1 **The thermal structure of cratonic lithosphere from global Rayleigh wave attenuation**

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Abstract

The resolution of and level of agreement between different attenuation models has historically been limited by complexities associated with extracting attenuation from seismic-wave amplitudes, which are also affected by the source, the receiver, and propagation through velocity heterogeneities. For intermediate- and long-period Rayleigh waves, removing the amplitude signal from focusing and defocusing effects is the greatest challenge. In this paper, three independent data sets of fundamental-mode Rayleigh wave amplitude are analyzed to investigate how three factors contribute to discrepancies between the attenuation models: uncertainties in the amplitude measurements themselves, variable path coverage, and the treatment of focusing effects. Regionalized pure-path and fully two-dimensional attenuation models are derived and compared. The approach for determining attenuation models from real data is guided by an analysis of amplitudes measured from synthetic spectral-element waveforms, for which the input Earth model is perfectly known. The results show that differences in the amplitude measurements introduce only very minor differences between the attenuation models; path coverage and the removal of focusing effects are more important. The pure-path attenuation values exhibit a clear dependence on tectonic region at shorter periods that disappears at long periods, in agreement with pure-path phase-velocity results obtained by inverting Rayleigh wave phase delays. The 2-D attenuation maps are highly correlated with each other to spherical-harmonic degree 16 and can resolve smaller features than the previous generation of global attenuation models. Anomalously low attenuation is nearly perfectly associated with continental cratons. Variations in lithospheric thickness are determined by forward modeling the global attenuation variations as a thermal boundary layer of variable thickness. Temperature profiles that satisfy the attenuation values systematically overpredict and underpredict Rayleigh wave phase velocity in cratons at short and long

periods, respectively. Introducing a low-velocity layer at depths 60-80 km and a high-velocity layer that begins at 200 km can resolve the discrepancy. The former is consistent with receiver-function detections of a mid-lithospheric discontinuity, and the latter may correspond to the Lehmann discontinuity.

Keywords: seismic attenuation, surface waves, tomography, lithospheric thickness, mid-lithospheric discontinuity, mantle temperature

1. Introduction

Seismic-wave attenuation ($1/Q$) offers great potential for revealing the physical and chemical state of the mantle, complementing the information provided by seismic velocity. For example, coherent upwelling flow between the lower and upper mantle has been inferred on the basis of agreement between the patterns of attenuation in the mantle transition zone, low shear velocities in the lower mantle, and hotspot locations at the surface (Romanowicz and Gung, 2002). Attenuation in the upper mantle can help to identify the presence of water (Rychert et al., 2008) and melt (Abers et al., 2014). It can also be used to place bounds on the range of mantle potential temperatures (Dalton et al., 2009) and to account for anelastic effects on seismic velocity, yielding more accurate estimates of temperature from wave speed (e.g., Goes and van der Lee, 2002).

However, challenges in isolating the signal of attenuation in seismic data have historically hindered the development of attenuation models, limiting both the resolution of the models and the level of agreement between different attenuation studies (see Fig. 17 of Dalton et al., 2008). In addition to attenuation, surface-wave amplitudes are affected by excitation at the earthquake source, local structure and the instrument response at the receiver, and propagation effects such as focusing, defocusing, and scattering. These factors, if not accounted for, can be erroneously mapped into attenuation, biasing the resulting values (Selby and Woodhouse, 2000; Dalton and Ekström, 2006; Yang et al., 2007). It has even been suggested that most of the observed seismic attenuation could be explained by elastic heterogeneities rather than dissipative mechanisms (Ricard et al., 2014), although wavefield simulations have shown that elastic scattering will enhance attenuation of fundamental-mode Rayleigh waves by at most a few percent (Dalton et al., 2014; Meschede and Romanowicz, 2015).

Source and receiver effects on amplitudes can be reasonably well estimated as long as each event is recorded by a large number of stations and each station records a large number of earthquakes (Ekström et al., 2006; Ferreira and Woodhouse, 2006; Dalton et al., 2014; Ma et al., 2016). Focusing effects, on the other hand, are controlled by gradients in seismic velocity and therefore depend on both the characteristics of velocity heterogeneity and the orientation of the propagation path relative to velocity heterogeneity. While full-waveform inversions, which have begun to produce high-resolution regional models of mantle attenuation (Zhu et al., 2013), can account for focusing effects in a self-consistent way, all other attenuation studies require the use of approximate theory. The great-circle ray approximation (GCRA; Woodhouse and Wong, 1986) treats the surface wave as a thin ray confined to the great-circle path, exact ray theory (ERT; Woodhouse and Wong, 1986; Wang and Dahlen, 1994; Larson et al., 1998) treats the surface wave as a thin ray that is allowed to bend according to elastic structure, and finite-frequency theory (FFT; Zhou et al., 2004) distributes the surface-wave sensitivity over a broad region surrounding the great-circle path. Dalton et al. (2014) evaluated the validity of each theoretical treatment by computing spectral-element synthetic seismograms using a 3-D velocity model and a 1-D attenuation model. They also showed that while failure to remove focusing effects from amplitudes introduced shorter-wavelength artifacts into attenuation models, the longer wavelengths can be faithfully recovered, confirming the findings of earlier observational studies (Romanowicz, 1995; Selby and Woodhouse, 2000).

Recently, three independent studies have produced global long-wavelength Rayleigh wave attenuation maps that exhibit a high level of agreement (Dalton and Ekström, 2006; Ma et al., 2016; Bao et al., 2016). While all three studies account for source, receiver, and focusing effects, they use different amplitude data sets and different approaches for the removal of focusing effects. The attenuation maps are dominated by tectonic-scale features

96 at short and intermediate periods, including high attenuation along mid-ocean ridges, low
97 attenuation beneath old seafloor, and low attenuation associated with stable continental
98 interiors. The correlation coefficients between each pair of 50-s maps are > 0.85 at
99 spherical-harmonic degree 12; to provide some context for this value, Dalton and Ekström
100 (2006) report correlation coefficients between their results and the early studies of
101 Romanowicz (1995) and Selby and Woodhouse (2000) in the range 0.24-0.44.

102 This general consistency between the three independent sets of attenuation maps
103 represents significant progress in global attenuation imaging. In this paper we analyze the
104 three observed amplitude data sets and solve for Rayleigh wave attenuation with both
105 regionalized and generic 2-D parameterization schemes. By treating consistently all three
106 data sets we can assess the robustness of features in the models and identify the origin of
107 the differences between the maps. The synthetic amplitudes generated by Dalton et al.
108 (2014) guide this analysis by allowing an investigation into how erroneous features can be
109 introduced into attenuation maps as a result of unmodelled propagation effects.

110 In the new attenuation maps, which provide a higher resolution than the earlier
111 studies, there is an excellent correlation between areas of very low attenuation and
112 Precambrian crust. Although these same areas are also characterized by fast seismic
113 velocity, challenges associated with separating the competing effects of temperature and
114 composition on velocity have made it difficult to extract reliable cratonic geotherms from
115 velocity models (e.g., Pedersen et al., 2009; Lebedev et al., 2009; Hirsch et al, 2015).
116 Attenuation, on the other hand, is typically considered to be much less sensitive to
117 compositional variations and should be a more useful proxy for mantle temperature,
118 especially if temperatures are low enough to minimize anelastic effects related to melt and
119 volatiles. Here, we demonstrate that the temperature structures that are compatible with
120 the Rayleigh wave attenuation values agree with the constraints provided by xenolith

thermobarometry and surface heat flow. These independent constraints on mantle temperature make it possible to determine the effect of mantle composition on shear velocity in cratonic lithosphere.

2. Rayleigh Wave Amplitudes and Focusing Effects

This study utilizes four different global data sets of fundamental-mode Rayleigh wave amplitudes (Table 1); three of these were measured from observed waveforms and one was measured from synthetic waveforms. The amplitudes used by Dalton et al. (2008) (hereinafter DED08) were measured using the algorithm of Ekström et al. (1997) applied to earthquakes with $M_w > 6.0$ that occurred during 1993-2005. The amplitude data used by Ma et al. (2016) (hereinafter MMM16) were measured by a cluster analysis technique applied to earthquakes with $M_w > 5.0$ that occurred during 1991-2007. The amplitude data used by Bao et al. (2016) (hereinafter BDR16) were measured using the mode-branch stripping technique of van Heijst and Woodhouse (1997) applied to earthquakes that occurred during 1991-2007. All three observed amplitude data sets account for the effects of the source, geometrical spreading, attenuation, and the receiver in a 1-D Earth using PREM (Dziewonski and Anderson, 1981) and the Global CMT solutions (Dziewonski et al., 1981; Ekström et al., 2012); the DED08 and BDR16 data were measured with respect to reference waveforms, whereas MMM16 computed and applied the corrections to their measurements. We pre-calculate and remove focusing effects with the velocity models used in the original studies for BDR16 and MMM16 and with the GDM52 maps of Ekström (2011) expanded in spherical harmonics to degree 20 for DED08 (Table 1). Finally, we remove paths for which the earthquake is not recorded at > 30 receivers or the station does not record > 30 earthquakes. Overall the three observed amplitude data sets are in strong agreement (Fig. S1). For source-receiver paths held in common between pairs of data sets, the variance

146 reductions are 82-93% and the correlation coefficients range between 0.92 and 0.96. The
147 fourth set of amplitude data was measured from spectral-element synthetic seismograms
148 that were generated using the software package SPECFEM3D_GLOBE (Komatitsch and
149 Tromp, 2002a,b). The wavefield simulations, described in detail by Dalton et al. (2014),
150 used a global 3-D elastic Earth model, a 1-D attenuation model, and 42 globally distributed
151 earthquakes. Focusing has been pre-calculated using ERT, GCRA, and FFT applied to phase-
152 velocity maps computed from the 3-D model used for the simulation.

153 Fig. 1 compares predictions of the focusing amplitude for 50-s Rayleigh waves. The
154 outliers tend to dominate the scatter plot and obscure the fact that for the majority of paths
155 the two sets of predictions are close to the 1:1 line. The comparisons in Fig. 1 also show that
156 the relationship between the GCRA and FFT predictions is nearly linear: both approaches
157 contain the assumption of propagation along the great circle and a focusing prediction that
158 varies linearly with the phase-velocity perturbation. Furthermore, the relationships
159 between the GCRA/ERT and FFT/ERT predictions are nonlinear due to the log-normal
160 distribution of ERT log-amplitudes versus the normal distribution of GCRA and FFT
161 predictions (Larson et al., 1998). The validity of each theoretical treatment for focusing
162 effects is assessed through comparison with the amplitudes measured from the spectral-
163 element synthetic seismograms, from which source and receiver effects have been removed
164 so that the remaining signal can be considered to have originated during propagation (Fig.
165 S2). The ERT predictions provide the best fit to the 50-s measurements (Dalton et al., 2014)
166 and reduce the variance of the measurements by 15%; GCRA and FFT do not provide
167 positive variance reduction at 50 s. For 75-s and 125-s Rayleigh waves, only FFT provides
168 positive variance reduction: 14% at 75 s and 27% at 125 s, demonstrating that the validity
169 of the theoretical treatments depends on period.

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Figure 1. Comparison of focusing-predicted amplitudes for 50-s Rayleigh waves using a degree-20 phase-velocity map and three approximate theories. Blue circles show all 4,749 paths. Yellow circles show 80% of paths with the smallest absolute difference between the two predictions. Red and dashed black lines are the best-fitting lines using orthogonal regression applied to the 80% subset of points and all points, respectively.

The effect of unmodelled focusing effects on the attenuation models obtained by inverting amplitude data is presented in Fig. 2. A synthetic amplitude data set that is sensitive to both attenuation and focusing effects is generated by multiplying the spectral-element amplitudes (Fig. S2) by amplitudes computed by integrating along great-circle paths through an input attenuation map that was created by scaling the 50-s phase-velocity map by a factor that produces attenuation variations with a similar magnitude to those observed for the real Earth. This synthetic data set is then inverted for attenuation structure with four different treatments of focusing effects: none, GCRA, ERT, and FFT. Figs. 2c,d demonstrate that (i) the short-wavelength attenuation features are most strongly impacted by the treatment of focusing effects, and (ii) using ERT at 50 s and FFT at 125 s produces an output attenuation map that is most similar to the input map. Importantly, all three theoretical treatments yield output maps that are greatly improved relative to the case where focusing effects are not removed.

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Figure 2. (a) Input attenuation map, created by scaling a 50-s phase-velocity map. (b) Output attenuation map obtained by inverting an amplitude data set that has sensitivity to lateral attenuation variations and unmodelled path effects such as focusing. Focusing effects predicted with ERT have been removed from the amplitude data. Both maps are expanded in spherical harmonics to degree 20. (c) Correlation coefficient at each degree between the input attenuation map and four output maps. Blue: focusing effects not removed from amplitude data. Red, green, and black show focusing effects removed with GCRA, ERT, and FFT, respectively. (d) Results for the same experiment performed for 125-s Rayleigh waves.

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204 3. Pure-Path Regionalized Inversions

205 As an initial exploration into the constraints on lateral attenuation structure that are
 206 suggested by the three observed data sets of Rayleigh wave amplitude, we use the
 207 amplitude measurements to estimate attenuation anomalies in the six tectonic regions
 208 identified by Jordan (1981) in the Global Tectonic Regionalization (GTR1) scheme. GTR1
 209 contains three oceanic regions, distinguished by seafloor age: young (A; 0-25 Myr),
 210 intermediate age (B; 25-100 Myr), and old (C; >100 Myr). The continents are divided into
 211 three regions on the basis of their tectonic behavior during the Phanerozoic: platforms

overlain by Phanerozoic cover (P), Phanerozoic deformation zones (Q), and exposed Archean and Proterozoic shields and platforms (S). We refer to regions P and S together as stable continents or cratons.

Each amplitude measurement $A_{ij}(\omega)$, corresponding to earthquake source i and seismic receiver j , is considered to depend on four factors

$$A_{ij}(\omega) = A_i^S(\omega) A_j^R(\omega) A_{ij}^F(\omega) A_{ij}^Q(\omega),$$

where ω indicates the angular frequency and the superscripts S , R , F , and Q refer to source, receiver, focusing/defocusing, and attenuation effects, respectively. The amplitude data are sensitive to perturbations in Rayleigh wave attenuation $\delta(Q^{-1})$ away from the value predicted by PREM integrated along the ray path $ds(\theta, \phi)$

$$A^Q(\omega) = \exp \left[\frac{-\omega}{2U(\omega)} \int \delta(Q^{-1})(\omega, \theta, \phi) ds(\theta, \phi) \right],$$

where $U(\omega)$ is group velocity, θ is latitude, and ϕ is longitude. Since the attenuation effect, unlike the focusing effect, accumulates linearly, using the great-circle path instead of exact ray theory or finite-frequency theory is appropriate for the relatively long-wavelength variations of interest here (e.g., Wang and Dahlen, 1994; 1995). The pure-path regionalized inversion determines, in addition to a frequency-dependent attenuation anomaly $[\delta(Q^{-1})]_k(\omega)$ in the k th tectonic region, a scalar correction for each source and each receiver:

$$\frac{-2U}{\omega} \ln[A_{ij}(\omega)/A_{ij}^F(\omega)] = \frac{-2U}{\omega} \{ \ln[A_i^S(\omega)] + \ln[A_j^R(\omega)] \} + \sum_{k=1}^6 X_{ij}^k [\delta(Q^{-1})]_k(\omega)$$

where X_{ij}^k indicates the path length through the k th tectonic region. Focusing effects, if accounted for, are pre-calculated and removed from each datum prior to inversion. The unknown parameters are solved for with least-squares minimization using Cholesky factorization; the only *a priori* constraint is requirement that the $\ln[A_j^R]$ terms sum to zero.

Attenuation anomalies obtained using the three different amplitude data sets (uncorrected for focusing effects) are largely consistent (Fig. 3a), although the three sets of values are offset slightly from each other. The offset can be attributed almost entirely to path coverage; when the inversion is repeated using only the paths shared in common by all three data sets the offset disappears. The pure-path attenuation results obtained from the SPECFEM amplitudes, for which the input attenuation model was uniform, show a very different pattern.

When focusing effects are removed from the observed amplitude data, the result is to increase the global range of estimated attenuation values (Fig. 3b) by enhancing attenuation in young oceans (A) and active continents (Q) and reducing attenuation in old oceans (C) and stable continents (regions P and S). While the sign of this adjustment is consistent for almost all types of focusing predictions, the magnitude of the correction depends on the approximate theory and phase-velocity map used. Removing focusing effects from the SPECFEM amplitudes adjusts the attenuation values in the same direction as for the observed data, which has the effect of reducing the global range by nearly a factor of two. The range of $1/Q$ values before and after the removal of focusing effects from the SPECFEM amplitudes is 0.0069-0.0112 and 0.0070-0.0094, respectively; the input Rayleigh wave attenuation at 50 s was 0.00866. The difference between the input and output attenuation values (with focusing effects removed) corresponds to an error of 6-10% for regions A, B, C, P, and Q and 18% for region S; in Section 5 these errors are assumed to

apply at all periods and used to constrain mantle thermal structure. The adjustment to a smaller global range after focusing effects are removed from the SPECfEM data is of the opposite sense from the adjustment to a larger global range observed with real data and is consistent with the uniform attenuation structure that was prescribed for the SPECfEM simulations. The ability of the pure-path inversion to recover the input anelastic structure reasonably well suggests that the attenuation anomalies obtained from the real data accurately represent the anelastic properties of the Earth's interior.

Fig. 3c shows the GTR1 attenuation anomalies obtained with the MMM16 amplitudes corrected for focusing effects using FFT (Table 2); results for BDR16 at 50 and 100 s and DED08 at 50, 75, 100, and 125 s are shown for comparison. Fig. S3a summarizes the variance reduction relative to that provided by only the source and receiver factors. The pure-path results provide the best fit to the data at short periods ($\sim 14\%$) and almost no variance reduction at the longest periods ($\sim 1.4\%$). Fig. S4 illustrates the depth sensitivity of Rayleigh waves of different periods to the Earth's intrinsic shear and bulk attenuation.

Fig. 3d shows phase-velocity perturbations for the GTR1 regions, obtained by inverting the Rayleigh wave time delays $\delta t_{ij}(\omega)$ measured by Ma et al. (2014) for pure-path phase-slowness anomalies δp_k in each GTR1 region

$$\delta t_{ij}(\omega) = \sum_{k=1}^6 X_{ij}^k \delta p_k(\omega).$$

Results obtained from inverting the eigenfrequency perturbations of Ritsema et al. (2011) are also shown, and variance reduction is plotted in Fig. S3b.

The attenuation and phase-velocity anomalies are anti-correlated with each other at short and intermediate periods. Excluding temporarily the 40-s results, the global range of attenuation and phase velocity is largest at 50 and 66 s and minimum at 200 s, although the

stable continents (P,S) are characterized by low attenuation and high velocity even at 200 s. The three oceanic regions (A,B,C) show a clear dependence on seafloor age, with the highest attenuation and slowest velocity at young ages. Stable continents reach their minimum attenuation anomaly and maximum velocity anomaly at 66 s. The 40-s results are consistent with a stronger influence of the crust on phase velocity than on attenuation. Fig. 3d shows that relative to PREM the three oceanic regions are characterized by a slight reduction in the velocity anomaly between 40 and 50 s whereas the three continental regions exhibit a velocity increase, which is large for the stable continents. The attenuation curves do not show such an obvious distinction between oceanic and continental regions in this period range. We also note that while intermediate-age oceans (B) and active continents (Q) exhibit very similar attenuation values at all periods, phase velocity for Q is much lower than for B at periods < 100 s. These differences between the oceanic and continental observations and between the phase-velocity and attenuation curves can be explained if the well-known velocity increase with depth that occurs at the continental Moho is not accompanied by a corresponding reduction in attenuation. Most mantle attenuation studies contain the assumption of a low-absorption crust (e.g., Durek and Ekström, 1996), which the results in Fig. 3c-d support.

The values of attenuation and phase velocity obtained using the GTR1 regionalization can be considered representative of the seismic properties of the crust and shallow mantle. This conclusion is evident from the fact that these values provide higher variance reduction at shorter periods and is to be expected given that the GTR1 regionalization scheme is guided by surficial properties such as seafloor age and tectonic behavior. The tendency of the values to converge at long periods reflects weakening correspondence between surface-tectonic features and the patterns of elastic and anelastic

309 heterogeneity at greater mantle depths. In Section 5, we explore the upper-mantle
310 temperature variations that are suggested by both sets of values.

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Figure 3. Rayleigh wave attenuation and phase velocity determined for the six tectonic regions of GTR1 (Jordan, 1981). All attenuation inversions also solve for source and receiver terms. (a) 50-s attenuation ($1/Q$) determined from the three amplitude data sets, with no correction for focusing effects. Solid lines: all available data used. Dashed lines: only paths shared by all three data sets used. Black curve shows the attenuation determined from the synthetic SPECFEM amplitudes. (b) Effect of accounting for focusing effects on the regional attenuation values. Separate results are shown for focusing effects predicted with ERT, FFT, and GCRA. Blue=BDR16 amplitudes. Grey=SPECFEM amplitudes. Attenuation anomalies are plotted as absolute deviations from the values in (a). (c) Regional attenuation anomalies, expressed as absolute perturbations to the PREM $1/Q$ values. See Fig. 6c for absolute attenuation values and error bars. Solid: MMM16 amplitudes with focusing effects removed using FFT. Dashed: DED08 amplitudes with focusing effects removed using GCRA. Dotted: BDR16 amplitudes with focusing effects removed using ERT. (d) Regional phase-velocity anomalies, expressed relative to PREM phase velocities. Solid: the travel-time data set of Ma et al. (2014). Dashed: the eigenfrequency-perturbation data set of Ritsema et al. (2011).

4. Maps of Rayleigh Wave Attenuation

The three sets of observed Rayleigh wave amplitudes A_{ij} are inverted for attenuation variations expanded in spherical harmonics to degree 16 and correction factors for each source i and each receiver j :

$$\frac{-2U}{\omega X_{ij}} \ln[A_{ij}(\omega)/A_{ij}^F(\omega)] = \frac{-2U}{\omega X_{ij}} \{\ln[A_i^S(\omega)] + \ln[A_j^R(\omega)]\} + \sum_{l=0}^L \sum_{m=-l}^l q_{lm} \overline{Y_{lm}}.$$

In this equation X_{ij} is the path length, q_{lm} are the spherical-harmonic coefficients to be determined, $\overline{Y_{lm}}$ is the path average of the spherical-harmonic function, and L is the maximum degree of the spherical-harmonic expansion. In each case, focusing effects can be pre-calculated and removed from the amplitudes prior to inversion. A smoothness constraint is applied by minimizing the squared gradient of the attenuation perturbation, as described by Dalton and Ekström (2006), using the same smoothness coefficient for all scenarios.

Fig. 4 shows the attenuation maps at 50 s and 100 s; variance reduction is summarized in Fig. S3a. These degree-16 maps are able to resolve specific features that were not present in previous global attenuation studies. Most notably, low attenuation is nearly perfectly associated with continental cratons. Fig. 5a shows the crustal types in CRUST1.0 (Laske et al., 2013) grouped by age. Very low attenuation beneath specific continental cratons can be seen in this comparison, for example the Amazonia craton in South America, the West Africa, Congo, and Kaapvaal cratons in Africa, the Baltic Shield and Russian Platform in western Eurasia, and the Siberian craton in eastern Eurasia. The attenuation models contain low attenuation in western Australia, although it appears to align more closely with the Pilbara craton and Proterozoic rocks in the northwest than it does with the Archean Yilgarn craton in the southwest. We also note that all three

attenuation maps contain a zone of very low attenuation just immediately northeast of the Caspian Sea in an area that is characterized as Phanerozoic in CRUST1.0.

In the 50-s maps, high attenuation is found beneath tectonically active western North America, eastern Africa, the Lau back-arc spreading center, and extending inland from the Nigerian coast, roughly aligned with the Cameroon Volcanic Line. Attenuation also exhibits a clear dependence on seafloor age, with higher attenuation at younger ages, as expected for a lithosphere that cools and thickens over time. This seafloor-age relationship is characterized by greater scatter than is the case for the age dependence of phase velocity, and the correlation coefficient between the 50-s MMM16 attenuation and GDM52 phase-velocity maps, which is -0.58 globally, is weaker in oceanic areas and stronger in continental areas: -0.49 for the 44% of pixels with crustal thickness 5-9 km, -0.67 for the 6% of pixels with thickness 28-32 km, and -0.75 for the 11% of pixels with crustal thickness 37-41 km. The weaker oceanic age-dependence for the attenuation maps could originate from imperfect corrections for focusing effects or it could also reflect differences in the sensitivity of attenuation and velocity to factors like melt and volatiles.

Figure 4. Degree-16 attenuation maps for 50-s and 100-s Rayleigh waves. These maps are determined using all possible paths for each data set. (a) MMM16 amplitudes, corrected for focusing effects using FFT. (b) BDR16 amplitudes, corrected for focusing effects using ERT at 50 s and FFT at 100 s. (c) DED08 amplitudes, corrected for focusing effects using GCRA. White diamonds show hotspot locations. The color scale shows absolute deviations in attenuation ($1/Q$) away from the attenuation values predicted by PREM: 0.00635 at 50 s and 0.00846 at 100 s.

The correlation coefficient between each pair of maps in Fig. 4 ranges between 0.77 and 0.80 at 50 s and 0.51 and 0.83 at 100 s (Fig. 5b). The lack of perfect correlation results from a combination of two factors: the different path coverage of the three data sets and the different corrections for focusing effects. When identical path coverage is used and focusing effects are not removed, the maps are nearly identical. This result demonstrates that

differences in the algorithms that measure the Rayleigh wave amplitudes have a negligible effect on the retrieved attenuation structure. Removing focusing effects from the paths held in common between the data sets reduces the correlation, as expected because focusing effects are treated slightly differently in the three cases. Fig. S5 illustrates how the use of different theories applied to a single phase-velocity map and data set influences the attenuation structure. When the same theory and phase-velocity map are used to predict focusing effects for all three data sets (i.e., focusing effects are treated consistently in the three cases), the correlation is stronger than when focusing effects are not removed (Fig. S6). Using all available paths instead of only the common paths changes the agreement between pairs of models in two ways (Fig. 5b). One, when focusing effects are not removed the correlation is weakened relative to when only common paths are used. Two, the correlation between maps with focusing effects removed is stronger than when focusing effects are not removed; using the same theory and phase-velocity map to predict focusing effects for all three data sets further strengthens this correlation (Fig. S6).

The analysis described in the previous paragraph makes clear that not removing focusing effects introduces artifacts into the maps, and removing focusing effects eliminates many of these artifacts. When only paths held in common are used, the different focusing corrections applied to each data set make the data sets less similar to each other and, as a result, the attenuation maps are less similar. When all paths are used, the artifacts, since they are not true features of the Earth, will depend on which paths are included; the attenuation maps without focusing effects removed show considerable disagreement. Removing focusing effects results in attenuation maps that are more representative of the Earth's true anelastic structure and thus more similar to each other. Therefore, consistent with the experiment performed on the SPECFEM amplitudes (Fig. 2), any of the three approaches for removing focusing effects improves the recovery of the input model relative

to not removing focusing effects at all. At 100 s the effects of path coverage and removal of focusing are similar for the BDR16 and MMM16 comparison. The comparisons that utilize the DED08 model indicate that removal of focusing effects results in a slightly weaker correlation, perhaps because GCRA is especially poor at predicting focusing effects at longer periods (Fig. S2c).

Figure 5. (a) Classification of crust according to CRUST1.0 (Laske et al., 2013). Dark, intermediate, and light blues indicate continental crust with early Proterozoic/Archean, mid-late Proterozoic, and younger ages, respectively. Seafloor is classified on the basis of age: 0-25 Myr, 25-100 Myr, and > 100 Myr according to Müller et al. (2008). (b) Correlation coefficient between each pair of degree-16 maps. Red curves correspond to maps determined with all possible paths for each data set; blue curves correspond to maps determined with identical path coverage. Solid and dashed lines correspond to maps determined from the raw and focusing-corrected amplitudes, respectively. Numbers of paths in common is given in parentheses.

5. Inferring Upper-Mantle Temperatures

5.1 Constructing Geotherms

The single largest factor contributing to the global attenuation values in the upper mantle is temperature; it is well established that temperature variations are large (e.g., Goes et al., 2000; Artemieva, 2009) and that attenuation is strongly controlled by temperature

(e.g., Jackson and Faul, 2010; McCarthy et al., 2011). Although attenuation can also be enhanced by the presence of melt (Jackson et al., 2004; McCarthy and Takei, 2011) and possibly water (e.g., Karato, 2003), we focus our interpretation below on stable continental areas where these factors likely contribute little to the observed attenuation. It is not straightforward to interpret the Rayleigh wave attenuation values in Figs. 3-4 in terms of temperature variations at a single depth, since they are a depth-integrated quantity. Rather, we investigate how well the attenuation variations can be explained by a simple thermal structure: a thermal boundary layer (TBL) that joins the mantle adiabat at some depth.

Temperature profiles are constructed by specifying the mantle adiabat, defined by a mantle potential temperature (T_p) and a temperature gradient with depth, and the depth at which the TBL intersects the adiabat. We consider 33 different adiabatic profiles, with 11 values of T_p (1100°C-1600°C) spaced every 50°C and three values of the adiabatic temperature gradient dT_a/dz (0.25°C/km, 0.40°C/km, 0.55°C/km). We consider 80 values for the intersection depth spaced every 5 km between 5 km and 400 km. The result is 2,640 different temperature profiles from the surface to 400-km depth. The advantage of this approach is that it is straightforward to map global variations in TBL thickness. Fig. S7 shows that our approach works particularly well in settings where the uppermost mantle is generally cold. Relative to our temperature profiles, realistic cratonic geotherms that incorporate radiogenic heat production in the crust and mantle (e.g., Rudnick et al., 1998) have slightly higher temperatures at depths < 50 km. However, since attenuation is very low at these temperatures, the temperature differences produce only minor differences in the predicted Rayleigh wave attenuation (Figs. S7a,b). Under hotter conditions, larger differences in Rayleigh wave attenuation are found (Figs. S7c,d) and suggest that our approach likely overestimates TBL thickness under hot conditions.

Temperature is scaled to isotropic shear velocity and attenuation using the parameterization of Jackson and Faul (2010), which is derived from torsional forced-oscillation experiments performed on dry, melt-free forsterite-90 olivine. In order to implement the frequency-dependent shear attenuation and seismic velocity inherent to that parameterization, we take advantage of the fact that Rayleigh waves of different periods sample different depths (Fig. S4) and assign a unique period at each depth according to: $\text{period (s)} = \text{depth (m)} / 1400$ (Forsyth, 1992). We test two values of grain size d (1 mm, 10 mm). At depths > 400 km PREM (Dziewonski and Anderson, 1981) is used. Three separate crustal structures are implemented; they were selected by identifying, for each of the six GTR1 regions, what is the most common crustal type in CRUST1.0 (Laske et al., 2013). All crustal parameters, including density and wave speed, are taken from CRUST1.0, and shear attenuation is fixed at 1/300 throughout the crust (e.g., Durek and Ekström, 1996). For comparison with oceanic regions, the shallow structure consists of 4.77 km of water, 0.50 km of sediment, and 7.00 km of crystalline crust. For comparison with orogenic/magmatic continents (Q) the crystalline crust is 30.0 km thick, and for comparison with the stable continents (P,S) the crystalline crust is 39.0 km thick. Each crustal structure is placed on top of each 1-D mantle seismic model, resulting in 23,760 1-D Earth models.

Figure 6. (a) Misfit for the GTR1 attenuation values as a function of T_P ; here, $d=1$ mm and $dT_a/dz=0.40^{\circ}\text{C}/\text{km}$. (b) Best-fitting intersection depth for the GTR1 attenuation values as a function of T_P . Large circles indicate the T_P value at which misfit is minimized. (c) Comparison of observed (solid) and predicted (dashed) attenuation; for each GTR1 region the best-fitting combination of T_P and intersection depth is used for the predictions. The error-bar length is estimated from the relative difference in the input and output attenuation values determined for each region using the 50-s SPEC-FEM amplitudes (see Fig. 3b and Section 3). (d) Variance reduction for the MMM16 amplitude data set. Variance reduction also includes source and receiver terms obtained from the GTR1 inversion and is expressed relative to the variance reduction provided by only the source and receiver terms. Black: GTR1 attenuation values. Red: TBL models with $d=1$ mm. Blue: TBL models with $d=10$ mm.

5.2 Determining TBL Thickness from Attenuation

For each TBL model, Rayleigh wave attenuation at periods of 40, 50, 66, 100, 133, and 200 s is predicted using Mineos and compared to the observed GTR1 values. Misfit is calculated for each model as the sum over all periods of the absolute value of the difference between observed and predicted attenuation. Fig. 6a summarizes the minimum misfit between observed and predicted attenuation at each T_p value for $d=1$ mm and $dT_a/dz=0.40^\circ\text{C}/\text{km}$. The misfit depends strongly on T_p for the four oceanic and tectonically active regions (A,B,C,Q), with well-defined minima at $T_p=1300^\circ\text{C}$ for regions A and B and $T_p=1250^\circ\text{C}$ for regions C and Q. The minimum misfit for the stable continents (P,S) occurs at $T_p=1200^\circ\text{C}$, although differences in the misfit calculated for $T_p \leq 1250^\circ\text{C}$ are slight. In all cases, the best-fitting intersection depth increases with T_p (Fig. 6b), as expected: a thicker TBL creates overall colder conditions in the uppermost mantle, which is necessary to balance the overall higher temperatures produced when a hotter adiabat is used. If $d=10$ mm instead of 1 mm, the best-fitting T_p values increase by 150-200 $^\circ\text{C}$ relative to their values at $d=1$ mm, but the best-fitting intersection depths are mostly unchanged. If $dT_a/dz=0.25^\circ\text{C}/\text{km}$ or $0.55^\circ\text{C}/\text{km}$ instead of $0.40^\circ\text{C}/\text{km}$, the best-fitting T_p values change by $\leq 50^\circ\text{C}$ and the best-fitting intersection depths change by ≤ 15 km. Figs. 6c,d show that modeling the GTR1 values as simple TBLs yields Rayleigh wave attenuation values that are highly similar to the observed values and can reduce the variance in the MMM16 amplitude data set almost as well as the GTR1 values themselves except for period=200 s, presumably because these deep-sampling, long-period Rayleigh waves are mostly sensing mantle structure that is unrelated to the TBLs.

Fig. 7 summarizes the distribution of geotherms that provide a good fit to attenuation values in the three oceanic and two cratonic regions in GTR1; we exclude tectonically active continents from our analysis because the presence of partial melt in the

mantle, for example in the Basin and Range (Plank and Forsyth, 2016) and Ethiopia (Bastow et al., 2008), would violate the assumption that temperature variations are controlling attenuation. For each region, acceptable geotherms are identified as those whose misfit between observed and predicted attenuation, summed over the six periods, is less than the summed misfit calculated using the error bars in Fig. 6c. The geotherms that fit the attenuation constraints from the oceanic regions agree reasonably well with predictions from the half-space-cooling model, with slightly thicker TBLs than the half-space-cooling model due to the fact that our simplified geotherms overpredict attenuation under hot conditions (Fig. S7) and thus the observed attenuation values will be mapped into slightly colder temperatures. The range of geotherms that satisfy the attenuation constraints for the GTR1 cratonic regions (Fig. 7b) have T_p values between 1100°C and 1300°C and intersection depths of 130-165 km for region P and 165-295 km for region S when $d=1$ mm. With this grain size, the geotherms for regions P and S overlap with the temperature and pressure conditions estimated from mantle xenoliths (Lee et al., 2011) with the exception of those samples from depths > 175 km. When $d=10$ mm, the best-fitting geotherms have similar intersection depths and higher T_p values (1300°C-1450°C) than for $d=1$ mm; with the higher T_p values obtained using the larger grain size even the xenoliths from depths > 175 km fall within the temperature fields estimated for regions P and S. Realistic cratonic geotherms that incorporate radiogenic heat production in the crust and mantle (e.g., Rudnick et al., 1998) agree well with the temperature fields, using $T_p=1300^\circ\text{C}$ and average crustal heat production = $0.65 \mu\text{W}/\text{m}^3$ and $0.75 \mu\text{W}/\text{m}^3$ when $d = 1$ mm.

Figure 7. (a) Distribution of best-fitting geotherms for GTR1 oceanic regions: A (red), B (green), and C (cyan). Dashed black lines show geotherms predicted by the half-space-cooling model with ages of 25 Myr, 85 Myr, and 125 Myr. (b) Distribution of best-fitting geotherms for GTR1 cratonic regions: P (blue) and S (grey). Dashed black lines show steady-state geotherms (Rudnick et al., 1998) calculated with surface heat flow = 41 mW/m² and radiogenic heat production = 0.02 µW/m³ in the lithospheric mantle and = 0.65 µW/m³ (shallower curve) and 0.75 µW/m³ (deeper curve) in the crust. Magenta symbols show constraints from xenolith thermobarometry in the Slave craton, South Africa, and Siberia (Lee et al., 2011). (c) Comparison of observed (black) and predicted phase velocity for GTR1 region S. Predicted values are obtained by converting the conductive geotherm for region S in (b) into seismic velocity with different assumptions. All predictions have a 39-km continental crust and $d=1$ mm. Magenta: isotropic velocity. Red: $V_{SH}-V_{SV}=0.15$ km/s at depths 39-200 km. Grey: As for red, with the addition of a low-velocity zone from 60-80 km. Blue: As for grey, with the addition of a strong high-velocity zone from 200-250 km. Green: As for grey, with the addition of a weak high-velocity zone from 200-350 km.

We next extend this approach to the global attenuation maps, which are expanded in spherical harmonics to degree 16 (Fig. 4). We use the maps derived from the MMM16 amplitude data set at the same six periods that are used for the GTR1 analysis and calculate the misfit between the observed and predicted attenuation values in each 1°x1° grid cell. Fig. 8a shows the best-fitting intersection depth in grid cells for which our approach is most appropriate: those that are characterized by Precambrian continental crust (Fig. 5a) in CRUST1.0 and for which the minimum misfit is < 0.005, a quality criterion determined from the GTR1 analysis described above. For crustal rocks that are Early-Mid Proterozoic and older in age the distribution of TBL thicknesses determined from these attenuation maps is

very similar, with mean intersection depth = 180-190 km. The TBL is considerably thinner for regions containing Late Proterozoic crustal rocks (130 km) as well as all younger continental regions (120 km), although as mentioned earlier our approach is likely not appropriate for areas that may contain melt. From the distributions of best-fitting geotherms in Fig. 7b, we calculate the standard deviation of intersection depths, which is 20 km for region P and 33 km for region S. Thus, 30 km is a reasonable uncertainty on the TBL values in Fig. 8a.

Figure 8. (a) Best-fitting intersection depth in pixels characterized by Precambrian crust and minimum misfit < 0.005. $d = 1$ mm. (b) Distribution of intersection depths for continental crust of different ages. “Active” describes crust younger than Precambrian. (c) Distribution of differences in lithospheric thickness between this study and PM13. (d) Distribution of differences in lithospheric thickness between this study and LITHO1.0.

6. Discussion

Estimating the thickness of the lithosphere from seismological constraints is not straightforward. Since the lithosphere is a layer defined by its rheology, the seismological constraints can serve only as a proxy, and in order to use them it is necessary to make assumptions about how the rheological properties will manifest as seismological properties. Commonly used approaches include the magnitude of and depth variations in absolute shear velocity (e.g., Priestley and McKenzie, 2013), the strength and orientation of seismic anisotropy (e.g., Yuan and Romanowicz, 2010), and receiver functions (e.g., Rychert et al., 2005). While recent results support the idea that receiver functions can detect the lithosphere-asthenosphere boundary (LAB) in continental areas where partial melt at the LAB enhances the impedance contrast (Hansen et al., 2015), LAB detection with receiver functions is less common and more controversial in cratonic settings (e.g., Abt et al., 2010; Foster et al., 2014); scattering interfaces with a negative velocity gradient with depth are typically interpreted as mid-lithospheric discontinuities in those areas. Shear velocities determined from surface waves provide excellent vertical resolution in the uppermost mantle; however, several studies have demonstrated that it is difficult to satisfy observations of phase velocity in cratons given realistic mantle geotherms and a uniform peridotite composition (Bruneton et al., 2004; Lebedev et al., 2009; Pedersen et al., 2009; Hirsch et al., 2015). Here, we have used Rayleigh wave attenuation to estimate lithospheric thickness by assuming that the lithosphere is defined as the thermal boundary layer across which heat transfer occurs by conduction. Since our approach works best in cratonic areas, our estimates of lithospheric thickness (Fig. 8a) are complementary to the LAB constraints in magmatic/orogenic areas that are provided by receiver functions. Furthermore, the use of attenuation instead of velocity avoids the need to make assumptions about mantle

composition and seismic anisotropy. Below we compare our results to other global studies of lithospheric thickness and discuss implications for shear-velocity structure.

6.1 Comparison with Other Studies

We compare our estimates of lithospheric thickness in cratonic areas to two global studies. Priestley and McKenzie (2013; hereinafter PM13) estimated temperature profiles from their global vertically polarized shear velocity (V_{sv}) model in addition to constraints from oceanic thermal models, garnet peridotite nodules, global shear attenuation, and viscosity. They chose the base of the lithosphere based on the depth at which temperature corresponds to a potential temperature of 1315°C. The LITHO1.0 model of Pasyanos et al. (2014) consists of ten layers within which the density and seismic properties are constant with depth; the lithospheric mantle is represented by a single layer. The thickness of the layers and their properties are allowed to vary geographically in order to satisfy constraints provided by Rayleigh and Love wave group- and phase-velocity maps.

Agreement between our results and PM13 is reasonably good (Fig. 8c). On average our estimates of lithospheric thickness are smaller by 20 km than PM13, and the two sets of values differ by < 50 km for 76%, 66%, and 64% of the Late Proterozoic, Early-Mid Proterozoic, and Archean crust, respectively. The histogram of differences contains a secondary peak at 50-60 km for the Early-Mid Proterozoic and Archean crust. The fact that the geographic locations corresponding to this peak are distributed across all continents suggests that shear velocity may systemically underestimate lithospheric thickness in certain areas of very old continental lithosphere. Agreement with LITHO1.0 is weaker, especially for the Early-Mid Proterozoic and Archean crust (Fig. 8d); in those areas our lithosphere is thinner by 70-90 km on average. The fact that we find better agreement with PM13 than LITHO1.0 is not surprising given the nature of the different approaches used.

Although PM13 rely mostly on shear velocity instead of attenuation and use a different criterion to identify the base of the lithosphere, their strategy of trying to map seismic properties into realistic TBL-type geotherms is similar to ours. On the other hand, treating the lithospheric mantle as a single layer of constant velocity and trying to simultaneously fit Rayleigh and Love wave velocities with isotropic shear-wave speed may result in larger uncertainties in the LAB depths in Pasyanos et al. (2014).

6.2 Implications for Shear-Velocity Structure

The temperature profiles that best fit the attenuation constraints in cratonic areas (Fig. 7b) struggle to also provide a satisfactory fit to the phase-velocity observations in these areas. Earlier studies have reached similar conclusions. Shear-velocity profiles lacking a low-velocity zone and characterized by flat or positive velocity gradients with depth have been observed in the central Baltic Shield (Bruneton et al., 2004; Zhu and Tromp, 2013), the East European craton (Zhu and Tromp, 2013), central Australia (Fishwick and Reading, 2008), and the Slave, Yilgarn, and South-Central Finland cratons (Pedersen et al., 2009).

For comparison to the phase-velocity observations, the conductive geotherm corresponding to surface heat flow = 41 mW/m² and crustal heat production = 0.75 μ W/m³ (deeper black dashed curve in Fig. 7b) is used rather than our simplified temperature profiles. Fig. 7c shows the comparison of observed and predicted phase velocity for GTR1 region S for scenarios with isotropy and radial anisotropy ($V_{SH}-V_{SV}=0.15$ km/s, where V_{SH} is the horizontally polarized shear velocity; Gaherty and Jordan, 1995; Gung et al., 2003; Yuan and Romanowicz, 2010) from the Moho to 200 km; the former predicts velocities that are too high at short periods, and the latter predicts velocities that are slightly high at short periods and too low at long periods. We experiment with what adjustments to the mantle velocities are necessary to match the observations. Reducing seismic velocities by 5% in the

depth range 60-80 km solves the problem at short periods. While the broad depth sensitivity of Rayleigh waves permits a range of velocity structures that can produce the low velocities necessary, our tests have shown that the velocity reduction must begin no deeper than 70 km in order to match the 40-s phase velocities. The magnitude and depth extent of the low-velocity layer in the shallow mantle is consistent with observations of a mid-lithospheric discontinuity in receiver functions (e.g., Hansen et al., 2015; Selway et al., 2015). Mechanisms proposed to explain this velocity reduction include the presence of minerals amphibole (Selway et al., 2015) and/or phlogopite (Hansen et al., 2015).

Matching the high velocities at long periods requires elevating V_{SV} above its geotherm-predicted value at depths > 200 km. The non-uniqueness of this high-velocity layer is illustrated in Fig. 7c: a 5% increase distributed over 50 km and a 2% increase distributed over 150 km both provide reasonable predictions of Rayleigh wave phase velocity. Although longer-period and/or higher-mode Rayleigh waves are needed to resolve the maximum depth extent of this feature, the observed phase velocities can more tightly constrain the shallow extent. Putting the top of the high-velocity layer at 150 km or 250 km results in phase-velocity predictions that are too high and too low, respectively, for periods 50-100 s. A velocity increase at ~ 200 km beneath continents was suggested by Lehmann (1959), and the Lehmann discontinuity has been the focus of numerous subsequent studies (e.g., Revenaugh and Jordan, 1991; Gaherty and Jordan, 1995; Gu et al., 2001), many of which favor seismic anisotropy as an explanation. In this study we cannot draw any conclusions about anisotropy, since the Rayleigh wave phase velocities constrain only V_{SV} . For the anisotropic models shown in Fig. 7c, the temperature-predicted shear velocity is assigned to isotropic (Voigt average) shear velocity with the assumption that $V_{SH}-V_{SV}=0.15$ km/s from the Moho to 200-km depth. Fig. 7c makes clear that the slight increase in V_{SV} that results from terminating this radial anisotropy at 200 km is too small to satisfy the

observations (Fig. 7c); furthermore, allowing $V_{SH} > V_{SV}$ to persist to greater depths only exacerbates the mismatch between observations and predictions. While this result would seem to indicate that the required velocity increase is not entirely anisotropic in nature (Vinnik et al., 2005), further analysis that incorporates observations sensitive to V_{SH} is required in order for definite conclusions to be drawn.

7.0 Conclusions

Three independent data sets of Rayleigh wave amplitudes are analyzed to assess the origin of differences between global models of attenuation. A fourth data set of synthetic amplitudes, generated using spectral-element simulations with 3-D elastic and 1-D anelastic Earth models, serves as a benchmark for how focusing and scattering effects can introduce bias into attenuation values and guides our treatment of real amplitude data. The fact that inversions of the synthetic amplitudes can resolve the input structures well (Fig. 2) and produce regionalized attenuation values very different from those found with real data (Fig. 3) serves as strong evidence that dissipative mechanisms, rather than scattering effects, are responsible for observed seismic attenuation, at least in the period band 40-200 s. Furthermore, while the synthetic amplitudes have been used to demonstrate that ERT and FFT better describe focusing effects at short and long periods (Fig. 1), respectively, the analysis of these data has also made clear that the effect on attenuation models of differences between the various focusing treatments is much smaller than the effect of removing versus not removing focusing effects.

The three amplitude data sets are highly consistent with each other, and any differences between the measurement algorithms have a negligible effect on the attenuation structure. Differences in the treatment of focusing effects degrade the level of agreement between the degree-16 maps only slightly. Differences in the path coverage are the single

biggest factor contributing to discrepancies between the maps, and in fact removing focusing effects from the amplitudes helps to counteract the detrimental impact of variable path coverage (Fig. 5b).

The regionalized attenuation and phase-velocity values are anti-correlated with each other (Figs. 3c,d), and the dependence on tectonic region disappears by 200 s for all regions but the stable continents. While the short-period phase velocities are strongly influenced by crustal structure, the short-period attenuation values are not, which is an advantage for constraining mantle attenuation structure. The degree-16 attenuation maps (Fig. 4) achieve a higher resolution than earlier global studies and clearly image very low attenuation associated with specific continental cratons. The frequency-dependent attenuation values at periods < 200 s are well fit by modeling the thermal structure of the cratonic upper mantle as a simple boundary layer, and the field of best-fitting geotherms overlaps with temperature constraints from mantle xenoliths (Figs. 6-7). By equating the boundary layer with the lithosphere in cratonic areas we show that the distribution of lithospheric thicknesses is peaked at 190 km for crust older than the Mid-Proterozoic, with a secondary peak at 250-300 km (Fig. 8). The geotherms that satisfy the attenuation and xenolith constraints, however, cannot match the observed cratonic phase velocities. A 5% velocity reduction at depths 60-80 km can remove the discrepancy at short periods and is consistent with receiver-function observations of a mid-lithospheric discontinuity in cratons. A velocity increase around 200-km depth is needed to satisfy the long-period phase velocities and may be related to the Lehmann discontinuity observed intermittently in continental areas.

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725

726 **Table 1.** Summary of the four data sets of fundamental-mode Rayleigh wave amplitudes.

727 The final column reports the approximate theory and phase-velocity maps with which

728 focusing effects have been pre-calculated and used in this study.

Data Set & Reference	Periods (s)	Number of Paths	Focusing Corrections
DED08 (Dalton et al., 2008)	50, 75, 100, 125	215,345 – 299,138	GCRA; maps of Ekström (2011)
BDR16 (Bao et al., 2016)	50, 100	390,360 – 393,464	ERT, FFT, GCRA; maps of BDR16
MMM16 (Ma et al., 2016)	40, 50, 66, 100, 133, 200	137,548 – 251,030	FFT; maps of MMM16
SPECFEM (Dalton et al., 2014)	50, 75, 125	4,749	ERT, FFT, GCRA; maps of input model

729

730 **Table 2.** Pure-path Rayleigh wave attenuation values ($1/Q$) for GTR1 (Fig. 3c).

	A	B	C	P	Q	S
40 s	0.0085	0.0056	0.0048	0.0022	0.0058	0.0011
50 s	0.0097	0.0073	0.0062	0.0031	0.0075	0.0018
66 s	0.0113	0.0090	0.0077	0.0044	0.0088	0.0030
100 s	0.0112	0.0100	0.0085	0.0057	0.0092	0.0049
133 s	0.0099	0.0090	0.0085	0.0058	0.0081	0.0052
200 s	0.0064	0.0067	0.0062	0.0046	0.0059	0.0045

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References

- Abers, G.A., K.M. Fischer, G. Hirth, D.A. Wiens, T. Plank, B.K. Holtzman, C. McCarthy, and E. Gazel, Reconciling mantle attenuation-temperature relationships from seismology, petrology, and laboratory measurements, *Geochem. Geophys. Geosyst.*, 15, doi:10.1002/2014GC005444, 2014.
- Abt, D.L., K.M. Fischer, S.W. French, H.A. Ford, H. Yuan, and B. Romanowicz, North American lithospheric discontinuity structure imaged by Ps and Sp receiver functions, *J. Geophys. Res.*, 115, 1978-2012, 2010.
- Artemieva, I.M., The continental lithosphere: Reconciling thermal, seismic, and petrologic data, *Lithos*, 109, 23-46, 2009.
- Bao, X., C.A. Dalton, and J. Ritsema, Effects of elastic focusing on global models of Rayleigh wave attenuation, *Geophys. J. Int.*, doi:10.1093/gji/ggw332, 2016.
- Bastow, I.D., A.A. Nyblade, G.W. Stuart, T.O. Rooney, and M.H. Benoit, Upper mantle seismic structure beneath the Ethiopian hot spot: Rifting at the edge of the African low-velocity anomaly, *Geochem. Geophys. Geosys.*, 9, doi:10.1029/2008GC002107, 2008.
- Bruneton, M., H. A. Pedersen, P. Vacher, I. T. Kukkonen, N. T. Arndt, S. Funke, W. Friederich, and V. Farra, Layered lithospheric mantle in the central Baltic shield from surface waves and xenolith analysis, *Earth Planet. Sci. Lett.*, 226, 41–52, doi:10.1016/j.epsl.2004.07.034, 2004.
- Dalton, C.A., and G. Ekström, Global models of surface-wave attenuation, *J. Geophys. Res.*, 111, B05317, doi:10.1029/2005JB003997, 2006.
- Dalton, C.A., G. Ekström, and A.M. Dziewonski, The global attenuation structure of the upper mantle, *J. Geophys. Res.*, 113, B09303, doi:10.1029/2007JB005429, 2008.

758 Dalton, C.A., G. Ekström, and A.M. Dziewonski, Global seismological shear velocity and
 759 attenuation: A comparison with experimental observations, *Earth Planet. Sci. Lett.*,
 760 284, 65-75, doi:10.1016/j.epsl.2009.04.009, 2009.

761 Dalton, C.A., V. Hjörleifsdóttir, and G. Ekström, A comparison of approaches to the prediction
 762 of surface-wave amplitude, *Geophys. J. Int.*, doi:10.1093/gji/ggt365, 2014.

763 Durek, J.J., and G. Ekström, A radial model of anelasticity consistent with long-period surface
 764 wave attenuation, *Bull. Seism. Soc. Am.*, 86, 144-158, 1996.

765 Dziewonski, A.M., and D.L. Anderson, Preliminary reference Earth model, *Phys. Earth Planet.*
 766 *Inter.*, 25, 297-356, 1981.

767 Dziewonski, A.M., T.A. Chou, and J.H. Woodhouse, Determination of earthquake source
 768 parameters from waveform data for studies of global and regional seismicity. *J.*
 769 *Geophys. Res.*, 86, doi: 10.1029/JB086iB04p02825, 2825-2852, 1981.

770 Ekström, G., A global model of Love and Rayleigh surface wave dispersion and anisotropy,
 771 25-250 s, *Geophys. J. Int.*, 187, 1668-1686, doi:10.1111/j.1365-246X.2011.05225.x,
 772 2011.

773 Ekström, G., J. Tromp, and E.W.F. Larson, Measurements and global models of surface wave
 774 propagation, *J. Geophys. Res.*, 102, 8137-8157, 1997.

775 Ekström, G., C.A. Dalton, and M. Nettles, Observations of time-dependent errors in the long-
 776 period instrument gain at global seismic stations, *Seismol. Res. Lett.*, 77, 12-22, 2006.

777 Ekström, G., M. Nettles, and A.M. Dziewonski, The global CMT project 2004-2010: Centroid-
 778 moment tensors for 13,017 earthquakes, *Phys. Earth Planet. Inter.*, 200-201, 1-9,
 779 doi:10.1016/j.pepi.2012.04.002, 2012.

780 Ferreira, A.M., and J.H. Woodhouse, Source, path and receiver effects on seismic surface
 781 waves, *Geophys. J. Int.*, 168, 109-132, doi:10.1111/j.1365-246X.2006.03092.x, 2006.

782 Fishwick, S., and A. M. Reading, Anomalous lithosphere beneath the Proterozoic of western

783 and central Australia: A record of continental collision and intraplate deformation?,
784 *Precambrian Res.*, 166, 111–121, doi:10.1016/j.precamres.2007.04.026, 2008.

785 Forsyth, D.W., Geophysical constraints on mantle flow and melt generation beneath mid-
786 ocean ridges. Mantle flow and melt generation at mid-ocean ridges. *AGU Geophysical*
787 *Monograph*. 71, 1–65, 1992.

788 Foster, K., K. Dueker, B. Schmandt, and H. Yuan, A sharp cratonic lithosphere-asthenosphere
789 boundary beneath the American Midwest and its relation to mantle flow, *Earth*
790 *Planet. Sci. Lett.*, 402, 82-89, 2014.

791 Gaherty, J.B., and T.H. Jordan, Lehmann discontinuity as the base of an anisotropic layer
792 beneath continents, *Science*, 268, 1468-1471, 1995.

793 Goes, S., and S. van der Lee, Thermal structure of the North American uppermost mantle
794 inferred from seismic tomography, *J. Geophys. Res.*, 107,
795 doi:10.1029/2000JB000049, 2002.

796 Goes, S., R. Govers, and P. Vacher, Shallow mantle temperature under Europe from P and S
797 wave tomography, *J. Geophys. Res.*, 105, 11,153–11,169, doi:10.1029/1999JP900300,
798 2000.

799 Gu, Y.J., A.M. Dziewonski, and G. Ekström, Preferential detection of the Lehmann
800 discontinuity beneath continents, *Geophys. Res. Lett.*, 28, 4655-4658, 2001.

801 Gung, Y.C., M. Panning, and B. Romanowicz, Anisotropy and lithospheric thickness, *Nature*,
802 422, 707-711, 2003.

803 Hansen, S.M., K. Dueker, and B. Schandt, Thermal classification of lithospheric
804 discontinuities beneath USArray, *Earth Planet. Sci. Lett.*, 431, 36-47, 2015.

805 Hirsch, A.C., C.A. Dalton, and J. Ritsema, Constraints on shear velocity in the cratonic upper
806 mantle from Rayleigh wave phase velocity, *Geochem. Geophys. Geosys.*, 16,
807 doi:10.1002/2015GC006066, 2015.

808 Jackson, I., and U.H. Faul, Grainsize-sensitive viscoelastic relaxation in olivine: Towards a
809 robust laboratory-based model for seismological application, *Phys. Earth. Planet.*
810 *Inter.*, 183, 151-163, 2010.

811 Jackson, I., U.H. Faul, J.D. Fitz Gerald, and B.H. Tan, Shear wave attenuation and dispersion in
812 melt-bearing olivine polycrystals: 1. Specimen fabrication and mechanical testing, *J.*
813 *Geophys. Res.*, 109, doi:10.1029/2003JB002406, 2004. .

814 Jordan, T.H., Global tectonic regionalization for seismological data analysis, *Bull. Seismol.*
815 *Soc. Am.*, 71, 1131-1141, 1981.

816 Karato, S., Mapping water content in the upper mantle, *Inside the Subduction Factory*, AGU
817 *Geophys. Monogr.*, 138, 135-152, 2003.

818 Komatitsch, D., and J. Tromp, Spectral-element simulations of global seismic wave
819 propagation-I. Validation, *Geophys. J. Int.*, 149, 390-412, doi: 10.1046/j.1365-
820 246X.2002.01653.x, 2002a.

821 Komatitsch, D., and J. Tromp, Spectral-element simulations of global seismic wave
822 propagation-II. 3-D models, oceans, rotation, and self-gravitation, *Geophys. J. Int.*,
823 150, 303-318, doi: 10.1046/j.1365-246X.2002.01716.x, 2002b.

824 Larson, E.W.F., J. Tromp, and G. Ekström, Effects of slight anisotropy on surface waves,
825 *Geophys. J. Int.*, 132, 654-666, 1998.

826 Laske, G., G. Masters, Z. Ma, and M. Pasyanos, Update on CRUST1.0—A 1-degree global
827 model of Earth's crust, *Geophys. Res. Abstr.*, 15, Abstract EGU2013-2658, 2013.

828 Lebedev, S., J. Boonen, and J. Trampert, Seismic structure of Precambrian lithosphere: New
829 constraints from broad-band surface-wave dispersion, *Lithos*, 109, 91-111, 2009.

830 Lee, C.-T. A., P. Luffi, and E. J. Chin, Building and destroying continental mantle, *Annu. Rev.*
831 *Earth Planet. Sci. Lett.*, 39, 59-90, doi:10.1146/annurev-earth-040610-133505,
832 2011.

833 Lehmann, I., Velocities of longitudinal waves in the upper part of the Earth's mantle, *Ann.*
 834 *Geophys.*, 15, 93-118, 1959.

835 Ma, Z., G. Masters, G. Laske, and M.E. Pasyanos, A comprehensive dispersion model of
 836 surface wave phase and group velocity for the globe, *Geophys. J. Int.*, 199,
 837 doi:10.1093/gji/ggu246, 2014.

838 Ma, Z., G. Masters, and N. Mancinelli, Two-dimensional global Rayleigh wave attenuation
 839 model by accounting for finite-frequency focusing and defocusing effect, *Geophys. J.*
 840 *Int.*, 204, 631-649, 2016.

841 McCarthy, C., and Y. Takei, Anelasticity and viscosity of partially molten rock analogue:
 842 toward seismic detection of small quantities of melt, *Geophys. Res. Lett.*, 38,
 843 10.1029/2011GL048776, 2011.

844 McCarthy, C., Y. Takei, and T. Hiraga, Experimental study of attenuation and dispersion over
 845 a broad frequency range: 2. The universal scaling of polycrystalline materials, *J.*
 846 *Geophys. Res.*, 116, doi:10.1029/2011JB008384, 2011.

847 Meschede, M., and B. Romanowicz, Non-stationary spherical random media and their effect
 848 on long-period mantle waves, *Geophys. J. Int.*, 203, 1605-1625, 2015.

849 Müller, R.D., M. Sdrolias, C. Gaina, and W.R. Roest, Age, spreading rates and spreading
 850 symmetry of the world's ocean crust, *Geochem. Geophys. Geosyst.*, 9, Q04006,
 851 doi:10.1029/2007GC001743, 2008.

852 Pasyanos, M.E., T.G. Masters, G. Laske, and Z. Ma, LITHO1.0: An updated crust and
 853 lithospheric model of the Earth, *J. Geophys. Res.*, 119, 2153-2173,
 854 doi:10.1002/2013JB010626, 2014.

855 Pedersen, H.A., S. Fishwick, and D.B. Snyder, A comparison of cratonic roots through
 856 consistent analysis of seismic surface waves, *Lithos*, 109, 81-95,
 857 doi:10.1016/j.lithos.2008.09.016, 2009.

858 Plank, T., and D.W. Forsyth, Thermal structure and melting conditions in the mantle beneath
859 the Basin and Range province from seismology and petrology, *Geochem. Geophys.*
860 *Geosyst.*, 17, 1312-1338, doi:10.1002/2015GC006205, 2016.

861 Priestley, K., and D. McKenzie, The relationship between shear wave velocity, temperature,
862 attenuation, and viscosity in the shallow part of the mantle, *Earth Planet. Sci. Lett.*,
863 381, 78-91, 2013.

864 Revenaugh, J., and T.H. Jordan, Mantle layering from ScS reverberations: 3. The upper
865 mantle, *J. Geophys. Res.*, 96, 19781-19810, 1991.

866 Ricard, Y., S. Durand, J.-P. Montagner, and F. Chambat, Is there seismic attenuation in the
867 mantle?, *Earth Planet. Sci. Lett.*, 388, 257-264, 2014.

868 Ritsema, J., A. Deuss, H.J. van Heijst, and J.H. Woodhouse, S40RTS: A degree-40 shear-
869 velocity model for the mantle from new Rayleigh wave dispersion, teleseismic travel
870 time and normal-mode splitting function measurements, *Geophys. J. Int.*, 184, 1223-
871 1236, doi:10.1111/j.1365-246X.2010.04884.x, 2011.

872 Romanowicz, B., A global tomographic model of shear attenuation in the upper mantle, *J.*
873 *Geophys. Res.*, 100, 12375-12394, 1995.

874 Romanowicz, B., and Y.C. Gung, Superplumes from the core-mantle boundary to the
875 lithosphere: implications for heat-flux, *Science*, 296, 513-516, 2002.

876 Rudnick, R. L., W. F. McDonough, and R. J. O'Connell, Thermal structure, thickness, and
877 composition of continental lithosphere, *Chem. Geol.*, 145, 395-411, 1998.

878 Rychert, C.A., K.M. Fischer, and S. Rondenay, A sharp lithosphere-asthenosphere boundary
879 imaged beneath eastern North America, *Nature*, 436, 542-545, 2005.

880 Rychert, C.A., K.M. Fischer, G.A. Abers, T. Plank, E. Syracuse, J.M. Protti, V. Gonzalez, and W.
881 Strauch, Strong along-arc variations in attenuation in the mantle wedge beneath

882 Costa Rica and Nicaragua, *Geochem. Geophys. Geosyst.*, 9, Q10S10,
 883 doi:10.1029/2008GC002040, 2008.

884 Selby, N.D., and J.H. Woodhouse, Controls on Rayleigh wave amplitudes: attenuation and
 885 focusing, *Geophys. J. Int.*, 142, 933-940, 2000.

886 Selway, K., H. Ford, and P. Kelemen, The seismic mid-lithospheric discontinuity, *Earth*
 887 *Planet. Sci. Lett.*, 414, 45-57, 2015.

888 van Heijst, H.J., and J.H. Woodhouse, Measuring surface-wave overtone phase velocities
 889 using a mode branch stripping technique, *Geophys. J. Int.*, 131, 209-230, 1997.

890 Vinnik, L., E. Kurnik, and V. Farra, Lehmann discontinuity beneath North America: No role
 891 for seismic anisotropy, *Geophys. Res. Lett.*, 32, doi:10.1029/2004GL022333, 2005.

892 Wang, Z., and F.A. Dahlen, JWKB surface-wave seismograms on a laterally heterogeneous
 893 earth, *Geophys. J. Int.*, 119, 381-401, 1994.

894 Wang, Z., and F.A. Dahlen, Validity of surface-wave ray theory on a laterally heterogeneous
 895 earth, *Geophys. J. Int.*, 123, 10.1111/j.1365-246X.1995.tb06888.x, 757-773, 1995.

896 Woodhouse, J.H., and Y.K. Wong, Amplitude, phase, and path anomalies of mantle waves,
 897 *Geophys. J. R. Astron. Soc.*, 87, 753-773, 1986.

898 Yang, Y., and D.W. Forsyth, Attenuation in the upper mantle beneath Southern California:
 899 Physical state of the lithosphere and asthenosphere, *J. Geophys. Res.*, 113, B033308,
 900 doi:10.1029/2007JB005118, 2008.

901 Yang, Y., D.W. Forsyth, and D.S. Weeraratne, Seismic attenuation near the East Pacific Rise
 902 and the origin of the low-velocity zone, *Earth Planet. Sci. Lett.*, 258, 260-268,
 903 doi:10.1016/j.epsl.2007.03.040, 2007.

904 Yuan, H., and B. Romanowicz, Lithospheric layering in the North American craton, *Nature*,
 905 466, doi:10.1038/nature09332, 1063-1069, 2010.

906 Zhou, Y., F.A. Dahlen, and G. Nolet, Three-dimensional sensitivity kernels for surface wave
907 observables, *Geophys. J. Int.*, 158, 142-168, doi:10.1111/j.1365-246X.2004.02324.x,
908 2004.

909 Zhu, H., and J. Tromp, Mapping tectonic deformation in the crust and upper mantle beneath
910 Europe and the north Atlantic ocean, *Science*, 341, 871-875,
911 doi:10.1126/science.1241335, 2013.

912 Zhu, H., E. Bozdag, T.S. Duffy, and J. Tromp, Seismic attenuation beneath Europe and the
913 North Atlantic: Implications for water in the mantle, *Earth Planet. Sci. Lett.*, 381, 1-
914 11, 2013.

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