GEOPHYSICAL RESEARCH LETTERS, VOL. 27, NO. 23, PAGES 3949-3952, DECEMBER 1, 2000

Frequency Dependent *Lg* Attenuation in South-Central Alaska

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Abstract. The characteristics of seismic energy attenuation are determined using high frequency *Lg* waves from 27 crustal earthquakes, in south-central Alaska. *Lg* time-domain amplitudes are measured in five pass-bands and inverted to determine a frequency-dependent quality factor, *Q(f)*, model for south-central Alaska. The inversion in this study yields the frequency-dependent quality factor, in the form of a power law: $Q(f) = Q_0 f^{\eta} = 220(\pm 30) f^{0.66(\pm 0.09)}$ ($0.75 \leq f \leq 12$Hz). The results from this study are remarkably consistent with frequency dependent quality factor estimates, using local *S*-wave coda, in south-central Alaska. The consistency between *S*-coda *Q(f)* and *Lg Q(f)* enables constraints to be placed on the mechanism of crustal attenuation in south-central Alaska. For the range of frequencies considered in this study both scattering and intrinsic attenuation mechanisms likely play an equal role.

1. Introduction

Regional differences in the attenuation of seismic waves were recognized prior to the advent of modern instrumented recordings. For example, shaking intensities of earthquakes in the western United States were observed to decrease at a faster rate, with epicentral distance, than those from comparable-sized earthquakes in the central and eastern United States [*Richter*, 1958, *Nuttli et al.*, 1979]. Direct phase and coda amplitude measurements, using modern digital seismic instruments, have confirmed that attenuation in the western United States is significantly higher than in the eastern United States [*Mitchell*, 1975, 1981; *Frankel et al*, 1990; *Benz et al.*, 1997]. In fact, global observations confirm that *S*-wave coda and *Lg* attenuation is higher for regions with active tectonism than for stable continental interiors [*Aki*, 1980a, 1980b; *Benz et al.*, 1997; *McNamara et al.*, 1996]. This study focuses on the frequency-dependence of the *Lg* wave quality-factor in south-central Alaska. *Lg* is commonly observed as the dominant phase on high-frequency, regional distance seismograms and provides a good measure of path-averaged crustal properties, such as shear-wave velocity and attenuation.

Lg amplitude is sensitive to lateral heterogeneity in the crust due to varying tectonic environments. In stable continental regions, such as northern Africa, *Lg* is observed at distances as great as 6000km [*McNamara and Walter*, 2000]. In contrast, when propagating through active tectonic regions, such as the Himalaya and Tibetan Plateau [*McNamara et al.*, 1996], or ocean basins [*Zhang and Lay*, 1995], *Lg* is completely attenuated due to scattering along tectonic faults and variations in the thickness of the crustal waveguide [*Kennett*, 1986].

The crustal properties of south-central Alaska, affecting the attenuation characteristics of *Lg* are a direct result of the continuing north-south convergence of the Pacific and North American tectonic plates. Major crustal-seismotectonic characteristics of the region include: active volcanism of the Cook Inlet and the Wrangell-St. Elias Range [*Nye*, 1999] (Figure 1), abundant intraplate seismicity [*Ratchkovsky et al.*, 1998; *Biswas and Tytgat*, 1988], active strike-slip motion along the Denali fault (~6mm/yr) [*Fletcher and Freymueller*, 1998], and the high elevations of the Alaska Range (Figure 1).

Numerous researchers have estimated frequency dependent *Lg* attenuation for continental crust throughout the world. The frequency-dependent quality factor, *Q(f)*, is commonly modeled using a power law of the form:

$$Q(f) = Q_0 (f/f_0)^{\eta}, \qquad (1)$$

where f_0 is a reference frequency (generally 1 Hz), Q_0 is Q at the reference frequency, and η is assumed constant over the frequencies of interest. The *Q(f)* function can vary significantly depending on the tectonic activity of the region. For example, several studies in the tectonically active Basin and Range province have documented low Q_0 and strong frequency dependent attenuation (e.g. $Q(f)=214f^{0.54}$, *Chavez and Priestly* [1986]). In contrast, numerous studies in the tectonically stable areas of the central and northeastern United States and eastern Canada found higher values of Q_0 and a much weaker frequency dependent attenuation (e.g. $Q(f)=670f^{0.33}$, [*Atkinson and Mereu*, 1992]; $Q(f)=1052f^{0.22}$, [*Benz et al.*, 1997]). In south-central Alaska, *Steensma and Biswas* [1988] noted a strong frequency dependent attenuation, from *S*-wave coda decay, measured on short period seismograms. Theoretical studies indicate that coda *Q* may not be directly comparable to *Lg Q* in strongly scattering media [*Frankel and Wennerberg*, 1987]; however, good correlation between coda *Q* and *Lg Q* estimates have been observed in regions such as the Tibetan Plateau [*Xie and Mitchell*, 1991; *McNamara et al*, 1996]. This study presents the first estimate of *Lg* frequency dependent *Q* in south-central Alaska using broadband instrumentation. Results from this study will be compared to previous *Lg* and *S*-wave coda frequency dependent *Q* studies to obtain further understanding of the attenuation properties of the crust in south-central Alaska.

2. Data Selection Criteria

The *Lg* waveforms used in this study were obtained from local and regional crustal earthquakes that occurred in central Alaska, and were digitally recorded as a part of the Broadband Experiment Across the Alaska Range (BEAAR) [*Meyers et al.*, 1999] (Figure 1). The sensor at each site was a broadband, active-feedback, three-component, Guralp CMG-3ESP that is flat to velocity from 0.30 to 50 Hz. Data were continuously recorded at a rate of 50 samples per second. Earthquake locations and magnitudes used in this study were obtained from the Alaska Earthquake Information Center (AEIC), June 1999, catalog of hypocenters [*McNamara et al.*, 1999] (Figure 1). All phase picks were reviewed and hypocenters recalculated to minimize travel time residuals. For this analysis *Lg* waveforms were restricted to

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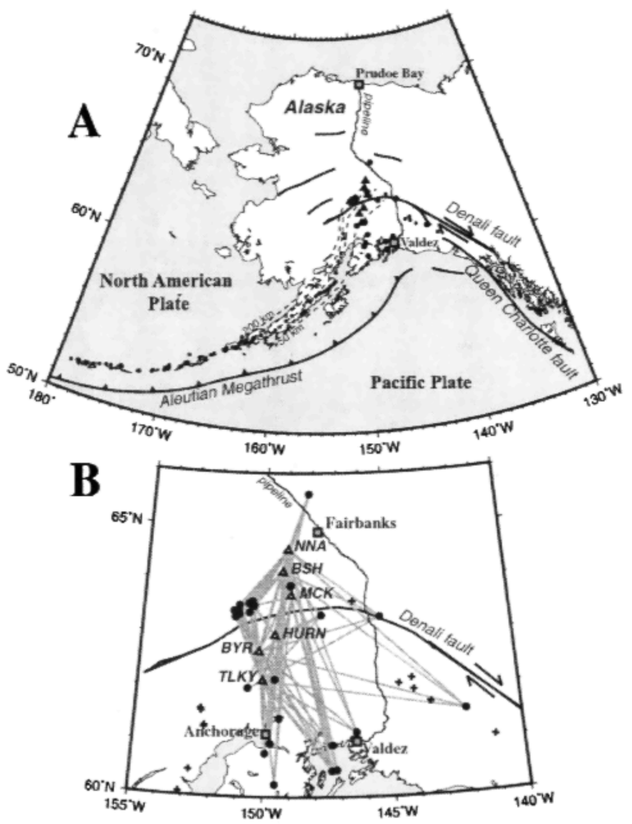

Figure 1: (a) Map of Alaska showing the BEAAR broadband stations (white triangles) and the distribution of regional events (solid circles) used in this study. The BEAAR array crosses major tectonic features including the Denali fault and the Alaska Range. Dashed lines are Pacific Plate depth contours. (b) *Lg* raypaths, passing the data selection criteria and used in the inversion for frequency-dependent quality factor, *Q(f)*. Black crosses are the locations of volcanoes.

well recorded (Ml≥2.0), crustal events (depth≤40 km, [*Biswas and Tytgat*, 1988]), with paths confined to south-central Alaska. For inclusion in the inversion, we also required that each station recorded at least two events and that each event be recorded by at least two stations.

A two step signal to noise analysis was performed to obtain high quality *Lg* amplitude measurements. First, the root-mean-square (RMS) amplitude of the first arriving body wave (*Pn*, windowed from 7.8-8.2km/s, or *Pg*, windowed from 5.8-6.5km/s), on the vertical component, velocity seismogram was required to be ten times greater than the RMS amplitude of the pre-event noise. The pre-event noise window (13.0-11.0km/s) was a similar length, and well ahead of the *P*-wave window. This step eliminated small and poorly recorded events. Second, we required an *Lg/P-coda* RMS amplitude ratio greater than two. *Lg* was windowed from 3.6-3.0 km/s and the *P-coda* (windowed from 5.8-4.8 km/s) was chosen as the scattered energy between faster crustal *P*-waves (*Pg*) and slower upper mantle *S*-waves (*Sn*). This step eliminated very few paths, since *Lg* is generally the dominant arrival on regional seismograms, and ensured that only amplitude measurements are included in the inversion where *Lg* is present at all distances. The goal was to eliminate paths crossing *Lg* blocking structures, such as ocean basins, that would bias a regional *Q* estimate for the continental crust.

Regional waveforms, that passed the signal-to-noise criteria were further processed to obtain *Lg* time-domain amplitude measurements. This included deconvolving the instrument response from the bandpass-filtered vertical component seismogram. The vertical component was selected for comparison with previous studies, that used only short-period vertical sensors, [*Hansen et al.*, 1998; *Steensma and Biswas*, 1988] and is supported by previous studies showing that energy in the *Lg* window is evenly scattered across all three-components of motion [*McNamara et al.*, 1996]. Finally, the RMS amplitude of *Lg* (3.6-3.0 km/s) was measured in five, one-octave, passbands, with center frequencies of 0.75, 1.5, 3, 6 and 12 Hz. Figure 2 shows an example of the filtered time domain *Lg* arrivals, from a single event (origin time 6/25/1999, 11:34:30.16, Ml=4.0, depth=9.35km, lat=63.447°, lon=-151.297°) recorded at station TLKY (dist=157km).

After applying the earthquake and waveform selection criteria to over 2000 regional seismograms, 105 high-quality *Lg* waveforms, from 27 events, recorded at six stations, with a distance range of 75-500 km, remained. Due to the restricted raypath coverage, the *Q(f)* determined in this study is only representative of the south-central portion of Alaska (Figure 1b).

3. Inversion Methods and Results

The inversion method used in this study to estimate the frequency dependence of *Lg* is described in detail by *Benz et al.*, [1997] and further implemented by *McNamara et al.*, [1996]. The observed *Lg* amplitude, *A*, at frequency *f* for the *j*th earthquake recorded at the *i*th station can be modeled as:

$$A_{ij}(f) = R_{ij}^{-\gamma} S_j(f) G_i(f)\, e^{-\pi f R_{ij}/Q\beta} \qquad (2)$$

where $S_j(f)$ is the source spectra, $G_i(f)$ is the site amplification, R_{ij} is the epicentral distance between the earthquake, *j*, and station, *i*, γ is the exponent for geometrical spreading, 0.5 in this study [*Benz et al.*, 1997], *Q* is the *Lg* quality factor at frequency *f*, and β

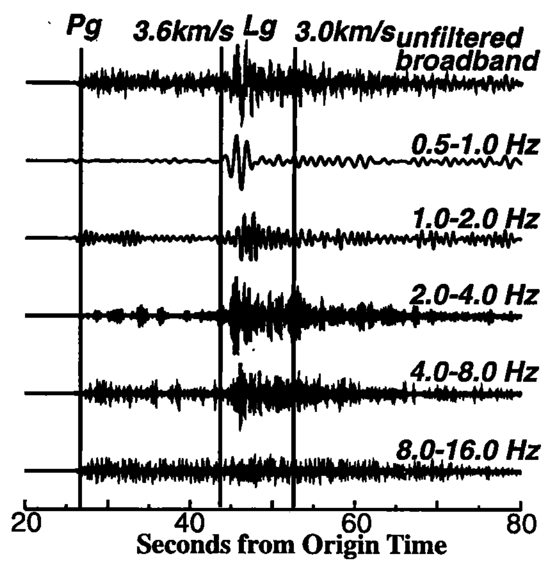

Figure 2: *Lg* from one earthquake (origin time 6/25/1999, 11:34:30.16, Ml=4.0, depth=9.35km, lat=63.447°, lon=-151.297°) recorded at station TLKY (dist=157km). Shown is the vertical component of the instrument corrected displacement seismogram. The seismogram is filtered in five passbands for amplitude measurement. Filtered traces are plotted on the same scale.

is the average shear-wave velocity for the crust, 3.5 km/s for this study. Taking the logarithm of (2) yields the following equation:

$$\ln A_{ij}(f) + \gamma \ln R_{ij} = \ln G_i(f) + \ln S_j(f) - \pi f R_{ij}/Q\beta. \quad (3)$$

When the left hand side of (3) is plotted with respect to distance, the right side of (3) describes a line where the receiver (G_i) and source (S_j) terms control the intercept and the Q term controls the slope. Using a data set with many source-receiver pairs, a system of linear equations can be set up based on equation (3). The system of equations is then solved using a singular-value decomposition inversion [e.g. *Menke*, 1990]. The inversion solves for both the source (S_j) and receiver terms (G_i) as well as a regionally averaged *Lg Q* for a single frequency passband, with center frequency, *f*.

By repeating the inversion, over five octaves, we obtain a *Q* estimate for each passband. Figure 3 shows the south-central Alaska *Lg* amplitudes corrected for the source (S_j) and receiver (G_i) terms plotted versus distance. The straight lines represent the best fitting *Q* for the particular frequency band. *Q* in south-central Alaska ranges from 198 at 0.75 Hz to 1190 at 12 Hz. A weighted least-squares regression analysis is then used to fit the frequency-dependent *Q* function, *Q(f)* (1), to the *Q* estimates. Taking the logarithm of both sides of (1) yields:

$$\ln Q = \ln Q_0 + \eta \ln(f) \quad (4)$$

where $\ln Q_0$ and η are the unknowns to be determined. The least-squares fit to the south-central Alaska *Lg Q* estimates, shown in Figure 4 (AKLg), is given in the form of a power law by:

$$Q(f) = 220(\pm 30) f^{0.66(\pm 0.09)} \quad (0.75 \le f \le 12 \text{ Hz}). \quad (5)$$

4. Discussion

Figure 4 shows apparent *Q(f)* functions, from several previous studies, obtained for a variety of tectonic regions. Tectonically stable continental regions, such as the northeastern United States, generally have the highest values of Q_0, and the weakest frequency dependence, (low η) (Figure 4; NEUS), while low Q_0 and high η values are observed in tectonically active regions (Figure 4; BRP, TP). As expected for south-central Alaska, the Q_0 and η determined from this study are more indicative of a tectonically active region.

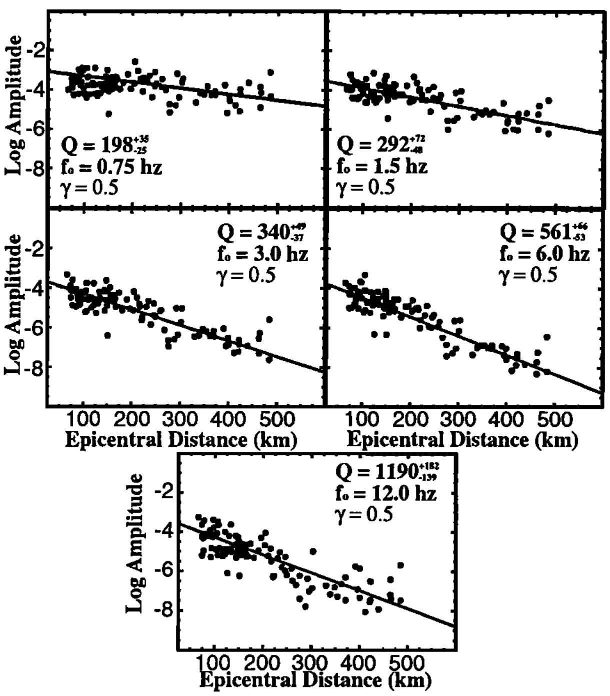

Figure 3: Best fit *Q*, determined from the inversion, over 5 different passbands.

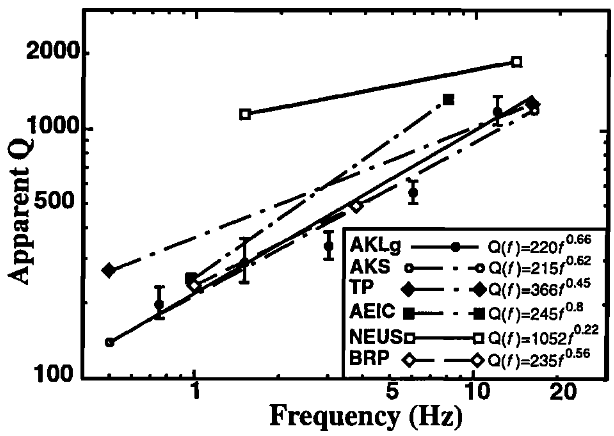

Figure 4: The weighted least-squares *Q(f)* fit (AKLg, solid line) to the *Q* estimates from this study (solid circles). Also shown are *Q(f)* estimates from Alaska, using different techniques (AKS) [*Steensma and Biswas*, 1988] (AEIC) [*Hansen et al.*, 1998], and several different regions for comparison, including the northeastern United States (NEUS) the Basin and Range (BRP) [*Benz et al.*, 1997] and the Tibetan Plateau (TP) [*McNamara et al.*, 1996].

An earlier attempt at characterizing the *Lg* quality factor in south-central Alaska was presented by *Hansen et al.*, [1998] (Figure 4 AEIC). The Q_0 from this study is comparable to the results of *Hansen et al.*, [1998] however, the frequency dependence is weaker (Figure 4). This difference is likely due to the limited data available to *Hansen et al.*, [1998]. Specifically, their study was only able to measure *Lg* amplitudes on short period (1Hz), vertical component seismograms within a limited epicentral distance range (100-300km).

A remarkably similar estimate of *Q(f)* in south-central Alaska was obtained by *Steensma and Biswas* [1988]. They measured the decay of *S*-wave coda and determined a frequency-dependent *Q* (Q_0=215, η=0.62) that falls well within the uncertainty of results obtained in this study (Figure 4, AKS). *Steensma and Biswas* [1988] measured *S*-wave coda decay from local events with focal depths ranging from 6-129km and event-station distances ranging from 2-435km. While shallower events (≤40km) with greater epicentral distances (75-500km) are used in this study, the results are comparable.

Seismic attenuation is generally caused by a combination of both scattering and intrinsic mechanisms. Scattering redistributes wave energy within the medium but does not remove energy from the overall wavefield. In contrast, intrinsic attenuation mechanisms convert wave energy to heat through friction, viscosity, and thermal relaxation processes. It is not yet clear which attenuation mechanism is dominant and the results from this study cannot independently distinguish between the two. The *Lg Q(f)* estimate, obtained from this study, is a combination of both intrinsic and scattering mechanisms and defines an apparent frequency-dependent attenuation for the whole crust.

The consistency of these results, *Lg Q(f)*, and the *S*-coda *Q(f)* results of *Steensma and Biswas* [1988], can allow some constraints to be placed on attenuation mechanisms. For example, several studies demonstrate that seismic coda is relatively in-sensitive to scattering, suggesting that intrinsic rock properties likely dominate attenuation [*Mayeda and Walter*, 1996; *Frankel and Wennerberg*, 1987]. *Mayeda et al.*, 1992 more precisely define the contribution of the individual attenuation mechanisms in a broad study of coda Q in Hawaii, Long Valley and central California. They conclude that, in all three regions, for frequencies less than or equal to 6.0 Hz, scattering dominates attenuation, whereas above 6.0 Hz, intrinsic rock properties are the dominant mechanism. Given this observation and the consistency between *S*-coda $Q(f)$ and Lg $Q(f)$, for the range of frequencies considered in this study (0.75-12Hz), both scattering and intrinsic mechanisms likely play and equal role in attenuating Lg energy.

South-central Alaska is currently overriding the active subduction of the Pacific Plate. Several related geologic factors may contribute to the apparent Lg Q within the crust of south-central Alaska. First, Lg amplitudes (≤6 Hz) likely decrease due to scattering along fractures, faults and sutures. The crust of south-central Alaska is highly fractured and faulted, consisting of numerous suture zones, bounding tectonic terranes, and large strike-slip faults, along which the crust is actively deforming [*Fletcher and Freymueller*, 1998; *McNamara et al.*, 1999]. Second, in regions with active crustal deformation and tectonic activity, the intrinsic rock properties of the crust may cause attenuation of Lg energy at frequencies >6 Hz. For example, interstitial fluids may be present in crustal fractures, from melt and metamorphism due to elevated temperatures from internal deformation and/or heating from the upward migration of volcanic material.

The thermal properties of south-central Alaska are not well known, however, there are several seismic and geologic observations that can be interpreted as evidence for high temperatures in the upper mantle and crust. For example, low upper mantle and crustal *P*-wave velocities [*Zhao et al.*, 1995] are consistent with high heat production. There is also an abundance of recent volcanism, extending from the Cook Inlet, into the Alaska range and eastward to the St. Elias range (Figure 1) [*Nye*, 1999]. These flows are due to melt from the subducting Pacific plate and would significantly heat the overlying crust of south-central Alaska. Crustal heating is likely to significantly increase crustal attenuation (*Frankel et al.*, 1990).

The goals of this paper are to document the frequency-dependent Lg Q of south-central Alaska and place some constraints on the mechanisms of attenuation. Also, results from this work, provide an attenuation function that can be used in a variety of scientific and engineering applications including local magnitude estimates, earthquake hazard assessment in populated areas and understanding the shaking intensity expected at structures, such as the Alaska Pipeline.

Acknowledgements. This paper benefited from comments by K. Mayeda, J. Lahr, M. Peterson, K. Shedlock and an anonymous reviewer. D.M expresses gratitude to AEIC, analysts: L. Rao, N. McGreggor, E. Clark, and M. LaFevers, who located the hypocenters used in this study, R. Hansen, K. Lindquist, and summer students E. Van Ark and J. Rubenstein who assisted in the preliminary research that led to these results. D.M would especially like to thank Liz Meyers, Doug Christensen and Geoff Abers for the successful deployment of the BEAAR array, and for early access and use of this very high quality seismic data. The BEAR project was supported by NSF grant EAR 9725168. This project was supported by USGS grant 98-HQ-GR-1007 to the Geophysical Institute, University of Alaska Fairbanks. Maps were creating using the GMT software [*Wessel and Smith*, 1991].

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(Received May 1, 2000; revised July 25, 2000; accepted September 26, 2000.)