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Radial Q_μ structure of the lower mantle from teleseismic body-wave spectra

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ABSTRACT

We have measured 150,000 P and 130,000 S wave spectral ratios up to 0.8 Hz using recordings of 250 deep (focal depth >200 km) earthquakes from 890 global and regional network stations. We have inverted these data to estimate the attenuation parameters t_P^* and t_S^* for P and S waves and a radial profile of the quality factor Q_μ for the lower mantle. On average, t_P^* increases by about 0.2 s and t_S^* increases by about 0.7 s between epicentral distances of 30° and 97°. The relatively strong increase of t_S^* ($t_S \approx 4t_P^*$) suggests that intrinsic shear attenuation is the cause of the overall trend in our data. The increase of t_P^* and t_S^* with distance is smaller than predicted by models PREM [12], QL6 [11], and QLM9 [16]. Assuming PREM values for Q_μ in the upper mantle, where the data lack resolving power, the P and S wave spectra are explained best if Q_μ increases from about 360 at PREM's 670-km discontinuity to 670 in the lowermost mantle. The high values for Q_μ can be reconciled with previously determined values by invoking a frequency-dependence of $Q_\mu(\omega)$ that is proportional to $\omega^{0.1}$. Data that are separated in 'Pacific' and 'circum-Pacific' subsets have slightly different trends. Estimates of t_P^* and t_S^* for the Pacific data, which sample the large low shear-velocity province of the Pacific, are higher than the circum-Pacific estimates. Thus, it appears that the Pacific large low shear velocity province has accompanying low Q_μ. The difference in Q_μ in the lowermost 1000 km of the mantle beneath the Pacific and beneath the circum-Pacific is at most 17%. Lateral variations of this magnitude are marginally resolvable given the uncertainties of our measurements.

1. Introduction

Although seismic constraints of the mantle have come primarily from studies on elastic velocities, it is well recognized that joint interpretations of seismic velocities and attenuation are critical for understanding the structure and dynamical state of the Earth's interior.

Global variations of attenuation in the upper mantle have been estimated using both surface waves (Billien et al., 2000; Dalton et al., 2008; Gung and Romanowicz, 2004; Romanowicz, 1995; Selby and Woodhouse, 2002) and body waves (e.g., Bhattacharyya et al., 1996; Reid et al., 2001; Warren and Shearer, 2002). These global-scale studies are consistent with regional-scale studies (Baqer and Mitchell, 1998; Hwang et al., 2009; Lawrence et al., 2006; Roth et al., 2000; Sheehan and Solomon, 1992) and indicate that the upper 200–300 km of the mantle beneath oceans and tectonically active regions is generally more attenuating than the mantle beneath stable, continental shields.

Except for the study by (Lawrence and Wysession, 2006b), wave attenuation in the lower mantle has been modeled using 1D profiles (Fig. 1). Whole-mantle profiles, constrained by normal-modes (Durek and Ekström, 1996; Dziewonski and Anderson, 1981; Resovsky et al., 2005; Roult and Clévédé, 2000; Widmer et al., 1991) and ScS/S waveforms (Lawrence and Wysession, 2006a), have a common low Q_μ layer in the uppermost mantle (80–200 km depth), intermediate Q_μ values in the transition zone (200–650 km), and the highest Q_μ values in the lower mantle. However, absolute values of Q_μ and the depth dependence of Q_μ in the lower mantle differ in these profiles. PREM (Dziewonski and Anderson, 1981) and QL6 (Durek and Ekström, 1996) indicate constant values of 312 and 355 in the lower mantle, respectively. (Oki and Shearer, 2008) resolve lower mantle Q_μ value of about 620 using S–P ratio method at short-period band (3–10 s). (Resovsky et al., 2005) constrain Q_μ to decrease in the lowermost 1000 km of the mantle. Q_μ in PAR3P (Okal and Jo, 1990) and QM1 (Widmer et al., 1991) decrease throughout the lower mantle while it increases in the lower 1000 km of the lower mantle in model QLM9 (Lawrence and Wysession, 2006a). The study of (Warren and Shearer, 2000) provide a high-frequency (0.16–0.86 Hz) estimate of Q_P from global P to PP spectra. Their frequency-independent Q_P is about 2600 in the lower mantle which is more than three times larger than the Q_P value of 780 in PREM. Variable approaches, data sets and measurement uncertainties are responsible for these differences and underscore that the basic radial structure of Q is still poorly constrained.

In this study, we follow a classical approach in which Q is determined from the epicentral distance variation of the body-wave

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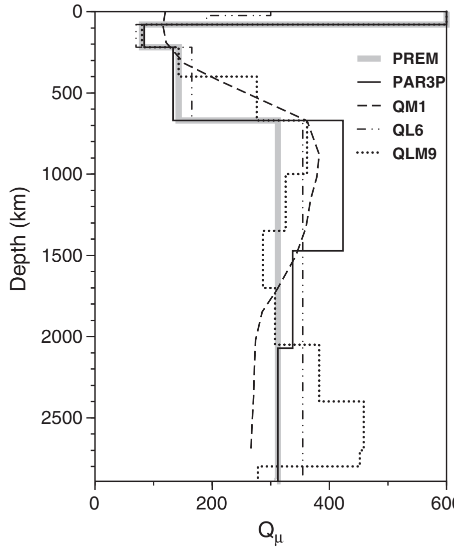

Fig. 1. Radial Q_μ structures of (thick grey line) PREM (Dziewonski and Anderson, 1981), (solid black line) PAR3P (Okal and Jo, 1990), (dashed line) QM1 (Widmer et al., 1991), (two-dot chain line) QL6 (Durek and Ekström, 1996), and (dashed dotted) QLM9 (Lawrence and Wysession, 2006a).

attenuation parameters t_P^* and t_S^*. The original study by (Teng, 1968) and subsequent studies of spectral ratios (e.g., Der and McElfresh, 1977; Der et al., 1982); (Solomon and Toksöz, 1970) and amplitude decay (e.g., Booth et al., 1974; Butler and Ruff, 1980) were mostly applied to band-limited, analog waveform data from the United States. Here, we measure t_P^* and t_S^* from a nearly two-decade long collection of digital waveforms from broadband seismic stations in regional and global networks. We describe our data in Section 2. We review methods and describe the epicentral variation of t_P^* and t_S^* and the inferred radial variation of Q_μ in Sections 3 and 4, respectively. Final conclusions are drawn in Section 5.

2. Data

2.1. Teleseismic body-wave spectra

The attenuation parameter t^* is defined as the ratio of the body-wave travel time t and the quality factor Q integrated along the ray path (e.g., Stein and Wysession, 2003):

$$t^* = \int_{\mathrm{ray}} \frac{dt}{Q}. \tag{1}$$

We use t_P^* and t_S^* and Q_P and Q_μ to denote the attenuation parameters and quality factors of P and S waves, respectively. Since the amplitude spectrum of a body wave is proportional to the attenuation function $\exp(-\frac{1}{2}\omega t^*)$, the ratio $R_{ij}(\omega)$ of spectra $O_i(\omega)$ and $O_j(\omega)$ for the same earthquake is related linearly to the t^* difference recorded at stations i and j (Teng, 1968):

$$\ln R_{ij}(\omega) = -\frac{\omega}{2}\Delta t_{ij}^*. \tag{2}$$

While we have previously investigated spectral ratios of P waves to map lateral variation of t_P^* (Hwang et al., 2009), we study here the variation of t_P^* and new measurements of t_S^* as a function of epicentral distance. If Δ_i and Δ_j are the epicentral distances of stations i and j, we associate

$$\Delta t_{ij}^* = t_{\Delta_i}^* - t_{\Delta_j}^* \tag{3}$$

to depth-dependent attenuation in the lower mantle, using the fact that P and S wave turning depths increase monotonously with epicentral distance (Fig. 2). Q_μ is related to Q_P by

$$Q_P^{-1} = LQ_\mu^{-1} + (1-L)Q_\kappa^{-1}, \tag{4}$$

where $L = \frac{4}{3}(V_S/V_P)^2$ and V_S and V_P are the S and P wave velocities (e.g., Anderson and Given, 1982). If shear attenuation is much larger than bulk attenuation (i.e., $Q_\mu \ll Q_\kappa$), then

$$Q_P^{-1} = LQ_\mu^{-1}. \tag{5}$$

Eq. (5) predicts that t_P^* and t_S^* differ by about a factor of 4.5 and that the Q_P/Q_μ ratio is 2.25 (depending on the velocity structure of the mantle) which is almost the same as the Q_P/Q_μ ratio of 2.27 estimated in this study.

2.2. Measurements

We measure Δt^* using more than 150,000 P wave and 130,000 S wave spectral ratios from about 250 events with magnitudes larger than 6. The events occurred between 1987 and 2005 and have been recorded by broadband seismometers from global (GSN and Geoscope) and numerous regional networks. The focal depths of the earthquakes are larger than 200 km so that the P and S signals are not complicated by surface reflections (i.e., pP, sP, sS) and not attenuated strongly by the uppermost mantle in the source region. We limit the analysis to epicentral distances larger than 30° to avoid waveform complexity due to strong velocity gradients in the upper mantle and to distances smaller than 97° to avoid the effects of diffraction along the core. A high-pass filter with a corner frequency of 120 s is applied to the vertical-component P wave and the transverse-component S wave.

We inspect all waveforms to ensure that the signals are well above noise level, and have low-amplitude coda and impulse onsets. Typically, P-wave and S-wave time windows are about 8 s and 27 s long, respectively. However, we adjust these to isolate waveforms with similar characteristics. We measure Δt^* for pairs of stations that have similar source azimuths to minimize the effects of rupture directivity on the spectra. Examples of waveforms and spectral ratios have been shown by (Hwang et al., 2009). The amplitude spectra are estimated up to 0.8 Hz using the multiple-taper spectral analysis of (Lees and Park, 1995). Δt^* and its uncertainty are estimated by linear regression.

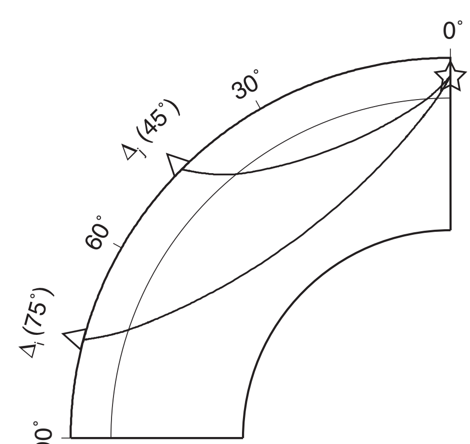

Fig. 2. Geometrical ray paths of P waves from a (star) 300-km deep earthquake source to (triangles) stations at epicentral distances of Δ = 45° and Δ = 75°.