Bulletin of the Seismological Society of America, Vol. 101, No. 5, pp. 2220–2230, October 2011, doi: 10.1785/0120100326

Coda Wave Attenuation for Three Regions of Georgia (Sakartvelo) Using Local Earthquakes

by Ia Shengelia, Zurab Javakhishvili, and Nato Jorjiashvili

Abstract Seventy-two local earthquakes were analyzed and the quality factor (Q_c) was calculated for three regions of Georgia using the single-scattering model in the frequency range of 1–32 Hz. Standard International Association of Seismology and Physics of the Earth's Interior (IASPEI) routines for the coda Q_c determination have been used. The coda window varied from 35 to 45 s and the lapse time was up to 63 s. The epicenter distances and the focal depths of the earthquakes were up to 50 km and 20 km, respectively. Local magnitudes ranged from 1.7 to 4.1. These earthquakes were recorded by three digital seismic stations equipped with broadband Guralp CMG40T seismometers. The Q_c values show strong frequency dependence for three regions. The value of Q_c at frequency 1 Hz varies from 35.1 ± 5.3 to 83 ± 8.2, and the frequency parameter, n, increases from 0.890 ± 0.062 to 1.095 ± 0.056 depending on the model of the coda generation and the considered region. Observed Q_c values reflect the average attenuation property of the whole crust beneath each region and are correlated with the seismicity pattern and geology in the studied regions.

Introduction

In the Earth, the scattering of seismic oscillations forms the tail of the seismogram (coda), which can be used to gain information about the path and the source. The seismic coda wave is caused by the scattering of seismic waves from numerous randomly distributed heterogeneities in the Earth's crust and the upper mantle (Aki, 1969; Aki and Chouet, 1975; Rautian and Khalturin, 1978). Seismic coda is a very useful and popular tool for solving many seismological problems, such as the structure of the Earth's crust and the mantle, the study of the physics of an earthquake source, seismic hazard estimation, and strong earthquake forecasting (Aki and Chouet, 1975; Rautian and Khalturin, 1978; Rautian *et al.*, 1981; Aki, 1985).

The first investigation of seismic coda was published in the 1960s (Aki, 1969). Aki and Chouet (1975) proposed the single back-scattering model to characterize coda waves and introduced seismic attenuation parameter coda Q or Q_c, which is the measure of the decay rate of coda waves within a certain frequency interval and is inversely proportional to the attenuation of the medium. Spatial and temporal variations of Q_c in the heterogeneous media were studied by numerous scientists in many regions around the world (e.g., Sato and Fehler, 1998). It was proven that the coda Q value is sensitive to the geological environment; in general, it is lower in tectonically active regions and higher in the stable regions. Jin and Aki (1988) state that Q_0 (coda Q_c at a frequency of 1 Hz) can be a useful parameter to specify the seismicity of the regions.

Scientific works show that the decay rate of coda amplitudes of small earthquakes, with an epicentral distance up to 100 km, is independent of the recording site of the path and source-to receiver distance, but depends on the frequency and the lapse time (the lapse time is defined as time elapsed after the origin time) (Aki and Chouet, 1975; Sato, 1977; Rautian *et al.*, 1981). The increase of coda Q_c with lapse time can be due to several reasons (Woodgold, 1994) such as nonzero source–receiver distance, nonisotropic scattering, multiple scattering, scattering from below the crust, and the increase of Q_c with the depth or the distance from both the source and the receiver.

The decay of the seismic energy in the coda wave is due to geometric spreading, scattering, and intrinsic attenuations. Q_c is a combination of scattering quality factor, Q_s (due to the scattering at heterogeneities in the Earth) and intrinsic quality factor, Q_i (due to the absorption by inelasticity of the medium). In general, the single-scattering model (Aki and Chouet, 1975; Sato, 1977) and the multiple-scattering model (Gao *et al.*, 1983; Wu, 1985; Fehler *et al.*, 1992) are used to estimate Q_c. The single-scattering models are based on the assumption that the scattering is a weak process and does not produce multiple scattering when primary waves encounter another scatterer. These models are applied worldwide because of their simplicity and the ease of application. In the single-scattering method the problem is the uncertainty of Q_c interpretation in terms of Q_s and Q_i. Several methods have been proposed to determine the relative contribution of scattering and intrinsic loss to the total attenuation. Wu (1985) introduced the radiative transfer theory for the multiple scattering problem. This theory enables the calculation of the energy envelopes of seismic waves, taking into account all orders of scattering; it can be used to separate the effects of scattering and intrinsic attenuation uniquely. Wu's method was improved later to the multiple lapse time window analysis by Fehler *et al.* (1992). Frankel and Wennerberg (1987) developed the energy-flux model, which differentiates between the direct and scattered waves for coda waves, assuming that coda energy is uniformly distributed within a region. Gao *et al.* (1983) have concluded that the coda waves at the short lapse time (less than 100 s) are attributed mainly to the single scattering, whereas coda waves recorded at the long time (more than 100 s) are due to multiple scattering.

In Georgia, coda wave investigations began in the late 1970s; it was the first country in the Caucasus to undertake such study (Shengelia, 1981). Data from 30 analog seismic stations of the regional network were used to study attenuation of coda waves in the lapse time of 10–3000 s and to solve some other seismological tasks.

In the present study we estimate the quality factor Q_c for three regions of Georgia (Fig. 1) using the data of the local earthquakes at different frequencies. In order to get estimates of regional Q_c we used the single-scattering models. In the current analysis all lapse times are up to 63 s, therefore the use of the single-scattering model is acceptable. We received the relationships $Q_c = Q_0 f^n$ for all regions and compared them with other tectonic regions of the world. No other such study has been carried out so far in Georgia.

Tectonic Setting and Seismicity

The Caucasus is a part of the Alpine folded system and is located between the Black and the Caspian Seas. The depressed zones surrounding the Greater Caucasus are found along the seas, and the Alpine folded structures are usually located between them. The territory of Georgia, as a component part of the Caucasian seismoactive region, belongs to the Mediterranean belt. Its seismotectonic activity is the result of the interaction between the Arabian and Eurasian plates. In Georgia the majority of active faults are hidden (overlapped by sediments of different thickness), except some regional thrusts. They are displaced on the surface as flexures or as clusters of regional faults (Gamkrelidze *et al.*, 1998; Fig. 1).

The territory of Georgia is characterized by moderate seismicity. The number of earthquakes and the maximum

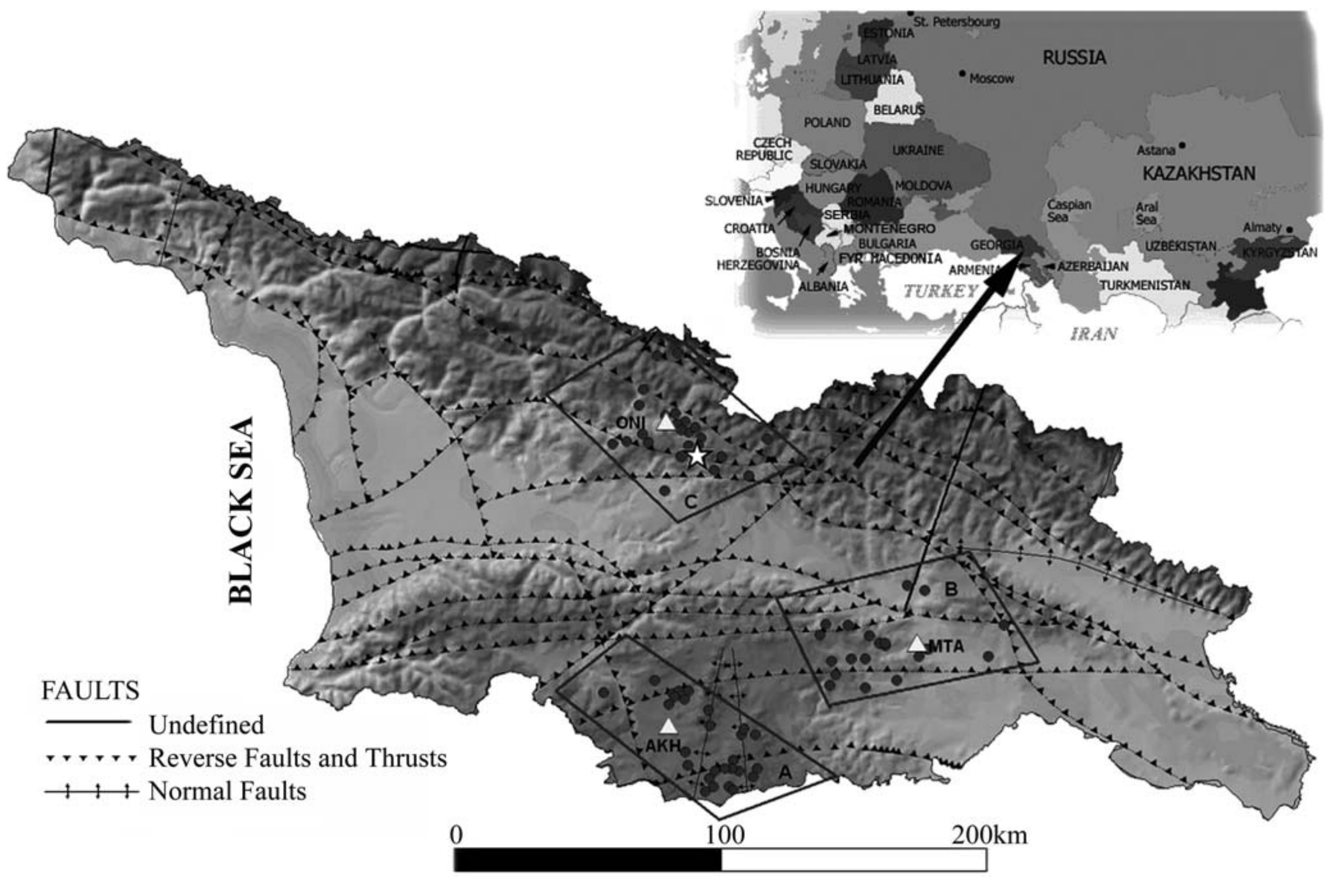

Figure 1. Map of active faults in Georgia (Gamkrelidze *et al.*, 1998). The epicenters of earthquakes (circles) and locations of stations (triangles) are shown with the borders of three regions marked by letters A, B, C. Racha earthquake (1991, *M* 7) is marked with a star.

intensity in Georgia are less than in neighboring Turkey and Iran, but strong and destructive earthquakes have often been observed in its territory. The instrumental period of seismology in Georgia began in 1899, when the first seismological station was installed in Tbilisi. The first seismogram was recorded on 6 December 1899. This gave the start to the development of seismology in the Caucasus. In the beginning of the last century, some seismic stations equipped with the low-gain mechanical instruments were installed in Georgia. From 1955 the network was equipped with highly sensitive seismographs of different types. In the 1990s about 40 seismic stations with analog recordings operated in Georgia, but due to well-known political and economic problems from 1995 to 2000, the number of seismic stations decreased sharply. From 2003 intensive development of the net of digital seismographs began. Now 17 seismic stations with digital seismographs are operating. Six of them are calibrated. In the near future, it is expected that the number of modem stations will increase, allowing us the possibility to solve many seismological problems by using digital data. The seismological database of the Institute of Earth Sciences, Ilia State University, includes information about 63,000 earthquakes from the whole Caucasus region. The data from Georgian, Azerbaijan, and Armenian seismic networks were collected and processed in the Institutes of Geophysics and of Earth Sciences of Georgia. The historical catalog documents data from the beginning of the Christian era. The information about the earthquakes of this period is taken from ancient Georgian annals (Vakhushti, 1855). The parameters of these earthquakes were determined on the basis of the macroseismic data analysis (Kondorskaya and Shebalin, 1982).

In order to estimate the relationship $Q_c(f)$ we selected three regions of Georgia, differing from each other seismically as well as by tectonic structure. These regions (the Javakheti Plateau, the metropolitan Tbilisi region, and the Racha region) considered for the study are marked as A, B, and C, respectively, on the map of Georgia (Figs. 1, 2). Tbilisi, the capital of Georgia (region B), is located between the Greater and the Lesser Caucasus, where the Mtkvari depression is especially narrow. Tbilisi and its adjacent territory have a complex tectonic structure. Several active faults surround Tbilisi; two of them intersect the north and south parts of the city. Tbilisi and its surrounding area are primarily covered by Quaternary sediments, with Middle Eocene outcrops along the Mtkvari River. The average thickness of the sediments is about 2–3 km (Sikharulidze, 1978).

The seismicity of the Tbilisi region is lower than that of the Greater Caucasus to the north and that of the Javakheti Plateau (Lesser Caucasus) to the south (Fig. 2). Several historical earthquakes with magnitude M 3.5–5.2 occurred in the Tbilisi region. The maximum effect from historical earthquakes in the Tbilisi area did not exceed macroseismic intensity of 7 on the Medvedev–Sponhouer–Karnik (MSK) scale. During the instrumental period, about 15 strong and moderate earthquakes affected the city with the intensity

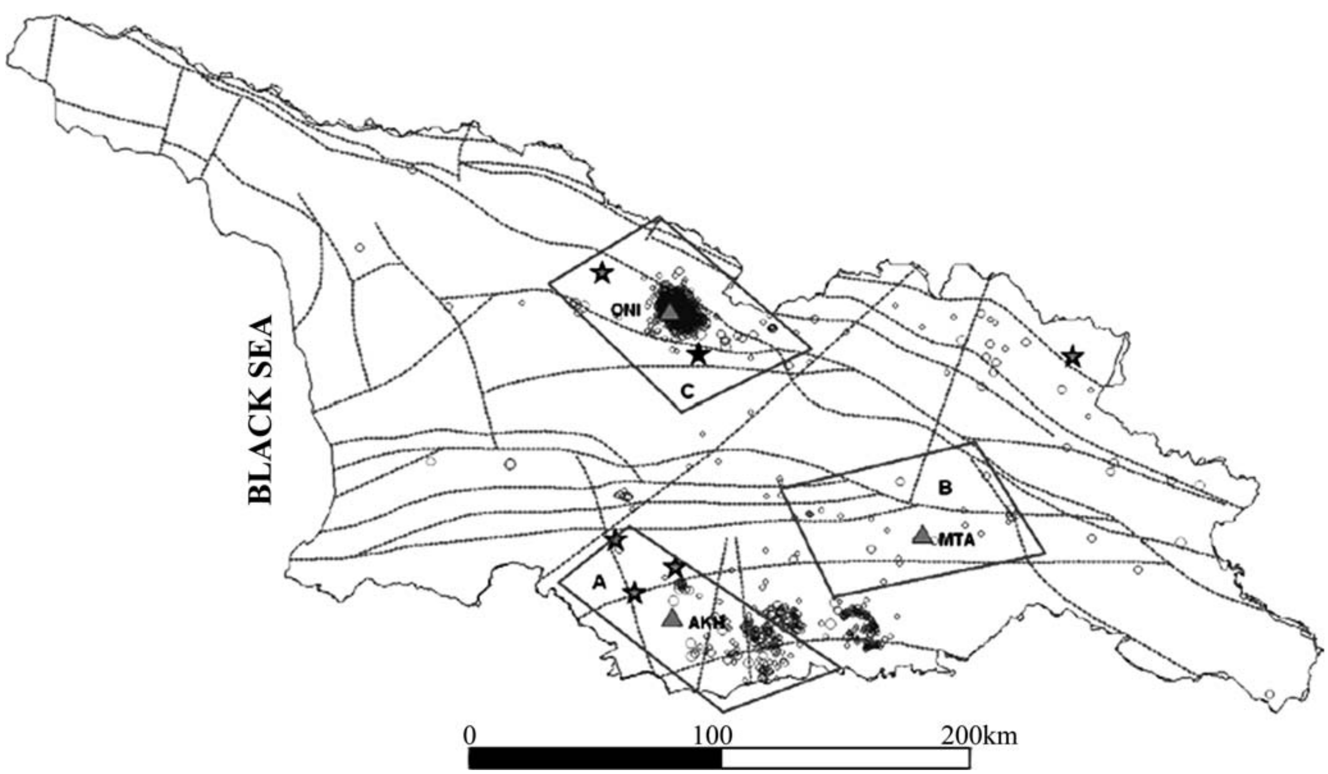

Figure 2. Seismicity in Georgia. Circles represent earthquakes with $M_L \geq 1.5$ that occurred during 2003–2009; light and dark stars indicate earthquakes ($M \geq 6.5$) during the historical and instrumental periods, respectively.

of V–VI MSK. Seismic activity in the city area was comparably low during the instrumental period. Activities of moderate and small earthquakes were observed near the north, south, and east boundaries of Tbilisi; but no significant activity occurred within the city.

During the last two centuries the largest earthquake in Tbilisi occurred on 25 April 2002, with the coda magnitude equal to 4.6. The source mechanism was the normal fault. There has been ongoing seismic activity within the city since this earthquake (Javakhishvili *et al.*, 2004).

The Javakheti Plateau (region A) differs sharply from the Tbilisi region geologically as well as seismically. The Javakheti zone, known as the South Georgian Volcanic Plateau in literature sources, is an area of intensive manifestation of Neogene-Quaternary volcanism. From the orographic point of view, the Javakheti Plateau includes the middle and high mountainous area where flat and weakly wavy plateaus, folded volcanic uplifts, high volcanic massifs, and mountain ridges are developed. From the structural point of view, the Javakheti Plateau occupies the central part of the Caucasian–Asia Minor segment of the Alpine-Himalaya belt; it is located in the zone of the Transcaucasian transverse uplift (Milanovski and Khain, 1963). The thickness of the volcanic cover changes within 1200–2000 m.

Among the seismic areas of the Caucasus, the Javakheti Plateau is notable for its frequent small-magnitude earthquakes (Fig. 2). Large earthquakes are distributed mainly in the peripheral part of the Javakheti Plateau along major faults. The central part of the Javakheti Plateau is divided into fault blocks. On this plateau, small earthquakes occur almost every day. In the last two centuries, six earthquakes with $M \geq 5.2$ occurred. For the instrumental period, the 1940 Tabatsquri earthquake (M 6.0) was the largest recorded in this region. The last large earthquake in the Javakheti Plateau region (M 5.9) was in 1986. Three historical earthquakes ($M \geq 6.5$) also occurred in this region in 1088, 1283, and 1899.

The Racha zone (region C), a late Alpine intermountain molasse depression, is located at the joint of two main structural units of the Caucasus region, the young mountain-folded structure of the Greater Caucasus and the Transcaucasian middle massif. This region is characterized by southward-directed thrusting of folded Paleozoic, Mesozoic, and Paleogene volcanic and sedimentary rocks over Neogene and Quaternary sediments of the Rioni Basin (Milanovsky and Khain, 1963). The thickness of the sedimentary layer is about 4 km (Sikharulidze, 1978).

During the instrumental period, the strongest earthquake in the Caucasus (M 7.0) occurred in Racha in 1991. The historical earthquake of 1350 (M 7.0) is also known. The Georgian catalog of seismicity for 1955–1990 shows some sparse activity within the Racha range; but after the 1991 earthquake the seismic rate increased in this territory, and seismic activity continues into the present. Here, from 1991 to the present, small earthquakes occur almost every day as they do in the Javakheti Plateau region. After the Racha earthquake in 1991, the following earthquakes were important: 4 July 1991, M 5.0; 2 February 2006, M 5.1; 7 September 2009, M 6.0; and 12 April 2010, M 4.2.

Data

For the territory of Georgia three (A, B, C) regions were selected (Figs. 1, 2). Digital seismic waveform data, recorded by stations AKH, MTA, and ONI, were used to study the attenuation property of coda waves in these regions of Georgia. All seismic stations are equipped with the broadband Guralp CMG40T seismometers at a sampling rate of 100 samples/s. At present, data from 11 seismic stations (including the data from Azerbaijan and Turkey stations) are transmitted to the Institute of Earth Sciences online. The database is run on a Linux platform, special seismic software SHM is selected, and the data are recorded in MiniSEED format. In order to process digital seismic waveform data we used the Seismic Analysis Code (SAC) and the Conversion Program Package for Seismological Data on PCs (see Data and Resources). The MTA and ONI stations are situated on sedimentary rock with sandstone and limestone under the stations, respectively. The AKH station is located on volcanic rock.

We used seismograms from the Seismic Monitoring Center network of Georgia. Seismograms of 72 earthquakes with good signal-to-noise ratio were selected and processed for coda Q_c calculation. Many events were discarded because of low signal-to-noise ratio as well as other sets of criteria for coda Q calculation. In general, we discovered practically no significant differences for the seismogram envelopes between north–south and east–west components on the coda portion; therefore, we collected north–south components of seismograms. Features of the data are the following: the epicentral distances are less than 50 km; the average epicentral distances are 26.4 km, 25.6 km, and 20.3 km, respectively, for regions A, B, and C. We did not use the two clusters of earthquakes between the regions A and B, because their epicentral distances are more than 50 km (Fig. 2). The local magnitude range of the earthquakes is 1.7–4.1 with shallow focal depth (up to 20 km); the period of data is from 2003 to 2010 (see Data and Resources for detailed data about most of these earthquakes). Hypocenters of some earthquakes were recalculated by Hypo 71 computer program (Lee and Valdes, 1989). Figure 1 shows the location of the seismograph stations and selected earthquakes. Table 1 shows parameters of the earthquakes used in this study.

Unfortunately, we had the data from only one station in each region, because we did not have other calibrated stations in the studied regions. But it should be noted that in general, for the small epicentral distances (especially for the epicentral distances less than 50 km), the temporal shape of (narrow-band) coda is independent of the azimuth of the seismic station. This independence was studied using the analog records for the territory of Georgia (Shengelia, 1981).

Table 1
Details of the Earthquakes Used in the Study for Regions A, B, C

Number	Region*	Date (yyyy.mm.dd)	Time (hh:mm:ss)	Magnitude M_L	Latitude (° N)	Longitude (° E)	h (km)
1	B	2003.06.05	22:53:06	3.3	41.76	44.50	14.20
2	B	2003.11.22	22:22:43	3.1	41.66	44.31	10.30
3	A	2004.02.05	02:11:53	3.5	41.18	43.80	13.10
4	B	2004.08.24	17:32:16	2.1	41.72	44.58	08.20
5	B	2004.11.03	22:32:57	1.7	41.59	44.64	03.60
6	B	2004.11.22	21:54:12	2.7	41.68	45.10	13.00
7	C	2005.01.27	07:04:36	3.7	42.38	43.70	02.50
8	A	2005.01.27	13:02:41	2.8	41.21	43.70	05.70
9	A	2005.02.07	18:54:04	2.4	41.49	43.49	11.50
10	C	2005.02.17	01:45:01	3.7	42.52	43.57	08.90
11	B	2005.06.05	09:24:06	2.6	41.68	44.75	08.30
12	C	2005.07.21	07:36:15	3.1	42.47	43.18	06.10
13	C	2005.09.23	07:53:30	2.5	42.82	43.50	10.70
14	C	2005.09.28	14:10:35	2.3	42.43	43.53	12.20
15	C	2005.10.04	11:47:06	2.6	42.56	43.55	09.50
16	C	2005.10.10	08:55:11	3.7	42.81	43.40	09.00
17	C	2005.11.02	02:17:58	3.2	42.48	43.25	06.20
18	C	2005.11.28	20:41:14	2.8	42.68	43.33	07.50
19	C	2005.11.30	00:10:56	2.3	42.62	43.31	09.50
20	A	2005.12.16	04:02:29	4.4	41.28	43.82	10.30
21	A	2006.01.29	06:33:36	2.4	41.24	43.85	11.90
22	C	2006.02.03	23:56:12	2.0	42.58	43.45	15.20
23	A	2006.02.12	22:00:22	3.1	41.25	43.81	09.00
24	A	2006.04.26	14:58:48	2.7	41.53	43.16	19.50
25	B	2006.09.13	23:12:12	2.3	41.56	44.48	10.80
26	B	2006.12.31	19:37:14	2.6	41.73	44.55	09.70
27	B	2007.01.02	03:12:11	2.5	41.95	44.69	09.20
28	C	2007.01.06	18:46:36	2.6	42.30	43.45	15.00
29	A	2007.06.29	19:44:38	3.4	41.16	43.68	13.50
30	A	2007.07.09	09:33:10	4.0	41.17	43.82	10.10
31	B	2007.07.22	01:34:51	2.5	41.80	44.39	16.50
32	A	2007.07.24	10:35:27	2.7	41.25	43.78	11.00
33	A	2007.07.24	19:31:22	4.1	41.18	43.72	14.20
34	A	2007.07.24	23:37:56	2.4	41.24	43.73	11.90
35	A	2008.04.01	08:00:00	3.9	41.53	43.51	15.20
36	B	2008.04.10	09:34:48	2.8	41.93	44.78	12.50
37	B	2008.06.18	08:54:33	3.6	41.61	44.33	07.50
38	C	2008.08.08	11:25:02	2.7	42.55	43.51	13.20
39	A	2008.11.18	23:39:17	1.9	41.58	43.42	08.00
40	A	2009.01.11	02:31:57	3.1	41.26	43.59	15.10
41	A	2009.01.11	02:44:07	3.1	41.31	43.58	17.00
42	B	2009.01.12	23:41:40	2.4	41.80	45.18	12.70
43	B	2009.01.19	14:46:30	2.6	41.67	44.48	16.60
44	A	2009.02.18	16:52:26	2.6	41.25	43.94	10.20
45	C	2009.02.21	19:14:29	3.7	42.43	43.74	12.90
46	C	2009.02.22	17:04:52	2.6	42.53	43.53	15.00
47	A	2009.02.26	08:20:13	2.8	42.36	43.88	18.90
48	A	2009.04.25	07:47:52	2.7	41.22	43.92	07.70
49	C	2009.05.15	19:33:57	1.9	42.48	43.36	11.20
50	C	2009.05.18	11:39:20	2.5	42.51	43.33	14.00
51	C	2009.06.09	00:35:24	2.7	42.50	43.97	15.50
52	A	2009.07.02	05:18:09	2.2	41.54	43.54	11.30
53	A	2009.07.02	12:15:37	2.5	41.54	43.59	13.40
54	A	2009.07.18	14:25:29	3.1	41.38	43.86	15.60
55	A	2009.08.08	09:11:36	2.4	41.40	43.87	15.70
56	A	2009.08.24	13:17:36	2.6	41.51	43.57	12.00
57	A	2009.08.29	15:27:01	2.4	41.38	43.93	10.30
58	C	2009.09.10	11:01:25	2.9	42.50	43.63	13.20
59	C	2009.09.11	05:59:21	3.0	42.47	43.57	10.50
60	C	2009.09.30	14:52:53	2.7	42.59	43.50	09.70
61	B	2009.10.07	22:50:12	2.3	41.73	44.58	09.40

(*continued*)

Table 1 (*Continued*)

Number	Region*	Date (yyyy.mm.dd)	Time (hh:mm:ss)	Magnitude M_L	Latitude (° N)	Longitude (° E)	h (km)
62	B	2009.10.11	18:03:11	2.9	41.71	44.79	07.60
63	A	2009.11.02	17:20:38	2.3	41.41	43.70	11.30
64	A	2009.11.10	02:43:38	3.3	41.52	43.52	14.80
65	B	2009.11.10	02:43:38	2.7	41.71	44.80	06.20
66	B	2009.12.04	17:45:14	3.1	41.57	44.30	08.80
67	A	2009.12.24	06:12:28	2.6	41.47	43.69	10.00
68	C	2009.12.27	21:18:41	2.1	42.47	43.62	10.40
69	C	2010.01.14	05:01:30	2.3	42.54	43.60	12.50
70	B	2010.01.28	08:45:13	2.8	41.80	44.28	11.60
71	B	2010.02.10	20:43:27	3.3	41.76	44.25	15.20
72	B	2010.05.02	19:45:35	2.6	41.67	44.41	12.30

*A, Javakheti Plateau; B, Tbilisi region; C, Racha region.

The Method and the Results

According to the single-scattering model, the coda wave amplitudes $A(f,t)$, obtained for each frequency band centered at f and at lapse time t measured from the earthquake origin time, are described as

$$A(f,t) = S(f)t^{-\alpha}\exp[-\pi f t/Q_c(f)], \tag{1}$$

where $S(f)$ represents the source factor at frequency f and is considered as a constant; α is the geometrical spreading parameter and is equal to 1.0, 0.5, or 0.75 for single scattering of body waves, surface waves, or diffusive waves, respectively (Aki, 1969; Aki and Chouet, 1975; Sato and Fehler, 1998); and $Q_c(f)$ is the quality factor.

The values of Q_c were determined using the data from all events listed in Table 1. The QCODA program (Valdes and Novelo-Casanova, 1989) is used for coda wave analysis and for the estimation of the quality factor Q_c in this study. The calculation of coda Q_c was performed in octave frequency intervals (1–2; 2–4; 4–8; 8–16; 16–32 Hz) with central frequencies of 1.5, 3, 6, 12, and 24 Hz for each region under the study. We used three methods available in the package: (1) The back-scattering model of Aki and Chouet (1975) with the source and receiver in the same location. In this model the scattering is weak (single scattering). Lapse times used for the Aki and Chouet (1975) model began at twice the *S*-wave travel time. (2) The single-scattering model of Sato (1977), in the time domain. Two lapse time intervals were examined: beginning at the *S*-wave travel time, as well as beginning at twice the *S*-wave travel time. (3) The single-scattering model in the frequency domain (Lee *et al.*, 1986), using the fast Fourier transform (FFT) for estimating the power spectrum of overlapping windows. Here again, lapse times beginning with the *S*-wave arrival, as well as beginning at twice the *S*-wave travel time were examined. Thus, we estimated coda Q_c using three different models, and for two of those models, two different lapse time intervals, for a total of five different estimates.

This program displays one seismic trace at a time and uses a windowed section of the seismogram, or the whole seismogram, which should be less than 10,000 samples long (or 100 s if the data were digitized at 100 samples/s). In our case, the windowed section is up to 63 s, and the coda window is from 35 s up to 45 s. The average coda window is about 42 s for each region. Unfiltered and filtered seismograms of the event that occurred on 9 November 2009 recorded at station ONI are shown in Figure 3.

For each earthquake group, we averaged all the Q_c values determined for each frequency. The average $\langle Q_c\rangle$ was calculated using the following expression (Hellweg *et al.*, 1995):

$$\langle Q_c\rangle = \frac{\sum(Q_{ci}/\sigma_i^2)}{\sum(1/\sigma_i^2)}. \tag{2}$$

The summation includes Q_{ci} values for all N earthquakes (for each region and for each frequency band) with standard deviations σ_i. Because each Q_{ci} was weighted by the inverse of its variance, the values with small variances contribute more to $\langle Q_c\rangle$ than values with larger variances. Following Hellweg *et al.* (1995), the variance of the mean was calculated using the expression

$$\sigma_m^2 = \frac{\sum[(1/\sigma_i^2)(Q_{ci} - \langle Q_c\rangle)^2]}{[(N-1)\sum(1/\sigma_i^2)]}, \tag{3}$$

where σ_m is the standard deviation of the mean.

As previously mentioned, we calculated the coda Q_c in five different ways using three methods. Among the five results from the QCODA package (Valdes and Novelo-Casanova, 1989), one result with the minimum error was selected as the appropriate coda Q_c for that propagation path. Namely, at frequency $f = 1.5$ Hz the Q_c values with an error $\leq 20\%$ and at other frequencies, the Q_c values with an error $\leq 10\%$ were selected for each trace. We also estimated the values of Q_c by the model of Aki and Chouet (1975), by Sato's model (1977), and by Lee's model (Lee *et al.*, 1986) separately. The Q_c values with the errors $\leq 20\%$ were selected at each central frequency for each method. In most cases, the method of Sato (1977), using a lapse time interval beginning at two times the *S*-wave travel time, resulted in errors exceeding 20%. Therefore, we only consider the results obtained using Sato's method at lapse times beginning

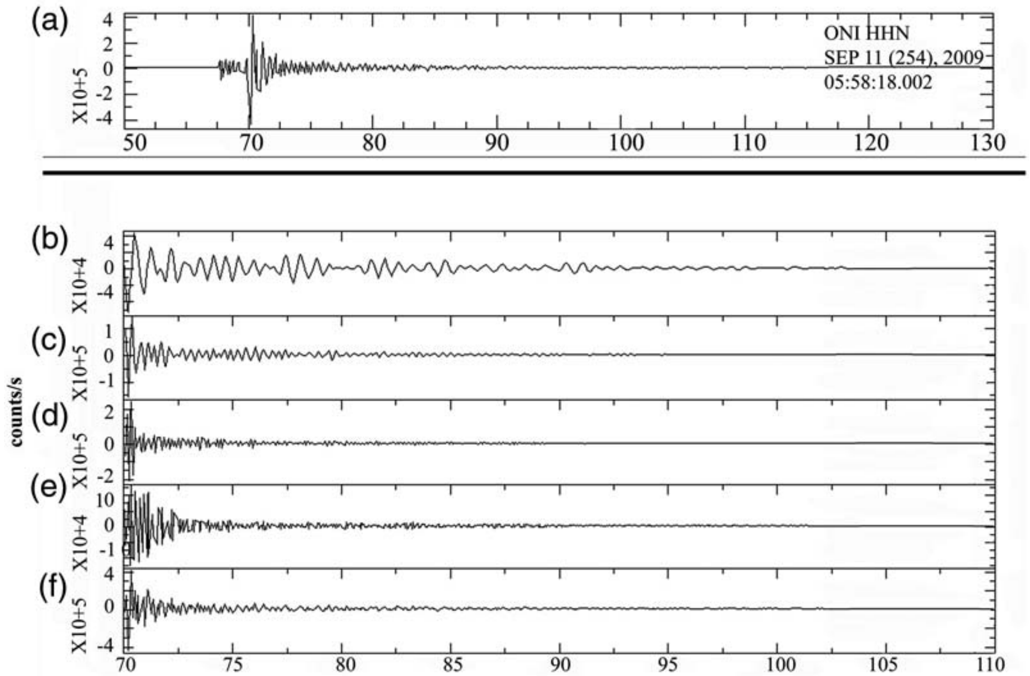

Figure 3. Example of the band-pass-filtered records for the 9 November 2009 event recorded at station ONI. (a) Original waveform. (b–f) Band-pass-filtered coda waves at central frequencies 1.5, 3.0, 6.0, 12.0, and 24.0. The selected window is from the *S* wave to the end of coda.

with the *S*-wave arrival. We denote the various models and lapse time intervals as follows: Aki and Chouet (1975), AC; Sato's and Lee's models, applied with lapse times beginning at the *S*-wave travel time and at twice the *S*-wave travel time, S1, L1, S2, L2, respectively. The mean values of Q_c and corresponding standard deviations are reported in Table 2. In general, for the B and C regions, Q_c values that were determined by the methods of Aki and Chouet (1975) and Sato (1977) had an error more than 20% at frequency $f = 24$ Hz. Thus, for regions B and C we established the relationships $Q_c(f)$ at central frequencies: $f = 1.5$; 3.0; 6.0, and 12.0 Hz using these two methods. At frequency $f = 24$ Hz, Lee's method (especially for the coda waves arriving immediately after the *S*-wave arrival) gives more appropriate Q_c values than the other two methods for all regions.

Q_c estimates for the three different regions of Georgia are found to be strong functions of frequency in the high-frequency range (from 1.5 Hz to 24.0 Hz). Estimated values of Q_c were fitted to a power law,

$$Q_c(f) = Q_0(f)^n, \tag{4}$$

where Q_0 is the quality factor at 1 Hz and n is the frequency parameter, which varies from region to region based on the heterogeneity of the medium (Aki, 1981). We estimated Q_c and its frequency dependence in all of the regions studied. We obtained the following estimates using all three models simultaneously (Fig. 4):

$$Q_c = (35.1 \pm 5.3)f^{1.095\pm0.056} \quad \text{(region A)}, \tag{5}$$

$$Q_c = (77.2 \pm 8.1)f^{0.937\pm0.044} \quad \text{(region B)}, \tag{6}$$

and

$$Q_c = (37.2 \pm 5.4)f^{1.089\pm0.054} \quad \text{(region C)}. \tag{7}$$

The Q_c functions determined for Sato's (1977) S1 model are given by the expressions

$$Q_c = (36.5 \pm 7.0)f^{1.081\pm0.071} \quad \text{(region A)}, \tag{8}$$

$$Q_c = (74.9 \pm 8.0)f^{0.944\pm0.053} \quad \text{(region B)}, \tag{9}$$

and

$$Q_c = (40.0 \pm 5.5)f^{1.057\pm0.060} \quad \text{(region C)}. \tag{10}$$

The relationships $Q_c(f)$ obtained using Aki and Chouet's (1975) AC model

$$Q_c = (36.2 \pm 7.7)f^{1.082\pm0.078} \quad \text{(region A)}, \tag{11}$$

$$Q_c = (86.2 \pm 10.8)f^{0.890\pm0.062} \quad \text{(region B)}, \tag{12}$$

and

$$Q_c = (41.0 \pm 5.4)f^{1.052\pm0.061} \quad \text{(region C)}. \tag{13}$$

Table 2

Mean Value of Q_c and Standard Deviation at Different Frequency Bands for Regions A, B, and C

Region*	1.5 Hz (1.0–2.0) $Q_c \pm \sigma_m$	N†	3 Hz (2.0–4.0) $Q_c \pm \sigma_m$	N†	6 Hz (4.0–8.0) $Q_c \pm \sigma_m$	N†	12 Hz (8.0–16.0) $Q_c \pm \sigma_m$	N†	24 Hz (16.0–32.0) $Q_c \pm \sigma_m$	N†
(a) All Three Methods Used Simultaneously										
A	92.7 ± 3.6	18	159.3 ± 3.9	25	266.1 ± 6.5	27	474.6 ± 11.8	22	1156.8 ± 34.4	16
B	136.7 ± 3.4	19	192.6 ± 7.2	22	409.0 ± 15.9	21	802.9 ± 17.5	20	1514.2 ± 38.9	15
C	74.1 ± 2.2	20	139.9 ± 5.2	23	275.7 ± 7.8	20	526.3 ± 14.7	20	1194.2 ± 33.3	19
(b) Sato's Method S1										
A	103.2 ± 3.0	11	157.8 ± 3.5	19	267.5 ± 7.7	25	477.0 ± 17.5	11	1151.3 ± 43.2	10
B	124.6 ± 3.7	14	186.9 ± 8.4	16	421.7 ± 10.8	13	783.4 ± 22.2	11	—	
C	71.3 ± 2.3	14	134.9 ± 5.4	18	251.9 ± 8.0	15	556.6 ± 15.9	15	—	
(c) Aki and Chouet's Method AC										
A	89.6 ± 3.9	16	156.3 ± 4.7	15	262.8 ± 10.2	18	480.8 ± 22.1	10	1139.7 ± 40.7	9
B	126.1 ± 4.2	12	198.6 ± 11.4	11	455.5 ± 21.3	14	779.0 ± 17.1	9	—	
C	69.5 ± 2.6	16	134.4 ± 5.1	15	261.8 ± 8.1	16	562.7 ± 16.5	13	—	
(d) Lee's Method L1 Applied at the *S*-Wave Travel Time										
A	102.9 ± 3.4	11	161.8 ± 4.5	18	274.9 ± 6.8	19	481.6 ± 12.4	14	1172.5 ± 34.5	13
B	136.7 ± 3.5	14	196.6 ± 5.2	17	422.7 ± 9.6	16	836.8 ± 19.0	13	1576.3 ± 26.0	11
C	82.6 ± 3.1	11	139.9 ± 5.4	20	266.4 ± 12.1	20	525.0 ± 16.2	20	1189.9 ± 37.7	16
(e) Lee's Method L2 Applied at Twice the *S*-Wave Travel Time										
A	100.6 ± 5.8	10	162.8 ± 4.0	18	284.9 ± 4.6	14	507.1 ± 12.3	12	1210.2 ± 49.0	10
B	133.3 ± 3.6	12	200.8 ± 8.5	13	466.4 ± 14.3	13	829.4 ± 17.4	12	1585.1 ± 27.1	11
C	82.1 ± 2.4	12	144.9 ± 5.5	19	286.8 ± 10.1	17	535.3 ± 17.8	17	1217.3 ± 42.1	14

*A, Javakheti Plateau; B, Tbilisi region; C, Racha region.
†N, number of earthquakes.

The relationships $Q_c(f)$ determined using Lee's model (Lee *et al.*, 1986), L1 and L2, are given by the expressions in equations (14), (15), (16), (17), (18), and (19), respectively:

$$Q_c = (37.0 \pm 5.6) f^{1.083\pm0.057} \quad \text{(region A)}, \tag{14}$$

$$Q_c = (79.0 \pm 7.4) f^{0.943\pm0.035} \quad \text{(region B)}, \tag{15}$$

$$Q_c = (36.7 \pm 6.6) f^{1.092\pm0.065} \quad \text{(region C)}, \tag{16}$$

$$Q_c = (38.8 \pm 6.9) f^{1.078\pm0.065} \quad \text{(region A)}, \tag{17}$$

$$Q_c = (83.4 \pm 8.2) f^{0.923\pm0.036} \quad \text{(region B)}, \tag{18}$$

and

$$Q_c = (39.1 \pm 6.9) f^{1.080\pm0.066} \quad \text{(region C)}. \tag{19}$$

The relationships $Q_c(f)$ show strong frequency-dependent behavior for the studied regions, no matter which method is used.

In the single-scattering model, for any given lapse time t, the seismic coda waves sample an ellipsoidal volume, with the earthquake source and the station as foci (Pulli, 1984). The area for coda wave generation is assumed to be elliptical, which can be represented as

$$\frac{x^2}{\alpha_1^2} + \frac{y^2}{\alpha_2^2} = 1, \tag{20}$$

where $\alpha_1 = v_s t/2$ and $\alpha_2 = \sqrt{\alpha_1^2 - \Delta^2}$ are the lengths of semi-major and semi-minor axes of the ellipsoidal and v_s is the S-wave velocity ($v_s = 3.2$ km/s, in our case). The

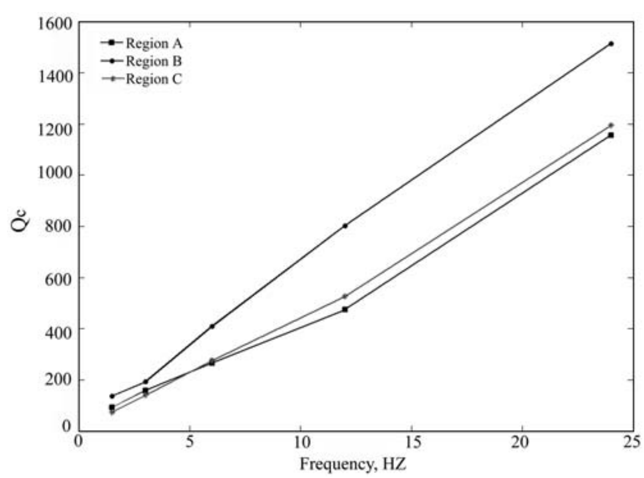

Figure 4. The frequency dependence of Q_c for different regions of Georgia.

Δ is the average epicentral distance of the events. The average lapse time is taken as $t_{\mathrm{start}} + W/2$, where t_{start} is the starting time of the coda window and W is the coda window length. The ellipsoid reaches the average depth of

$$h = h_{\mathrm{av}} + \alpha_2, \tag{21}$$

where h_{av} is the average focal depth of the earthquakes (Pulli, 1984; Parvez *et al.*, 2008). So the Q_c values represent the average attenuation property of the ellipsoids with areas of 7643 km^2, 6789 km^2, and 7071 km^2; the average depths of the ellipsoidal volumes reach 56 km, 55 km, and 56 km under stations MTA, AKH, ONI, respectively. The average depths of the ellipsoidal volumes under every station are almost the same, and because the crust thickness is about 50 km in the Caucasus, we can suppose that estimated Q_c values are characteristic of the Earth's crust.

Discussion and Conclusions

As described earlier in the Introduction, previous work using data from analog seismic stations of the regional Georgia network examined attenuation of coda waves, using records of earthquakes from the Caucasus region (Shengelia, 1981). The seismic stations were equipped with short period SKM-3 type and long period SKD type (instrument response was practically flat to displacement between 0.6 and 1.2 s for SKM-3 type and between 0.5 and 20 s for SKD type) analog seismometers. It was found that values of Q_c varied from 65 to 570 depending on the lapse time, the bandwidth of the seismometer, and the model of coda generation. As the lapse time increased, the Q_c increased also.

In the present study we investigated the coda attenuation for three regions of Georgia, using the single-scattering model of coda generation for local earthquakes waveforms. The QCODA program (Valdes and Novelo-Casanova, 1989) was used for the estimation of the quality factor Q_c. The calculation of coda Q_c was performed in octave frequency intervals with center frequencies of 1.5, 3, 6, 12, and 24 Hz. In Georgia, several studies have been carried out concerning the attenuation for body waves, surface waves (Sikharulidze, 1978), and coda waves (Shengelia, 1981), but the values of Q were estimated only for certain frequencies using the analog recordings. For example, $Q_c = 65$–85 for the lapse time 10–100 s at the frequency $f = 0.7$–1.25 Hz, $Q_c = 140$–200 for the lapse time 100–500 s at the frequency $f = 0.5$–0.8 Hz, and $Q_c = 415$–570 for the lapse time 500–1400 s at the frequency $f = 0.25$–0.5 Hz for all of Georgia and adjacent its territory. Sarker and Abers (1999) estimated Q for the western Greater Caucasus and obtained $Q_c = 753 \pm 107$; their estimates are dominated by the attenuation at 8–12 Hz. We received a similar result for region B at $f = 12.0$ Hz.

Here we find Q_c estimates for different regions of Georgia to be a strong function of the frequency in the high-frequency range. These dependencies are described by the expressions in equations (5)–(19). For the territory of Georgia the relationship $Q_c(f) = Q_0(f)^n$ was established for the first time. Though, in 1984, during the expedition (organized by the Institute of Geophysics of the Georgian National Academy of Sciences) the ChISS apparatus was installed in Akhalkalaki at the same location where the broadband instrument is placed now (Shengelia, 1985). The ChISS instrument was designed by Zapolskii (1971) and was similar to instruments used by Aki and Chouet (1975). The vertical component of the velocity of the ground was passed through a system of band-pass filters and was recorded on photographic paper. Data from eight channel systems, covering the range 0.6 to 42 Hz, were used. The calculation of coda Q_c was performed in octave frequency intervals with center frequencies of 0.6, 1.2, 3.0, 6.0, 9.0, 18.0, 27.0, 36.0, and 42.0 Hz. The coda was analyzed in a time window of 20–120 s. The relationship for the Javakheti Plateau $Q_c = (48.0)f^{0.98}$ was established. In the present work these relationships are given by equations (5), (8), (11), (14), and (17), but here the lapse time is up to 63 s.

It should be noted that when Q_c values (derived from analog data) were compared by Shengelia (1981), it was found that in the lapse time 10–3000 s Q_c values were the lowest in the Caucasian region, compared with the regionsof Crimea, the Carpatians, the South and North Tien Shan, Kirghizia, the Altai, Baikal, and Kamchatka. This was explained by the geological age of the folding. Relatively late folding

Table 3
Mean Values of Q_0 and n for Different Regions of the World

Region	Q_0	n	Coda Window
A (this study)	35.1 ± 5.3	1.095 ± 0.056	35–45 s
B (this study)	77.2 ± 8.1	0.937 ± 0.044	35–45 s
C (this study)	37.2 ± 5.4	1.089 ± 0.054	35–45 s
Northwest Himalaya (Kumar *et al.*, 2005)	158	1.05	40 s
Southwest Anatolia (Şahin and Alptekin, 2006)	39 ± 4	0.860 ± 0.043	50 s
West Anatolia (Akinci *et al.*, 1994)	50.7	1.01	30 s
Garhwal Himalya, India (Mandal *et al.*, 2001)	30 ± 0.8	1.21 ± 0.03	50 s
Southeast Sicily (Giampiccolo *et al.*, 2002)	79	0.9	40 s
Andaman Islands (Parvez *et al.*, 2008)	122	0.75	40 s
Yunnan Province, China (Li Bai-ji *et al.*, 2004)	49	0.95	30 s
Southern Netherlands (Goutbeek *et al.*, 2004)	90	0.93 ± 0.26	35 s

(such as in the Caucasus) is more disturbed and inhomogeneous; accordingly, Q_c values in such regions are smaller than in those regions where the folding age is older and the medium is more consolidated, that is, the number of scattering discontinuities is relatively small (Rautian *et al.*, 1981).

In this paper, we cannot directly compare the estimated Q_c values with the values derived earlier, because the frequency bands as well as the lapse times considered are different. But we can compare our Q_c estimates with Q_c estimates derived from other regions of the world. For comparison, it is desirable to have almost similar lapse times, because Q_c values appear to increase with increasing lapse time (Rautian *et al.*, 1981). Table 3 gives values Q_0 (Q_0 is Q_c at $f = 1$ Hz) and n (frequency parameter) for various seismically active regions of the world.

It seems that the regions we investigated (especially regions A and C) are characterized by lower Q_0 and higher n than most of the available values of the world (Table 3). Many researchers have shown that both n and Q_0 parameters can characterize the level of tectonic activity of a seismic region. In general, the Q_0 value is low in seismically active regions; it is a measure of medium heterogeneities. Also the degree of frequency dependence n is higher ($n \sim 1$) for tectonically active regions than for tectonically stable regions (Aki, 1980; Akinci *et al.*, 1994; Singh *et al.*, 2001). So our estimates of Q_0 and n indicate that the crust beneath the studied region is probably more heterogeneous than beneath most other regions. The lower coda Q_c may be due to some heterogeneity in the medium causing the loss of the energy.

The most noticeable features of our results are the following: regions A and C show low coda Q_c for all frequency bands, corresponding to the epicentral areas of large earthquakes ($M \geq 6.5$). In the territory of Georgia, the highest seismicity is observed in these regions. It should be noted that the seismicity of the Racha region increased sharply after the 1991 earthquake (M 7.0). For both the Javakheti Plateau region and the Racha region the n coefficient is almost the same; the differences between the Q_c values are not considerable at all frequencies. This shows the uniform distribution of small-scale scatter over these two regions and a similar level of tectonic activity in the regions A and C. In the Tbilisi region B, which is seismically not as active, Q_c values are relative higher. Also in regions A and C, Q_0 values are lower than in region B. This means that these regions are more heterogeneous compared with region B. This observation conforms well to the existing geology of the considered regions. In the part of the Greater Caucasus where the Racha region is situated, deformations are concentrated, while the Javakheti Plateau consists of numerous faults and cracks (Milanovsky and Khain, 1963; Gamkrelidze *et al.*, 1998).

Although the present study is based on a limited data set, still the estimated relationships $Q_c(f) = Q_0(f)^n$ give some information about seismic wave attenuation characteristics of the three studied regions of Georgia. The estimation of Q_c in Georgia and in the Caucasus will be an important parameter for the prediction of large earthquakes, for the assessment of the seismic hazard, and for better understanding of the tectonics, the seismicity, and the engineering seismology (Jin and Aki, 1988). In the future, when enough data will be available, we intend to determine Q_c values for the whole territory of Georgia and to accomplish the associated mapping.

Data and Resources

Digital seismic waveform data were processed using Seismic Analysis Code (SAC) (http://www.llnl.gov/sac, last accessed April 2011) and The Conversion Program Package for Seismological Data on PCs (http://www.orfeus-eu.org/Software/conversion.html; last accessed April 2011). See http://seismo.iliauni.edu.ge (last accessed April 2011) for detailed data about the earthquakes in regions A, B, and C.

Acknowledgments

We thank Pier Luigi Bragato and an anonymous reviewer for their critical reviews and valuable comments. This work was supported by Georgian National Science Foundation Grant ST06/5-083.

References

Aki, K. (1969). Analysis of seismic coda of local earthquakes as scattering waves, *J Geophys. Res.* **74,** 615–631.

Aki, K. (1980). Scattering and attenuation of shear waves in the lithosphere, *J. Geophys. Res.* **85,** no. B11, 6496–6501.

Aki, K. (1981). Source and scattering effects on the spectra of small local earthquakes, *Bull. Seismol. Soc. Am.* **71,** 1687–1700.

Aki, K. (1985). Practical approaches to earthquake prediction and warning, *Earthquake Prediction Res.* **3,** 219–230.

Aki, K., and B. Chouet (1975). Origin of coda waves: Source, attenuation, and scattering effects, *J. Geophys. Res.* **80,** 3322–3342.

Akinci, A., A. G. Taktak, and S. Ergintav (1994). Attenuation of coda waves in western Anatolia, *Phys. Earth Planet. In.* **87,** no. 1–2, 155–165.

Fehler, M., M. Hoshiba, H. Sato, and K. Obara (1992). Separation of scattering and intrinsic attenuation for the Kanto-Tokai region, Japan, using measurements of *S*-wave energy versus hypocentral distance, *Geophys. J. Int.* **108,** 787–800.

Frankel, A., and L. Wennerberg (1987). Energy-flux model of seismic coda: Separation of scattering and intrinsic attenuation, *Bull. Seismol. Soc. Am.* **77,** 1223–1251.

Gamkrelidze, I., T. Giorgobiani, S. Kuloshvili, G. Lobjanidze, and G. Shengelaia (1998). Active deep faults map and the catalogue for the territory of Georgia, *Bull. Georgian Acad. Sci.* **157,** 80–85.

Gao, L. S., L. C. Lee, N. N. Biswas, and K. Aki (1983). Comparison of the effects between single and multiple scattering on coda waves for local earthquakes, *Bull. Seismol. Soc. Am.* **73,** 377–389.

Giampiccolo, E., G. Tusa, H. Langer, and S. Gresta (2002). Attenuation in southeastern Sicily (Italy) by applying different coda methods, *J. Seismol.* **6,** 487–501.

Goutbeek, F. H., B. Dost, and T. Van Eck (2004). Intrinsic absorption and scattering attenuation in the southern part of the Netherlands, *J. Seismol.* **8,** no. 1, 11–23, doi 10.1023/B:JOSE.0000009511.27033.79.

Hellweg, M., P. Spudich, J. B. Fletcher, and L. M. Baker (1995). Stability of coda *Q* in the region of Parkfield, California: View from the U.S. Geological Survey Parkfield dense seismograph array, *J. Geophys. Res.* **100,** 2089–2102.

Javakhishvili, Z., T. Godoladze, M. Elashvili, T. Mukhadze, and I. Timchenko (2004). The Tbilisi earthquake of April 25, 2002 in context of the seismic hazard of Tbilisi urban area, *Bollettino di Geofisica Teorica ed Applicata* **45,** no. 3, 165–185.

Jin, A., and K. Aki (1988). Spatial and temporal correlation between coda Q and seismicity in China, *Bull. Seismol. Soc. Am.* **78,** 741–769.

Kondorskaya, N. V., and N. V. Shebalin (1982). *New catalogue of the strong earthquakes in the USSR from ancient times through 1977*, Report SE31, World Data Center A for Solid Earth Geophysics, 608 pp.

Kumar, N., I. A. Parvez, and S. H. Virk (2005). Estimation of coda wave attenuation for NW Himalayan region using local earthquakes, *Phys. Earth Planet. In.* **151,** 243–258.

Lee, W. H. K., and C. M. Valdes (1989). User Manual for HYPO71PC, *International Association of Seismology and Physics of the Earth's Interior.* Chapter 9, IASPEI Software Library, **1,** 203–236.

Lee, W. H. K., K. Aki, B. Chouet, P. Jonson, S. Marks, J. T. Newberry, A. S. Ryall, S. W. Stewart, and D. M. Tottijngham (1986). A preliminary study of coda Q in California and Nevada, *Bull. Seismol. Soc. Am.* **76,** 1143–1150.

Li, B.-J., J.-Z. Qin, X.-D. Qian, and J.-Q. Ye (2004). The coda attenuation of the Yao'an area in Yunnan Province, *Acta Seismologica Sinica* **17,** no. 1, 47–53.

Mandal, P., S. Padhy, B. K. Rastogi, H. V. S. Satyanarayana, M. Kousalya, R. Vijayraghavan, and A. Srinivasan (2001). Aftershock activity and frequency-dependent low coda Q_c in the epicentral region of the 1999 Chamoli earthquake of M_w 6.4, *Pure Appl. Geophys.* **158,** 1719–1735.

Milanovsky, E. E., and V. E. Khain (1963). *Geology of the Caucasus*, Moscow State University Publishing House, USSR, Moscow, 357 pp. (in Russian).

Parvez, I. A, A. K. Suntar, M. Mridula, S. K. Mishra, and S. S. Rai (2008). Coda Q estimates in the Andaman Islands using local earthquakes, *Pure Appl. Geophys.* **165,** 1861–1878.

Pulli, J. J. (1984). Attenuation of coda waves in New England, *Bull. Seismol. Soc. Am.* **74,** 1149–1166.

Rautian, T. G., and V. I. Khalturin (1978). The use of the coda for determination of the earthquake source spectrum, *Bull. Seismol. Soc. Am.* **68,** 923–948.

Rautian, T. G., V. I. Khalturin, M. C. Zakirov, A. G. Zemtsova, A. P. Proskurin, B. G. Pustovitenko, L. G. Sinelnikova, A. G. Filina, and I. S. Shengelia (1981). *Experimental Investigation of Seismic Coda*, Nauka Publishing House, Moscow, 142 pp. (in Russian).

Şahin, Ş., and Ő. Alptekin (2006). Frequency dependent attenuation of seismic waves in the crust and upper mantle in southwest Anatolia, *J. of the Earth Science Application and Research Centre of Hacettepe University* **27,** no. 2, 53–62.

Sarker, G., and G. A. Abers (1999). Lithosperic temperature estimates from seismic attenuation across range fronts in southern and central Eurasia, *Geology* **27,** 427–430.

Sato, H. (1977). Energy propagation including scattering effects: Single isotropic scattering approximation, *J. Phys. Earth* **25,** 27–41.

Sato, H., and M. Fehler (1998). *Seismic Wave Propagation and Scattering in the Heterogeneous Earth*, AIP Press/Springer-Verlag, New York, 308 pp.

Shengelia, I. (1981). Seismic coda of earthquakes in the Caucasus. Its properties and application in order to solve some tasks of seismology, *Ph.D. Thesis*, Tbilisi State University, Tbilisi, Georgia, 176 pp. (in Russian).

Shengelia, I. (1985). Seismic coda envelopes for earthquakes of Javakheti plateau, *Proc. of the Geophysical Institute of Georgia* **3,** 25–29 (in Russian).

Sikharulidze, D. (1978). *The Earth's Structure According Surface Waves*, Metsniereba Publishing House, Tbilisi, Georgia, 247 pp. (in Russian).

Singh, D. D., A. Govoni, and P. L. Bragato (2001). Coda Q_c Attenuation and source parameter analysis in Friuli (NE Italy) and its vicinity, *Pure Appl. Geophys.* **158,** 1737–1761.

Valdes, C. M., and D. A Novelo-Casanova (1989). User Manual for Q coda, Chapter 10, *International Association of Seismology and Physics of the Earth's Interior*, IASPEI Software Library, vol. **I,** 237–255.

Vakhushti, B. (1855). *Description of the Georgian Kingdom*, S. Kaukhchishvili (Editor), Tbilisi, Georgia, 323 pp. (in Georgian).

Woodgold, C. R. D. (1994). Coda Q in the Charlevoix, Quebec Region: Lapse time dependence and spatial and temporal comparisons, *Bull. Seismol. Soc. Am.* **84,** 1123–1131.

Wu, R. S. (1985). Multiple scattering and energy transfer of seismic waves-separation of scattering effect from intrinsic attenuation-I. Theoretical modelling, *Geophys. J. Roy. Astron. Soc.* **82,** 57–80.

Zapolskii, K. K. (1971). The frequency selective seismograph station ChISS, in *Experimental Seismology*, Nauka Publishing House, Moscow, 20–36 (in Russian).

Institute of Earth Sciences
Ilia State University
77, Nutsubidze Str., 0177
Tbilisi, Georgia
ia_shengelia@iliauni.edu.ge
zurab_javakhishvili@iliauni.edu.ge
(I.S., Z.J.)

Department of Seismic Hazard and Risk Analysis
Institute of Earth Sciences
Ilia State University
77, Nutsubidze Str., 0177
Tbilisi, Georgia
nato_jorjiashvili@iliauni.edu.ge
(N.J.)

Manuscript received 30 November 2010