Q_{M1D}: A 1D model of shear attenuation in the mantle from differential body waves

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Abstract

We introduce a new 1D model of shear attenuation in the mantle (Q_{M1D}) constructed from measurements of differential attenuation of S and ScS body waves. We build a dataset of over 90,000 measurements of $\delta t^*_{\text{ScS-S}}$ from seismograms with high-quality S and ScS waveforms from deep earthquakes between 20–75° epicentral distance from 1990–2024. To ensure that measurements are robust, we require that values of differential attenuation are similar between the instantaneous frequency matching and waveform matching methods. Using this dataset, we perform a Bayesian inversion to construct 1D profiles of mantle shear attenuation Q_μ and its uncertainties. Our dataset has a high sensitivity to the mid-mantle (750–2200 km) corresponding to the turning depths of the S phases, which maps into lower uncertainty in the ensemble distribution at these depths. Q_{M1D} suggests that the mantle is more attenuating than PREM on average; but, notably, that there is a high attenuating layer around 750 km depth and that there are low attenuating layers around 900 km and 2000 km depth. Due to the source-receiver distribution for the phases, our dataset has a sampling bias in the mantle around the edge of the Pacific. These weakly attenuating zones in the mantle may correspond to a viscosity jump around 900 km depth and a further viscosity jump in the deep mantle around 2000 km depth, whilst the thin strongly attenuating layer may

correspond to regions of slab stagnation caused by a potential thin layer of low viscosity beneath the mantle transition zone.

1 Introduction

Over the past decades, elastic tomography models of the mantle have been numerous and are becoming increasingly detailed [e.g. Dziewonski and Anderson, 1984; Ritsema et al., 2011; Cui et al., 2024]. However, it is equally important to understand the anelastic structure since it results in variations in the amplitude of seismic phases [e.g. Romanowicz, 1995; Dalton and Ekström, 2006; Zaroli et al., 2010; Hansen et al., 2021] – which may otherwise be mapped into elastic structure [e.g. Karato, 1993; Chaves and Ritsema, 2016] – but also because the anelastic structure is thought to be much more sensitive to temperature and grain size variations in the mantle than elastic structure is [e.g. Jackson et al., 1992, 2002; Matas and Bukowinski, 2007; Faul and Jackson, 2015]. Knowledge of both elastic and anelastic structure will therefore result in important additional constraints on mantle dynamics, but also in distinguishing between the chemical and thermal composition of the mantle [e.g. Romanowicz, 1995; Cooper, 2002; Gung and Romanowicz, 2004].

Whilst the large-scale anelastic structure of upper mantle, i.e. attenuation, is relatively well understood from body and surface waves [e.g. Romanowicz, 1995; Bhattacharyya et al., 1996; Reid et al., 2001; Romanowicz and Gung, 2002; Gung and Romanowicz, 2004; Dalton et al., 2008; Adenis et al., 2017; Karaoğlu and Romanowicz, 2018] – where an anti-correlation is found between 3D perturbations in velocity and attenuation – there have been significantly fewer studies on attenuation in the lower mantle. The large-scale attenuation structure of the lower mantle is predominantly constrained by normal mode observations [e.g. Dziewonski and Anderson, 1984; Moulik and Ekström, 2025] and is typically assumed to be less attenuating than the upper mantle [e.g. Warren and Shearer, 2002; Souriau et al., 2012; Nguyen et al., 2025]. There have been attempts to distinguish relative attenuation within and outside the two antipodal Large Low Shear Velocity Provinces (LLSVPs) using body waves [Romanowicz, 1995; Hwang and Ritsema, 2011; Liu and Grand, 2018] to provide much needed constraints on the grain size, temperature, and longevity of these structures. A recent study by Talavera-Soza et al. [2025] used anelastic normal mode splitting function observations to construct a 3D model of attenuation of the mantle. Their model, QS4L3, is in good agreement with attenuation observations from body waves and surface waves in the upper mantle, but appears to differ from the few studies of the lowermost mantle which use body waves. In particular, normal modes find LLSVPs to be weakly attenuating – suggesting large grain sizes and high temperatures, which potentially implies a long-lived structure [Talavera-Soza et al., 2025] – whilst body waves studies have suggested the opposite [Romanowicz, 1998; Lawrence and Wysession, 2006; Hwang and Ritsema, 2011; Liu and Grand, 2018]. It is important to note that the datasets used in each of the studies of the lower mantle are typically small and that the frequency content differs by a few orders of magnitude between body waves and normal modes. Given there is significant variability and scatter in measurements of attenuation [e.g. Bhattacharyya, 1998], the nature of anelasticity in the lowermost mantle remains unclear. In this study, we therefore aim to elucidate the anelastic structure of the lower mantle.

Attenuation t^* (with units of seconds) of a body wave phase is defined as the shear quality factor Q_μ (dimensionless) integrated along the raypath

$$t^* = \int \frac{\mathrm{d}s}{\beta\, Q_\mu} = \int_s \frac{\mathrm{d}t}{Q_\mu} \tag{1}$$

where s defines the raypath and β is the shear velocity. This is also sometimes related to the path averaged quality factor Q_{phase} as $t^* = T_{\text{phase}}/Q_{\text{phase}}$ for the total travel time T_{phase} of a given body wave phase. It is not possible to directly measure t^*, but it is possible to measure the relative attenuation between a target phase with t^*_{TARG} and a reference phase with t^*_{REF} as

$$\delta t^*_{\text{REF-TARG}} = t^*_{\text{REF}} - t^*_{\text{TARG}} = \int_{\text{REF}} \frac{\mathrm{d}t}{Q_\mu} - \int_{\text{TARG}} \frac{\mathrm{d}t}{Q_\mu} \tag{2}$$

There have been many methods used over the past decades to make measurements of relative attenuation of body waves. These can be categorised as frequency domain or time domain methods, but the principle of both methods is the same: when a signal is attenuated, the higher frequency content is more attenuated than the lower frequency content resulting in a broadened signal [e.g. Teng, 1968; Quan and Harris, 1997]. In the frequency domain, the spectral ratio method is by far the most commonly used method [e.g. Sheehan and Solomon, 1992; Bhattacharyya et al., 1996; Roth et al., 1999; Fisher et al., 2003; Lawrence and Wysession, 2006; Hwang and Ritsema, 2011; Suetsugu et al., 2019; Soto Castaneda et al., 2021; Nguyen et al., 2025], which relates the spectral amplitude ratio to the relative attenuation between two signals. A more sophisticated frequency domain method – the instantaneous frequency matching method (IFM) – applies the causal attenuation operator to the target signal until its instantaneous frequency matches that of the reference signal [e.g. Matheney and Nowack, 1995; Quan and Harris, 1997; Ford et al., 2012; Durand et al., 2013; Sun et al., 2025]. In the time domain, the waveform matching (WM) method finds the optimal relative attenuation that matches the waveforms between the target signal which has been attenuated by the causal attenuation operator and the reference signal [e.g. Chan and Der, 1988; Bhattacharyya, 1998; Reid et al., 2001; Oki and Shearer, 2008; Bezada et al., 2023]. This is also sometimes done in a joint inversion for the amplitude ratio, time delay and source parameters [e.g. Reid et al., 2001; Garcia et al., 2013; Pejić et al., 2017]. Additionally, implicit inversion of Q_μ structure has also been performed by waveform modelling of amplitude ratios [e.g. Revenaugh and Jordan, 1991; Chaves and Ritsema, 2016; Konishi et al., 2017, 2020; Borgeaud and Deschamps, 2021], but this has significant trade-offs with elastic structure. Each of these methods has its own advantages and disadvantages – and no one method works in every situation [e.g. Bhattacharyya, 1998; Ford et al., 2012] – so the most appropriate method(s) to use depends on the context of the investigation. It is important to note that all of the methods described above result in inaccurate estimates of attenuation when there are interfering phases in the target or reference seismic waveforms.

The choice of which reference and target signals to use can also be broadly categorised into three different types:

1. The first is to estimate the relative attenuation of the target and reference signals for two different seismic phases recorded on the same station for the same event; e.g. ScSScS-ScS [Revenaugh and Jordan, 1991], sS-S and sScS-ScS [Flanagan and Wiens, 1994], and ScS-S [Durand et al., 2013]. Any differential attenuation is then attributed to the difference between the two raypaths travelled by the two phases, since the source and receiver contributions to attenuation are assumed to be similar for both raypaths.

2. The second is to estimate the attenuation of the same seismic phase for the same event recorded on different stations [e.g. Warren and Shearer, 2002; Hwang et al., 2009]. In this case, the 'source term' – a contribution to the measurement from source effects – is assumed

to be negligible since the raypaths travelled by a phase are similar near the source. However, the ‘station term’ – the contribution to the measurement due to near-station effects – must be accounted for. One important consideration in using this method is that rupture directivity modifies the source frequency spectra as a function of the take-off angle [e.g. Bernard et al., 1996], so any inferred attenuation difference between pairwise stations includes both differential attenuation structure and source frequency spectra variations [e.g. Warren and Shearer, 2006]. To overcome this, station pairs can be chosen so that they are azimuthally similar [e.g. Hwang and Ritsema, 2011].

3. The third is to estimate the ‘absolute’ attenuation of an observed phase relative to synthetic waveforms, in which case both source and station effects must be accounted for in the synthetic modelling in order to get accurate results [e.g. Zhu et al., 2022; Sun et al., 2025]. Equivalently, by assuming a source model and the station response [e.g. Brune, 1970; Scherbaum, 1990], it is possible to estimate the ‘absolute’ attenuation of a signal by directly fitting its frequency spectra [e.g. Rietbrock, 2001; Eberhart-Phillips and Chadwick, 2002].

Source and station terms can be inverted for – as well as Q_μ, but at additional computational expense – assuming that these terms are consistent for a given source or station, and that individual sources and stations are represented in the dataset multiple times [e.g. Roth et al., 1999; Warren and Shearer, 2002; Hwang et al., 2009]. Historically, ScS reverberations have been used to make whole-mantle estimates of average Q_{ScS} in corridors between pairs of sources and receivers [e.g. Chan and Der, 1988; Revenaugh and Jordan, 1991; Isse and Nakanishi, 1997; Gomer and Okal, 2003; Kanamori and Rivera, 2015; Chaves and Ritsema, 2016], which minimises the effect of source and station terms. However, this choice of phases involves unpaired raypaths travelling through the upper mantle and crust, so accurate upper mantle and crustal models are required to correct for these effects or the results could be significantly biased [e.g. Revenaugh and Jordan, 1991; Isse and Nakanishi, 1997]. Additionally, S reflections off the underside of Earth’s surface have highly complex mini-max Fresnel zones and so poorly match ray theoretical approximations [e.g. Shearer, 1993; Schultz and Gu, 2013]. It is thus difficult to make corrections for unpaired raypaths through the crust. This is also the case when comparing synthetics to data to estimate attenuation [e.g. Sun et al., 2025].

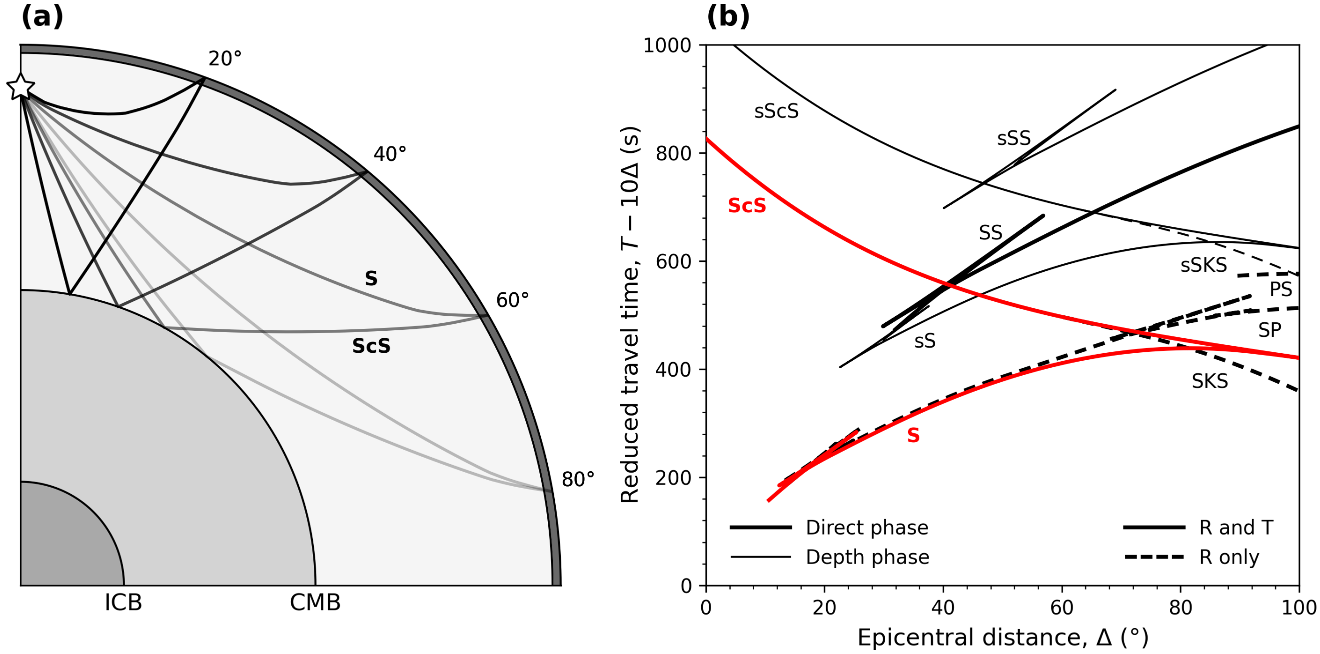

Figure 1: **(a)** S and ScS raypaths for a range of epicentral distances. The turning depths of S and differential incident and take-off angles of S and ScS are shown in Figure S1. **(b)** The reduced travel times of seismic phases observed around S and ScS (highlighted in red). The visibility of phases on the transverse (T) and radial (R) components for an isotropic model is indicated as R and T (solid line) or R only (dashed line).

Ford et al. [2012] and Durand et al. [2013] used S and ScS – which minimises the effects of source and station terms, and has no unpaired raypaths travelling through the crust – to estimate the shear attenuation profile of the mantle. S turns in the mid-mantle, whilst ScS is reflected off the core-mantle boundary (CMB, Figure 1a). At long epicentral distances, the raypaths of S and ScS, and the difference in take-off and incident angles between S and ScS, are very similar (Figures 1a & S1), but the travel time differential becomes too small to distinguish between them on seismograms (Figure 1b). At short epicentral distances, the difference in take-off and incident angles becomes large, and so the assumption that the near-source terms cancel out becomes strained [e.g. Ford et al., 2012]. However, some have suggested that this effect is negligible for frequency domain methods, even for phases with opposite take-off angles [e.g. Flanagan and Wiens, 1994]. The differential attenuation of ScS and S is sensitive to the whole depth range of the mantle, but significantly less sensitive in the uppermost mantle and crust where the raypaths are most similar. Assuming a radially isotropic Earth, the relative attenuation of ScS and S would therefore predominantly depend on the turning depth of the S raypath (Figures 1a & S1a-b).

In this study, we compile a dataset of over 90,000 high-quality S and ScS waveforms and estimate their relative attenuation $\delta t^*_{\text{ScS-S}}$ to make inferences about 1D mantle shear attenuation. We evaluate three commonly used methods to estimate differential attenuation (spectral ratio, instantaneous frequency matching, and waveform matching). After weighing the advantages and disadvantages of each of the methods, we disregard the spectral ratio method, and analyse and compare our dataset of S and ScS waveforms with the instantaneous frequency matching (IFM) and waveform matching (WM) methods. Finally, we perform a Bayesian inversion to fit a 1D whole-mantle model of Q_μ to our dataset, which also includes estimates of uncertainty in the profile.

2 Methods

2.1 Data

We inspected seismograms of earthquakes larger than magnitude 6.0 and deeper than 75 km depth between 1990–2024 from the EarthScope SAGE event catalogue. We only downloaded seismograms for stations between epicentral distances 20–80°, since S is not distinguishable from surface waves at shorter distances and ScS at longer distances. We ensure that S phases at short epicentral distances do not appear to be contaminated by triplication, which may result in phase broadening (discussed in Section 4.1). 75 km is chosen as the event depth threshold since the arrival time difference between S and ScS and their respective depth phases (sS and sScS) is approximately twice the dominant period of S ($\sim$14 s). The instrument response is removed and the traces output as velocity, before being downsampled to 10 Hz and rotated to the ZRT orientation.

To avoid issues of seismic phase interference in our measurements, we disregard seismograms for which the arrival time of ScS is within 25 s of S, sS, SS or SSS as calculated for the Preliminary Reference Earth Model (PREM) [Dziewonski and Anderson, 1984]. We then compute the signal-to-noise (SNR) of the seismic phases, defined as the ratio between the maximum amplitude in a window of ±15 s around the predicted S or ScS arrival and the root mean square of the window 120–20 s before the predicted S arrival. We then discard any seismograms for which the SNR of S or ScS are less than 5, or if the maximum amplitude is at the edge of the 30 s window around either phase.

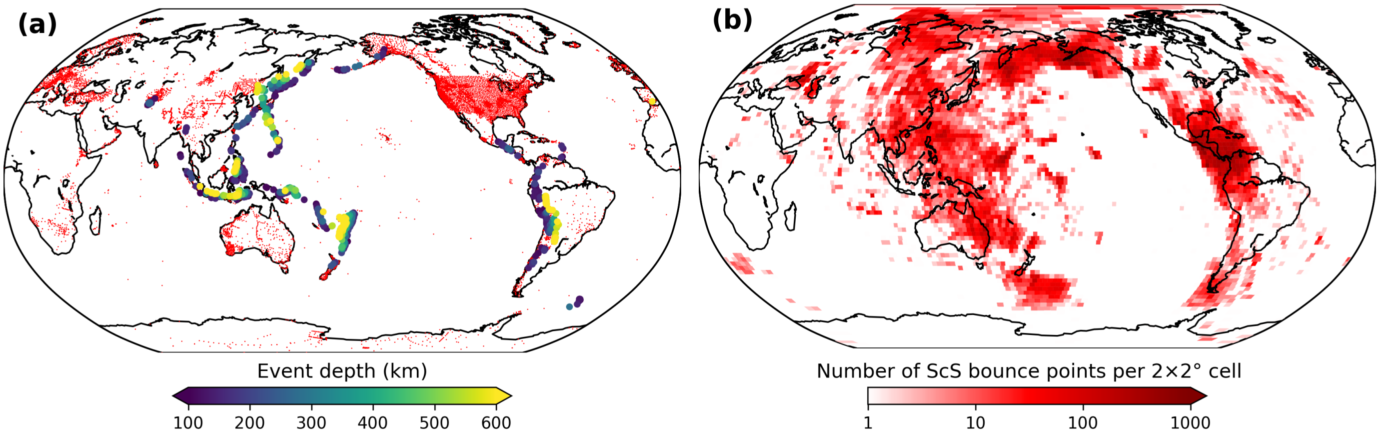

Figure 2: **(a)** Locations of earthquakes (circles, coloured by event depth) and stations (red dots) used in this study. **(b)** Number of ScS bounce points per $2 \times 2°$ quasi-equal area cell.

The sources and receivers represented in our final dataset are shown in Figure 2a. For a quasi-equal area grid of the core-mantle boundary (CMB), we plot the number of ScS bounce points per $2 \times 2°$ cell in Figure 2b, which shows that coverage of the mantle of our dataset is strongly biased by the source-receiver distribution around the edge of the Pacific.

2.2 Estimating differential attenuation of body waves

2.2.1 Spectral ratio method

The spectral ratio method of estimating relative attenuation has been known for over 60 years [e.g. Asada and Takano, 1963; Kovach and Anderson, 1964], in which the gradient of the spectral ratio between two signals can be related to the relative attenuation between the signals assuming frequency independence of Q. This method suffers from significant technical challenges pertaining to the windowing and tapering of signals, spectral gaps, and the choice of frequency range over which to estimate the gradient of the spectral ratio [e.g. Bhattacharyya, 1998; Ford et al., 2012; Tanaka and Tkalčić, 2015]. Whilst these shortfalls have been resolved to some extent – e.g. using multi-tapering of seismic phases to reduce spectral leakage [Thomson, 1982; Lees and Park, 1995] – we find that the calculated δt^* values depend strongly on the setup parameters and SNR of the data. Notably, there was often very low similarity between the spectral ratio and its best linear fit, as previously observed by Isse and Nakanishi [1997]. To improve the quality of fit, some studies have also stacked data within a vertical column from ScS multiples and surface reflections to increase the

SNR of the waveforms [e.g. Gomer and Okal, 2003; Suetsugu et al., 2019], though this assumption is only reasonable for short epicentral distances and not practical for a broader global study. Others have performed joint fitting of spectral phase and amplitude ratios [e.g. Eilon and Abers, 2017].

2.2.2 Instantaneous frequency matching method

As attenuation preferentially diminishes higher frequency spectra, the average and peak frequencies will appear lower for more attenuated waves [e.g. Matheney and Nowack, 1995; Quan and Harris, 1997]. Matheney and Nowack [1995] derived that the differential attenuation between two signals is equivalent to the value t^*_{CAO} of the causal attenuation operator that matches the instantaneous frequency of the target signal f^*_{TARG} to that of the reference signal f_{REF} (Figure 3a-e). The causal attenuation operator $D(\omega)$ takes the form

$$D(\omega) = \exp\left[\frac{-\omega t^*_{\mathrm{CAO}}}{2}\left(1 - \frac{2i}{\pi}\ln\frac{\omega}{\omega_0}\right)\right] \tag{3}$$

where t^*_{CAO} is the relative attenuation factor between two phases, ω is angular frequency, and ω_0 is a reference angular frequency (usually taken to be 1 Hz) [e.g. Chapman et al., 1988; Aki and Richards, 2002]. Since $t^*_{\mathrm{CAO}} \geq 0$ s, whichever signal has the lowest instantaneous frequency is taken as the reference and the other as the target signal (Figure 3c-e).

The instantaneous amplitude (envelope) $a(t)$ and instantaneous phase $\phi(t)$ of a signal $y(t)$ represented as a complex function are given as

$$y(t) = a(t)\exp\left[i\phi(t)\right], \qquad a(t) = \sqrt{y(t)^2 + y^*(t)^2}, \qquad \phi(t) = \tan^{-1}\left[\frac{y^*(t)}{y(t)}\right] \tag{4}$$

where $y^*(t)$ is the Hilbert transform of a signal $y(t)$, so the instantaneous frequency f_{I} is given by

$$f_{\mathrm{I}}(t) = \frac{1}{2\pi}\frac{\mathrm{d}\phi(t)}{\mathrm{d}t} = \frac{1}{2\pi}\frac{y(t)\frac{\mathrm{d}y^*(t)}{\mathrm{d}t} - y^*(t)\frac{\mathrm{d}y(t)}{\mathrm{d}t}}{a^2(t) + \epsilon^2} \tag{5}$$

where ϵ is a small damping factor added to ensure numerical stability [e.g. Matheney and Nowack, 1995].

It can be shown that the instantaneous centre frequency of the signal f_{C} is related to the instantaneous frequency,

$$f_{\mathrm{C}} = \frac{\int_0^\infty f A(f)\,\mathrm{d}f}{\int_0^\infty A(f)\,\mathrm{d}f} \equiv \frac{\int_0^\infty f_{\mathrm{I}}(t) a(t)^2\,\mathrm{d}t}{\int_0^\infty a(t)^2\,\mathrm{d}t} \tag{6}$$

where $A(f)$ is the Fourier power spectrum [e.g. Barnes, 1993; Dasios et al., 2001], so the time-averaged instantaneous centre frequency $f_{\mathrm{C}}(t)$ is

$$f_{\mathrm{C}}(t) \approx \frac{\int_{t-W}^{t+W} f_{\mathrm{I}}(t') a(t')^2\,\mathrm{d}t'}{\int_{t-W}^{t+W} a(t')^2\,\mathrm{d}t'} \tag{7}$$

for a window of length $2W$ [Matheney and Nowack, 1995].

We perform an iterative search across t^*_{CAO} values with $W = 5$ s, beginning from 0–10 s in increments of 1 s and decreasing by factors of 10 until we reach a precision of 10^{-2} s. For S and ScS, if $f_{\mathrm{S}} < f_{\mathrm{ScS}}$ (i.e. S is more attenuated than ScS) then S is the reference phase and ScS is the target phase, so $\delta t^*_{\text{ScS-S}} = -t^*_{\mathrm{CAO}}$; otherwise $\delta t^*_{\text{ScS-S}} = t^*_{\mathrm{CAO}}$ [e.g. Ford et al., 2012; Durand et al., 2013].

2.2.3 Waveform matching method

Time domain methods have also been used to estimate relative attenuation of seismic phases [e.g. Bhattacharyya, 1998; Reid et al., 2001]. The amplitude spectrum of S and ScS can be written as

$$A_{\mathrm{S}}(\omega, \Delta) = S(\omega)\; R(\omega)\; I(\omega)\; G_{\mathrm{S}}(\Delta)\; D(\omega, t^*_{\mathrm{S}})\; \exp[-i\omega t_{\mathrm{S}}] \tag{8}$$

$$A_{\mathrm{ScS}}(\omega, \Delta) = S(\omega)\; R(\omega)\; I(\omega)\; G_{\mathrm{ScS}}(\Delta)\; D(\omega, t^*_{\mathrm{ScS}})\; \exp[-i\omega t_{\mathrm{ScS}}] \tag{9}$$

where ω is the angular frequency, Δ is the epicentral distance, $S(\omega)$ is the source spectrum, $R(\omega)$ is the site response, $I(\omega)$ is the instrument response, $G(\omega, \Delta)$ includes the source radiation pattern and geometric spreading, $D(\omega)$ is the causal attenuation operator as defined in Equation 3, and t_{S} and t_{ScS} are the travel times of S and ScS, respectively [e.g. Reid et al., 2001]. Assuming that the source spectrum, site response and instrument response are similar for both phases, we can relate the spectra of S and ScS as

$$A_{\mathrm{ScS}}(\omega, \Delta) = G_{\text{ScS-S}}(\Delta)\; A_{\mathrm{S}}(\omega, \Delta)\; D(\omega, \delta t^*_{\text{ScS-S}})\; \exp[-i\omega \delta t_{\text{ScS-S}}] \tag{10}$$

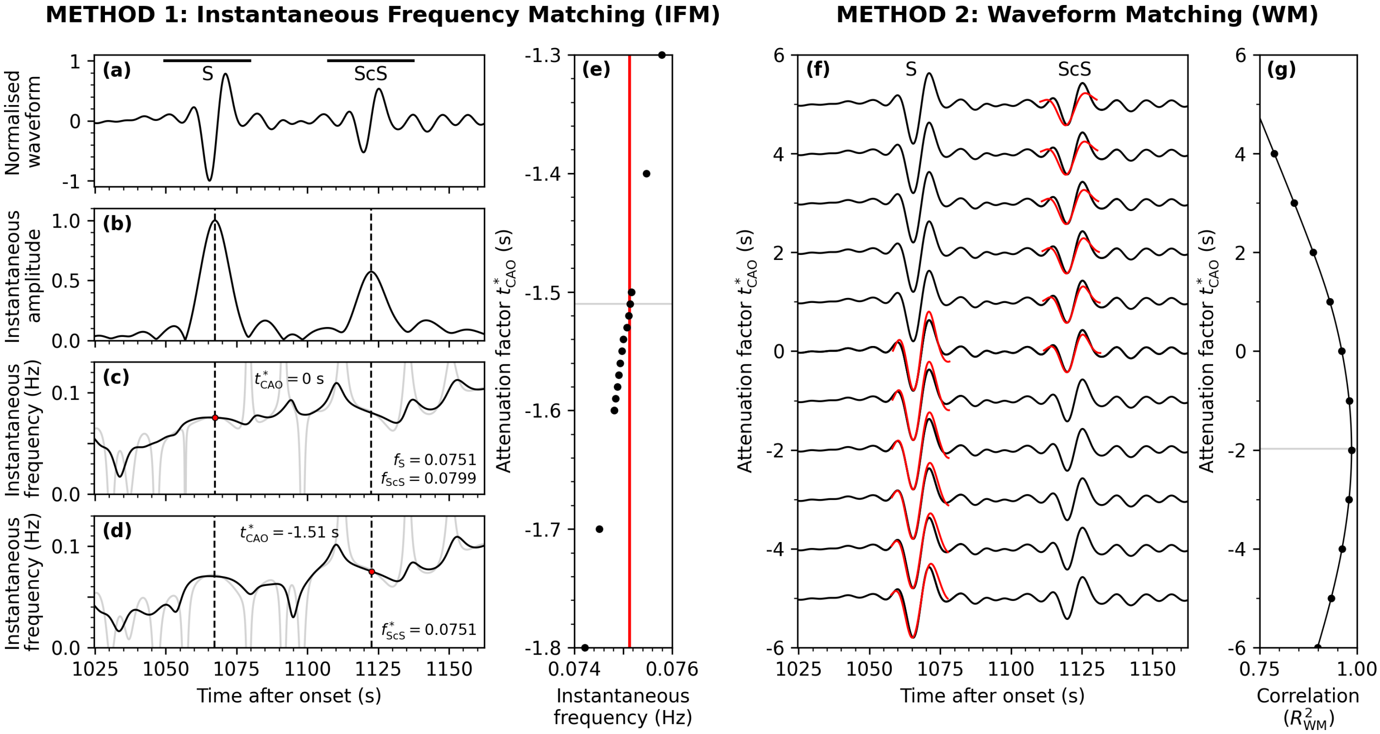

Figure 3: An overview of the two methods used to estimate differential attenuation between S and ScS in this study. Method 1 **(left)** is instantaneous frequency matching (IFM): **(a)** The transverse velocity seismogram for a Mw 6.6 earthquake occurring on 1997/11/28 in South America at 600 km depth, recorded at Southern California Seismic Network station PLM, bandpass filtered between 0.02–0.12 Hz. The black horizontal bar marks a window of ±15 s around the predicted S and ScS arrival times from PREM. The corresponding **(b)** instantaneous amplitude and **(c)** instantaneous frequency of the trace. The dashed lines mark the peaks of instantaneous amplitudes which are used as the reference points to estimate the instantaneous frequencies of S and ScS. The black line is the instantaneous frequency which has been weighted by the instantaneous amplitude, whereas the light grey line is the unweighted instantaneous frequency of the waveform. **(d)** The instantaneous frequency of the trace which has been convolved with the attenuation operator with $t^*_{\mathrm{CAO}} = -1.51$ s. The negative value indicates that S is more attenuated than ScS in this trace. **(e)** A range of values of t^*_{CAO} are trialled iteratively until the instantaneous frequency of the attenuated phase, f^*_{ScS} (black circles), matches the instantaneous frequency of the unattenuated reference phase, f_{S} (red line). Method 2 **(right)** is waveform matching (WM): **(f)** Waveforms which have been synthetically attenuated with different attenuation operator values (red) are overlaid on the original waveforms (black). For positive values of t^*_{CAO}, S waveforms have been attenuated and overlaid on the original ScS waveforms; for negative values of t^*_{CAO}, ScS waveforms have been attenuated and overlaid on the original S waveforms. **(g)** The correlation R^2_{WM} between the attenuated and reference signals is calculated for a range of attenuation operator t^*_{CAO} values. The optimum relative attenuation between the two phases can then be estimated from interpolating between these values. The values of $\delta t^*_{\text{ScS-S}}$ for this event-station pair from IFM and WM are -1.51 s and -1.98 s, respectively.

where $G_{\text{ScS-S}}(\Delta) = G_{\text{ScS}}(\Delta)/G_{\text{S}}(\Delta)$ is a constant for a given event-station pair, $\delta t^*_{\text{ScS-S}} = t^*_{\text{ScS}} - t^*_{\text{S}}$ and $\delta t_{\text{ScS-S}} = t_{\text{ScS}} - t_{\text{S}}$. For S and ScS, the assumption that the source spectra of both signals is the same is reasonable, as long as the take-off angles of the two phases are in the same lobe of the radiation pattern [e.g. Flanagan and Wiens, 1994].

In the time domain, the relative attenuation operator can therefore be found simply by minimising the misfit between the reference and attenuated target waveforms (Figure 3f-g), which we quantify using the Pearson correlation coefficient R_{WM}. We apply the attenuation operator to the target waveforms for values of t^*_{CAO} between 0–10 in increments of 1 s for both S and ScS as the target signal (since either phase can be more attenuated), and calculate the associated correlation coefficient allowing the attenuated target waveform to shift in time to maximise similarity with the reference waveform. From this, we then interpolate between the observed correlation coefficients to identify the value of $\delta t^*_{\text{ScS-S}}$ as t^*_{CAO} with the maximum value of R_{WM} (Figure 3g). Some studies instead jointly determine the relative attenuation, amplitude ratio, time delay and source parameters in a non-linear parameter inversion [e.g. Reid et al., 2001; Garcia et al., 2013; Pejić et al., 2017], but this additional complexity introduces trade-offs which are unnecessary unless the source terms for both phases are not similar.

Assuming that the geometric spreading factor, source radiation pattern, and reflection and transmission coefficients are well known, it is also possible to use a simplified approach to this method of directly inferring δt^* from the envelope ratios of two signals bandpass filtered in a narrow window. For example, this is a commonly used approach to image attenuation in the uppermost inner core from envelope ratios of PKiKP and PKIKP [e.g. Cao and Romanowicz, 2004; Waszek and Deuss, 2013], but has also been used in mantle imaging by comparing envelope amplitudes of S multiples with synthetic waveforms [Zhu et al., 2022] and comparing the amplitude ratios of s to ScS [Zhang et al., 2019]. Hansen et al. [2021] used a similar method to correct for relative attenuation effects between P and PcP waveforms to detect ultra-low velocity zones in the Antarctic region, but did not investigate spatial distributions or patterns of $\delta t^*_{\text{PcP-P}}$ values. Some studies using the spectral ratio method have also used an implicit form of this method to simply verify that the attenuated waveforms (using the attenuation factor determined by spectral division as t^*_{CAO} in the causal attenuation operator) matched the reference phase better than the unattenuated waveform, and discarded data for which a worse match was recovered, except for weakly attenuating data for which this requirement was waived [e.g. Fisher et al., 2003; Lawrence and Wysession, 2006].

Waveform matching is less precise than frequency domain methods – which can be seen from the wide range of t^*_{CAO} values that fit the waveforms almost equally well in Figure 3g compared to the apparent high precision of the instantaneous frequency matching method in Figure 3e. However, waveform matching does not suffer from the same issues as frequency domain methods with regards to spectral holes – 'missing' amplitude in the spectral content – since it is taken as an average measurement over a filter band [e.g. Tanaka and Tkalčić, 2015]. Bhattacharyya [1998] warns that this time domain method can be contaminated by interfering phases resulting in a measurement error of up to 2 s, and that they found frequency domain methods to be more accurate in synthetic tests, but we address this by disregarding waveforms for which phase pairs are predicted to be too close to distinguish (as discussed in Section 2.1).

2.2.4 Combined fitting

We thus opt for a joint fitting approach with the instantaneous frequency matching (IFM) and waveform matching (WM) methods, as likewise recommended by studies that compared frequency and time domain methods [e.g. Der and Lees, 1985; Dasios et al., 2001]. Agreement between the two methods, which have different drawbacks and strengths, significantly increases confidence in the values we obtain. Figure 4a shows the measurements of differential attenuation for IFM and WM, where the points are coloured by the similarity R^2_{IFM} between the reference and attenuated target waveforms from the IFM method. (We note that, by definition of the methodology, $R^2_{\mathrm{WM}} \geq R^2_{\mathrm{IFM}}$). There are a number of results clustered around $\delta t^*_{\mathrm{IFM}} = 0$ s, due to the way that S and ScS are trialled as the target and reference phases, but these mainly occur for waveforms with low similarity (Figure 4b). If we apply additional criteria to our dataset for a minimum similarity R^2_{C} between the reference and attenuated target waveforms, we find exceptionally good correlation Pearson correlation $R > 0.9$ between the two methods (Figure 4b-c). Based on this analysis, we set a further criteria that $R^2_{\mathrm{IFM}} \geq R^2_{\mathrm{C}} = 0.9$, and we require that the difference between the two methods $\Delta t^* = |\delta t^*_{\mathrm{WM}} - \delta t^*_{\mathrm{IFM}}|$ is less than 3 s (Figure 4c). The measurement of differential attenuation is then taken as the average of the two methods.

Following Ford et al. [2012], we apply a bandpass filter between 0.02–0.12 Hz on the waveforms before estimating attenuation. We then also require that measurements of the weighted instan-

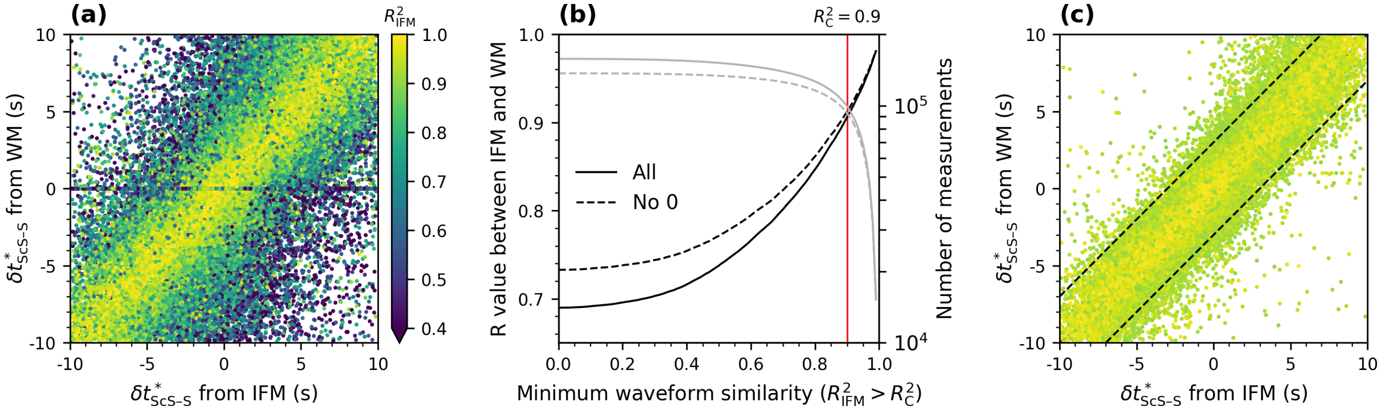

Figure 4: **(a)** A comparison of $\delta t^*_{\mathrm{ScS\text{-}S}}$ measurements from the instantaneous frequency matching (IFM) and waveform matching (WM) methods. The colour of the data point corresponds to the correlation R^2_{IFM} between the waveforms of the original reference phase and the attenuated phase using the value of $\delta t^*_{\mathrm{ScS\text{-}S}}$ obtained by the IFM. **(b)** The Pearson correlation value R between measurements of $\delta t^*_{\mathrm{ScS\text{-}S}}$ from the two methods as a function of minimum waveform similarity ($R^2_{\mathrm{IFM}} > R^2_{\mathrm{C}}$). **(c)** The dataset for measurements with a minimum waveform similarity R^2_{C} of 0.9, i.e. $R^2_{\mathrm{IFM}} > 0.9$. The dashed lines are the imposed condition that $\Delta t^* = |\delta t^*_{\mathrm{WM}} - \delta t^*_{\mathrm{IFM}}| < 3$ s.

taneous frequencies of both S and ScS are between 0.03–0.11 Hz, which excludes fewer than 20 measurements in total (less than 0.02% of measurements). We take this as evidence that our choice of bandpass preserves the frequency spectra of the data whilst minimising noise.

3 Results

3.1 Preliminary data analysis

We measured δt^* between S and ScS using both IFM and WM (see Methods). Our final dataset is shown in Figure 5, separated into six event depth ranges from 75–200 to 600–700 km and plotted as a function of epicentral distance.

The most obvious feature of the data is that there is a significant amount of scatter in the measurements ($\sigma_{\mathrm{noise}} \sim 3$ s) which does not vary with the threshold similarity of waveforms R^2_{C}. We therefore average the values of differential attenuation and travel time residuals (corrected for attenuation) at the ScS bounce point on a quasi-equal area cell map of the CMB (Figure 6). The travel time residuals show the expected global scale pattern of LLSVPs, with ScS raypaths travelling through the LLSVPs being more delayed relative to S than predicted by PREM (Figure 6b). However, no such large scale features are immediately obvious for the measurements of differential

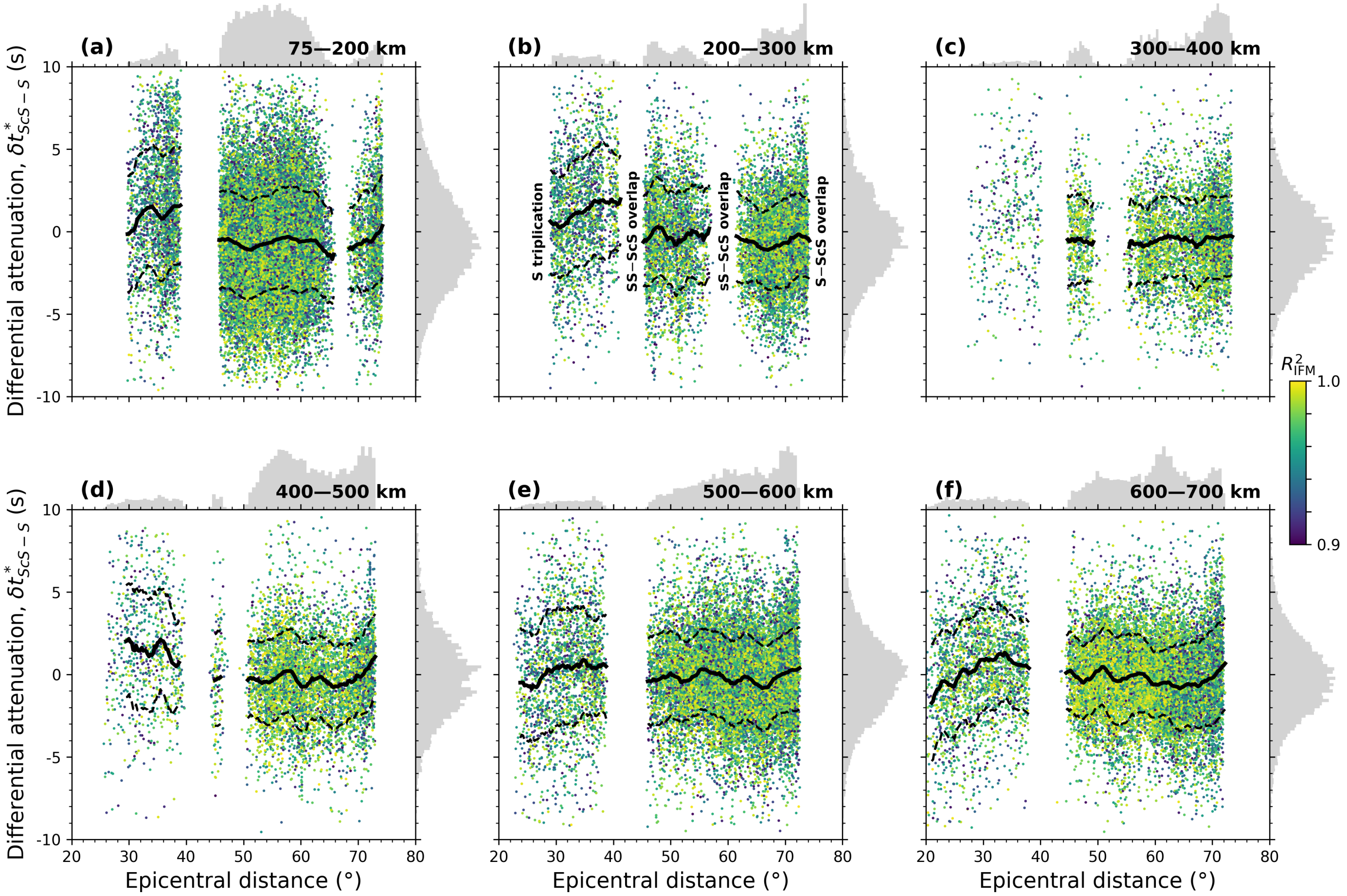

Figure 5: The mean value of differential attenuation $\delta t^*_{\text{ScS-S}}$ between the IFM and WM methods for each waveform within our dataset. The data are plotted in six event depth ranges: **(a)** 75–200, **(b)** 200–300, **(c)** 300–400, **(d)** 400–500, **(e)** 500–600, **(f)** 600–700 km. The colour of each data point is the waveform similarity R^2_{IFM} between the attenuated phase and the reference phase calculated in the IFM method. The mean (solid black) and standard deviation (dashed black) of each depth range have been overlaid, using a 2° rolling window. Normalised histograms of the epicentral distance and differential attenuation values have been projected on the upper and right sides of the subplots, respectively, for each of the event depth ranges. The gaps in data are a result of SS and sS overlapping with ScS at shorter and longer epicentral distances, respectively.

attenuation (Figure 6a). We also plot the measurements of differential attenuation separated by turning depth of the S raypath in map view (Figure S3). In this representation, there is a clear global distribution of positive $\delta t^*_{\text{ScS-S}}$ between 800–1000 km turning depth of S (approximately 20–30° epicentral distance), and a clear global distribution of negative $\delta t^*_{\text{ScS-S}}$ for turning depths 1000–2000 km (where 2000 km is the deepest turning depth of S in our dataset, around 75° epicentral distance). That is to say, ScS is more attenuated than S between 20–30° and S is more attenuated than ScS between 30–75°. This matches with the overall distribution of measured instantaneous frequencies of S and ScS (Figure S4), $f_{\mathrm{S}} = 72.1 \pm 9.0$ mHz and $f_{\mathrm{ScS}} = 72.8 \pm 9.4$ mHz, which shows that on average S is more attenuated than ScS.

Before analysing the dataset further, we first benchmarked our aggregated observations (Figure 7a) against other published datasets of $\delta t^*_{\text{ScS-S}}$ (Figure 7b). Lai et al. [2019] used waveform matching of empirical wavelets and Sun et al. [2025] used the instantaneous frequency matching method to estimate the absolute attenuation of shear phases relative to synthetics computed for tomographic model S40RTS [Ritsema et al., 2011]. From these datasets, we then constructed the relative attenuation if both measurements were available at a given event-station pair (for Lai et al. [2019]) or epicentral distance (for Sun et al. [2025]). Lawrence and Wysession [2006] used the spectral ratio method to estimate the relative attenuation of ScS and S, which is directly comparable to our measurements, but is also the least consistent with our measurements. Whilst there are differences between the datasets, there are some reassuring similarities: (i) the relative attenuation is more positive at shorter epicentral distances than at longer epicentral distances; (ii) there is a consistent increase in relative attenuation between 65–75°; and (iii) a consistent minimum in relative attenuation around 65–70°. However, there are clear offsets in the averages between datasets and also in the amplitude of our dataset compared to the others. It is difficult to ascertain whether this is due to lateral variations in the data sampling distributions or the accuracy of the measurement techniques used in the studies, or both.

3.2 Forward modelling

Since measurements of $\delta t^*_{\text{ScS-S}}$ represent the cumulative attenuation difference between two raypaths, interpreting the data is not a straightforwardly intuitive process. Our first step was therefore to perform forward modelling to establish how variations in Q_μ influence the predicted $\delta t^*_{\text{ScS-S}}$. Figure 8 shows how perturbations in Q_μ in different depth layers affect the predicted $\delta t^*_{\text{ScS-S}}$ as a function of epicentral distance. At shallow depths, from the surface to ~750 km depth, there is very little deviation from a uniform Q_μ for strong positive or negative perturbations due to the negligible difference between the S and ScS raypaths. Towards the bottom of the mantle, below ~2200 km depth, positive and negative perturbations in layers of the uniform Q_μ profile result in approximately linear negative and positive shifts in $\delta t^*_{\text{ScS-S}}$, respectively. This occurs due to ScS spending approximately the same time at each depth, which varies little with epicentral distance

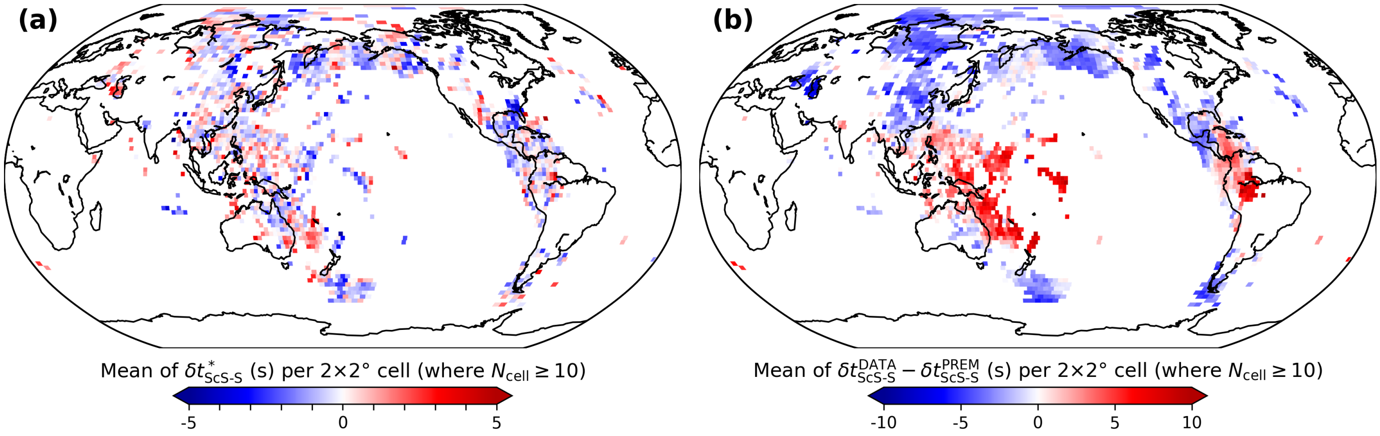

Figure 6: Mean values of **(a)** $\delta t^*_{\mathrm{ScS\text{-}S}}$ and **(b)** $\delta t^{\mathrm{DATA}}_{\mathrm{ScS\text{-}S}} - \delta t^{\mathrm{PREM}}_{\mathrm{ScS\text{-}S}}$ per $2 \times 2°$ quasi-equal area cell where grid cells with fewer than 10 ScS bounce points have been omitted.

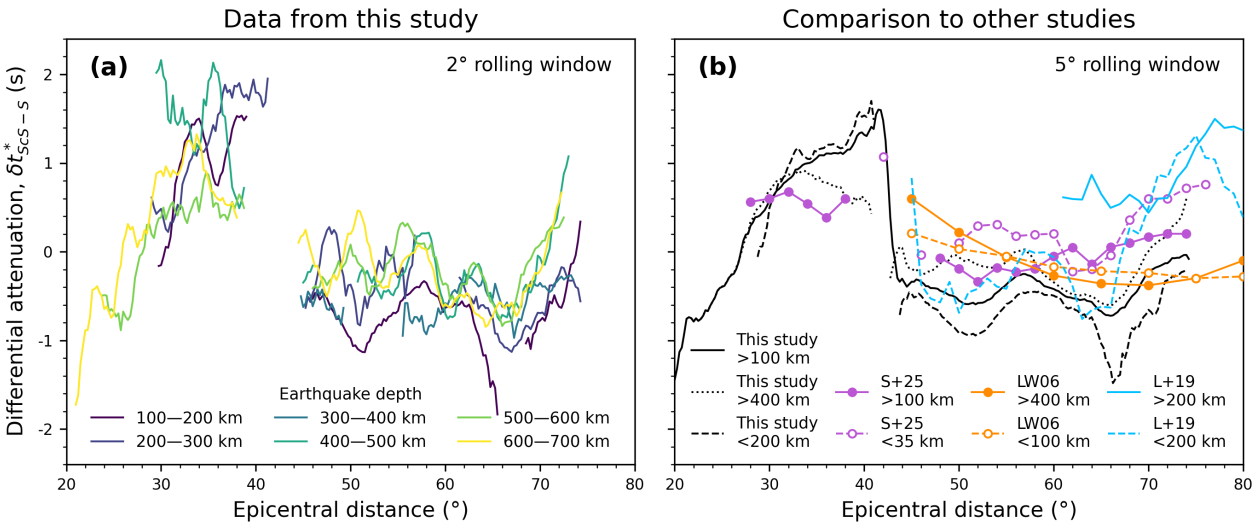

Figure 7: **(a)** The mean value of $\delta t^*_{\mathrm{ScS\text{-}S}}$ for measurements in this study as a function of epicentral distance and earthquake depth, using a 2° rolling window. **(b)** Comparison of our dataset to other published datasets, using a 5° rolling window. Lawrence and Wysession [2006] made measurements of $\delta t^*_{\mathrm{ScS\text{-}S}}$ for deep ($>$ 400 km, solid orange) and shallow ($<$ 100 km, dashed orange) earthquakes. Lai et al. [2019] made measurements of absolute attenuation for various phases; where two measurements were made for the same event-station pair, these were binned and averaged in the same way as our dataset but with a minimum threshold of 40 data per bin (blue). A comparison of the $\delta t^*_{\mathrm{ScS\text{-}S}}$ values constructed from the dataset of Lai et al. [2019] to our dataset is shown in Figure S5. Sun et al. [2025] also made measurements of attenuation for a range of S and ScS multiples relative to PREM; estimates of $\delta t^*_{\mathrm{ScS\text{-}S}}$ were constructed from the aggregated mean values for each of S and ScS for a 5° rolling window and uncorrected to PREM (purple).

(Figure S1a). However, at intermediate depths, between ∼750–2200 km, there are strong non-linear effects, with some epicentral distance ranges shifted positively and others shifted negatively. We thus refer to our final model, which is a 1D model with sensitivity mostly to the mid-mantle region, as Q_{M1D}. In a radially isotropic Earth, the relationship between the $Q\mu$ profile with depth and resulting $\delta t^*_{\text{ScS-S}}$ with epicentral distance can therefore be approximately interpreted as a direct mapping, since the depth in the Q_μ profile is largely determined by the turning depth of the S raypath.

We include epicentral distances less than 40° despite potential triplication of S arrivals, as the waveforms are not strongly affected by this effect [e.g. Wang et al., 2018] and there are relatively few measurements at this epicentral distance range (Figure 5). Since we require WM and IFM to produce similar values, $\Delta t^* < 3$ s, and the attenuated waveform matches the unattenuated reference waveform, $R^2_{\mathrm{C}} = 0.9$, this likely has a negligible effect on our dataset. We also visually inspected S waveforms in this epicentral distance range for several events to confirm the absence of triplications. At epicentral distances much longer than 75°, the separation between S and ScS waveforms in the seismograms becomes too small to robustly interpret (Figure 1c).

3.3 Inversion for 1D Q_μ

We next perform an inversion for 1D Q_μ, in which we compute the forward model for earthquake depths of 100 to 700 km in intervals of 50 km and epicentral distances of 20 to 80° in intervals of 0.5°. We then compute the misfit by comparing the measured $\delta t^*_{\text{ScS-S}}$ with the model corresponding to the nearest depth and epicentral distance of the event-station pair for the measurement. Before the inversion, we pre-compute the raypaths of S and ScS using TauP [Crotwell et al., 1999] through ObsPy [Beyreuther et al., 2010] for PREM, which makes each computation of the forward model extremely efficient. One full iteration – to compute $13 \times 121 = 1573$ forward models and the corresponding misfit for 90,000 measurements – takes on the order of 0.5 s on a single processor. As such, it becomes feasible to perform a Markov chain Monte Carlo (McMC) inversion which is able to explore a wide parameter space and, most importantly, produce robust estimates of the associated uncertainties [e.g. Bodin and Sambridge, 2009]. Estimates of uncertainty derived from Bayesian inversion methodologies are also more reliable and realistic compared to those from linearised inversions [e.g. Rudolph et al., 2020], such as the one by Sun et al. [2025].

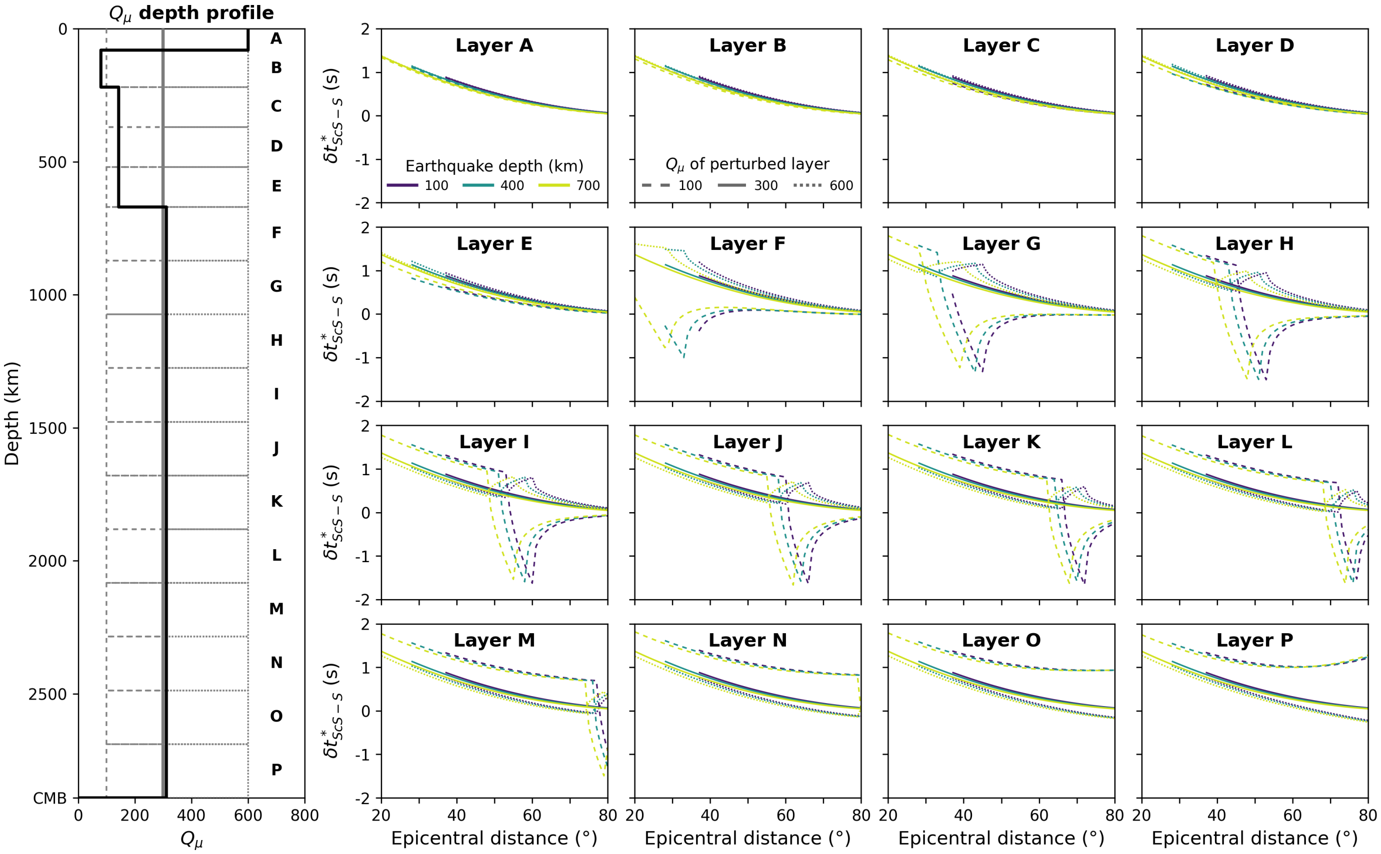

Figure 8: A depth profile **(left)** has been split up into 16 approximately equal-depth ranges and labelled **Layers A–P**. The Q_μ profile for PREM is plotted for reference (black line). Subplots **(right)** show the predicted $\delta t^*_{\text{ScS-S}}$ profiles as a function of epicentral distance for a perturbation in each layer only for event depths of 100 km (blue), 400 km (teal green) and 700 km (light green). The value of Q_μ in all other depth ranges is kept at 300. Each perturbed layer is allowed the value of 100 (dashed) or 600 (dotted), which is compared to a uniform profile of $Q_\mu = 300$ (solid).

We parameterise the Q_μ profile as a function of cubic B-splines in depth, such that

$$Q_\mu(r) = \sum_k p_k f_k(r) \tag{11}$$

where $f_k(r)$ are the cubic B-splines functions at radius r and p_k are the corresponding weighting factors of the basis functions. Since the weights of cubic B-splines are strictly positive, it means that all Q_μ profiles will be positive for $p_k > 0$. We thus allow p_k to vary between 1 and 1000 (the prior distribution). From our initial analysis of the data, we expect a strongly heterogeneous region between 500–1000 km depth where our data shows strong variability. As such, we chose to use basis functions with 32 depth knots to ensure adequate depth resolution. For the uppermost

400 km, we use knots with fixed separation of 50 km with p_k chosen to be PREM-like, with the remaining knots distributed such that the spacing between nodes increases linearly down to the CMB (Figure 9a) following Sun et al. [2025].

The misfit $\Phi(m)$ of a model m of Q_μ is defined as

$$\Phi(m) = \sum_i \left(\frac{p_i(m) - d_i}{\sigma_{\text{noise}}} \right)^2 \tag{12}$$

where p is the predicted value of $\delta t^*_{\text{ScS-S}}$ for the model m, d is the observed value of $\delta t^*_{\text{ScS-S}}$, and σ_{noise} is the estimated value of the data noise (chosen as 5 s). This is larger than the standard deviation of the overall dataset of $\delta t^*_{\text{ScS-S}}$ measurements ($\sigma_{\text{noise}} \sim 3$ s), but this was chosen to ensure the parameter space was well explored in a manner similar to simulated annealing [e.g. Weber, 2000]. For a proposed model m', one randomly chosen weighting factor is perturbed by a step length drawn from a Gaussian distribution, $p'_k = p_k + \mathcal{N}(\Delta p_k)$. The forward model scheme is then run for the 1573 combinations of event depth and station epicentral distance using the proposed model, and the misfit is calculated for the dataset. The natural logarithm of the acceptance probability is [e.g. Bodin and Sambridge, 2009]

$$\ln\left[\alpha\big(m' \big| m\big)\right] = \min\left[0, \tfrac{1}{2}\big(\Phi(m') - \Phi(m)\big)\right] \tag{13}$$

The acceptance probability is then compared to a randomly drawn value $w = [0, 1)$: if $\alpha > w$ the proposed model m' is accepted, otherwise it is rejected and a new proposal is drawn from the current model m. The form of the acceptance probability means that all proposed models which decrease the misfit are accepted, whilst those which increase the misfit can sometimes be accepted depending on the value of w. The posterior distribution, the probability distribution of Q_μ profiles which fit our dataset, is then taken to be the sparsely sampled distribution of models m after a sufficient burn in period to remove transient effects of the starting conditions [e.g. Bodin and Sambridge, 2009]. We find that for the setup presented here – 32 depth knots with the upper 6 knots fixed to PREM-like values – the inversion converges after a burn in period of ∼2000 iterations. The inversion ensemble is then taken as every 500th iteration from 50 parallel inversions after a burn in period of 2500 iterations and run for 12000 iterations. The ensemble distribution therefore comprises of 1000 uncorrelated models from the posterior distribution, from which we calculate the mean and standard deviation of each of the weighting factors p_k.

We also explored a range of other inversion setups, including:

1. Changing the number of depth knots and their relative separation (Figures S7 & S8). We found that, regardless of the number of knots, the resulting Q_μ profile is very consistent although the profiles with fewer knots led to broader features. 32 depth knots was a balance between sufficient depth resolution and minimising overfitting noise.

2. Parameterising Q_μ in logarithmic space, similar to parameterisation used by Sun et al. [2025]. However, we found that this parameterisation was less stable due to the strong variations in Q_μ required to fit our observed data unless we used a large number of depth splines.

3. Adding additional terms to the misfit if the predicted values of values of absolute attenuation t^*_{S} were higher than those of PREM. This test was a result of recent measurements of absolute attenuation of t^*_{S} which were approximately equal to or less attenuating than PREM synthetics across all measured epicentral distances [Zhu et al., 2022; Sun et al., 2025].

4. Using different starting models to initiate the inversion. Due to the relatively high value of σ_{noise} we use, the inversion was always well converged by ∼2000 iterations for any uniform or arbitrary starting models of Q_μ.

Our final model, Q_{M1D}, is taken as the mean of the distribution of Q_μ profiles in the inversion ensemble with the associated uncertainty in the model taken as the uncertainty in the weighting factor of each depth knot (Figure 9b). Our model shows a strongly attenuating layer ($Q_\mu < 100$) at 750 km depth and two weakly attenuating layers ($Q_\mu \sim 600$) at 900 and 2000 km depth. Between the two weakly attenuating layers, the mid-mantle is relatively strongly attenuating ($Q_\mu \sim 150 - 250$). We then also compare our model to several previously published models (Figure 9c-d): PREM [Dziewonski and Anderson, 1984], QM1 [Widmer et al., 1991], QL6 [Durek and Ekström, 1996], QDW14 [De Wit et al., 2014], REM1D [Moulik and Ekström, 2025], QS4L3 [Talavera-Soza et al., 2025], QLM9 [Lawrence and Wysession, 2006], QHR11 [Hwang and Ritsema, 2011], QMSF [Zhu et al., 2022] and Q_{L1D} [Sun et al., 2025]. PREM, QM1, QL6, QDW14, REM1D and QS4L3 are constructed from measurements of quality factors of normal modes; whilst QLM9, QHR11, QMSF,

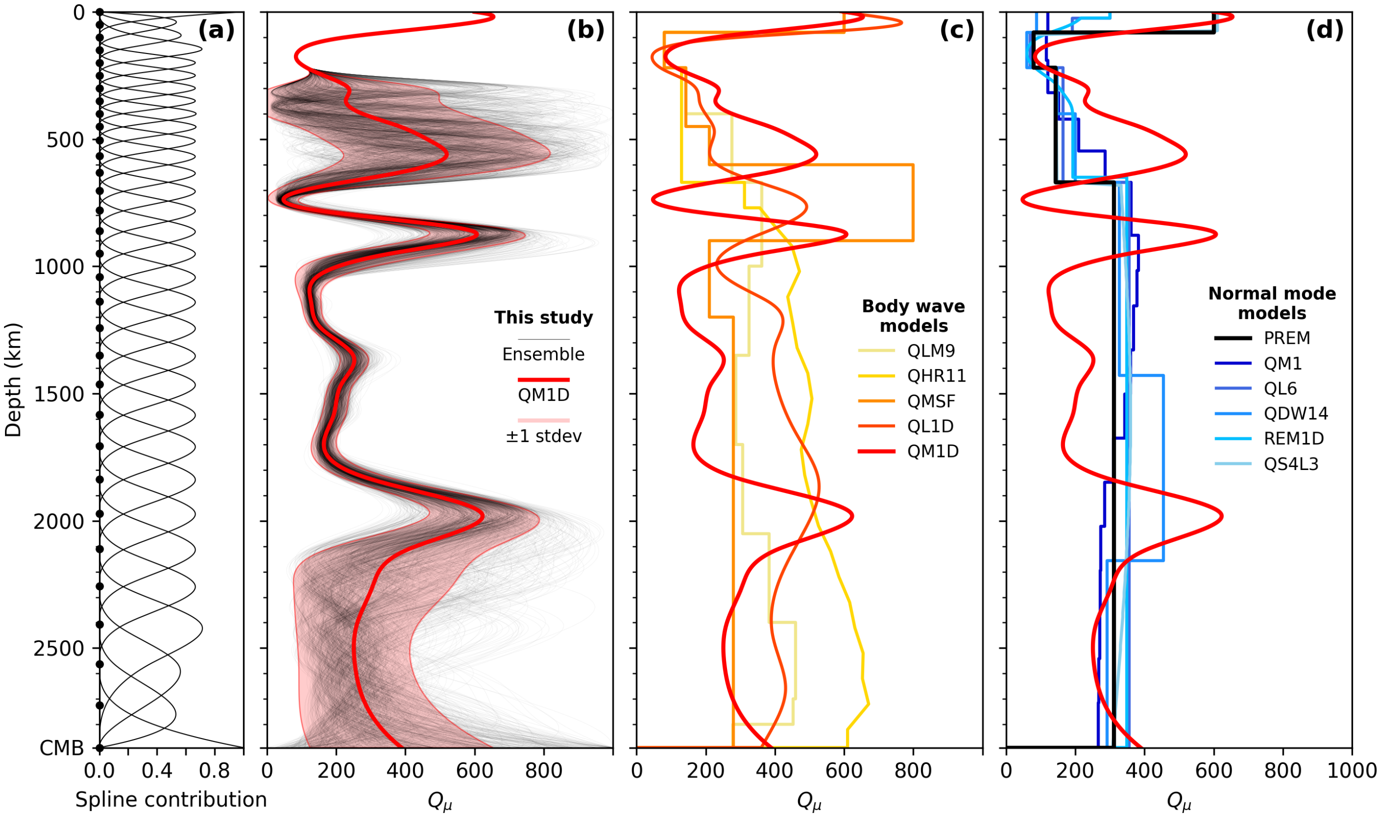

Figure 9: **(a)** The cubic B-splines of the basis functions used in our modelling with 32 depth knots. **(b)** The ensemble of models from our inversion (thin black lines), with the mean and standard deviation of the ensemble (red and pink, respectively). The mean of the ensemble is taken to be our model Q_{M1D} with its corresponding uncertainties for each depth parameter. A comparison of our model with **(c)** other short period body wave models (warm colours) and **(d)** various long period normal mode models (cool colours) of shear attenuation: black [PREM, Dziewonski and Anderson, 1984], dark blue [QM1, Widmer et al., 1991], blue [QL6, Durek and Ekström, 1996], medium blue [QDW14, De Wit et al., 2014], sky blue [REM1D, Moulik and Ekström, 2025], light blue [QS4L3, Talavera-Soza et al., 2025], pale yellow [QLM9, Lawrence and Wysession, 2006], yellow [QHR11, Hwang and Ritsema, 2011], orange [QMSF, Zhu et al., 2022], dark orange [Q_{L1D}, Sun et al., 2025], red [Q_{M1D}, this study].

Q_{L1D} and Q_{M1D} are from measurements of attenuation of body waves. Predictions of $\delta t^*_{\mathrm{ScS\text{-}S}}$ for these Q_μ profiles are plotted against the running average of our measurements, which show that only Q_{M1D} matches the trends in our dataset (Figure 10).

3.4 Comparison with other body wave lower mantle models

Our model Q_{M1D} differs significantly from previous models, notably around 500–1000 km depth. Whereas Zhu et al. [2022] and Sun et al. [2025] observe a weakly attenuating region around 750 km

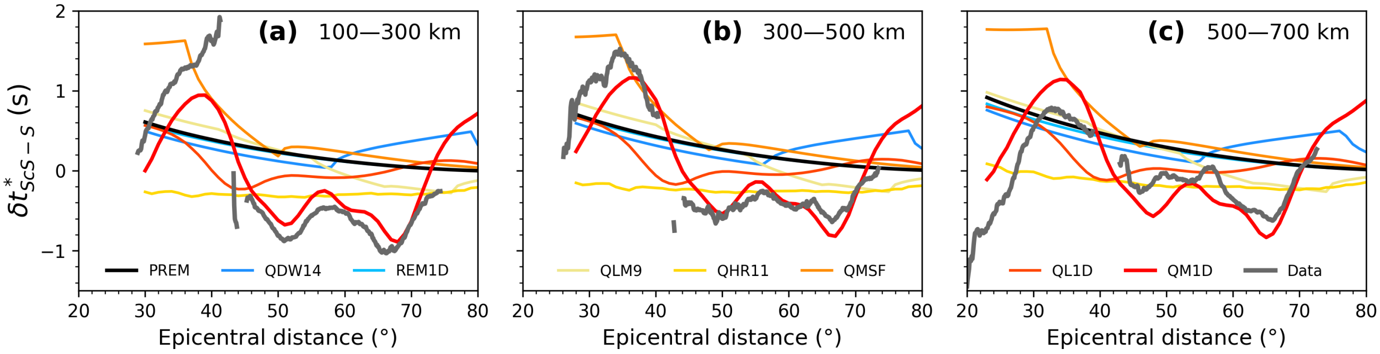

Figure 10: A comparison of predictions and observations of $\delta t^*_{\text{ScS-S}}$ for a few different Q_μ profiles from Figure 9. Running averages of $\delta t^*_{\text{ScS-S}}$ measurements are shown for event depth ranges **(a)** 100–300, **(b)** 300–500 and **(c)** 500–700 km, using a 5° rolling window (grey lines).

depth, our model prefers a thin strongly attenuating layer. There are also two very distinct low attenuation layers around 900 and 2000 km depth, the latter of which has not been robustly observed in any of the other models. These features in our Q_μ profile result in a significantly better fit to our dataset of $\delta t^*_{\text{ScS-S}}$ measurements than any of the other models, which are broadly flat across the 20–80° epicentral distance range (Figure 10).

As well as direct comparisons between our observations of $\delta t^*_{\text{ScS-S}}$ with those from other studies (Figure 7), we also compared our model Q_{M1D} to observations of other differential phase pairs, SS-S and SSS-SS, from previous studies (Figure S6a-d). Predictions from Q_{M1D} for these observations show a distinctly poor match, which is likely due to the poor resolution of the shallow mantle by the dataset we used to construct our model, but also due to the unmatched upper mantle and crustal contributions to the measurements for the differential phases. Similarly, several studies have found that their measurements are less attenuating than PREM, in contradiction with our results (Figure S6e-g). Zhu et al. [2022] find that, on average, the amplitudes of SS, SSS and SSSS phases are larger than those predicted in synthetics for PREM, implying that any model should be less attenuating than PREM overall. Hwang and Ritsema [2011] similarly found that their t^*_{data} was consistently below the predicted value of t^*_{PREM} for P and S waves. More recently, a similar analysis on a larger scale was completed by Sun et al. [2025] for S, SS, SSS, SSSS, ScS, ScS_2, ScS_3, Sdiff, and their respective depth phases. Whilst again there was a significant scatter in their data, there was a consistent trend that the phases were less attenuated than predicted by PREM. Sun et al. [2025] also suggested that there may be a low attenuation layer around 2000 km depth (although this is not clear from their model Q_{L1D}). Indeed, adding a more pronounced weakly attenuation layer in their preferred model Q_{L1D} at this depth improves the fit to their dataset of t^*_{SS} measurements (Figure S6f).

Our model Q_{M1D} will also have important implications several previous studies which invert for attenuation in the upper mantle using the assumption that the lower mantle contributes negligible attenuation [e.g. Warren and Shearer, 2002; Souriau et al., 2012; Nguyen et al., 2025]. Several studies have also assumed that a smoothly varying velocity profile from the base of the mantle transition zone (MTZ) to the top of the D" region implies lack of strong attenuation heterogeneity in this region [e.g. Flanagan and Wiens, 1994]. Our model therefore recommends caution to future studies assuming that the effect of attenuation in the lower mantle can be neglected.

Zhao et al. [2015] investigated the Pacific region by creating an empirical source function for and comparing the broadening of S. On average, as a function of epicentral distance, the S waves which turned within the lowermost 900 km were more attenuating than S waves which turned around 900 km above the CMB. This matches with our observation of a low attenuation region around 2000 km (approximately 900 km above the CMB) with stronger attenuation below this depth. This similarity could suggest that our observation of a low attenuation layer at 2000 km depth may be a globally present feature, rather than just around the Pacific where our dataset is spatially biased.

3.5 Comparison with normal mode models

Nearly all Q_μ profiles derived from measurements of quality factors of long period normal mode feature a roughly uniform weakly attenuating lower mantle with Q_μ between 300–400 [e.g. Dziewonski and Anderson, 1981; Moulik and Ekström, 2025], except for QR05 [Resovsky et al., 2005] and QDW14 [De Wit et al., 2014] which suggest that there is a weakly attenuating layer between 1500–2000 km depth. In our model, we find a more strongly attenuating mid-mantle (Q_μ = 150–250), flanked by two regions of low attenuation (Q_μ = 500–700). Given the poor depth resolution of normal mode models [e.g. Talavera-Soza et al., 2025], most models are constructed using a very coarse parameterisation of the lower mantle with the notable exception of QM1 [Widmer et al., 1991]. It is therefore plausible that the weakly attenuating layer observed by Resovsky et al. [2005] and De Wit et al. [2014] is the same as the one we find around 2000 km depth. The lowermost 600 km of the mantle in our model is also consistent with the long period Q_μ profiles, although we note that Q_{MID} is poorly constrained in this region and there is a large uncertainty in our model (Figure 9b). Finally, there do not appear to be any similarities between normal mode models and a more attenuating mid-mantle or for a weakly attenuating layer at 900 km depth, as suggested by Q_{M1D}.

Attenuation is generally considered to be frequency dependent, i.e. $Q = Q_0 f^{\alpha}$ [e.g. Anderson and Given, 1982], with a value of $\alpha \sim 0.1 - 0.3$ derived seismically [e.g. Lekić et al., 2009; Zaroli et al., 2010; Romanowicz and Mitchell, 2015] and experimentally [e.g. Jackson and Faul, 2010; McCarthy et al., 2011]. Over narrow frequency ranges, it is typically assumed that attenuation is frequency independent – for example, this is an assumption of the spectral ratio method – although Moulik and Ekström [2025] find that they do not need to invoke frequency dependence of attenuation to satisfy observations over the range 1–3200 s period, i.e. $\alpha = 0$. If attenuation is frequency dependent, $Q\mu$ measured from body waves (higher frequency) would be expected to be higher than those from normal modes (lower frequency) [e.g. Hwang and Ritsema, 2011; Sun et al., 2025]. Assuming a dominant period of 300 s for the observations of normal modes and a uniform layer of attenuation of the whole lower mantle with $Q_{\mu} = 312$ from PREM, we can compare this with our model Q_{M1D} from our dataset with a dominant period of 10–20 s to estimate values of α at each depth range of the mantle. In the two layers of weak attenuation at 900 and 2000 km depth we therefore obtain a value of α between 0.14 and 0.30, whilst in the mid-mantle between these two layers α is between -0.27 and -0.07. For the thin layer of strong attenuation at 750 km depth and in the lowermost mantle, α is -0.75 to -0.48 and -0.08 to 0.04, respectively. These values are consistent with previously reported ranges [e.g. Lekić et al., 2009], including that attenuation is independent of frequency in the lowermost mantle [e.g. Moulik and Ekström, 2025], except for the highly negative values of α we observe for the strongly attenuating thin layer beneath the MTZ.

We also computed values of the Q of normal modes for both PREM and our model Q_{M1D} using MINEOS [Woodhouse, 1988; Masters et al., 2011] with all other parameters kept the same as PREM, with the assumption that $\alpha = 0$. Comparing these predictions with the observed measurements from normal modes [Deuss et al., 2013] (Figure S12), we find that there is relatively good agreement between the different 1D profiles of Q_{μ} with the measurements, suggesting that a heterogeneous joint model of normal mode and body wave observations may be possible.

4 Discussion

4.1 Uncertainty in the data and methodology

Various effects – such as de/focussing, anisotropy, and scattering – can cause a decrease in amplitude or a broadening of seismic waveforms. Such effects result in a measured attenuation ('apparent attenuation') which is larger than the true attenuation due to anelastic absorption of energy ('intrinsic attenuation') [e.g. Cormier, 2020]. In the following sections, we discuss these effects and the potential influence they have on our observations and interpretations.

4.1.1 Effect of de/focussing of elastic structure

Chaves and Ritsema [2016] used synthetic modelling to demonstrate that measurements of ScS/ScS$_2$ amplitude ratios are dominated by focussing effects due to elastic structure, i.e. velocity, rather than 3D variations in attenuation. Similarly, Pang et al. [2023] found that focussing effects accounted for up to 20% of the differential attenuation between station pairs of shallow events measured using the spectral ratio method. However, they noted that the complex azimuthal patterns observed in shallow events were almost non-apparent for events deeper than 150 km.

At short epicentral distances, the layered velocity structure in the upper mantle results in triplication of S waves. This would likely lead to broadening of the S waveform and an apparent increase in t^*_{S} and thus a more negative $\delta t^*_{\mathrm{ScS\text{-}S}}$, which matches the trend in relative attenuation we observed (Figure 5). However, our relative attenuation trend appears to be robust across all event depths and extends well into epicentral distances for which triplications are not predicted (model predictions are truncated in Figure 10 when triplications are predicted for PREM). Additionally, the number of measurements at the shortest epicentral distances are far outnumbered by those at longer epicentral distances (Figure 5) and so any trend in our data due to triplication will have a weak effect on the misfit during the inversion process. Nonetheless, we only include data in the misfit for which PREM does not predict a triplication in S, resulting in a final dataset of 90,000 $\delta t^*_{\mathrm{ScS\text{-}S}}$ measurements from 894 events.

4.1.2 Effect of anisotropy

Presence of anisotropy in the mantle can also lead to the broadening of the SH phase, which could be misinterpreted as being due to anelastic attenuation [Cormier, 2020]. Anisotropy is thought only to be significant in the uppermost and lowermost mantle [e.g. Cormier, 2020]. Since S and ScS have similar raypaths near to the receiver in the upper mantle, any potential broadening would affect both phases and approximately cancel out. However, only ScS passes through the lowermost mantle and so it is possible that there is a small contribution from anisotropy in this region to broadening of ScS, resulting in a larger apparent differential attenuation between ScS and S (i.e. more positive $\delta t^*_{\text{ScS-S}}$). When we compare our dataset with other published values, we find that our average values of $\delta t^*_{\text{ScS-S}}$ are in fact more negative. Therefore, it us unlikely that anisotropy affects our dataset noticeably. Indeed, Ford et al. [2012] and Durand et al. [2013] – who both investigated the effect of correcting for anisotropy on measurements of deep earthquakes targetting the CMB near Central America – concluded that the effect was small (on the order of tenths of a second) and that corrections did not correlate with 3D velocity models. As such, we do not correct for any potential waveform broadening caused by anisotropy in this study.

4.1.3 Effect of scattering

Elastic structure on the order of wavelength of the seismic waves can cause the elastic redistribution (i.e. scattering) of energy which results in an effective pulse broadening and increase in apparent attenuation or spectral holes in the signals [e.g. Cormier, 2020]. Since scattering is a redistribution of energy, it should not be misinterpreted as anelastic attenuation which is absorption of energy. Previous studies have found that, whilst such scatterers may broaden waveforms, this would require significant variations in density and elastic parameters of 6–9%, which is much larger than observed in tomographic models [e.g. Ricard et al., 2014]. Small-scale structure may be present in the mid-mantle [e.g. Kaneshima, 2016; Waszek et al., 2018], which would result in more scattering of S than ScS on average and therefore lead to the more negative values of $\delta t^*_{\text{ScS-S}}$ we observe. However, most seismic observations of scatterers have been located within or near the D" layer in the lowermost mantle [e.g. Shearer, 2007; Rost et al., 2015; Rost and Frost, 2025] and there are significantly fewer studies which suggest there are small-scale scatterers in the mid-mantle. Therefore, since scatterers in the D" would only affect ScS, which would result in a more negative $\delta t^*_{\text{ScS-S}}$, it seems unlikely that scattering has had a significant impact on our measurements. Various studies have attempted to distinguish between intrinsic attenuation and scattering from apparent attenuation using radiative transfer theory [e.g. Fehler et al., 1992; Carcolé and Sato, 2010; Ogiso, 2019; Akande et al., 2019], but this is beyond the scope of this paper.

4.2 Comparison to mineralogy

Whilst in the upper 400 km of the mantle there is increasing evidence for an anti-correlation between 3D models of shear velocity and attenuation [e.g. Romanowicz, 1995; Dalton et al., 2008; Karaoğlu and Romanowicz, 2018; Talavera-Soza et al., 2025] – suggestive of related causal physical mechanisms such as temperature, grain size, water content or partial melt [e.g. Jackson et al., 2002] – there does not appear to be any similar correlation between 1D models of shear velocity and attenuation in the lower mantle. We compared our observations of differential attenuation $\delta t^*_{\text{ScS-S}}$ with shear wave velocity variations (V_{S}) from 3D tomographic models SEMUCB-WM1 [French and Romanowicz, 2014], REVEAL [Thrastarson et al., 2024] and GLAD-M35 [Cui et al., 2024] at the turning point of S (Figure S9) and at the bounce point of ScS (Figure S10) of our dataset. There is a weak anti-correlation between shear velocity at the turning point of S with differential attenuation ($R \sim -0.08$) and a weaker anti-correlation between shear velocity at the bounce point of ScS with differential attenuation ($R \sim -0.03$). The strongest – but still weak – anti-correlation is found between travel time residuals between observed and predicted $\delta t_{\text{ScS-S}}$ and differential attenuation (Figure S11, $R \sim -0.09$), which suggests that there is no coupled dominant physical mechanism driving perturbations in velocity and attenuation in the lower mantle.

Previous studies have interpreted low attenuation layers in their Q_μ profiles as evidence for regions of high viscosity [Lawrence and Wysession, 2006; Sun et al., 2025] (Figure 11). Several studies have found evidence for an increase in viscosity in the mantle around 1000 km depth from geodynamical modelling [e.g. Kido and Čadek, 1997; Forte and Mitrovica, 2001; Höink et al., 2012; Rudolph et al., 2015], experimental mineralogy [e.g. Marquardt and Miyagi, 2015], and seismic tomography [e.g. Fukao and Obayashi, 2013; French and Romanowicz, 2015]. Some have additionally identified an increase in viscosity around 2000 km depth [e.g. Ricard and Wuming, 1991; Forte and Mitrovica, 2001; Forte et al., 2010; Brown et al., 2026], matching our observation of a low

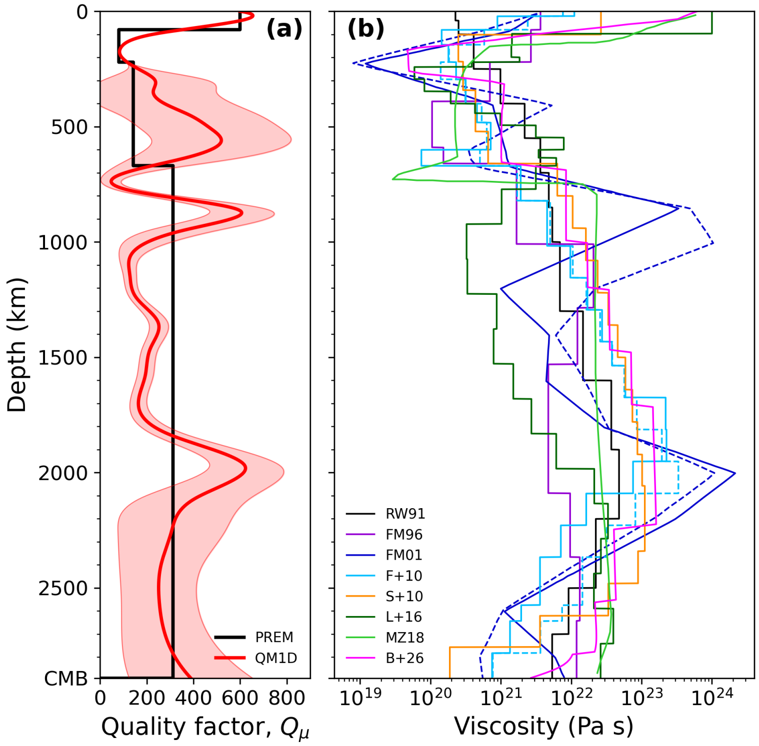

Figure 11: A comparison of radial profiles of **(a)** shear Q_μ for PREM and Q_{M1D} with **(b)** mantle viscosity: black [Ricard and Wuming, 1991], purple [Forte and Mitrovica, 1996], dark blue [Forte and Mitrovica, 2001], light blue [Forte et al., 2010], orange [Steinberger et al., 2010], dark green [Lau et al., 2016], light green [Mao and Zhong, 2018] and pink [Brown et al., 2026]. Dashed lines indicate alternative models.

attenuation layer at this depth, which Glišović et al. [2015] interpreted as the region of maximum grain size within the mantle. The thin highly attenuating layer in our model at 750 km depth is also consistent with previous suggestions of a thin low viscosity layer at the bottom of the MTZ which is thought to potentially control slab stagnation [e.g. Mitrovica and Forte, 2004; Mao and Zhong, 2018; Rudolph et al., 2020], although evidence for this is mixed [e.g. Liu et al., 2021]. Mao and Zhong [2021] further find that geodynamic models with a thin layer beneath the MTZ only in regions of slab subduction can explain observed slab stagnation equally well as geodynamic models including a global layer, which may explain why the layer is distinctly visible in our model (with data coverage predominantly in regions of subduction, Figure 2b) but not in others, e.g. Q_{L1D}.

4.3 Future developments and suggestions

ScS-S is fundamentally limited by the source-receiver geometries for which ScS can be measured without being contaminated by other phases (Figure 1b), resulting in a spatial sampling biased around the edge of the Pacific (Figure 2b). Other studies which have addressed the problem of spatial bias have typically measured absolute attenuation values of different shear phases [e.g. Lai et al., 2019; Zhu et al., 2022; Sun et al., 2025]. However, measuring the absolute attenuation is difficult and involves strong assumptions about the attenuation contributions from the source and receiver regions [Sun et al., 2025]. As such, other pairs of differential phases are required to increase the spatial sampling and ensure less uncertainty in measurements of δt^*. ScSScS and SS may be one such potential avenue to increase the spatial sampling of our dataset. While this phase combination will still have the same constraints of depth sensitivity – i.e. predominantly the mid-mantle – it would have a more global distribution than the spatially biased $\delta t^*_{\text{ScS-S}}$ dataset we present in this study. Such improvements in spatial sampling are required to make progress towards a 3D model of attenuation of the mantle from body waves.

Figure S7 shows the inversion results for a number of different depth splines with 24, 32, 50 and 90 knots for comparison with the 32 knots we used in our new model, Q_{M1D}. We recommend any future study using a spline parameterisation for Q_μ to implement an irregularly spaced arrangement of knots to ensure sufficient resolution around the 900 and 2000 km layers of low attenuation and the 750 km depth layer of high attenuation. Alternatively, it may be preferable to perform a transdimensional McMC in which the number of knots is also an unknown variable [e.g. Bodin and Sambridge, 2009], but this would create additional computational challenges if the objective is a 3D model of attenuation.

5 Conclusions

Using a dataset of 90,000 high-quality measurements of differential attenuation between S and ScS body waves, we have constructed a new 1D model of shear attenuation Q_μ in the mantle, Q_{M1D}. We find a thin strongly attenuating layer around 750 km depth, and two weakly attenuating layers at 900 and 2000 km depth. We speculate that key features in our attenuation profile are related to features observed in viscosity profiles of the mantle. The low attenuation layers at 900 and 2000 km depth could be related to increases in global viscosity in the deep mantle, and the high attenuation layer at 750 km depth could be related to a thin layer of low viscosity beneath the mantle transition zone in the presence of subducted slab material. Our model also suggests the mid-mantle is more attenuating than usually assumed from models based on normal modes, including the widely-used 1D reference model PREM.

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Data and code availability

The facilities of IRIS Data Services (www.iris.edu) were used to download waveforms and related metadata. IRIS Data Services are funded through the Seismological Facilities for the Advancement of Geoscience (SAGE) Award of the National Science Foundation under Cooperative Support Agreement EAR-1851048. Data were downloaded from data centres EarthScope, GFZ, NOA, RESIF, ORFEUS, GeoNet, AusPass, USP, KOERI and UIB-NORSAR. Additionally, data from the F-net network were made available by National Research Institute for Earth Science and Disaster Resilience [Okada et al., 2004]. Data from NIED was downloaded using HinetPy [Tian, 2024]. Earthquake parameters were used from the Global CMT Project (www.globalcmt.org). We used TauP [Crotwell et al., 1999] and ObsPy [Beyreuther et al., 2010] to do the analysis in this project, and matplotlib [Hunter, 2007] and basemap (https://matplotlib.org/basemap/) to make the figures. Codes and datasets will be made available on GitHub and Zenodo upon publication.

Competing interests

Authors declare that they have no competing interests.

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Supplementary materials for

Q_{M1D}: A 1D model of shear attenuation in the mantle from differential body waves

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SUPPLEMENTARY

In this supplementary are:

- Figures S1–S12

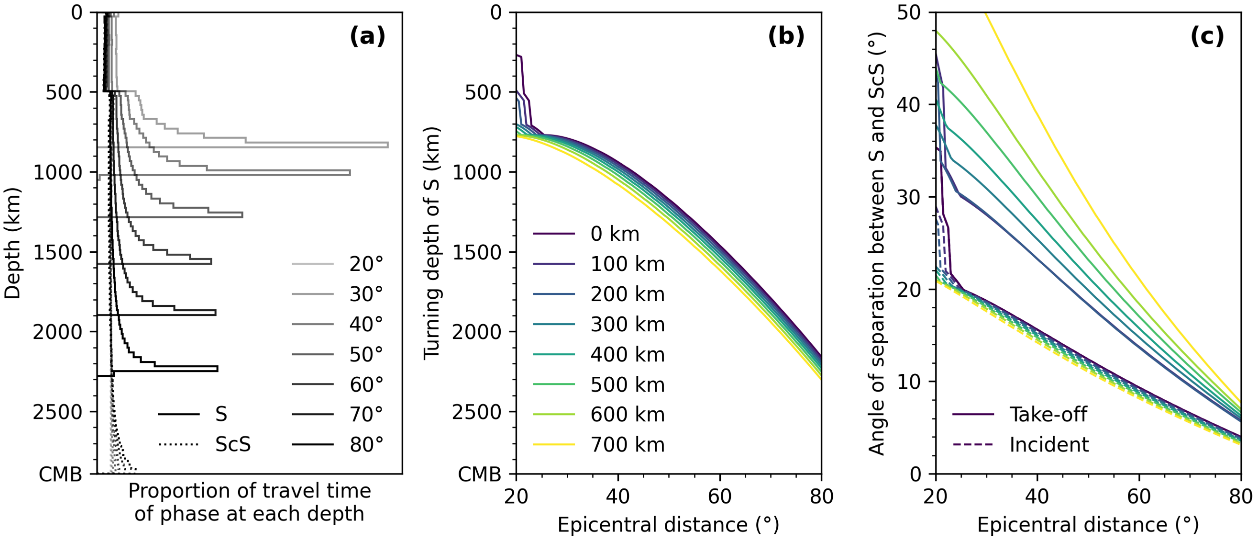

Figure S1: **(a)** The proportion of travel time of each of S and ScS body waves spent at each depth as a function of epicentral distance. **(b)** The turning depth of S waves as a function of epicentral distance. **(c)** The difference in angle between S and ScS raypaths at take-off (from the event) and incident (at the receiver).

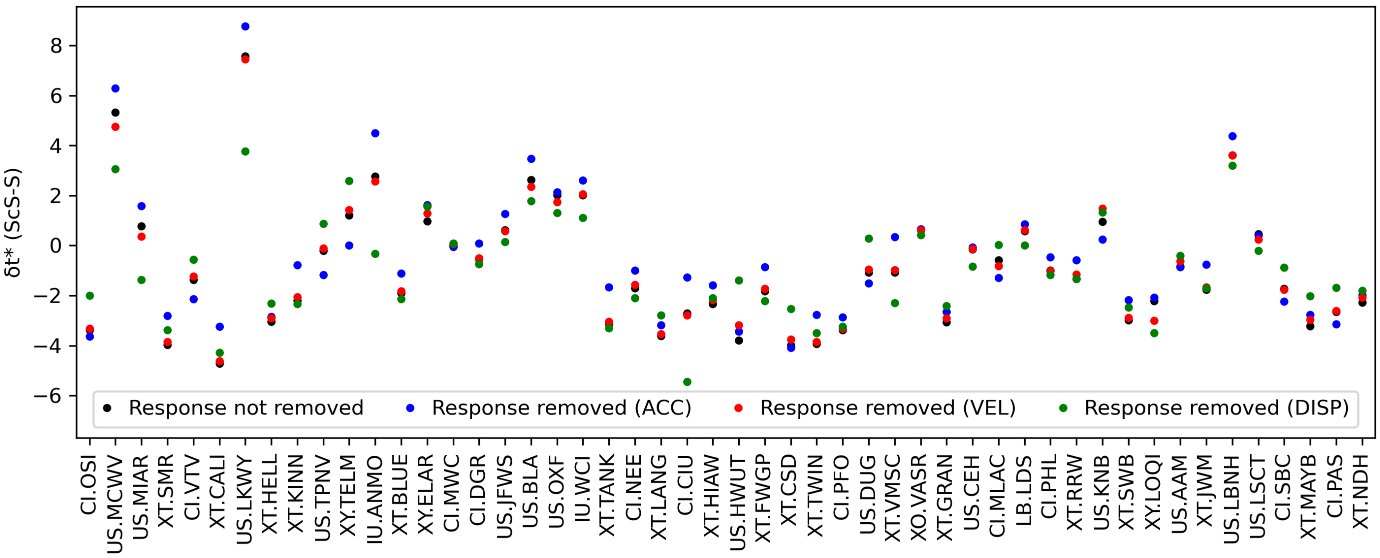

Figure S2: $\delta t^*_{\mathrm{ScS-S}}$ measurements for a selection of 50 stations for the event on 1997/11/28, as used by Ford et al. [2012], for data which has been downloaded. Values are for data with station response not removed (black) and removed and processed as acceleration (blue), velocity (red), and displacement (green).

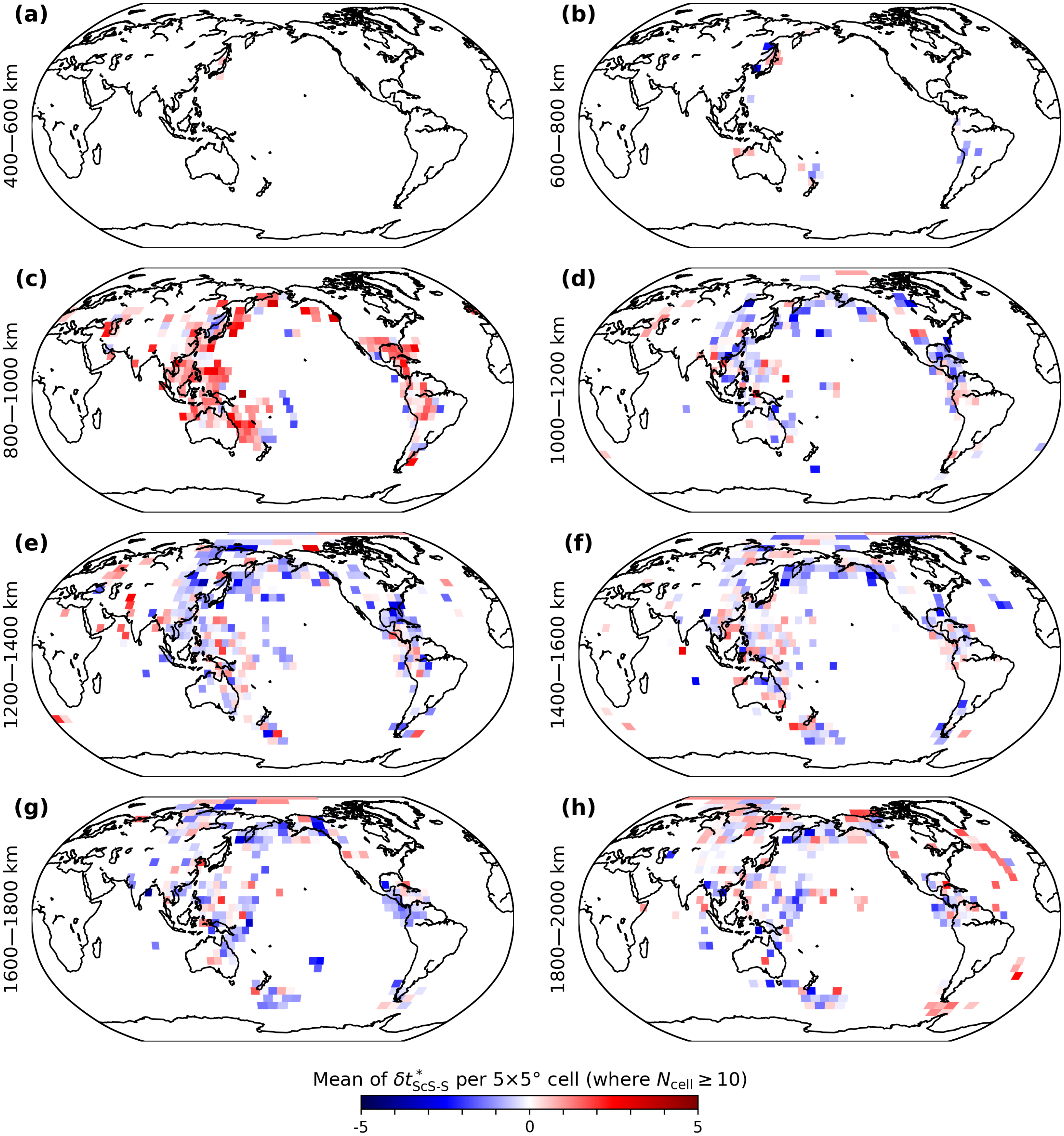

Figure S3: Mean value of $\delta t^*_{\text{ScS-S}}$ per $5 \times 5°$ quasi-equal area cell where grid cells with fewer than 10 ScS bounce points have been omitted, binned according to the turning depth of the S raypath: **(a)** 400–600, **(b)** 600–800, **(c)** 800–1000, **(d)** 1000–1200, **(e)** 1200–1400, **(f)** 1400–1600, **(g)** 1600–1800 and **(h)** 1800–2000 km.

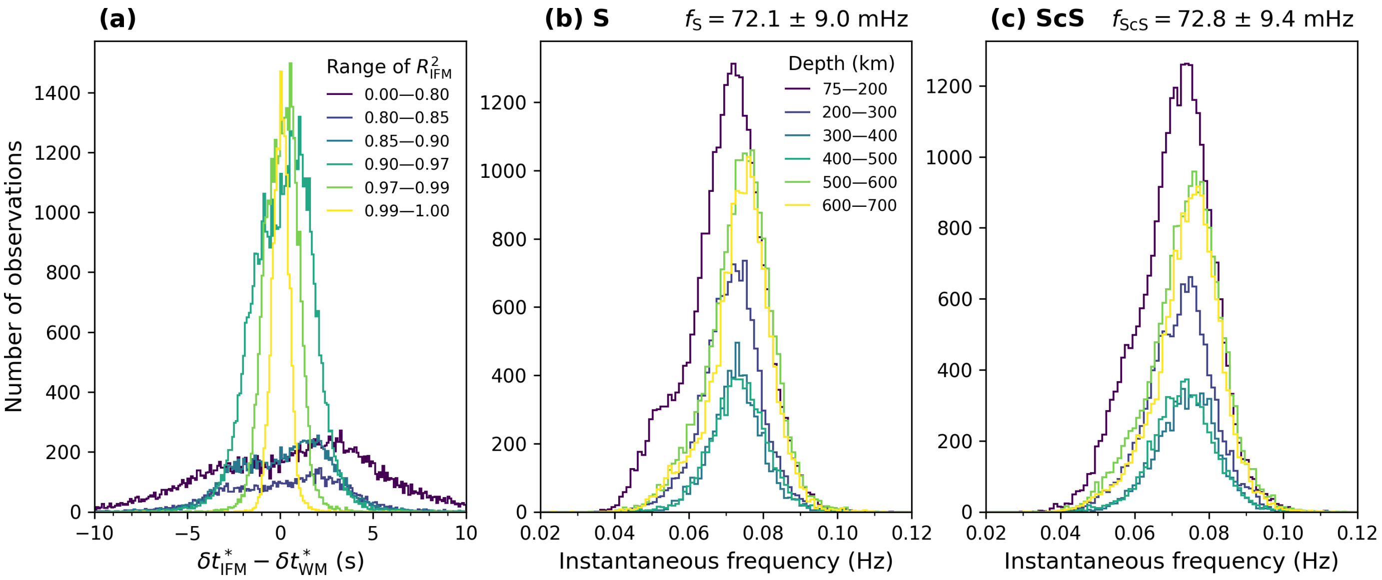

Figure S4: Histograms of **(a)** the difference between measurements of relative attenuation from the instantaneous frequency matching (IFM) and waveform matching (WM) methods, and instantaneous frequency measurements of **(b)** S and **(c)** ScS, binned by event depth. On average, $f_{\mathrm{S}} < f_{\mathrm{ScS}}$, indicating that S is more attenuated than ScS.

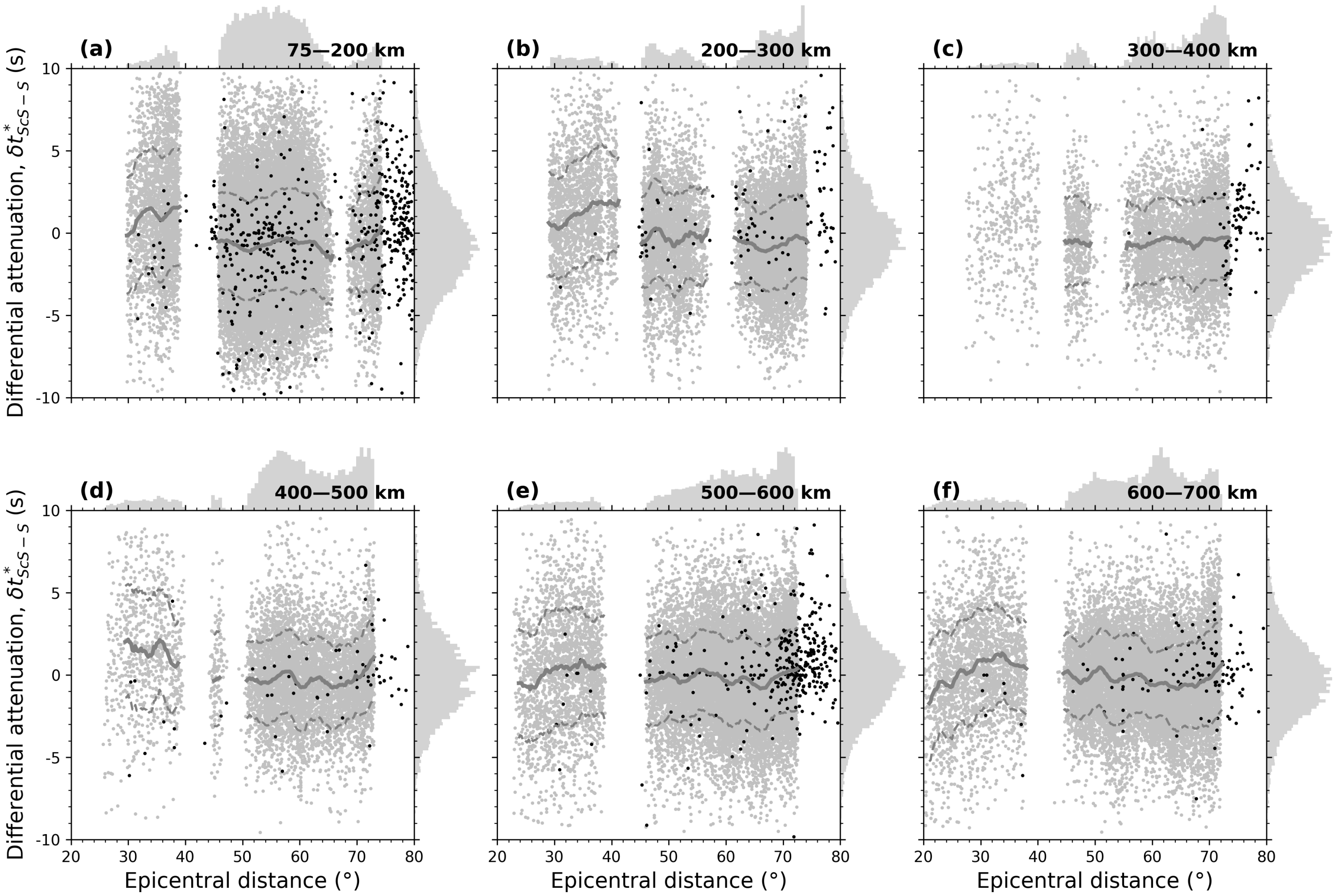

Figure S5: The dataset of differential attenuation $\delta t^*_{\text{ScS-S}}$ measurements reproduced from Figure 5 (grey dots), with constructed measurements of differential attenuation $t^*_{\text{ScS}} - t^*_{\text{S}}$ from the published dataset of Lai et al. [2019] (black dots). Aggregated trends of these measurements for events deeper and shallower than 200 km are plotted in Figure 7 in the main paper (solid and dashed blue lines, respectively).

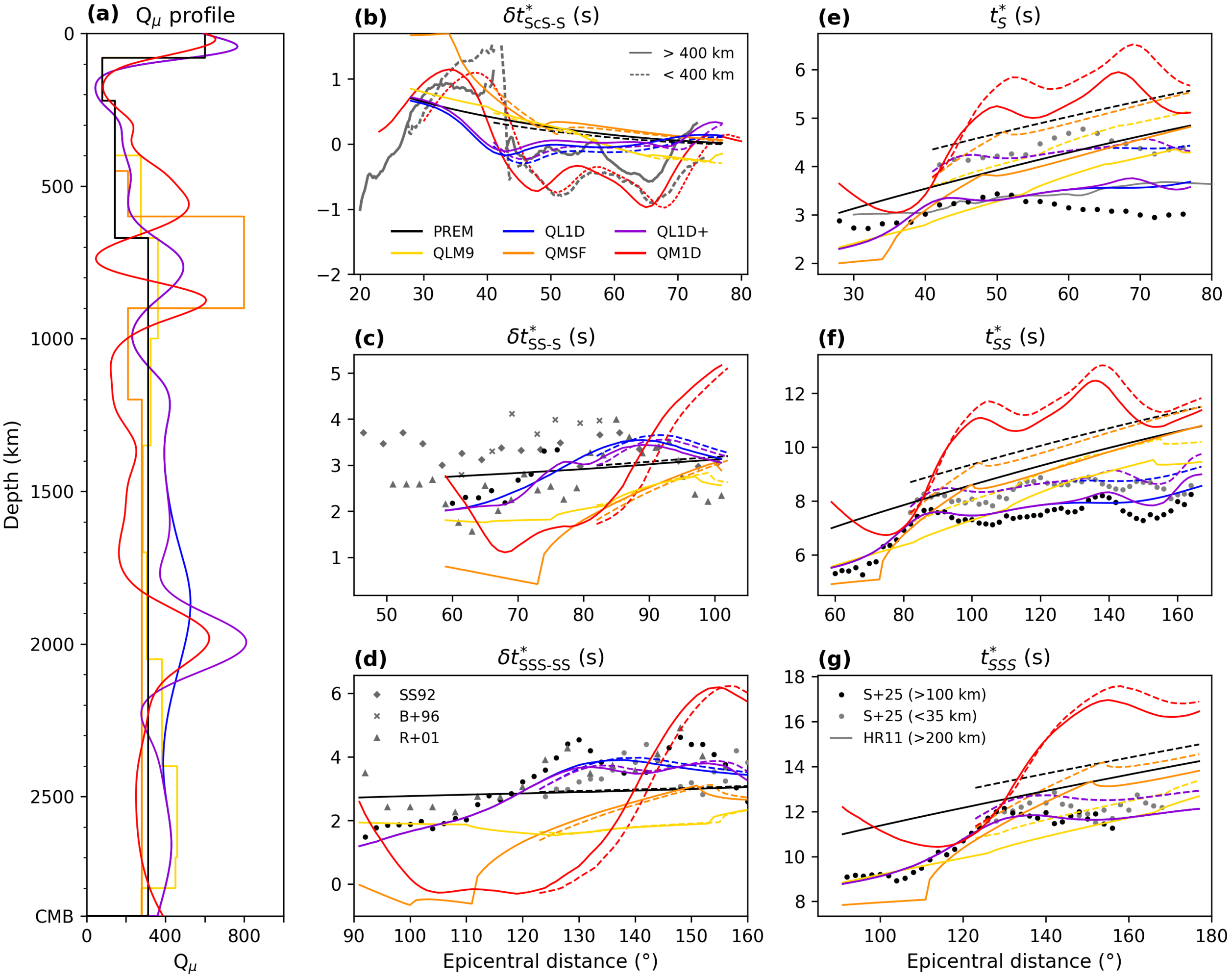

Figure S6: **(a)** Q_μ profiles of PREM (black), QLM9 (yellow), QMSF (orange), Q_{M1D} (red), Q_{L1D} (blue) and a modified version of Q_{L1D} with a Ricker wavelet superimposed at a depth of 2000 km (purple). **(b)** Predictions of relative attenuation $\delta t^*_{\mathrm{ScS\text{-}S}}$ for each of the models overlaid on data from this study, using a 5° rolling window, for deeper ($>$ 400 km, solid) and shallower ($>$ 400 km, dashed) events. Predictions of differential attenuation **(c)** $\delta t^*_{\mathrm{SS\text{-}S}}$ and **(d)** $\delta t^*_{\mathrm{SSS\text{-}SS}}$ and absolute attenuation **(e)** t^*_{S}, **(f)** t^*_{SS} and **(g)** t^*_{SSS} for each of the Q_μ profiles. The mean measurements of $\delta t^*_{\mathrm{SS\text{-}S}}$, $\delta t^*_{\mathrm{SSS\text{-}SS}}$, t^*_{S}, t^*_{SS} and t^*_{SSS} for deep (black dots) and shallow (grey dots) events were constructed from the Sun et al. [2025] dataset. Datasets from other studies are also included for reference where aggregated data was available: t^*_{S} by Hwang and Ritsema [2011] (grey line, note that data has been fixed to match PREM at 30°); $\delta t^*_{\mathrm{SS\text{-}S}}$ by Sheehan and Solomon [1992] (grey crosses) and Bhattacharyya et al. [1996] (grey diamonds); and $\delta t^*_{\mathrm{SS\text{-}S}}$ and $\delta t^*_{\mathrm{SSS\text{-}SS}}$ by Reid et al. [2001] (grey triangles).

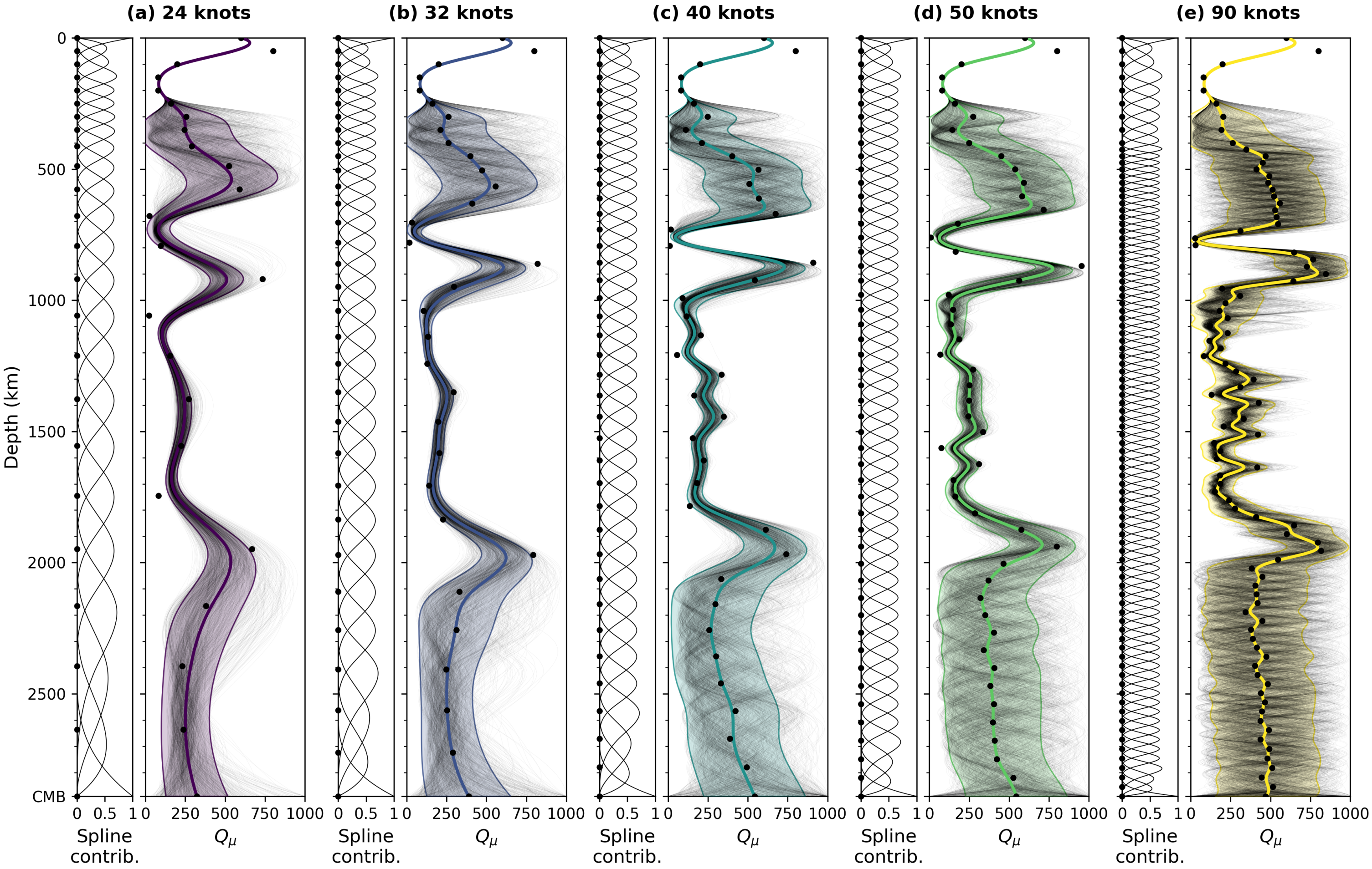

Figure S7: Cubic B-spline basis functions and inversion results for **(a)** 24, **(b)** 32, **(c)** 40, **(d)** 50 and **(e)** 90 depth knots.

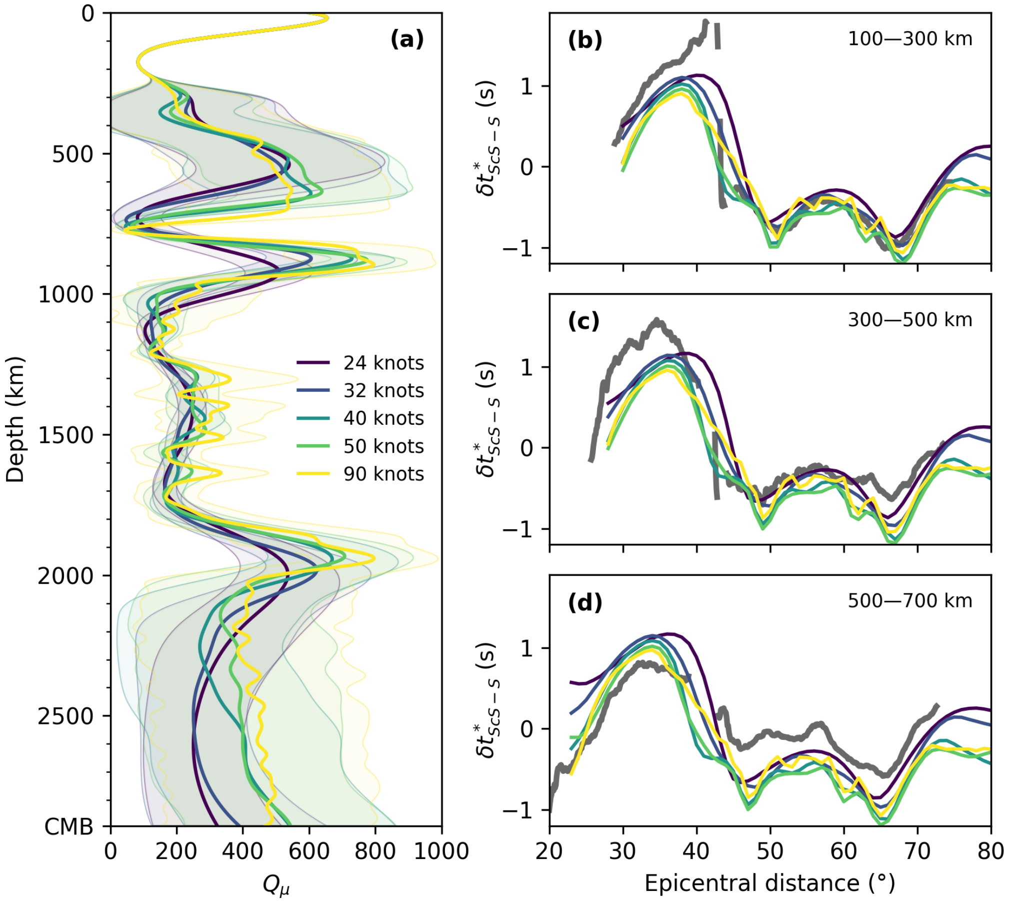

Figure S8: **(a)** Comparison of mean inversion Q_μ profiles for the different numbers of depth knots in Figure S7. Predictions of $\delta t^*_{\text{ScS-S}}$ for each of the models overlaid on data from this study for event depths **(b)** 100–300, **(c)** 300–500 and **(d)** 500–700 km, using a 5° rolling window.

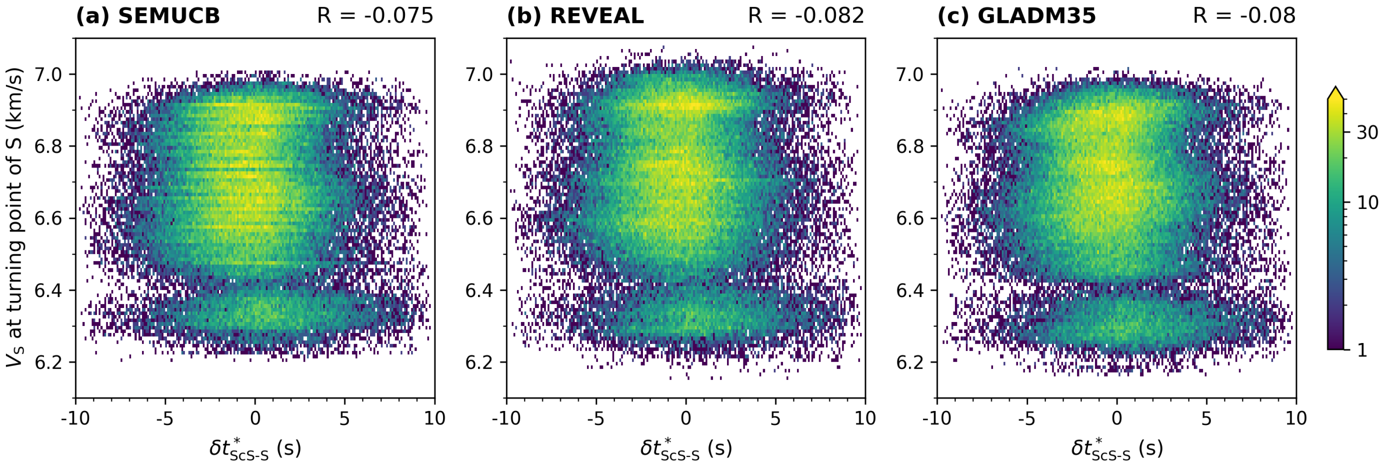

Figure S9: Plots of our measured values of $\delta t^*_{\text{ScS-S}}$ compared to the velocity at the turning depth of the S raypaths for three full waveform tomographic models of the mantle: **(a)** SEMUCB-WM1 [French and Romanowicz, 2014], **(b)** REVEAL [Thrastarson et al., 2024] and **(c)** GLAD-M35 [Cui et al., 2024]. The Pearson correlation coefficient R is computed for each comparison.

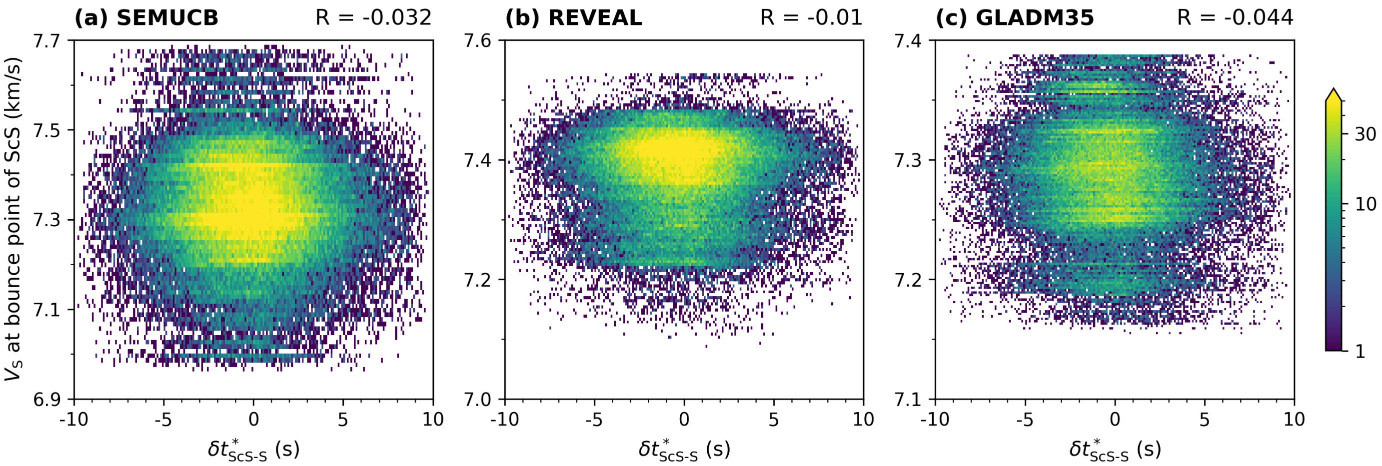

Figure S10: Plots of our measured values of $\delta t^*_{\text{ScS-S}}$ compared to the velocity at the bounce point of the ScS raypaths for three full waveform tomographic models of the mantle: **(a)** SEMUCB-WM1 [French and Romanowicz, 2014], **(b)** REVEAL [Thrastarson et al., 2024] and **(c)** GLAD-M35 [Cui et al., 2024]. The Pearson correlation coefficient R is computed for each comparison.

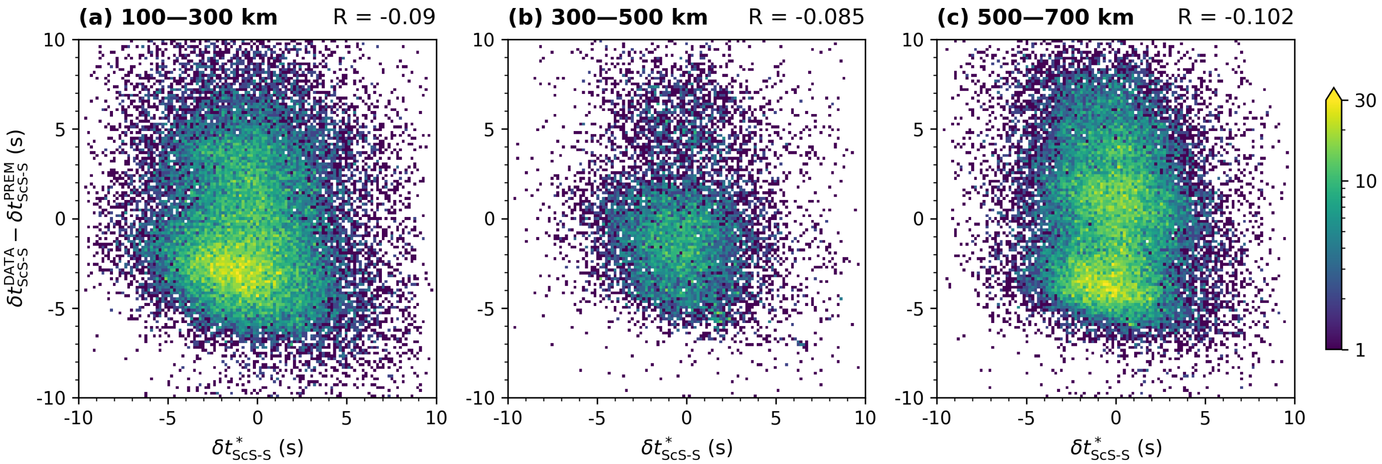

Figure S11: Plots of our measured values of $\delta t^*_{\text{ScS-S}}$ compared to the difference between the observed travel time differences (corrected for attenuation) and predicted travel time differences for PREM $\delta t^{\text{DATA}}_{\text{ScS-S}} - \delta t^{\text{PREM}}_{\text{ScS-S}}$, for the same event depth ranges as Figure S8: **(a)** 100–300, **(b)** 300–500 and **(c)** 500–700 km. The Pearson correlation coefficient R is computed for each comparison.

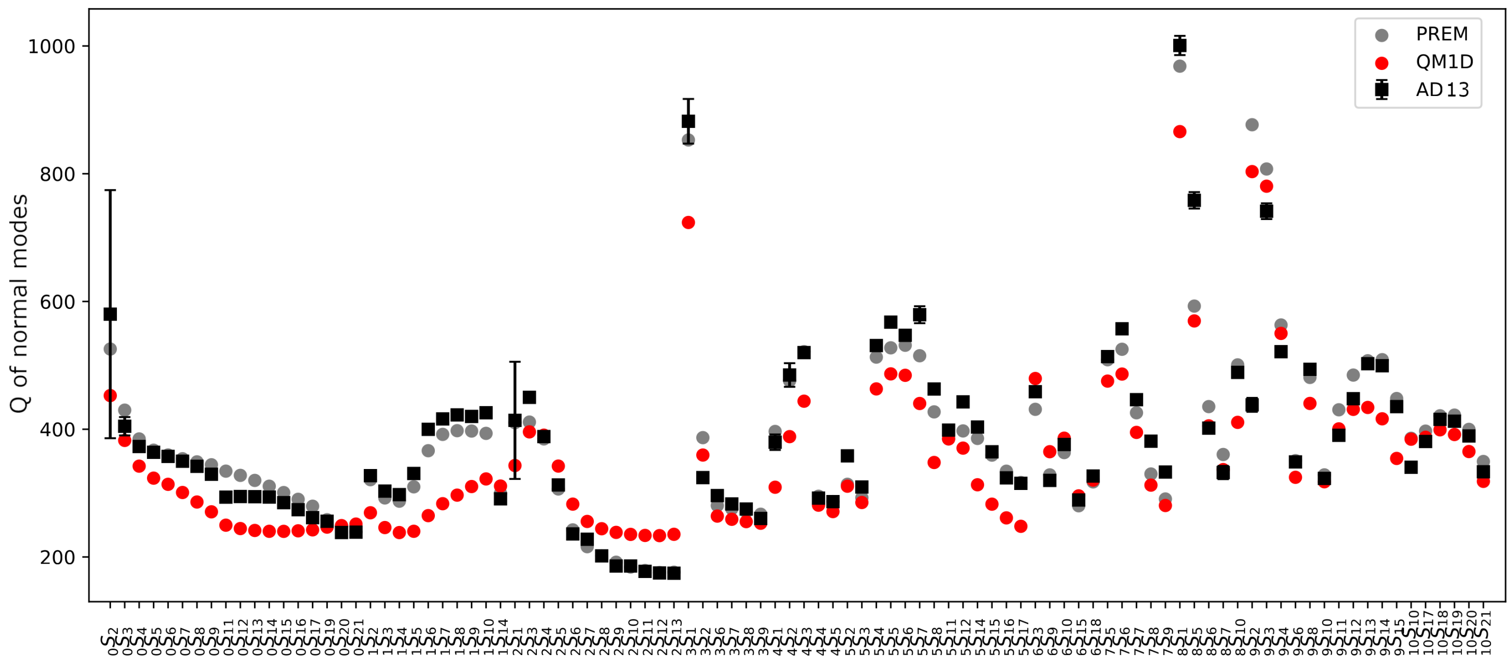

Figure S12: Values of quality factor Q of normal modes for both PREM (grey) and our model Q_{M1D} (red) computed using MINEOS [Woodhouse, 1988; Masters et al., 2011], with all other parameters kept the same as PREM and assuming that $\alpha = 0$. Measured values of Q of normal modes from Deuss et al. [2013] (black) are overlaid.

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