Variation of coda wave attenuation in the Alborz region and central Iran

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SUMMARY
More than 340 earthquakes recorded by the Institute of Geophysics, University of Tehran (IGUT) short period stations from 1996 to 2004 were analysed to estimate the S-coda attenuation in the Alborz region, the northern part of the Alpine–Himalayan orogen in western Asia, and in central Iran, which is the foreland of this orogen. The coda quality factor, $Q_c$, was estimated using the single backscattering model in frequency bands of 1–25 Hz. In this research, lateral and depth variation of $Q_c$ in the Alborz region and central Iran are studied. It is observed that in the Alborz region there is absence of significant lateral variation in $Q_c$. The average frequency relation for this region is $Q_c = 79 \pm 2^{1.07\pm0.08}$. Two anomalous high-attenuation areas in central Iran are recognized around the stations LAS and RAZ. The average frequency relation for central Iran excluding the values of these two stations is $Q_c = 94 \pm 2^{0.97\pm0.12}$. To investigate the attenuation variation with depth, $Q_c$ value was calculated for 14 lapse times (25, 30, 35, … 90s) for two data sets having epicentral distance range $R < 100$ km (data set 1) and $100 < R < 200$ km (data set 2) in each area. It is observed that $Q_c$ increases with depth. However, the rate of increase of $Q_c$ with depth is not uniform in our study area. Beneath central Iran the rate of increase of $Q_c$ is greater at depths less than 100 km compared to that at larger depths indicating the existence of a high attenuation anomalous structure under the lithosphere of central Iran. In addition, below $\sim 180$ km, the $Q_c$ value does not vary much with depth under both study areas, indicating the presence of a transparent mantle under them.

Key words: Coda waves; Seismic attenuation.

1 INTRODUCTION
The attenuation of short-period $S$ waves, expressed as the inverse of the quality factor ($Q^{-1}$), helps in understanding the physical laws related to the propagation of the elastic energy of an earthquake through the lithosphere. Seismic waves in the Earth attenuate with distance at rates greater than predicted by geometrical spreading. The contributing factors are intrinsic attenuation due to the medium anelasticity, and scattering attenuation associated with the inhomogeneities. Knowledge of the relative contributions of scattering ($Q_s^{-1}$) and intrinsic ($Q_i^{-1}$) attenuation is important for appropriate subsurface material identification, tectonic interpretations and quantification of the ground motion (e.g. Hoshiba 1993; Akinci et al. 1995; Del Pezzo et al. 1995; Bianco et al. 1999, 2002). Attenuation inferred from the decay rate of the coda (Aki & Chouet 1975; Singh & Herrmann 1983; Sato & Fehler 1998) is a combination of scattering and intrinsic attenuations. The intrinsic attenuation is associated with small-scale crystal dislocations, friction, and movement of interstitial fluids. The scattering attenuation, associated with an elastic process of redistributing wave energy by reflection, refraction and conversion at irregularities in the medium, is often characterized by an exponential attenuation quality factor, $Q_i$. The latter process is not true anelasticity but has virtually indistinguishable effects that are not accounted for by simple earth models. Unlike $Q_i$ defined for anelastic processes, $Q_i$ is not a measure of energy loss per cycle but, rather, a measure of energy redistribution.

Short period coda waves recorded during the occurrence of local earthquakes, which arrive at station after the arrival of all direct phases, are assumed to be the superposition of backscattered primary $S$ waves generated by the numerous heterogeneities distributed randomly in the Earth’s crust and upper mantle. These waves arrive at the station on different time intervals and form a coda. Therefore, the decay of these waves with time, in a seismogram, provides the average attenuation characteristics of the medium instead of the property of a single path connecting the source to the station. As these waves are the result of numerous heterogeneities distributed randomly, they cannot be explained by any deterministic approach in which a number of parameters are required to describe a small portion of seismogram. However, they can be solved by applying the statistical method in which a small number of parameters are sufficient to describe the properties of coda waves. Aki (1969) and Aki & Chouet (1975) are pioneers in this field and they proposed...
a single backscattering model to use the coda waves of local earthquakes for the estimation of quality factor \(Q_c\) of coda waves in a region. In the single-backscattering model, the coda is considered as a superposition of backscattered wavelets from randomly distributed heterogeneities (Aki 1969; Aki & Chouet 1975; Rautian & Khalturin 1978). Gao et al. (1983) represented another attenuation model for coda wave, which is known as the multiple scattering model. However, this model is mainly used for long (>100 s) lapse times (Kumar et al. 2005), which is beyond the scope of this paper.

The attenuation of seismic waves in the lithosphere is an important property for studying the regional earth structure and seismotectonic activity. Numerous studies have been carried out in different parts of the world to determine the seismic wave attenuation properties of the medium (Frankel 1991). These studies analyse the coda waves of local earthquakes for estimation of \(Q_c\) values, using the single backscattering model. In this study, for understanding of variation of \(Q_c\) values in the Alborz region and central Iran laterally and with depth, data sets for epicentral distance \(R < 100\) km (data set 1) are selected for each station and computed \(Q_c\) values are used to check the spatial variation of \(Q_c\). Following similarity in \(Q_c\) values of some stations, a common \(Q_c\) relation over these areas is presented. The results are compared with measured coda waves in a few representative areas of the world, such as the Himalayan collision zone, and available \(Q_c\) measurements in tectonically active areas of Iran. To investigate attenuation effects at different depths of each tectonic area, two data sets for each area comprising seismograms recorded at distance ranges of \(R < 100\) km and \(100 < R < 200\) are produced and \(Q_c\) values in 14 lapse times (25, 30, 35, ... 90) are studied. From this, depth-dependent variations are estimated and compared with other available studies in our study areas. Some possible implications are discussed at the end.

2 TECTONIC SETTING

The Alpine–Himalayan seismic belt is recognized as one of the seismically active areas of the world. The Iranian plateau, situated on this belt, has experienced several major and destructive earthquakes in the recent past. Deformation and seismicity due to shortening in Iran accommodates the northward motion of the Arabian shield into Eurasia. The seismicity within Iran suggests that much of the deformation is concentrated in the Zagros, Alborz and Koppeh Dagh mountains, and in east Iran. These areas surround central Iran and the Lut desert, which are virtually aseismic and behave as relatively rigid blocks (Fig. 1a; Jackson & McKenzie 1984). This view is supported by the limited GPS data (Vernant et al. 2004; Masson et al. 2005).

The Alborz is an active, E–W trending mountain belt 100 km wide and 600 km long, which was formed when a piece of the Gondwana collided with Eurasia in the Late Triassic (Sengor et al. 1988). This mountain range is bounded by the Talesh mountains to the west, ∼37°N 49°E and by the Koppeh Dagh mountains at about 56°E to the east, and consists of several sedimentary and volcanic layers of Cambrian to Eocene ages that were deformed during the late Cenozoic collision (Stocklin 1974; Berberian 1983; Alavi 1996). The Alborz is extremely steep, with the flanks abruptly joining the plains along major thrust faults on both northern and southern sides (Berberian & Yeats 1999). The Quaternary volcano of Damavand (height 5671 m) is located in the centre of the belt (Fig. 1b). The Alborz mountain range shows strong tectonic activity with several destructive earthquakes in the past including the catastrophic 1990 June 20 (\(M_w\) 7.3) Rudbar earthquake (denoted by star in Fig. 1a).

Figure 1. (a) Location of different tectonic areas in Iranian plateau. Topography, active faults location (Hessami et al. 2003) and location of earthquake epicentres (\(M_w > 4\)) during the period 1964–2006 (Engdahl et al. 2006) are plotted to show variation of seismicity and structure across Iran. Seismicity is concentrated in the Zagros (Z), Talesh (T), Alborz (A) and Koppeh Dagh (K) mountain belts, which surround relatively aseismic central Iran (C) and the Lut desert (L). Central Iran consists of a mosaic of various tectonic blocks, known as Yazd (Y), Tabas (Tb) and Great Kavir (GK) blocks from south to the north. Solid lines represent the active faults and white star denotes catastrophic 1990 June 20 (\(M_w\) 7.3) Rudbar earthquake. The study region (boxed portion) is shown in greater detail in (b). (b) IGUT station locations over volcanic and intrusive rocks and salt extrusions in Alborz and central Iran. The stations are being operated by three provincial networks of IGUT, namely, Tehran, Semnan and Mazandaran (Sari). Inverted triangles and triangles, respectively, show stations located in Alborz and central Iran while a thick solid line separates the two areas. Stations located on border (solid line) receiving data from both regions simultaneously are shown by both symbols. The location of Damavand volcano in centre of the Alborz mountain range is shown too.
which caused 80 km of coseismic left-lateral surface ruptures in the western Alborz (Berberian et al. 1992). Several active faults affect the central Alborz (i.e. see Berberian 1983; Berberian et al. 1993, 1996; Trifonov et al. 1996; Berberian & Yeats 2001; Allen et al. 2003). Most are parallel to the range and accommodate the present-day oblique convergence across the mountain belt.

Recent large earthquakes occurring in this region suggest that the seismicity is connected with major faults of recent age that cut across the regional Quaternary lineaments.

Central Iran consists of a mosaic of various tectonic blocks, known as the Yazd, Tabas and Great Kavir blocks from the south to the north (Fig. 1a). Most of the seismic deformation has been concentrated within the deformational zones among these rigid blocks. According to Berberian (1979), central Iran is not a linear seismic zone. It is characterized by scattered seismic activity with large magnitude earthquakes, long recurrence periods and seismic gaps along several Quaternary faults. The earthquakes in central Iran are generally shallow and are usually associated with surface faulting. The area covered in this study, which hereafter will be called central Iran, is in fact the northern part of central Iran, that is, the Great Kavir block. The Tertiary Great Kavir basin of central Iran is an intracontinental basin filled with highly evaporitic sediments. The area has a unique abundance of exposed salt diapirs observed today in grand scale in two large and many small salt domes (Fig. 1b) within the basin (Jackson et al. 1990).

3 DATA

We use digital ground-motion data from the IGUT short period stations. These short period networks have been operated by the Institute of Geophysics, University of Tehran for the last 13 yr in the Alborz region and the northern part of central Iran. In this study, we have used the data collected during 1996–2004 to calculate Qc for the Alborz region and central Iran, defined as a region bounded between 34°–36.6°N and 49.5°–54°E. The epicentral distance, depth and magnitude of recorded seismograms range from 20 to 200 km, 1 to 30 km and 2 to 5, respectively.

A total of 345 events that occurred in the study region are located using data from 18 stations of three provincial networks of IGUT, namely, Tehran, Semnan and Mazanadan (Fig. 1b). The networks are equipped with a three-component short-period Kinemetrics SS-1 seismometer (i.e. 1-Hz corner frequency) and a 24-bit digitizer. The instrumental velocity response is flat in the frequency band of 1–25 Hz. The sampling rate is 50 samples per second. Most of the stations are installed on rock (Ghods & Sobouti 2005). The distribution of IGUT stations are shown in Fig. 1(b). The ray path distribution for epicentral distance ranges of less than 100 and between 100 and 200 km are plotted in Fig. 2 for each tectonic area. Ray paths are dense enough to present a reliable quality factor in the Alborz mountain range and central Iran. 1746 records of the well relocated events having a signal-to-noise ratio of greater than 2 are included in our study. The number of stations whose data are used for Qc estimation for each event ranges from 1 to 18.

4 METHODOLOGY AND DATA ANALYSIS

The coda waves of 1746 seismograms of 345 local earthquakes that passed the quality condition have been analysed to estimate the quality factor (Qc) for Alborz and central Iran regions by employing the well-known single backscattering model (Aki 1969, Aki & Chouet 1975). To estimate Qc values, we follow the method described by Rahimi & Hamzehloo (2008) and a brief outline of the method is presented here. According to this method, the coda amplitude can be approximately expressed by the following formula:

\[
A(f, t) = C(f) \exp(-\pi ft/Qc).
\]

where C(f) is the coda source factor which is considered as constant, a is the geometrical spreading factor (a = 1 for body waves), and Qc is the quality factor of coda waves representing the average attenuation properties of the medium for a given region. A(f, t) is the time series of amplitudes of the filtered seismograms that is calculated using eq. (2):

\[
A(f, t) = \sqrt{x(f, t)^2 + H[x(f, t)]^2}.
\]

Here, x(f, t) is coda wave of filtered seismograms and H is its Hilbert transform function. Taking the natural logarithm of eq. (1) and rearranging the terms, we get,

\[
\ln[A(f, t), t] = c - bt,
\]

where c = ln C(f) and b = \(\pi f/Qc\).

Quality factor Q remains more or less constant from \(\sim 10^{-4}\) to \(\sim 1\) Hz because of the superposition of different Debye peaks, which represent effects of different attenuation mechanisms (Lay &

Figure 2. The ray path distribution for epicentral distance ranges less than 100 (left-hand panel) and between 100 and 200 km (right-hand panel) for each tectonic area. Dark grey ray paths are for the Alborz region. Light grey ray paths are for central Iran.
window lengths were selected at 2 s for the estimation of attenuation at different average lapse times. These lapse time windows are taken from 25 to 90 s with a variation of 5 s to estimate nine central frequencies and 14 coda window lengths. These windows are located in central Iran. In the next section the results are discussed in detail.

Filtered seismograms are used for the detailed study of the decay of coda wave amplitudes with time to estimate $Q_c$ values at nine central frequencies and 14 coda window lengths. These windows are taken from 25 to 90 s with a variation of 5 s to estimate the attenuation at different average lapse times. These lapse time window lengths were selected at $2t_r$ to avoid the data of the direct S waves, where $t_r$ is the traveltime of the S wave (Havskov & Ottemoller 2003). The low cut-off and high cut-off of nine frequency bands are given in Table 1. An increasing frequency band was used for increasing central frequency to avoid ringing and to take constant relative bandwidths for getting an equal amount of energy into each band, as suggested by Rautian & Khalturin (1978) and Havskov & Ottemoller (2003).

Figs 3–5 show the estimation procedure of $Q_c$ in the central frequencies of 1, 3, 6 and 12 for ANJ, VRN and KIA for lapse times window length of 30 s. These stations have, respectively, high, moderate and low $Q_c$ values. In these figures, the logarithmic plot of geometrical spreading corrected and smoothed coda amplitude as a function of lapse time is shown for selected windows of seismograms and then fitted by eq. (3) at four central frequencies. The slope of this least-squares straightline, corresponding to each plot, provides the $Q_c$ value for each central frequency using eq. (3). In this procedure, only amplitudes with a signal-to-noise ratio greater than 2 and estimates of $Q_c$ with correlation coefficients of the linear regression greater than 0.5 will be considered. For the estimation of attenuation relation in the form $Q_c = Q_0 \pm \Delta Q_c f^{2\pm\Delta n}$, the least-square technique between the logarithm of the selected $Q_c$ values and the logarithm of frequency is done, where $Q_0$ is $Q_c$ at 1 Hz, $n$ is the frequency relation parameter and the terms after the ± symbol represent errors in the estimated parameters. For the three stations mentioned in Figs 3–5, the power law plot is shown in Fig. 6 for lapse time 30 s and for epicentral distance ranges $R < 100$ and $100 < R < 200$ km. It is observed that $Q_0$ values for $R < 100$ km is less than that for $100 < R < 200$ km and the $n$ value does not change significantly and its variation lies within the error ranges. Station RAZ has the lowest and station VRN the highest $Q_c$ value, although both of them are located in central Iran. In the next section the results are discussed in detail.

### Table 1. Central frequency components of bandpass filter with low and high cut-off frequencies.

<table>
<thead>
<tr>
<th>Low cut-off (Hz)</th>
<th>Central frequency (Hz)</th>
<th>High cut-off (Hz)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.00</td>
<td>1.5</td>
<td>2.00</td>
</tr>
<tr>
<td>1.33</td>
<td>2.0</td>
<td>2.67</td>
</tr>
<tr>
<td>2.00</td>
<td>3.0</td>
<td>4.00</td>
</tr>
<tr>
<td>2.67</td>
<td>4.0</td>
<td>5.33</td>
</tr>
<tr>
<td>4.00</td>
<td>6.0</td>
<td>8.00</td>
</tr>
<tr>
<td>5.33</td>
<td>8.0</td>
<td>10.67</td>
</tr>
<tr>
<td>8.00</td>
<td>12.0</td>
<td>16.00</td>
</tr>
<tr>
<td>10.67</td>
<td>16.0</td>
<td>21.33</td>
</tr>
<tr>
<td>16.00</td>
<td>20.0</td>
<td>24</td>
</tr>
</tbody>
</table>

Wallace 1995). However, from $\sim 1$ Hz onwards $Q$ estimated by any method (e.g. $Q_c$ estimated using coda of local earthquakes) increases with increasing frequency. From eq. (1) it is obvious that as $Q_c$ increases attenuation decreases. This would mean that from $\sim 1$ Hz onwards attenuation would decrease with increasing frequency. This is counterintuitive, as observation shows that with increasing frequency attenuation increases. This can be explained by the fact that the term $f$ (frequency) in the exponential factor representing attenuation of seismic wave amplitude in eq. (1) overcomes the effect of increasing $Q_c$ with increasing $f$ and hence the attenuation factor $\pi f t / Q_c$ increases with increasing $f$.

Figure 3. Figure showing original seismogram (lower left-hand panel) recorded at station ANJ and the coda window (lower right-hand panel) used for estimation of $Q_c$. The four upper left-hand panels show the coda filtered at central frequencies of 1, 3, 6 and 12 Hz, respectively. The corresponding right-hand panels show the coda envelope (grey line) and the least-square line fit (black line). $R^2$ represents the correlation coefficient for the least-squares line fit and $Q$ gives the estimated $Q_c$ value.
5 RESULTS

5.1 Lateral variation of $Q_0$ in Alborz and central Iran

We calculated $Q_0$ in each station separately to investigate the lateral variation of upper lithosphere attenuation structure in our study area that includes two tectonically different regions. Based on the fact that small-scale lateral variation in medium property is expected to be more common in the upper lithosphere, we concentrated on the study of coda waves with shorter lapse times and for smaller epicentral distances because they are less affected by deeper parts of the lithosphere and the heterogeneity of the upper lithosphere is better manifested in them. So, data sets including epicentral distance...
range of 0–100 km are selected for each station and \( Q_0 \) for the first lapse time window (= 25 s) were calculated. The obtained results including the quality factor at reference frequency 1 Hz, that is, \( Q_0 \), attenuation parameter, \( n \), and the result of statistical analysis including correlation coefficients and standard deviations are presented in Table 2. For two stations (LAS and SHR) located on the border of two areas, two separated data sets from ray paths received from Alborz and central Iran are included, so two values are presented for each one in Table 2. To compare the results in all stations distributed all over the region, \( Q_0 \) values are plotted in Fig. 7. In general, it could be seen that some lateral variation is observed over the regions and the average value of \( Q_0 \) in the seismically less active part of the study area, that is, central Iran, is greater than that in the more active Alborz region.

5.1.1 Alborz

Comparison of \( Q_0 \) values obtained from data in 11 stations of the Alborz region (inverted triangles in Fig. 7) shows that the calculated \( Q_0 \) values are not uniform and roughly increased towards the edges (e.g., at GLO, PRN, LAS and SHR). The maximum \( Q_0 \) value (= 97) is observed in one of these stations (LAS) located in the southern border of Alborz, whereas the minimum \( Q_0 \) values (59 and 63) are seen in SHM and KIA located in the middle of the region, between stations LAS and GLO.

The variation of \( Q_0 \) values may be caused by differences either in geological environment or lithospheric structure in the region. However, to discuss the existence of such differences in Alborz, we should be certain that the average epicentral distance of each data set is similar. We check this parameter because the rays that propagate to the longer distances penetrate to greater depths where the attenua-

![Figure 6. The power law plot of ANJ, VRN and KIA for lapse time 30 s and for epicentral distance ranges \( R < 100 \) km and \( 100 < R < 200 \) km.](image)

![Figure 7. Map of \( Q_0 \) at epicentral distance of 0–100 km for 25 s time window lengths. Triangles and inverted triangles show stations located in central Iran and Alborz, respectively. Stations located on the border (solid line) receive data from both regions simultaneously and thus they have two \( Q_0 \) values, shown by both symbols. Inset shows scale for \( Q_0 \).](image)

### Table 2. The quality factor at reference frequency 1 Hz, that is, \( Q_0 \), attenuation parameter, \( n \), and the result of statistical analysis including correlation coefficients, \( R^2 \), and standard deviations (presented after ±) for epicentral distance range of 0–100 km and lapse time window 25 s.

<table>
<thead>
<tr>
<th>Station (Alborz)</th>
<th>( Q_0 ) ± σ</th>
<th>( n ) ± σ</th>
<th>Average distance (km)</th>
<th>( N )</th>
<th>( R^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>SHM</td>
<td>59 ± 2</td>
<td>1.08 ± 0.07</td>
<td>47</td>
<td>39</td>
<td>0.97</td>
</tr>
<tr>
<td>KIA</td>
<td>62 ± 1</td>
<td>1.02 ± 0.06</td>
<td>45</td>
<td>24</td>
<td>0.98</td>
</tr>
<tr>
<td>ALA</td>
<td>68 ± 2</td>
<td>1.14 ± 0.10</td>
<td>57</td>
<td>52</td>
<td>0.93</td>
</tr>
<tr>
<td>DMV</td>
<td>69 ± 2</td>
<td>1.11 ± 0.07</td>
<td>51</td>
<td>116</td>
<td>0.98</td>
</tr>
<tr>
<td>GZV</td>
<td>69 ± 2</td>
<td>1.09 ± 0.12</td>
<td>53</td>
<td>14</td>
<td>0.91</td>
</tr>
<tr>
<td>FIR</td>
<td>71 ± 2</td>
<td>1.08 ± 0.07</td>
<td>60</td>
<td>82</td>
<td>0.96</td>
</tr>
<tr>
<td>AFJ</td>
<td>79 ± 2</td>
<td>1.05 ± 0.08</td>
<td>59</td>
<td>73</td>
<td>0.96</td>
</tr>
<tr>
<td>PRN</td>
<td>83 ± 2</td>
<td>1.06 ± 0.08</td>
<td>66</td>
<td>116</td>
<td>0.96</td>
</tr>
<tr>
<td>SHR</td>
<td>82 ± 2</td>
<td>1.02 ± 0.11</td>
<td>72</td>
<td>31</td>
<td>0.93</td>
</tr>
<tr>
<td>GLO</td>
<td>80 ± 2</td>
<td>1.06 ± 0.06</td>
<td>73</td>
<td>23</td>
<td>0.97</td>
</tr>
<tr>
<td>LAS</td>
<td>98 ± 2</td>
<td>0.93 ± 0.13</td>
<td>68</td>
<td>32</td>
<td>0.92</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Station (central Iran)</th>
<th>( Q_0 ) ± σ</th>
<th>( n ) ± σ</th>
<th>Average distance (km)</th>
<th>( N )</th>
<th>( R^2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>RAZ</td>
<td>54 ± 2</td>
<td>1.10 ± 0.08</td>
<td>48</td>
<td>20</td>
<td>0.96</td>
</tr>
<tr>
<td>LAS</td>
<td>57 ± 1</td>
<td>1.15 ± 0.05</td>
<td>52</td>
<td>18</td>
<td>0.98</td>
</tr>
<tr>
<td>QOM</td>
<td>78 ± 2</td>
<td>1.01 ± 0.11</td>
<td>58</td>
<td>24</td>
<td>0.93</td>
</tr>
<tr>
<td>SHR</td>
<td>85 ± 2</td>
<td>0.98 ± 0.12</td>
<td>80</td>
<td>18</td>
<td>0.92</td>
</tr>
<tr>
<td>SFB</td>
<td>86 ± 2</td>
<td>1.18 ± 0.09</td>
<td>80</td>
<td>17</td>
<td>0.94</td>
</tr>
<tr>
<td>HSB</td>
<td>88 ± 2</td>
<td>0.98 ± 0.13</td>
<td>65</td>
<td>48</td>
<td>0.91</td>
</tr>
<tr>
<td>ANJ</td>
<td>95 ± 2</td>
<td>0.95 ± 0.18</td>
<td>68</td>
<td>9</td>
<td>0.83</td>
</tr>
<tr>
<td>MHD</td>
<td>99 ± 3</td>
<td>0.88 ± 0.17</td>
<td>64</td>
<td>22</td>
<td>0.87</td>
</tr>
<tr>
<td>VRN</td>
<td>101 ± 2</td>
<td>0.89 ± 0.10</td>
<td>64</td>
<td>60</td>
<td>0.96</td>
</tr>
</tbody>
</table>

Note: Number of observations, \( N \), and average epicentral distance for each station are also presented.
in Alborz. While an anomalous geological feature can easily destroy the observed order, the order is evident from lack of a strong lateral variation of $Q_c$ over the Alborz region. In addition, the vicinity of SHM and LAS, which are, respectively, showing the minimum and maximum $Q_c$ values confirms the idea, because there is a vast common area monitored by both of them. Therefore, it seems that the observed lateral variation in Alborz is mainly because of the difference in data set properties. The existence of similar fracture and faulting and as a result similar seismicity (Fig. 1a) have made a uniform structure in this study area and therefore we believe that there is no significant lateral variation detectable by coda waves in the upper part of the lithosphere in the Alborz region. Because of this we merge all data sets to obtain the average $Q_c$ frequency relation $Q_c = 79 \pm 2f^{1.97} \pm 0.08$ for this region. We also observe that the observed $n$ values in different stations (Table 2) are similar (within the error bars) over the region and therefore variation of this parameter does not present useful reliable information in this study.

5.1.2 Central Iran

From Fig. 7 it can be seen that stations located in central Iran, except for LAS and RAZ, show $Q_c$ values greater than the average value for the Alborz region (= 74) and they vary from 78 to 101. The general increase of $Q_c$ in central Iran in comparison with Alborz is expected because of clear differences in topography, amount of fracture and faulting and level of seismicity (Fig. 1a). Table 2 shows $Q_c$ and the average epicentral distance for each station. The systematic variation between two parameters is observed at all stations, however, the available average distance this time gives us an extra chance to see if the increase of $Q_c$ value in central Iran in comparison to Alborz is observable in similar average distances or not. The higher $Q_c$ value in central Iran in comparison to Alborz is seen in all stations except LAS and RAZ. These stations, respectively, show $Q_c$ values equal to 54 and 56, which are smaller than the minimum observed value in Alborz (= 59 in SHM). The low $Q_c$ value in LAS and RAZ is not only due to small average distance, but also due to distinct geological features observed around these stations as described below.

While it may seem strange that $Q_c$ in LAS is smaller than GZV and DMV (located in Alborz with similar average distance), the presence of salt extrusions in the study area can justify the anomalous low $Q_c$ value in this station (Fig. 1b). The vast salt domes positioned beside this station at the surface and the buoyant salt layers with the thickness of 6 km under 3 km deposition as the source of salt domes at depth around this area (Jackson et al. 1990) caused high attenuation properties around LAS and make this station distinct from the others. The low $Q_c$ value in RAZ has a completely different reason; it is due to high seismic activity around RAZ. This seismic area in central Iran is positioned at the western edge of the study area where lots of earthquakes occur and many faults are present (Figs 1a and b). Comparison of RAZ and SHM (in Alborz) with similar average distance confirms the idea that low $Q_c$ value is not only because of smaller epicentral distance, but also due to high seismic activity of the region. So, in the study of lithosphere attenuation structure in central Iran, two anomalous areas could be distinguished and even though both of them are located near borders of central Iran, they do not follow the general trend of attenuation properties in the area. Hereafter, to study attenuation in central Iran we will exclude LAS and RAZ. Merging other station data sets we obtain an average attenuation relation of $Q_c = 94 \pm 2f^{0.97} \pm 0.12$. Besides, the attenuation relation for LAS and RAZ are $Q_c = 57 \pm 1f^{1.15} \pm 0.05$ and $Q_c = 54 \pm 2f^{1.10} \pm 0.08$, respectively. The study of attenuation variation with depth in central Iran, which will be presented in next section also, will be performed without these anomalous stations.

5.2 Attenuation variation with depth

Observations by Havskov et al. (1989), Gupta et al. (1998), Kumar et al. (2005) and Mukhopadhyay et al. (2007, 2008) from different regions of the world indicate that $Q_c$ increases with an increase in lapse time. The larger the lapse time, the larger the volume of material with larger lateral and depth extent from which coda waves reach a given station. Hence, the coda at a larger lapse time carries information about volume of material with larger lateral and depth extent. Most workers, however, agree that variation of $Q_c$ with depth affects it more strongly (Abubakirov & Gusev 1990; Ibañez et al. 1990; Hoshiba 1991; Del Pezzo and Patané 1992; Wennerberg 1993; Giampiccolo et al. 2002, 2004; Mukhopadhyay & Tyagi 2007; Mukhopadhyay et al. 2008). Hence, to investigate the variation of attenuation with depth, $Q_c$ value was calculated for 14 lapse time window lengths, which are taken from 25 to 90 s with variation of 5 s. Another parameter which is considered in this section is epicentral distance. The consideration of distance is due to the fact that stations at shorter epicentral distances sample the coda waves from correspondingly shorter zones while those at a greater epicentral distance cover a larger and deeper area. Thus, to study the attenuation properties of study areas with depth in more detail, two data sets with epicentral distances $R < 100$ km (data set 1) and $100$ km $< R < 200$ km (data set 2) were produced and the $Q_c$ value in 14 lapse time windows were calculated. The obtained results beside the result of statistical analysis are presented in Table 3. Fig. 8(a) shows variation of $Q_c (Q_c$ in 1 Hz) with lapse time in central Iran (triangles) and in Alborz (inverted triangles) using the rays with epicentral distances $R < 100$ km. As can be seen, $Q_c$ values in lapse times of 25–45 s increase uniformly for two regions. Observed $Q_c$ values in central Iran are more than that for the Alborz region, similar to what we have seen in previous section for 25 s. This means the Alborz region is more attenuative compared to central Iran. An interesting point in Fig. 8(a) is a change in increasing trend of $Q_c$ in central Iran at 45 s. The figure shows that after 45 s the slope of $Q_c$ for central Iran decreases and this decrease continues until the last lapse time, 90 s. This change causes the average attenuation in central Iran to become more than that in Alborz in the last lapse times (80–90 s).

Fig. 8(b) shows attenuation variation with lapse time in a second data set, 100 $< R < 200$ km, for both regions. It is observed that $Q_c$ values for lapse time less than $\sim 35$ s are similar for both the regions. At larger lapse times $Q_c$ for Alborz is larger than that for central Iran. This is the opposite of the trend observed for $R < 100$ km (Fig. 8a). The rate of increase of $Q_c$ with lapse time decreases for both the data sets for lapse time greater than 55 s. For central Iran it becomes more or less constant. This shows that at deeper regions, the lithosphere below central Iran is more attenuative compared to that below Alborz, whereas at shallower depth the opposite trend is observed. The attenuation parameter, $n$, reported in Table 3, does not vary significantly (i.e. values are comparable within the error limits) over lapse times in both data sets of two study areas.

5.3 Comparison of $Q_c$ with other observations

Seismic wave attenuation studies worldwide (e.g. Aki 1980; Pulli & Aki 1981; Roecker et al. 1982; Van Eck 1988; Akinci et al. 1994; Hellweg et al. 1995; Gupta et al. 1998; Biescas et al. 2007; Mukhopadhyay & Tyagi 2007) have shown that $Q_c$ is normally
high in stable regions and low in active tectonic regions, while its frequency dependence, \( n \), has an inverse behaviour. The results of our estimates in the seismically less and more active parts of the study area, that is, central Iran and Alborz, show that, in general, \( Q_0 \) and \( n \) values are consistent with the idea of the tectonic indicators (Table 3).

We compare our results to other similar studies performed in Alborz and central Iran. Motazedian (2006) measured the attenuation of \( S \) waves (\( Q_s \)) in northern Iran, including our study areas, from acceleration records of Building and Housing Research Center (BHRC) stations. The frequency-dependent relation reported by Motazedian (2006) is \( Q_s = 87.9^{1.44} \). The values of the \( Q_0 \) and \( n \) given by him are similar with our estimated values for shorter lapse times. Recently, some research has been conducted in Iran to calculate the \( Q_s \) relations in northern part of Zagros continental collision zone (Rahimi & Hamzehloo 2008) and east central Iran (Ma’hood & Hamzehloo 2009). The \( Q_s \) values for north of Zagros (Silakhor area) were estimated before and after the 2006 Darb-e-Asatine earthquake of \( M_w 6.1 \) using its foreshocks and aftershocks and are reported separately. The \( Q_s \) frequency-dependent relation for the region is \( Q_s = 117^{0.88} \) and \( Q_s = 99^{0.84} \) for foreshocks and aftershocks, respectively.

In Fig. 9 we compare reported \( Q_s \) values obtained through aftershocks with our results. As these relations were estimated for earthquakes that occurred exactly before and after the main event, these cannot be considered as tectonic detectors to compare with other regions because of the temporal change of stress in the main shock area. However, for both of the reported frequency relations, two tectonic indicators show that Silakhor area is less active than Alborz and central Iran.

The second region, east central Iran, is an intracratonic active zone in the western rim of the Lut block. The obtained \( Q_s \) relation for west of Lut is \( Q_s = 101^{0.94} \). Comparing these \( Q_s \) and \( n \) values with those obtained in this study, it is clear that these values are similar to the values obtained for central Iran and from Fig. 9 it is observed that for the same frequencies the \( Q_s \) values are similar to that for central Iran. The seismicity of the region west of Lut is similar to central Iran (Fig. 1). Seismically, Lut and central Iran are less active than Alborz. Mirzaei et al. (1998) have reported them as a major seismotectonic province with the common name of central-east Iran.

We have compared some of the results of \( Q_s \) studies obtained by various researchers worldwide (Fig. 9). These studies generally show low values of \( Q_0 \) (e.g. <200) for seismotectonically active regions, high \( Q_0 \) values (e.g. >600) for seismically inactive stable regions and intermediate values for moderate regions. Both of our calculated \( Q_s \) values for two regions fall in the active part. The obtained values are comparable with the \( Q_s \) frequency trend of the northwestern Himalayas (\( Q_s = 113^{1.01} \); east central Iran (\( Q_s = 101^{0.94} \); NW Himalayan (\( Q_s = 93^{1.11} \); southwestern British Columbia, Canada (\( Q_s = 110^{0.73} \); South Korea (\( Q_s = 95^{0.94} \); South Africa and New York (Fig. 9). These values are less than the obtained values for Hong Kong (\( Q_s = 256^{0.94} \); east Canada (\( Q_s = 480^{0.94} \); Koyna-Warna (\( Q_s = 169^{0.77} \)) and are more than the values of northeastern India (\( Q_s = 52^{1.32} \); SE Sabalan, NW Iran (\( Q_s = 49^{0.96} \) and southern California (Fig. 9). Of all the \( Q_s \) values, the \( Q_0 \) and \( n \) values of the South Korea is most similar to central Iran, possibly because of the similarity in tectonic activities between South Korea and central Iran. From Fig. 9, it could be seen that our \( n \) value falls within the range of values obtained for tectonically active regions by these and other research groups. However, our \( n \) values are close to 1, like those reported for tectonically active regions (Mukhopadhay & Tyagi 2007).
Figure 8. Variation of average $Q_0$ ($Q_c$ in 1 Hz) with lapse time and depth in central Iran (triangles) and in Alborz (inverted triangles) for epicentral distance ranges (a) $R < 100$ km and (b) $100 < R < 200$ km.

Figure 9. Comparison of $Q_c$ of the Alborz Region and central Iran with those reported from other regions of the world. The plots for other regions are, respectively, obtained from Mukhopadhyay et al. (2006), Frankel (1991), Mandal & Rastogi (1998), Woodgold (1990), Kumar et al. (2005), Zelt et al. (1999), Hazarika et al. (2009), Maka et al. (2004), Yun et al. (2007), Rahimi & Hamzehloo (2008, 2009) and Ma’hood & Hamzehloo (2009).

6 DISCUSSION

Two data sets with epicentral distances $R < 100$ km and $100 < R < 200$ km and different lapse times allow us to measure the coda wave decay at different depths of two tectonically different study regions. A rough estimate of maximum depth through lapse time and average epicentral distance is achieved using the single backscattering model of Aki (1969) and Aki & Chouet (1975). In this model, the estimated attenuation of coda wave is the average decay of amplitudes of backscattered waves on the surface of ellipsoid volume.
having earthquake source and station as foci (Pulli 1984; Gupta et al. 1998; Mukhopadhyay & Tyagi 2007; Mukhopadhyay et al. 2008). The observed $Q_0$ reflects the average attenuation properties of the volume of ellipsoid with a maximum depth $h = a_2 + h_0$, where $a_2 = \sqrt{a_1^2 - (\Delta/2)^2}$ is the small semi-axis of the ellipsoid with $\Delta$ as the average epicentral distance and $h_0$ is the average focal depth of the earthquakes (about 10 km for our study region) (Pulli 1984; Havskov et al. 1989; Canas et al. 1995). The large semi-axis of the ellipsoidal volume is $a_1 = \beta t/2$ for lapse time $t$ and velocity $\beta$ of the $S$ wave. We have taken $\beta = 3.5$ km s$^{-1}$. The average lapse time is taken as $t = t_0 + W/2$, where $t_0$ is the starting time of the coda window and $W$ is the lapse time window length.

To calculate the maximum depth of scatterers responsible for the generation of coda waves, in each lapse time and in each data set it is required to have $\Delta$ and $t_0$ related to four data sets including rays with epicentral distances $R < 100$ km and $100 < R < 200$ km in both study areas. The values of $\Delta$ and $t_0$ in the first data set in central Iran are, respectively, 67 km and 36.9 s and in Alborz 62 km and 33.3 s. The values in the second data set are 137 km and 75.7 s in central Iran and 136 km and 76.1 s in Alborz. The small difference between $\Delta$ and $t_0$ for both areas in each data set allows us to take an average value for each parameter in each data set and have the same $\Delta$ and $t_0$ for both areas. While this work enters some errors in evaluating the depth values, it eases comparing the attenuation of two areas in depth. $\Delta$ and $t_0$ as average values in two areas are taken as 64.5 km and 35.2 s in first data set and 136.5 km and 75.9 s in second one.

By considering these values, the lapse time 45 s ($W = 45$) in the first data set, which displays a clear change in attenuation variation curve of central Iran, shows the maximum depth of $\sim 106$ km (Fig. 8a). This result is in correlation with the result of teleseismic $P$-wave velocity tomography carried out by Paul et al. (2009) on a temporary local network present in central Iran and Alborz. The obtained tomogram shows a distinct velocity contrast between central Iran and the lower velocity Alborz region in depths less than 100 km. The study of the $S$-velocity structure of crust and upper mantle beneath the Iranian plateau (Shadmanamen & Shomali 2009) showed similar results for the study area. It is well known that regions of higher velocity show higher $Q$ (i.e. lower attenuation). Our results show this to be true for the study area. In addition, the uniform increase of $Q_0$ before 45 s lapse time in data set 1 ($R < 100$ km) means that attenuation in the lithosphere of both regions decreases with increasing depth; that is, with increasing depth, the medium becomes more homogeneous (Mukhopadhyay et al. 2008).

The observed change in the curve for central Iran in the first data set (Fig. 8a), which was interpreted as a change in attenuation structure after $\sim 100$ km ($W = 50$ s), continues to depth 147 km ($W = 90$ s). This change causes the average attenuation at maximum depth of 147 km in central Iran to become more than that in Alborz. This observation implies the existence of a high attenuation (low velocity) anomalous structure under the lithosphere of central Iran. Reported velocity tomograms of the upper mantle beneath the Iranian plateau (Alinaghi et al. 2007; Shadmanamen et al. 2009) show an anomalously low velocity anomaly beneath central Iran and Alborz but they are not able to distinguish any differences between these two adjacent areas. However, the velocity tomogram presented by Paul et al. (2009) shows an anomalously low velocity anomaly just under central Iran between 150 and 200 km depth.

Fig. 8(b) shows attenuation variation with lapse time and depth in the second data set including epicentral distance 100 $< R < 200$ km. In this data set the $Q_0$ at shortest lapse time ($W = 25$ s) represents the average $Q_0$ over a maximum depth of 149 km. The $Q_0$ curve in this figure at and before lapse time 40 s (163 km) shows that at such depth the attenuation in the Alborz and central Iran regions is similar and decreases rapidly with increasing depth, as $Q_0$ versus lapse time has a steep slope. However, this steep slope changes to a gentler slope at 45 s for both curves and becomes more or less a flat trend after 55 s (178 km) for central Iran. In other words, after $\sim 180$ km, the $Q_0$ value does not vary much with depth. From Fig. 8(b) it can be said that in this part of the mantle attenuation becomes more or less constant for central Iran and it decreases less rapidly compared to lithosphere for the Alborz region. This could also indicate the presence of a transparent mantle (Aki 1973; Flatte and Wu 1988) after $\sim 180$ km under study area. The existence of a transparent mantle beneath parts of Himalaya has been reported by Mukhopadhyay et al. (2008) in the source region of the 1999 Chamoli Earthquake, India.

7 CONCLUSION

We analysed local earthquake waveforms recorded on the short-period seismic network of IGUT in Alborz and central Iran to study lateral and depth variation of attenuation structure using coda waves. The study of $Q_0$ in Alborz showed no significant lateral variation and the average $Q_0$ frequency relation is given as $Q_0 = 79 \pm 2^{1.07 \pm 0.08}$ for this region. In central Iran two anomalous high-attenuation areas were distinguished around LAS and RAZ and a separate attenuation relation was presented for these two stations. The high attenuation of LAS is due to the vast salt domes and layers around this station. RAZ is located in a seismically active area in the western border of central Iran and high attenuation is due to high heterogeneity. Excluding those, we found similar lithosphere attenuation structure over other parts of central Iran and the average $Q_0$ frequency relation for this region is given by $Q_0 = 94 \pm 2^{0.07 \pm 0.12}$. The observed relations show low values of $Q_0$ (less than 200), which is a fingerprint for tectonically active and highly heterogeneous regions. To investigate the attenuation variation with depth, the $Q_0$ value was calculated for 14 lapse time windows for two data sets. We found $Q_0$ increases with increasing lapse time in both the areas; however, the rate of increase is not uniform. As $Q_0$ and $Q_0$ are inversely proportional to attenuation, it means that attenuation decreases with increasing depth; that is, with increasing depth, the medium becomes more homogeneous. After $\sim 100$ km the rate of increase of $Q_0$ decreases for central Iran and beyond $\sim 180$ km it becomes more or less constant. This coincides with an anomalous low-velocity structure under lithosphere of central Iran. Even for the Alborz region the rate of increase of $Q_0$ with depth becomes smaller beyond this depth. After $\sim 180$ km, the $Q_0$ value does not vary much with depth. This indicates the presence of a medium at those depths where attenuation does not vary much with depth and the presence of a transparent mantle under both study areas.

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