

Q_P and Q_S in the upper mantle beneath the Iberian peninsula from recordings of the very deep Granada earthquake of April 11, 2010

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[1] Granada (Southern Spain) is a place of rare and enigmatic very deep focus earthquakes, the last one on April 11, 2010, with magnitude of 6.3 and depth of 620 km. We use regional broadband recordings to estimate Q_P and Q_S in the mantle for frequencies between 0.25 and 8 Hz, computing the spectra of the direct P- and S-waves with their early P- and S coda. We use the spectral decay method, constraining crustal Q to values given in the literature. We obtain robust estimates of Q_P in 6 frequency bands (0.25, 0.5, 1, 2, 4 and 8 Hz) and of Q_S in 4 bands (0.25, 0.5, 1, 2 Hz). Q_P in the mantle ranges from 13 at 0.25 Hz to 346 at 8 Hz and Q_S from 59 at 0.25 to 183 at 2 Hz. The frequency dependence is well fitted by $Q = Q_0 f^\alpha$ with α equal to 0.6 for Q_S and 1.0 for Q_P , and Q_0 equal to 109 for Q_S and 63 for Q_P . The Q_P/Q_S ratio is less than 1. These are extreme values within the ranges of mantle Q , Q_P/Q_S and α values reported in the literature, indicating strong scattering attenuation and absence of melt. We propose that such values, rather than being an exception, may approximate the average upper mantle, with solid olivine composition and small-scale heterogeneity. **Citation:** Mancilla, F., E. Del Pezzo, D. Stich, J. Morales, J. Ibañez, and F. Bianco (2012), Q_P and Q_S in the upper mantle beneath the Iberian peninsula from recordings of the very deep Granada earthquake of April 11, 2010, *Geophys. Res. Lett.*, 39, L09303, doi:10.1029/2012GL050947.

1. Introduction

[2] In 1954, an unexpected deep ($h = 630$ km) earthquake [$M_w = 7.9$] beneath Southern Spain pointed out the existence of deep seismic activity far from any active subduction boundary known at that time [Hodgson and Cock, 1956; Chung and Kanamori, 1976]. Together with the absence of subsequent seismic activity in the same source zone, this quake is recognized as the most peculiar deep earthquake of the last century [Frohlich, 2006]. Three small deep-focus earthquakes ($\sim 4 < M < 5$) followed this event in 1973, 1990 and 1993 [Buforn *et al.*, 2004] with hypocenters very close to the 1954 event at less than 25 km epicentral distance. On

April 11th, 2010, a larger deep earthquake [$h = 620$ km; $M_w = 6.3$] occurred, just in the same source region. Different from the 1954 event, the 2010 earthquake was recorded at a dense regional network of seismic broadband stations, providing an unprecedented amount of on-scale high quality seismograms. Hereafter we denote this earthquake with the acronym DSE.

[3] The scarcity and spatial isolation of Southern Spain deep earthquakes led to the hypothesis of a detached piece of lithosphere sinking into the mantle [e.g., Buforn *et al.*, 1997]. Early tomographic images [Blanco and Spakman, 1993] show a high velocity body between 250 and 650 km depth, suggesting a scenario of slab detachment. More recently, Spakman and Wortel [2004] describe the 3-D geometry of a continuous slab extending to 660 km depth, supporting models of westward slab roll-back in the Betic-Rif-Alboran region [Lonergan and White, 1997; Faccenna *et al.*, 2004; Jolivet *et al.*, 2008]. Ongoing eastward subduction has been proposed by Gutscher *et al.* [2002], however no signature of such a process shows up in the regional seismotectonic patterns and the GPS velocity field [Stich *et al.*, 2005, 2006]. Other subduction geometries have been proposed by other authors to explain the formation of the Gibraltar arc and the occurrence of deep seismicity [e.g., Zeck, 1996], as well as entirely different geodynamic models, ranging from convective removal of the lithospheric root [Platt and Visser, 1989] to delamination of the upper mantle [Seber *et al.*, 1996; Calvert *et al.*, 2000].

[4] All these models have in common the very limited spatial extent of the lithosphere recycling processes beneath the Betic-Rif chain and the Alboran Sea in line with the tight concentration of very deep hypocenters. This represents a fundamental difference to the typical locations of very deep earthquakes in much larger and more active slabs. In particular, the deep Spanish earthquakes provide an interesting opportunity to study upper mantle Q from regional P- and S-wave recordings. Using deep regional earthquakes to study upper mantle attenuation allows for extending attenuation measurements to higher frequency (>1 Hz) compared to teleseismic body waves (<1 Hz), and surface waves (<0.05 Hz). At those high frequencies, effects as scattering, focussing/defocussing or 3D propagation should become more important than or comparable to intrinsic attenuation. While in a typical subduction setting, such waves often carry a significant imprint of the mantle wedge or along-slab propagation, the limited horizontal extension of the upper mantle high-velocity anomaly beneath southern Spain [Spakman and Wortel, 2004] suggests that here the inferred Q values should be more representative of an average upper mantle outside the subduction system. In particular, the

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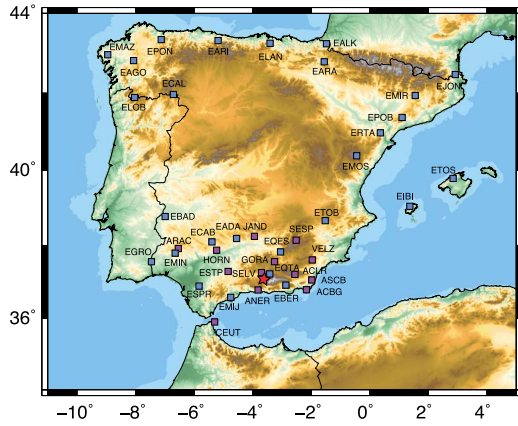


Figure 1. Map of the Iberian peninsula showing the station locations (blue squares for stations sampling at 100 sps and purple squares for stations sampling at 50 sps), and the epicenter of the 2010 very deep Granada earthquake (star).

stations north of the epicenter are not expected to sample the subduction zone (see Figure 1). Except for *Okal* [2001], who reports a Q_μ of 479 between 1 Hz and 3 Hz using a single recording of the 1993 deep Spanish earthquake, there are no previous studies of the attenuation parameters of the

upper mantle beneath Spain. We will use P and S waves data generated by the DSE, to contribute in defining the seismic space-averaged total attenuation in the upper mantle beneath Iberia, for useful comparisons with subduction zones around the world.

2. Data and Method

[5] We analyse seismograms of DSE at 47 broad-band 3-components seismic stations (Figure 1), managed by the Instituto Andaluz de Geofísica (IAG, University of Granada) and Instituto Geográfico Nacional (IGN). Epicentral distances range from 29 km to 877 km. Seismograms are sampled at 50 sps for IAG data and 100 sps for IGN data. Examples of the seismic traces utilized in this paper are reported in Figure 2. High frequency (>2 Hz) waves in the early coda of S-waves result to be contaminated by the later P-coda radiation, and thus excluded from the analysis.

[6] We measure the decay of spectral amplitude of the ground velocity (for P and S waves) with source-to-station distance [Sato and Fehler, 1998, pp. 110–111], which, as well known, depends on the total-Q (Q_P and Q_S for P and S waves, respectively). The spectral amplitude is estimated through Fast Fourier Transform. The amplitude spectral density is calculated for time-windows 60 seconds long, starting respectively at the P- and S- wave onset. We use spectra of vertical component for P-waves and log-averaged

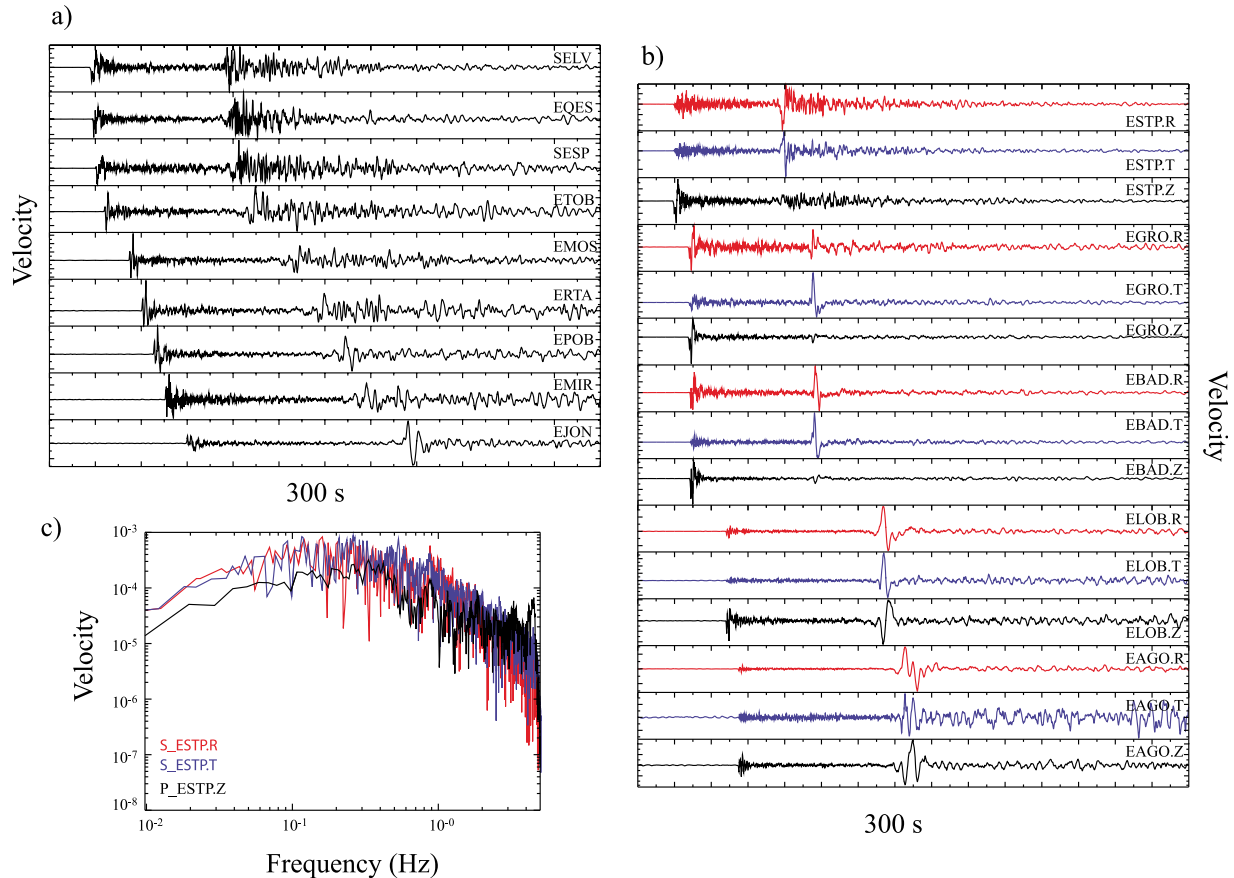


Figure 2. Data examples: (a) vertical components of velocity seismograms for stations along a profile toward the NE; (b) 3-components velocity seismograms along a profile toward the NW; (c) velocity spectrum of the P-wave measure in Z component, and for the S-wave in R and T components for the station ESTP. (See Figure 1 for station locations.)

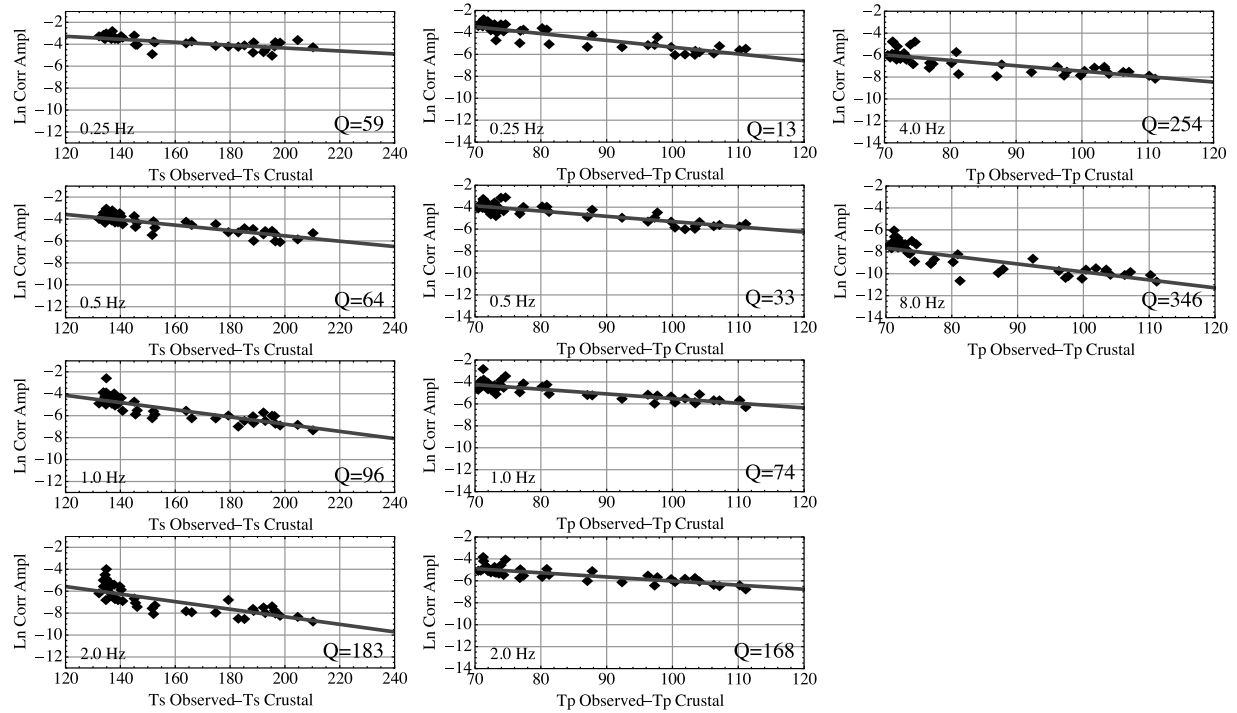


Figure 3. $Q_{P,S}^M$ estimates for P and S waves at each frequency band, using the spectral decay method. Observations are marked by squares.

spectra of the two horizontal components for S-waves. Pre-event noise spectra are calculated and compared with signal spectra. Signal-to-noise ratio results always greater than 5 at 8 Hz, and much greater than this value for decreasing frequency. The spectral Log-average is evaluated over the horizontal components and finally the integral of the spectral amplitude is calculated in 6 frequency bands, respectively from 0 to 0.35, from 0.35 to 0.7, from 0.7 to 1.4, from 1.4 to 2.8, from 2.8 to 5.6 and from 5.6 to 12 Hz. Central frequencies for each band are 0.25, 0.5, 1.0, 2.0, 4.0 and 8.0 Hz. To estimate the contribution of the seismic attenuation in the crust, we use total-Q for S-waves in southern Spain calculated by *Akinci et al.* [1995] ($Q_S^C = 34.8f^{1.0}$). We assume that total-Q for P-waves is a half of total-Q for S-waves in the crust, an average of the experimental values through the world [*Sato and Fehler*, 1998, Figure 5.3], and use this estimate as a constraint for calculating the correspondent quality factors, Q_P^M and Q_S^M in the mantle. The equation which associates the integral of the spectral amplitude for P and S waves (as a function of frequency $S_{P,S}(f)$) to the quality factor is

$$S_{P,S}(f, T_{P,S}^M, T_{P,S}^C) = K \cdot (T_{P,S}^M + T_{P,S}^C)^{-1} \cdot \exp(-\pi f T_{P,S}^M / Q_{P,S}^M) \cdot \exp(-\pi f T_{P,S}^C / Q_{P,S}^C) \quad (1)$$

where K includes source-, site-, instrument-transfer functions and velocity model; T is the travel time (M and C indicate respectively mantle and crust and subscripts P and S respectively compressional and shear waves). The left hand side of this equation represents the measured quantities as a function of travel time (calculated through ray-tracing). Equation (1) for all the stations, and for P

and S waves, represents a largely overdetermined system of N equations for 2 unknowns (K and $Q_{P,S}^M$) that can be solved, after Log-linearization, with an optimization algorithm. We use here the weighted least squares approach, described by the following matrix equation for any value of the frequency, f :

$$\mathbf{m}(f) = (\mathbf{G}(f)^T \cdot \mathbf{W} \cdot \mathbf{G}(f))^{-1} \cdot \mathbf{G}(f)^T \cdot \mathbf{W} \cdot \mathbf{d}(f) \quad (2)$$

where $\mathbf{m}(f)$ is the vector containing the unknowns, $\mathbf{G}(f)$ is the 2-column coefficient matrix (first column elements = 1; second columns elements = $-\pi f (T_{i,P,S}^M)$ where $T_{i,P,S}^M$ is the travel time in the mantle at the i -th receiver). $\mathbf{W}$ is the matrix containing the weights. After many tests for different weighting matrices, we observe that unweighted- are comparable to weighted-solutions and selected $\mathbf{W}$ = unity. $\mathbf{d}(f)$ is the data matrix, whose elements ($i = 1, \dots, N$) are

$$d(f)_{i,P,S} = \text{Log}(S_{i,P,S}(f)) + \text{Log} \left[(T_{P,S}^C + T_{P,S}^M) + \pi f T_{P,S}^C / Q_{crust} \right] \quad (3)$$

We solved system (2) for each frequency, separately for P and S. Errors on the estimates of Q_P and Q_S are obtained using as Covariance matrix the diagonal matrix associated with the variances in the spectral estimates.

3. Results and Discussion

[7] The importance of the physical properties on the seismic attenuation parameters have been demonstrated in experimental studies applied to environmental and petrological conditions of the upper mantle [e.g., *Jackson et al.*,

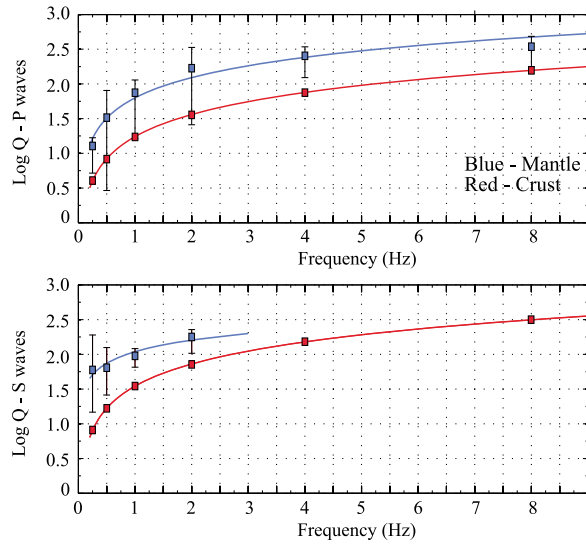


Figure 4. Pattern of (top) Q_P and (bottom) Q_S as a function of frequency. With blue squares, the estimates calculated in the present study for $Q_{P,S}$ in the mantle together with their error bars and the blue line is the obtained fitting function ($Q = Q_0 f^\alpha$). In red the values of $Q_{P,S}$ in the crust used for the crustal correction in the calculations [Akinci et al., 1995].

2002; Faul and Jackson, 2005; Aizawa et al., 2008; Faul et al., 2004]. Parameters as temperature [Jackson et al., 2002; Faul and Jackson, 2005], partial melting [Faul et al., 2004] and the water content [Aizawa et al., 2008] play a crucial role in the attenuation of the seismic waves propagating in the upper mantle. The Q_P^M calculated in the frequency range between 0.25 and 8 Hz and the Q_S^M values for frequencies from 0.25 to 2 Hz (Figure 3) are low ($Q_P^M = 74$ and $Q_S^M = 96$ at 1 Hz), similar to the quality factors measured in high attenuation crustal domains. Low Q_P and Q_S values in the upper mantle are commonly associated with the mantle wedge in active subduction zones, instead of the cold oceanic subducting lithosphere where substantially higher Q values are found [e.g., Wiens et al., 2008]. Q_P^M and Q_S^M for Southern Spain are similar to those measured in the mantle wedge of subduction zones like Andes [Schurr et al., 2003; Badi et al., 2009], Marianas [Pozgay et al., 2009], Philippines sea-Izu Bonin subduction [Shito and Shibutan, 2003], Japan arc [Tsumura et al., 2000] and Costa Rica and Nicaragua [Rychert et al., 2008].

[8] From the results displayed in Figure 3, we obtain a frequency dependence well expressed by $Q_{P,S}^M = Q_0 f^\alpha$. Q_0 is Q at 1 Hz and α the dependence with the frequency of Q . $Q_{P,S}^M$ and α has been estimated using least squares minimization. The results show that for Q_P^M , $Q_0 = 63 \pm 1$ and $\alpha = 1.0 \pm 0.1$ (correlation coefficient = 0.98) taking into account the whole frequency range (Figure 4). The restricted frequency range available for Q_S^M makes the estimate of the Q_S^M frequency dependence less stable than that for Q_P^M . Results for Q_S^M give $Q_0 = 109 \pm 1$ and $\alpha = 0.6 \pm 0.1$ (correlation coefficient = 0.95, Figure 4). The above results for α are higher than the estimates obtained by Shito et al. [2004] for the upper mantle beneath Japan and at the upper bound of values reported by Cheng and Kennett [2002]. From lab-experiment on solid olivine samples, $\alpha = 0.2 - 0.4$ [e.g., Jackson et al., 2002, 2004], tending to 0 when partial

melting is present [Faul et al., 2004]. Higher α values suggest a solid olivine composition of the mantle path above the source in Southern Spain and the presence of strong scattering attenuation [e.g., Shito and Shibutan, 2003].

[9] Q_P^M/Q_S^M ratio calculated in the present paper in the range between 0.25 Hz and 2 Hz is less than 1.0 (0.21, 0.51, 0.77 and 0.91 at 0.25 Hz, 0.5 Hz, 1.0 Hz and 2.0 Hz, respectively). Taking into account that Q_P/Q_S for a Poisson solid with null bulk attenuation (Q_K^{-1}) is close to 2.25 [Jackson, 2007], the present result showing significantly smaller ratio ($Q_P^M/Q_S^M < 1.0$) suggests the presence of melt and/or scattering attenuation in the mantle beneath Spain [Budiansky et al., 1983]. Q_P^M/Q_S^M values found in Southern Spain are lower than: a) those calculated in Central Alaska by Stachnik et al. [2004] ($Q_P/Q_S = 1.1 - 1.4$), b) those calculated in Tonga-Fiji region [Roth et al., 1999] (Q_P/Q_S values between 1.7 and 2.2), c) those calculated in the Philippines mantle wedge [Shito and Shibutan, 2003], and at the lower bound of values reported for the Marianas subduction system ($\sim 0.8 - 2.1$ [Pozgay et al., 2009]).

[10] To support our initial hypothesis that our estimates of $Q_{P,S}^M$, Q_P^M/Q_S^M and α approximate average mantle conditions, we repeat the calculations excluding several stations that potentially may register waves propagating through the high velocity anomaly. Therefore we exclude seven stations at short distances that show the largest positive residuals in the regression for Q_S^M at 2 Hz (Figure 3). For this data set, we found Q_P^M in the mantle ranging from 13 at 0.25 Hz to 346 at 8 Hz and Q_S^M from 59 at 0.25 to 183 at 2 Hz. α becomes 0.7 for Q_S^M and remains 1.0 for Q_P^M , while Q_P^M/Q_S^M ratios become 0.20, 0.47, 0.66 and 0.76 at 0.25 Hz, 0.5 Hz, 1.0 Hz and 2.0 Hz, respectively. These estimates are very close to the results for the full data set, suggesting that the average values are not significantly contaminated by propagation through the subduction system.

4. Conclusions

[11] In the present paper we estimate attenuation parameters ($Q_{P,S}^M$) for the upper mantle beneath the Iberian peninsula using the spectral decay method for direct P and S waves in the frequency range of 0.25 to 8 Hz by using a complete and dense data set from broad band seismograms of the 11 April 2010 deep ($h = 620$ km) Spanish Mw = 6.3 earthquake. We found a highly attenuative mantle with very low values of $Q_{P,S}^M$ in the frequency range analyzed, similar to those measured in active volcanic arcs. A large frequency dependence of the quality factor is observed, with $\alpha \sim 1.0$ for Q_P^M and ~ 0.6 for Q_S^M . Q_P^M/Q_S^M ratios are remarkably low (< 1.0). Our estimates are close to the extremes of the range of previously reported $Q_{P,S}^M$, Q_P^M/Q_S^M and α for the upper mantle.

[12] The combination of small Q_P^M/Q_S^M and large α favours an interpretation in terms of scattering attenuation as the dominant mechanism, compared to a classic mantle wedge setting, where melting processes, thermal perturbations and an environment rich in fluids should translate into much smaller values of $\alpha \sim 0$. Our data suggest a solid, randomly heterogeneous propagation medium producing high attenuation (low values of Q_P^M and Q_S^M). We propose that such conditions may be characteristic of the upper mantle outside subduction systems.

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