A note on the differences in magnitudes estimated from strong motion data and from Wood-Anderson Seismometer

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Magnitude dependent differences between estimates of local magnitude, M_L (Richter[12,13]) based on strong motion data and on the response of Wood-Anderson Seismometer are analyzed. It is suggested that the principal cause for differences for intermediate and small magnitudes ($M < 5.5$) could be associated with anelastic attenuation; indicating Q values between 100 and 200 in Southern California. For large magnitudes the observed differences are interpreted to result from saturation of strong-motion amplitudes relative to other long period estimates of magnitude.

INTRODUCTION

Since its introduction in 1935, the local earthquake magnitude scale, M_L, has been used in various scaling relations to estimate the amplitudes of strong ground motion for earthquake engineering applications (e.g. Trifunac[15,16]; Trifunac and Lee[21,22],). It was defined in terms of the logarithm of the peak response, A (in mm) of Wood-Anderson Seismometer (WAS), with static magnification $V_s = 2800$, natural period $T_n = 0.8$ sec and fraction of critical damping $\zeta = 0.8$ and in terms of the empirical attenuation law '$\log_{10}A_0(R)$' in which R is epicentral distance (in km) (Richter[12,13]). With accumulation of strong motion accelerograph data near earthquake sources, for epicentral distances less than 50 to 100 km, especially after 1971 (Lee and Trifunac[8]; Lee[7] Trifunac[17]) it became possible to begin computing the estimates of M_L using strong motion data (Trifunac and Brune[20]; Trifunac and Brady[18]). With subsequent increase in the number of recorded strong motion accelerograms, studies could be carried out on the differences between such estimates and the published magnitude M_p for the same earthquakes (Luco[10]; Trufunac[17]; Lee[7]). These studies lead to the improved form of attenuation equation for use with the estimates of M_L from strong motion data.

Our recent studies confirmed that not only the distance dependent corrections, but also magnitude dependent corrections are required for the estimates of M_L from the recorded strong motion accelerograms (Luco[10]; Trifunac[17]). The correction, $D(\bar{M}_L^{SM}) \equiv \bar{M}_L^{SM} - M_p$, where $\bar{M}_L^{SM}$ is determined from synthetic Wood-Anderson seismometer response and M_p is the published magnitude for the same earthquake (Lee[7]), describes the observed systematic differences between near source ($R < 50$ to 100 km) and more distant ($50 < R < 600$ km) estimates computed from WAS. The required corrections are as large as 1.5 magnitude units for $M_p \sim 3.5$, they decrease with M_p to zero near $M_p = 6.5$, and become negative for $M_p > 6.5$ (Trifunac[17]).

The assumption, in this and in our previous works, has been that the attenuation equation $\log_{10}A_0(R)$ in the definition of M_L scale (Richter[12]) should not be changed for $\Delta > 100$ km. Our proposed extension and modification of this empirical law (Trifunac[17]) has been presented only for use with strong motion data, for the distance range $\Delta < 100$ km and for the magnitudes typically greater than 3 to 4 i.e. for the levels of shaking for which the Wood-Anderson seismometer would usually go off scale. We have also assumed that the attenuation of strong motion amplitudes for use with definition of M_L^{SM} should be based on the analysis of attenuation of spectral amplitude of recorded strong motion. However, such approach is neither necessary nor unique, as one could search for the best 'new' attenuation equation, subject to a constraint that, for example, $D(\bar{M}_L^{SM})$ is to be minimized. Such approach was considered for M_L data recorded by Wood-Anderson Seismometers in Southern California (Hutton and Boore[4]).

For engineering risk estimates it is important to have homogeneous data on magnitudes of past earthquakes and for time intervals as long as possible. It is also essential to have some common, generally excepted relative scaling parameter. Since M_L has served this role for many years, changes in its definition would result in scaling proplems which are too many and too complex to enumerate here. Also lumping many different physical effects together in one attenuation equation, which is determined by a regression analysis, produces results which are specific to the region contributing the data. When such scaling is to be repeated or introduced in another region one has to do the complete regression analysis again. After such analyses have been completed, assuming that adequate data base has been gathered (this

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often takes many years), it may not be obvious how the resulting magnitude estimates may be related or even compared, and how the differences in physical, region specific parameters may have contributed to the observed differences.

The purpose of this note is to analyze the plausible physical causes for $D(\bar{M}_L^{SM})$ and to explore what could be the explanations for its amplitudes. Understanding of these causes will help to interpret regional variations of magnitude estimates, how these variations depend on geology and tectonics, and how other different magnitude scales can be computed and related to a chosen reference scale. Fundamental and the essential first step in all engineering studies of earthquake risk must address the question of the homogeneity, uniformity and compatibility of the earthquake scaling parameters (magnitude or intensity) with their use in scaling the strong ground motion amplitudes and their duration. The published seismological catalogues may have been tested for homgeneity and uniformly of coverage, but may not use the required or compatible definitions of magnitude for earthquake engineering scaling needs. The resulting biases in the end result, in earthquake design spectra, for example, may be not only large, but also difficult to identify and interpret, since the experts who may not be familiar with all fine details involved in engineering characterization of strong ground motion might be responsible for preparing the data on seismicity.

EXAMPLES OF POSSIBLE PHYSICAL CAUSES FOR $D(\bar{M}_L^{SM})$

In the following, several plausible models are considered to explain the amplitudes and the shape of $D(\bar{M}_L^{SM})$. Only those phenomena, which produce large and probable contributions to $D(\bar{M}_L^{SM})$, will be considered. As this function has been determined by a regression analysis, it represents an empirical sum total of all factors leading to the differences between M_L and $\bar{M}_L^{SM}$, and so, it must also depend on other physical, and data recording and processing characteristics which will not be considered here.

Dispersion of large peaks

Strong motion displacement may contain large (impulsive) peaks (Fig. 1a). Such peaks are common during intermediate and small earthquakes and at stations with small epicentral distance. Dispersion, scattering, reflection, refraction and attenuation of high frequencies will eventually disperse such peaks, as the waves travel to greater distances. The resulting motion will become 'smoother' and will contain longer periods (Fig. 1b). As the estimation of M_L is directly related to the peaks in the ground motion it is clear that WAS will sample different features of strong motion in these two examples. Strong motion accelerographs record strong shaking at epicentral distances less than 50 to 100 km. WAS is more sensitive and records motions between about 50 km and 600 km (Fig. 2).

To evaluate a 'bias' which these large impulsive peaks of ground motion may introduce into the magnitude estimates we consider the following hypothetical example. Suppose that at two recording stations, one far from and the other close to the source, the ground displacement,

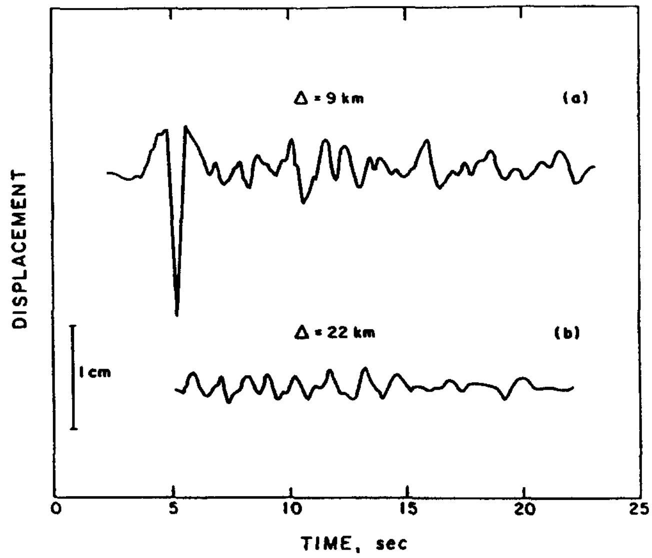

Fig. 1 Comparison of strong motion displacement records at two epicentral distances. Note the displacement pulse, at 5 seconds, in the top trace. These records represent the same (radial) component of motion, are along the same azimuth and have been recorded during the same earthquake.

$d(t)$, has been 'recorded' such that only dispersion is present. Attenuation of any kind is assumed to be absent. Then taking the strong motion part of motion, T, (Trifunac and Brady[19]; Trifunac and Westermo[23,24]) at both stations one can assume that $\int_0^T d^2(t)dt$ is the same at both stations, where T represents duration. Then the ratio of the root mean square displacements at the two stations will be equal to $(T_2/T_1)^{1/2}$ and also to $(N_2/N_1)^{1/2}$ where N_2 and N_1 represent the number of peaks of WAS at these two stations, during the respective durations T_2 and T_1. Then, using the results of Gupta and Trifunac[2], arbitrarily normalizing all peaks of WAS for say $N = 100$ (distant station), to one, and assuming that the station closer to the source records 4, 6, 8, 10 or 50 peaks, the ratios of the peak of the ground displacement (and of WAS) at or near the source to that of distant station would be as shown in Table 1. In this table ε represents a measure of the width of recorded spectra (Gupta and Trifunac [2]). A spectrum is 'narrow' for small ε and 'white' for $\varepsilon = 1$. Typical earthquake spectra combined with the transfer function of the WAS would have ε between 0.6 and 0.8.

Table 1. Approximate Peak amplitude of a short impulsive with N = 4,6,8,10 or 50 peaks normalized to the root mean square amplitude of a function and relative to a "long" record with N = 100 peaks. When ε = 0 there is only one frequency present while for ε = 1 the spectrum of motion is "white". (After Gupta and Trifunac,[2]).

N	$\varepsilon = 0.4$	$\varepsilon = 0.6$	$\varepsilon = 0.8$	$\varepsilon = 1.0$
4	3.00	2.93	2.74	2.03
6	2.70	2.66	2.53	2.05
8	2.48	2.46	2.37	2.00
10	2.32	2.29	2.22	1.93
50	1.31	1.31	1.30	1.26
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Table 1 and this approximate analysis suggest that a short (e.g. 4 peaks) impulsive ground motion will lead to

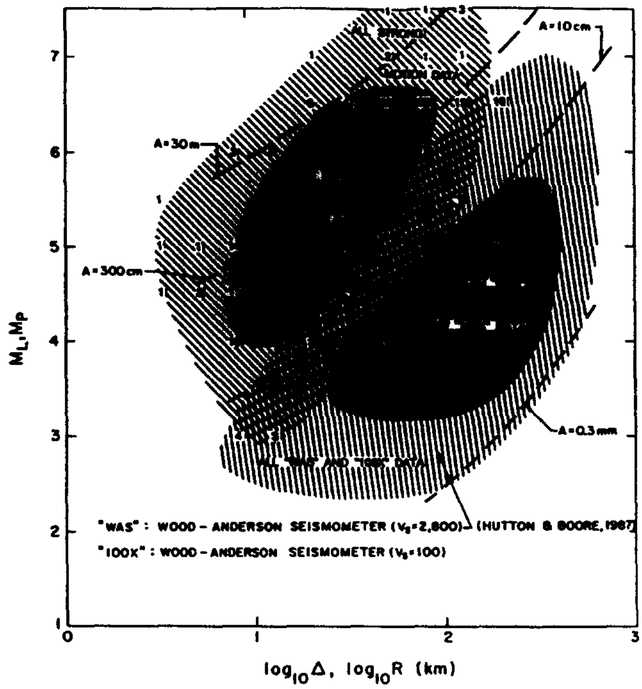

Fig. 2 Comparison of magnitude-distance distributions of the available data for recordings with strong motion accelerographs and with the Wood-Anderson seismometers (WAS). The matrix of numbers indicates the number and the distribution of strong motion data which is used in this work. A represents the amplitude recorded by WAS (or by its mathematical equivalent, for $A \geqslant 10$ cm)

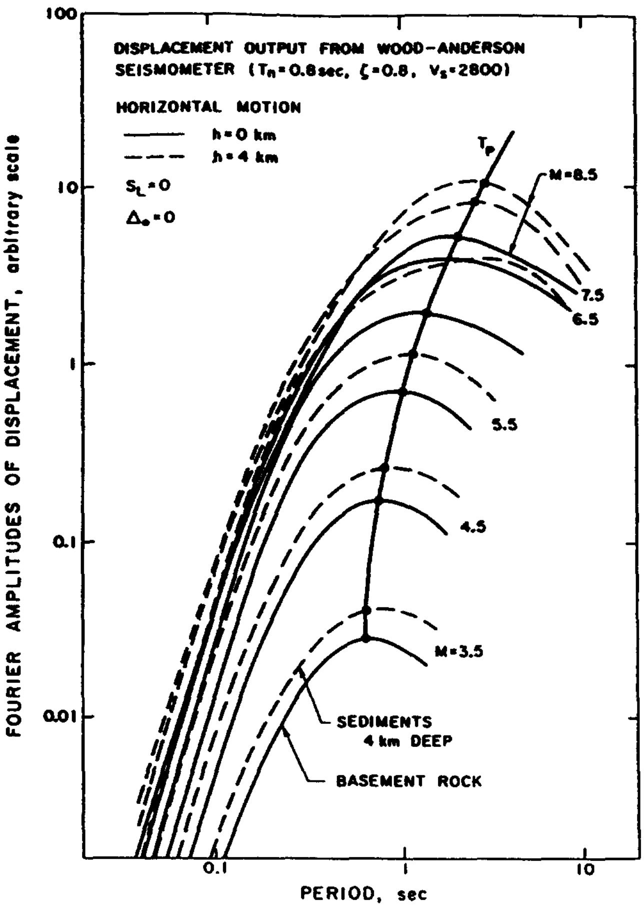

Fig. 3 Expected Fourier amplitude spectra (arbitrary scale) of displacement recorded by the Wood-Anderson seismometer (natural period $T_n = 0.8$ sec; fraction of critical damping $\zeta = 0.8$ and static magnification $V_s = 2800$), at epicenter ($\Delta_0 = 0$ km) on 'rock' local soil conditions $S_L = 0$) and for recording site on basement rock ($h = 0$ km) or on sediments 4 km deep.

peak displacement amplitudes at most 2 to 3 times larger than for the 'longer' and 'smoother' motion at greater distance ($N = 100$), for the same 'total energy' in the record. This suggests that M_L estimated from strong motion data at small epicentral distance and with displacements containing large pulses (Fig. 1a) may be larger by 0.3 to 0.5 magnitude units, whenever such impulses are present in ground motion.

Attenuation of high frequencies

We presented an empirical description of the frequency dependent attenuation of strong motion for epicentral distances less than 50 to 100 km (Trifunac and Lee[22]). Beyond about 100 km, there is shortage of recorded strong motion data (Lee[7]), and at present the frequency dependent attenuation there cannot be determined empirically. For engineering estimation of strong motion amplitudes, we use $\log_{10}A_0(R)$ attenuation function of Richter[12] on an interim basis, because it should represent good overall average of attenuated amplitudes, centered around 1 *Hz*.

To evaluate approximately the consequences of the attenuation on $D(\bar{M}_L^{SM})$, one might consider the relative attenuation law of the form $\exp(-(\omega\Delta/2Q\beta))$, where ω is frequency, Δ representative distance, β-shear wave velocity and Q- the quality factor. This form of attenuation has been considered by many investigators and it is known that in California Q may be bettween 150 and 300 for frequencies near 1 *Hz*. For sediments, typical results from several studies suggest low Q near the surface, from 10 to 50, increasing to 25 to 100 between 500 m and 2 km depth (McDonald, *et al.*[11]. Tullos and Reid,[25] Hauge,[3] Ganley and Kanasewich,[1] Johnson and Silva,[5] Joyner *et al.*[6]).

One contribution to $D(\bar{M}_L^{SM})$ can come from attenuation of high frequency waves, especially for smaller earthquakes. These waves would be 'seen' by near strong motion stations, say at distance Δ_1, but would be attenuated before reaching a more distant station, say at distance Δ_2, where WAS would record the motion. Assuming that the main contribution to M_L comes from the periods of ground motion near the peak, at T_p, of the Fourier amplitude spectra of the imput motion, multiplied by the transfer function of WAS, the contribution to $D(\bar{M}_L^{SM})$ from the attenuation would be $\exp[(\pi/T_pQ\beta)(\Delta_2 - \Delta_1)]$. Since T_p depends on the magnitude and the recording site conditions (Fig. 3), the result will also depend on these two. Table 2 illustrates this for $Q = 100$, 200 and 500, for $\Delta_2 - \Delta_1 = 100$ and 200 km and for $\beta = 1.5$ and 2.5 km/sec. It is seen that in the examples considered, $\log_{10}(A_1/A_2) \sim D(\bar{M}_L^{SM})$, with A_1 and A_2 representing the expected peak displacement amplitudes at Δ_1 and Δ_2 respectively. This table shows $D(\bar{M}_L^{SM})$ as large as 1.7 for $M = 3.5$ and for $Q = 100$ (Fig. 4).

Table 2. $\log_{10}(A_1/A_2)$, *where* A_1 *and* A_2 *are the expected peak response amplitudes at distances* Δ_1 *and* Δ_2 *for* $\Delta_2 - \Delta_1 = 100$ *and 200 km, for* $Q = 100$, *200, and 500, for* $\beta = 1.5$ *and 2.0 km/sec and for recordings on basement rock (r) or sediments (s), and for magnitude* $M = 3.5 \ldots 8.5$.

M		T_p(sec)	$\Delta_2 - \Delta_1 = 100$ km, $\beta = 1.5$ km/sec, $Q = 100$	$Q = 200$	$Q = 500$	$\Delta_2 - \Delta_1 = 200$ km, $\beta = 2.5$ km/sec, $Q = 100$	$Q = 200$	$Q = 500$
3.5	r	.65	1.40	0.70	0.28	1.68	0.84	0.34
	s	.70	1.30	0.65	0.26	1.56	0.78	0.31
4.5	r	.80	1.14	0.57	0.23	1.36	0.68	0.27
	s	.85	1.07	0.54	0.21	1.28	0.64	0.26
5.5	r	1.0	0.91	0.45	0.18	1.09	0.55	0.22
	s	1.20	0.76	0.38	0.15	0.91	0.45	0.18
6.5	r	1.50	0.61	0.30	0.12	0.73	0.36	0.15
	s	1.80	0.51	0.25	0.10	0.61	0.30	0.12
7.5	r	1.80	0.51	0.25	0.10	0.61	0.30	0.12
	s	2.50	0.36	0.18	0.07	0.44	0.22	0.09
8.5	r	2.20	0.41	0.21	0.08	0.50	0.25	0.10
	s	3.00	0.30	0.15	0.06	0.36	0.18	0.07

Saturation of strong motion amplitudes with respect to M_p and high-pass filtering by WAS

Figure 3 shows estimates of the Fourier amplitude spectra of ground motion on rock or on sediments 4 km deep, for magnitudes 3.5, 4.5, ... 7.5 and 8.5 (Trifunac and Lee[21] Trifunac[16,17]) and multiplied by the transfer funtion of WAS plotted with arbitrary vertical scale. Peak amplitudes of the time functions corresponding to these spectra would yield local magnitude M_L. Perusal of Fig. 3 shows that the relative vertical separation of the two consecutive spectra, differing by one magnitude unit, decreases from small towards large magnitudes. The expected peak amplitudes of the time functions are proportional to the area under these spectra, but to a first approximation, may be assumed to be proportional only to the amplitudes of the spectra near T_p. Assuming that the peak of $M_p = 5.5$ should correspond to M_L, the amplitudes of spectra in Fig. 3 suggest that for $M_p = 6.5$ and 7.5 M_L determined from strong motion data should be 0.5 and 1.2 magnitude units smaller, respectively, because of the high-pass filtering effect of WAS and because of the saturation of strong motion amplitudes with increasing M_p (Luco,[10] Trifunac[14,16,17]). Combining these results with the effects of frequency dependent attenuation results in further reduction of the expected peak amplitudes for $M \geq 5.5$ as shown by the dashed lines in Fig. 4. It is seen that the amplitudes and the shape of $D(\bar{M}_L^{SM})$ predicted in this way are close to the estimates determined from the averaged differences between $\bar{M}_L^{SM}$ and M_p, over 1/2 magnitude intervals for data in Southern California (Trifunac[17]; Lee[7]).

The saturation of strong motion amplitudes with increasing magnitude has been treated consistently in all of our previous analyses (e.g Trifunac[14,16]) in terms of the 'published' magnitude M_p. $M_p \sim M_L$ for most strong motion data for $M_p \leq 6.5$ and most determinations of M_L in Southern California have resulted in magnitude less than 6.5 (Hutton and Boore[4]). Therefore $D(\bar{M}_L^{SM})$, presented by Lee[7] and Trifunac[17] (see also solid points in Fig. 4), can be used to determine $M_L - \bar{M}_L^{SM}$ for M_L less than 6 to 6.5. For $\sim 6 < M_L < 6.5$, saturation of M_L versus other long period magnitudes may yield some differences between $\bar{M}_L^{SM} - M_L$ and $\bar{M}_L^{SM} - M_p$. However, at present, there is neither enough strong - motion data nor has anyone conducted a careful systematic study of the long period magnitude estimates for all earthquakes in our data base, to make analysis of such differences feasible. The strong motion data base employed by Trifunac[7] and Lee[17] has been recorded since 1933. Changes in instrumentation and in the practices employed in computing various long period and larger (say $M > 6.5$) magnitudes have changed so much during the past 55 years that an investigation of the associated consequences is well beyond the scope of this brief note. Most determinations ($\sim$95 percent, Hutton and Boore[4]) of local magnitude in Southern California are for $M_L < 5.5$ (see Fig. 2) and for that range $\bar{M}_L^{SM} - M_L$ should be equal to $\bar{M}_L^{SM} - M_p$ (Fig. 4).

DISCUSSION AND CONCLUSIONS

The above analysis shows that it is possible to explain the amplitudes of $D(\bar{M}_L^{SM})$ in California, in terms of anelastic attenuation with characteristic distance of about 100 to 200 km and with Q between about 100 and 150. Allowing for the effects of large impulsive peaks in strong ground motion at small epicentral distances, and essentially for all recordings of small magnitude events, would result in estimates of Q between about 150 and 200. This range of Q is not inconsistent with other estimates of shallow Q values in Southern California.

Our interpretation of $D(\bar{M}_L^{SM})$ is dominated by the overall average value of anelastic attenuation for the

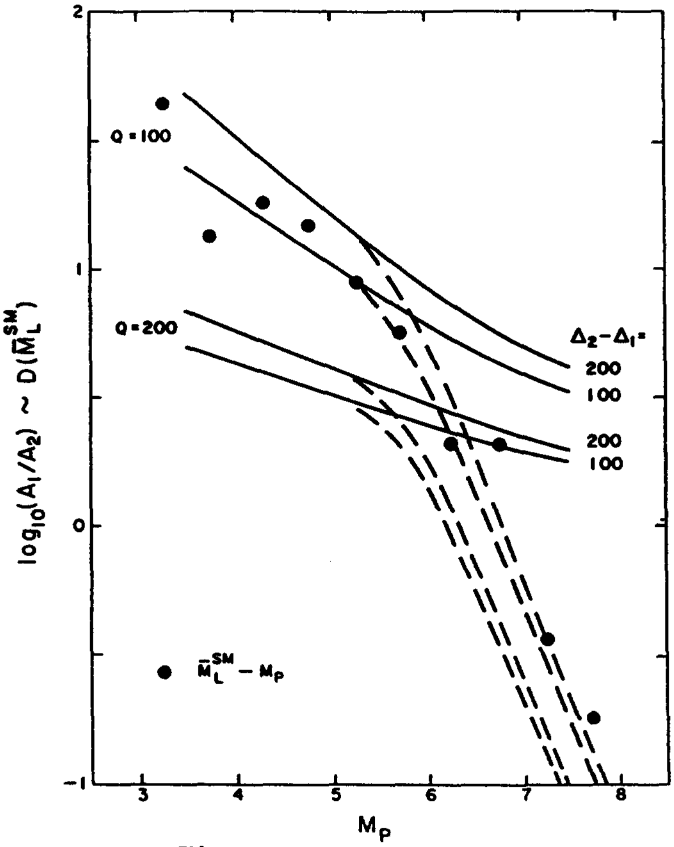

Fig. 4 $D(\bar{M}_L^{SM})$ versus published magnitude, M_p, for $Q = 100$ and 200, and for $\Delta_2 - \Delta_1 = 100$ and 200. Solid lines include only the effects of attenuation modeled in terms of Q. Dashed lines reflect the saturation of M_L^{SM} versus M_p. Solid points represent the available data.

entire region, sampled by the earthquake waves which contribute to the peaks on WAS. It can be seen from Fig. 4 that $D(\bar{M}_L^{SM})$ is quite sensitive to Q values, suggesting that, after all required refinements have been considered to improve the description of anelastic attenuation, the above procedure could be useful as a tool to infer some regional average estimates of Q (provided $Q \leqslant 500$). For $M \geqslant 5.5$, the effect of saturation of strong motion amplitudes relative to M_p results in faster decreasing $D(\bar{M}_L^{SM})$ with respect to M_p. The details of this saturation rate are not precisely known in our scaling laws (e.g. Trifunac[14,16]) and, so, our inferences on the details of the shape of $D(\bar{M}_L^{SM})$ are only approximate. In all our previous work we approximated this saturation by a parabola versus M^p, to some M_{max}, beyond which no growth of strong motion amplitudes has been considered. However our regression analyses have indicated M_{max} in the range typically well above 7.0 which is outside the range where enough strong motion data is currently available, to control well the result of the analyses.

A comparison has been made (Trifunac[17]) of the differences between our attenuation function $Att(\Delta_0)$ and $\log_{10}A_0(R)$ (Richter[12]) for $\Delta < 100$ km. This showed that $Att(\Delta_0)$ is more negative than $\log_{10}A_0(R)$, by ~ 0.2 for $\Delta = 1$ km, by about 0.8 for $10 < \Delta < 20$ km, with this difference gradually decreaing as $\Delta \to 100$ km (~ 0.2 at $\Delta = 50$ km). Similar (0.1 to 0.4 magnitude units) differences can be observed between $\log_{10}A_0$ of Hutton and Boore[4] and $\log_{10}A_0(R)$. In fact for $10 < \Delta < 50$ km and for the source depth between 10 and 20 km, the shapes of $-\log_{10}A_0$ and $Att(\Delta_0)$ are very similar, with $Att(\Delta_0)$ more negative by about 0.1 to 0.3 magnitude unites, relative to $-\log_{10}A_0$. Considering, for example, as some have suggested, that the Wood Anderson seismometers have static magnifications closer to $V_s = 2000$ rather than the nominal value $V_s = 2800$, would further reduce these differences. This suggests that $D(\bar{M}_L^{SM})$ greater than about 0.5 might not be absorbed easily, into a 'new' empirical strong motion 'attenuation' equation.

Assuming that $Att(\Delta_0)$ is representative of strong motion amplitudes and that $D(M_L^{SM})$ depends mostly on Q, taking large Q would reduce $D(M_L^{SM})$ to zero for $\bar{M}_L^{SM}) \leqslant 5.5$. Thus M_L (station) $-M_L$ (earthquake), averged and plotted for numerous earthquakes versus distance should be closer to zero for all stations on basement rock. Hutton and Boore[4] show three examples of M_L (station) $-\mathrm{M}_L$ (earthquake) for Pasadena, Riverside and Palomar stations in Southern California. It is interesting to observe that using our $Att(\Delta_0)$ for $10 < \Delta_0 < 100$ km would reduce these residuals to essentially zero for $10 < \Delta < 100$ km at Palomar and at Pasadena. Palomar station is on and is surrounded by the basement rocks. At Riverside using $Att(\Delta_0)$ would enlarge the residuals for small Δ by as much as 0.3 for $\Delta = 0$ (if one took the average station correction only for the residuals with $\Delta > 200$ km). For a small distance range, the earthquakes contributing to these residuals are typically smaller than for $M_L = 3$, and, so, this would imply $D(\bar{M}_L^{SM}) \sim 0.3 - 0.5$ for $M_p \sim 3$. With increasing distance, the average magnitude contributing to the averages of M_L (station) $-$ M_L (earthquake) comes from progressively larger earthquakes, and at $\Delta \sim 300$ km (for $M_p \sim 5$) the residuals at this station would become zero. It is also interesting that the Riverside station is surrounded by sediments about 10 km wide and 20 km long. Further and more detailed analyses of geologic setting of other Southern California stations and of their residuals M_L (station) $-M_L$ (earthquake), should show to what extent our model might help in explaining the observed trends in station residuals.

A part of the fluctuations in the residuals of M_L (station) $-M_L$ (earthquake) versus epicental distance comes from estimation of attenuation equation via regression of peak responses. Physically, amplitudes of peak response of Wood Anderson Seismometer are closely related to the peak ground velocity and peak ground displacement. As the attenuation of the peak amplitudes with distance is not necessarily a smooth function (e.g. focusing points of wave arrivals from different discontinuities in the crust), it may not easily be described by a polynomial in Δ. Hence, station residuals of M_L (station) $-M_L$ (earthquake) versus Δ for permanent seismological stations may lead to irregular fluctuation which may not be repeatable (distribution of earthquake azimuths, and thus orientation of incident waves, may fluctuate over long periods of time) but should depend on the geology surrounding the stations. M_L (station) $-M_L$ (earthquake) residuals for strong motion stations are subject to even larger fluctuations, which are also caused by the geologic setting of the stations, because these stations are not as 'permanent' and so the number of recordings per station is much smaller.

Analysis of the distribution of the recorded data with magnitude and epicentral distance (Fig. 2) shows the mean distance between strong motion stations and seismological stations recording earthquakes in the range between 3.5 and 4.5 magnitudes is about 150 km. For magnitudes between 4.5 and 5.5 this mean distance increases to about 250 km. For stong motion accelerographs recordings events with magnitudes between 3.5 and 4.5 the typical epicentral distance ranges from 0 to 15 km. For events with magnitudes between 4.5 an 5.5 this distance extends from 15 to about 40 km. Thus, having reached the strong motion station, the seismic waves must propagate additional 100 to 200 km before they are recorded by Wood-Anderson seismometers. Therefore, the attenuation along this propagation path, with considerable energy travelling through shallow surface layers, might be one of the more obvious and plausible causes for the observed $D(\bar{M}_L^{SM})$ amplitudes.

In this note, our interpretation of the physical causes which determine the amplitudes of $D(\bar{M}_L^{SM})$ is neither unique nor does it cover all possible contributing factors. The effects of anelastic attenuation (Q), and of saturation of strong motion amplitudes with M_p, are larger and the more plausible candidates to explain the observed trends. The effect of impulsive peaks in ground displacement is just one example, from a family of many similar and related amplitude trends, which could be considered to understand how different wave types, in different ranges of distance, and along different propagation paths, for example, could influence magnitude determinations. Investigations of such trends would not only require interdisciplinary analysis of large bodies of data, but would also require going back to old recordings and old procedures, if this is feasible. Such analyses are beyond the scope of this brief note. If the reader of this work is convinced that the anelastic attenuation for small M_p, and saturation of strong motion amplitudes for larger M_p, could be the main contributing mechanism in determining $D(\bar{M}_L^{SM})$, and also realizes that future improvement in describing $D(\bar{M}_L^{SM})$ could serve as a tool for the analyses of average regional variation in Q, then the aim of this brief note has been accomplished.

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