Bull. Ind. Soc. Earth. Tech., Paper No. 329, Vol. 30, No. 1, Mar. 1993, pp. 19-25

AN EXAMPLE OF COMPUTING M_L^{SM} FOR THE DHARAMSHALA EARTHQUAKE OF 1986 IN INDIA

by

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ABSTRACT

For epicentral distances R < 100 km, the definition of the local magnitude scale, M_L must be modified to reflect the near source geometry, and to reconcile the differences between the strong motion magnitude M_L^{SM} (based on strong motion data) and the distant seismological estimates using M_L, M_S (local and surface wave magnitude) and other magnitude scales. The procedure for computing M_L^{SM} is illustrated for Dharamshala earthquake of 1986 in India.

INTRODUCTION

The purpose of this brief note is to emphasize the need for systematic and unbiased computation of local magnitude, using strong motion data. Development of carefully calibrated, uniform and homogeneous catalogues on earthquake magnitudes is the essential and the first step in creating the data base for engineering seismic risk studies (Anderson and Trifunac, 1978), for development of site specific spectra in design of important structures (Trifunac and Todorovska, 1989), and for microzonation work based on the Uniform Risk Spectrum Technique (Trifunac, 1988, 1990). Computation of magnitude from strong-motion data also offers physical advantages, which are associated with short propagation path, and thus the uncertainties in scaling and calibrating the attenuation equation can be reduced.

In seismological estimates of earthquake magnitude, in many parts of the world, the "standard", "published" magnitude definitions are used without detailed verification of whether the chosen attenuation law is acceptable for the region in question. Yet, many studies show significant regional differences when attenuation laws are actually calibrated by the data recorded locally. These differences are particularly important for engineering estimates of strong motion shaking, since high frequency waves are affected most by the regional differences in Q (Trifunac and Lee, 1990; Lee and Trigunac, 1992, Trifunac 1992).

Following major destructive earthquakes, it is important to asses, as soon as possible, the levels of destruction, so that the rescue operations can be initiated without a delay. Since all local seismological stations typically go off scale, the magnitude determination as obtained from distant recording, and the associated inaccuracies in attenuation laws can result in misleading predictions, during the first precious hours following a destructive earthquake. In south east Europe, for example, we found systematic bias in seismological estimates of magnitude, as large as one magnitude unit (Trifunac and Herak, 1992) by performing detailed comparison of seismological regional estimates of magnitudes with the strong motion magnitude M_L^{SM} (Trifunac,1991a, Lee et al.,1990, Trifunac and Herak,1992).

STRONG MOTION MAGNITUDE M_L^{SM}

Computation of magnitude, using strong motion data, requires a separate attenuation equation, and definition of the representative distance to the source (Trifunac, 1991a), since the original Richter's definition of M_L, in 1935(Richter, 1935;1958) was presented only for epicentral distances $R > 25$ km. Later, Gutenberg & Richter (1942) extended this attenuation law to $R < 25$ km, by using less sensitive instruments (static magnification $V_s = 4$, damping $\zeta \approx 1$, and natural period $T_n \approx 10$ sec.). The final table on the attenuation law $\log_{10}A_o(R)$, published in 1958, reflects these later changes. We used this attenuation law for computation of M_L^{SM} since 1970(Trifunac and Brune,1970), but it was not until Luco's (1982) study that it became clear that, at small epicentral distances, $\log_{10}A_o(R)$ must be modified. Jennings and Kanamori(1983) also noted that $\log_{10}A_o(R)$ must be modified for small R, but their failure to include the local geologic site effects in the analysis resulted in biased estimates of this attenuation law.

Near the source, when the distance to the fault plane becomes small relative to the source dimensions, the attenuation equation becomes multivalued and depends on the source geometry, and size. However, fortunately, for the computation of M_L^{SM}, high frequency strong motion waves are mostly sensitive to the local movements on the fault, and this enables one to reduce the family of magnitude dependent attenuation equations to one which is independent of magnitude (Trifunac, 1991a). This results in the final definition of M_L^{SM},

$$M_L^{SM} = \log_{10} A_{synthetic} - Att(\Delta_o) - b_2(M)(2 - s) - D(\overline{M}_L^{SM}), \quad (1)$$

where $A_{synthetic}$ is the peak response of Wood-Anderson seismometer (natural period $T_n = 0.8$ sec., fraction of critical damping $\zeta = 0.8$, and static magnification $V_s = 2800$) computed from the recorded strong motion accelerograms. $Att(\Delta_o)$ (Table 1) is the "new" attenuation equation which is different from $\log_{10}A_o(R)$ for $R < 100$ km and which coincides with $\log_{10}A_o(R)$ for $R > 100$ km. $b_2(M)$ (see Table 2) represents correction for different geologic site conditions (most strong motion accelerographs are typically on $s = 0$ i.e. sediments, while most seismological stations are an $s = 2$, basement rock; see Trifunac and Brady (1975) for definition and description of the parameters s). $D(\overline{M}_L^{SM})$ (Table 3) represents systematic departure of M_L^{SM} relative to the distant magnitude estimates. It could be associated with different Q along different propagation paths in strong motion recording and in more distant seismological recording (Trifunac, 1991b), but this must be studied further before we understand all the causes resulting in $D(\overline{M}_L^{SM})$. The last two terms in equation (1) cannot be absorbed in $Att(\Delta_o)$.

To define M_L^{SM}, one first computes

$$M = \log_{10} A_{synthetic} - Att(\Delta_o) \quad (2)$$

with $Att(\Delta_o)$ as in Table 1, where R is epicentral distance, H is focal depth and $\Delta_o = (R^2 + H^2)^{1/2}$. With this M, we next compute

$$\overline{M}_L^{SM} = M - b_2(M)(2 - s) \quad (3)$$

where $b_2(M)$ is given in Table 2. Finally,

$$M_L^{SM} = \overline{M}_L^{SM} - D(\overline{M}_L^{SM}) \quad (4)$$

where $D(\overline{M}_L^{SM})$ is given in Table 3, in terms of M_p or $\overline{M}_L^{SM}$. When only strong motion data is available, one can use Table 3(b). When seismological estimates provide the "published" magnitude M_p, Table 3(a) can also be used.

To illustrate this computation, we consider $A_{synthetic}$ in Table 1 of Jain et.al. (1992), for six strong motion recordings of the Dharamshala earthquake of 1985 in India. The first two columns in Table 4 present the station names and the station code names. The third column shows $\log_{10}A_o$ (in mm) for the peaks of two horizontal components of computed response of Wood-Anderson seismometer to excitation by the recorded strong motion accelerographs. R is epicentral and Δ_o is hypocentral distance (in km), assuming H = 10 km (see column 4). Column 5 gives $Att(\Delta_o)$ from Table 1 and column 6 gives M in equation (2). Assuming s = 0 (sediments) for all stations, column 7 then gives $\overline{M}_L^{SM} - b_2(M)(2-s)$. Columns 8 and 9 give M_L^{SM} is equation (4) with $D(\overline{M}_L^{SM})$ in terms of $\overline{M}_L^{SM}$ and M_p respectively. The resulting average M_L^{SM} are 5.23 and 5.31 respectively. For comparison, columns 10 and 11 give $\log_{10}A_o(R)$ and M_L computed as in Jain et al. (1992). M_p for this earthquake is 5.4

DISCUSSION AND CONCLUSIONS

From seismological viewpoint, the "published" magnitude M_p = 5.4 for this earthquake is in excellent agreement with M_L = 5.54 and with M_L^{SM} = 5.23 to 5.31 (if s = 0) and 5.45 to 5.53 (if s = 2 at all stations). However, $\log_{10}A_o(R)$ cannot be reconciled with the near field data on attenuation in California, and I suspect might not survive the similar test in India, when sufficient strong motion data becomes available there to carry out such tests. Thus, the use of $\log_{10}A_o(R)$ on a routine basis will result in systematically based results on M_L for distances R < 100 km, and so should not be used in earthquake engineering estimation of seismic risk. We used $Att(\Delta_o)$ in Europe and found no difficulties or discrepancies with strong motion data there. At small epicentral distances (less than 100 km), the effects of different geological environment are probably small, especially if correct local site characterization (s = 0 for sediments, s = 2 for basement rock and s = 1 for intermediate sites is used via equation (3)). Systematic differences and biases of regional magnitude estimates of the order of 0.2 to 0.3 are often acknowledged and accepted in routine seismological work. In engineering estimates of seismic risk, this corresponds to amplitude factors of 2 in the uniform risk spectrum amplitudes, and thus should be eliminated or reduced, whenever possible.

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Table 1.* $Att(\Delta_0)$ for $\Delta_0 = \sqrt{R^2 + H^2}$

Δ_0	$Att(\Delta_0)$	Δ_0	$Att(\Delta_0)$	Δ_0	$Att(\Delta_0)$
1	-1.62	110	-3.08	340	-4.21
5	-2.08	120	-3.13	360	-4.30
10	-2.30	130	-3.18	380	-4.38
15	-2.42	140	-3.23	400	-4.45
20	-2.51	150	-3.28	420	-4.52
25	-2.58	160	-3.33	440	-4.58
30	-2.63	170	-3.38	460	-4.63
35	-2.68	180	-3.43	480	-4.69
40	-2.71	190	-3.48	500	-4.73
45	-2.75	200	-3.53	520	-4.78
50	-2.78	220	-3.63	540	-4.82
60	-2.83	240	-3.73	560	-4.85
70	-2.88	260	-3.83	580	-4.88
80	-2.93	280	-3.93	600	-4.90
90	-2.98	300	-4.02		
100	-3.03	320	-4.12		

Table 2.* $b_2(M)$ in equation (3)

M:	3.5	4.5	5.5	6.5	7.5
$b_2(M)$:	0.10	0.10	0.11	0.12	0.13

* from Trifunac (1991a), for strong motion data in western United States.

Table 3(a).* $D(\overline{M}_L^{SM})$ in equation (4) versus M_p

M_p	3.1	3.5	4.0	4.5	5.0	5.5	6.0	6.4	6.8	7.0	7.4	7.7	8.0
$D(\overline{M}_L^{SM})$:	1.70	1.60	1.45	1.26	1.05	0.81	0.53	0.29	0.03	-0.11	-0.40	-0.63	-0.87

Table 3(b).* $D(\overline{M}_L^{SM})$ in equation (4) versus $\overline{M}_L^{SM}$

$\overline{M}_L^{SM}$	4.6	4.8	5.0	5.5	6.0	6.2	6.4	6.6	6.8	7.0	7.1	7.2	7.3
$D(\overline{M}_L^{SM})$:	1.60	1.53	1.45	1.27	1.07	0.94	0.78	0.58	0.38	0.18	0.08	-0.02	-0.12

* from Trifunac (1991a), for strong motion data in western United States.

Table 4

Comparison of M_L^{SM} for Dharamashala 1966 earthquake in India with M_L, computed from strong motion data

Station	Station code	$\log_{10} A(mm)$	$\frac{R(km)}{\Delta_0(km)}$*	$Att^{**}(\Delta_0)$	M	$\overline{M}_L^{SM}$ (Assuming $s = 0$ in $-b_2(M)(2-s)$)	M_L^{SM} (using $D(\overline{M}_L^{SM})$ in Table 3b)	M_L^{SM} (using $D(\overline{M}_L^{SM})$ in Table 3a)	$\log_{10} A_0(R)^{***}$	M_L
Dharamshala	DHAR	3.99 4.28	3.54 10.61	-2.31	6.30 6.59	6.08 6.37	5.13 5.42	5.21 5.50	-1.38	5.37 5.66
Shahpur	SHAH	4.01 4.38	10.38 14.41	-2.40	6.41 6.78	6.19 6.56	5.24 5.61	5.32 5.69	-1.51	5.52 5.89
Kangra	KANG	3.92 4.26	10.68 14.56	-2.41	6.33 6.67	6.11 6.46	5.16 5.50	5.24 5.58	-1.52	5.44 5.78
Nagrota Bagwan	NAG BAG	4.14 3.58	13.66 16.84	-2.45	6.59 6.03	6.37 5.81	5.42 4.86	5.50 4.94	-1.57	5.71 5.15
Baroh	BARO	3.74 3.69	21.60 23.80	-2.56	6.30 6.25	6.08 6.03	5.13 5.06	5.21 5.16	-1.76	5.52 5.47
Sihunta	SIHU	3.65 3.77	22.81 24.91	-2.58	6.23 6.35	6.01 6.12	5.06 5.18	5.14 5.26	-1.81	5.46 5.58
						AVE = 6.18	AVE=5.23	AVE 5.31		AVE=5.54

Note:

* Focal depth $H = 10$ km is assumed

** From Table 1

*** From Richter (1958)