Investigation of Attenuation of the Lg-Wave Amplitude in the Caribbean Region

by M. Hosseini,* S. Pezeshk, A. Haji-Soltani, and M. Chapman

Abstract The focus of this study is to determine the frequency-dependent quality factor function \( Q(f) \) for the Caribbean region. The analysis considers the Lg portion of 2685 three-component waveforms. Waveforms are selected from 116 earthquakes that occurred between 2006 and 2013 with moment magnitude \( M_w \) ranging from 4.6 to 7.0. Spectral amplitudes over 12 distinct passbands from 0.1 to 12.8 Hz are calculated only for waveforms with a signal-to-noise ratio of 5 or better. In the regression model, the vertical component and the geometric mean of two horizontal components are used to estimate \( Q(f) \). A geometrical spreading function with spectral amplitude decay of \( R^{-0.5} \) is used for distances beyond 100 km. The following quality factor functions for the assumed geometrical spreading are obtained: \( Q_H = 310 f^{0.54} \) for the horizontal components, and \( Q_V = 235 f^{0.65} \) for the vertical components.

Introduction

Currently, there are limited ground-motion prediction equations (GMPEs) for the southeastern United States and the northern Caribbean region. One possible approach for developing GMPEs for this region is to estimate ground motions by using a stochastic procedure (Atkinson and Boore, 1995, 1998, 2006; Frankel et al., 1996; Toro et al., 1997; Boore, 2003; Pezeshk et al., 2011). Critical to any stochastic simulation is the selection of seismological input parameters such as the frequency-dependent quality factor function \( Q(f) \). The purpose of this study is to determine the frequency-dependent quality factor function \( Q(f) \) for the Caribbean region. The shaking intensity of earthquakes and instrumental seismic recordings in the different tectonic environments show that areas of active tectonics, like the Caribbean and the western United States (WUS) regions, have higher attenuation (lower quality factor) of seismic waves than the stable continental regions such as the central and eastern United States (CEUS) regions (Aki, 1980a,b; Singh and Herrmann, 1983; Frankel et al., 1990; Benz et al., 1997; Erickson et al., 2004; Zandieh and Pezeshk, 2010; Zhou et al., 2011; McNamara et al., 2012). McNamara et al. (2012), based on studies by Aki (1980a,b), Gregersen (1984), Frankel (1991), and Benz et al. (1997), suggested these observations from different tectonic regions indicate a highly fractured crust in tectonically active regions that absorb high-frequency seismic waves, differences in crustal temperature, and variations in crustal structure.

Lg waves carry the most prominent energy for continental paths at regional distances (Båth, 1954). Lg was first identified as surface waves (Press and Ewing, 1952), which are \( S \) waves trapped in the crustal waveguide. The amplitude of Lg waves in the continental crust is a function of crustal structure and the physical properties of the crustal material (Mitchell, 1995). Lg waves are attenuated more rapidly in active tectonic regions, in contrast with the stable tectonic blocks (Aki, 1980a,b; Zhou et al., 2011). The dominant frequency of Lg is in the range of 0.5–5.0 Hz, with group velocity of approximately 2.8–3.7 km/s.

The Caribbean study region is made up of over 7000 islands, islets, coral reefs, and cays. The independent countries of the region are Cuba, Dominican Republic, Haiti, the Bahamas, Jamaica, and Trinidad and Tobago. The Caribbean lithospheric plate mainly consists of an anomalously thick, oceanic plateau located between two major continental regions. It is a geologically complex region that exhibits a variety of plate boundary interactions, including subduction (Lesser Antilles and central America) and strike slip on northern and southern boundaries, and seafloor spreading in the Cayman Trough (Mattson, 1977; Jackson, 2002). The Port-au-Prince region of Haiti was struck by an earthquake of \( M_w \) 7.0 on 12 January 2010. Damages caused by this disastrous earthquake, in which more than 200,000 people were killed, were estimated to be around $8 billion (Calais et al., 2010).

Many studies have been performed to evaluate the attenuation characteristics of seismic waves in various regions of the world (e.g., Nuttli, 1973; Mitchell, 1975; Bollinger, 1979; Chen and Pomeroy, 1980; Nicolas et al., 1982; Atkinson and Mereu, 1992; Atkinson and Boore, 1995; Benz et al., 1997; Atkinson, 2004; Allen et al., 2007; Zandieh and Pezeshk, 2010). A number of attenuation studies have also been carried out for the Caribbean region. Molnar and Oliver (1969) investigated the average attenuation of high-frequency \( S_n \) shear waves propagating across the concave side of the

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Antilles are for both oceanic and continental crust. They reported anomalous propagation of shear waves near the Lesser Antilles. Rial (1976) estimated the shear-wave $Q$ for propagation paths through the anomalous zone to be about 400 for the entire path. According to Rial (1976), seismograms for paths through the anomalous region feature frequencies of 0.5 or less, and $Q$ is about 80 or less, agreeing with Molnar and Oliver (1969). Frankel (1982) reported $Q$ of about 400 for Rayleigh waves for Caribbean region. Ambe and Lynch (1993) investigated coda quality factor $Q_c$ for the eastern Caribbean. He reported $Q_c$ in the range from 152 to 239 at 1.5 Hz in the eastern Caribbean, increasing to approximately 1236–3455 at 16 Hz. Ambe and Lynch (1993) estimated $Q_0$ in the range of 97–145, with $\eta$ in the 0.82–1.09 range in the linear logarithmic regression $Q = Q_0 f^{\eta}$. Recently, McNamara et al. (2012) estimated $Q$ using approximately 850 observations of $Lg$ waves in the Hispaniola Island region. They used a hinged-trilinear geometrical spreading function consistent with Atkinson (2004) and Motazedian and Atkinson (2005). Odum et al. (2013) selected 27 sites as the representative of the near-surface material to study the site parameters in Puerto Rico. They compared their results with the observed data for the 16 May 2010 $M_w$ 5.8 Puerto Rico earthquake recorded in eight specific PRSMP stations. The stations are located at distances from 0.5 to approximately 12 km from the surveyed sites (Odum et al., 2013). A general geologic description of PRSMP stations and their National Earthquake Hazards Reduction Program site classifications can be found in Odum et al. (2013). Also, Odum et al. (2013) provide $V_{330}$ obtained by reflection/refraction (body wave) and refraction microtremor (surface wave) surveys at sites near PRSMP stations.

Data Selection and Preprocessing

For each earthquake, associated seismograms at a specific station are selected based on the availability of high-quality data. Only those seismograms recorded at hypocentral distances larger than 100 km and less than 1000 km are used for the regression analyses. Figure 2 illustrates the distribution of earthquakes in magnitude and distance. Waveforms are recorded by broadband stations from the Caribbean (CU), Cayman Islands (CY), Instituto Sismológico Universitario (DR), Global Seismograph (IU), and Puerto
Rico (PR) seismic networks. Waveforms are downloaded 30 s prior to and 600 s after the origin time of each earthquake, for all three components. A 30 s window before the event is used for signal-to-noise ratio control. The data preference is for broadband seismograms with high sample rates. Sampling rates are in the range of 20–100 points per second. The majority of dataset records have a sampling rate of 40 points per second and above. Based on the sampling rate of each specific waveform, the Nyquist frequency ($f_{\text{nyq}}$) beyond which Fourier amplitudes are not used) ranges from 10 to 50 Hz. Table 1 presents the Nyquist frequency for different stations. The majority of stations have a constant sampling ratio according to time series from events. In contrast, about a third of the stations have different Nyquist frequency values due to different sampling rate of time series from different events. Maximum and minimum of $f_{\text{nyq}}$ is provided for these stations.

A fast Fourier transform (FFT) is used to derive the amplitude at 12 frequency bands, centering on 0.25, 0.35, 0.5, 0.7, 1.0, 1.4, 2.0, 2.8, 4.0, 5.6, 8.0, and 11.2 Hz. The lower limit of the first frequency band starts at 0.2 Hz and the upper limit of the last frequency band ends at 12.8 Hz. The bandwidth doubles every two intervals. For example, the first frequency band covers 0.2–0.3 Hz, the second band covers 0.3–0.4, the third covers 0.4–0.6, and so on.

After applying the FFT to the time series, amplitudes are averaged for frequencies falling in each frequency band, and their average amplitude is reported as the amplitude associated with the center frequency of that specific frequency.

Figure 1. Maps of (a) the Caribbean region and (b) the study area, showing locations of earthquakes (stars) and broadband stations (inverted triangles). Only those earthquakes and stations in the Caribbean plate (inside the rectangle bordered by the dashed line in (b)) are used in this study. The color version of this figure is available only in the electronic edition.
band. Signal-to-noise considerations are implemented by considering noise in a 20 s window starting from 30 s prior to event time. The geometric mean of the two horizontal components is used along with the vertical component. FFT amplitudes for the noise window at the same 12 frequency centers are calculated and are compensated for the difference between data and noise window lengths. The data selection required signal-to-noise ratios of five or greater. FFT amplitudes are calculated for a data window capturing the \( Lg \) wave. Considering an \( Lg \) wavespeed of 3.50 km/s (McNamara et al., 2012), \( Lg \)-wave arrival is defined as

\[
T_{Lg} = T_0 + \frac{r}{3.50},
\]

in which \( T_{Lg} \) is the arrival time of the initial onset of the \( Lg \) phase, \( T_0 \) is the earthquake origin time, and \( r \) is the epicentral distance in kilometers. The duration window for the Fourier analysis of the \( Lg \) phase is defined by examination of the integral of the squared acceleration time series. The duration of the \( Lg \) window \( T_d \) is defined according to

\[
\int_{T_{Lg}}^{T_{Lg}+T_d} a^2 \, dt = 0.75 \int_{T_{Lg}}^{T_{Lg}+150} a^2 \, dt,
\]

in which \( a \) is the ground acceleration. The \( Lg \) signal duration \( T_d \) is defined as the time at which the integral of the squared acceleration time series (starting at \( T_{Lg} \)) reaches 75% of its value at \( T_{Lg} + 150 \) s. In the FFT analysis, we use the \( Lg \) window from \( T_{Lg} \) to \( T_{Lg} + T_d \) to obtain the amplitudes in 12 frequency bands.

Figure 3 shows the location of stations AGPR and MPR, which recorded an event marked by the star. Seismograms from the 4 February 2008 earthquake with a reported magnitude of 5.5 recorded by these two stations are illustrated in Figure 4. At each station, all three components (BHE, BHN, and BHZ in order from top to bottom) are plotted, along with an \( Lg \)-wave window represented with vertical lines.

We performed data analysis separately for the geometric mean of amplitudes for the two horizontal components, as well as amplitudes for the vertical component, and each are reported separately. The next section provides details on the data analysis for the path effect study.

### Data Analysis

Following Atkinson and Mereu (1992) and Zandieh and Pezeshk (2010), the spectral amplitude generated at the earthquake hypocenter (source amplitude) travels across the path between the source and the location of the recording seismograph. The source amplitude undergoes two major changes, one resulting from the path effect and the other from the local site geology at the location of the seismograph. The path effect is modeled by a combination of geometrical spreading and anelastic attenuation functions. Local site geology may amplify or deamplify the amplitude. The observed spectral amplitude is given by:

\[
\log|O_{i,j}(f)| = \log[A_i(f)] - B(R_{i,j}) \log(R_{i,j})
\]

\[
- \log(e) \pi f Q(f) \times \rho R_{i,j} + \log[S_j(f)],
\]

in which

<table>
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<th>Station Name</th>
<th>( f_{nyq} ) (Hz)</th>
<th>Station Name</th>
<th>( f_{nyq} ) (Hz)</th>
<th>Station Name</th>
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in which $O_{i,j}(f)\) is the observed spectral amplitude of earthquake $i$ at station $j$ at frequency $f$; $A_i(f)$ is the source spectral amplitude of earthquake $i$ at unit hypocentral distance; $B(R_{i,j})$ is the geometrical spreading coefficient; $R_{i,j}$ is the hypocentral distance; $e$ is the Napier’s constant (2.7183); $Q(f)$ is the quality factor, which is a function of the frequency; and $S_j$ is the site (receiver) term for station $j$. It should be noted that the source spectral amplitude at the hypocenter location is considered to be equal for all of the observations at different stations and that the site (receiver) term $S_j$ is independent of the event.

Geometrical Spreading

For a whole space, the concept of the geometrical spreading comes from the law of energy conservation where energy density on the surface of common-centered spheres with various diameters should decrease as the diameter increases. The geometrical spreading term, $B(R_{i,j})\log R_{i,j}$ defines the logarithmic decay of amplitude at a specific frequency. Atkinson and Mereu (1992) modeled the geometrical spreading function using a hinged-trilinear functional form, in which the decay rate is different in three distance segments. The hinged-trilinear functional form of the geometrical spreading used here is given by

$$B(R_{i,j}) \log(R_{i,j}) = \begin{cases} b_1 \log R_{i,j} & R_{i,j} \leq R_1 \\ b_1 \log R_1 + b_2 \log R_{i,j}/R_1 & R_1 \leq R_{i,j} \leq R_2 \\ b_1 \log R_1 + b_2 \log R_2/R_1 + b_3 \log R_{i,j}/R_2 & R_{i,j} \geq R_2 \end{cases}$$

Motazedian and Atkinson (2005) and McNamara et al. (2012) used $b_1 = 1.0$, $b_2 = 0.0$, and $b_3 = 0.5$ with hinge points $R_1 = 75 \text{ km}$ and $R_2 = 100 \text{ km}$. Our data is at distances greater than 100 km, and we assume the same model for geometrical spreading.

System of Equations

Rearranging equation (3) by considering a known geometrical spreading gives

$$\log[O_{i,j}(f)] + B(R_{i,j}) \log(R_{i,j}) = \log[A_i(f)] - \frac{\log(e)\pi f}{Q(f) \times t} R_{i,j} + \log[S_j(f)].$$

in which the left side consists of known parameters and the right side consists of unknown arguments. Equation (5) can be cast into a standard matrix formation,

$$Gm = d.$$

Equation (6) represents a typical linear inversion problem that can be solved using least squares, maximum-likelihood, or generalized inversion methods (e.g., Aki and Richards, 1980; Menke, 1984; Lay and Wallace, 1995; Aster et al., 2013). The matrix $G$ is an $m \times n$ forward operator matrix; $n$ is the number of unknowns (source terms, receiver terms, and the quality factor) and $m$ is the number of observations. Such a system of equations has a unique solution when the number of observations ($m$) is more than or equal to the number of unknowns ($n$). The solution for $m$ is found using the singular value decomposition (SVD) procedure. The matrix $G$ can be expressed as the multiplication of three matrices:
\( G = U S V^\dagger \),

in which \( S \) is a diagonal matrix containing singular values of the matrix \( G \) on its diagonal and has the same size as \( G \). Matrices \( U \) and \( V \) are \( m \times m \) and \( n \times n \) unitary square matrices, and the columns of each of them form a set of orthonormal vectors. The prime superscript for \( V \) denotes the conjugate transpose. After finding the rank of the \( G \) matrix, its pseudoinverse can be calculated as

\[
G^{-\dagger} = V_k S_k^{-1} U^\dagger,
\]

in which the subscript \( k \) denotes the consideration of the rank of \( G \) in associated matrices, which includes removing problematic singular values from \( S \) and their associated columns from \( U \) and \( V \). Therefore, using the SVD procedure the vector \( m \) can be written as (Menke, 1984)

\[
m = G^{-\dagger} d.
\]

Based on equation (5), if the total number of earthquakes is \( p \) and the total number of stations is \( q \), then the matrices in equation (6) can be written as

\[
G = \begin{bmatrix}
1 & 0 & \ldots & 0 & 1 & \ldots & 0 & 0 & -\log(e)\pi f R_{11}/\beta \\
1 & 0 & \ldots & 0 & 0 & 1 & \ldots & 0 & -\log(e)\pi f R_{12}/\beta \\
\vdots & \vdots & \ddots & \vdots & \vdots & \ddots & \vdots & \vdots & \vdots \\
0 & 0 & \ldots & 1 & 0 & 0 & \ldots & 1 & 0 & -\log(e)\pi f R_{p(q-1)}/\beta \\
0 & 0 & \ldots & 1 & 0 & 0 & \ldots & 1 & 0 & -\log(e)\pi f R_{pq}/\beta \\
\end{bmatrix}_{pq \times (p+q+1)}
\]

\[
m = \begin{bmatrix}
\log A_1(f) \\
\log A_2(f) \\
\vdots \\
\log A_{p-1}(f) \\
\log A_p(f) \\
\log S_1 \\
\log S_2 \\
\vdots \\
\log S_{q-1} \\
\log S_q \\
\end{bmatrix}_{(p+q+1) \times 1}, \text{ and}
\]

\[
d = \begin{bmatrix}
\log [O_{11}(f)] + B(R_{11})\log(R_{11}) \\
\log [O_{12}(f)] + B(R_{12})\log(R_{12}) \\
\vdots \\
\log [O_{p(q-1)}(f)] + B(R_{p(q-1)})\log(R_{p(q-1)}) \\
\log [O_{pq}(f)] + B(R_{pq})\log(R_{pq}) \\
\end{bmatrix}_{pq \times 1},
\]

Equations (9) and (10) are the basic equations for our inversion when the geometrical spreading term is known. Each row of the forward operator \( G \) refers to an individual observation. The first \( p \) columns are related to earthquakes; columns \( p+1 \) to \( p+q \) address the receiver terms; and the very last column with index \( p+q+1 \) is related to the frequency-dependent attenuation term. In an inversion problem, eigenvalues of \( G \) affects the stability of the inversion and the accuracy of the results, that is, the ratio of largest eigenvalue to the smallest one, which is called the condition number, is an indicator of stability of the inversion. The smaller the condition number, the better the accuracy will be. In the case of the current study, when source terms are considered known, the SVD technique might not improve the accuracy all the time. However, in case of unknown source terms, improvement was observed in most of the cases where the interplay between the source and site (receiver) terms are resolved effectively.

For example, in the inversion for the quality factor at the 8.0 Hz frequency for the vertical component, \( G \) is a 145 \times 16 matrix (145 observed data and 16 unknown model parameters); and, after SVD, the eigenvalues are plotted in Figure 5.

For the current example, the results of the SVD technique become the same as those from the least square solution (Fig. 6).

**Source Model**

The source acceleration Fourier amplitude spectrum is defined as (Brune, 1970, 1971; Boore, 1983, 2003)

\[
A(f) = \frac{R_{\delta q} F V M_0 (2\pi f)^2}{4\pi \rho \beta^2} \left( 1 + \left( \frac{f}{f_0} \right)^2 \right),
\]

in which \( M_0 \) is the seismic moment; \( R_{\delta q} \) is the radiation pattern average value of 0.55 for shear waves; \( F \) is the free-surface amplitude amplification equal to 2; \( V \) is the coefficient for partitioning into two horizontal components, \( 1/\sqrt{2} \); and \( \rho \) is the density, assumed to be 2800 kg/m\(^3\) (Boore, 1983, 2003). The parameter \( f_0 \) is the source corner frequency given by

\[
f_0 = 4.906 \times 10^3 \beta \left( \frac{\Delta \sigma}{M_0} \right)^{1/3},
\]

in which \( \beta \) is the shear-wave velocity at the source (taken as 3.51 km/s for this study) and \( \Delta \sigma \) is the stress drop. The quality factor estimation is performed utilizing a wide range of stress drops (100–600 bars with 100 bars interval). It was observed that the assumed value of stress drop has negligible effect on the estimated quality factors at all frequencies; therefore, a typical stress drop of 100 bars is selected for all events.
Results and Discussion

Using equations (5)–(12), quality factors for vertical and horizontal components are estimated. Figure 7 illustrates the obtained frequency-dependent $Q(f)$; a straight line in logarithmic scale with equation $Q = Q_0 f^n$ is fitted to the quality factor estimates. The resulting equations are $Q^H = 310 f^{0.54}$ for the horizontal component, and $Q^V = 235 f^{0.65}$ for the vertical component.

To make a visual comparison of the observed data with the model obtained for the path effect, we reorder equation (5) to estimate the observed path effect. The observed path effect is derived by removing the source and site (receiver) terms from recorded amplitudes. The observed path effect is referred to as normalized amplitudes by Atkinson and Mereu (1992) and Zandieh and Pezeshk (2010), and is given by

$$\log[BC\_OBS(f, R_{i,j})] = \log[A_i(f)] - \log[O_{i,j}(f)] + \log[S_j(f)].$$

(13)

in which $\log[BC\_OBS(f, R_{i,j})]$ is the observed path effect. The predicted path effect is calculated by the following equation, assuming a geometrical spreading function and an estimated quality factor for different frequencies:

$$\log[BC\_PRE(f, R_{i,j})] = B(R_{i,j}) \log(R_{i,j}) + \frac{\log(e) \pi f}{Q(f) \times \beta} R_{i,j}.$$  

(14)

in which $\log[BC\_PRE(f, R_{i,j})]$ is the predicted path effect and can be plotted along with the observed path effect given in equation (13). Furthermore, residuals are determined by

$$Res(f, R_{i,j}) = \log[BC\_OBS(f, R_{i,j})] - \log[BC\_PRE(f, R_{i,j})].$$

(15)

Figures 8 and 9 show observed and predicted path effects and residuals for frequencies of 1.0 and 4.0 Hz, associated with horizontal and vertical components. Other frequencies show the same trend and no irregular behavior is observed. There is no apparent trend in residuals versus distance; a straight line is fitted to the residuals, and the equation of the line, as presented in the residual plots, shows a minimum intercept and slope.

Summary and Conclusions

Only data with epicentral distances greater than 100 km were used. We assumed a geometrical spreading of $R^{-1.0}$ ($b_1 = 1.0$) for distances less than 75 km; for distances from 75 to 100 km, no decay is presumed ($b_2 = 0$); and $R^{-0.5}$ ($b_3 = 0.5$) is assumed beyond 100 km. The following quality factor function and geometric pairs are obtained: $Q^H = 310 f^{0.54}$ for the horizontal component, and $Q^V = 235 f^{0.65}$ for the vertical component.

Figure 10 shows the attenuation models in the Caribbean region and the surrounding region, ranging from Jamaica and Cuba in the west to Puerto Rico and the Lesser Antilles in the east, compared to those in the WUS and CEUS regions. Both WUS and the Caribbean region have higher attenuation (lower $Q$ factor) than the CEUS. Table 2 presents parameters used by various studies plotted in Figure 10. According to the observations, $Q_0$ in this study is close to those from the Hispaniola Island region (McNamara et al., 2012) and Basin and Range Province (Benz et al., 1997).

Hough and Anderson (1988) and McNamara et al. (2012) pointed out the attenuation properties of the $Lg$ phase differ from those of the direct $S$ wave because the $Lg$ phase samples the entire crust, including the deeper crust, which is likely to be characterized by lower attenuation (higher $Q$ factor), whereas the direct $S$ waves are more controlled.

Figure 6. Solution of the singular value decomposition technique compared with the results of the conventional least squares. Parameters are the natural logarithm of the site (receiver) terms obtained from horizontal components at 8.0 Hz.
Figure 7. The quality factor versus frequency for (a) horizontal and (b) vertical components.

Figure 8. Fit quality between the predicted and observed path effects (top) and trend of the residuals (bottom) for frequencies (a) 1.0 Hz and (b) 4.0 Hz for horizontal amplitudes. A line is fitted to the residuals, and its equation is provided to quantitatively investigate the trend of the residuals.
by the upper crust (lower $Q$ factor). Therefore, the $Q$-factor model developed by Motazedian and Atkinson (2005) for the Puerto Rico region using local earthquakes as deep as 200 km might give a higher $Q$ factor (lower attenuation) than the shallow crustal attenuation models in the Caribbean region.

**Data and Resources**

All waveforms used in this study are archived and available for download from the Incorporated Research Institutions for Seismology Data Management Center. The data are all from broadband seismometers. Seismograms and all station instrument responses were received automatically using Standing Order of Data software. All three components of the waveforms are utilized in the analysis and only those stations were selected that simultaneously possessed components in three directions. Correction for the instrument response was performed using a modified version of the Engineering Seismology Toolbox developed by Assatourians and Atkinson (2008). Processing and inversion was performed using an automated package developed at the University of Memphis, Department of Civil Engineering, as a part of the Next Generation Attenuation-East project.
Acknowledgments

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Table 2

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