The oceanic and cratonic upper mantle: Clues from joint interpretation of global velocity and attenuation models

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1. Introduction

An important application of seismological models is inferring the physical and chemical state of the Earth's interior. The majority of such models on a global scale contain three-dimensional variations in shear-wave speed. However, shear-wave velocity can be influenced by a number of factors. For example, anomalously low velocity can be produced by high temperature (e.g., Ritzwoller et al., 2004), low Mg# (Lee, 2003) and partial melt (Hammond and Humphreys, 2000a; Jackson et al., 2004); it is also expected that the presence of water reduces velocity indirectly through anelastic dispersion (Karato and Jung, 1998). A unique interpretation of seismological models remains challenging because each of these factors, and any combination of them, could result in anomalously low wave speeds.

The Earth's anelasticity is manifest in the attenuation of seismic waves with time and distance traveled. Anelastic processes are strongly dependent on temperature, leading to an increasing temperature derivative with increasing temperature (Cooper, 2002; Jackson et al., 2002). The presence of partial melt will enhance attenuation if the mechanism is grain-boundary sliding (Faul et al., 2004) and have no effect in the seismic band if the mechanism is melt squirt (Hammond and Humphreys, 2000b). There are few experimental constraints on anelastic effects related to major-element composition or volatile content, but it is expected that attenuation will be insensitive to compositional variations and enhanced by water (Karato and Jung, 1998; Aizawa et al., 2008).

Given that seismic velocity and attenuation have different and complementary sensitivities to temperature, composition, melt, and water, jointly interpreting these two sets of observations should help improve and reduce some of the ambiguity in traditional interpretations of wave speed. There have been a number of attempts to do this at the global scale (e.g., Romanowicz, 1990, 1995; Billien et al., 2000; Artemieva et al., 2004; Gung and Romanowicz, 2004; Dalton et al., 2009) and at regional scales (e.g., Roth et al., 2000; Lawrence et al., 2006; Yang et al., 2007). The success of such studies has been limited by two separate factors: (1) difficulties associated with accurately imaging attenuation in the mantle, and (2) insufficient experimental measurements, made at the appropriate conditions, to be used for quantitative interpretations.

Seismic surface waves provide the best constraints available on global upper-mantle structure. A surface-wave travel time is a direct measure of upper-mantle elastic velocity, and the decay of a surface-wave amplitude directly reflects upper-mantle anelasticity. However, the amplitudes are affected by factors in addition to intrinsic attenuation: focusing and defocusing due to gradients in elastic velocity, and uncertainty in the knowledge of the source excitation and instrument response. For these reasons, much of the effort in attenuation studies has been concerned with developing techniques to account for the extraneous effects and isolate the signal of attenuation in surface-wave amplitudes (e.g., Romanowicz, 1990;
Durek et al., 1993; Selby and Woodhouse, 2000; Dalton and Ekström, 2006.

We have recently developed a new global model of shear attenuation in the upper mantle from a large data set of fundamental-mode Rayleigh wave amplitudes (Dalton et al., 2008). We consider Rayleigh waves in the period range 50–250 s and, after data selection, have 30,000–50,000 amplitudes at each period. To minimize the contaminating effects on amplitudes, we invert the amplitude data set simultaneously for the coefficients of the attenuation model as well as for terms that correspond to each of the other factors: frequency-dependent amplitude correction factors that allow for uncertainty in the source amplitude and instrument amplitude, and Rayleigh wave phase-velocity maps to account for focusing effects (Dalton and Ekström, 2006). The new model, QRFSI12, is parameterized with degree-12 spherical harmonics, providing lateral resolution of ~2000 km.

We have also recently measured shear modulus and attenuation on samples of pure melt-free olivine (forsterite-90) at upper-mantle temperatures (1000–1300 °C) and seismic frequencies (0.001–1 Hz). These measurements are made simultaneously, thus constraining the anelastic effect on wave speeds as well as the frequency and temperature dependence of attenuation at the appropriate frequencies (Jackson et al., 2002). The experimental observations quantitatively demonstrate the extent to which increasing temperature, decreasing frequency, and decreasing grain size all lead to smaller shear modulus and higher dissipation. Faul and Jackson (2005) derived a model, by fitting these experimental data, that describes the temperature, frequency, and grain-size dependence of modulus and attenuation. The model allows extrapolation to laboratory conditions to upper-mantle conditions, which is particularly important for grain size and pressure, both of which were too small in the experiments (3–165 μm and 300 MPa, respectively).

In this paper, we jointly analyze seismological models of shear velocity and attenuation together with the experimentally based model of Faul and Jackson (2005). In plots of attenuation versus velocity at fixed depths, the seismological model values and the experimental predictions form distinct trends. In regions where the seismological model trend coincides with the experimentally predicted trend, the range in velocity and attenuation can be ascribed mostly to lateral temperature variations (Dalton et al., 2009). In regions where the seismological trends are oblique to the experimental relationship, contributions to velocity and/or attenuation in addition to lateral temperature variations are required. We investigate possible mechanisms for the seismological velocity and attenuation values that fall outside of the experimental range, including laterally variable anelastic dispersion, the presence of melt beneath young seafloor, and compositionally distinct continental lithosphere.

2. Description of observations

As described in Section 1, the joint interpretation of seismic velocity and attenuation models may facilitate improved interpretations of the features in these models. For this analysis, we compare QRFSI12 with shear-velocity model S362ANI (Kustowski et al., 2008). Fig. 1 shows slices of the models at depths of 100 and 400 km. The details of QRFSI12 are described in Dalton et al. (2008). The robust features include a strong correlation with surface tectonics for depths shallower than 200 km and, at greater depths, zones of high attenuation located in the southeastern Pacific and eastern Africa, with low attenuation beneath a number of subduction zones in the western Pacific. The strong anticorrelation between velocity and attenuation is apparent in Fig. 1 and is discussed in more detail in Dalton et al. (2009).

An example of the predicted effects of lateral temperature variations on shear-wave velocity and attenuation at 100-km depth, calculated using the model of Faul and Jackson (2005), is shown in Fig. 2. As expected, increased temperature leads to lower wave speed and enhanced attenuation. The influence of grain size becomes important at high temperatures: a small grain size of 1 mm results in a more pronounced decrease in velocity and increase in attenuation with increasing temperature. Within the elastic regime (here, T<950 °C), no attenuation is predicted, and velocity is controlled entirely by the anharmonic derivatives, thereby being independent of grain size, activation volume, and frequency. When the two relationships are combined, a predicted relationship between velocity and attenuation given lateral temperature variations is obtained (Fig. 2c). Minimal attenuation is associated with velocities larger than 4.6–4.7 km/s at 100 km, and in general the curves in Fig. 2c depend little on grain size and activation volume. To illustrate effects related to uncertainty in the appropriate frequency relationship for attenuation, curves are also shown for shorter periods of 10 and 50 s.

In Fig. 3, the seismological velocity and attenuation values are overlain on the laboratory-derived predictions for depths of 100, 150, 200, and 250 km (Dalton et al., 2009). We sampled the global shear-velocity model S362ANI and the global shear-attenuation model QRFSI12 (Dalton et al., 2008) at 5762 points evenly spaced across the globe. Isotropic (Voigt average) velocities are used, and S362ANI is truncated at spherical-harmonic degree 12 to achieve the same spatial resolution as QRFSI12. The seismological points indicate a negative correlation between velocity and attenuation at all depths; correlation coefficients, using degrees 1–12, are −0.77, −0.58, −0.51 and −0.45 at 100, 150, 200, and 250 km, respectively. Each of these values is significant at >99% confidence levels.

To investigate regional trends in the velocity–attenuation relationship, we have drawn contours that enclose 75% of points from individual tectonic regions. Using the GTR1 tectonic regionalization scheme (Jordan, 1981) as a guide, we have defined “old continents” by grouping exposed Precambrian shields and platforms together with platforms overlain by undisturbed Phanerzoic cover. The oceans have been divided by age in two groups: >70 Myr and <70 Myr. To determine the contours, the areas shown in Fig. 3 are divided into small grid cells, and the contours connect cells with a similar point density. Each grid cell spans 0.025 km/s in velocity and 0.00125 units in attenuation. The best-fitting lines through each group of points have been determined by orthogonal regression; the 95% confidence bounds on these lines are reported by Dalton et al. (2009).

Focusing first on the seismological values, at 100-km depth the three tectonic regions are characterized by different velocity and attenuation values. Young oceans exhibit low velocities and high attenuation, whereas high velocity and weak attenuation describe old-continental points. The best-fitting lines through these points have very similar regression coefficients $dQ^{-1}dV_0$, resulting in nearly parallel lines. At 150 km, the two oceanic regions exhibit greater overlap than at 100 km and have similar regression coefficients. Old continents, characterized by higher velocities and lower attenuation, remain distinct from the oceanic areas. The best-fitting line through the old-continental points is also considerably less steep than is the case for the oceanic regions, a difference that is statistically significant at >95% confidence. These trends persist at 200 km. At 250 km, the best-fitting slopes for the oceanic and old-continental regions agree more closely and overlap within the 95% confidence bounds.

The velocity–attenuation relationships predicted by the model of Faul and Jackson (2005) are shown by the black curves in Fig. 3. The predicted curves indicate the expected values due to lateral temperature variations in dry melt-free forsterite-90 olivine (e.g., Fig. 2c). The seismic effects of variable composition and the presence of melt or water are not included and, as discussed below, may help to explain differences between the seismological values and the experiment-derived predictions. To first order, there is considerable overlap between the seismological values and the experimental predictions, in both magnitude and trend. At 100 km, the oceanic and old-continental regions exhibit slopes ($dQ^{-1}dV_0$) that are slightly shallower than the predicted curves. Either the range of seismological attenuation values is smaller or the range of seismological velocities is larger than predicted.
At 150 km, the old-continental points maintain a shallower-than-predicted slope, whereas the oceanic regions more closely match the magnitude and trend predicted by the mineral-physics model. This pattern holds at 200-km depth. At 250 km, however, it is no longer true that old-continental regions deviate significantly from the predicted curves.

We have established that the results in Fig. 3 are robust with respect to how strongly the attenuation model is damped. In constructing QRFSI12, the squared gradient of the attenuation perturbations is minimized, and we have found that, as expected, the strength of the horizontal-smoothness constraint influences both the data variance and the magnitude of lateral attenuation variations. Stronger damping results in a smoother, weaker model that provides a worse fit to the data (Dalton et al., 2008). Concerning the patterns in Fig. 3, weaker damping results in a wider spread of attenuation values and slightly larger values for the slope \(dQ^{-1}/dV_2\) of the best-fitting lines. However, the relationships between young oceans, old oceans, and old continents that are apparent in Fig. 3 are found for attenuation models determined with a wide range of horizontal-damping coefficients. Furthermore, the relationships between the seismological and mineral-physics models are also robust with respect to strength of damping, as the slopes of the seismological best-fitting lines do not vary considerably for models constructed with reasonable values of the smoothness coefficient. We note that the relationships in Fig. 3 are also found when the velocity and attenuation models are truncated at a lower spherical-harmonic degree; in other words, the patterns are not dominated by the small-scale features in the models.

From the comparison of seismological and laboratory-predicted values in Fig. 3, it is clear that laterally varying temperature in dry melt-free forsterite-90 olivine can explain much of the variability in global seismic models for oceanic regions at depths 250 km and old-continental regions at 250 km. This simple explanation is not sufficient for oceanic regions at 100-km depth and continental regions at depths <250 km, as indicated by the less favorable agreement between seismological values and predictions. As discussed by Dalton et al. (2009), laterally variable temperature likely does exert a strong influence over the seismic properties of these regions, especially in the shallow mantle. However, we infer from Fig. 3 that other factors, such as composition and partial melt, may also be important. In the following sections, we investigate explanations for the discrepancies. In particular, we focus on the shallower-than-predicted slopes for oceanic regions at 100-km depth and for cratonic regions at depths \(\leq\) 200 km.

3. Identification of outliers

We identify outliers as seismological points that fall outside of the range defined by the mineral-physics predictions. In Figs. 4–7, points that fall to the left and right of the experimentally defined range are colored red and blue, respectively, while points falling within the experimental range are shaded grey. The experimental range is defined to include the mineral-physics predictions for activation volumes between \(V = 12 - 20 \times 10^{-6} \text{ m}^3/\text{mol}\) and grain-size values of 1–50 mm. The experimental range is allowed to be slightly wider than these bounds to allow for smaller and larger grain-size and activation-volume parameters as well as experimental uncertainties. The geographical location of each of those points is plotted on the accompanying maps. At 100-km depth (Fig. 4), 49.5% of all points fall within the experimental range; 11.5% of the points fall to the left of the range and are characterized by lower-than-experimental velocity and/or lower-than-experimental attenuation, and 39% fall to the right. The majority of the anomalously low-velocity/low-attenuation points are geographically located beneath oceanic crust of age <70 Myr. The
At 150 km (Fig. 5), almost all of the oceanic points fall within the experimental range, with only ~25% of the young-oceanic points characterized by lower-than-experimental velocities and/or lower-than-experimental attenuation. In contrast, 72% of the old-continental points have larger velocity and/or higher attenuation than the experimental range. This trend persists at 200-km depth, where ~50% of old-continental points fall to the right of the experimental range while nearly all of the oceanic points are located within the experimental range. At 250 km, a handful of old-continental points (15%) are located to the right of the range, but the vast majority of all points agree with the mineral-physics model.

At 100 km, outliers located to the left of the experimental range are almost all located along the global mid-ocean-ridge system. This is especially true for the East Pacific Rise, the Pacific Antarctic Ridge, and the Southeast Indian Ridge. Areas of the Mid-Atlantic Ridge near the Azores hotspot and centered on the equator also exhibit anomalously low velocity and/or low attenuation. Small clusters of these points can be found in the northeastern Pacific and centered on the Red Sea. Beneath the oceans, outliers located to the right of the experimental trend comprise much of the western Atlantic, offshore Africa, and the northern central Pacific. Some of the anomalously high-velocity/high-attenuation points that are adjacent to continental areas could result from smearing of the continental properties into the oceanic regions, given that the global models used for this analysis have a relatively coarse resolution (spherical-harmonic degree 12). However, many of the outliers are far from any continental region (e.g., the central Pacific) and are not likely to be artefacts. Within the old continents, 84% of the points are located to the right of the experimental range. The remaining 16% of old-continental points that fall within or to the left of the experiments are generally located adjacent to tectonically younger areas. The few oceanic points with values to the right of the experimental range almost all adjoin an old continent, perhaps indicating some smearing of the continental seismic properties into the nearby ocean basins. Outliers located to the left of the experimental trends are found mostly in oceanic regions <70 Myr. As with 100 km, the old-continental points that fall within the experimental range are generally adjacent to tectonically younger areas. The few oceanic points with values to the right of the experimental range almost all adjoin an old continent, perhaps indicating some smearing of the continental seismic properties into the nearby ocean basins. Outliers located to the left of the experimental trends are found mostly in oceanic regions <70 Myr. As with 100 km, many of these points are located near the East Pacific Rise, Pacific Antarctic Ridge, and equatorial Mid-Atlantic Ridge. Iceland, the eastern Pacific, and the central Indian Ridge also have lower velocity and/or lower attenuation than the mineral-physics model predicts at 150 km.

At 150 km, almost all outliers located to the right of the experimental range (i.e., higher velocity and/or higher attenuation than the mineral-physics model) are found within old-continental regions: only 1.3% and 8.1% of young and old oceanic regions, respectively, fall to the right of the experimental trend, whereas 72% of the old-continental points do. As is the case for 100 km, the old-continental points that fall within the experimental range are generally adjacent to tectonically younger areas. The few oceanic points with values to the right of the experimental range almost all adjoin an old continent, perhaps indicating some smearing of the continental seismic properties into the nearby ocean basins. Outliers located to the left of the experimental trends are found mostly in oceanic regions <70 Myr. As with 100 km, many of these points are located near the East Pacific Rise, Pacific Antarctic Ridge, and equatorial Mid-Atlantic Ridge. Iceland, the eastern Pacific, and the central Indian Ridge also have lower velocity and/or lower attenuation than the mineral-physics model predicts at 150 km.

At depths of 200 and 250 km, the number of outliers shrinks, in part because the width of the experimental range expands. Nearly all of the young and old oceanic points fall within the experimental range; at 200 km, exceptions include the Pacific Antarctic Ridge, northeastern Pacific, equatorial mid-Atlantic Ridge, and a swath of the central Indian Ridge. Approximately 50% of the old-continental points fall outside the experimental trends at 200 km. Those outliers are almost all associated with anomalously high-velocity points, generally associated with oceanic crust older than 70 Myr and old-continental areas.

In (c), the black curves represent predictions of attenuation made at periods of 10 and 50 s to illustrate the effect of frequency.
with cratons with tectono-thermal ages > 1100 Myr (Artemieva, 2006). At 250 km, more than 85% of old-continental points fall inside the experimental range, with the few outlier points located in northern Canada, central Australia, South America, and western Europe.

4. Discussion

In Sections 2 and 3, we presented a comparison of the seismological models and the experimentally derived model of Faul and Jackson (2005). The following observations can be drawn from this comparison (e.g., Figs. 3–7): (1) Oceanic and old-continental regions are characterized by different values of $dQ^{-1}/dV$ at 150 and 200 km, and both regions undergo a change toward a steeper slope that provides better agreement with the mineral-physics model. This slope-steepening occurs in the depth range 100–150 km for the oceans and 200–250 km for the old continents. (2) At 100-km depth, almost all points that fall to the left side of the experimental range are located along or nearby mid-ocean ridges. Points that fall to the right side of the experimental range are located within continental cratons, beneath old seafloor (some of which is adjacent to continental cratons), and in the central Pacific. (3) At 150 km, points to the left of the experimental range are found along sections of mid-ocean ridge and in the northeastern Pacific. Points to the right of the experiments are located almost exclusively in continental cratons and oceanic areas that adjoin these continental regions. (4) At 200 and 250 km, the number of outliers continues to shrink, with most of the low-velocity/low-attenuation outliers found near mid-ocean ridges and high-velocity/high-attenuation outliers contained within continental cratons.

If we assume that the mineral-physics model is appropriate for the Earth’s upper mantle, then explanations are required for the following: lower seismological velocities and/or lower attenuation than predicted for parts of the mid-ocean ridge system and northeastern Pacific; and higher velocities and/or higher attenuation than predicted for many old-continental areas, associated with portions of old oceanic seafloor, and the central Pacific at 100-km depth. In the following sections, we discuss several possible origins for these anomalies, including artefacts of the seismological analysis as well as physical mechanisms. In particular, we focus on the mid-ocean ridges and continental cratons and delay a
We recognize that the mineral-physics model may not be appropriate for the upper mantle, given the conditions of the experiments from which it was derived (e.g., Jackson et al., 2002). For example, experimental grain size and pressure are too small, requiring extrapolation of both variables. Dislocation-related anelasticity may be underestimated since the synthetic olivine samples were prepared with minimal dislocation densities so that grain-boundary processes could be studied. Also, the experiments were performed only on samples of pure olivine; effects related to the coexistence of olivine with other mineral phases (e.g., Sundberg and Cooper, 2007) and major-element compositional variations were not considered. However, these experiments represent the best quantitative laboratory constraints currently available and offer an opportunity for direct comparison with seismological models. We note that the experimental results on which Faul and Jackson (2005) based their model have been recently reanalyzed and now indicate higher attenuation at low temperatures.

Fig. 4. (a) Identifying seismological shear velocity and attenuation points that fall inside and outside of the range defined by the mineral-physics model of Faul and Jackson (2005). The experimental range is shown by the green polygon and includes the velocity and attenuation values predicted for a range of grain sizes and activation volumes (see Fig. 3). Grey points are located within the experimental range, red points fall to the left of the range and blue points to the right. The map shows the geographical location of all points. Triangles indicate old-continental points, squares indicate oceanic regions ~70 Myr, and circles show oceanic regions N70 Myr. For a depth of 100 km. (b) Estimated changes in velocity as a result of corrections for laterally variable anelastic dispersion. Coloring of points and position of polygon are same as in a. Period = 75 s assumed for the calculations.
than the earlier work (Jackson and Faul, submitted to Phys. Earth Planet. Inter.). Furthermore, ongoing experimental work is exploring the effects of dislocation density and water on seismic properties (e.g., Jackson et al., 2009; Aizawa et al., 2008).

4.1. Anelastic dispersion

The Earth’s anelastic properties cause the attenuation of seismic waves and result in a frequency dependence of seismic velocity. Because seismic models such as S362ANI are determined from seismic data that span a wide range of frequencies, it is necessary to account for this frequency dependence and to reference all the data sets to a common frequency. This is an important step if one desires to separate elastic wave speed from the anelastic contribution and be able to make confident inferences about mantle temperature, composition, and physical state. In the development of S362ANI, this was done by assuming a one-dimensional attenuation model (QL6; Durek and Ekström, 1996) and applying the following dispersion correction

\[
v(\omega_2) \approx v(\omega_1) \left[ 1 + \frac{1}{\eta Q} \ln \left( \frac{\omega_2}{\omega_1} \right) \right],
\]

where \( \omega_2 \) and \( \omega_1 \) indicate the angular frequencies of interest and \( Q^{-1} \) is the relevant attenuation parameter (Liu et al., 1976).

![Fig. 5](image_url)

As in Fig. 4 but for a depth of 150 km.
frequency difference between \( \omega_1 \) and \( \omega_2 \) and/or a higher attenuation parameter both lead to a more substantial dispersion correction. For S362ANI, the reference frequency \( \omega_1 = 6.28 \, s^{-1} \) (period = 1 s).

For 1-D model QL6, \( Q_\mu = 70 \) in the depth range 80–220 km, and \( Q_\mu = 165 \) for depths 220–670 km. In the Earth’s upper mantle, lateral variations in attenuation are large (\( Q_\mu = 40–1000 \) at 100 km; Figs. 1 and 3), and ideally the velocity dispersion correction would account for three-dimensional variations in \( Q_\mu^{-1} \) in addition to the 1-D structure of QL6. To investigate how the velocities in S362ANI might be altered by allowing for a laterally variable dispersion correction, we perform a simple calculation using Eq. 1. We note that this back-of-the-envelope approach is only approximate, and ultimately this step should be formally applied to the data or the sensitivity kernels prior to the inversion for a velocity model.

In panel b of Figs. 4–7, we show the results of this approximate correction for laterally variable dispersion. For each of the 5762 locations, the parameter \( v(\omega_2) \) is calculated twice: once using the \( Q_\mu \) value from QL6, which is everywhere the same at a fixed depth, and once using the \( Q_\mu \) value from QRFSI12, which varies laterally and with depth. The difference between these two \( v(\omega_2) \) values is considered to be the amount by which the dispersion correction was under- or over-estimated in S362ANI, and the velocities are then shifted by this amount. Thus, if the \( Q_\mu \) from QRFSI12 is smaller than the QL6 value, the original dispersion correction to the 1-s reference period was too

Fig. 6. As in Fig. 4 but for a depth of 200 km.
small and velocities calculated assuming the higher attenuation value should be even larger than they are in S362ANI; these points are shifted to the right in panel b of Figs. 4–7. On the other hand, if attenuation in QRFSI12 is lower than the QL6 value, the original dispersion correction to the 1-s reference period was too large; accounting for this over-correction results in shifting velocities to the left in Figs. 4–7. In panel b of Figs. 4–6, it can be seen that points for which attenuation is less than 1/70 (0.014) are shifted toward lower velocities relative to panel a, and points for which attenuation is larger than 0.014 are shifted to higher velocities. In panel b of Fig. 7, the difference occurs below and above the attenuation value of 0.006 (1/168).

At depths of 100 and 150 km, the net effect of the laterally variable dispersion correction is to increase the number of points that fall within the experimental range and to decrease the deviation in velocity between the old-continental points and the experimental range. The effect is larger for the old-continental points than for the oceanic points: after the correction, 3% more oceanic points and 8% more cratonic points are located within the experimental range at 100 km. For 150 km, there are fewer oceanic points in the experimental range (1.5%) but considerably more old-continental points (8.3%). At 200 km, 5.5% fewer oceanic points and 18.1% more old-continental points fall within the experimental range. At 250 km, the lateral dispersion correction results in slightly fewer points within the
experimental range for all regions because the $Q_v$ value assumed in the construction of S362ANI (165 at this depth) is generally larger than the QRF5112 values. We calculate the lateral dispersion correction assuming a period of 75 s ($\alpha_w = 0.0838$ s$^{-1}$) at all depths, since the shallow mantle in S362ANI is mostly constrained by the intermediate-period surface waves (cf. Fig. 5 in Kustowski et al., 2008).

The purpose of the calculations performed in this section is to demonstrate, both qualitatively and quantitatively, how velocities in S362ANI might be altered if laterally variable attenuation was included in the anelastic dispersion correction. Using three-dimen-

sional attenuation values from QRF5112, we found that at 100 and 150 km, many low-velocity points would be shifted toward higher wave speeds, and many high-velocity points would be shifted toward lower wave speeds. This results in a larger number of points that fall inside the experimental range, and for those points located outside the experimental range, a smaller velocity or attenuation discrepancy that needs to be explained.

4.2. Mid-ocean ridges

While the points that fall to the right of the experimental range at 100-km and 150-km depth include old continents as well as areas beneath seafloor >70 Myr, almost all of the points to the left of the experimental range are centered on portions of the mid-ocean-ridge system (Figs. 4–5). These points are characterized by lower-than-

experimental velocities and/or lower-than-experimental attenuation. Using the thermally based mineral-physics model as reference, it is difficult to find mechanisms that could produce low attenuation at the mid-ocean ridge: cooler temperatures would lower attenuation but also increase velocity; enhanced water content relative to the dry olivine samples in the experiments would elevate attenuation; and the presence of partial melt would either elevate (Faul et al., 2004) or have minimal influence on (Hammond and Humphreys, 2000b) attenuation. We thus consider explanations for the anomalously low velocities in Figs. 4–5.

Given the geographical locations of these points near the mid-ocean ridge, it seems likely that the presence of partial melt is responsible for the velocity reductions. Compositional effects or enhanced water content could also decrease velocities, but their influence should be observable beneath oceanic crust of all ages (assuming they are controlled by melt extraction at the ridge), whereas the points of interest are located beneath young seafloor. Both the volume of melt present at depth as well as the details of the melt geometry control the way in which seismic-wave propagation is affected by the melt. Hammond and Humphreys (2000a,b) investigated the seismic effects of melt squirt and found that, for realistic melt geometries, shear velocity would be reduced and shear attenuation mostly unaffected. For a mantle with 1% partial melt, they reported ~8% decrease in shear-wave speed relative to the melt-free case.

Jackson et al. (2004) and Faul et al. (2004) measured velocity and attenuation of melt-bearing olivine in the laboratory at seismic frequencies. They observed enhanced attenuation as well as reduced velocity due to the presence of melt and inferred that the mechanism for these effects was elastically accommodated grain boundary sliding. The measured increase in attenuation was found to be large, more than doubling dissipation relative to the melt-free case. Since we do not observe such extreme attenuation values with the seismological models, our results indicate better consistency with the melt-squirt mechanism for velocity reduction. Yang et al. (2007) reached a similar conclusion from their regional seismic study of the East Pacific Rise. They called upon ~0.5% melt in the depth range 25–100 km beneath seafloor of age 2–6 Myr to satisfy their velocity data set.

The low velocities may indicate the presence of small amounts of melt beneath portions of the mid-ocean-ridge system to depths of 100 and 150 km. The velocity reductions with respect to the experimental range are ~0.05–0.1 km/s, which corresponds to melt fractions of 0.1–

0.3%, using the results of Hammond and Humphreys (2000a). The presence of melt at depths of at least 150 km is consistent with seismic and magnetotelluric results from the MELT experiment (e.g., The MELT Seismic Team, 1998; Evans et al., 1999). In addition, those studies reported melt extending laterally for several hundred kilo-

meters away from the ridge. Our results also indicate low velocities in a swath surrounding the East Pacific Rise, Pacific Antarctic ridge, and Southeast Indian ridge (Figs. 4–5). Given the coarse resolution of the global seismic models, it is difficult to accurately estimate the area in which melt would be present, and there are likely trade-offs between the amount of melt and the volume of mantle in which it is present. Partial melt at depths greater than expected for the dry-peridotite solidus can be explained by the addition of water (Asimow and Langmuir, 2003), carbonated peridotite (Dasgupta and Hirschmann, 2006), and/or a minimum in water solubility (Mierdel et al., 2007).

4.3. Cratonic lithosphere

In this section, we consider explanations for the old-continental points that fall to the right of the experimental range in Figs. 4–7. These points indicate higher velocity and/or higher attenuation than predicted; both possibilities are considered here. In particular, the dual influences of temperature and composition on the density and seismic velocity of the cratonic upper mantle has been a subject of interest for several decades (e.g., Jordan, 1978, 1979; Lee, 2003; Schutt and Lesher, 2006; Begg et al., 2009), and we focus on understanding what extent seismic effects due to major-element composition can explain the higher-than-experimental velocities.

Constraints on the temperature of continental lithosphere come primarily from two sources: surface heat-flow measurements (e.g., Pollack and Chapman, 1977) and mantle-xenolith chemistry (e.g., Rudnick et al., 1998, Artemieva (2006) compared xenolith geotherms with the heat-flow-derived geotherms of Pollack and Chapman (1977) and found that some cratonic xenoliths indicate a geotherm consistent with surface heat flow of 40–45 mW/m$^2$ whereas others are consistent with lower heat flow of 35–38 mW/m$^2$. In Fig. 8a, we reproduce these two end-member geotherms and consider them representative of xenolith constraints on the thermal state of cratonic lithosphere. Following Pollack and Chapman (1977), we assume a four-layer model with an upper and lower crust, a lithospheric mantle, and an asthenosphere. Heat flux from the lower crust and mantle is assumed to comprise 60% of the surface heat flow, with the remaining 40% derived from the upper crust. For a surface heat flow of 45 mW/m$^2$, this yields upper-crustal heat production of 1.8 $\mu$W/m$^2$; for surface heat flow of 37 mW/m$^2$, the heat production is 1.48 $\mu$W/m$^2$. In Table 1, we detail our assumptions about crustal and mantle radiative heat production and thermal conductivity. The hotter geotherm intersects the mantle adiabat at ~170-km depth (Fig. 8a), while for the cooler profile the intersection is located deeper, ~310 km.

Rudnick et al. (1998) have also compared xenolith P-T constraints with theoretical geotherms, using a similar approach. However, in that paper surface heat flow is fixed at 41 mW/m$^2$ and crustal heat production is varied in order to match the xenolith data. Ultimately they conclude that, assuming 0.03 $\mu$W/m$^2$ for heat production in the lithospheric mantle, crustal heat production must be fairly low: 0.4 to 0.5 $\mu$W/m$^2$. Geotherms calculated with those values are similar to the profiles in Fig. 8a.

We use the Perple_X software package (Connolly, 2005; www.perplex.ethz.ch) to determine mineral modes and compositions and shear-wave velocity for the two thermal models in Fig. 8a. The thermal effect on velocities is straightforward at a fixed pressure and results in lower wave speeds for the hotter geotherm (Fig. 8b), with the temperature sensitivity approximately described as $-3.4 \times 10^{-4}$ km s$^{-1}$ C$^{-1}$. To investigate the effects of composition on shear-wave speed, the calculation is repeated for several major-element bulk
compositions, which are summarized in Table 2. Pyrolite ( McDonough and Sun, 1995 ) is a model composition considered to represent primitive, fertile mantle that has not been depleted by melt extraction. The composition labeled “Slave craton” is the average composition of spinel peridotites from the Slave craton as sampled by the Jericho kimberlite ( Kopylova and Russell, 2000 ). Many cratonic xenoliths have depleted major-element compositions indicative of high degrees of partial melting ( e.g., Walter, 1998 ). For example, the removal of partial melt depletes the residue of certain oxides ( FeO, Na 2 O, CaO, and Al 2 O 3 ) and minerals ( garnet and clinopyroxene in particular), and it enriches the residue in olivine ( e.g., Boyd, 1989 ; Walter, 1998 ). We realize that it would be impossible to select one composition as representative of all cratonic upper mantle, chosen because of its depleted nature ( Table 2 ). As compared to pyrolite, this composition has a higher Mg# ( 100 × Mg/(Mg + Fe) ) and lower Al 2 O 3, Na 2 O, and CaO. The mineral modes calculated by Perple_X reflect the different bulk compositions, with less olivine and orthopyroxene and more clinopyroxene and garnet for pyrolite than “Slave craton.”

Two primary ways to increase seismic velocity through compositional effects are by increasing the Mg# of olivine and increasing the Al 2 O 3 content of the bulk rock. Forsterite has a higher shear modulus and lower density than fayalite ( e.g., Lee, 2003 ), and the aluminum content strongly controls the availability of garnet, a mineral characterized by large velocities ( e.g., Stixrude and Lithgow-Bertelloni, 2005 ). For the “Slave craton” composition, the net seismic effect is a slight velocity increase relative to pyrolite of −0.5% (0.02 km/s) at all depths ( Fig. 8b ). This higher velocity for “Slave craton” reflects a combination of factors: one, a velocity increase due to greater olivine content in general and larger olivine Mg# in “Slave craton” relative to pyrolite; two, a velocity decrease due to less garnet in “Slave craton”; and three, a velocity decrease due to the additional orthopyroxene, which is characterized by lower velocities than olivine and garnet, in “Slave craton.”

Deviations in the seismic velocity of many old-continental points from the experimental range in Figs. 4–7 are clearly larger than 0.02 km/s, especially at 100 and 150 km, where the deviations are as large as 0.05–0.1 km/s, even after the correction for anelastic dispersion ( Section 4.1 ). To explain such high velocities by composition would require a more depleted composition or higher Al contents than “Slave craton”. Periodotite xenoliths from Kaapvaal,

\[
\begin{array}{lcccccc}
\text{Layer} & \text{Thickness (km)} & \text{Heat production (W/m}^2\text{)} & \text{Thermal conductivity (W/m K)} \\
\hline
\text{Upper crust} & 10 & 1.80 or 1.48 & 2.5 \\
\text{Lower crust} & 30 & 0.25 & 2.5 \\
\text{Lithospheric mantle} & 80 & 0.01 & 3.0 \\
\text{Asthenosphere} & \text{–} & 0.084 & 3.0 \\
\end{array}
\]

Table 2: Bulk compositions assumed for and mineral modes determined from Perple_X calculations and used for velocity predictions in Fig. 8b. The mineral modes and shear-wave speeds vary with temperature and pressure, and the values presented here are only for a single temperature-pressure coordinate. Abbreviations: olv = olivine, opx = orthopyroxene, cpx = clinopyroxene, gt = garnet.
Jericho, and Siberia show Mg\# in the range 90–94 and Al\(_2\)O\(_3\) contents between 0.5 and 2.0 wt.\% (e.g., Kopylova and Russell, 2000; Boyd et al., 1997; Boyd, 1989; Lee, 2003). Motivated by these observations, we investigate the seismic properties of a hypothetical composition with higher Mg\# and higher (but still within the cratonic xenolith range) Al\(_2\)O\(_3\) than “Slab craton”; it is referred to as “Slave, more Mg & Al” in Table 2 and Fig. 8b. The differences between this hypothetical composition and “Slave craton” result in higher garnet content and olivine Mg\# and less abundant pyroxene. The hypothetical composition also produces higher shear-wave speeds than “Slave craton” at all depths by ~0.04 km/s (Fig. 8b).

For comparison with the seismological values, in Fig. 8b we have plotted the mean shear-wave speeds of the old-continental points that fall to the right of the experimental trend at depths of 100, 150, 200, and 250 km. Three sets of averages are shown: the mean of all such old-continental points, the mean of the largest 50% of these old-continental points, and the mean of the largest 10% of these old-continental points. Relative to the Perple\(_X\) velocity predictions, the mean of all the points (green stars) falls on or to the left of the velocity profiles calculated for the cold geotherm and pyrolite, suggesting that many of the high-velocity old-continental points can be explained by reasonable cratonic compositions and temperatures that fall within the range prescribed by the xenolith P-T constraints. Considering now the mean of the 50% highest-velocity points (cyan stars), they are consistent with a pyrolitic or “Slave craton” composition and the cold geotherm at depths of 100, 200, and 250 km. However, at 150 km, the seismological values require larger velocities than can be obtained by these conditions, and this is also true for the 10% highest-velocity points at all depths (magenta stars). To explain these large velocities, the seismic model values require either very depleted compositions with abundant garnet or colder temperatures than suggested by the xenolith P-T data (Fig. 8a). If we assume the “Slave craton” composition and a temperature derivative \(\frac{\Delta V}{\Delta T} = -3.4 \times 10^{-4}\) km\(s^{-1}\)°C\(^{-1}\), the temperatures required to produce these high velocities must be 100–200 °C colder than the 37-mW/m\(^2\) geotherm (magenta stars in Fig. 8a). Such low temperatures are not consistent with xenolith P-T constraints (Rudnick et al., 1998) or most surface-heat flow data (e.g., Nyblade and Pollack, 1993).

We note that the calculation of seismic properties with Perple\(_X\) accounts for elastic effects only; the anelastic contribution to velocities, which reduces velocities at high temperatures and in the presence of melt (Jackson et al., 2002; Faul et al., 2004), is neglected here. As such, the predicted wave-speed curves in Fig. 8b represent maximum values and would, in some cases, be lower, thereby widening the gap between the model and the seismological values.

To summarize this section, many of the old-continental points that fall outside of the experimental range in Figs. 4–7 have shear-wave speeds that are consistent with reasonable cratonic temperature and composition. Some of these points, however, may require a more depleted composition than the range explored here, and future work will investigate whether xenolith compositions and seismic-velocity models can be reconciled in these areas.

A complication arises when the seismological attenuation and velocity models are considered together. For the cold geotherm (37 mW/m\(^2\)), the mineral-physics model (Faul and Jackson, 2005) predicts minimal attenuation at 100 and 150 km and \(Q_v^{-1}\) values of 0.006 and 0.004 at 200 and 250 km, respectively. However, the seismological values in Figs. 4–5 indicate larger attenuation associated with all high-velocity points, in contrast to the mineral-physics predictions. According to the model of Faul and Jackson (2005), to obtain \(Q_v^{-1} = 0.005\) requires temperatures of 1050 °C and 1090 °C at 100 and 150 km, respectively. While these temperatures overlap with the hotter geotherm in Fig. 8a, the high velocities at 150 and 200 km could not be achieved at such warm conditions, even with a highly depleted composition.

It is difficult to produce the non-zero seismological attenuation at the cold temperatures required to explain the high velocities. We note that reanalysis of some of the experimental data on which Faul and Jackson (2005) based their model indicates larger attenuation at low temperatures than previously realized (Jackson and Faul, submitted to Phys. Earth Planet. Inter.), which may help to explain some of this discrepancy. It is also possible that non-zero attenuation values associated with the high velocities in Figs. 4–7 could be artefacts of the attenuation-imaging process. Attenuation tomography is especially challenging because amplitudes, which are the data used in most studies, require a complex interpretation, as mentioned in Section 1. In particular, the effects of multiple scattering are not accounted for in the construction of QRF512 and could bias the model toward high attenuation in certain regions. As compared to short-wavelength body waves, this effect should be much weaker for the long-wavelength surface waves used to develop QRF512. However, it is possible that intermediate- and large-scale heterogeneity (such as individual cratonic blocks and the ocean-continent interface) could be effective scatterers of long-wavelength energy, and the redirection of that energy would masquerade as seismic attenuation. Future work will try to address this issue using accurate three-dimensional wave-propagation modeling as well as comparison of global models with regional and local data sets.

It is clear that many of the old-continental points (>50%) fall outside of the experimental range at depths of 100–200 km, and we have suggested that a more depleted bulk composition may help explain the very high velocities. However, at 250 km, only a small fraction of the old-continental points fall outside the experimental range (~15%), suggesting that chemically distinct cratonic upper mantle is not required by the seismic data at depths >200 km. A chemical boundary layer beneath cratons extending to ~175 km has been suggested by Lee et al. (2005) based on xenolith thermobarometric data and major-element compositions. The seismological model values appear to support the existence of such a layer to depths of 200 km globally.

5. Conclusions

We present a comparison of seismological models and experimental measurements of shear-wave velocity and attenuation on a global scale. We show that attenuation and velocity are strongly anti–correlated in the upper mantle, and that there are distinct regional depth-dependent trends in the relationship between the two quantities. We assume that predictions of the experimentally derived model describe the seismic effects of lateral temperature variations in dry melt-free olivine. Thus, regions for which the seismological and experimental trends agree suggest that lateral temperature variations in dry melt-free olivine control much of the seismic variability. We explore explanations in addition to lateral temperature variations for those regions in which the seismological and experimental trends do not align. In oceanic regions, the seismic models agree well with the experimental predictions at depths \(\geq 150\) km, and we attribute the many low-velocity outliers at 100 km to small amounts of partial melt. In cratonic regions, the seismological trend is oblique to the experimental trend for depths \(<200\) km. We are able to explain some of this mismatch by laterally variable anelastic dispersion and some by increased velocity due to chemically depleted lithosphere. It remains difficult to reconcile the highest velocities, which indicate relatively cold conditions, with observations of non-zero attenuation, which require elevated temperatures.

References


