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Short Note

M_L and M_W Scales in the Iranian Plateau Based on the Strong-Motion Records

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Abstract The availability of a large amount of strong-motion data recorded by the National Strong Motion Network of Iran (NSMNI) has motivated this study to develop relations for routine determination of M_L and M_W from digital horizontal components of the strong-motion records. The dataset comprises 861 two-component horizontal acceleration time series recorded for 125 earthquakes with magnitudes of 4.5 and larger. The M_L scale is based on the horizontal synthesized Wood–Anderson seismograms. We have applied the Monte Carlo technique to evaluate distance correction curves for use in determining the local magnitude, M_L, in Iran and in its northern, eastern, and Zagros subregions. Results indicate that the distance correction curves show trilinear behavior for geometrical spreading. The resulting coefficients evaluated for Iran as a whole are: $R_1 = 96 \pm 5$ km; $R_2 = 131 \pm 5$ km; $n_1 = 1.01 \pm 0.02$; $n_2 = -0.14 \pm 0.1$; $n_3 = 0.14 \pm 0.03$; $k = 0.00020 \pm 0.00008$. The distances less than R_1 correspond to attenuation of the direct waves. Between R_1 and R_2 is the distance where the multiply reflected and refracted shear waves from Moho dominate the arrivals. n_1, n_2, and n_3 are the coefficients of geometrical spreading for distances from the source to R_1, R_1 to R_2, and beyond R_2. k is the coefficient of inelastic attenuation. For estimating the M_W scale from the strong-motion data, we used the method proposed by Andrews (1986). To find the best correlation between the moment magnitudes measured from the strong-motion data and those measured from teleseismic data, we examined several time windows (e.g., whole trace, *S*-wave coda, and source time durations). The regressions show that the M_W estimates from different time windows are all equally well correlated with the corresponding reported values with nearly identical standard deviations. Finally, relations between the estimates of local and moment magnitudes for the regions show that for earthquakes with magnitudes larger than about 6.0, the M_L scale gradually becomes saturated and, therefore, it gives smaller values than those obtained by the M_W scale. However, for smaller earthquakes, the M_L scale overestimates the M_W scale. This discrepancy occurs mainly because the frequency contents of the waveforms employed in these scales are different.

Introduction

The area under study extends from 25 to 40 degrees north latitude and from 44 to 62 degrees east longitude (Fig. 1). It is a region of a conspicuous series of folded ranges formed during the Alpine earth movements of the Tertiary Period. Iran, as a region with a high level of seismic activity, has experienced numerous catastrophic earthquakes throughout its long history (Ambraseys and Melville, 1982; Berberian, 1983). The seismotectonic characteristics in Iran are complex and show distinct variations in different regions of the country. Seismicity and seismotectonics of Iran have been discussed by several authors (e.g., Stöcklin, 1968; Nowroozi, 1971; Takin, 1972; Nowroozi, 1976; Shoja-Taheri and Niazi, 1981; Ambraseys and Melville, 1982; Berberian 1983). Stöcklin (1968) recognized nine structural zones with different tectonic styles in Iran. A more simplified division was proposed by Takin (1972), who divided the entire country into four divisions. Based on relocated

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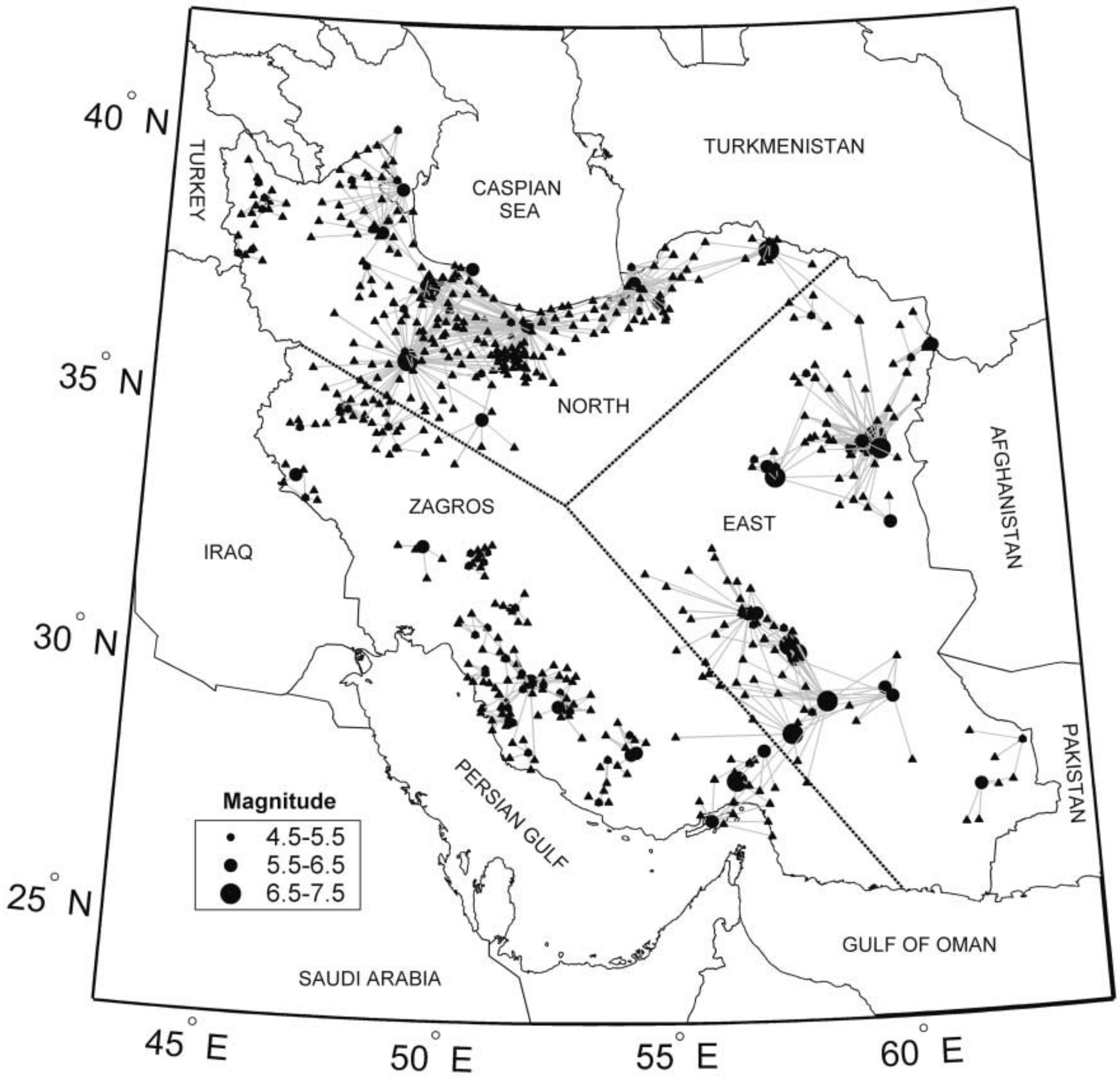

Figure 1. Earthquake epicenters and recording stations.

earthquakes, Nowroozi (1976) constructed a seismotectonic map of Iran including 23 seismotectonic provinces. Discussing the seismicity of Iran, Shoja-Taheri and Niazi (1981) divided the Iranian plateau into three distinct tectonic subregions of Albourz, east-central, and Zagros.

The availability of a large amount of the strong-motion data that have been recorded by the National Strong Motion Network of Iran (NSMNI) since its inception in 1973 has provided opportunities for conducting a broad range of studies about earthquakes in Iran. The main objective of the present study is to provide relationships for routine determination of M_L and M_W magnitudes of the events that are recorded by NSMNI. Such study of earthquakes in Iran, is needed because the attenuation relations introduced for other seismic regions (e.g., Espinosa, 1980; Bakun and Joyner, 1984; Hutton and Boore, 1987; Alsaker *et al.*, 1991; Baumbach *et al.*, 2003), in general, do not result in unbiased estimates of magnitudes when applied to the strong-motion data of Iran. The data used in this study were registered within 250 km from their sources. It would be more desirable, of course, to extend this range of distance by adding the Wood–Anderson synthetic seismograms from higher-gain networks. But inadequacy and nonuniform distribution of such supplementary data in Iran and its different seismic regions has confined our study to the use of the strong-motion data.

Local magnitude, M_L, was the first magnitude scale to describe the size of an earthquake. It was first developed by Richter (1935) for earthquakes in southern California. The local magnitude is based on the amplitude recorded by the Wood–Anderson torsion seismograph with a natural period of 0.8 sec, a damping constant of $h = 0.8$, and a static magnification of $V = 2800$. M_L is now widely used in tectonic environments that can be different from southern California. Therefore, it is necessary to calibrate M_L so that it gives magnitudes that are consistent with other magnitude scales such as body-wave magnitude, m_b, or moment magnitude, M_W. The local magnitude, M_L, has the most direct relevance to engineering application because M_L is determined within the period range of greatest engineering interest (typically, 0.2–3 sec). This feature, plus the fact that the source distances for M_L scale are shorter, in general, than those used for other magnitude scales, means that M_L is of particular interest for most engineering applications, including the determination of seismic-design criteria for major projects.

On the other hand, the moment magnitude, M_W, which is estimated from seismic moment M_0 (Kanamori, 1977), is not directly related to the strength of shaking in the frequency range of most interest in engineering, although it represents other important characteristics of the earthquake such as the fault dimension and the duration of the strong ground motion.

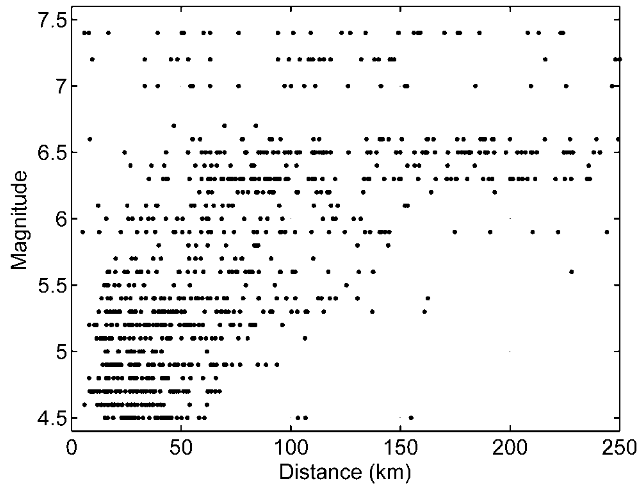

Figure 2. Distribution of the magnitudes of the selected earthquakes versus distance.

Strong-Motion Data

The data used in this study comprise 861 two-component horizontal acceleration time series recorded for 125 earthquakes with magnitudes of 4.5 and larger. We have selected only those earthquakes that are registered by at least three stations within 250 km of epicentral distances. In this study the Iranian plateau is divided into three tectonic subregions of northern Iran, eastern Iran, and Zagros. We employed the strong-motion data to develop relations for M_L and M_W determination for the whole country and also for each of the previously mentioned subregions. The epicenters and their recording stations are plotted in Figure 1. Distribution of the magnitudes of the selected earthquakes versus distance is shown in Figure 2. One unavoidable characteristic of such distribution of the strong-motion data is that small earthquakes are more often recorded at shorter distances, whereas larger earthquakes generally have more distant recording. Numbers of the selected events and their corresponding records for each of the subregions and also for the whole region of Iran are listed in Table 1.

Table 1
Number of Events and Records in Different Magnitude Range

	Magnitude					
	4.5–5.4		5.5–6.4		6.5–7.4	
	Events	Records	Events	Records	Events	Records
Iran	84	411	28	264	13	186
East	12	47	11	70	9	102
North	26	159	8	136	3	80
Zagros	46	205	9	58	1	4

We processed the uncorrected strong-motion data to make baseline and instrument corrections, and high-pass filtering. The filtering was performed in the frequency domain with the response that of a zero-phase Butterworth filter of four poles with different corner frequencies depending on the magnitude of each earthquake. The uncorrected data were originally digitized at the rate of 200 samples/sec, but we processed the data at the rate of 100 samples/sec.

The M_L Scale in the Iranian Plateau

According to Richter (1935, 1958), M_L is given by

$$M_L = \log A - \log A_0, \tag{1}$$

where A is the maximum trace amplitude in millimeters measured from a Wood–Anderson seismogram and, $-\log A_0$ is the empirically derived attenuation curve. The attenuation curve depends on the amplitude decay of the signal due to anelastic attenuation, scattering, and geometrical spreading along the event-station path. The attenuation curve strongly depends on the regional earth structure, and hence, it is applicable only where the curve is obtained. By definition, the constraint on the previous equation is that the local magnitude of a shock recorded by a hypothetical Wood–Anderson instrument at 100 km from the epicenter record with amplitude of 1 mm is 3.0. With this constraint. equation (1) can be given as

$$M_L = \log A + n \log(R/100) + k(R - 100) + 3.0, \tag{2}$$

where n and k are coefficients for geometrical spreading and anelastic attenuation, respectively, and R is the hypocentral distance.

In a more general form, equation (2) reads

$$M_L = \log A + n \log(R/R_{ref}) + k(R - R_{ref}) + K_{ref}, \tag{3}$$

where R_{ref} is a reference distance that could be different from 100 km, and K_{ref} is a constant that is given in connection with Richter's M_L definition. Recently, from very large data bases, Hutton and Boore (1987) have developed new attenuation terms for southern California. The connection to Richter's M_L is achieved by anchoring $-\log A_0$ values to those of Hutton and Boore (1987), resulting in the following relation

$$K_{ref} = 1.110 \log(R_{ref}/100) + 0.00189(R_{ref} - 100) + 3.0 \quad (4)$$

which gives a value of 2.8 for 60 km and 3.0 for a distance of 100 km.

To prepare the strong-motion data for the M_L study, we convert each horizontal component into synthesized Wood–Anderson seismogram (Nigam and Jennings, 1969), and read the maximum zero-to-peak amplitudes of the wave trains; select a reference distance for connecting to Richter's M_L definition.

Performing a linear regression analysis on the synthesized data to develop a relation for Iran similar to that of equation (3), we obtained values of $n = 1.26$ and $k = -0.0033$. Negative value of k, which is not physically acceptable, indicates that the validity of assuming a simple shape for the attenuation curve, with a constant geometrical spreading, n, for both near-source and regional distances is questionable. The wave-propagation studies indicate that the expected shape of amplitude decay for simple layered crustal models is complex (e.g., Burger *et al.*, 1987; Ou and Herrmann, 1990; Atkinson and Mereu, 1992). Layering in the crust causes direct-wave amplitudes to decay more steeply than R^{-1}. Then, as the direct arrivals are joined by postcritical reflection off the Moho and intracrustal discontinuities, there may be distance ranges where amplitudes actually increase with distance, between approximately 50 and 200 km. Atkinson and Mereu (1992) have shown that the attenuation curve in southeastern Canada has three distinct sections. Their results indicate that at distances less than 70 km, corresponding to attenuation of the direct wave, amplitudes decay slightly more rapidly than $1/R$. Between 70 and 130 km, the amplitudes are approximately constant. Beyond 130 km, amplitude decays at a rate that is consistent with $R^{-0.5}$.

Based on theoretical studies of wave propagation in layered crustal models in eastern North America (Burger *et al.*, 1987; Ou and Herrmann, 1990), we regress the data for the previously mentioned subregions and for the whole plateau of Iran, allowing n to take separate values in three different distance ranges For the hinged trilinear form of attenuation, the general form of M_L becomes

(i) $R \le R_1$,

$$(M_L)_{ij} = \log A_{ij} + n_1 \log R_{ij} + kR_{ij} + C \quad (5a)$$

(ii) $R_1 < R \le R_2$,

$$(M_L)_{ij} = \log A_{ij} + n_1 \log R_1 + n_2 \log(R_{ij}/R_1) + kR_{ij} + C \quad (5b)$$

(iii) $R > R_2$,

$$(M_L)_{ij} = \log A_{ij} + n_1 \log R_1 + n_2 \log(R_2/R_1) + n_3 \log(R_{ij}/R_2) + kR_{ij} + C, \quad (5c)$$

where $C = K_{ref} - k\, R_{ref} - n_1 \log R_{ref}$. The indices i and j stand for ith earthquake and jth record.

To estimate the coefficients k, n_1, n_2, n_3, R_1, and R_2 we used the Monte Carlo technique by generating random numbers confined within the predefined ranges for each of these parameters, testing all possible combinations of the attenuation parameters, and searching for the combination that minimizes the average residual errors. The predefined windows for the parameters and their corresponding values obtained for all regions are given in Table 2.

The attenuation curves are plotted in Figure 3. The Richter's attenuation term, $-\log A_0$, revised by Hutton and Boore (1987), is also shown for comparison.

For each earthquake the residuals between the mean and estimated magnitude from each record are:

$$\mathrm{Res}_{ij} = \overline{M}_i - M_{ij} \quad (6)$$

Table 2
The Parameters Estimated for M_L Attenuation Curves

	R_1	R_2	n_1	n_2	n_3	k
	Predefined Windows of Parameters*					
	70–120 km	100–160 km	0.8–1.3	−0.3–0.3	0.1–0.5	0.001–0.002
	Results					
Iran	96 ± 5	131 ± 5	1.01 ± 0.02	−0.14 ± 0.13	0.14 ± 0.03	0.00020 ± 0.00008
East	88 ± 2	122 ± 4	1.08 ± 0.02	0.15 ± 0.10	0.22 ± 0.08	0.00055 ± 0.00020
North	102 ± 3	138 ± 4	1.03 ± 0.04	−0.24 ± 0.05	0.15 ± 0.04	0.00017 ± 0.00006
Zagros	79 ± 4	115 ± 5	0.88 ± 0.02	−0.25 ± 0.03	0.23 ± 0.10	0.00022 ± 0.00009

*For Zagros region, predefined windows of R_1 and R_2 are, respectively, 50–110 and 80–150 km.

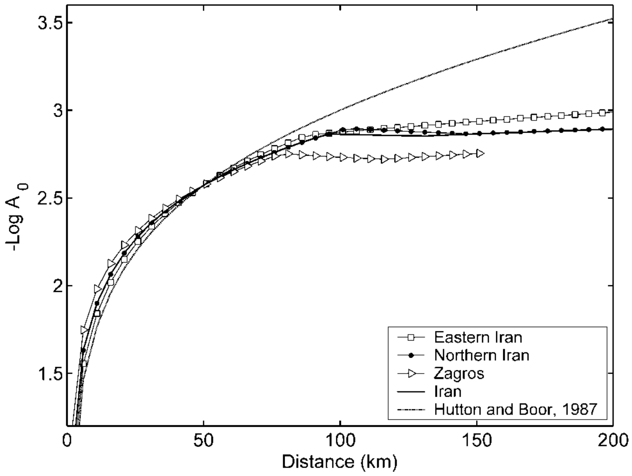

Figure 3. Trilinear distance correction, $-\log A_0$, from the present study, along with the distance correction, $-\log A_0$, of the revised Richter's relation (Hutton and Boore, 1987) for southern California.

The resulting residuals from the attenuation relations developed in this study and also the corresponding residuals from the revised Richter's relation (equation 4) are shown in Figure 4. Clearly, the residuals and their mean values, because of our trilinear distance attenuation relations, have uniform distribution with distance and fit their zero baselines very closely. Using the revised Richter's relation (equation 4) to our dataset, however, results to a noticeable overestimate of M_L values at distances larger than about 80 km.

Calibration of M_W for Earthquakes in Iran

M_W derived from the seismic moment M_0 is a more reliable estimate of the magnitude because it is calculated using longer periods, so the attenuation is not greatly affected by near-source structure. Kanamori (1977) derived a relationship for calculating M_W with respect to M_0. It is given by

$$M_W = 2/3 \log M_0 - 10.7 \tag{7}$$

Moment-tensor analysis for earthquakes larger than M_W 5.0 has been routinely carried out by Harverd University (e.g., Dziewonski *et al.*, 1981) and the U.S. Geological Survey (USGS) (e.g., Sipkin, 1986) since the late 1970s and has produced a large catalog of M_W values for the larger global events. With the use of the available strong-motion dataset in Iran, it is now possible to routinely calculate regional and local seismic moments by extending the magnitude threshold down to M_W 4.5. To compute M_W from the strong-motion data, we employed the method proposed by Andrews (1986). In this method, the integrals of the squared velocity and squared displacement are applied to measure seismic moment, M_0, by means of the following formulas (Keilis-Borok, 1960)

$$M_0 = \frac{4\pi\rho\beta^3}{\Re_{\theta\varphi} FP} G(R)\Omega_0, \tag{8}$$

where ρ is density (2.7 g/cc), β is shear velocity (3.5 km/sec), F is the amplification of free surface(2), $\Re_{\theta\varphi}$ is the mean of radiation pattern (0.55), P is the coefficient of energy partition between two horizontal components (0.7), $G(R)$ is geometrical spreading for distance R and is defined by Herrmann and Kijko (1983) as:

$$G(R) = \begin{cases} R & , R \le 100 \text{ km} \\ (100R) & , R > 100 \text{ km} \end{cases}$$

and $\Omega_0 = 2I_D^{3/4} I_V^{-1/4}$ (Andrews, 1986) is the low-frequency spectrum level according to Brune's ω-squared model (Brune, 1970, 1971).

$$I_V = \int_0^{+\infty} V^2(t)dt, \quad I_D = \int_0^{+\infty} D^2(t)dt. \tag{9}$$

For inelastic attenuation correction we assumed $Q = 153 f^{0.88}$ (Chin and Aki, 1991). To find an appropriate time duration in I_V and I_D such that it gives best correlation between the moment magnitudes measured from the strong-motion data and those from the teleseismic data, we examined five different time spans of S-wave coda, T_s (which is related to the source duration and is magnitude dependent, McGuire and Hanks, 1980; Atkinson and Silva, 2000), 75%, 95%, and also the whole trace durations. The first four time windows begin with the onset of S waves, whereas the whole trace window begins with the onset of P waves.

As seen in Figure 5, the regressions made between the calculated magnitudes, M_{WO}, using all the data for the whole region of Iran, and the reported magnitudes, M_{WH}, give similar results for different time windows with nearly identical standard deviations, σ. Thus, for the sake of practical conveniences in routine determination of moment magnitudes from the strong-motion data, the whole trace window is preferred for the regression. The result of regression for the three subregions of northern Iran, eastern Iran, Zagros, and also for the whole plateau of Iran are given by relations (9).

(i) Iran

$$M_{WO} = (0.92 \pm 0.03)M_{WH} + (0.50 \pm 0.17), \quad \sigma = 0.148 \tag{9a}$$

(ii) East

$$M_{WO} = (0.90 \pm 0.04)M_{WH} + (0.65 \pm 0.24), \quad \sigma = 0.116 \tag{9b}$$

(iii) North

$$M_{WO} = (0.95 \pm 0.05)M_{WH} + (0.30 \pm 0.29), \quad \sigma = 0.129 \tag{9c}$$

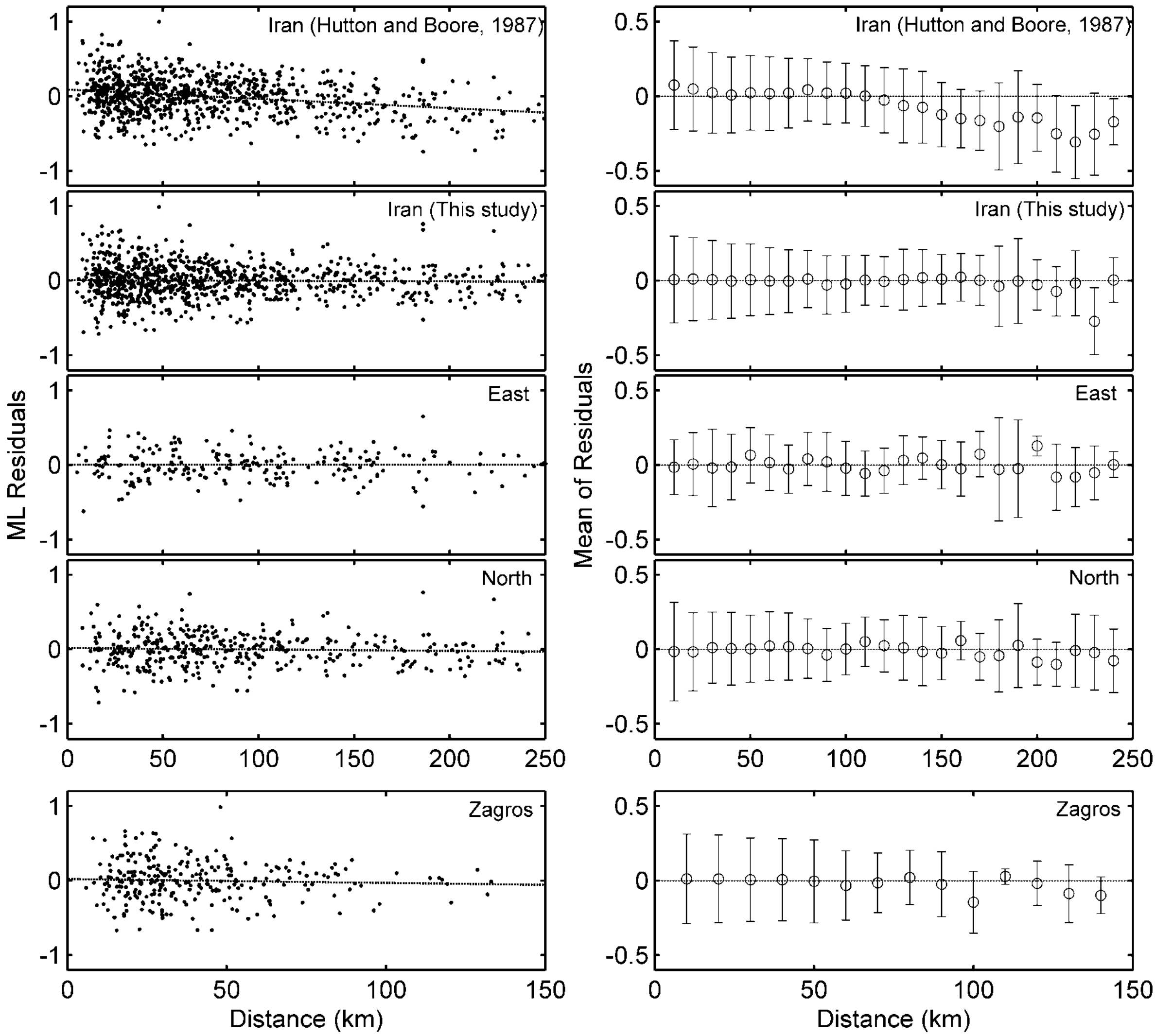

Figure 4. The M_L residuals and their mean values obtained by using the trilinear distance attenuation relations developed in this study are compared with the residuals from using the revised Richter's relation (Hutton and Boore, 1987).

(iv) Zagros

$$M_{WO} = (0.94 \pm 0.12)M_{WH} + (0.40 \pm 0.63), \quad \sigma = 0.184 \quad (9d)$$

M_L versus M_W

We can now show the relationship between the M_L and M_{WO} scales estimated from the strong motion data. Figure 6 depicts the fits between the two scales for the regions. The results show that for earthquakes with magnitude larger than about 6.0, the M_L scale gradually becomes saturated and gives smaller values than those obtained by the M_W scale; for smaller earthquakes, the M_L scale overestimates the M_W values. This discrepancy is mainly because the frequency ranges employed in these two scales are different.

Conclusions

A large amount of strong-motion data, comprising 1722 horizontal component acceleration time series recorded by 125 earthquakes with magnitudes between 4.5 and 7.4, have been used in developing new M_L and M_W scales for routine and reliable determination of magnitudes from the strong-motion data recorded in different seismic regions in Iran. We have selected only those earthquakes that are registered by at least three stations within 250 km of epicentral distances. In this study the Iranian plateau is divided into three

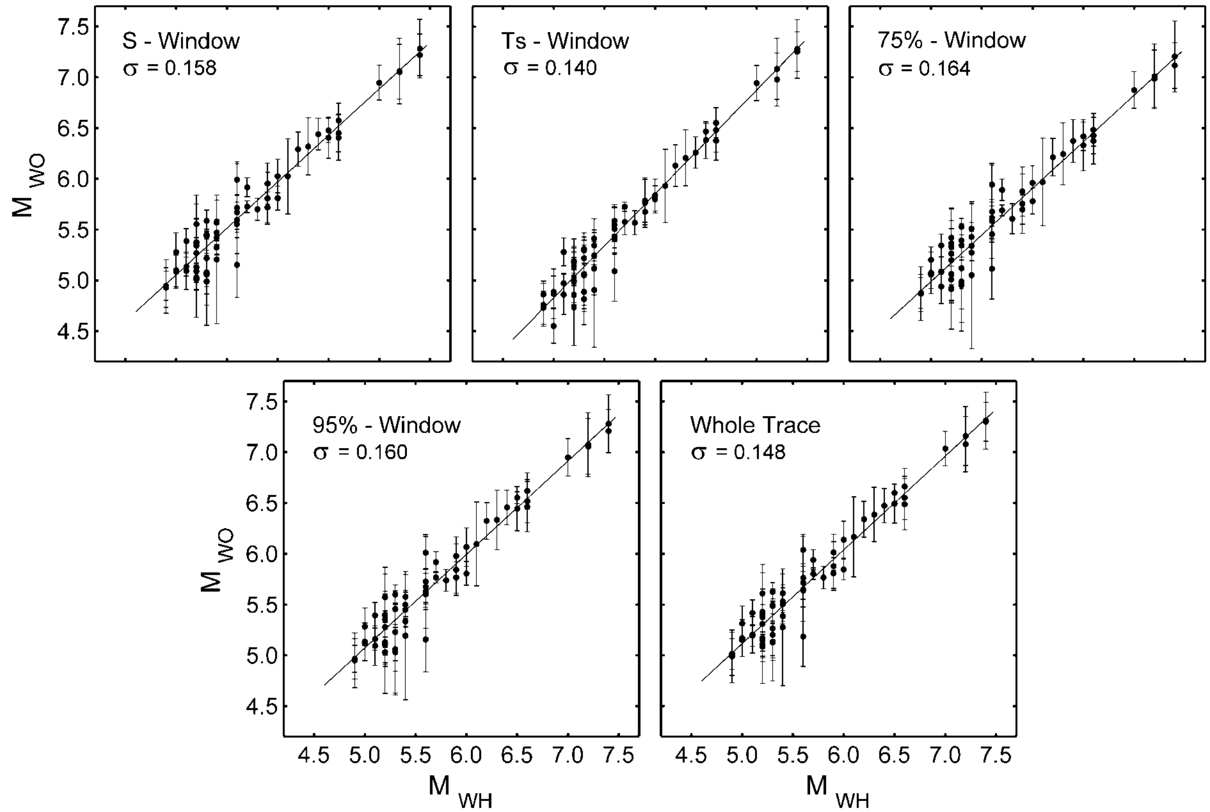

Figure 5. Regressions between the moment magnitudes calculated using the strong-motion data and those reported by Harvard for different time windows.

tectonic subregions of northern Iran, eastern Iran, and Zagros. Performing linear regression analysis on the data to develop relations similar to that of equation (3) results in a negative value of the inelastic attenuation constant k for each of the seismic zones in Iran. Negative value of k indicates that the validity of assuming a simple shape for the attenuation curves, with a constant geometrical spreading, n, for near and regional source distances in Iran, is questionable. Thus, to overcome this discrepancy we regressed the data, allowing n to take separate values in three different distance ranges. For estimating the coefficients of k, n_1, n_2, n_3, R_1, and R_2, we employed the Monte Carlo technique by generating random numbers confined within the predefined ranges for each of the parameters. Testing all possible combinations of the parameters, we searched for the combination that minimizes the average of the residual errors. The uniform distribution of the residuals about their baselines show that the trilinear distance attenuation relations developed in this study provide more reliable estimates of M_L values than those from linear relations. Applying linear distance attenuation relations to our dataset generally results in noticeable overestimation of M_L values at distances larger than about 80 km.

For the M_W scale the Andrews (1986) technique was used. It was shown that the M_W values estimated from the five previously mentioned time windows are all equally well correlated with their corresponding reported M_W values with nearly identical standard deviations, σ. Thus, for the sake of practical convenience in routine determination of moment magnitudes from the strong-motion data, the whole trace window was chosen.

The correlation made between the local and moment magnitude scales shows that for earthquakes with magnitude larger than about 6.0, the M_L scale gradually becomes saturated and it therefore gives smaller values than those obtained by the M_W scale. However, for smaller earthquakes, the M_L scale overestimates the M_W scale. This discrepancy is mainly because the frequency contents of the waveforms used in these two scales are different.

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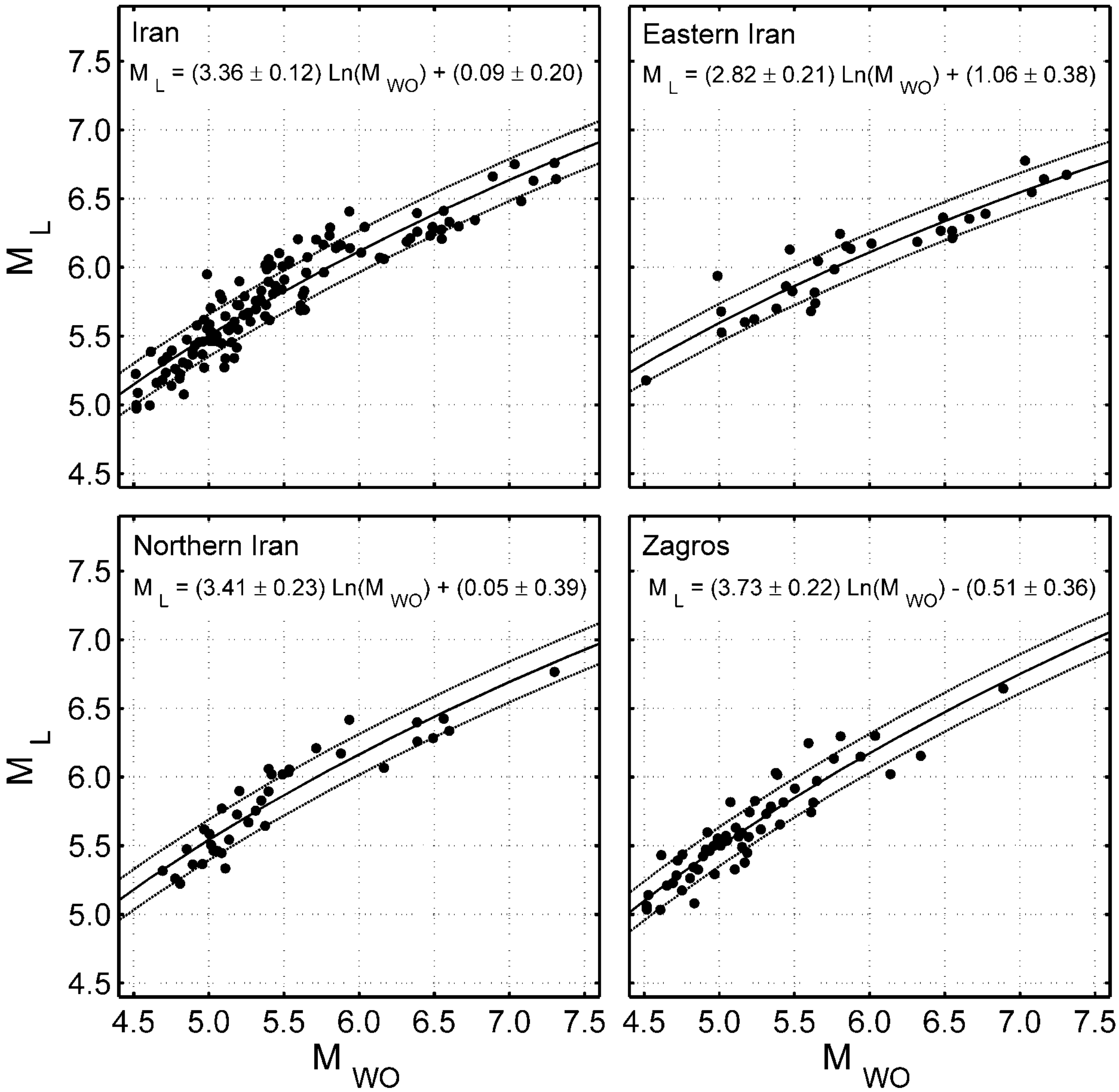

Figure 6. M_L versus M_W scales that were found from recorded strong-motion data.

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