Local and regional *P*-wave spectral attenuation model for Iran

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SUMMARY

A robust frequency-dependent local and regional *P*-wave attenuation model is estimated for continental paths in the Iranian Plateau. In order to calculate the average attenuation parameters, 46 337 vertical-component waveforms related to 9267 earthquakes, which are recorded at the Iranian Seismological Center (IRSC) stations, have been selected in the distance range 10–1000 km. The majority of the event's magnitudes are less than 4.5. This collection of records provides high spatial ray path coverage. Results indicate that the shape of attenuation *P*-wave curve versus distance is not uniform and has three distinct sections with hinges at 90 and 175 km. A trilinear model for attenuation of *P*-wave amplitude in the frequency range 1–10 Hz is proposed in this study. Fourier spectral amplitudes are found to decay as $R^{-1.2}$ (where R is hypocentral distance), corresponding to geometric spreading within 90 km from the source. There is a section from 90 to 175 km, where the attenuation is described as $R^{0.8}$, and the attenuation is described well beyond 175 km by $R^{-1.3}$. Moreover, the average quality factor for *Pg* and *Pn* waves (Q_{Pg} and Q_{Pn}), related to anelastic attenuation is obtained as $Q_{Pg} = (54.2 \pm 2.6)f^{(1.0096 \pm 0.07)}$ and $Q_{Pn} = (306.8 \pm 7.4)f^{(0.51 \pm 0.05)}$. There is a good agreement between the results of the model and observations. Also, the attenuation model shows compatibility with the recent regional studies. From the results it turns out that the amplitude of *P* waves attenuates more rapidly in comparison with the global models in local distances.

Key words: Body waves; Seismic attenuation; Wave propagation; Wave scattering and diffraction.

1 INTRODUCTION

Investigation of seismic wave attenuation has an important role in the studies of engineering seismology and seismology. In engineering seismology, the accuracy of seismic hazard maps—of the regions strongly depends on how well the used ground motion prediction equations (GMPE) predict the attenuation of the seismic waves at different distances. Also, simulation of large earthquakes, which have engineering impact on different structures, is concerned with the attenuation of shear and surface waves. In seismology, the waveform modelling and simulation of the seismic wave propagation with different applications, including evaluating and studying structure of the Earth, in local, regional and global scales, are related to the seismic waveform attenuation. Also, the studies of source parameter determination, earthquake magnitude calculation and design or optimization of the seismic networks require the knowledge of seismic wave attenuation.

In assessing the capabilities of seismic networks, the main concern is the ability of the network to detect and locate the small events. Schorlemmer & Woessner (2008) estimated the detection threshold of the seismic network by calculating the probability of detecting an earthquake using past performance of the existing network. D'Alessandro *et al.* (2011), on the other hand, used the simulation methods to evaluate the detection threshold of the seismic network. Also, Kværna & Ringdal (2013), quantified the event detection capability of the seismic network in a specific target area. They considered the ensemble of reported events in the area, and downscale each event magnitude with the observed signal-to-noise ratio (SNR) at the station. In such studies, since the purpose is to estimate the minimum detection magnitude of the network, attenuation characteristics of the seismic waves generated by small events is of great importance.

In strong ground motion modelling, the distance of concern is usually less than 200 km from epicentre of the earthquake. So an important issue in using these attenuation models is the distance limit of the studies. Because of that, most GMPE were developed for this distance range (Douglas & Edwards 2016; Zafarani *et al.* 2018 and NGA West2 models). However, in the case of evaluation detection threshold of a seismic network, with regard to high dynamic range of the modern seismic sensors, and with the help of the array signal processing, it is possible to detect the arrivals of the small events in the regional and global distances. As an example, Gibbons & Ringdal (2006) showed that by the seismic array-based processing method, a small ($m_b = 2.6$) seismic event can be detected at a distance as large as 2300 km. Thus, in such

studies, it is necessary to study the frequency-dependent attenuation of the seismic waves in a wide distance range, from tens to thousands of kilometre.

Reduction of seismic monitoring thresholds requires the use of regional seismic data to constrain the event locations. Global models used in real-time monitoring could not consider geologic complexity of the crust and upper mantle in regional scale. So, some studies like regional seismic travel time (RSTT) model (Myers *et al.* 2010) have started to capture the major effects of 3-D crust and upper-mantle structure on RSTT.

The local and regional structures of the earth have great effects on the attenuation of seismic waves. The attenuation variations cause the systematic variations in seismic wave amplitudes. The seismic wave amplitudes generally decrease, and it changes the frequency content with increasing travel distance through the earth. Some phenomena like geometrical spreading, scattering attenuation and intrinsic attenuation are the most important factors in the attenuation of seismic waves (Sato *et al.* 2012).

S-wave attenuation models in local distances are somehow more widespread than *P*-wave attenuation models. Although, the attenuation of shear waves has a more significant role in engineering studies and seismic hazard estimation, the *P*-wave attenuation models are more important for constraining something specific in seismology. The *P*-wave attenuation curves are not only useful for obtaining source parameters accurately, but also they are helpful for characterizing the earth properties (e.g. Jimenez *et al.* 2005; Chiou *et al.* 2010; Nakajima *et al.* 2013). *P*-wave amplitude–distance curves are also being used in seismic monitoring, determining the type of events and yielding the estimation of earthquakes at regional distances (Wu *et al.* 2016).

Regional short-period *Pn* waves are the most prominent phases of small to medium size earthquakes in the regional distances. The body-wave magnitude (*mb*) provides an empirical estimate of the earthquake size that is directly related to the strength of ground shaking. *mb* scale is based on *P* waves' amplitude measured in the first few seconds on a specific model of a short-period seismograph. The short period improves detection of the smaller events, and discriminates more accurately between the tectonic earthquakes and underground nuclear explosions (Bormann & Wielandt 2013). The International Data Center of the Comprehensive nuclear-Test-Ban Treaty Organization routinely reports *mb* of the seismic events. However, the prior knowledge of regional attenuation is necessary to accurately retrieve the *mb*. Otherwise, its neglect would result in a bias (Bormann 2012).

The first amplitude–distance relationships for body waves were developed by Gutenberg (1945a,b). In these studies, theoretical amplitudes of the seismic waves were determined from geometric spreading of the seismic energy in an earth model and then corrected for attenuation. Evernden (1967) calculated the amplitude decay of *Pn* waves in the eastern and western United States to obtain the magnitude of *Pn* body wave in the regional and near-regional distance ranges (200–2100 km). Veith & Clawson (1972) developed an empirical surface-focus amplitude–distance curve based on the geometrical spreading, and theoretical deep-focus amplitude–distance curves incorporating the geometric spreading and attenuation, with an emphasis on the teleseismic distance ranges. These curves have been used as correction factors for computing the magnitude (*mp*). Sereno Jr *et al.* (1988) also studied *Pn* waves and estimated the frequency-dependent attenuation of the regional wave for continental paths to the NORESS array in Norway at ranges between 200 and 1400 km. Sereno Jr & Given (1990) presented the results of a theoretical study of the sensitivity of *Pn* attenuation to the elastic and anelastic structure of the uppermost mantle for frequencies between 1 and 15 Hz and distances between 200 and 1000 km. Atkinson & Mereu (1992) shear wave attenuation model was derived from the analysis of approximately 1000 digital short-period vertical-component seismograms, that were being recorded from 1980 to 1990 on the Eastern Canada Telemetered Network. Atkinson & Boore (1995) developed the shear wave amplitude attenuation relations for ground motions by using eastern North American earthquakes at the local and regional distances. Priestley & Patton (1997) established the regional magnitude scales, *mb* (*Pn*) and *mb* (*Lg*) by determining the attenuation rates of *Pn* and *Lg* waves. Atkinson (2004, 2015) redefined and improved their shear wave attenuation model by new seismic data. Romanowicz & Mitchell (2007) reviewed and interpreted both the global and many of the regional variations of intrinsic attenuation for both *P* and *S* waves.

The seismic *Pg* waves, upgoing *P* waves from a source in the upper crust, are the first arriving energies observed on seismograms in the local distances. Energy density of *Pg* waves is confined to the crust and is propagated with a group velocity typically between 5.8 and 6.4 km s⁻¹. *Pn* waves are the first arrivals in the regional distances (beyond 200 km) that are refracted *P* waves at Moho discontinuity and are propagated in the uppermost mantle with a group velocity between 7.5 and 8 km s⁻¹ (Storchak *et al.* 2003).

Anelastic absorption and scattering attenuation are usually expressed by the inverse of the quality factor (Q^{-1}). This representation helps to understand the physical laws related to the elastic energy propagation of an earthquake through the lithosphere (Knopoff & Hudson 1964; Aki 1980). In Iran, many studies have been done for quantifying the amplitude–distance curve and frequency-dependent quality factor of coda (*Qc*) and body waves (*Qp* and *Qs*) in different tectonic regions of the country. In these studies, *P*, *S* and multiple lapse time windows and coda wave decay rate have been analysed using the ground-motion records of the local distance earthquakes (e.g. Rahimi & Hamzehloo 2008; Mahood & Hamzehloo 2009; Rahimi *et al.* 2010; Mahood 2014; Meghdadi & Shoja-Taheri 2014; Kaviani *et al.* 2015). Also several works (e.g. Askari *et al.* 2009; Rezapour & Rezaei 2011; Motaghi & Ghods 2012; Meghdadi & Shoja-Taheri 2014) have determined the seismic attenuation model for Iran in local distance by data obtained from local networks. However, there are few detailed studies on frequency and distance dependent of the *P*-wave attenuation parameters in regional scale in the Iranian Plateau.

This study addresses the frequency- and distance-dependent attenuation of *P* waves for the earthquake located in the Iranian Plateau at local and regional distances. The emphasis of the study is not only on the determination of *Qp*, but also on studying the distance dependent variation of *P*-wave attenuation in different frequency ranges. In this regard, we first present the data set used in this study and the signal

processing procedure we followed to prepare the data (Section 2). In Section 3, the formulation of attenuation of the seismic waves is reviewed in brief. The results of the study, the discussion on the results and validation of the attenuation model are provided in Section 4.

2 DATA SET AND SIGNAL PROCESSING

2.1 Waveform data

In order to monitor seismic events of the Iranian plateau, seismic local telemetry networks of the Iranian Seismological Center (IRSC) were installed in the middle of 1990s (web address: <http://irsc.ut.ac.ir> 2019). This network is composed of several provincial subnetworks. At the end of 2018, IRSC had 121 stations with 20 local subnetworks around the country. This network has the official responsibility of determining the location and magnitude of the earthquakes in Iran in a timely manner. This network consists of 71 broad-band stations equipped with Trillium-40 s, Trillium-120s, CMG3ESP-120s and CMG3T-360s, 49 stations equipped with SS1 short-period and one long-period boreholes station (Naserieh *et al.* 2019). Depending on the type of the stations, waveforms have been digitized with a sampling rate of 50 or 100 samples per second. The location of IRSC stations is presented in Fig. 1(a) with yellow and red triangles for broad-band and short-period stations respectively. IRSC has scarce coverage of the South East part of the country. To remedy this gap and having better ray coverage of the events in the region, the detections of CHBR station have been also considered in this study. CHBR is a broad-band station from Broadband Iranian National (BIN) seismic network, which has been established by the International Institute of Earthquake Engineering and Seismology (IIEES, Ansari & Hosseini 2014).

The data bank of this study consists of 46 337 waveforms from 9267 events in 2016. Fig. 1(a) depicts the distribution of the events used in this study. Most of the seismicity within the country is concentrated on Zagros region in west and south west, Alborz in the north and central and eastern part of the country. This pattern of seismicity is also reflected in the model of seismotectonic provinces, as presented by Mirzaei *et al.* (1998) and shown in the key map in Fig. 1(a).

Fig. 1(b) depicts the ray coverage of the events considered in this study. High spatial ray path coverage makes it possible to determine a robust model of the local and regional *P*-phase attenuation in the region of the study. Fig. 1(c) represents the distance–magnitude distribution of the events in data set of this study. The magnitudes of the events range are from M_L 1.5 to 5.5 with epicentral distance from 10 to 1600 km. Most of the earthquakes occurred along the known faults in depth range of 5–20 km in the upper crust (Fig. 1d).

In attenuation relation, distance from event location is an important parameter; so for evaluating the event location in data bank, and to access the uncertainty of the epicentre location, a comparison is made with the locations provided by ISC Reviewed Bulletin (International Seismological Centre, ISC On-line Bulletin 2019). The histogram of the differences between the epicentral location in data bank and Reviewed ISC bulletin is shown in Fig. 2. The mean and the standard deviation of the distribution are 8.5 and 9 km, respectively. 90 per cent of the location difference, for magnitudes smaller than 4 is 27 km, and for those greater is 12 km. Also, 70 per cent and 90 per cent are 9.0 and 15 km respectively for all the events, and the mode of the location difference is about 5 km.

2.2 Signal analysis of waveform data

To create a data base of *P*-wave amplitudes, at the first step, the mean and the linear trend were removed from each waveform, and all vertical components of the waveforms were corrected for the instrument response. In the second step, all waveforms were resampled to 50 Hz. Finally, all the records were bandpass filtered between 1 and 10 Hz using Butterworth filter of order 2 in order to remove long-period noise of the records.

To extract *P* wave from the whole waveform of the earthquake, the nominal time window of *P* waves, as proposed by Kim & Ottemoller (2017), were used. The duration of *P*-wave window is defined as

$$T(R) = GV_{\text{centre}} \pm Tw(R) \quad (1)$$

where GV_{centre} is the central group velocity (epicentral distance (km)/travel time (s)) and $Tw(R)$ is the time window length at epicentral distance (R), which is scaled by distance as

$$Tw(R) = Tw_{\text{ref}} \times \left(\frac{R}{R_{\text{ref}}} \right) \quad (2)$$

where $Tw_{\text{ref}} = 15$ s and $R_{\text{ref}} = 1000$ km were used for the *P* windowing, as suggested by Kim & Ottemoller (2017). According to the studies of Al-Lazki *et al.* (2003), Lu *et al.* (2012) and Motaghi *et al.* (2015), the average *Pn* velocity beneath the Iranian plateau is about 8.0 km s^{-1} , and the maximum velocity perturbations are approximately 3–4 per cent. Kianimehr *et al.* (2018) found a smoothly increasing *Pg* velocity by depth with velocities of 5.8 km s^{-1} at shallow depth and velocities of 6.4 km s^{-1} at the deepest crustal levels of Iran. So, the *Pg* and *Pn* wave windows have been calculated by assuming the average velocities of *Pg* and *Pn* as 6.1 and 8 km s^{-1} , respectively. In this paper, we refer to the time window containing *Pn*, *Pg* or both as *P* window. The waveform of each event was time windowed from *P* phase onset, using 5 per cent cosine taper to reduce the spectral artefact of sharp truncation of a signal. Examples of selected *P* window of the waveforms in different distances are shown in Fig. 3.

Figure 1. (a) Epicentre of the events used in the study in addition to broad-band (yellow triangles) and short-period (red triangles) stations of IRSC and CHBR broad-band station (green triangle) of BIN seismic network, (b) ray path coverage, (c) distance–magnitude distribution of the events in data set and (d) distribution of cumulative frequency of events depth.

Figure 2. The histogram of differences between the epicentral location in data set and Reviewed ISC bulletin. (Brown vertical dashed line ($\bar{x} = 8.5$) shows the mean of differences and the standard deviation of distribution is 9 km; and red line indicates cumulative frequency of distance differences. Brown horizontal dashed line ($\gamma = 0.9$) shows 90 per cent percentile of location difference. Red dotted and red dashed lines are cumulative frequency of distance differences for magnitude smaller than 4 and greater than 4, respectively.)

2016-06-25, 20:14:15.2, 28.168N, 51.584E, h=17.4km, Mn4.1

Figure 3. Vertical records of LIN station ($R = 861$ km), GAR station ($R = 473$ km), JHRM station ($R = 199$ km) and AHRM station ($R = 82$ km) from M_n 4.1 earthquake on 2016 June 24 at 20:14. Grey window shows selected P window of each waveform.

Figure 4. The distributions of the number of selected records over hypocentral distance in the data set in frequencies 2.84 and 5.66 Hz.

For each waveform, SNR was estimated by calculating the ratio between the root mean square (rms) of P -wave amplitude in the time window length $T(R)$, and the RMS of pre- P arrival amplitude, which considered as the noise signal with the same length of window, $T(R)$. We discarded those records for which the SNR was < 2 .

The smoothed Fourier spectrum of the velocity waveform of P window was calculated by computing the average amplitudes at the midpoint of 0.1 log frequency wide sliding window in the central frequencies of 1.13, 1.42, 1.79, 2.25, 2.84, 3.57, 4.5, 5.66, 7.13 and 8.97 Hz. A data base of Pg and Pn amplitudes in various frequencies, hypocentral distances and magnitudes was compiled for calculating the attenuation of P waves.

Fig. 4 exemplifies the distributions of the number of selected recordings counted over distance bins 100 km width for two frequencies. In frequencies 2.84 and 5.66 Hz, number of events in each distance range 100 km is reasonable to calculate the model. The maximum number of selected records is about 2200 and in hypocentral distance range 200–300 km. The number of selected records above 300 km decreases with distances, and distances above 850 km are less than 100 records. At higher frequency, the data base is slightly sparser due to SNR threshold but it could be negligible.

3 SPECTRAL ATTENUATION CURVE

According to the study of Fukushima & Tanaka (1990), Atkinson & Mereu (1992) and Meghdadi & Shoja-Taheri (2014), the relation between amplitude, source and distance term could be generally considered as:

$$\log A_{ij}(f) = \log A_{i0}(f) - b(f) \log(R_{ij}) - c(f) R_{ij} + d_j(f) \quad (3)$$

where A_{ij} is the observed spectral amplitude of event i at station j , A_{i0} is the source amplitude of event i , R_{ij} is hypocentral distance for i th event and j th station, f is the frequency and b , c and d are the constant coefficients that should be determined for each frequency(f). The second and the third terms in eq. (3) are respectively related to the geometrical spreading and the anelastic attenuation, and $d_j(f)$ is the station correction term for station j . The source amplitude of the event is unknown but could be retrieved.

3.1 Geometrical spreading (b) and the anelastic attenuation (c)

As the seismic wave moves away from the source, it spreads over an increasingly larger area. This leads to a decrease in its intensity, commonly called ‘geometric spreading’. Seismic waves propagating through the Earth are attenuated with a distance at rates greater than the predicted ones by the geometrical spreading. Intrinsic anelasticity and scattering attenuation are two other factors in attenuation of seismic waves. Intrinsic anelasticity is related to small-scale crystal dislocations, friction and movement of the interstitial fluids. The scattering attenuation is associated with elastic process of redistributing wave energy by reflection, refraction and conversion at irregularity in the medium (Sato *et al.* 2012).

In local and regional distances, ray paths differ due to the properties of the crustal heterogeneity, geological layers, environmental conditions and also the depth of earthquakes. Pg waves have impulsive high-frequency onset and the whole ray paths are in the upper crust. Therefore, the most effective factors in reducing Pg wave amplitude are to be found within the crust. However, most of the path of Pn phases,

the small amplitude impulses or emergent onsets with lower frequency content, are confined in the uppermost mantle, so the amplitudes of Pn wave are more affected by the uppermost mantle (Asudeh 1982, Kim & Ottemoller 2017). Since the properties of the upper crust and uppermost mantle are different, the decay rates of Pg and Pn amplitudes are distinct (the case of anelastic attenuation and geometrical spreading). It could be shown theoretically that the geometrical spreading coefficient of the body waves in a homogeneous half space is 1 for distances less than 100 km (Aki 1980). But the finding of seismic waves attenuation studies in different part of the world show the attenuation is clearly steeper than R^{-1} (e.g. Atkinson 2004; Herrmann & Malagnini 2004).

Moho depth changes from 56 km beneath the western part of Central Iran to 40 km in Central Iran and 38 km in the East. Also the crustal thickness beneath CHBR station located in Makran is 31 km (Nasrabadi *et al.* 2019). The crust has an average thickness of 42 km beneath Zagros fold, thrust belt and the Central domain. Also, it is thicker beneath the hanging wall of the Main Zagros Reverse Fault with a greater maximum Moho depth in the Central (69 km) than in the Northern Zagros (56 km) (Paul *et al.* 2010). Rajaei *et al.* (2007) using receiver functions from a temporary seismic network, obtained a crustal thickness of 46–48 km in south of Alborz, 55 km in its central part and about 44 km in north of Alborz. Motavalli-Anbaran *et al.* (2011) showed that the crust thickens under the Kopet-Dagh Mountains up to 48 km with a strong slope reaches 37 km underneath the Kopet-Dagh foreland basin.

The Pn phase is the first phase to arrive when the distance between a station and an earthquake exceeds 150 km in the continental regions with a Moho depth of about 35–50 km (Shearer 2019); so if we consider the average Moho depth as about 45 km in Iran, the first arrivals should change from Pg to Pn at the distance of around 150 km.

3.2 Attenuation model and regression

For determining Pg and Pn wave attenuation curve parameters, the hinged trilinear attenuation model of Atkinson & Mereu (1992) was considered as the following equation:

$$\log A_{ij}(f) = \begin{cases} \log A_{i0}(f) - b_1(f) \log\left(\frac{R_{ij}}{R_{ref}}\right) - c_1(f)(R_{ij} - R_{ref}) + d_j(f) & R_{ij} \leq R_1 \\ \log A_{i0}(f) - b_1(f) \log\left(\frac{R_1}{R_{ref}}\right) - b_2(f) \log\left(\frac{R_{ij}}{R_1}\right) - c_1(f)(R_{ij} - R_{ref}) + d_j(f) & R_1 < R_{ij} \leq R_2 \\ \log A_{i0}(f) - b_1(f) \log\left(\frac{R_1}{R_{ref}}\right) - b_2(f) \log\left(\frac{R_2}{R_1}\right) - b_3(f) \log\left(\frac{R_{ij}}{R_2}\right) - c_1(f)(R_2 - R_{ref}) - c_2(f)(R_{ij} - R_2) + d_j(f) & R_{ij} > R_2 \end{cases} \quad (4)$$

To include the effect of the layered media in the geometrical spreading, we allowed b to take separate frequency-dependent values in each of the three distance ranges: $R_{ij} \leq R_1$, corresponding to the attenuation of the direct wave (Pg); $R_1 < R_{ij} \leq R_2$, corresponding to a transition zone as the direct P waves (Pg) is joined by the post-critical reflections from mid-crustal interfaces and Moho discontinuity; $R_2 < R_{ij}$, corresponding to the attenuation of the refracted P waves (Pn) from Moho discontinuity. Also, we considered c_1 for crust related to anelastic properties of the crust layer and c_2 for anelastic properties of the upper mantle of the case-study region, and $d_j(f)$ is the station correction term for station j that could be related to site response or the lateral variations of attenuation in site (local conditions), or both.

Any model where a source term is multiplied by an attenuation term (or added, considering the logarithm) has non-unique solution due to one undetermined degree of freedom. If all station spectra are multiplied by one arbitrary function of frequency, and all event spectra divided by the same function, the fit would not change (Andrews 1986). This undetermined degree of freedom can be fixed by some constraints. Therefore, we set constraints to determine coefficients of the attenuation model. Site response for each of the stations is obtained by averaging the residuals of station, applying the constraint that the average site term be equal zero. This is because nearly all seismic stations are on rock in contrary to strong motion stations that could be located on rock or soil. Also it was assumed a reference distance as source-attenuation trade-off, where the attenuation is one at $R = R_{ref}$ (and the obtained sources are therefore scaled to the reference distance) in all frequencies. We have set $R_{ref} = 1$ km based on minimum distance sampled by the data set. So in attenuation model eq. (4), A_0 is the spectral source amplitude at $R = R_{ref}$. These assumptions could be both implicit in the model and imposed to the solution.

In this study, the regression method (Atkinson & Mereu 1992) has been used for estimating the source levels (the spectral source amplitude at $R = R_{ref}$), station corrections and the attenuation parameters. In this method, in order to calculate the accurate attenuation parameters, and also lessen the influence of the outlying points on the solution an absolute value of the errors (L1 norm) was considered as the residual function. The coefficients of eq. (4) should be estimated in such a way that the residual function is being minimized.

In regression analysis that is based on Atkinson & Mereu (1992), the events were categorized into 0.5 units of magnitude, and optimization analysis was done in each category. Also an overall source level for each magnitude set is being estimated. We could also determine the station correction term for each station j . For determining the P -wave attenuation curve parameters, the hinged trilinear attenuation model (eq. 4) is being considered and various combinations of the model parameters are being used to estimate the source term of earthquake (at $R = R_{ref}$) sets iteratively.

The initial values for $R_1 = 80$ km and $R_2 = 150$ km have been selected based on the studies of Motaghi & Ghods (2012). Also all $d_j(f)$ values, (site terms), have been assumed equal to zero in the first iteration. We have used starting point combinations with the ranges of b_1 from 1.0 to 1.5, b_2 from -1 to 1 and b_3 from 0.5 to 2. The anelastic attenuation coefficients c_1 and c_2 were unconstrained, and also c_1 was assumed to be constant in all layers of the crust and c_2 is the same for upper mantle. The solution is the combination of b_1 , b_2 , b_3 , c_1 and c_2 which minimize the value of the following residual function

$$e_{ij} = ([\log A_{i0}(f)]_j - \overline{\log A_{i0}(f)})^2. \quad (5)$$

Figure 5. Average residual and its standard deviation by station, at 2.84 Hz. Red and blue dashed lines represent residuals average and 90 per cent significant level, respectively.

Table 1. The estimated coefficients of *P*-wave attenuation model by the method.

f (Hz)	b_1	b_2	b_3	c_1	c_2	σ
1.13	1.17	-0.80	1.31	0.0040	0.00053	0.47
1.42	1.17	-0.80	1.31	0.00399	0.00056	0.52
1.79	1.17	-0.80	1.32	0.00410	0.00055	0.48
2.25	1.18	-0.81	1.32	0.00421	0.00054	0.51
2.84	1.19	-0.81	1.32	0.00412	0.00054	0.49
3.57	1.20	-0.81	1.33	0.00448	0.00051	0.47
4.5	1.21	-0.82	1.33	0.00515	0.00058	0.51
5.66	1.22	-0.82	1.33	0.00535	0.00054	0.52
7.13	1.23	-0.82	1.34	0.00588	0.00055	0.65
8.97	1.24	-0.83	1.34	0.006	0.00052	0.68

$\overline{\log A_{i0}(f)}$, is the mean source amplitude for each category of the earthquake magnitude (at $R = R_{\text{ref}}$), obtained from averaging the $[\log A_{io}(f)]_j$ values over all stations recording the earthquake at $R = R_{\text{ref}}$.

We plotted the residuals average of each station and average of all residuals in Fig. 5 that is not much different from zero. To study the behaviour of the residuals better, the null hypothesis in which the mean of the residuals is zero was being tested using Chi-square test (Corder & Foreman 2014). At 90 per cent significant level, the null hypothesis is not rejected.

Then, the above process is repeated by using these site terms and this constraint. Finally, we optimized eq. (4) by grid search, and imposed the constraints on R_1 [80–120 km], R_2 in [150–190 km] and the interval as 5 km.

These regressions provide estimates of the station correction term for each frequency at each station, the source term for each category of the earthquake magnitude, and the coefficients b and c . The results of the optimization and the regression coefficients were obtained for the data in different frequencies as Table 1. The result of the optimization shows that the attenuation curve have hinges in $R_1 = 90$ km and $R_2 = 175$ km.

Because of the large number and the wide magnitude range of the events in our data base, and in order to compare the attenuation model fitted to the data visually, it is needed to normalize all of the amplitudes to the source level and station correction term. By subtracting A_{i0} and $d_j(f)$ from each of the observed amplitudes, it would be possible to estimate the amplitudes normalizing to a common level and the actual attenuation form (leaving just path effects in the data). So the normalized spectral amplitudes, (A_{ij}^n) , for earthquake i at station j is thus defined as

$$\log A_{ij}^n(f) = \log A_{ij}(f) - \log A_{i0}(f) - d_j(f). \quad (6)$$

The normalized spectral amplitude and their fitted curves against the distance in frequency 1.42 and 3.57 Hz are plotted in Fig. 6.

4 RESULT AND DISCUSSION

According to the researches (e.g. Ou & Herrmann 1990; Atkinson & Mereu 1992), not only is the shape of amplitude decay for the simple layered model complex, but assuming a simple shape for attenuation curve with constant geometrical spreading b for both near source and regional distance is also questionable (Meghdadi & Shoja-Taheri 2014). Atkinson (2004) showed that the attenuation curve of Fourier spectral amplitudes for the shear wave window of the earthquakes in eastern Canada and northeastern United States has three distinct sections. Shoja-Taheri *et al.* (2008) have shown that the attenuation curve for determining the local magnitude M_L in eastern Iran has distinct sections of hinged trilinear form at 106 and 347 km. Motaghi & Ghods (2012) have calculated the decay of shear wave amplitude by regressing the hinges at distances of 80 and 160 km for north of Iran (the central Alborz region).

Figure 6. The normalized spectral amplitude against the distance (grey circle), and the fitted curve of *P*-wave attenuation (red line) in frequency 1.42 Hz (left) and 3.57 Hz (right). Vertical dashed lines shows the hinge points of the model.

Figure 7. Comparing *P*-wave attenuation curve in this study and the global model (Veith & Clawson 1972).

Similar to the attenuation of shear waves, Fig. 6 shows that the attenuation of *P* wave is not constant with the distance and it is possible to use two-hinge model to describe the attenuation of *P* waves, as describe by eq. (4). Fig. 6 depicts the *P*-wave attenuation model of Iran with two hinge points at a distance range of 70–100 and 150–190 km. At the distances less than ≈ 90 km, the amplitude decays nearly constant. The amplitudes show no significant changes between ≈ 90 and ≈ 175 km. The amplitudes decline at a constant rate beyond ≈ 175 km. Fig. 7 also provides a comparison between *P* attenuation model of Iran and the global *P* attenuation model of Veith & Clawson (1972). The attenuation model of Iran is higher in short hypocentral distances less than 200 km in comparison with the global model.

Some studies (e.g. Kaviani *et al.* 2015; Amiri Fard *et al.* 2019), have pointed out that the high seismic attenuation found in some regions of Iran can be ascribed to, e.g. high degree of heterogeneity, high thermal anomalies, strong lateral variation in the crustal thickness

Table 2. The final anelastic attenuation coefficients of *P*-wave attenuation model.

f (Hz)	c_1	c_2
1.13	0.0040	0.00054
1.42	0.00409	0.00055
1.79	0.00415	0.00055
2.25	0.00411	0.00055
2.84	0.00422	0.00054
3.57	0.00468	0.00051
4.5	0.00525	0.00056
5.66	0.0053	0.00056
7.13	0.00568	0.00055
8.97	0.006	0.00052

Figure 8. The estimated *P*-wave attenuation model in 1.42, 2.25, 3.57, 5.66 and 8.97 frequencies.

and/or high pore-pressure. Rahimi *et al.* (2010) noted low values of Q in Northwest of Iran are because of geothermal activities of volcanic environments. Therefore, our observation of high attenuation can be explained by high intrinsic attenuation due to the presence of fluids, partial melt, or high temperatures in the crustal rocks. Both the intrinsic attenuation due to the concentration of fluids in the porous, young sedimentary deposits and scattering attenuation by the sedimentary basin or along faults can play significant roles in reducing the energy of the *P* waves propagating across this region.

However, the attenuation rate of the *P* waves in Iran is similar to the global model for the distances more than 400 km. The global model of Veith & Clawson (1972) has not enough resolution to show the constant attention between 70 and 180 km (the second term in eq. 4).

Records with distances less than about 90 km have been used to calculate *Pg* amplitude–distance curve and those with distances greater than about 175 km have been utilized to determine *Pn* amplitude–distance curve. There are approximately strong post-critical reflections from the Conrad and Moho discontinuities above 90–175 km such as *Pb* phase (an upgoing *P* wave from a source in the lower crust or a *P* wave bottoming in the lower crust) and *PmP* phase (*P* reflection from the outer side of the Moho). All coefficients of the attenuation model 4 and the variability (σ) in different frequencies are shown in Table 1.

The results of regression show that there are no significant changes in the values of b_1 , b_2 and b_3 in different frequencies. So we could consider b_1 , b_2 and b_3 being independent of frequency. This result is similar to Atkinson & Mereu (1992) results in eastern Canada and northeastern United States. With these results, the regressions were repeated, by imposing the constraints that b_1 , b_2 and b_3 are independent of frequency. The obtained coefficients are: $b_1 = 1.2 \pm 0.04$, $b_2 = -0.8 \pm 0.03$ and $b_3 = 1.3 \pm 0.04$, $R_1 = 90 \pm 5$ km and $R_2 = 175 \pm 5$ km. c_1 and c_2 are frequency dependent as presented in Table 2. Fig. 8 shows the calculated *P*-wave attenuation models in different frequencies.

The estimated values of b_1 (the first coefficient of geometrical spreading) are a little higher than the theoretical values of the body waves geometrical spreading 1. The partitioning of the compressional and kinetic energies in the layered media makes the attenuation rate higher than 1. So, the spectral amplitudes of *Pg* waves decay as $R^{-1.2}$ (where R is a hypocentral distance) within 90 km of the source. The identical values were somehow found in the studies of Atkinson (2004), Motazedian (2006) and Motaghi & Ghods (2012).

The negative b_2 for the second segment suggests that the strong post-critical refractions from the Conrad and Moho discontinuities, that appear at a distance greater than 90 km, flattens the attenuation curve, and leads to a little apparent attenuation between 90 and 175 km.

A similar negative behaviour of b_2 was also discovered in the past studies (e.g. Atkinson 2004; Motazedian 2006; Motaghi & Ghods 2012; Meghdadi & Shoja-Taheri 2014).

The average b_3 , the average geometrical spreading coefficients for distances beyond 175 km, is more than 1. The numerical studies of Pn (Serenio Jr *et al.* 1988) indicated that the geometrical spreading rate lies between R^{-1} and R^{-2} , for typical upper-mantle structures, which is in agreement with $b_3(f)$ estimated in this study.

The quality factor of P waves (Qp) could be calculated by the anelastic coefficient of P -wave attenuation model (e.g. Atkinson & Mereu 1992; Motaghi & Ghods 2012; Meghdadi & Shoja-Taheri 2014). The $Qp(f)$ as a function of the anelastic term of attenuation is

$$Qp = \frac{\pi f}{Ln(10)c(f)\alpha} \quad (7)$$

where α is the average P -wave velocity and $c(f)$ is the anelastic attenuation. The parameters $c_1(f)$ and $c_2(f)$ (the anelastic coefficients) are related to the quality factors of Pg and Pn waves. The Q values were approximated for the frequencies above 1.0 Hz with the form $Q = Q_0 f^k$ (Aki 1980), and by a straight line, where Q_0 is the value of Q at 1 Hz and k is a numerical constant, named the frequency parameter (Singh & Herrmann 1983). The least-squares method is employed to estimate the power function of Qpg and Qpn as:

$$Qpg = (54.2 \pm 2.6) f^{(1.0096 \pm 0.07)} \quad (8)$$

$$Qpn = (306.8 \pm 7.4) f^{(0.51 \pm 0.05)} \quad (9)$$

The computed Q_0 value (the Q value at 1 Hz) is 54.2 for Pg waves, and the frequency parameter is 1.0096. The Q_0 value less than 100 and the frequency parameter about 1 are the evidences of high average anelastic attenuation of Pg waves in the Iranian Plateau (Mahood 2014). These values are in agreement with the values reported for active seismotectonic regions (e.g. Mahood & Hamzehloo 2009; Mahood 2014; Farrokhi & Hamzehloo 2017; Amiri Fard *et al.* 2019). The computed Q_0 value for Pn waves is also 306.8 and the frequency parameter is 0.51.

4.1 Validation of the attenuation model

Fig. 9 shows the mean residual between the observed and predicted spectral amplitudes of the events versus the distance in distinct frequencies. The red lines show 95 per cent of the confidence interval of each frequency assuming that the residuals follow the normal distribution with zero mean. As shown, the variation of the residuals increases with frequency and this can partially confirm that the model is more robust in lower frequencies. In addition, we performed the hypothesis test by the null hypothesis of the zero mean of residual (Corder & Foreman 2014); the null hypothesis was not rejected. This analysis confirms that the estimated parameters are reasonable and stable for the study region and it could be concluded that the attenuation model of eq. (4) provides a satisfactory description of the data.

The mean residuals between the observed and predicted spectral amplitudes of the events versus the magnitude and depth in frequency of 2.84 Hz, as well as the linear fit obtained using the maximum likelihood method and mean line are plotted in Fig. 10. The mean of the residuals are -0.018 and 0.0169 for the magnitude and depth, respectively, which are not significantly different from zero. The slope of trend line for magnitude is 0.0254. This means that the model overpredicts the average spectral amplitude values at high magnitudes and down estimates for low magnitudes but all are negligible. Although shallow earthquakes attenuate more rapidly than deep events (the error for the deeper earthquake is larger and negative), the slope of the maximum likelihood straight line for depth is so low (0.0042) which could be considered as zero. In addition, we performed the hypothesis test by the null hypothesis of the zero mean of the residual. At 90 per cent significant level, the null hypothesis is not rejected.

Station corrections term correlates with site conditions. All the stations used in this study are sited on hard rock (bedrock) (Naserieh *et al.* 2019). Variability in site response and local changing attenuation among the rock sites and the lateral variations of attenuation in each site (local conditions) are evaluated based on the analysis of regression residuals for vertical component on a station-by-station basis. We considered the residual function as the log of the observed Fourier amplitude minus the log of the Fourier amplitude predicted by the model. So, to determine the relative site response and the local changing attenuation for each vertical component of the station (difference value with zero), the residuals were averaged on each station. Large average residual for a station is expressing a significant site response. The majority of the stations have the residual values very close to zero. The positive residual in some stations (e.g. GLO and TAB stations) is related to the amplifying effect of geology beneath the station or smaller attenuation along the path. The negative residual in other stations (like MRD and DMV station) is dependent on lower seismic velocity, and larger attenuation of local condition. The residual values of four stations (in frequency 2.84 Hz) are plotted in Fig. 11.

Table 3 and Fig. 12 (left) show comparison of the estimated Qpg values with reported quality factor of Pg wave in different tectonic regions of the Iranian Plateau. The estimated quality factor values for Pg waves in this study are in agreement with the results of the studies in different regions of Iran.

It is worth mentioning that the results of our study have been compared with the other ones in different regions of the world. The estimated value of Qpg is being compared with the results of the previous studies in SAR and tectonically SSR in Table 3 and Fig. 12 (middle). Various observations in different regions of the world have indicated a notable difference in the rate of the amplitude decay between the

Figure 9. Estimation of the residuals means (residuals defined as $\log(\text{obs.}) - \log(\text{per.})$) of the amplitude spectra for frequencies at several distance ranges centred at the black circles; 95 per cent confidence level for each frequency by the assumption that the residuals follow the normal distribution with zero mean are also shown by the red lines.

Figure 10. Estimation of the residuals means (residuals defined as $\log(\text{per.}) - \log(\text{obs.})$) of the amplitude spectra in frequency of 2.84 Hz centred at the black circles. Left: residual versus local magnitude, and right: residual versus depth (km). Solid black, blue dashed and red dashed lines represent the bias, the maximum likelihood best fits and 90 per cent significant level of hypothesis test, respectively.

Figure 11. The distribution of residuals for MRD, TAB, DMV and GLO stations at 2.84 Hz.

Table 3. The comparison of estimated Q_p with some of the estimated Q_{pg} and Q_{pn} in different tectonic regions of Iran and the World.

Region	Reference/local region	Q_p	Frequency range (Hz)	Maximum distance (km)
Iran	Nowroozi (2006)/Northeastern Iran	$Q_{pg} = 62f^{0.87}$	1–24	200
	Mahood <i>et al.</i> (2009)/East Central Iran	$Q_{pg} = 40f^{0.99}$	1.5–24	100
	ShojaTaheri & Farrokhi (2009)/Northeastern Iran	$Q_{pg} = 43f^{0.79}$	1.5–36	120
	Mahood (2014)/Eastern Iran	$Q_{pg} = 50f^{1.01}$	1.5–24	120
	Farokhi & Hamzehloo (2017)/Alborz region	$Q_{pg} = 62f^{1.03}$	0.3–24	130
	Farokhi & Hamzehloo (2017)/North Central Iran	$Q_{pg} = 48f^{0.95}$	0.3–24	130
	Amiri Fard <i>et al.</i> (2019)/Western Iran	$Q_{pg} = 56f^{1.12}$	1–24	170
	Vidals Basurto <i>et al.</i> (2014)/Mexico	$Q_{pg} = 34f^{0.82}$	0.6–12.6	120
Different regions of the World	Sereno <i>et al.</i> (1988)/Scandinavia	$Q_{pn} = 325f^{0.48}$	1–15	1400
	Kim <i>et al.</i> (2004)/Central South Korea	$Q_{pg} = 330f^{0.54}$	1–24	160
	Sharma <i>et al.</i> (2008)/India	$Q_{pg} = 59f^{1.04}$	1.5–18	55
	Console & Rovelli (1981)/Italy	$Q_{pg} = 80f^{1.1}$	0.1–10	200
	This study/Iranian Plateau	$Q_{pg} = 54f^{1.01}$	1–10	175
Iranian Plateau		$Q_{pn} = 306f^{0.51}$	1–10	1000

Figure 12. Left: comparison of the estimated Q_{pg} in this study and the reported quality factor in different tectonic regions of Iran; middle: comparison of Q_{pg} in different SAR and SSR of the World and right: comparison of the estimated Q_{pn} of Iran and Scandinavia.

tectonically active and the stable regions. Fig. 12 (middle) shows that the attenuation values in this study are in a good agreement with the tectonically active regions of the world.

For the first time, in this study, the quality factor of Pn wave was determined for Iranian Plateau in a distance range up to 1000 km. Fig. 12 (right) shows the comparison between the quality factor of Pn wave of Iran and Scandinavia.

5 CONCLUSIONS

The vertical-component seismic waveforms of 9267 earthquakes in the magnitude range of M_n 1.5–5.5, reported by the IRSC were processed and analysed in order to obtain a P -wave amplitude attenuation model in local and regional distances. The results of this study indicate that the attenuation curve for the amplitudes of P wave in the frequency domain has three distinct sections in the Iranian Plateau. It is found that the rate of the amplitude decline for distances less than 90 km is higher than those above 175 km. The shape of the attenuation curves was determined by trilinear attenuation functions with hinges at distances 90 and 175 km. Fourier spectral amplitudes are found to decay as $R^{-1.2}$ (where R is hypocentral distance) corresponding to the geometric spreading, within 90 km from the source. There is a section from 90 to 175 km where the attenuation has been described as $R^{0.8}$, and above 175 km, the attenuation is described well by $R^{-1.3}$, corresponding to the geometric spreading in three sections. We have found that all geometrical spreading coefficients are independent of frequency. The anelastic attenuation (including scattering attenuation) of P waves are related to the quality factor being estimated as $Q_{pg} = (54.2 \pm 2.6)f^{(1.0096 \pm 0.07)}$ and $Q_{pn} = (306.8 \pm 7.4)f^{(0.510 \pm 0.05)}$. The estimated Q values for P waves are similar to the tectonically active regions of the Iranian Plateau. The agreement between the residual mean of the observed and predicted spectral amplitudes confirms that the estimated parameters of the attenuation curve model is reasonable for the study region. The result implies that the Iranian Plateau has higher attenuation values and higher heterogeneities within the earth's crust in comparison with the global models in local distances. The observation of high attenuation can be explained by high intrinsic attenuation due to the presence of fluids, partial melt, or high temperatures in the crustal rocks. Interpretation of the results has helped to parametrize the observed P -wave amplitude attenuation which is useful for obtaining the accurate source parameters and characterizing the earth properties.

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