

*Short Notes**Lg Q* in the Eastern Tibetan Plateau

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Abstract *Lg* spectra are collected from the 1991–1992 Tibetan Plateau Passive Experiment to measure *Q* values. Using a standard two-station method that virtually eliminates source and site effects, I obtain a model of $Q_0 = (126 \pm 9)$ and $\eta = (0.37 \pm 0.02)$ in a frequency range between 0.2 and 3.6 Hz, where Q_0 and η are *Lg Q* at 1 Hz and its power-law frequency dependence, respectively. The estimated Q_0 value is among the lowest ever reported for continental areas; it qualitatively supports the observation by McNamara *et al.* (1996) that *Lg* can not be observed inside the plateau beyond about 700 km, a limiting distance that is much shorter than those in the other low Q_0 (~ 200) regions, such as Iran and the western United States. The low Q_0 value may be the cause of *Lg* blockage across the northern boundary of the plateau and may indicate abnormally high temperature and fluid content in the Tibetan crust.

Quantitatively, the estimated Q_0 value is lower by a factor of 3 than the value of 366 estimated by McNamara *et al.* (1996), who used data from the same experiment. Since there are several differences in the data processing and inversion procedures used in this and the previous studies, I investigated the effects of these differences on the *Q* estimates. I conclude that the most probable cause of the discrepancy is in the different inverse methods used. The previous inversion solved for a large number of free parameters that include the source and site terms. In this study only two free parameters (Q_0 and η) are solved for, thus avoiding the instability caused by parameter trade-offs.

Introduction

The seismic *Lg* wave can be treated as multiple supercritical *S*-wave reflections or many overtone surface waves traveling in the continental crust. The attenuation rate, or *Q*, of *Lg* generally correlates with the tectonic environment. The *Lg Q* has often been observed to fit a power-law frequency dependence:

$$Q_{Lg}(f) = Q_0 f^\eta, \quad (1)$$

where Q_0 and η are *Q* at 1 Hz and its power-law frequency dependence, respectively. *Lg Q*₀ values in tectonically active regions are typically lower than about 300. For example, *Lg Q*₀ values were estimated to be about 150 in California (Herrmann, 1980; Nuttli, 1986), 200 in Iran (Nuttli, 1980), and 200 to 267 for the various regions in the western United States (Xie and Mitchell, 1990; Xie, 1998). These values are generally about a factor of 2 to 5 lower than the values found in the stable central and eastern United States.

The highest *Lg* attenuation rates ever documented are those observed in the Tibetan Plateau. Ruzaijin *et al.* (1977) and Ni and Barazangi (1983) reported that, on short-period

seismograms observed over paths crossing the northern and southern boundaries of the plateau, the *Lg* wave is either abnormally weak, or absent. These phenomenon are qualitatively known as partial or complete blockage of *Lg* by regions along, or behind, these boundaries. McNamara *et al.* (1996) used seismic data from a portable network deployed inside the plateau to study *Lg* attenuation. They found that for paths that lie completely inside the plateau, the *Lg* phase could be observed out to a limiting distance of about 600 to 700 km, beyond which *Lg* was absent owing to a high attenuation, or low *Q*. They also quantitatively fitted the *Lg* amplitudes observed over paths in eastern Tibet, with a $Q_{Lg}(f)$ model of $Q_0 = 366 \pm 37$ and $\eta = 0.45 \pm 0.06$. A puzzling aspect of that model is that the estimated Q_0 value is considerably higher than the value of about 200 estimated for Iran and the values below 267 for the western United States, whereas the 1-Hz *Lg* signal can be observed at distances out to 1400 km in Iran (Nuttli, 1980) and out to more than 2000 km in the western United States (Xie and Mitchell, 1990; Xie, 1998). In this article, I report a new analysis of the *Lg* data used by McNamara *et al.* and a resulting $Q_{Lg}(f)$

model that is characterized by a much lower Q_0 value for eastern Tibet.

Data Processing

The raw waveform data used in this study are the same as those used by McNamara *et al.* (1996) and are collected from 11 broadband PASSCAL seismic stations deployed during the 1991–1992, passive Tibetan Plateau experiment. Detailed specification of the stations and seismic events that were recorded during this experiment are given by McNamara *et al.* (1996). Station locations are also shown in Figure 1 of this article. The data selection and processing procedures in this study somewhat differ from those used by McNamara *et al.* (1996). The data processing procedure used by Xie and Mitchell (1990) and Xie (1998) is used in this study to calculate the amplitude spectra from the vertical component L_g waveforms. Specifically, a 20% cosine taper

window is used in the Fast Fourier Transform (FFT) to obtain L_g spectra. The two corners of the window are located around group velocities of 3.0 and 3.5 km/sec, respectively, and are allowed to be slightly adjustable to best isolate the L_g waveform. A noise reduction procedure, in which the L_g power spectra is subtracted by a moving window average of the pre- P noise spectra, is applied. Only those spectral estimates with signal-to-noise (S/N) ratio greater than 2.0 are kept for further analysis. The original purpose of the spectral analysis in this study was to find L_g source spectral parameters of the events recorded. During an initial analysis of the L_g spectra from only a few events, it was found that the interstation spectral decay rate of L_g could not be fit by a Q_0 near 366. Rather, a Q_0 of no higher than 150 was required by the subset of L_g spectra. Subsequently, a systematic effort was undertaken to obtain all L_g amplitude spectra from the 11 stations and events that were shown in figure 2 of McNamara *et al.* (1996). It is found that there is a fairly large

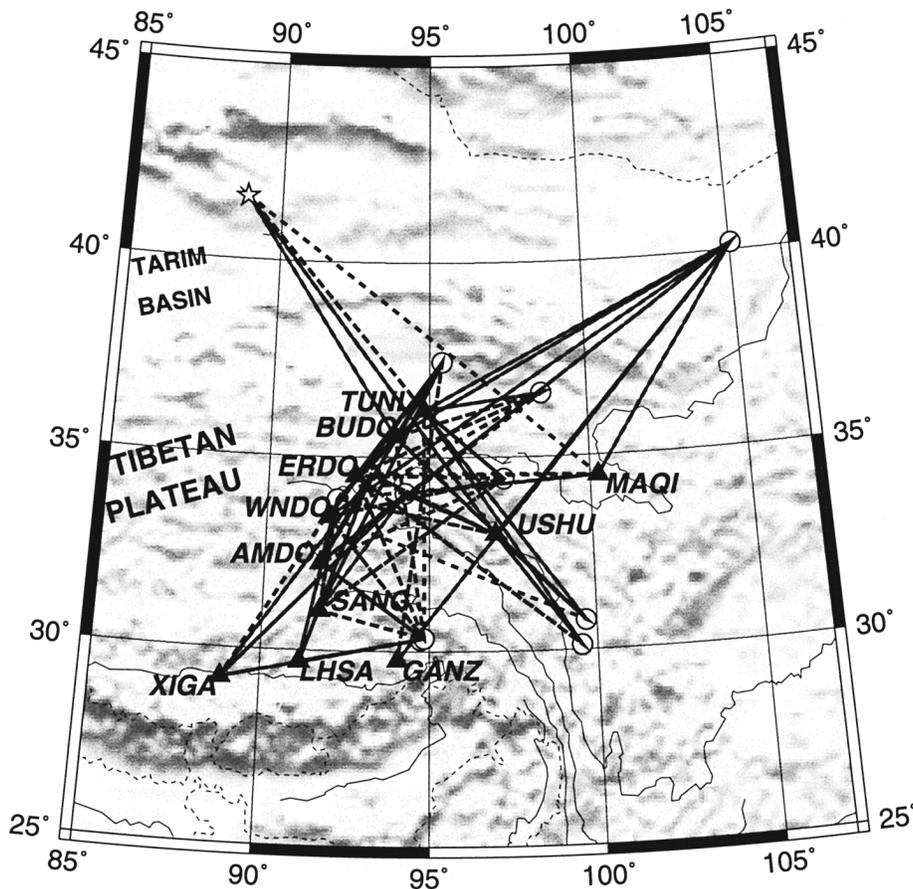


Figure 1. Locations of the PASSCAL stations deployed during the 1991–1992 passive Tibetan Plateau experiment (solid triangles), earthquakes (open circles), and explosion (star) used in this study. Solid paths are those satisfying a $(\delta\theta)_{\max}$ (the maximum allowable difference between the event-to-station azimuths of two stations; see text) of 12° when two-station pairs are selected. Dashed paths are those satisfying a $(\delta\theta)_{\max}$ of 30° . More information of the stations and events can be found in McNamara *et al.* (1996). Note the similarity of the 74 paths plotted here and the 106 paths plotted in figure 7 of McNamara *et al.* (1996).

number (37) of two-station pairs that recorded the same events along about the same great-circle paths. This permitted an areal averaged $Q_{Lg}(f)$ model to be developed using the two-station spectral ratio method, which is described in the next section.

Inverse Method

The two-station spectral ratio method is commonly used for measuring interstation phase delay or amplitude decay of seismic waves (e.g., Aki and Richards, 1980). A version of this method to measure interstation $Q_{Lg}(f)$ was given by Xie and Mitchell (1990). In this method the source effects in the recorded Lg spectra are cancelled by taking ratio of the spectra from a pair of stations that are aligned along the same great circle path from the source. Spectral ratios from many combinations of the two-station pairs are then averaged to suppress the station site responses. The averaged (stacked) spectral ratio is then used to measure path Q , with a minimal error caused by the source and site effects. In the following paragraphs, we present the method and its potential errors more rigorously following Xie and Mitchell (1990), with minor adaptations.

The $Q_{Lg}(f)$ in the study area is parameterized as being laterally homogeneous. When there is an i th event recorded by two stations, $j1$ and $j2$, along the same great circle path, we denote the Lg amplitude spectra recorded at the l th discrete frequency and station j as $A_l^{i,j}$ ($j = j1, j2$), and define a scaled logarithmic spectral ratio

$$D_l^{i,j1,j2} = \frac{V_{Lg}}{\pi(R^{i,j2} - R^{i,j1})} \ln \left[\frac{\sqrt{R^{i,j1}} A_l^{i,j1}}{\sqrt{R^{i,j2}} A_l^{i,j2}} \right] \quad (2)$$

where Vg is the Lg group velocity and $R^{i,j1}$ and $R^{i,j2}$ are the epicentral distances. For each of the station pairs ($j1, j2$), we denote the number of events that are recorded along the same great circles as $I(j1, j2)$. We average all available $D_l^{i,j1,j2}$ to obtain the stacked spectral ratio (SSR):

$$D_l = \frac{1}{N_d} \sum_{j1,j2} \left(\sum_{i=1}^{I(j1,j2)} D_l^{i,j1,j2} \right), \quad (3)$$

where

$$N_d = \sum_{j1,j2} I(j1, j2) \quad (4)$$

is the total number of available $D_l^{i,j1,j2}$ at the l th frequency. Xie and Mitchell (1990) show that the mathematical expectation of D_l is given by

$$E(D_l) = \frac{f_l}{Q_{Lg}(f_l)} + \frac{V_{Lg}}{N_d} \sum_{j1,j2} \frac{1}{\pi \Delta^{j1,j2}} \left[\sum_{i=1}^{I(j1,j2)} \ln \left(\frac{C_l^{j1}}{C_l^{j2}} \right) \right], \quad (5)$$

where the term C_l^j ($j = j1, j2$) describes the station site response.

Ignoring the second term in the above equation (e.g., assuming C_l^j are all unity), D_l can be used to fit a $Q_{Lg}(f)$ model:

$$\ln(D_l) = -\ln Q_0 + (1-\eta) \ln(f_l), \quad (6)$$

where the power-law frequency dependence (equation 1) has been assumed. In general, ignoring the second term of equation (5) subjects the estimated $Q_{Lg}(f_l)$ to a systematic bias

$$\delta \left(\frac{1}{Q_{Lg}(f_l)} \right) = \frac{V_{Lg}}{\pi f_l N_d} \sum_{j1,j2} \frac{1}{\Delta^{j1,j2}} \left[\sum_{i=1}^{I(j1,j2)} \ln \left(\frac{C_l^{j1}}{C_l^{j2}} \right) \right]. \quad (7)$$

$Q_{Lg}(f_l)$ (or Q_0 and η) estimated using equation (6) is also subject to a random error that decays in proportion to $1/\sqrt{N_d}$.

Results

Two-station pairs aligned to the same event-to-station azimuth can seldom be obtained. In practice the approximations of the SSRs (equation 2) are obtained by requiring that the differences between the azimuths to the two stations are smaller than a preset maximum allowable value, $(\delta\theta)_{\max}$. The choice of $(\delta\theta)_{\max}$ is less restrictive for the Lg than for many other phases, since the Lg contains a minimal source radiation pattern in realistic, 3D structures (e.g., Xie, 1998). In this study two different values of $(\delta\theta)_{\max}$ of 30° and 12° are used. These values result in 37 and 22 two-station pairs, respectively. SSRs are formed and used to fit $Q_{Lg}(f)$ models using equation (6). Figure 1 shows the path coverage by the two sets of SSRs, and Figure 2 shows the SSRs and the fit of the $Q_{Lg}(f)$ models. SSRs obtained with $(\delta\theta)_{\max} = 30^\circ$ are fitted by

$$Q_{Lg}(f) = (126 \pm 9) f^{(0.37 \pm 0.02)} \quad 0.2 \text{ Hz} \leq f \leq 3.6 \text{ Hz}. \quad (8)$$

SSRs obtained with a $(\delta\theta)_{\max}$ of 12° are very similar to those with a $(\delta\theta)_{\max}$ of 30° (Fig. 2) and are fitted by a $Q_{Lg}(f)$ with Q_0 of (134 ± 10) and η of (0.32 ± 0.02) , respectively. These values are virtually the same as those in equation (8) but are theoretically more subject to random errors and bias owing to a smaller $N_d Q_{Lg}(f)$ in equation (8) is therefore my preferred model.

The upper bounds of the bias in Q_l estimates, caused by nonunity values of C_l^j in equation (7), can be estimated following Xie and Mitchell (1990). For the preferred $Q_{Lg}(f)$ model, we have $N_d = 37$, $V_{Lg} \approx 3.5$ km/sec, and $\Delta^{j1,j2}$ between 200 and 500 km (Fig. 1). Also, some combinations of $\ln(C_l^{j1}/C_l^{j2})$ cancel. Contributions to nonunity values of C_l^j by site response and focusing/defocusing are extensively discussed by Xie and Mitchell (1990) and Xie (1998). A fairly conservative estimate of the upper bound of C_l^j is 2.0.

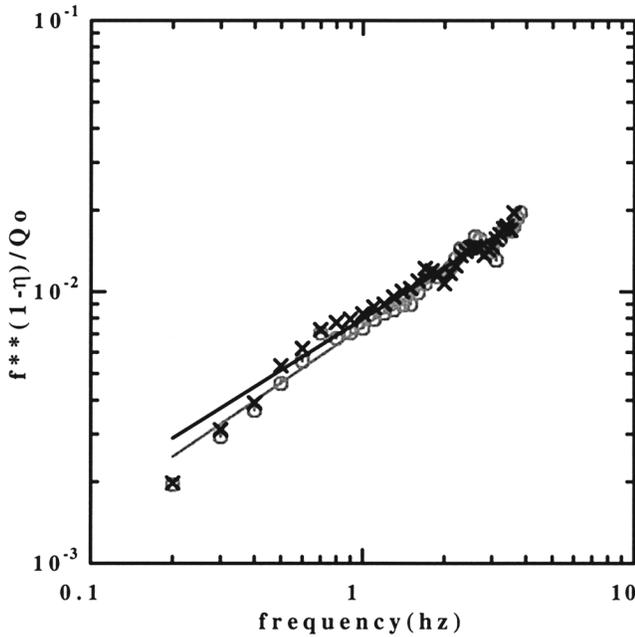


Figure 2. Stacked spectral ratios (SSRs) from many two-station pairs plotted in Figure 1, and the fit of best Q models (straight lines). Black and gray symbols are SSRs obtained using a $(\delta\theta)_{\max}$ of 30° and 12° , respectively. The Q models from fitting both sets of SSRs are similar. The Q model written on the top of the panel is from fitting the black symbols. $f = 0.2\text{--}3.6$ Hz; $Q_0 = 126 \pm 9$; $\eta = 0.37 \pm 0.02$.

Substituting the terms on the right-hand side of equation (7) by these estimates, at $f_i = 1.0$ Hz, we have

$$\delta \left(\frac{1}{Q_0} \right) < 0.006. \quad (9)$$

Such a bias is small. As an independent and empirical test of effects of the nonunity C_i^j , I form new SSRs by selecting eight pairs of spectral ratios under a criterion that, when the summations in equations (3), (5), or (7) are taken, the effect of $\ln(C_i^{j1}/C_i^{j2})$ completely cancel (for example, station pair AMDO-BUDO is sampled by reversed geometry, resulting in such a cancellation). The new SSRs are fitted by a Q_0 of (122 ± 18) , which is virtually the same as that in equation (8) except the uncertainty slightly increased owing to a smaller N_i . These demonstrate the robustness of the $Q_{Lg}(f)$ estimated using the two-station method. A drawback of this method is that it requires Lg spectra to be recorded at two stations that are approximately located on the same great circles, leading to lower cutoff frequencies at the more distant stations. As a result, the highest frequency used in this study is below 4 Hz, much lower than that of about 12 Hz used by McNamara *et al.* (1996).

Discussion and Conclusion

McNamara *et al.* (1996), using data from the same seismic experiment as that used in this study, observed a limiting

distance of about 700 km, beyond which Lg disappears owing to a high attenuation. That limiting distance is at least a factor 2 smaller than the maximum distances of Lg observation in other low Q_0 regions, where the estimated Q_0 values are below 267 (see Introduction). The Q_0 value estimated in this study is much lower than 267 and is therefore qualitatively in agreement with the observed short limiting distance in Tibet.

Quantitatively, the Q_0 estimate of about 126 by this study is lower by a factor of 3 than the value of 366 estimated by McNamara *et al.* (1996). To explore the cause of this large discrepancy, I note that the signal processing and inversion procedure of this study differs from that of McNamara *et al.* in several aspects. These differences include (1) different upper frequency limits as mentioned in the last section; (2) different components of the seismograms used: vertical and horizontal components are used in this and the previous studies, respectively; (3) different methods used to measure Lg amplitude: straightforward Fourier amplitude spectra are used in this study, whereas maximum amplitudes of bandpass filtered, time domain signals were used in the previous study; (4) different data censoring criteria used: the Lg spectra used in this study are selected based on the S/N ratios calculated using pre- P noise, whereas the previous study used S/N ratios calculated using the mean level of 50 sec of Lg coda; (5) nonidentical sampling areas by data used, and (6) different algorithms used for the inversion of $Q_{Lg}(f)$. Difference on the upper frequency limits used should not affect the Q_0 estimate since it is measured around 1 Hz, a frequency that is much lower than the high-frequency limits in either study. The effect of the different components used in estimating $Q_{Lg}(f)$ has been extensively investigated by several authors, including McNamara *et al.* (1996), and has been found to be very small. Different methods used for amplitude measurements should make little difference in Q measurement as long as the windowing and bandpass filtering do not cause significant processing artifacts and appropriate geometrical spreading terms (GSTs) are used. A GST with distance decay of $\Delta^{-0.83}$ was used by McNamara *et al.*, treating the time domain Lg signal as an Airy phase. In this study, a GST with $\Delta^{-0.5}$ is used in the frequency domain. To ensure that the choice of GST, in practice, does not significantly affect the $Q_{Lg}(f)$ estimates, I reinverted the SSRs obtained using the GST with $\Delta^{-0.83}$ to see its effect. The estimated Q_0 and η values are (150 ± 14) and (0.37 ± 0.02) , respectively. Obviously, such changes are too small to account for a Q_0 discrepancy of a factor 3.

To test the possibility that different data censoring criteria has somehow biased the $Q_{Lg}(f)$ model obtained in this study, I reran the inversion by selecting the 22 pairs of SSRs from seismograms on which the Lg has distinctly larger amplitudes than its preceding phases and coda. This selection procedure excludes any emergent Lg signals. The resulting Q_0 and η estimates are (125 ± 15) and (0.22 ± 0.03) . I do not favor this model since the emergent appearance of Lg is expected at the more distant stations when Q is low; there

is no reason to exclude the respective Lg spectra in Q estimates, whose robustness and reliability are dependent on the number of available two station pairs, Nd . In any case, the Q_0 estimate of about 125 obtained here is unchanged from that in equation (8).

Different nonidentical sampling areas could change Q_0 estimates if they caused much difference in the geographical areas sampled and if Q_0 values vary drastically in these areas. The 37 pairs of SSRs (from 74 amplitude spectra) used in this study roughly sample the area covered by the network used (Fig. 1), whereas the 106 records used by McNamara *et al.* (in the 2–4 Hz band; their figure 7) mostly sampled the same area, with a small fraction (about 20% or less) of paths also sampling the areas that are adjacent to the network but still lying inside the eastern Tibetan plateau. It is very unlikely that the relatively minor difference in sampling areas could explain a factor of 3 difference in the Q_0 estimates.

Contrary to the first five differences, the difference in the inverse algorithms used can potentially result in drastically different Q_0 estimates. In the method used by McNamara *et al.* (1996), Lg waveforms were filtered into five discrete frequency bands. Maximum Lg amplitudes were then read from the filtered time series and used as data in the inverse scheme. In each frequency band, there were no more than 106 amplitude data and many unknown free parameters to be solved for. These unknowns included the $Q_{Lg}(f)$ value, plus twenty parameters describing the sources and eight parameters describing the site responses. In all there were more than 100 free parameters to be solved for, making the inverse

problem less stable. The Q_0 value should have a significant trade-off with the source and site parameters because the observed Lg amplitudes could be fit by simultaneously increasing the estimated Q_0 value and decreasing the source and/or site terms. By contrast, in this study effects of the source radiation and site responses are virtually canceled by taking the SSRs, leaving only two free parameters, Q_0 and η , that are solved for.

The Q_0 value of about 126, estimated for the eastern Tibetan plateau in this study, is the lowest ever documented for any continental areas. It is comparable or slightly lower than the recently reported values near the study area by Phillips *et al.* (2000) and Fan and Lay (2001). An implication of this low value is that the observed blockage of Lg for paths crossing the boundaries of the plateau (see Introduction) may be largely, or even entirely, attributed to the low Q_0 values in the plateau (see Fig. 3 for a detailed discussion). Another implication of the low Q_0 is that the crust in Tibet may be characterized by higher-than-normal temperature and fluid content, which are responsible for the low Q_0 values. Future research should be directed toward analyzing more seismic data from recent seismic experiments in the plateau to resolve details of the lateral variations of $Q_{Lg}(f)$ in the plateau.

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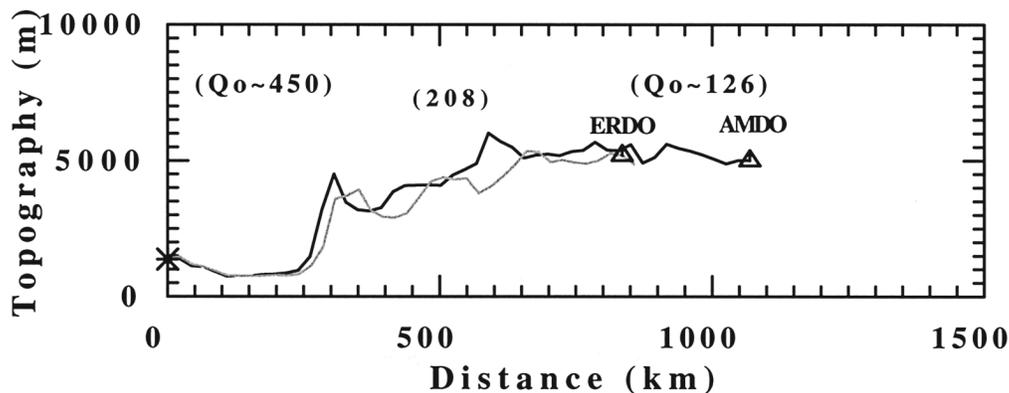


Figure 3. An example showing how low Q values alone can cause an Lg blockage across the northern boundary of Tibetan Plateau. Plotted are topographic profiles from the 052192 Lop Nor explosion (asterisk; shown as star in Fig. 1) to station ERDO (curve in gray) and AMDO (curve in black). The observed Lg amplitudes show a partial blockage at ERDO (McNamara *et al.*, 1996, their figure 10) and complete blockage at AMDO. Numbers in parentheses are Lg Q_0 values: the left segments of the profiles are in the Tarim Basin where Q_0 is 450 or higher from previous works (e.g., Xie *et al.*, 1996). Between ERDO and AMDO, Q_0 should be close to 126 (this study). Using (1) these Q_0 values, (2) the spectral amplitude in the expected Lg window at AMDO, and (3) the known source level of the 052192 explosion (Xie *et al.*, 1996), I estimate that Q_0 between the topographic boundary and station ERDO is 208 or lower. This means that a low average Q_0 value in the plateau of between 126 and 208 is capable to block Lg at AMDO, and a strong scattering at the topographic boundary is not required. This is consistent with the simulation of Kennett (1986), who shows that the scattering can not fully account for the blockage.

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