A two-layer attenuation model for the upper mantle at short periods

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Abstract. Estimates of maximum mean P wave attenuation (described by the quality factor $Q_{\alpha}$) are used to make a two-layer model of $Q_{\alpha}$ in the period range 0.125–1 s for the northwest Pacific subducting region. A shallow highly attenuating layer ($Q_{\alpha} \approx 150$) of only 85 km thickness overlays a less attenuating layer to a depth of 450 km ($Q_{\alpha} > 1000$). The added bandwidth made available by the use of array data yields key information for testing the validity of textbook models predominantly based on extrapolation from longer period data. We find that seismic attenuation at short periods appears to behave differently to long-period attenuation. This reversal is represented by a modification to the absorption band model obtained from longer period data.

Introduction

Seismic waves radiated from an earthquake source decrease in amplitude at a rate greater than that required by geometric spreading alone. This may be due to an intrinsic energy loss in the form of heat to anelastic processes operating on lattice defects, grain boundaries, or fluid-rock interfaces. Alternatively the presence of inhomogeneities in the propagating medium produce perturbations along the raypath, e.g., reflections, refractions, and diffractions, and lead to a reduction in amplitude by energy being transferred further down the seismogram in the form of a seismic coda (for a more general discussion of seismic attenuation see Toksöz and Johnston, 1981). Thus, spatial variations in the measured attenuation may be related to the intrinsic properties of the medium or to variations in the amplitude of local heterogeneity, although in practice these two are difficult to separate with narrow-band data. These fluctuations give an alternative means to seismic velocity of characterising Earth structure, and may, for example, be more sensitive to heterogeneities in the mantle due to thermal effects or the presence of partial melt.

Much of our current understanding of the upper mantle is based on inversions of attenuation structure from long-period seismic data ($\tau > 1$ s). Figure 1 shows a model for seismic attenuation developed from such data at different depths due to the presence of seismic absorption bands peaked at different periods for different depths [Anderson, 1989]. In this paper we examine short-period ($\tau < 1$ s) P-wave radiation from earthquakes in the northwestern Pacific seismic zone. The new local short-period estimates of seismic attenuation behave differently to the global long-period attenuation models, represented by a modification to the absorption band model (dashed line in Figure 1).

Data

The data consist of earthquakes recorded teleseismically at the four arrays in Warramunga (Australia, WRA), Yellowknife (Canada, YKA), Gauribidanur (India, GBA) and Eskdalemuir (Scotland, EKA). The arrays were designed for studying methods of detection

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Figure 1. The absorption band model for the globally-averaged $Q_{\beta}$ structure of the uppermost 500 km of the Earth [Anderson, 1989]. The model has been extrapolated to frequencies outside the range of observations (‘original study’). From observations above 1 Hz (‘this study’) a simple modification to the model (dashed absorption band at 300 km depth) may be needed, at least in the northwest Pacific subduction zone.
and identification of underground explosions, and have a broader bandwidth than conventional short-period seismometers [Bache et al., 1985]. In order to optimise the resolution and make best use of the limited spatial coverage of the arrays in a single study we concentrate on a relatively large tectonic region. Figure 2 shows the earthquakes used in this case study, ranging in depth from 16 km to 442 km.

The use of array data allows signal to noise enhancement by summing individual records after applying a time correction to each array seismograph dependent on the azimuth and incident angle of the incoming phase to the array [Bache et al., 1985]. After such processing coherent energy is seen above the background noise to as much as 8 Hz in some cases. Figure 3 shows the signal-to-noise enhancement that can be obtained from the use of array data compared to a single station in the array.

Estimation of $t^*$

For teleseismic studies at these frequencies seismic attenuation is usually expressed as $t^*$, the ratio of travel time $T$ to mean quality factor $Q$. Values of $t^*$ of 1 s, commonly accepted in long-period studies, are clearly inappropriate at short periods. Figure 3 (from the present work) shows that such values are inconsistent with the observed signal energy at frequencies higher than about 2–3 Hz at teleseismic distances. As a result the best estimate of $t^*$ has been dropping steadily with increasing signal-to-noise ratio at these frequencies in recent years [Der et al., 1985, Bache et al., 1986]. In this study we examine this effect in more detail by using earthquakes at different depths. For a more detailed description of the method used in this study see Sharrock et al., 1995.

After summing the data in the time domain the observed spectrum is determined by standard fast Fourier Transform routines. The spectrum is then corrected for a simple power law far-field source model ($\omega^{-2}$, see Aki and Richards, 1980) and the resulting decay in spectral amplitude with increasing frequency is assumed to be entirely due to attenuation. This procedure is justified because the corner frequencies of the teleseismic earthquakes used here ($m_b > 5.4$) are below 1 Hz. Due to the difficulty in separating intrinsic attenuation and scattering at these frequencies (1–8 Hz) no attempt to do so has been made, or to determine a frequency-dependent $Q$.

The mean path attenuation is then proportional to the slope of the source-corrected spectrum above the corner frequency (Figure 3). The observed mean path attenuation $t^*$ is a conservative (upper bound) estimate because other possible far-field source models (e.g., $\omega^{-8}$) would result in a lower $t^*$. From the line fit we estimate a typical uncertainty in the estimate of this upper bound to $t^*$ of about 20%.

**Interpretation of $t^*$**

One of the advantages of using $t^*$ is that the contribution from different parts of the raypath sum linearly, for example as:

$$t^* = t_0^* + t_1^* + t_2^* + t_3^* + t_4^*$$

(1)

where $t_0^*$ and $t_1^*$ are due to the crustal region beneath the source and receiver, respectively, $t_2^*$ and $t_3^*$ are due to paths in the upper mantle and, where appropriate, $t_4^*$ is due to paths in the lower mantle.

![Figure 3. The spectrum of the array data shows signal amplitude to frequencies of at least 4 Hz, while a $t^*$ of 1 s requires a spectral decay shown by slope $A$, reducing the available bandwidth considerably. Summing the instrument in the array also greatly enhances the signal-to-noise ratio. The amplitude spectrum of a single instrument (dashed line) is enhanced by the delay and sum of the 20 instruments in the array by a factor of as much as $\sqrt{20}$ (solid line). The noise is shown as the dotted line.](image-url)
The components \( t_{w}^{*} \) and \( t_{u}^{*} \) are equivalent for a spherically symmetric Earth, and \( t_{d}^{*} \) is assumed to be negligible because seismic attenuation in the lower mantle is thought to be very small compared to the other regions considered. The sites of the arrays and their element spacing have been carefully chosen to minimise the effects of crustal structure. Nevertheless the effects of the known crustal structure at the receiver arrays have been examined using a local horizontally-layered model. The effect of local scattering at the arrays, mainly due to reverberations in the crustal layers, was found to be small compared to \( t_{d}^{*} \), and hence near-receiver effects can be considered negligible. In summary, equation (1) can now be written as a sum of attenuation near the source region and attenuation in the upper mantle (\( t_{u}^{*} \)):

\[
t^{*} = t_{s}^{*} + t_{u}^{*}
\]  

(2)

**Estimation of \( t_{d}^{*} \)**

Obviously, further analysis requires the separation of the effects of \( t_{w}^{*} \) and \( t_{u}^{*} \). In order to do this we use the depth variation of earthquakes in the northwest Pacific seismic zone. The International Seismology Centre depth estimates were all independently verified by examination of depth phases, and corrected where necessary. We found \( t_{d}^{*} \) in the range 0.1–0.6 s, consistent with other studies of body wave attenuation [Der and Lees, 1985, Bache et al., 1986, Walck, 1988].

A plot of \( t^{*} \) against focal depth (Figure 4) shows that the deeper earthquakes (85–450 km) have a lower mean path attenuation than the shallower earthquakes (< 85 km). Since the upper mantle paths traveled by both deep and shallow earthquakes are similar at the teleseismic distances used in this study, this difference in observed attenuation must be due to the source region. Therefore, the effect of the source region can be calculated from the difference in \( t^{*} \) above and below 85 km. Above 85 km we get a \( t^{*} \) estimate of \( t^{*} = 0.36 \pm 0.015 \) s, where the standard adjusted error is \( \sigma_{n-1}/n \), for \( n \) data with a standard deviation of \( \sigma_{n-1} \). Below 85 km we get \( t^{*} = 0.23 \pm 0.076 \) s, yielding a \( t_{u}^{*} \) of about 0.1 s to a maximum depth of 85 km, equivalent to \( Q_{a} \) of about 150. The mean \( t_{w}^{*} \) from 85–450 km, after correcting shallow earthquakes for \( t_{s}^{*} \), was found to be 0.24 s (\( \sigma_{n-1} = 0.11 \) s), corresponding to \( Q_{a} \) of over 1000. This \( Q_{a} \) is significantly lower in the upper 85 km than it is in the depth range 85–450 km at frequencies greater than 1 Hz. This result is the opposite to that obtained from low-frequency studies where a highly attenuating region below 85 km underlies a less attenuating lid above this depth [Kanamori, 1970, Niazi, 1971].

**Discussion**

The averaged \( Q_{a} \) structure of the northwest Pacific subduction region is different to that of the averaged Earth. The values have been converted to estimates of the S-wave quality factor \( Q_{s} \) using the following relationship for pure shear in a Poisson solid [Anderson and Archambeau, 1964]:

\[
Q_{a} = \frac{0}{4} Q_{s}
\]

(3)

There is a suggestion that \( Q_{a} \) approaches the value of \( Q_{s} \) at short periods, and in this case the above relation gives a minimum estimate of \( Q_{s} \). This might suggest that shear dissipation has a lesser effect at shorter periods, but, in fact, this convergence of compressional and shear attenuation at short periods is due to scattering [Cormier, 1989].

The absorption band model shown in Figure 1 is a globally-averaged model, and does not represent the extremes of seismic attenuation observed in the Earth, such as hot active subduction zones. On the other hand, the northwest Pacific model proposed here represents an area of lateral heterogeneity in \( Q \). Therefore, the absorption band model should predict higher \( Q \) than that observed in the northwest Pacific model at short periods. Our model suggests that in the period range 0.125–1 s the attenuation drops rapidly with depth, so that at 300 km depth \( Q_{s} \) is already greater than 1000. A simple modification to the absorption band model (dashed absorption band in Figure 1) resolves this discrepancy, with the result that observed attenuation should become strongly dependent on frequency above 1 Hz.

In retrospect we should perhaps not be surprised at the apparent reversal in our (short-period) results of the usual (long-period) pattern of a poorly attenuating lithosphere overlaying a more highly attenuating asthenosphere. For example, it is well known that in general materials behave in a more perfectly elastic fashion at high strain rates (or high frequencies in cyclic loading), so relatively lower \( Q_{a} \) at 85–450 km depth might be expected by extrapolation from the results at lower frequencies. In contrast the highly attenuating lithosphere at frequencies above 1 Hz may be best explained by increased scattering off geometric heterogeneities at

![Figure 4. A graph of \( t^{*} \) versus depth for each earthquake-station path in the case study. The mean with adjusted standard error bars of the two layers of the short-period model are shown.](image-url)
Conclusions

The array data used here can provide a useful means of extending the observable seismic band to periods below 1 s for selected tectonic regions. In this study, we have determined a simple model for the P-wave attenuation structure in the period range 0.125–1 s for the northwest Pacific subducting region. The observed values of $t^*$ are significantly less than 1 s as so commonly applied to short-period data. We find an attenuating region 85 km in depth with a mean $Q_A$ of 150, underlain by a mean $Q_A$ for the upper mantle of over 1000. This pattern is the opposite to that found at longer periods. A simple comparison with extrapolated long-period data shows that the attenuation drops rapidly with depth in the period range 0.125–1 s. This observation leads to a simple modification to the absorption band model, suggesting strong frequency dependence for seismic attenuation at short periods.

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