlo method of Hoshiba (1991) to calculate the shape and relative amplitudes of the \( r^2E(r) \) versus \( r \) curves for the time intervals used in our data analysis and for various media parametrizations (Hoshiba et al. 1991). Energy versus time and source–receiver distance were calculated for chosen values of \( g + \omega/QV \) and seismic albedo. We choose \( V = 4 \text{ km s}^{-1} \) for crustal S-wave velocity in the Kanto–Tokai region. The energy was integrated over the same three time windows we used for analysing our data. The result was a set of three curves for each value of \( g + \omega/QV \) and seismic albedo. An example of a set of curves for two time windows is shown in Fig. 1. By laying the \( r^2E(r) \) versus \( r \) curves obtained from data on the theoretical curves calculated for different values of albedo and \( g + \omega/QV \), we can adjust the media parameters until the best match is obtained between the data and the theory.

An alternative method allows us to determine \( g \) and \( g + \omega/QV \) directly. It is based on the observation that the ratio of the integrated energy in the early time interval, 0–15 s, to that in the later time intervals is most influenced by the amount of scattering, which is parametrized by \( g \). In

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**Figure 8.** Geometric spreading-corrected integrated energy \( r^2E(r) \) versus distance obtained for shallow earthquakes in the Kanto–Tokai region of Japan. Individual measurements and the average over 15 km windows with 7.5 km spacing are shown. Error bars are 1 standard deviation spreads calculated for each 15 km window. Data are for the 2–4 Hz band and vertical component of motion. Results for three integration time windows are shown: 0–15, 15–30 and 30–100 s.
addition, the ratio of integrated energy in the early time interval at short distance to that in the same time interval at long distance is dominated by the total attenuation, $g + \omega/QV$. We thus measure the following two ratios from the modelling results of Hoshina et al. (1991):

$$R_1 = \log_{10} \left[ \frac{E_1(r = 50 \text{ km})}{E_1(r = 150 \text{ km})} \right],$$

$$R_2 = \log_{10} \left[ \frac{r^2 E_1(r = 50 \text{ km})}{r^2 E_1(r = 150 \text{ km})} \right].$$

(5)

where $E_1(r = 50 \text{ km})$ is the energy in the first time interval, 0–15 s, measured at 50 km distance and $E_3(r = 150 \text{ km})$ is the energy in the third time interval, 30–100 s, measured at 150 km. For each of the two ratios, there is a range of possible values of $g$ and $g + \omega/QV$ (and hence of $R_2$) that are represented by the lines drawn in Fig. 10. The point of intersection of the lines determined from measurements of $R_1$ and $R_2$ defines the values of the two attenuation coefficients. Since the lines representing possible values of $g$ and $g + \omega/QV$ consistent with each measured ratio intersect at relatively high angle, the values of the

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**Figure 9.** $r^2 E(r)$ for vertical component of motion in three frequency bands: 1–2, 2–4 and 4–8 Hz. Three curves shown for each band represent the results for integration times of 0–15, 15–30 and 30–100 s after the S-wave onset. Fits to the data using the method of Hoshina (1991) and the values of $g$ and $g + \omega/QV$ listed in Table 1 are shown.
parameters are well constrained by measuring these two ratios alone. By placing error bounds on the ratios determined from the data, a range of possible values of $g$ and $g + \omega/Q_i V$ can be determined from Fig. 10.

Table 1 lists the values of the ratios $R_1$ and $R_2$ estimated for the three frequency bands: 1–2, 2–4, and 4–8 Hz. Fig. 11 shows the ranges of acceptable values of $g$ and $g + \omega/Q_i V$ plotted on a figure similar to Fig. 10. We have assumed that the uncertainties in $R_1$ and $R_2$ are ±0.25. Fig. 11(a) shows the ratios for the vertical data and 11(b) the horizontal data. The similarity between the two figures shows the independence of the results on the component of motion analysed. Values of $g$ and $g + \omega/Q_i V$ are listed in Table 1, along with the calculated values of seismic albedo $B_o$, mean free path $1/g_i$, $Q_i^{-1}$, $Q_s^{-1}$, and total attenuation, $Q_i^{-1} + Q_s^{-1}$.

The fits to the data calculated using Hoshiba’s (1991) method for the values of $g$ and $g + \omega/Q_i V$ listed in Table 1 are shown in Fig. 9. The correlation between the observed data and the calculations shows that the graphical method of determining $g$ and $g + \omega/Q_i V$ yields excellent results. The graphical method does not take into account the data in the 15–30 s time window. In Fig. 9, we see that the fit to data in the 15–30 s window is not good for the 1.5 Hz band. We are unable to find any values of $g$ and $g + \omega/Q_i V$ that could simultaneously fit data from all three time windows for the 1–2 Hz band. Our inability to fit these data is presumably due to the simplifying assumptions used in our model: homogeneous distribution of scatterers and isotropic scattering. It is reasonable to assume that there is some change in the density of scatterers across the Mohorovičić discontinuity. This first-order discontinuity is not accounted for in the theory. An additional problem with the 1–2 Hz band may be that both surface and body waves are mixed, which is not accounted for in the theoretical modelling. Tsujiura’s (1978) conclusion that site amplification for direct $S$-waves are the same as those obtained from coda waves for frequencies between 0.75 and 24 Hz in the Kanto–Tokai region leads us to believe that there is little surface wave contamination in the coda waves.

**DISCUSSION**

In general, forward scattering increases as frequency increases. If we include strong forward scattering in our
Table 1. Values of ratios, \( R_1 \) and \( R_2 \) (equation 5) determined from data and values of media parameters obtained from Fig. 10. Results for the vertical and north–south horizontal components of motion are shown. \( Q_s^{-1} \) is obtained from equation (1), \( Q_v^{-1} \) from equation (6).

<table>
<thead>
<tr>
<th>Center Freq. (Hz)</th>
<th>( R_1 )</th>
<th>( R_2 )</th>
<th>( g )</th>
<th>( g + \omega/VQ_s )</th>
<th>( B_0 )</th>
<th>( g^{-1} )</th>
<th>( Q_1^{-1} )</th>
<th>( Q_2^{-1} )</th>
<th>( Q_3^{-1} )</th>
<th>( Q_4^{-1} )</th>
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<tr>
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<td>(km)</td>
<td>(km/s)</td>
<td>(km)</td>
<td></td>
<td>(km)</td>
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<td></td>
<td></td>
<td></td>
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<tr>
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<td>.75</td>
<td>.30</td>
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<td>.009±.007</td>
<td>.45±.5</td>
<td>250</td>
<td>.0021</td>
<td>.0017</td>
<td>.0038</td>
<td>.0041</td>
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<tr>
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<td>1.15</td>
<td>.70</td>
<td>.006±.005</td>
<td>.019±.007</td>
<td>.34±.06</td>
<td>150</td>
<td>.0026</td>
<td>.0014</td>
<td>.0040</td>
<td>.0033</td>
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<tr>
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<td>1.10</td>
<td>.01±.005</td>
<td>.030±.007</td>
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<td>.0011</td>
<td>.0033</td>
<td>.0022</td>
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In theory, the energy integral \( E_1 \) that includes the direct arrival will increase relative to \( E_2 \) and \( E_3 \) when compared to results obtained from an isotropic scattering model. If we use the reference curves derived using the isotropic scattering assumption to fit data for a non-isotropic scattering case, we will obtain an albedo smaller than 1 even if there is no intrinsic absorption. Thus, large forward scattering leads to an underestimation of seismic albedo using our method and we should remember that the albedos we determined may reflect the non-isotropy in scattering.

Our determination of media properties was made using a total of only three measurements from the 0–15 and 30–100 s data. The ability of the model obtained from these data to fit the entire curves for these two lapse times and to fit the data in the 15–30 s time window gives us confidence in the uniqueness of the result obtained by the ratio method we used. The close fit between the data and the model for the 2–4 and 4–8 Hz bands is rather surprising given all the assumptions of the model. As a further test to see where model assumptions become important, we analysed vertical component data using the following two sets of lapse-time windows: 0–15, 15–30, and 30–45 s; 0–10, 10–20, and 20–30 s. The data were analysed in exactly the same manner as was previously done; only the lapse-time windows were changed and a set of model curves similar to Fig. 10 were developed for the new windows. Table 2 summarizes the results obtained using the different sets of lapse-time windows. The media attenuation parameters obtained for the 2–4 and 4–8 Hz bands were nearly identical for all three sets of lapse-time windows giving us increased confidence in our results. For the 1–2 Hz band, the attenuation extinction coefficient \( g + \omega/Q_s^{-1} \) was the same for all sets of windows but \( g \) decreased from 0.008 km\(^{-1}\) for the shortest set of lapse time windows (0–10, 10–20, 20–30 s) down to 0.004 km\(^{-1}\) for the longest set (0–15, 15–30, 30–100 s). This indicates an apparent increase in the mean free path with increasing lapse times while the total attenuation remains constant for this band. Since the total attenuation measured by our approach is dominated by the decay in energy versus source–receiver distance of the earliest arriving waves, we should not expect much dependence in measured total attenuation on the lapse-time windows used. The increase

![Range of Media Parameters for Vertical Component Data](image1)

![Range of Media Parameters for Horizontal Component Data](image2)

Figure 11. Ranges of possible values of \( g \) and \( g + \omega/Q_s^{-1} \) for three frequency bands studied in the Kanto–Tokai region. (a) Range for vertical component data. (b) Range for north–south component data.
in the scattering mean free path and corresponding decrease in scattering attenuation predicted using longer lapse-time windows may be due to decreased heterogeneity in the upper mantle, where the long lapse-time waves have certainly sampled.

The values of total attenuation $Q^{-1}$ found in our study are of the same order with those found by Aki (1980a), who used a single station method to measure attenuation of direct S-waves in the Kanto–Tokai region; however there is a difference in the frequency dependence of the $Q^{-1}$ found by the two studies. This difference might be due to the poor results we have obtained for the 1–2 Hz band.

The values of $g$ estimated in this study are in the middle of the range of $g_n$, the backscattering coefficient, estimated by Sato (1978) for the Kanto region by modelling coda-wave excitation. The values of $g$ in the present study are smaller by a factor of 2 to 4 than of the $g_n$ estimated by Aki (1980b) for the Kanto region using coda-wave excitation. Thus, there is some coincidence between the scattering coefficient estimated using analysis based on multiple isotropic scattering and the backscattering coefficient estimated using single scattering analysis applied to coda-wave excitation for the Kanto region. This coincidence may be due to the small value of $g$ found for the region.

We estimated coda attenuation $Q^{-1}$ (Tsujjura 1978) from our data using the same earthquakes used to calculate site amplifications (see Fig. 2). Our measurements were made over the lapse times from 60 to 105 s, using data only at lapse times greater than two times the S-wave traveltime to each station. A value of $Q^{-1}$ that gave the best fit to

$$V_j(t, f) = \frac{C_j}{t} e^{-\lambda P/Q},$$

where $C_j$ is a combination of the site amplification for receiver $j$ and the source factor for earthquake $i$, and $V_j(t, f)$ is the FFT of a window of the seismogram at receiver $j$ for event $i$ centred at time $t$ averaged over the frequency band centred at frequency $f$. The values we found are listed in Table 1. The values of $Q^{-1}$ are remarkably close to the values of $Q^{-1}$ for the 2–4 and 4–8 Hz frequency bands, in agreement with Frankel & Wennerberg (1987). For the 1–2 Hz band, $Q^{-1}$ and $Q^{-1}$ are similar. In a previous study of attenuation in the northeastern United States, Toksöz et al. (1988) found that $Q^{-1}$ is similar to $Q^{-1}$ and that both are smaller than scattering $Q^{-1}$. We find a similar result. Mayeda et al. (1991) reported that coda attenuation is similar to scattering attenuation and that both are much larger than intrinsic attenuation in California.

Sato (1984, 1990) explained both S-wave attenuation and coda excitation only by scattering due to random inhomogeneities on the basis of the Born approximation for the elastic wave equation, where frequency-dependent non-isotropic scattering and mode conversion scattering were included, but only single scattering was considered. To reconcile these results with those of Sato (1984, 1990) it is necessary to develop a method for the measurement of seismic albedo which includes frequency-dependent non-isotropic scattering of elastic vector waves including mode conversions and multiple scattering.

The value of scattering coefficient found for the Kanto–Tokai region is a factor of 10 smaller than those found for Alaska by McSweeney et al. (1991) and for California by Mayeda et al. (1991). It is also significantly smaller than that found for the northeast United States and Canada by Toksöz et al. (1988). As previously mentioned, these results were obtained by comparing with a theory where energy is integrated over an infinite time and the results so obtained may overestimate the seismic albedo.

The similarity between the site amplifications and energy versus distance relationships calculated for the vertical and horizontal components of motion is striking. No previous study of site amplification using the coda-wave method has been made which used horizontal component data.

We can make a consistency check on the validity of the assumption made when using the coda-wave method to determine site amplifications that elastic energy is uniformly distributed after an earthquake within a spherical volume that expands at half the S-wave velocity. Fig. 12 shows the distribution of seismic energy at $t = 60$ s and $t = 100$ s after an earthquake in a material with $g = 0.007$, which is similar to the values found in this study. The results were calculated using the method described by Hoshiba et al. (1991). The figure shows that the energy is uniformly distributed to at least a distance of 120 km at both times, which is the maximum source–receiver distance we used for calculating site amplifications.
CONCLUSIONS

We have investigated the causes of attenuation of seismic energy in the Kanto–Tokai region of Japan using a new method called multiple lapse-time window analysis. The technique of Wu (1985) has been improved using the observation made by Hoshiba (1991) about the importance of the time limits over which seismic energy is integrated resulting in a new method that yields robust estimates of $g$ and $g + w/Q,V$, from which we calculate seismic albedo $B_{p}$, intrinsic attenuation $Q^{-1}$, scattering attenuation $Q^{-1}$, and total attenuation $Q^{-1}$. Our method is based on the assumption that scatterers are homogeneously distributed and that scattering is isotropic. The method relies on the measurements of two ratios made from the integrated energy versus distance curves. The reliability of using the ratios to arrive at estimates of media parameters has been confirmed by the good correlation between the integrated energy versus distance curves obtained from the data and theory. For the 1–2 Hz band, we were unable to completely fit the data for all of the time windows with a single model. We believe that this misfit between data and model is due to the simplifying assumptions in the theory that scattering is isotropic and scatterers are uniformly distributed.

Scattering attenuation is smaller than intrinsic attenuation in the Kanto–Tokai region. For the higher frequency bands, the value of $Q^{-1}$ is nearly the same as the value of intrinsic attenuation, $Q^{-1}$.

We have found site amplifications for stations in the Kanto–Tokai region using the coda-wave technique. The technique yielded stable results for frequency bands centred at 1.5, 3, and 6 Hz. The site amplifications were the same for the vertical and north–south components of ground motion.

Integrated energy versus source–receiver distance was similar for the horizontal and vertical components of motion. This results in our estimates of media attenuation parameters being particle motion independent.

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