Global variation of body-wave attenuation in the upper mantle from teleseismic P wave and S wave spectra

Y. K. Hwang, J. Ritsema, and S. Goes

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We constrain the spatial variation of P-wave ($t^p_*$) and S-wave ($t^s_*$) attenuation by inverting 190,000 teleseismic P- and S-wave spectra up to 0.8 Hz. These spectra are derived from 250 deep earthquakes recorded at 880 broadband global and regional network stations. The variance and ratios of $t^p_*$ and $t^s_*$ values are consistent with PREM’s upper mantle velocity and Q structures and conventional $t^p_*$ and $t^s_*$ values. High attenuation is resolved beneath stations in tectonically active regions characterized by high heat flow. Low attenuation marks stable continental regions. The maps of $t^p_*$ and $t^s_*$ correlate well with the variations of P- and S-wave velocity tomography.

1. Introduction

Models of the elastic velocity structure of the mantle have advanced our knowledge of mantle dynamics [e.g., Romanowicz, 2008], but are by themselves insufficient to obtain complete descriptions of the physical state of Earth’s interior. Anelasticity models can provide important complementary information. Anelasticity has a much stronger sensitivity to temperature and water content than elastic velocities, a lower sensitivity to composition and a different sensitivity to melt [e.g., Anderson, 1967; Karato and Jung, 1998; Hammond and Humphreys, 2000; Jackson et al., 2002; Faul et al., 2004; Shito et al., 2006].

A number of studies have mapped the global variation of seismic wave attenuation in the upper mantle using surface waves [e.g., Romanowicz, 1995; Billien et al., 2000; Gung and Romanowicz, 2004; Selby and Woodhouse, 2002; Dalton and Ekström, 2006; Dalton et al., 2008] and body waves [e.g., Bhattacharyya et al., 1996; Reid et al., 2001; Warren and Shearer, 2002; Lawrence and Wysession, 2006]. Here we add a new estimate of attenuation in the upper mantle from teleseismic P-wave and S-wave spectra.

Using globally distributed stations, we invert ratios of body-wave spectra for the P-wave and S-wave attenuation parameters $t^p_*$ and $t^s_*$. We compare our maps of $t^p_*$ and $t^s_*$ to surface-wave Q tomography and attenuation maps inferred from a thermal interpretation of shear-velocity tomography.

2. Spectral Analysis of P and S Waves

The attenuation parameter $t^*$ is defined as the ratio between the body-wave traveltimes $t$ and the quality factor $Q$ along the (ray) path $L$ [e.g., Stein and Wysession, 2003]:

$$t^* = \int \frac{dr}{Q}.$$  \hspace{1cm} (1)

If we write the spectrum $O(\omega)$ as the product of the source spectrum $S(\omega)$ and the attenuation function $e^{-\omega t^*/2}$,

$$O(\omega) = S(\omega)e^{-\omega t^*/2};$$ \hspace{1cm} (2)

the logarithm of the spectral ratio $R_{ij}$ between $O(\omega)$ and $O_j(\omega)$,

$$\ln R_{ij}(\omega) = -\frac{\omega}{2} \Delta t^*_{ij},$$ \hspace{1cm} (3)

is linearly related to the difference between the attenuation parameters at stations $i$ and $j$. Here, $\Delta t^* = t^* - t^*_j$.

To isolate the influence of intrinsic attenuation on $t^*$ from other sources such as crustal amplification, scattering, focusing and defocusing, we use large number of spectral ratio measurements. Our data set comprises 190,000 P- and S-wave spectral ratios from broadband recordings of 250 earthquakes with magnitudes larger than 6. The earthquake focal depths are larger than 200 km to ensure short source-time functions and to avoid interference of the direct P- and S-waves with the surface reflections pP, sP, and sS. We analyze the spectra at teleseismic distances (30°–85°) to avoid waveform complexities from triplication in the transition zone and diffraction along the core-mantle boundary. We select 10–30 s long segments of P-wave and S-wave signals with impulsive onsets, low-amplitude coda, high signal-to-noise ratios, and similar waveforms for the same earthquakes. To minimize the variations of spectra due to varying source azimuths, we measure $\Delta t^*_j$ for station pairs that have similar azimuths.

We determine $\ln R(\omega)$ up to a frequency of 0.8 Hz using the multiple-taper spectral analysis method of Lees and Park [1995]. $\Delta t^*_j$ (for P-waves) and $\Delta t^*_j$ (for S-waves) and 2σ uncertainties are estimated by linear regression of $\ln R(\omega)$. We apply a correction using the results of Hwang and Ritsema [2011] to account for the systematic increase.

1Department of Geological Sciences, University of Michigan, Ann Arbor, Michigan, USA.
2Department of Earth Science and Engineering, Imperial College London, London, UK.
The correlation of $L06308 = 3.5$ ratio values reviewed by to after correction $t = 4.5$ structures and versus \[1982\]. Variations do and $D$ and (b) by a joint $V$ values are determined by are affected by wave scattering and $t$ uncertainties and mea-

Within overlapping circles with radii of $D$ and $t$ values have been averaged using Spatial variations of (a) $P$ in the teleseismic distance range by about 0.2 s Measurements. High attenuation characterizes tectonically active collision zones, rift zones and back-arc regions, while low attenuation is found below stable continental cores. For example, $t^S$ is relatively high in the tectonically-active western North America and low in

\[\Delta t^* = \frac{3}{4} \int \left( \frac{V_P^2}{V_S^2} \right) ds \Delta t_P^* \] (4) using velocity structures ($V_P$ and $V_S$) of PREM and assuming that $\Delta r^*$ is due to laterally varying $Q$ in the upper mantle only and that bulk attenuation is negligible [e.g., Anderson and Given, 1982].

Figure 3a shows the global distribution of $t^S$ in a map that has been smoothed by cap-averaging. High attenuation characterizes tectonically active collision zones, rift zones and back-arc regions, while low attenuation is found below stable continental cores. For example, $t^S$ is relatively high in the tectonically-active western North America and low in

\[\Delta t^* \] measurements and uncertainties are given by Hwang et al. [2009] and Hwang and Risema [2011].

3. Results
3.1. Lateral $t_P^*$ and $t_S^*$ Variations
[s] Since $t_P^*$ and $t_S^*$ are affected by wave scattering and near-surface 'site-responses', we investigate the average values of $t_P^*$ and $t_S^*$ within overlapping circles with radii of 3° (Figures 1a and 1b). The averaging of the data brings out the large-scale patterns of $t_P^*$ and $t_S^*$ that reflect global tectonics and that are similar to the global heat flow variations (Figure 1c).

[9] The spatial variations of $t_P^*$ and $t_S^*$ are similar and the ratio of $t_P^*$ and $t_S^*$ variances (~4) is consistent with the expected ratio of 4.5 for the upper mantle $Q$ structure of PREM [Dziewonski and Anderson, 1981] and the conventional value of 3.5 for the $t_P^*/t_S^*$ ratio [e.g., Cormier, 1982] (Figure 2). This indicates that variations in $t_P^*$ and $t_S^*$ do indeed reflect the lateral variation of intrinsic attenuation in the upper mantle.

3.2. Joint Inversion for $t_S^*$
[10] In Figure 3a, we show the map of $t_S^*$ by a joint inversion of $\Delta t_P^*$ and $\Delta t_S^*$. To relate the $\Delta t_P^*$ data to $\Delta t_S^*$ in the upper mantle, we have used

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Figure 3. Spatial variation of $t_S^f$ in the upper mantle estimated (a) by joint inversion of $\Delta t_P^f$ and $\Delta t_S^f$, (b) from the surface-wave $Q$ model of Dalton et al. [2008] ($t_P^f$), and (c) from the thermal interpretation of S20RTS ($t_T^f$). The correlation coefficient between $t_S^f$ (Figure 3a) and $t_P^f$ (Figure 3b) and between $t_S^f$ (Figure 3a) and $t_T^f$ (Figure 3c) are about 0.3.

The platforms of central and eastern North America. A similar contrast is also apparent in Europe: $t_S^f$ is higher in western Europe than in the Baltic shield region. Station density is lower in other regions but a pattern consistent with tectonics persists. For example, $t_S^f$ is low in the East African Rift region and high at stations within the western and southern cratons of Africa. In addition, $t_S^f$ is high in the back-arc regions of the western Pacific subduction zones.

3.3. Comparison With Seismic Tomography

We compare the map of $t_S^f$ based on the spectral ratios of P- and S-waves with the $t_S^f$ variation computed by integrating through two $Q$ models for the upper 400 km of the mantle using (1). In Figure 3b, we show the distribution of $t_P^f$, and denote it as $t_P^f$, predicted by the model QRFSI12 [Dalton et al., 2008] for the upper mantle. QRFSI12 is a spherical harmonic degree-12 model of shear attenuation derived using fundamental-mode Rayleigh-wave amplitudes in the long-period range (50–250 s). The data set of Rayleigh-wave amplitudes are corrected for source, instrument, and focusing effects.

In Figure 3c, we show $t_T^f$, and denote it as $t_T^f$, for the $Q$ structure based on a thermal interpretation of S20RTS [Ritsema et al., 1999] shear-velocity anomalies with respect to the Ocean Reference Model of Ritsema and Allen [2003]. For the conversion from $dV_S$ to temperature anomalies we assume that the mantle has a homogeneous pyrolitic composition and that below a PREM lithospheric structure, the average velocity profile corresponds to a mantle adiabat with a potential temperature of 1300°C. Elastic velocities are calculated using a finite-strain approach [Cammarano et al., 2003; Goes et al., 2005] with a correction for anelastic effects using an Arrhenius-type pressure and temperature-dependent $Q$ formulation [Karato, 1993; Goes et al., 2005]: $Q_P(T, P) = Q_0 \exp\left(gT_m(P)/T\right)$, where $T$ and $P$ are absolute temperature and pressure, respectively, $Q_0 = 0.1.10^{1.5}$, $g (= 40)$ is a scaling factor, and $T_m$ is the peridotite solidus. The conversion yields temperature, and corresponding $V_P$, density, $Q_S$ and $Q_P$. Regional models under North America, Europe and Australia converted in a similar manner yielded temperatures that could reconcile observed $V_P$, $V_S$ and surface heat flow [Goes et al., 2000, 2005; Goes and van der Lee, 2002]. The long-wavelength thermal structure inferred from S20RTS has reasonable temperatures varying between 600°C and 1450°C at 100 km depth and 1200–1550°C at 300 km depth (Figure 4).

4. Discussion

There is a remarkable similarity between $t_S^f$, $t_P^f$, and $t_T^f$. This indicates that surface-wave amplitudes and body-wave spectra are affected by the same long-wavelength
variation in attenuation even though the wavelengths and propagation directions of surface-waves and body-waves are entirely different.

[15] The variations in $t_\nu^*$ values are larger than the variations in $t_\phi^*$ and $t_\phi^+$. For example, the contrast between western North America and stable North America and between western Europe and the Baltic region is about 0.7 s in $t_\phi^*$ but about 0.3 s and 0.5 s for $t_\phi^+$ and $t_\phi^+$, respectively. However, these differences are to be expected given the uncertainties originating from averaging $(t_\phi^*)$, the regularization of the inverse problem $(t_\phi^*)$, and uncertainties of the velocity-temperature conversion $(F^*)$.

[16] The correlation between patterns in $t_\phi^+$, surface heat flow, tectonics and shear-velocity anomalies suggest attenuation is largely the result of thermally activated creep. The conclusion that temperature exerts the main control on global $Q_S$ and $V_S$ structures is consistent with other studies [Artemieva et al., 2004; Dalton et al., 2009; Dalton and Faul, 2010]. Our analyses illustrate that the maps of $t_\phi^+$ and $t_\phi^+$ can be explained by variations of intrinsic attenuation consistent with a temperature variation as that depicted in Figure 4.

[17] Other factors such as the presence of melt below mid-ocean ridges and a melt-depleted composition of cratonic roots likely have additional influence [Artemieva et al., 2004; Dalton et al., 2009; Dalton and Faul, 2010]. The back-arc high $t_\phi^+$ and $Q_S^*$ anomalies that coincide with low shear velocities, which we have interpreted as high temperatures, may partially reflect high water content compatible with an interpretation of regional $V_P$, $V_S$, and $Q_S$ below the Izu-Bonin arc [Shirot et al., 2006]. To better distinguish between different mechanisms requires an imaging of $t_S^*$, $t_P^*$, and seismic velocities at more similar resolution and scale than the models we compared here.

5. Conclusions

[18] New maps of $t_\phi^+$ and $t_\phi^*$, derived from 190,000 teleseismic, global P-wave and S-wave spectra, exhibit a coherent large-scale spatial variation that is consistent with heat flow and tectonic variations. The ratio of $t_\phi^*$ to $t_\phi^*$ is consistent with the PREM ratio of 4.5 and the conventional $t_\phi^*$ to $t_\phi^*$ ratio of 3.5. Moreover, a joint inversion of the P-wave and S-wave spectral ratios yields lateral variation of $t_\phi^*$ that is similar to the predicted $t_\phi^*$ variation for a recent surface-wave $Q$ model $(t_\phi^*)$ of the upper mantle and a thermal interpretation of shear-velocity anomalies in the upper mantle $(t_\phi^+)$. Combined these observations indicate that the large-scale pattern in $t_\phi^*$ and $t_\phi^+$ reflects variations in intrinsic shear attenuation.

[19] The high correlation between $t_\phi^*$ and $t_\phi^*$ indicates that coherent patterns of attenuation can be constrained from large data sets of horizontally and vertically propagating waves. The similarity between $t_\phi^*$, $t_\phi^*$, and $t_\phi^+$ suggests that the patterns of Figure 3 predominantly reflects variable attenuation in the upper few hundred kilometers of the mantle. The patterns are consistent with a thermal structure of the mantle as inferred from shear velocity anomalies.

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References


S. Goes, Department of Earth Science and Engineering, Imperial College London, London SW7 2AZ, UK. (s.goes@imperial.ac.uk)

Y. K. Hwang and J. Ritsema, Department of Geological Sciences, University of Michigan, Ann Arbor, MI 48109, USA. (ykhwang@umich.edu; jritsema@umich.edu)