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USING STRONG MOTION ACCELEROGRAMS FOR ESTIMATION OF LOCAL MAGNITUDES OF EARTHQUAKES IN HIMALAYAN REGION

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ABSTRACT

To have uniformity with the available data on past earthquakes in a region, it is necessary that the local magnitudes estimated from strong-motion accelerograms match with the published magnitude values based on the records of conventional seismographs. For this purpose, the magnitude estimates based on strong-motion data need an empirical correction to account for the anelastic attenuation of seismic waves recorded on conventional seismographs at large distances and for the saturation of strong-motion amplitudes during very large magnitude earthquakes. Such correction factors available for other parts of the world would not be, in general, suitable for earthquakes in the Himalayan region. Therefore, using available strong-motion data base from six earthquakes, appropriate correction factors are developed to calibrate the method of computing local magnitudes from strong-motion accelerograms in the Himalayan region.

KEYWORDS: Local Magnitude, Strong Motion Accelerogram, Himalayan Region

INTRODUCTION

The Richter's local magnitude of an earthquake is defined in terms of the maximum amplitude of the record on a standard Wood-Anderson seismograph (WAS) with natural period as 0.8 s, damping ratio as 0.8 of critical damping, and static magnification equal to 2800 (Richter, 1935, 1958). If A is the maximum trace amplitude in millimeters on a WAS at an epicentral distance of R km, the local magnitude M_L is equal to $\log A - \log A_0(R)$ where $\log A_0(R)$ is an attenuation function for seismic waves with distance (Richter, 1935; Gutenberg and Richter, 1942). For distances less than about 25 km, the standard WAS normally goes out of scale even for small magnitudes (≈ 4.5). Therefore, several investigators (e.g., Trifunac and Brune, 1970; Kanamori and Jennings, 1978; Luco, 1982; Uhrhammer and Bolt, 1991; Gupta and Rambabu, 1993; etc.) have proposed to evaluate the local magnitude by synthesizing the response of WAS using strong-motion accelerograms. Due to short distances for recording of strong-motion data, their use for computation of magnitude has the advantage that they are not much influenced by the propagation path effects.

The early studies (Trifunac and Brune, 1970) to compute the local magnitudes from strong-motion data used the Richter's attenuation function $\log A_0(R)$ only. But, Luco (1982) found that $\log A_0(R)$ is not suitable to describe the attenuation of strong-motion amplitudes in the near field, particularly near 20 km. He, therefore, proposed a modification in the Richter's attenuation function for smaller distances. Jennings and Kanamori (1983) also suggested modifications in $\log A_0(R)$ for small R, but their failure to include the local geological site effects in the analysis resulted in biased estimate of this attenuation. Depending upon the source-to-site distance, magnitude and the geological condition at a site, the peak values of the response of WAS may occur at different wave periods. Therefore, using the frequency - dependent attenuation function due to Trifunac and Lee (1990) and the knowledge of the frequency where most energy is concentrated in the strong-motion as filtered by the WAS, Trifunac (1991a) has defined an attenuation function $Att(\Delta_0)$ in place of the Richter's function, where Δ_0 is the hypocentral distance. This new attenuation function represents much faster decay of strong ground motion than the Richter's function for distances up to 35 km.

In the present study, Trifunac's attenuation function $Att(\Delta_0)$ is used to compute the strong-motion magnitudes of six significant earthquakes in the Himalayan region. However, the mean value of the strong-motion magnitude thus obtained from several accelerograms for each of the earthquakes is, in general, found to be in large variance with the corresponding published magnitude (Richter's local magnitude up to around magnitude 6.5 and the surface-wave magnitude for larger earthquakes). Such differences are mainly due to the anelastic attenuation of seismic waves recorded on conventional seismographs at long distances and the saturation effects of strong-motion amplitudes for larger magnitudes. To account for these differences, Trifunac (1991a) has introduced an empirical correction factor, which he has defined for different values of the conventional magnitude, using strong-motion data for the western United States (US). Similar correction factors obtained by Lee et al. (1990) for Yugoslavia are quite different from those for the western US. Such differences may be ascribed to the difference between the anelastic attenuation characteristics, the wide variations in the distance range at which the strong-motion accelerograms are recorded and regional differences in computing the magnitudes (Trifunac and Herak, 1992). To account for the effects of all these factors, in the present study, appropriate correction factors are developed to estimate the strong-motion local magnitudes for earthquakes in the Himalayan region. Strong-motion acceleration data from six earthquakes with a wide range of magnitude has been used for this purpose.

THE DATA BASE USED

The two horizontal components of a total of 99 strong-motion records obtained from six earthquakes in different parts of the Himalayan region have been used to compute the strong-motion local magnitudes in the present study. These accelerograms have been recorded by three strong-motion arrays of about 40 to 50 instruments each, installed in the 'Kangra' region of Himachal Pradesh, 'Shillong' region of northeast India and the 'Uttarkashi' area of Uttar Pradesh. The Kangra array has recorded one event, the Shillong array has recorded four earthquakes and the Uttarkashi array has also recorded one earthquake. The details of these six earthquakes and the number of records obtained from them are listed in Table 1. Each of these earthquakes has been assigned different magnitude values by different agencies. For example, the Kangra earthquake of 26 April 1986 is assigned magnitude of 5.5 by United States Geodetic Survey (USGS) and 5.7 by India Meteorological Department (IMD), respectively. As the USGS estimate is based on teleseismic data, the magnitude of 5.7 by IMD is considered to be a better value of local magnitude for this earthquake. Similarly, the four earthquakes recorded by Shillong array are also assigned different magnitudes by USGS, IMD and some other agencies. The various values of magnitude reported for the Uttarkashi earthquake of 19 October 1991 are $M_L = 6.5$, $M_S = 7.0$ and $M_W = 6.8$. The magnitude values listed in Table 1 as published magnitude, M_p, are those finally reported by IMD, which is the official agency for estimating the earthquake parameters in India.

Table 1: Details of the Six Earthquakes and the Number of Three-Component Strong-Motion Records Obtained from Them in the Himalayan Region

EQ. #	Name of EQ.	Date D M Y	Epicentre Lat (N)	Epicentre Long (E)	M_p	MMI Max.	H (km)	No. of Records
1.	Kangra	26 04 86	32.175	76.287	5.7	7	7	09
2.	Meghalaya	10 09 86	25.564	92.200	5.5	6	28	12
3.	N.E. India	18 05 87	25.479	93.598	5.7	5	50	14
4.	N.E. India	06 02 88	25.500	91.460	5.8	7	15	18
5.	Burma Border	06 08 88	25.384	94.529	7.2	7	91	33
6.	Uttarkashi	19 10 91	30.738	78.792	6.5	7	19	13

Two independent estimates of the strong-motion local magnitude are obtained from the two horizontal components of each of the records from the six earthquakes listed in Table 1. The original accelerograms are recorded on 70 mm photographic film. These were digitized using a semi-automatic Calcomp-9000 digitizer and were processed to correct for the high-frequency digitization errors, dynamic response of the recording transducers, and the baseline distortions (Chandrasekaran and Das, 1993). Such corrected data, which are available at equally spaced time intervals of 0.02 s, have been used to compute the strong-motion magnitudes in the present study.

THE METHODOLOGY

To use the strong-motion accelerograms for computing the local magnitude, it is necessary first to synthesize the response of the WAS to a given accelerogram as input excitation. The WAS can be modelled as a single-degree-of-freedom system with natural period $T_n = 0.8 s$, fraction of critical damping $\varsigma = 0.8$ and static magnification $V_s = 2800$. However, due to distortion of the taught-wire suspension in a WAS, the actual magnification may be smaller than the ideal value of 2800 (Uhrhammer and Collins, 1990), which may lead to somewhat underestimation of the conventional magnitude. But, for synthesizing the WA seismograms from strong-motion records, one can realize the ideal value of the magnification V_s. Thus, the displacement record, $x(t)$, of a WAS to a horizontal ground acceleration, $\ddot{z}(t)$, can be obtained by solving the following equation of motion:

$$\ddot{x}(t) + 2\varsigma\omega_n\dot{x}(t) + \omega_n^2 x(t) = -V_s\ddot{z}(t) \tag{1}$$

where $\omega_n = 2\pi / T_n$ is the angular natural frequency of the WAS. The solution of this equation can be obtained in terms of the Duhamel integral as follows

$$x(t) = \frac{-V_s}{\omega_n\sqrt{1-\varsigma^2}} \int_0^t \ddot{z}(\tau) e^{-\varsigma\omega_n(t-\tau)} \sin\left[\omega_n\sqrt{1-\varsigma^2}(t-\tau)\right] d\tau \tag{2}$$

This integral can be solved very efficiently by using the method due to Nigam and Jennings (1968), as used to compute the response spectra of ground acceleration. The same method has been followed in the present study to compute the synthetic Wood-Anderson seismograms from strong-motion accelerograms.

If $A_{synthetic}$ is the peak amplitude in mm of the computed Wood-Anderson seismogram, the local magnitude would conventionally be defined as

$$M_L = \log A_{synthetic} - \log A_0(R) \tag{3}$$

However, the strong-motion local magnitude in the present study is defined using improved attenuation function $Att(\Delta_0)$ rather than $\log A_0(R)$, as mentioned before. The procedure used can be summarized as follows:

- Compute M from

$$M = \log A_{synthetic} - Att(\Delta_0) \tag{4}$$

- Correct M for local geological site conditions to get $\overline{M}_L^{SM}$ as

$$\overline{M}_L^{SM} = M - b_2(M)(2-s) \tag{5}$$

  where $s = 0$ for recordings on soft ground and $s = 2$ for recordings on firm ground. The coefficient $b_2(M)$ is given by Trifunac (1991a) in a tabular form, which is obtained empirically using strong-motion data from California region of United States.

- The strong-motion local magnitude, M_L^{SM}, is finally defined as

$$M_L^{SM} = \overline{M}_L^{SM} - D\left(\overline{M}_L^{SM}\right) \tag{6}$$

  where $D\left(\overline{M}_L^{SM}\right)$ is another empirical correction factor which is introduced to account for the observed systematic differences between the strong-motion magnitudes, $\overline{M}_L^{SM}$, and the published magnitudes, M_p, which are mostly based on relatively distant recordings on conventional seismographs. The correction factor $D\left(\overline{M}_L^{SM}\right)$ is also listed in a tabular form by Trifunac (1991a) as a function of the conventional magnitude M_p for the western US region.

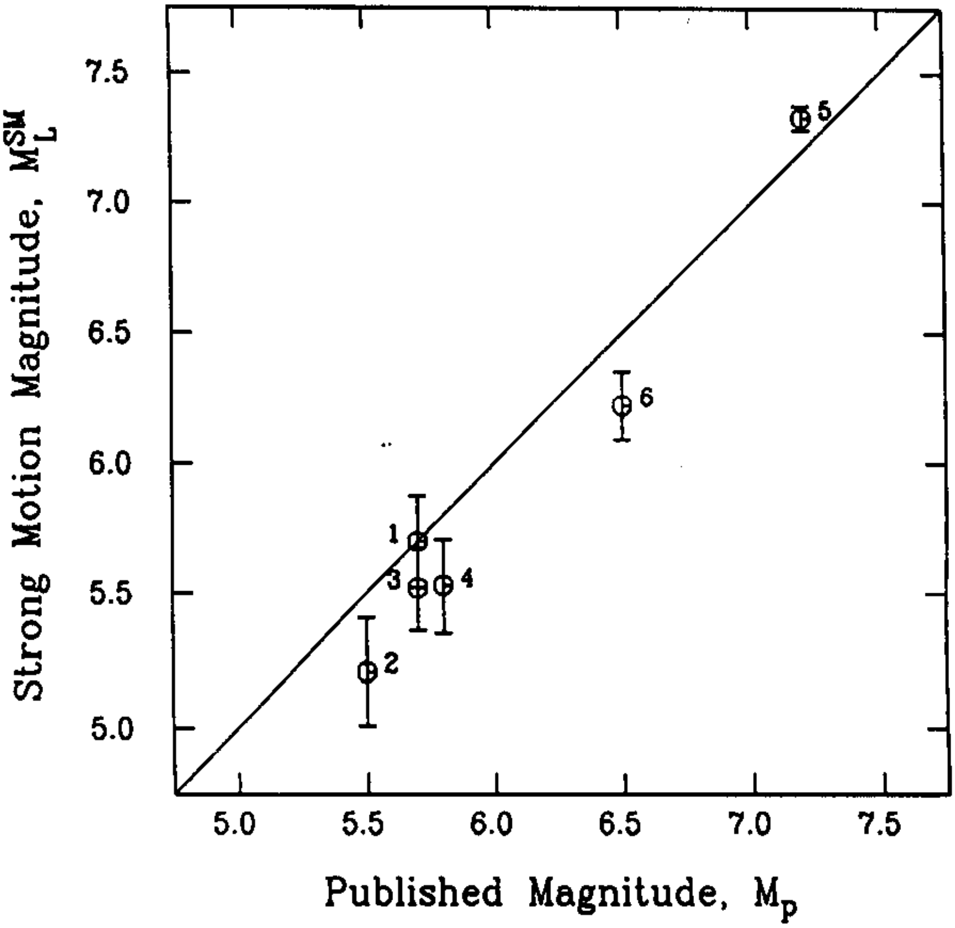

Fig. 1 Comparison of mean ± one standard deviation ranges of strong-motion magnitudes with the corresponding conventional magnitudes

RESULTS AND DISCUSSION

Using the methodology described in the preceding section, strong-motion local magnitudes, M_L^{SM}, are computed for the horizontal components of all the strong-motion records from the six earthquakes as listed in Table 1. For a given earthquake, the values of M_L^{SM} obtained from different accelerograms are found to vary very widely. This variation may be ascribed to some extent to the variations in the hypocentral distance, local site condition, the propagation path effects and the change in the radiation pattern with azimuthal direction. But, some of the magnitude values, which lie far away from the predominant band of the variation, are not considered to be reliable and are thus deleted from the analysis. The mean plus and minus one standard deviation ranges of M_L^{SM} as obtained from only good quality of data are plotted in Figure 1 versus the published magnitude M_p. It is seen that, in general, these values of M_L^{SM} based on the correction factors $D\left(\overline{M}_L^{SM}\right)$ for western US due to Trifunac (1991a) systematically underestimate the corresponding M_p. For four earthquakes (EQ. # 2, 3, 4 and 6) out of six, M_L^{SM} is underestimated by 0.2 to 0.3 magnitude units. Therefore, in the following, the results are examined critically and analyzed in detail to arrive at the correction factors $D\left(\overline{M}_L^{SM}\right)$, which are appropriate for the Himalayan region.

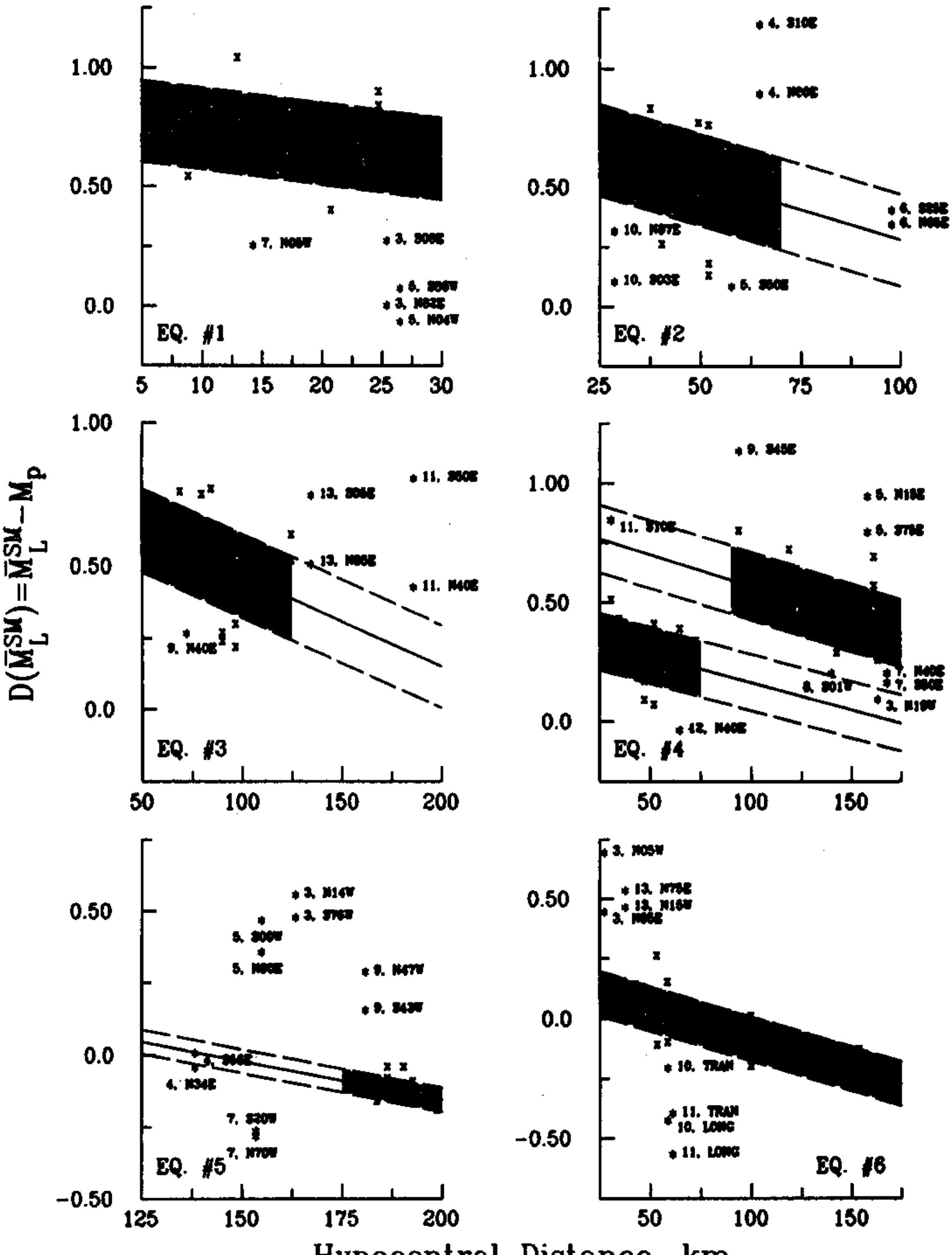

Fig. 2 Observed attenuation of the correction factor $D\left(M_L^{SM}\right)$ for six earthquakes in the Himalayan region

Figure 2 presents for the six earthquakes considered in this study the plots of $D\left(\overline{M}_L^{SM}\right)$ needed for exact matching of M_L^{SM} with M_p as a function of the hypocentral distance. Though the observed values of $D\left(\overline{M}_L^{SM}\right)$ are seen to be scattered quite widely, the plots in Figure 2 show well-defined decreasing trend of $D\left(\overline{M}_L^{SM}\right)$ with distance for all the six earthquakes. For magnitudes up to around 5.5, the contribution to $D\left(\overline{M}_L^{SM}\right)$ comes mainly from the additional anelastic attenuation suffered by seismic waves used to evaluate M_p compared to the strong-motion recordings at smaller distances. For larger magnitudes, the effect of saturation of strong-motion amplitudes relative to M_p results in increasingly smaller values of $D\left(\overline{M}_L^{SM}\right)$ with increase in M_p. Thus, as the distances of strong-motion recordings approach closer to those for the conventional seismographs used for evaluation of M_p, the differences $D\left(\overline{M}_L^{SM}\right)$ reduce to small values. The focal depth of the six earthquakes considered in this study vary over a wide range of about 7 to 91 km. But, the distance range of recordings is mostly much higher than the focal depths for all the six earthquakes. Thus, the hypocentral and epicentral distances do not differ so significantly, and the influence of focal depths on $D\left(\overline{M}_L^{SM}\right)$, if any, is not expected to be significant.

Table 2: Regression Coefficients and the Standard Deviations for the Relationship of Equation (7) Describing the Dependence of Correction Factor $D\left(\overline{M}_L^{SM}\right)$ on the Hypocentral Distance for the Six Earthquakes in the Himalayan Region

EQ. #	M_p	Distance Range (km)	Regression Coefficients		Standard Deviation σ
			C_1	C_2	
1.	5.7	8.8 – 26.5	0.8058	0.006297	0.1725
2.	5.5	28.7 – 67.4	0.7832	0.005034	0.1927
3.	5.7	59.2 – 124.5	0.7814	0.003152	0.1455
4.	5.8	30.7 – 72.5	0.3939	0.002283	0.1186
		93.8 – 167.9	0.8325	0.002636	0.1414
5.	7.2	177.2 – 198.7	0.3866	0.002719	0.0415
6.	6.5	27.3 – 153.6	0.1658	0.002502	0.0920

In Figure 2, the values of $D\left(\overline{M}_L^{SM}\right)$ corresponding to the good quality of data as mentioned before are plotted by crosses and those of poor quality by asterisks. The poor quality of data are those which lie in isolated positions relative to the trend of the majority of data. Our serial numbers corresponding to the poor quality of data points along with the component of recording are also indicated in Figure 2. As mentioned before, the reasons for such outliers may be many and quite differing from case to case. For example, for EQ. # 1, the under-estimation of $\overline{M}_L^{SM}$ for N05W component of accelerogram 7 can be assigned mainly to the directivity effects, whereas for accelerograms 3 and 5, the path effects dominate. For EQ. # 4, there are two distinct sets of data points, which describe the attenuation of $D\left(\overline{M}_L^{SM}\right)$ at smaller and larger distances, respectively. Contrary to the expectation, the distant strong-motion records are characterized by larger values of $D\left(\overline{M}_L^{SM}\right)$. This perhaps is due to the focussing of seismic waves as a result of reflection from some deeper inhomogeniety, probably Mohorovicic discontinuity. The wide scattering of data for EQ. # 5 is mainly due to highly heterogeneous geology and very rugged topography at smaller distances. Similar effects are also exhibited for EQ. # 6. Though the aim of the present study is not to investigate the causes for such spurious observations, it is important to understand the plausible causes for poor quality of observations before they are deleted from the analysis. From the results in

Figure 2, it is quite clear that a single recording cannot be considered adequate to get a reliable estimate of the magnitude of an earthquake. For practical applications, a stable mean value of the magnitude can be obtained only by using several good quality records from an earthquake.

To investigate the dependence of $D\left(\overline{M}_L^{SM}\right)$ on the hypocentral distance, following form of least-squares regression relation is fitted to each of the six earthquakes using only good quality of data points:

$$D\left(\overline{M}_L^{SM}\right) = C_1 - C_2\Delta_0 \pm \sigma \quad (7)$$

The mean regression equations thus obtained along with plus and minus one standard deviation ranges are also plotted in Figure 2. The shaded portions indicate the distance ranges for which the data has been used to obtain the least-squares relations for different earthquakes. The regression coefficients C_1 and C_2 and the standard deviation σ describing the attenuation of $D\left(\overline{M}_L^{SM}\right)$ with distance for all the six earthquakes are listed in Table 2. The published magnitude M_p is normally obtained from conventional seismographs at large distances (greater than about 100 km), whereas the strong-motion data are recorded at comparatively much smaller distances. Thus, the records of conventional seismographs suffer additional anelastic attenuation compared to those synthesized from strong--motion accelerograms, which is not accounted by the attenuation function $Att(\Delta_0)$ used to compute $\overline{M}_L^{SM}$. If Δ_1 is the hypocentral distance of recording the strong-motion accelerograms and Δ_2 that for the conventional seismographs, then in principle, $D\left(\overline{M}_L^{SM}\right)$ is given by $C_2(\Delta_2 - \Delta_1)$. However, as mentioned before, for magnitudes greater than about 5.5, the saturation of strong-motion amplitudes also contributes significantly to $D\left(\overline{M}_L^{SM}\right)$. For the first four earthquakes in the present study, which can be considered to be not much affected by the saturation effect, the observed values of $D\left(\overline{M}_L^{SM}\right)$ can be reconciled with $(\Delta_2 - \Delta_1)$ around 100 to 150 km. For EQ. # 5, the distance range of recording the strong-motion data is comparable to that for the conventional seismograms, and hence the observed $D\left(\overline{M}_L^{SM}\right)$ can mainly be assigned to the saturation of strong-motion amplitudes. In case of EQ. # 6, for $(\Delta_2 - \Delta_1) \approx 100$ km, the attenuation coefficient C_2 of Table 2 predicts $D\left(\overline{M}_L^{SM}\right)$ of about 0.25, which is probably counter balanced by the saturation effect, resulting in negligible net value of $D\left(\overline{M}_L^{SM}\right)$ for this earthquake. Because the details of saturation are not known precisely, it is not possible to use directly the attenuation coefficients C_2 to predict the correction factors $D\left(\overline{M}_L^{SM}\right)$ for all the magnitudes.

Table 3: Correction Factors $D\left(\overline{M}_L^{SM}\right)$ Needed to Match $\overline{M}_L^{SM}$ with M_p for the Six Earthquakes in the Himalayan Region

EQ. #	M_p	Distance Range (km)	No. of Records	$\overline{M}_L^{SM}$	Observed $D\left(\overline{M}_L^{SM}\right)$
1.	5.7	8.8 – 26.5	13	6.34	0.64
2.	5.5	28.7 – 67.4	17	6.01	0.51
3.	5.7	59.2 – 96.7	19	6.22	0.52
4.	5.8	30.7 – 72.5	14	6.07	0.27
5.	7.2	177.2 – 198.7	12	7.07	-0.13
6.	6.5	53.3 – 99.8	12	6.47	-0.03

Table 3 lists along with M_p the distance ranges of strong-motion recordings, the number of good quality of records available and the mean-values of the magnitude $\overline{M}_L^{SM}$ computed using Equations (4) and (5). The last column of this table gives the correction factors $D\left(\overline{M}_L^{SM}\right)$ actually required to match the strong-motion magnitude with M_p. The observed values of $D\left(\overline{M}_L^{SM}\right)$ for the six earthquakes analysed in the present study are plotted as a function of M_p in Figure 3 along with the Trifunac's (1991a) recommendation for western US and that of Lee et al. (1990) for Yugoslavia. It is seen that neither of the two available corrections factors are suitable for the Himalayan region. However, it is seen that the trend of the curve for western US matches quite well with the observed data for the Himalayan region. In general, the present values fall below the Trifunac's curve and close matching for EQ. # 1 and 5 can be considered incidental or due to the peculiar conditions under which the strong-motion data of these earthquakes are recorded. Thus, to get $D\left(\overline{M}_L^{SM}\right)$ for Himalayan region, the Trifunac's curve is shifted down to a position which gives minimum value for the sum of the squared differences from the observed values of $D\left(\overline{M}_L^{SM}\right)$ for EQ. # 2, 3, 4 and 6. This is plotted by solid curve in Figure 3. To compute the strong-motion local magnitudes for Himalayan earthquakes, it is recommended to use the $D\left(\overline{M}_L^{SM}\right)$ values corresponding to this curve.

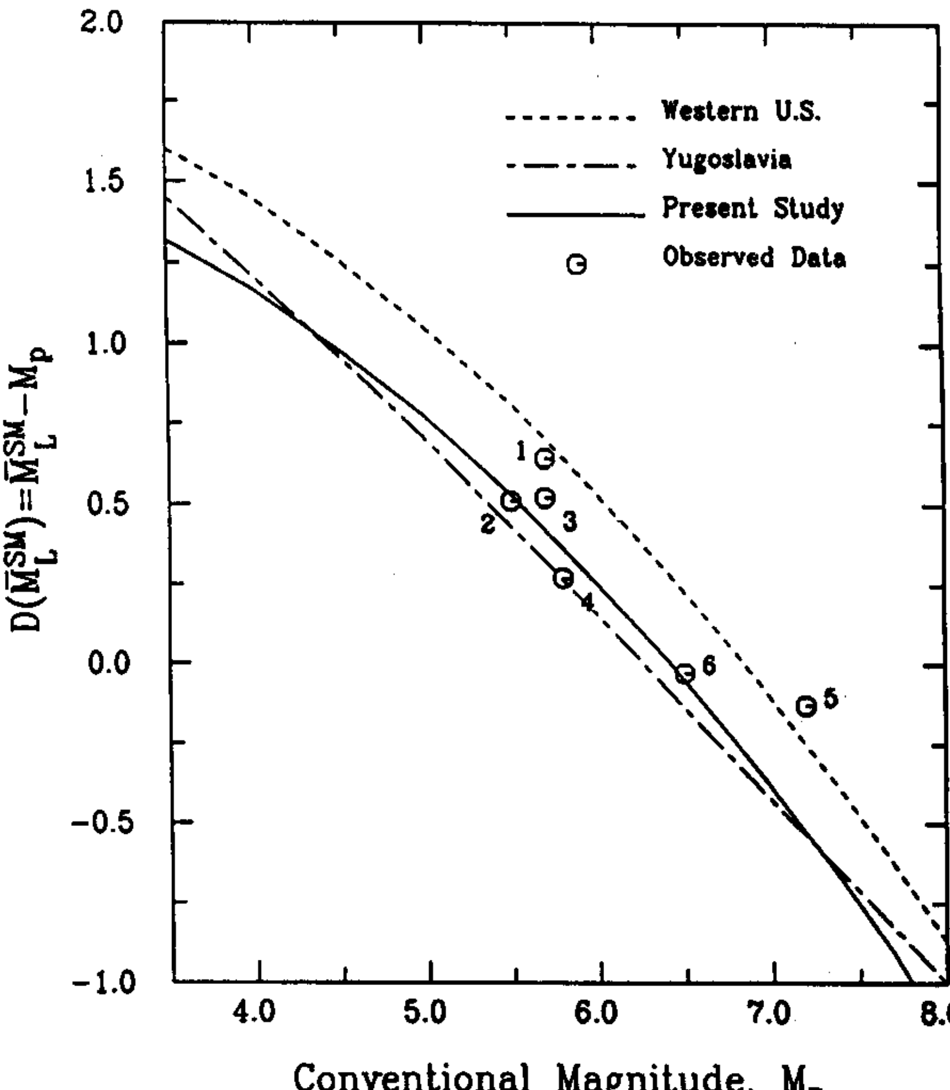

Fig. 3 Trifunac's correction factor $D\left(M_L^{SM}\right)$ modified to match the observed data in the Himalayan region

CONCLUSIONS

For the purpose of engineering seismic risk analysis, it is very important to have reliable and accurate estimation of the earthquake magnitudes. Errors in magnitude of the order of 0.3 may be equivalent to a factor of about 2 in the ground motion amplitudes. The magnitude values listed in the catalogues of past earthquakes in a region are mostly based on Wood-Anderson Seismographs at very long epicentral distances (100 to 600 km) compared to the distance range (less than about 50 to 100 km) at which the strong-motions are normally recorded. The magnitude values determined from very distant recordings may sometimes be in large errors due to uncertainties in the attenuation law used. Thus the Wood-Anderson records synthesized from strong-motion accelerograms provide a useful alternative for getting more reliable estimates of local magnitudes.

At small distances, the Wood-Anderson seismograph goes out of scale even for low to moderate magnitude earthquakes. On the other hand, for large magnitude earthquakes, the strong-motion amplitudes get saturated at close distances. Further, the distant recordings on conventional seismographs suffer from significant anelastic attenuation effects compared to the strong-motion records. Therefore, to account for the anelastic attenuation and the saturation effects, the magnitude estimates from strong-motion data need an empirical correction to get results close to the original Richter's definition of local magnitude scale or to the surface-wave magnitudes for larger events. This is necessary to maintain continuity with the magnitude data of past earthquakes which will have to be used for a long time till data with better estimates of magnitude (say, moment magnitude) are collected for a sufficiently long period. The available correction factors based on the data from other countries are not found to be suitable for the Himalayan region due to the differences in the geology and the tectonic setup. Therefore, using the strong-motion data recorded from six earthquakes, the method of computing local magnitudes has been calibrated for Himalayan region. But, due to a very limited data base, the present recommendations are considered to be of preliminary nature. Nevertheless, the proposed results would be useful to estimate the strong-motion local magnitudes till more data are available to make significant improvements.

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