Lateral variations of crustal seismic attenuation along the INDEPTH profiles in Tibet from Lg Q inversion

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1 Fourier spectra of crustal guided Lg waves are collected to study seismic attenuation, or Q, along the INDEPTH profiles. The 1 Hz Lg Q (Q0) values are measured between many pairs of two stations and input to a back projection algorithm to obtain laterally varying Q0 models. Along the INDEPTH III profile in central Tibet, the Q0 model consists of low and nearly constant values of ~90. These low values are consistent with abnormally high temperature as well as partial melts in the crust. Along the INDEPTH II profile in southern Tibet the Q0 values increase southward. They are extremely low (~60) over the northernmost segment but increase by a factor of 2 over a distance of ~100 km between the Indus-Yalong Suture and the Kangmar Dome. Farther south and into the high Himalayas, Q0 values are higher than 300. Regional Rayleigh waves observed along the northern INDEPTH II profile are used to infer a low-velocity, low-Q layer at midcrustal depths (between ~15 and 30 km) in southern Tibet. The aqueous fluid trapped in the upper crust and a midcrust partial melting zone, associated with the underthrusting of the Indian lithosphere, are the likely causes of the high Lg attenuation in southern Tibet. Low Q, rather then a systematic Lg conversion along a dipping Moho, is the main cause of the previously observed Lg blockage over paths crossing the southern Tibetan boundary.

INDEX TERMS: 7205 Seismology: Continental crust (1242); 7219 Seismology: Nuclear explosion seismology; 7255 Seismology: Surface waves and free oscillations; KEYWORDS: Lg Q, crustal attenuation, Tibet, Lg Blockage, crustal melting


1. Introduction

[2] The Lg is typically the most prominent seismic phase on high-frequency seismograms observed over continental paths at regional to teleseismic distances. It can be modeled as a sum of higher-mode surface waves [Knopoff et al., 1973], or as many supercritically reflected S rays [Bouchon, 1982] traveling in the continental crustal wave guide. The Lg group velocity and quality factor (Q), which is inversely proportional to the Lg attenuation coefficient) reflect those of the crustal average. While the Lg velocity typically varies by less than 20% from one region to another, its Q varies by at least a factor of 3 on major continents such as North America [Nuttli, 1973]. The Lg Q is often assumed to obey a power law frequency dependence:

\[ Q_{Lg}(f) = Q_0 f^{\eta}, \]

where Q0 and \( \eta \) are Lg Q at 1 Hz and its power law frequency dependence, respectively. The Lg Q0 value correlates well with the intensity of recent tectonic activity in the crust. It is typically ~200 in the tectonically active western North America, and is higher than 650 for the stable, central and eastern North America [e.g., Mitchell, 1995].

[3] The regions in and around the Tibetan Plateau seem to exhibit the most severe Lg attenuation of all the continental regions in the world. Earlier studies observe drastic Lg amplitude reductions over paths crossing the boundaries of the Plateau [Rucaykin et al., 1977; Ni and Barazangi, 1983]. These amplitude reductions are so severe that the Lg waves disappear into the coda of preceding body waves. This phenomenon of complete Lg elimination has been known as

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Lg blockage.” Ruzaikin et al. [1977] proposed two viable mechanisms that may be responsible for this phenomenon. The first mechanism involved drastic lateral variations of the crustal structure, such as a crustal thickening under the high Himalayas, that disrupt the wave guide and cause the Lg to be systematically converted into other waves such as the mantle Sn. The second mechanism involved a very low (two-digit) crustal Q. Both mechanisms seemed to have been possible as the Plateau was created by the India-Asia continental collision, resulting in unusual velocity and Q structures. The identification of the true cause of the Lg blockage should significantly improve our knowledge of the complex structure of the Tibetan crust.

Numerous works have since been conducted to quantify the effect of a possible crustal thickening on the Lg propagation [e.g., Kennett, 1986] and to measure Lg Q. Before the 1990s, reliable Lg Q measurements in and around the Tibetan Plateau were not practical owing to a lack of digital seismic stations inside the Plateau. Stations outside the plateau only provided long recording paths over which much of high-frequency (∼1 Hz) Lg content has been attenuated. These paths also tended to cumulatively sample various geographic zones belonging to the interior, boundary and exterior of the plateau so it was difficult to measure Q in any specific zones, such as a boundary zone. Major improvements of the quality and geographic coverage of seismic data have occurred since the early 1990s with the installation of the Chinese digital station Lhasa, and with three collaborative PASSCAL experiments conducted in Tibet. Using data from the first of these experiments (the 1991–1992 Sino-U.S. Tibetan Plateau experiment), McNamara et al. [1996] reported that the Lg could be observed for paths located entirely inside the Plateau, out to some maximum distances of ∼700 km. They also estimated an average Lg Qo of 366 ± 37 for much of the interior regions of the Plateau. Reese et al. [1999] and Phillips et al. [2000] estimated Lg Qo for subregions of the Tibetan Plateau that are generally above 200, a value similar to those estimated for the western U.S. and Iran, where Lg can be observed out to ∼2000 km. Xie [2002] pointed out that the maximum distance of ∼700 km to observe Lg in the plateau qualitatively required Qo values that were lower than 200. He reanalyzed the 1991–1992 data used by McNamara et al. [1996] with a reliable two-station method and obtained an average Lg Qo of 126 ± 9 for a large area of eastern Tibet. He also demonstrated that the previously reported Lg blockage across the northern boundary of the Plateau could be largely or fully attributed to low intrinsic Q in the Tibetan crust. The low Qo value and inference by Xie [2002] are physically plausible given that the Tibetan crust is generally associated with high heat flows and widespread fluids, and are supported by very recent Q estimates by Fan and Lay [2002, 2003] and S. Phillips (personal communication, 2004).

In this paper we present new measurements of Lg Q using data from the INDEPTH (International Deep Profiling of Tibet and the Himalaya) II and III experiments [Nelson et al., 1996; Haines et al., 2003] in southern and central Tibet,
respectively. Stations deployed during both experiments roughly delineate great-circle profiles, thus permitting inversions for the lateral variations of $Lg Q$. We report a trend of $Q_0$ variation in which it is very low ($\sim 100$) in central Tibet, decreasing southward in southern Tibet to extremely low values of $\sim 60-70$, and then increasing southward to higher than 300 in the southernmost Tibet in the general regions of high Himalayas south of the Indus-Yalong Suture (IYS) and Kangmar Dome. We infer that the low $Lg Q$ is the primary reason for the $Lg$ blockage across the southern boundary of the Plateau. We also use long-period Rayleigh waves observed along the northern INDEPTH II profile to infer a midcrustal low-velocity, low-$Q$ layer. We suggest that the low $Lg Q$ in southern Tibet is mainly caused by widespread aqueous fluids and partial melts in the middle crust.

2. Data

[6] The INDEPTH II experiment was conducted in southern Tibet between May and October 1994 [Nelson et al., 1996]. Nine broadband and six short-period stations were deployed roughly along a profile in a NNE direction, starting from the high Himalayas, crossing the IYS and ending about 150 km north of it (Figure 1). The INDEPTH III experiment was conducted in central Tibet between July 1998 and June 1999. Forty-seven broad or intermediate band and 15 short-period stations were deployed along a profile across the Bangong-Nujiang Suture (BNS) in a NNW direction (Figure 1) [cf. Rapine et al., 2003]. Stations in both experiments have three components with sampling rates of 20 s$^{-1}$ or higher. We only use vertical component $Lg$ waveforms, which are Fourier transformed with a 20% Cosine taper window with the two corners set at group velocities of 3.0 and 3.5 km/s (slightly adjustable [cf. Xie, 1998]), respectively. Instrument responses are then removed to yield ground displacement spectra. The lowest frequencies that we use are about 0.2 Hz. The highest frequencies are typically controlled by the minimum signal/noise ratio of 2 allowed in this study; they can be as low as $\sim 1.5$ Hz at the more distant stations. Figure 2 shows a time domain record section of seismograms along the INDEPTH II profile in southern Tibet, from an event in the north (94236 in Figures 1 and 3) that is approximately in line with the profile. The $Lg$ amplitude is large at the northernmost station (BB05), but decreases very rapidly along the first half of the profile. The $Lg$ amplitude reduction from BB05 to BB14 is a factor of $\sim 10$ despite the short (150 km) distance between the stations. The selected $Lg$ amplitude spectra plotted on the bottom of Figure 2 show that the amplitude decay at high frequencies is faster along the profile, causing the spectral corner to shift lower with increasing distance. This phenomenon is qualitatively diagnostic of low $Q$ because a finite $Q$ is a measure of energy loss per wavelength and as such, causes faster amplitude decay at higher frequencies.

3. Methodology

[7] Stations deployed during each INDEPTH experiment are fit by a great circle profile under a least squares criterion. A primary objective of this study is to invert for the variations of $Lg Q_0$ along these best-fit profiles (e.g., profile A-A’ in Figure 3). The inversions are composed of two steps. In the first step, interstation $Q_0$ are measured from selected pairs of two stations. These measurements are then used in the second step to map the lateral variations of...
Lg $Q_0$ using a back projection method. We describe these methods below.

3.1. Two-Station Method to Measure Interstation $Q$

In an ideal situation for applying this method, two recording stations are aligned exactly with the source (Figure 4a) and the Earth’s velocity structure is 1D. Denoting the two stations as stations $i$ and $j$ at which the Lg spectra $A_i(f)$ and $A_j(f)$ can be collected, one can define and calculate a scaled spectral ratio

$$ R(f) = \left( \Delta_i^{1/2}/\Delta_j^{1/2} \right) [A_i(f)/A_j(f)], $$

where $\Delta_i$ and $\Delta_j$ are the epicentral distances and the square root of their ratio cancels the effect of geometrical spreading in 1D. The $Q_0$ and $\eta$ over the interstation path $\Delta_{ij}$ can be estimated using $R(f)$ and the following relationship [Xie and Mitchell, 1990b],

$$ (V_{Lg}/\pi\Delta_{ij}) \ln(R(f)) = f^{1-\eta}/Q_0, $$

or its logarithm:

$$ \ln \left[ \frac{V_{Lg}}{\pi\Delta_{ij}} \ln(R(f)) \right] = (1-\eta) \ln f - \ln Q_0, $$

where $V_{Lg}$ is the typical Lg group velocity of 3.5 km/s.

Figure 3. Detailed view of the INDEPTH II stations (triangles), regional events (stars), and the great-circle profile (A-A') that best fits the stations.

Figure 4. Schematics showing (a) an ideal recording geometry for the application of a two-station method for $Q$ measurement and (b) a more practical geometry in which the source-to-station azimuths vary by an amount $\delta\theta$. 

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measurements are shown as gray lines. The interstation paths used for two-station Q with nonzero station method (equation (3) or (4)) is used. In a geometry complications cause various potential errors when the two-focusing and defocusing, and nonunity site response. These complications to velocity, the true Earth has a 3D structure which causes into cells of a constant height of 0.1°/C176/A-A Strip area that encloses the great circle profile (Figure 5).

**Figure 5.** Strip area that encloses the great circle profile A-A’ and INDEPTH II stations shown in Figure 3. The strip has a finite width of 0.6° and is divided by the dashed lines into cells of a constant height of 0.1°. Q among cells is variable. The interstation paths used for two-station Q measurements are shown as gray lines.

[9] The real situation for applying the method is more complicated. In terms of recording geometry, a perfect alignment of stations and source shown in Figure 4a is usually not possible, particularly with passive seismic experiments. A more practical recording geometry is shown in Figure 4b where the azimuths from the source to the two stations are different by an angle, δθ. In terms of seismic velocity, the true Earth has a 3D structure which causes complications to Lg propagation such as forward scattering, focusing and defocusing, and nonunity site response. These complications cause various potential errors when the two-station method (equation (3) or (4)) is used. In a geometry with nonzero δθ (Figure 4b), the Q0 and η values estimated for the interstation path Δij may be erroneous because they may contain effects of Lg attenuation outside the path. Additionally these values may also contain errors resulting from a nonisotropic source radiation pattern. These errors can be minimized if we require the angle δθ to be smaller than a threshold value δθ,max. A practical δθ,max comes from the consideration that in the real 3D velocity structures, Lg contains forward scattered waves that arrive at the recording stations with a spread of propagation directions. Der et al. [1984] estimated that these directions are deflected from the great-circle by as much as ±15° using array data. These deflections cause an inherent error in the interstation Q measurement that is very similar to that caused by nonzero δθ. Therefore it is reasonable to set a δθ,max at 15°.

[10] Potential errors in two-station Lg Q measurements caused by focusing and defocusing in 3D media and nonunity site responses are more difficult to suppress. We may quantify the total error by a factor of 1 + δx that is unaccounted for in equations (3) and (4) which assume the ideal sampling geometry and 1D structure. In Appendix A we show that when δx is small, the relative error in the estimated interstation Q0 is related to δx by (equation (A15))

\[
\delta Q_0 / Q_0 \approx 1.1 (Q_0 / \Delta_{ij}) \delta x.
\]

Therefore for the purpose of reducing the relative error in Q0 estimates, it is desirable to use two station pairs with separations Δij that are as large as possible.

### 3.2. Model Parameterization and Selection of Two-Station Pairs

[11] Stations in each INDEPTH experiment occupy a narrow strip with a finite width, W, along the best-fitting great circle profiles (e.g., profile A-A’ in Figure 5). W is about 0.6° and 0.4° for the INDEPTH II and III profiles, respectively. We parameterize the laterally varying Lg Q0 structure by discretizing the narrow strips into cells with a constant height, H, of 0.1° (Figure 5). Q0 inside each cell is assumed to be constant. Under this parameterization, the inversion of lateral variation of Lg Q0 along profile A-A’ becomes one that solves for the Lg Q0 values in the cells. The input of this inversion is composed of interstation Q0 measurements from a combination of two-station pairs.

[12] The choice of two-station combinations is made under two sampling principles: first the interstation distances, Δij, should be as large as possible to reduce measurement errors in the interstation Q0 (last section). Second, Lg spectra from all stations should be sampled with equal weight. In the ideal case when the profile is sampled by an even number of 2N stations, the simplest combination that satisfies both considerations is one in which station n (n = 1, 2, 3… N) is paired with N + n, forming N pairs. In practice, complications such as an odd number of stations occur. We form two-station combina-

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<th>Longitude, °E</th>
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tions that are as close as possible to the ideal case with some minor compromises as described later.

3.3. Inversion for Lateral Variations of $Q$

When there are a total of $N$ two-station paths over which $Q_0$ values are measured, we denote the distance and $Q_0$ along the $n$th path ($n = 1, 2, ..., N$) by $D_n$ and $Q_n$, respectively (in the notation of the last section the distance is $D_n$, but the subscript $n + N$ is dropped here for conciseness; likewise we drop subscript “0” for $Q$ here). Denoting the length over which the $n$th path intersects the $m$th cell by $D_{mn}$ (see Figure 5), the following linear relationship exists:

$$D_n = Q_n - \sum_{m=1}^{M} D_{mn}/Q_m + e_n, \quad n = 1, ..., N,$$

where $M$ is the total number of cells along profile A-A’, $Q_m$ is the $Q_0$ value inside the $m$th cell to be inverted for, and $e_n$ is a small residual. Equation (6) represents a set of $N$ standard linear equations that can be solved for to obtain the unknown $Q_m$ under the least squares criterion:

$$\sum_{n=1}^{N} e_n^2 = \text{min.}$$

In this study we solve for these equations using an iterative back projection method [e.g., Humphreys and Clayton, 1988] used by Xie and Mitchell [1990a]. We choose this method because equation (6) is a special, 1D case of equation (11) of Xie and Mitchell [1990a], who dealt with a 2D case. Extensive computer programs for solving for the unknown $Q_m$ model, and for estimating the resolution and random errors associated with the model, have already been implemented for the 2D case by Xie and Mitchell [1990a].

### Table 2. Two-Station $Q_0$ Measurements

<table>
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<th>Event</th>
<th>Station Pair</th>
<th>$Q_0$</th>
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<td>236</td>
<td>BB05-BB34</td>
<td>74 ± 7</td>
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<tr>
<td></td>
<td></td>
<td>BB08-BB20</td>
<td>80 ± 7</td>
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<td></td>
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<td>BB36-BB23</td>
<td>99 ± 9</td>
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<td></td>
<td></td>
<td>BB10-SP25</td>
<td>92 ± 10</td>
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<tr>
<td></td>
<td></td>
<td>BB05-BB14</td>
<td>65 ± 11</td>
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<td></td>
<td></td>
<td>BB10-SP27</td>
<td>143 ± 11</td>
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<tr>
<td></td>
<td>180</td>
<td>BB08-BB18</td>
<td>78 ± 14</td>
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<tr>
<td></td>
<td></td>
<td>BB08-BB20</td>
<td>78 ± 13</td>
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<td></td>
<td></td>
<td>BB14-BB23</td>
<td>136 ± 69</td>
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<td>181</td>
<td>BB05-BB14</td>
<td>63 ± 24</td>
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<td>BB05-BB34</td>
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<td></td>
<td></td>
<td>BB08-BB20</td>
<td>79 ± 15</td>
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<td>BB08-BB18</td>
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<td>ST14A-ST30</td>
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<td>ST16A-ST32</td>
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<td>ST20-ST26</td>
<td>76 ± 19</td>
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Zhao and Xie [1993]. The adaptations of these programs required for the current 1D case are minor. Details of these adaptations are given in Appendix B.

4. Results

4.1. \( Lg Q_0 \) Along the INDEPTH II Profile

Three regional events with \( m_b \) values between 5.1 and 5.9 (Table 1) occurred approximately in line with profile A-A’ (Figures 1 and 3) through the INDEPTH II stations. The numbers of recording stations vary with events. In all, there are 9 broadband and 2 short-period stations that provide data from at least one event. For each event we form a combination of two-station pairs for interstation \( Q_0 \) measurements following the sampling principles described in the earlier section. Minor compromises have to be made to avoid using pairs whose spectral ratios are unstable. For example, for event 94236 the ideal combination should contain a station pair BB14-SP27 (Figures 3 and 5), but that pair does not yield a stable \( Q_0 \) measurement with a reason that is explained in Appendix A. So we replaced that pair by two alternative pairs, BB05-BB14 and BB10-SP27. We formed a total of 15 two-station pairs which are shown in Figure 5 and listed in Table 2. The maximum difference in source-to-station azimuths, \( \delta a \), is 12°, which is below the threshold value of 15° set in an earlier section. To stabilize the spectral ratios, we applied a 17-point running average to smooth the raw \( Lg \) spectra prior to forming the spectral ratios. This smoothing reduces the frequency resolution from about 0.024 Hz carried by raw spectra to 0.4 Hz, which is acceptable since here we seek a stable measurement of \( Q_0 \) rather than \( \eta \) (equation (1)). Figure 6 shows examples of the linear regression fitting of \( Q_0 \) values based on equation (4). The interstation \( Q_0 \) values in Table 2 are generally low (~60–80), except for the southernmost pairs involving stations SP27, SP25 and BB23. For the latter pairs the interstation \( Q_0 \) values are near or above 100, indicating an increase of \( Q_0 \) toward the Himalayas.

The two-station \( Q_0 \) values are input to the back projection algorithm to invert for a laterally varying \( Q_0 \) model and its resolution and random errors. The results are shown in Figure 7, where the \( Lg Q_0 \) model along profile A-A’ may be divided into three segments with distinctly different features. Along the northernmost segment from A and to just north of the IYS, the model contains low and nearly constant \( Q_0 \) values between ~65 and 70. Along the central segment across the IYS, the \( Q_0 \) values increase from ~75 at 20 km near the IYS to ~90 at the IYS, and to ~190 at ~20 km south of the Kangmar Dome. This segment is geologically a part of the Tethyan Himalayas. Along the southernmost segment which traverses into the high Himalayas \( Q_0 \) increase steeply toward south, from about 200 at 40 km south of the Kangmar Dome to above 500 at 50 km from point A. The southern end of the \( Q_0 \) model is not shown because of concerns about errors as elaborated below.

In Figure 7, the lateral resolution of the \( Q_0 \) model varies between 0.6° and 1.3° along the entire profile. This implies that the N-S variation of the true \( Lg Q_0 \) is smeared out over lengths of at least 70 km by the model. The random
errors in Figure 7 are generally small except along the southernmost segment, where there are two causes for the large errors. The first cause is the sparser path coverage. For example, within 50 km of point A the profile is only sampled by one pair of stations (SP27 and BB14 from event 236), the errors become larger than 300 so we do not show the $Q_0$ model there. A second cause of the larger errors in the south comes from a fundamental problem in mapping a laterally varying $Q$ model: the errors in $Q$ estimates tend to increase where $Q$ itself becomes higher, such as along the southern INDEPTH II segment. We describe this fundamental problem in detail in Appendix A. A concern has been raised that regardless of the cause of the error bars in Figure 7, the fact that they are larger in the south makes it less certain that the true $Q_0$ along the profile indeed increases southward. To address this concern, we use many pairs of two stations south of station BB14, which is approximately located at the IYS, to estimate an approximate average $Q_0$ of $191 \pm 40$ for the southern INDEPTH II profile. This estimate is unambiguously higher than the respective estimate of $64 \pm 10$ made for the northern profile (north of IYS; see a later section). A further support for the southward increase of $Lg Q_0$ can be found in the work of Ni and Barazangi [1983, Figure 15], where the $Lg$ propagations is efficient under the Himalayas, but inefficient farther north in southern Tibet.

4.2. $Lg Q_0$ Along the INDEPTH III Profile

[17] Figure 8 shows the great-circle profile B-B’ through INDEPTH III stations, and locations of two well recorded, $m_b \sim 4.4$ earthquakes that are approximately in line with the profile. This profile is sampled more densely, by more than 40 stations, than the INDEPTH II profile which is sampled by only 11 stations. Following the two sampling principles mentioned in an earlier section, a total of 32 two-station pairs are formed for interstation $Q_0$ measurements. The maximum difference in the source-to-station azimuths, $\delta\theta$, is about $8^\circ$, satisfying the criteria of $\delta\theta < 15^\circ$ set earlier. The 17-point smoothing used in the INDEPTH II study is also applied here to obtain stable $Q_0$ measurements. Examples of interstation spectral ratios and the linear regression fit for $Q_0$ values are shown in Figure 9. The $Q_0$ values are subsequently used in the back projection inversion to obtain a laterally varying $Q_0$ model, which is shown in Figure 10 together with its resolution and random errors. The model along profile B-B’ contains $Q_0$ values that fluctuate around a low level of about 90 with no significant N-S variations.
In particular, there is no detectable change in $Q_0$ across the major Bangong-Nujiang Suture (BNS). This Mesozoic suture is probably a surface feature and no longer has any major tectonic significance. The lack of lateral variations in the $Q_0$ model is not an artifact of limited resolutions which vary between about 0.3° and 1.5° (Figure 8) and are comparable or better than those along the INDEPTH II profile (Figure 7). The random errors in Figure 8 vary between ~10 and 15% along much of the profile. Toward the two ends, the errors increase mainly because of the sparser spatial coverage by interstation paths there. Within the level of random errors, the entire $Q_0$ model in Figure 8 can be fit by a line of a constant $Q_0$ of ~89.

4.3. Frequency Dependence of $Lg Q$

Stable measurements of interstation $\eta$ values are much more difficult to obtain than those of $Q_0$ values, so no attempt is made to invert for the lateral variations in $\eta$. Instead, average $\eta$ values are estimated along the northern segment of the INDEPTH II profile and the entire INDEPTH III profile, over which the $Lg Q_0$ models show little lateral variations (Figures 7 and 10). To obtain a stable $\eta$ estimate along the northern INDEPTH II profile, we reprocessed the $Lg$ spectra from all three events and northern stations starting from BB14, which is near the IYS. The smoothing scheme used by Xie [1998] is applied to raise the resolution of the resulting spectra to 0.1 Hz from about 0.4 Hz used before. All possible pairs of two stations are used to estimate the interstation spectral ratios, regardless of the interstation distance $\Delta_{ij}$. This results in 17 sets of spectral ratios whose average is used to fit for a single set of $Q_0$ and $\eta$, as shown in Figure 11. The same reprocessing procedure is applied to obtain a set of average spectral ratios and the best-fit $Q_0$ and $\eta$ for the entire INDEPTH III profile, also shown in Figure 11. Between 0.3 and 1.6 Hz, the average ratios along the INDEPTH III profile closely follow a linear trend predicted by a power law frequency dependence model with a small $\eta$ of $0.0 \pm 0.1$. The averaged ratios along the northern INDEPTH II profile, on the other hand, are not well approximated by a power law frequency dependence. A moderate $\eta$ model of $0.5 \pm 0.1$ is obtained, but the ratios between 0.3 and 0.7 Hz significantly depart from the straight line delineated by the model. This complication may have three possible causes: (1) the fluctuations of spectral ratios resulting from 3D structural complications and site responses, (2) abnormal depth variations of shear $Q$ that can be mapped into an apparent frequency dependence of $Lg Q$ in a complex and nonlinear manner [Mitchell and Xie, 1994] (note that the complexity may worsen in the presence of a midcrustal low-velocity layer); and (3) a localized peak of shear $Q$ near 0.7 Hz caused by a single relaxation time of ~1.4 s [e.g., Anderson, 1989]. Only cause 3 implies that the observed frequency dependence of $Lg Q$ reflects a true frequency dependence of shear $Q$. With the INDEPTH II data, we do not have sufficient constraint to indicate which cause plays a dominant role. Consequently, we can not infer whether the shear $Q$ in any portion of the crust is inherently dependent on frequency.

5. Discussion

5.1. Low $Q$ as the Main Cause of $Lg$ Blockage Through Southern Tibet

The most striking feature in the $Lg Q_0$ models reported in this paper is the extra low $Q_0$ values between 60 and 70 found along the northern segment of the INDEPTH II profile, between the vicinity of Kangmar.
Dome and ~100 km north of Lhasa (Figure 7). These \(Q_0\) values are among the lowest ever reported in continental regions and cause a factor of 10 reduction of \(L_g\) amplitude over a distance of 150 km, as shown in Figure 2. In that figure at the southern stations such as BB23, \(L_g\) is still observable owing to the attenuated and weak regional \(P_g\) wave coda. Were there a stronger coda from preceding phases, \(L_g\) would have not be visible. Low, two-digit \(L_g\) \(Q_0\) values in southern and central Tibet (Figures 7 and 10) explain well why over far regional or teleseismic paths crossing southern Tibet the \(L_g\) appears to be completely eliminated, or “blocked” [e.g., Ruzaitkin et al., 1977; Ni and Barazangi, 1983]. Over those paths mantle body waves are prominent because first, they generally have downgoing take-off angles, hence are not significantly affected by the low crustal \(Q\) near the sources. Second, these waves are often observed in the distance range between about 15° and 30°, in which travel time triplications of deep-turning mantle body waves tend to cause large amplitudes. The stronger body waves should have more prominent coda in the expected \(L_g\) time window, making the severely attenuated \(L_g\) invisible.

In the Introduction we mentioned two mechanisms of the \(L_g\) blockage proposed by Ruzaitkin et al. [1977], including a systematic conversion of \(L_g\) to other waves and a low crustal \(Q\). We do not believe the conversion mechanism plays a dominant role for three reasons elaborated below: (a) As mentioned earlier, the rapid \(L_g\) amplitude reduction along the northern INDEPTH II profile (Figure 2) is accompanied by a frequency shift which is compatible with a low \(Q\). This shift is incompatible with a systematic \(L_g\)-to-body wave conversion along a dipping Moho because such a conversion is a boundary transmission process independent of frequency. (b) In terms of locations of the low \(Q_0\) regions, Figures 7 and 10 show that these regions onset to the north of the Kangmar Dome and extend northward through central Tibet. These regions are characterized by a relatively flat surface topography with elevations at ~5000 m. A systematic \(L_g\)-to-body wave conversion should occur in regions where the crustal thickness changes abruptly, as caused by topographic changes in both the Earth’s surface and deep crustal interfaces including Moho [e.g., Kennett, 1986]. In southern Tibet, such changes occur in an area located south of the

Figure 9. Examples of linear regression fit for interstation \(Q_0\) values using spectral ratios from pairs of two stations of the INDEPTH III experiment. Plotted in the same manner as in Figure 6.
Kangmar Dome, between the main thrust belt (MTB (Figure 1)) and the high Himalayas. In this 100- to 150-km-wide zone, the surface elevation increases from about sea level at the MBT to as high as ~8 km along the peaks of the Himalayas such as Mount Everest, while the crust thickens from ~45 km to ~70 km as determined by analyses of gravity data and receiver functions [e.g., Zhao et al., 1993; Nelson et al., 1996; Kind et al., 1996, 2002; Yuan et al., 1997]. The region of high Himalayas is sampled by the southern INDEPTH II profile where \( Lg Q_0 \) values are estimated to be higher than 300 (Figure 7), indicating a much weaker \( Lg \) attenuation than the crust farther north. Note that we have discussed extensively the reliability of this observation. (c) After decades of efforts, none of the numerical simulations succeeded in demonstrating that a systematic conversion of \( Lg \) waves in southern Tibet is capable of eliminating them [e.g., kennett, 1986; Zhao et al., 2003]. In summary, while a systematic \( Lg \) conversion resulting from crustal thinning in southern Tibet probably plays a role in the attenuation of \( Lg \) waves, that role seems to be secondary compared to that of the low crustal \( Q \). Our argument is in line with those made by Xie [2002] and Fan and Lay [2002] for northern Tibet based on observations with lower spatial resolutions.

5.2. Interpretation of Regional Rayleigh Waves in Southern Tibet

[21] The lowest \( Q_0 \) values of ~60 are found along the northern INDEPTH II profile in a region north of the Kangmar Dome. The location coincides with a band of bright spots found from reflection surveys [Nelson et al., 1996; Brown et al., 1996]. These bright spots are located at 10–15 km depth and are interpreted as marking the top of a low \( P \) velocity layer. Studies of seismical receiver functions [Kind et al., 1996], and surface waves [Cotte et al., 1999; Rapine et al., 2003] in southern Tibet found that a layer with low \( S \) wave velocities starts at similar depths. Makovsky and Klemperer [1999] analyzed the P-to-S reflected phase from the bright spots and suggested that they were caused by aqueous fluid concentrations which are underlain by partial melts. Given that seismic \( Q \) is more sensitive to the presence of fluids than is velocity [e.g., Mitchell, 1995], the coincidence of the low \( Lg Q_0 \) values with the bright spots suggests that the layers of enriched fluids are the main cause of the low \( Q_0 \). However, since the \( Lg Q_0 \) only reflects the crustal average of shear \( Q \) [Mitchell and Xie, 1994], the low \( Q_0 \) model cannot be directly used to infer a low \( Q \) layer. Depth variations of velocity and \( Q \) may be studied using long-period Rayleigh waves. The three \( m_b \sim 5–6 \) regional events used in the \( Lg \) study (Table 1) are large enough to generate a regional, fundamental-mode Rayleigh wave along the INDEPTH II profile (Figure 12). A visual inspection of this wave from all events suggests that it has dominant periods between about 8 and 15 s.

[22] If the velocity structure is approximately 1D then the interstation phase velocity and \( Q \) measured using the Rayleigh wave, both as functions of period, can be used to invert for layered velocity and \( Q \) models [e.g., Herrmann, 1987; Mitchell, 1995]. We use program package SURF by Herrmann [1987] for Rayleigh wave data processing and inversion (for more discussion on the theories involved [cf. Mitchell, 1995]). We limit our analysis to the northern INDEPTH II profile where \( Lg Q_0 \) is low and velocity structure is simpler as discussed in previous sections. Frequency-domain analyses (not shown) suggest that the Rayleigh wave has a decreasing amplitude with increasing period above 15 s. Above ~30 s the fundamental mode becomes unseparable from higher modes. We use the same 17 two-station pairs that are used to estimate \( \eta \) (Figure 11) to estimate the average interstation Rayleigh wave phase

![Figure 10. \( Lg Q_0 \) model and the associated resolution and random errors along the INDEPTH III profile obtained using the back projection inversion. Plotted in the same manner as in the top panel of Figure 7, except that the vertical scale is expanded because the model is less variable. The abscissa is the distance along profile B-B' counted from the southernmost point B. See Figures 1 and 8 for map views of the BNS and points B, B'.](image)

![Figure 11. Average spectral ratios (ASR) values. Black crosses are ASR from 17 pairs of two stations along the northern INDEPTH II profile. Gray circles are ASR from all pairs of two stations along the INDEPTH III profile. The ASRs are used to obtain best-fit \( Q_0 \) and \( \eta \) values under the power law frequency dependence (black and gray straight lines, respectively). Deviations of the ASRs from the best-fit straight lines indicate how much \( Lg Q(f) \) departs from the power law frequency dependence.](image)
velocity, which is subsequently inverted for a layered shear velocity model shown in Figure 13. Above a depth of ~30 km the model has resolutions of ~10 km and contains a midcrustal low-velocity zone starting at the depth of ~12 km, where the shear velocity \(v_s\) is about 3.9 km/s. From there the \(v_s\) value decreases monotonically with depth until its reaches a minimum of 2.55 km/s at a depth of ~21 km, below which \(v_s\) rebounds and becomes larger than 3.5 km/s at the depth of ~30 km. At greater depths the model has poor resolution owing to the difficulty of modal attenuation.

Figure 12. Record section of broadband (unfiltered) seismograms across the INDEPTH II profile from event 94236. Plotted in the same manner as in Figure 2, where high-pass filtered seismograms are plotted. Solid and dashed guidelines mark arrivals of group velocities of 3.0 and 3.5 km/s (typical velocities of fundamental Rayleigh and \(L_g\) waves), respectively.

Figure 13. (left) One-dimensional shear velocity model for the northern INDEPTH II profile and the associated error bars, obtained by inverting average interstation Rayleigh wave phase velocity using the program package SURF [Herrmann, 1987]. A combination of 17 pairs of two stations north of the IYS, used in Figure 11, are used here to obtain individual interstation phase velocities, whose average are used in the inversion. The starting velocity model used in the inversion is the southern Tibet model by Rapine et al. [2003, Figure 6]. (right) resolution kernels at four different depths for the velocity model.
measurements are subject to larger errors than the theoretically predict that the Rayleigh wave $Q$ values above the depth of receiver functions, and by those previously obtained by Rapine et al. [2003] at greater depths. Dashed lines indicate the depths above which 80% of the Rayleigh wave energy is distributed.

At 15 s, 80% of the Rayleigh wave energy is distributed above a depth of 16 km. At 30 s that fraction of energy is distributed above 28 km, a depth that is almost twice as large. Therefore a highly attenuative (low-$Q$) layer, located in the midcrustal depth range between about 15 and 30 km, is a likely cause of the increasing attenuation of Rayleigh wave with periods.

5.3. Cause of Lateral Variations of $Lg Q$

[24] Southern Tibet is underlain by a midcrustal partial melt horizon resulting from collisional crustal thickening [e.g., Nelson et al., 1996]. The partial melting is thought to be responsible for multiple geophysical anomalies, including midcrustal low-velocity, low-crustal $Q$ observed in previous and this studies, high heat flows and crustal electrical conductivities [e.g., Wei et al., 2001; Li et al., 2003]. The partially molten horizon is underlain by the Indian crust with a stiff mantle lid. Water released from the underthrust Indian sediments and metasediments probably causes wet midcrust to melt at a lower melting temperature. Enriched radiogenic elements from subducted Indian crust provide additional heat source in the crust. This melting process lowers the viscosity of the midcrust and helps the Indian plate continue to underthrust beneath southern Tibet. The bright spots and partially molten horizon terminate in the south at the Kangmar Dome [Nelson et al., 1996], beyond which the Indian crust has not subducted to a sufficient depth to release water. This explains well why the $Lg Q_0$ values increase to the south of Kangmar Dome (Figure 7).

[25] Low $Lg Q_0$ values of ~90 are estimated along the INDEPTH III profile. Along this profile there is no observation of bright spots that would indicate abundant fluids in a midcrustal layer. However, the geothermal gradient and electrical conductivity remain high throughout much of the Plateau [e.g., Wei et al., 2001]. There are wide spread Cenozoic volcanisms in the Tibetan plateau, with most recent basaltic volcanisms distributed in the north [e.g., Ding et al., 2003]. There are two possible causes for the observed low $Q$ in central Tibet, including a high-temperature and partial melts. Our current knowledge of how these causes affect $Q$ values is still very limited. A number of authors [e.g., O'Connel and Budiansky, 1977; Kampfmann and Berckhemer, 1985; Mitchell, 1995; Hammond and Humphreys, 2000; Faul et al., 2004] have proposed that for intrinsic $Q$, fluid-bearing material may be characterized by low $Q$ values that have little or even negative frequency dependencies. In dry and hot materials on the other hand, thermally activated processes may dominate attenuation and may cause moderate frequency dependencies of $Q$. Assuming that these proposed patterns are generally true, the lack of frequency independence of $Lg Q$ along profile INDEPTH III (Figure 11) would suggest melt-bearing layers at or below midcrustal depths.

[26] It is interesting to compare the $Lg Q$ in Tibet with those reported in other continental plateaus, such as the Altiplano Plateau in the central Andes. Although the evolution history and tectonic setting of the Altiplano Plateau are not the same as those of southern Tibet, the two regions have similar features such as high heat flows and pronounced midcrustal low-velocity layers [e.g., Beck and Zandt, 2002]. In particular, Baumont et al. [1999] estimated a model of $Q_{Lg}(f) = 53 f^{0.76}$ for the region of Altiplano and

![Figure 14.](image)
Eastern Cordillera, a model that is similar to that obtained for southern Tibet in this study (Figure 11). Baumont et al. [1999] suggested that the low \( Lg \) \( Q \) in the Altiplano is dominantly caused by small-scale scattering by crustal heterogeneity on the basis of the significant frequency dependence in their \( Q \) model. However, as discussed earlier in this paper, a frequency dependence of \( Lg \) \( Q \) may arise from multiple reasons, such as a complex, nonlinear sampling process in which a depth-varying shear \( Q \) is mapped into an apparent frequency dependence of \( Lg \) \( Q \). It is possible that the similar \( Lg \) \( Q \) models for both Altiplano and southern Tibet are grossly caused by low intrinsic \( Q \) in partially molten crustal layers.

6. Conclusions

\[ Lg \] spectra recorded along the INDEPTH II and III seismic profiles are collected to study attenuation or \( Q \) in the Tibetan crust. A two-station method and a back projection algorithm are used to obtain models of laterally varying \( Lg \) \( Q \) (\( Lg \) \( Q \) at 1 Hz). Along the INDEPTH III profile in central Tibet, \( Q \) values are found to be low (~90) and nearly constant. Along the INDEPTH II profile in southern Tibet the \( Q \) values increase toward the south: they are ~60–70 between ~100 km north of Lhasa and the Indus-Yalong Suture, and reach above 300 between the Kangnmar Dome and the high Himalayas. The \( Lg \) \( Q \) along the INDEPTH III profile is found to be nearly frequency-independent. Fundamental mode Rayleigh wave recorded along the northern INDEPTH II profile is analyzed to constrain the depth variations of velocity and \( Q \) in southern Tibet. The phase velocity is best fit by a low shear velocity zone in the depth range between ~15 and 30 km. An increase of apparent Rayleigh wave attenuation at periods longer than ~15 s suggests that the shear \( Q \) in the low-velocity layer is also very low.

We interpret the extremely low \( Lg \) \( Q \) region in the southern Tibetan Plateau as being caused by a partially molten crust at midcrustal depths, possibly with a layer of enriched aqueous fluid located above it, starting at depths of 10 to 15 km. This interpretation is consistent with the idea that as the Indian lithosphere underthrust southern Tibet, sediments and metasediments reach H2O-saturated melting and dehydration occurs. In central Tibet, the low \( Q \) values are likely caused by partial melting and high temperature in the crust. The phenomenon of \( Lg \) blockage crossing the southern Tibetan boundary [Ruzaitkin et al., 1997; Ni and Barazangi, 1983] is mainly caused by unusually low \( Q \), or high attenuation, in the crust beneath southeastern Tibet.

Appendix A: Factors Affecting Errors in \( Q \) Estimates

The fact that errors in the estimated \( Q \) are inherently dependent on the measuring distance and frequency, and on \( Q \) itself, is often overlooked [Xie, 2004]. In this appendix we develop mathematical formulations that describe these dependencies in general, and apply these formulations to address the various error issues encountered in this study. Under the assumption that the anelastic property of the Earth’s Green’s function can be represented by a finite \( Q \), the amplitude spectra of a seismic phase, \( A(f) \) at frequency \( f \) and distance \( \Delta \) can be represented by

\[
A(f) = S(f)G(\Delta) \exp\left(-\frac{\pi f \Delta}{Q(f) V}\right) X(f),
\]

where \( S(f) \) and \( G(\Delta) \) represent our idealized models of the source spectra and geometrical spreading, \( V \) is the group velocity and \( Q(f) \) is the quality factor. The term \( X(f) \) represents modeling errors such as those caused by (a) complex ray deflections, focusing/defocusing and site responses associated with the 3D Earth structure whose details are unknown, and (b) deviations of the real source spectra from the idealized source model. \( X(f) \) can be random or systematic in nature. When \( X(f) \) is close to unity (the modeling error is small) and when two recording stations \( i \) and \( j \) are aligned with the source along the same great-circle, the interstation \( Q(f) \) can be approximately estimated using the two-station method. In this method, the recorded spectra at the two stations are used to calculate a spectral ratio \( R(f) \), defined as

\[
R(f) = \left[G(\Delta_i)/G(\Delta_j)\right] \times \left[A_i(f)/A_j(f)\right],
\]

where subscripts denote station numbers. From equations (A1) and (A2) we have

\[
\frac{V}{\pi \Delta_{ij}} \ln[R(f)] = \frac{1}{Q(f)} - \frac{V}{\pi \Delta_{ij}} \ln[E(f)],
\]

where \( \Delta_{ij} \) denotes the interstation distance, \( \Delta_i - \Delta_j \). \( E(f) = X(f)/X(f) \) is a new error term. If \( X(f) \) (and hence \( E(f) \)) are not unity we can ignore them in equation (A3) to obtain \( Q_i(f) \), an estimate of the true \( Q(f) \):

\[
\frac{V}{\pi \Delta_{ij}} \ln[R(f)] = \frac{1}{Q_i(f)}.
\]

The error in \( Q_i(f) \) is

\[
\delta Q_i(f) = Q_i(f) - Q(f).
\]

From equations (A3) and (A4), \( \delta Q_i(f) \) causes a respective error in the term \( E(f) \):

\[
\hat{\delta}\left(\frac{f}{Q_i(f)}\right) = \frac{f}{Q_i(f)} - \frac{f}{Q(f)} = \frac{-V}{\pi \Delta_{ij}} \ln[E(f)].
\]

If the modeling errors are small (\( X(f) \) and \( E(f) \) close to unity), then we may express \( E(f) \) by

\[
E(f) = 1 + \delta x,
\]

where \( \delta x \) is a small quantity \( \delta x \ll 1 \) and is assumed to be independent of \( f \) for simplicity. Taking the first-order approximation of equation (A6) under equation (A7) and then multiplying both sides by \( f/Q(f) \) results in

\[
\frac{\delta Q_i(f)}{Q_i(f)} = \frac{V}{f \pi} (Q(f)/\Delta_{ij}) \delta x.
\]
Equation (A8) gives the relative errors of \( Q(f) \) estimates when they are obtained at individual frequencies. An example of such \( Q(f) \) estimates is that obtained using long-period surface waves, which will be discussed more later. At higher frequencies \((f \sim 1 \text{ Hz})\) \( Q(f) \) is often assumed to obey a power law frequency dependence under the “absorption band” model:

\[
Q(f) = (f/f_0)^\eta Q_0,
\]

where \( f_0 \) is a reference frequency at which \( Q(f) \) equals to \( Q_0 \) \cite{Anderson, 1989}. Inserting equation (A9) into equation (A6) and introducing a new variable

\[
f' = f/f_0,
\]

we have

\[
\delta(f'^{1-\eta}/Q_0) = \frac{V}{f_0 \pi \Delta_{ij}} \ln(E(f)),
\]

where the error in \( f'^{1-\eta}/Q_0 \) arises from the error in \( Q_{0e} \) and \( \eta_k \), the estimates of the true \( Q_0 \) and \( \eta \):

\[
Q_{0e} = Q_0 + \delta Q_0; \quad \eta_k = \eta + \delta \eta.
\]

Taking a first-order approximation of equation (A11) using relations (A7) and (A12) we have

\[
(f'^{1-\eta} \ln f'/Q_0)\delta \eta + \left( f'^{1-\eta} \frac{1}{Q_0} \delta Q_0 = \frac{V}{f_0 \pi \Delta_{ij}} \delta \eta \right).
\]

There are various ways to obtain \( Q_{0e} \) and \( \eta_k \) and there may be a slight tradeoff between \( \delta Q_0 \) and \( \delta \eta \). We take a simple approach in which \( \delta Q_0 \) can be estimated at \( f' = 1 \) and the tradeoff goes away:

\[
\delta Q_0 = \frac{V}{f_0 \pi} (Q_0/\Delta_{ij}) \delta \eta.
\]

Equation (A14) is of the same form of equation (A8). Both equations quantify the relative error in the estimate of \( Q(f) \) when the estimate is conducted at a frequency of interest. In the case of equation (A8) this frequency is arbitrary whereas in the case of equation (A14) it is the reference frequency \( f_0 \) specified by the power law equation (A10). We first discuss the meaning of equation (A14) which states that for a given modeling error \( \delta \eta \), the relative error in \( Q_{0e} \), \( \delta Q_0/Q_0 \), increases with the ratio of \( Q_0/\Delta_{ij} \). As a special case we consider the \( \log Q_0 \) measurement for which \( f_0 = 1 \text{ Hz} \) (equation (1)), and \( V = 3.5 \text{ km/s} \). Equation (A14) becomes

\[
\delta Q_0 \approx 1.1(Q_0/\Delta_{ij}) \delta \eta \quad (\text{For Lg}).
\]

If \( \delta \eta \approx 0.2 \) and we wish to measure \( Q_0 \) with relative errors below the level of 20%, equation (A15) requires \( \Delta_{ij} \) to be no less than \( Q_0 \). Although equation (A15) is a first-order approximation under the condition of small \( \delta \eta \) (equation (A7)), the numerical simulation by \textit{Xie} \cite{1998, Figure 2} shows that it is approximately valid for a larger \( \delta \eta \) of 0.4. In this paper when \( \log Q_0 \) along the northern INDEPTH II profile is estimated, typical \( \Delta_{ij} \) is \( \sim 120 \text{ km} \) which is about twice of the typical \( Q_0 \) of 60 (Figures 5–7; Table 2). The estimated random error in \( Q_0 \) is at the level of 10–20%, which means \( \delta \eta \) is at the level of 0.2 to 0.4 according to equation (A15). If over the southern profile typical \( \delta \eta \) and \( \Delta_{ij} \) values remain at the same level but \( Q_0 \) increases by a factor of more than 3 (at the level of ~200), then the random error in \( Q_0 \) should increase by a similar factor. This explains why in this study the southern station pairs (those involve stations SP27 and BB23) tend to be associated with larger random errors in the interstation \( Q_0 \) measurements. For example station pair BB14-SP27 did not yield stable \( Q_0 \) measurement; errors between BB14-BB23 in Table 1 tend to be larger. From equation (6) which is used to map the laterally varying \( Q \) using interstation \( Q \) measurements, an error in the \( n \text{th} \) \( Q_0 \) measurement will cause errors in the discrete \( Q_m \) models (note subscript “0” is dropped) by

\[
\delta(1/Q_0) = \frac{1}{Q_0} \sum_{i=1}^{M} \frac{\Delta_{mu} Q_m}{Q_m} \delta Q_m, \quad n = 1, \ldots, N,
\]

where \( \delta \) is used to represent errors and a first-order approximation has been applied to the right hand side. Obviously, a larger \( \delta Q_m \) (or \( \delta(1/Q_m) \)) will cause all \( \delta Q_m \) values to increase. However, for those cells in which \( Q_m \) are larger and \( \Delta_{mu} \) not small, \( \delta Q_m \) will be larger because of the weighting of \( Q_m^{-2} \). We therefore see two inherent causes for the error bars in \( Q \) model to increase with \( Q \) itself: (a) larger interstation \( Q \) measurements tend to be associated with larger errors, and (b) larger \( Q_m \) are associated with larger errors (\( \delta Q_m \)) when the laterally varying \( Q_m \) are inverted for. These causes are responsible for the exceptionally large error bars along the southern INDEPTH II profile where \( Q_m \) are large (Figure 7).

\[\text{[30]}\] When no power law is assumed for \( Q(f) \), such as the case when Rayleigh wave attenuation coefficient (or \( Q(f) \)) is estimated in this study, equation (A8) gives the error in the estimated \( Q(f) \) in the same form of equation (A14), except \( f \) is an arbitrary rather than a reference frequency. For Rayleigh wave the typical \( f \) is \( \sim 0.1 \text{ Hz} \) which is about an order of magnitude lower than the \( \log f_0 \) of \( \sim 1 \text{ Hz} \). If \( \Delta_{ij} \) and \( V \) remain similar, we expect the errors in Rayleigh wave \( Q(f) \) estimates to be an order of magnitude larger than that of \( \log Q_0 \) according to equations (A8) and (A14).

Appendix B: Adaptations of the Back Projection Method

\[\text{[31]}\] In this study we solve for the laterally varying \( Q_m \) model using the iterative back projection method \cite[e.g.,][]{Humphreys and Clayton, 1988} used by \textit{Xie and Mitchell} [1990a]. We choose this method because equation (6) is a special, 1D case of equation (11) of \textit{Xie and Mitchell} [1990a], who dealt with a 2D case. Extensive computer programs for the back projection calculations to obtain the \( Q_m \) model and its resolution and random errors have already been implemented for the 2D case by \textit{Xie and Mitchell} [1990a, equations (11)–(16)] and \textit{Zhao and Xie} [1993]. The adaptations of these programs to the current 1D case are minor. The first adaptation is that, for interiteration smooth-
[16] As in previous 2D cases, after inverting for the $Q_m$ structure, its lateral resolution is approximately estimated by calculating “point spread function (p.s.F)” [Humphreys and Clayton, 1988; Xie and Mitchell, 1990a], which is more easily done in the 1D case. The random errors associated with the final $Q_m$ values, caused by random errors associated with the input $Q_m$ measurements, are estimated in this study by a numerical simulation that is similar to that described by Xie and Mitchell [1990a, p. 165], with one modification described in the following. In the simulation procedure by Xie and Mitchell [1990a], a random noise series of $N$ members was generated by computer, with the specification that the $n$th member has a randomly generated sign and an absolute value $\delta Q_m$, with $\delta Q_m$ being the known standard error associated with the $Q_m$ measurement. This randomly generated noise series was then added to the $Q_m$ values to generate a new input to the back projection inversion of $Q_m$ yielding in an estimate of the random errors in $Q_m$. This simulation procedure was repeated as many as 20 times to stabilize the estimated error in $Q_m$. In the current study all simulation steps are the same as those used by Xie and Mitchell [1990a], except that the $n$th member of the computer-simulated noise is generated to follow a Gaussian noise with a zero mean and a standard deviation of $\delta Q_m$. The modification is based on the consideration that the random errors in $Q_m$ may be distributed closer to a Gaussian noise than to a two-status ($\pm \delta Q_m$) distribution.

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