BUREAU OF MINERAL RESOURCES, GEOLOGY AND GEOPHYSICS

RECORD

Record 1982/29

PROCEEDINGS OF THE WORKSHOP ON AUSTRALIAN EARTHQUAKE MAGNITUDE SCALES

BMR, CANBERRA, 21 MAY 1982

Compiled

by

D. Denham

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CONTENTS

SUMMARY

The Workshop On Australian earthquake magnitude scales, which was held in Canberra on 21 May 1982 was triggered by two main factors. The first was the problems in South Australia resulting from an effective change in the magnitude scale in the mid-1970s, and the implications of this change in terms of earthquake risk assessment. The second was the wide disparity between magnitudes determined by the Australian National University and the Phillip Institute of Technology, for earthquakes in south-eastern Australia.

This record describes the proceedings of the workshop and contains accounts of the magnitude determination methods used at all the main seismological centres in Australia. It was decided, at the meeting, to set up of a database, using 1980/81 Australian earthquakes, which can be used to derive regional attenuation factors, and also unified duration magnitude scales for the Australian region.

1. A REVIEW OF MAGNITUDE SCALES AND SOME PROBLEMS IN THE AUSTRALIAN REGION

by

D. Denham

INTRODUCTION

This workshop was triggered by two main factors. The first was the problems in South Australia resulting from an effective change in magnitude scale in the mid-1970s (Stewart, 1975) and the implications of this change in terms of earthquake risk and the costs of construction. The second was the wide disparity between ANU (Australian National University) and PIT (Phillip Institute of Technology) derived magnitudes for earthquakes in southeastern Australia.

Magnitudes of earthquakes have a special place in seismology. A reliable, standard measure of the size of an earthquake is an obvious need, not only for classification purposes but also for all kinds of tectonophysical and engineering applications. It is therefore remarkable that, in the history of seismology, magnitude scales were not developed until half a century after the early installations of instrumental equipment.

In fact, even when Charles F. Richter developed the famous Richter scale in 1935, it was rumoured that the main impetus for this work was to keep the newsmen happy.

Perhaps the delay in developing a magnitude scale illustrates the difficulties inherent in the problem and points to the confusion that still exists. Ironically, in Richter's classic book on Elementary Seismology (Richter, 1958) he appropriately heads a major section in his chapter on 'Magnitude, Statistics, Energy' with the word 'CONFUSION'. Båth (1966) summarised the position well when he wrote

> 'Today it is an equally well-known as deplorable fact that several different magnitude scales are in current use, and it is not always easy to know which scale is used. Moreover, there are significant and consistent differences in magnitudes determined

at different centers and supposedly in the same scale . . . There are two major reasons for such discrepancies: definition of magnitude, and its calculation'.

This statement still holds true today, and when one considers that since 1967 an average about one magnitude research paper per week has been published, it must be one of the most studied parameters in seismology.

In spite of these difficulties I believe that we, as seismologists, have an obligation to put our house in order and to provide consistent data for engineers, tectonophysicists, and others who need to know the sizes of earthquakes. So let us first look at the requirements for a magnitude scale and some of the inherent limitations. The essential requirements are that a magnitude scale must be simple to use and fast to apply. The desirable requirements are:

1) A magnitude scale should be uniform from region to region and station to station.
2) A magnitude scale should represent a physical quantity - say, E (energy release), and be applicable to earthquakes at all depths.
3) A magnitude scale should be useful - if it is not useful it should be discarded.

The main limitations are caused because each magnitude is represented by a single number (consequently limited to reconnaissance studies only), and, as the term implies, it can be applied only to an order of magnitude and hence is only a crude indicator.

Let us now examine some of the main magnitude scales and see how they relate to the above criteria.

SURFACE-WAVE MAGNITUDES, Ms

Gutenberg (1945a) developed the M_S magnitude scale based on 20s surface waves. The amplitude-distance factor was evaluated to include the level of the effects of geometrical spreading, absorption, and dispersion, and the level of the scale was adjusted to agree with the Richter magnitude M_L - or so he thought.

Gutenberg's formula was:

$$M_S = \log A + 1.656 \log \Delta^\circ + 1.818 \text{ with A in microns} \qquad (1)$$

This was amended by IASPEI (International Association of Seismology and Physics of the Earth's Interior) at Zurich in 1967 to:

$$M_S = \log\left(\frac{A}{T}\right)_{max} + 1.66 \log\Delta + 3.3 \qquad (2)$$

which is now commonly used.

$(A/T)_{max}$ refers to the maximum horizontal component of surface waves in the period range $T = 20 \pm 3$ s; it is applicable only to shallow earthquakes and only in the distance range 20-160 degrees.

There are four main restrictions involving use of the M_S scale:

1) the scale saturates at about $M_S \sim 8$;
2) we need horizontal long-period instruments to determine M_S and these are not in common use;
3) it cannot be used for earthquake distances less than 20 degrees; and
4) it only applies to shallow-depth earthquakes.

The saturation problem cannot be overcome because of the physical nature of earthquakes, but Marshall & Basham (1973) have attempted to extend the formula down to a distance of one degree and in the period range 10-40 seconds.

Their formula is:

$$M_S = \log A + B(\Delta) + P(T) \qquad (3)$$

where $B(\Delta)$ and $P(T)$ are given in tabular form for different propagation paths and are strongly dependent on the regional lithospheric structure.

Restriction 2) has been overcome by USGS (United States Geological Survey) by using Rayleigh waves instead of Love waves. However, their values for M_S are about 0.2 smaller than the M_S values obtained strictly in accordance with Equation 2.

Some deep earthquakes do not produce surface waves, and therefore Restriction 4) cannot be overcome, but Båth (1977) produced a set of corrections ranging from 0.1 to 60 km to 0.4 at 100 km and greater. Thus it appears that corrections for depth can be made for some earthquakes.

BODY-WAVE MAGNITUDES, m_b

Gutenberg (1945b) introduced the body-wave magnitude m_b based on P, PP, and S waves from shallow focus earthquakes, and calibrated it to agree with M_S. Gutenberg's formula is:

$$m_b = \log\left(\frac{A}{T}\right) + q(\Delta) + 0.1(m_b - 7) + Cr \qquad (4)$$

The correction term $0.1(m_b-7)$ was included to provide the agreement with M_S. However, this was modified by Gutenberg & Richter (1956) to read:

$$m_b = \log\left(\frac{A}{T}\right)_{max} + q(\Delta,h) \qquad (5)$$

This formula was not altered by the 1967 IASPEI Zurich meeting and still applies. The ISC (International Seismological Centre) & USGS values are usually smaller than other agencies because they use only the first three cycles whereas other agencies use the maximum number of cycles in the P-wave group.

In recent years calibration functions for $q(\Delta,h)$ have been extended to include earthquakes out to distances of 170 degrees, and several studies have been made so that an m_b value can be determined for short distances (see Båth, 1981).

The latter procedure is very difficult because $q(\Delta,h)$ values at short distances depend heavily on variations of crustal and upper mantle structure from one region to another. For example, the low velocity layer discussed in the Gutenberg & Richter (1956) study, which causes low amplitudes and large q values around 10 degrees, is not a worldwide phenomenon and certainly does not exist in Western Australia.

Although the m_b scale can be easily used from 10-170 degrees it has two main disadvantages: it cannot be used for distances less than 10 degrees without special regional studies because of lithospheric heterogenity, and, the scale saturates at a low level of about $m_b \sim 6$, because the high frequency parts of the seismogram are used to determine m_b.

LOCAL MAGNITUDES, M_L

The local magnitude scale M_L was developed by Richter (1935) for earthquakes in southern California. However, because it was defined for attenuation rates in southern California, the distance factors he determined are not necessarily appropriate for other parts of the world. In fact, it is essential to develop pertinent formulas for each area and not to simply adopt formulas from other structurally different areas if successful scales are to be developed.

Thus, both Drake (1974) and White (1968) found that with the Richter distance factors, magnitudes increased with distance over the Australian continent. Drake (1974) reduced Richter's factors by 0.1 from 100-249 km, 0.2 from 250-499 km, and 0.3 from 500 km and greater, using the Riverview station to correct for this phenomenon. White (1968) derived the following formula:

$$m_L = 2.07 + \log Up + 1.31 \log \Delta^\circ + 0.08\Delta^\circ + Ci \qquad (6)$$

where Up is the maximum peak-to-peak ground velocity in microns/s. and Ci a station correction.

White's formula was applied to S-wave data on Benioff SPZ seismograms in the South Australian network and was related to M_L by:

$$m_L = 0.7 + 0.71\, M_L \qquad (7)$$

Stewart (1975) developed a new scale (M_N):

$$M_N = 4.85 + \log Ag + 0.0003f\Delta/2.3 + 0.84 \log \Delta - 2.89 \log f + 2.45(\log f)^2 + Ci \qquad (8)$$

where Ag is the ground amplitude in mm at f Hz, at Δkm epicentral distance, and Ci is a station correction.

This formula has the dubious distinction of being the most complicated formula for calculating local magnitudes yet published.

Stewart produced two relations for converting $M_N \rightarrow M_L$. In his 1972 thesis (Stewart, 1972) he obtained the following relation:

$$M_L = 1.33M_N - 0.73 \quad (9)$$

However, this was obtained incorrectly and subsequently revised (Stewart, 1975) to -

$$M_L = 1.05\ M_N \quad (10)$$

This revision caused problems in South Australia because the change in magnitude scale effectively changed the earthquake recurrence relationships and hence the earthquake risk assessments.

The M_L scale is very useful, particularly in earthquake risk studies, because it is based on the S phases, which cause most damage to buildings because of their direction of vibration and frequency range. However, the M_L scale can only be used for events closer than about 8 degrees, and it saturates at about $M_L \sim 7$.

The saturation factor thus applies to all scales so far discussed and causes problems for large earthquakes. The first approach to overcome this difficulty was to use longer period surface waves (100-200s) to quantify large earthquakes. However, the significant breakthrough came only when Kanamori (1978) proposed a new magnitude scale (M_w) based on seismic moment M_o. M_o is based on the physics of the earthquake source and is defined as:

$$M_o = \mu A\overline{D} \quad (11)$$

where μ is the rigidity, A is the surface area displaced, and $\overline{D}$ is the average displacement on that surface.

If we start with the seismic wave energy E, which can be expressed as the difference between the strain energy drop ΔV and the heat loss due to friction (assuming frictional stress σ_f = final stress σ_1, and σ_o = initial stress), then:

$$E = \Delta V - \sigma_f \bar{D}A \qquad (12)$$

$$= \frac{(\sigma_o+\sigma_1)}{2} \bar{D}A - \sigma_1 \bar{D}A \qquad (13)$$

$$= \frac{(\sigma_o-\sigma_1)\bar{D}A}{2} \qquad (14)$$

From (11) and (14) we get:

$$E = \frac{\Delta\sigma}{2\mu} M_o \qquad (15)$$

where $\Delta\sigma = \sigma_o - \sigma_1$

Assuming a constant stress drop of 30 bars (3MPa) and a value for μ of 30 GPa we get:

$$E = \frac{10^{-4}}{2} M_o$$

But, $\log E = 11.8 + 1.5M_s$ (Richter, 1958).

So, by replacing M_s by M_w we get:

$$11.8 + 1.5M_w = \log M_o - 4.3 \qquad (16)$$

So,

$$M_w = \frac{\log M_o}{1.5} - 10.7 \text{ (dyn-cm)} \qquad (17)$$

or

$$M_w = \frac{\log M_o}{1.5} - 6.0 \text{ (N-m)} \qquad (18)$$

This magnitude scale (M_w) can be used for all ranges of magnitude. The only problem is one of calculating M_o. This is usually defined in terms of the low frequency amplitude spectral level Ω_o as:

$$M_o = 4\Pi\rho\beta^3\Omega_o\Delta/0.85$$

where β is the S-wave velocity, and ρ is the density.

We can usefully compare some of the main characteristics of the four magnitude scales by looking at them in terms of the dimensions of the earthquake source - as is shown in the table below.

Magnitude Scale	Period (s)	λ (km)	M_{max}
M_L	0.3-3	0.3-10	~7
M_s	~20	~60	~8
m_b	~1	~10	~6
M_w	10-∞	∞	∞

DURATION MAGNITUDES

It is a characteristic feature of seismic records that each phase is followed by a tail of decaying amplitude, especially pronounced in short period local or regional records of the large amplitude Sg or Lg waves. This phenomenon is probably associated with scattering in the crust.

Emphirically, $\log \left(\frac{A}{T}\right)_{max} \alpha \log \tau$

where τ is the duration of the record.

So we can write

$M_\tau = a\log \tau + b(\log\tau)^2 + c\Delta + dh + e$, where a, b, c, d and e are constants for a given instrument at a given station and h the depth in km. Some 20 or more formulae of this type are now in the literature (Båth, 1981). Once a formula has been established the method is simple to use. It does not depend on the calibration of the seismograph (provided it remains constant) and it is not affected by clipping during large earthquakes. It also has a weak distance dependence term and therefore M_τ should not depend strongly on regional attenuation differences.

The formula breaks down during double earthquakes, and it can only be used for distances less than 10 degrees.

A typical formula is

$$M_L = 2.00 \log \tau + 0.0035\Delta \ -0.87 \qquad (20)$$

which was derived by Lee & others (1972) for California, (τ is in seconds). In this relation the b and d coefficients are put at zero.

One other problem is that the amplitude of the earthquake coda sometimes decays very slowly and the end of the disturbance is difficult to measure. In practice it is common to measure the time from the initial P-wave arrival to the time when the trace amplitude decays to twice the background level before the earthquake.

DISCUSSION

The main methods of determining magnitudes of earthquakes have now been briefly reviewed. Although there are several approaches to the determination of magnitude, I believe that we are in a better position to resolve the problems now, than we were, say, ten years ago. This is because of the great breakthroughs that have been made in our understanding of the processes that take place at the earthquake focus.

We certainly should not be deterred by earlier difficulties because we have a duty to try and establish a uniform scale or set of scales for Australian earthquakes.

2. CURRENT MAGNITUDE DETERMINATION PRACTICE AT THE UNIVERSITY OF ADELAIDE

by

R. McDougall

TELESEISMIC EARTHQUAKES

(a) Log $(A/T)_{max}$ is calculated routinely from the short period vertical ADE (WWSSN) seismograms for subsequent determination of m_b. The maximum amplitude of the entire P-wave train is used, as outlined in the WDC-A Manual of Seismological Observatory Practice (Willmore, 1979), Section OP 3.4.1

(b) M_S is not calculated routinely. When required, it is determined using the 1967 IASPEI recommendation, using amplitudes and periods (18s<T<22s) from the long-period horizontal components of the WWSSN station.

LOCAL EARTHQUAKES

(a) M_N

$$M_N = 4.85 + \text{Log Ag} + 0.84 \log\Delta + 0.0003\Delta/2.3 + 2.89 \log f + 2.45(\log f)^2 + C \qquad \text{(Stewart, 1975)}$$

where Ag = peak-to-peak maximum vertical ground amplitude in the S-wave train (in mm).

f = corresponding wave frequency (in Hz)

Δ = epicentral distance (in km).

c = station correction.

M_N is valid over the range $0<\Delta<600$ km. The reported value of M_N is the mean of M_N from each station recording the event. The standard deviation in M_N is ~0.2 per event.

(b) M_D

$$M_D = a_o + a_1 \log\tau + a_2\Delta \qquad \text{(Parham, 1981)}$$

where a_o, a_1, a_2 are station coefficients

τ = trace duration from initial onset to time when signal amplitude becomes equal to that of background noise (in s).

Δ = epicentral distance (in km)

M_D is correlated to M_N such that on average $M_D=M_N$. Hence, M_D is also limited to the range $0<\Delta<600$ km although the small distance dependence of M_D ($a_2 \sim 5 \times 10^{-4}$) implies that it may be able to be extended beyond this range. The reported value of M_D is the mean of M_D from each station recording the event. The standard deviation in M_D is ~ 0.1 per event.

(c) M_L

$M_L = 1.33\ M_N - 0.73$ (Stewart, 1972)

Until recently the above relationship was used to convert M_N (and M_D) to Richter magnitude. The more recent conversion proposed by Stewart is:

$M_L \approx 1.05\ M_N$ (Stewart, 1975)

This does not appear to offer any significant improvement over the original, albeit incorrectly derived, formula.

3. CURRENT STATION CALIBRATION PRACTICE AT THE UNIVERSITY OF ADELAIDE

by

R. McDougall

MOTOR CONSTANT DETERMINATION

The motor constant (G) of the seismometer calibration coil is determined using a weight lift/current pulse method.

$$G = \frac{9.8\ m\ x_i}{i\ x_m}$$

where m = mass lifted (in g).
i = current pulsed (in mA)
x_i = pen deflection due to current pulse (in mm).
x_m = pen deflection due to mass lift (in mm).

FREQUENCY RESPONSE DETERMINATION

Stations are calibrated using a whole system, constant current, varying frequency sinusoidal input to the seismometer calibration coil, the theory of which is outlined in the WDC-A Manual of Seismological Observatory Practice, Section INST 4.4.2.

At each frequency used in the calibration, the system's velocity sensitivity and gain are calculated.

A sixth-order polynomial of log(gain) vs. log(frequency) over the range [1 to 10] Hz is fitted to the resulting gain values for use in determining absolute ground amplitudes for M_N.

4. TO SCALE MAGNITUDES FROM THE UNIVERSITY OF TASMANIA NET

by

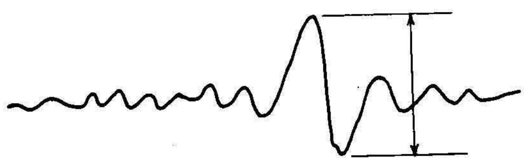
R.G. Underwood

GENERAL

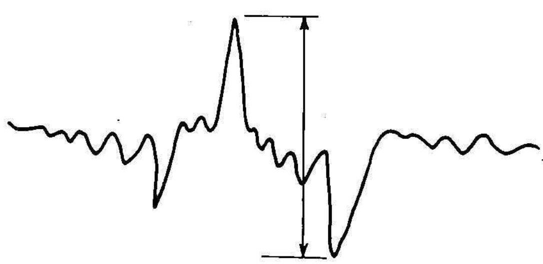
Magnitudes were first defined from Wood Anderson (W-A) horizontals in standard adjustment. We have to try to convert our peak scalings into what a vertical W-A would have recorded, then enter the nomogram that corrects for distance, and so get Magnitude Equivalent Wood-Anderson for Net Verticals (MEWANV).

SCALING

1. Find the peak oscillation on the seismogram. This will need to be



and not just the distance from the highest peak to the lowest, i.e. not



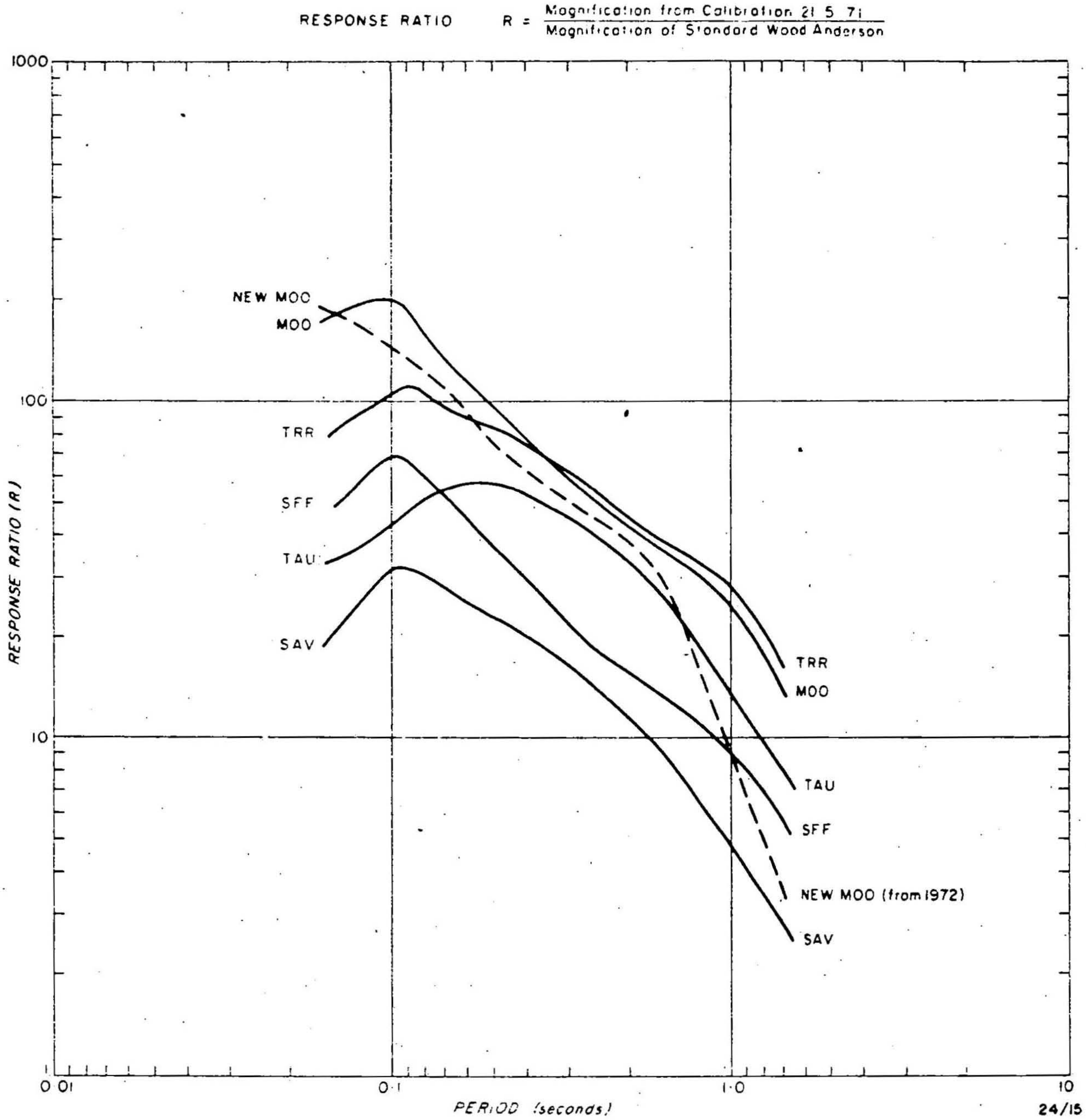


Fig. 1 Response ratio for University of Tasmania network

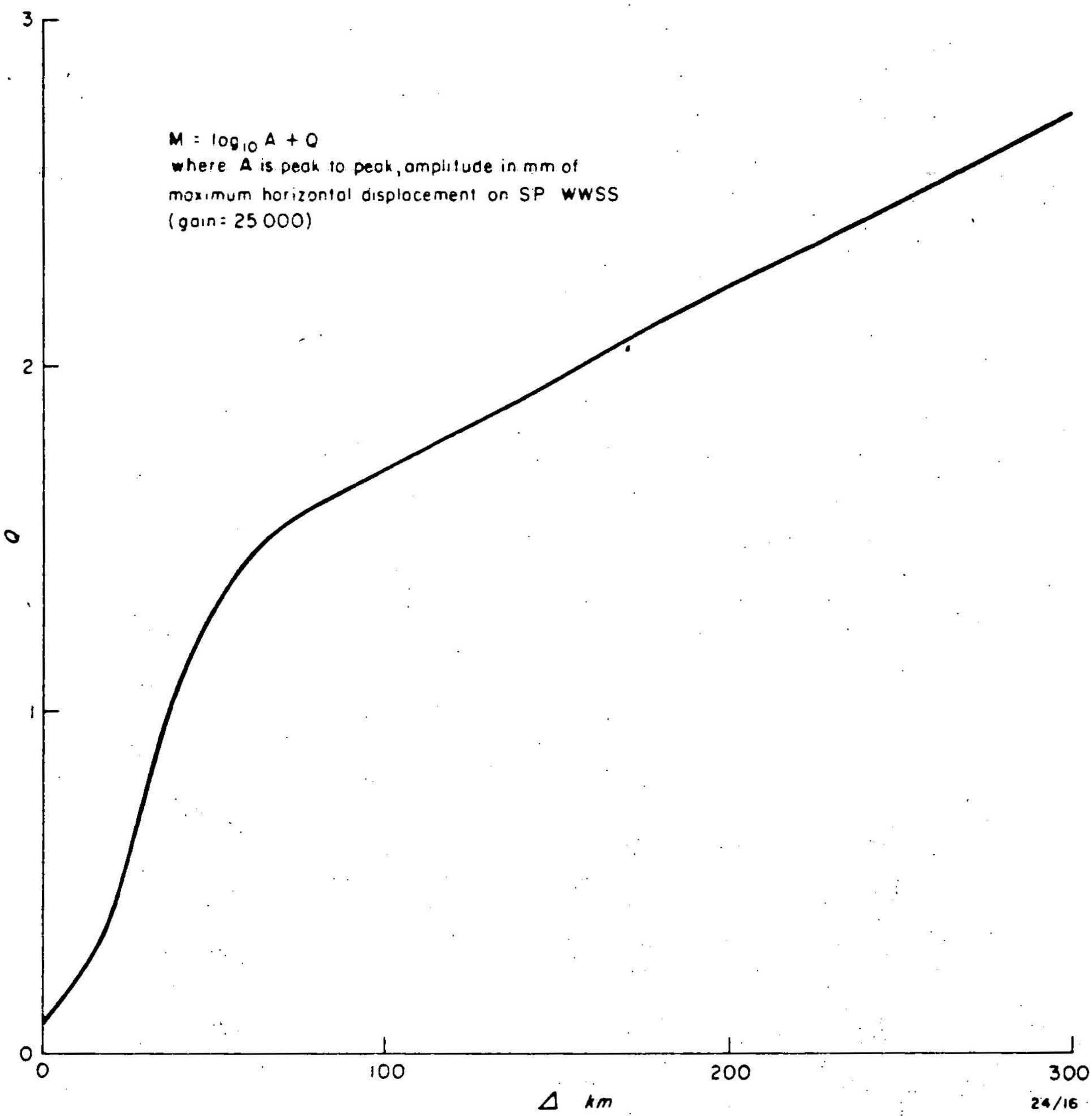


Fig.2 Magnitude calculation from TAU (WWSS)

Scale its peak to mean amplitude in millimetres, and scale its period. Write these on the card.

2. From the period, get the Response Ratio, R, from the graph (Fig. 1).

3. From the log, find the pen-amplifier gain, P, in volts/cm.

4. Compute the peak-mean amplitude that a W-A would have shown -

$$B_{(mm)} = \frac{\text{Amplitude (peak-mean in mm)}}{R} \times P$$

5. Check to see whether some longer period but less high peak on the seismogram might not give a bigger value of B by having a lower value of R. Take the biggest value of B.

6. Determine the epicentral distance, preferably from the adopted location, but even from S-P.

7. Use the nomogram* given by Gutenberg & Richter (1942) with B and Δ, and find M for each station. Write it on the card.

8. Take an average of the most consistent ot the estimates. Call it MEWANV.

COMMENTS

1. Watch out for alterations in instruments or communications; these will probably alter the R values. Ask for a recalibration.

2. Scaling the daily calibration pulses should be a good way to keep track of the calibrations.

3. To get an R curve from a new calibration, use

$$R = \frac{\text{Magnification from calibration}}{\text{Magnification of standard W-A}}$$

$$= \frac{\text{Velocity sensitivity from calibration}}{\text{Velocity sensitivity of standard W-A}}$$

---

*If period is to difficult, assume period of 0.3 seconds which implies the following R values -

TAU	48	TRR	65	MOO	65	STG	21
SFF	65	SVR	11	SAV	26	SPK	12 1/2

4. The W-A values are:

f Hz	T (s)	Mag	V (s^{-1})
0	D.C	0	0
0.5	2	424	135
0.7	1.43	776	177
0.8	1.25	965	192
0.9	1.11	1160	205
1.0	1.0	1340	214
1.1	0.91	1520	220
1.2	0.83	1680	220
1.25	0.8	1755	224
1.5	0.67	2040	216
2.0	0.5	2380	190
4.0	0.25	2710	108
10.0	0.1	2790	44 1/2
∞	0	2800	0

5. MEWANV is only an experiment - don't despair if they don't come out reasonable, and don't rely on them.

<u>MAGNITUDE FROM WWSS</u>

1. Scale A, peak-to-peak amplitude (in mm) of largest excursion on record of each of SPE and SPN.

2. Using Δ distance, find Q from graph (Fig. 2).

3. $M = \log_{10} A + Q$.

4. Q is derived from Table 22-1 in Richter (1958) and incorporates gain, and a factor of 2 for peak-mean.

5. This scale is used only when network verticals are unavailable (e.g. clipped).

5. SCALING FACTORS FOR ANU NET

by

K.J. Muirhead

The methods described by Underwood in this Record (Section 4) outline the methodology used to at both ANU and the University of Tasmania. The table below gives the scaling factors employed at the ANU as of July 1982.

To get the scaling factor, the gain of the station is divided by 2800. However, since trace amplitudes are given in 1/2 mm, the figure obtained for the scaling factor should be doubled.

e.g. Gain of station = 7×10^4
Relative to W-A = 25
Scaling factor = 50

$\therefore$ Scale reading of 50 is equivalent to 1 mm on a W-A.

STATION	CAL (UNITS)	CM	GAIN	÷2800	SCALE FACTOR	COMMENTS
CAN	24 now 46	1.20	98 000	35	70	
KHA	51 now 24	2.5	205 000	73	146	
TAO	12	0.6	49 000	17.5	35	
DRT	12	0.6	49 000	17.5	35	
SBR	23	1.15	94 000	34	68	
CBR +			74 000	26	53	calc from sinusoid
WAM +			74 000	26	53	from cal- box
MEG	23	1.15	94 000	34	68	
IVN	19	0.95	78 000	28	56	after 12/5/79
LER	14	0.7	57 000	20	41	
BWA +			140 000	50	100	NB after 13/4/79
YOU +			400 000*	142	284	after 7/6/79
CAH	15	0.75	61 000	22	44	

* Signal level is lower since this check. + No calibration signal

6. CANBERRA OBSERVATORY GROUP : CURRENT SEISMOLOGICAL PRACTICE

by

R.S. Smith

SEISMOGRAPH CALIBRATIONS

1. All seismographs (except TOO long period seismographs) are fitted with electromagnetic calibration coils - those for the Willmore Mk II's were made in BMR workshops.

2. All stations are calibrated about annually with a function generator, and the calibrator motor constants (G) checked by performing lift tests. We have investigated variations (up to ±15%) in G determinations and found them to be caused by excessive displacements between calibration coils and magnets.

3. Calibration pulses are applied daily (automatically or manually) from constant-current D.C. sources, to monitor for changes in system gains.

4. In a few cases, magnifications have been unknown, or those adopted have been in error by up to a factor of two, for extended intervals. More effort needs to be spent on past and current documentation to avoid such errors.

MAGNITUDE DETERMINATIONS

BMR practice has been described by McGregor & Ripper (1976)

1. Local Magnitude, M_L

(a) A magnitude is determined for earthquakes at distances up to about 1500 km using:

$$M_L = \log\left(\frac{A}{R}\right) - \log A_o$$

or $$M_L = \log A - \log A_o - \log R$$

where A = mean-to-peak trace amplitude (in mm) for S-wave.

A_o = the Richter attenuation factors as extended by Eiby & Muir (1961)

R_1 = the ratio between the seismograph magnification and the standard (V_o = 2800) W-A magnification at the S-wave period.

(b) The mean of the amplitudes on the horizontal components is used at TOO, ASP, and MTN, and the vertical component is used at other stations.

(c) The magnitudes are sometimes designated M_L'(SH) or M_L'(SV) but more usually just as M_L.

(d) We find that magnitudes determined at greater distances (>300 km) tend to be higher (by about 0.3) than those at short distances (<300 km), and that M_L(BMR) < M_L(ANU), generally by about 0.3.

(e) We have observed that the horizontal amplitudes of S-waves are sometimes up to 1.5 or 2.0 times the vertical amplitudes but have not carefully investigated this effect.

2. <u>Body-Wave Magnitude, m_b</u>

(a) $m_b = \log (A/T) + Q$

where

A = maximum (mean-to-peak) ground amplitude in microns

T = the period in seconds

Q = the attenuation factors of Gutenberg & Richter (1956).

The maximum double-trace amplitude of the P-wave envelope is scaled provided it occurs in the first six seconds. Only well defined wave packets of sinusoidal character are considered. Period (T) and amplitude (A) are reported to the USGS and Log (A/T) to the ISC.

3. Surface-Wave Magnitudes, M_S

(a) No routine reports are made.

(b) Long-period components are in operation only at TOO, ASP, and RIV.

(c) M_S is usually calculated only for inclusion in special investigations and published papers where the methods employed are explained (usually the IASPEI, or Marshall & Basham (1973) formulae).

4. Each station estimate is included in phase lists, and the mean of all estimates is listed in the Earthquake Data File.

5. Conversions of trace to ground amplitudes are made by computer using a table of seismograph magnifications.

6. A computer routine for calculation of M_L values from amplitude and period readings was developed but is not in routine use.

7. MUNDARING GEOPHYSICAL OBSERVATORY DETERMINATION OF MAGNITUDES

by

P.J. Gregson

SURFACE-WAVE MAGNITUDE, M_S

(a) The procedure used at Mundaring is the IASPEI recommendation and is as described in McGregor & Ripper (1976, p5).

(b) $M_S = \log\left(\frac{A}{T}\right) + 1.66 \log \Delta + 3.3$

where

A = maximum combined horizontal mean-to-peak amplitude in microns (10^{-6}m) of surface waves with period near 20s.

T = the wave period (seconds) and is $17<T<22$.

(c) The M_s magnitude is not used routinely and is normally only determined from MUN and NWAO seismograms. KNA horizontals can be used where 20-second surface waves are sometimes recorded.

BODY-WAVE MAGNITUDE, m_b

(a) The procedure used at Mundaring is as described in McGregor & Ripper (1976, p7).

(b) $m_b = \log A/T + Q$

where

A = maximum mean-to-peak ground amplitude in microns of the P-wave train.

Q = depth/distance factor. For shallow earthquakes, TABLE 3.2 in the ISC Manual of Seismological Observatory Practice is used (Willmore, 1979).

(c) The m_b magnitude is not used routinely.

LOCAL MAGNITUDES

1. Richter magnitude, M_L

(a) The true Richter magnitude is determined from using the Mundaring W-A seismographs. See McGregor & Ripper (1976, p2).

(b) $M_L = \log A - \log A_o$

where

A = the maximum trace amplitude (in mm) measured from mean-to-peak on the seismogram (The arithmetic mean of the two horizontal components is used).
and A_o the Richter distance factors (Richter, 1958).

(c) The nonogram (after C.F. Richter 1935) reproduced in the Manual of Seismological Observatory Practice (Willmore, 1979) is used to determine M_L.

(d) 0.14 is added to the value determined from the nonogram as the magnification of the Mundaring W-A is 2040 (not the standard 2800).

(e) The Mundaring W-A can be used for earthquakes in the southwest seismic zone in the range of M_L = 3 to 5.

2. WWSS (MUN), M_L

(a) The Richter formula is extended to determine magnitudes from the mean-to-peak S-wave trace amplitudes on the WWSS horizontal components.

(b) Likewise, in the vicinity of Kununurra, the KNA horizontals are used.

(c) $M_L = \log A - \log A_o - \log R$

where R is the ratio between the seismograph magnification and the standard W-A seismograph at the period of ground movement.

3. Magnitude from SV waves

(a) There are numerous earthquakes in the southwest seismic zone (SW Western Australia) with Richter magnitudes less than 3 which are recorded on the station network of vertical seismographs (i.e. MUN HG-Z, KLB, BAL, NWAO, and WLP).

(b) A magnitude is derived using:

Magnitude $= \log A - \log A_o - \log R$

where

A = the mean-to-peak amplitude of the S wave on the vertical component.

A_o = the standard Richter attenuation factor.

R = ratio between the seismograph magnification and the standard W-A magnification at the S-wave period.

(c) The magnitude used is the mean of the values determined from each station.

(d) Erroneously, this magnitude has been quoted as M_L. It does however, approximate M_L when determined at stations greater than 200 km.

(e) Using this method, we have found that magnitudes at close stations are low compared with magnitudes at the more distant stations. We are working on a relationship for distance/correction and have very preliminary results (see Fig. 3).

(f) SV waves are also used for Western Australian regionals which are recorded on vertical seismographs at KLG, MEK, NAU, MBL, KNA, WBN, and MUN but are not recorded on horizontal components.

DURATION MAGNITUDE

Very preliminary work has been done on relating the duration of recordings at seismograph stations with magnitudes of earthquakes in the southwest of Western Australia. The magnitude determined (M) approximates the Richter magnitude. The method for determining M is shown below

Method: First Approximation

Area - Southwest seismic zone

Data used - Earthquakes 17 km S Cadoux
Jan 1 to Feb 10, 1982
Magnitude range ML: 3 to 4.9

Station constants: KLB - 4.1; NWAO - 5.0; MUN HG-Z - 4.7

$M_T = 2.1 \log t + \log A - \log A_o + C$

where

M_T = duration magnitude approximating M_L

t = time in seconds between P-wave arrival and trace amplitude to drop consistently below amplitude A

A = trace amplitude (centre-peak) in mm

A_o = Richter standard amplitude for distance *

C = station constant

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*Since A_o is less than 1, its logarithm is negative, and the table shows values for $- \log A_o$.

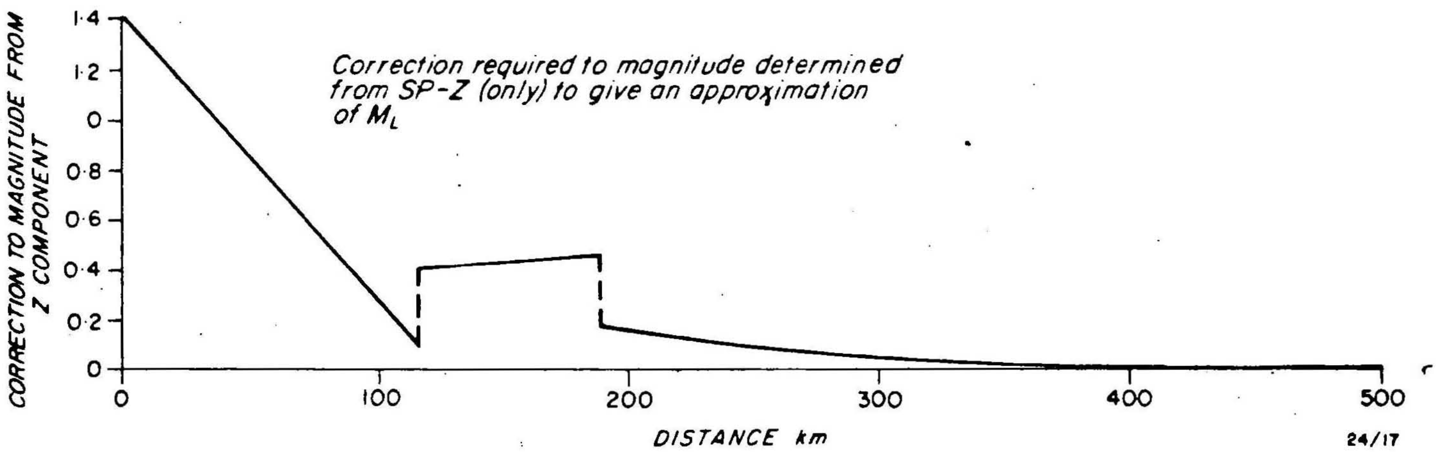

Fig.3 Magnitude correction for earthquakes in southwest Western Australia

The table below shows how - log A_o depends on the distance v.

Distance v - log A_o

v (km)	$-\log A_o$	v (km)	$-\log A_o$	v (km)	$-\log A_o$
0	1.4	150	3.3	390	4.4
5	1.4	160	3.3	400	4.5
10	1.5	170	3.4	410	4.5
15	1.6	180	3.4	420	4.5
20	1.7	190	3.5	430	4.6
25	1.9	200	3.5	440	4.6
30	2.1	210	3.6	450	4.6
35	2.3	220	3.65	460	4.6
40	2.4	230	3.7	470	4.7
45	2.5	240	3.7	480	4.7
50	2.6	250	3.8	490	4.7
55	2.7	260	3.8	500	4.7
60	2.8	270	3.9	510	4.8
65	2.8	280	3.9	520	4.8
70	2.8	290	4.0	530	4.8
80	2.9	300	4.0	540	4.8
85	2.9	310	4.1	550	4.8
90	3.0	320	4.1	560	4.9
95	3.0	330	4.2	570	4.9
100	3.0	340	4.2	580	4.9
110	3.1	350	4.3	590	4.9
120	3.1	360	4.3	600	4.9
130	3.2	370	4.3		
140	3.2	380	4.4		

MAJOR EARTHQUAKES

We have had problems in the past in determining the magnitude of large earthquakes at close stations. For example, the M_L = 6.2 earthquake at Cadoux on 2 June 1979 saturated all seismographs in Western Australia (including the SRO) except KNA.

Since then we have installed a very insensitive short-period vertical seismograph in the Mundaring Office with a magnification of about 80.

SEISMOGRAPH CALIBRATIONS

All Western Australian seismographs were calibrated at installation. Calibrations are checked when the opportunity arises during service visits and when an appropriate interval of time has elapsed since the previous check. The calibration procedure can be completed in about an hour.

8. CURRENT MAGNITUDE DETERMINATION PRACTICE AT PIT

by

Gary Gibson

The PIT seismograph network in Victoria covers a relatively small area, and earthquake-to-seismograph distances rarely exceed a few hundred kilometres. Most of the earthquakes being located are quite small, and only about one event per year exceeds magnitude 4.0.

It is assumed that magnitude is a numerical value used to describe the 'size' of an earthquake. Earthquakes are very complex events, and size may be measured in a number of ways. Different methods yield magnitude values which do not correlate precisely, and which may be best related using non-linear functions.

It is impossible to define a simple magnitude scale that takes into account all aspects of earthquake size, such as rupture area, fault displacement, and stress drop. However, at the present stage of seismological observation, it is difficult to envisage a standardised method for earthquake size specification other than magnitude. Magnitude values may be correlated with total energy release, peak power, earthquake moment or other physical quantities.

It is known that attenuation of seismic wave amplitudes with distance varies considerably over the earth. Magnitude should be defined so that it is a property of the earthquake source, and not a measure of ground motion at an arbitrary distance. It is essential that engineers and others involved in seismic risk studies are aware that variations in attenuation occur, and that magnitudes alone are insufficient to estimate actual ground motion.

Magnitudes should be defined so that values conform as nearly as possible to those that would be obtained using the original Richter definition, corrected for local attenuation. This was based on the use of a particular type of instrument recording earthquakes in a particular area (California), but is as good as any other basis for numerical values, and is in common use.

Because teleseismic earthquakes are not normally studied at PIT, m_b or M_s magnitudes are not routinely calculated. The magnitudes computed are Richter M_L values calculated as described below, or other scales defined to give numerical values as near as possible to the Richter M_L values.

PEAK GROUND MOTION MAGNITUDE, M_L

M_L values are obtained from measurements of peak seismogram amplitude and the frequency at this peak, and at PIT are computed by the routine earthquake location program.

The response of each seismograph is parametrically defined by transducer otuput, damping, and natural frequency, and the recording system gain and filter orders and frequencies. This information is stored on a file which includes any changes that have been made to the seismographs, and is automatically accessed using the earthquake date. A calibration pulse at the start and end of each record is used to detect changes in system response, but these are rare.

The measured zero to peak amplitude is in millimetres from an analogue record, or in counts from a digital record. This is from the larger of P or S-wave motion. It is usually from the vertical motion, but for triaxial instruments, it is from that component which gives the highest peak value. The frequency corresponding to this peak is measured or estimated.

The ground displacement corresponding to this is then computed. It is assumed that this is the displacement that would give peak response on the standard Wood-Anderson (W-A) seismograph as used in the Richter M_L definition (gain 2800, period 0.8 sec, and damping 0.8), and this response in millimetres and its logarithm are computed.

The Richter attenuation factor for California is computed using the earthquake hypocentral distance. This is used instead of epicentral distance with an assumed earthquake depth of 16 kilometres in order to allow for deep earthquakes, or shallow nearby earthquakes. A function is used to approximate the Richter tabular values. This is based on that given by McGregor & Ripper (1976), but corrected for use with hypocentral distance. Extrapolation of the function for distances less than 16 kilometres and greater than 600 kilometres is under investigation.

For earthquakes in central Victoria, no significant variation of M_L with distance has been found, either using PIT or BMR seismograms. This suggests that attenuation in this area is similar to that in California. Attenuation is probably higher in the southwest of the State, and lower in the north-east. We are investigating the use of alternative peak ground motion attenuation functions for these areas.

Peak ground motion magnitudes have an inherent azimuthal variation due to the seismic wave propagation pattern, and the earthquake M_L is taken to be the average from all seismograph M_L values. This will lead to systematic errors when an earthquake is outside the network, where the earthquake will have a similar azimuth to each seismograph. Site corrections are currently not applied.

DURATION MAGNITUDE, M_D

The duration in seconds from the first motion of an earthquake to when the recorded value drops to double its level prior to that first motion is taken to be a measure of earthquake size. For each seismograph, an empirical relation is obtained using non-linear least squares which relates duration (D) and hypocentral distance (R) to the earthquake M_L value obtained as described above. This is of the form:

$$M_L = A_o + A_1 (\log_{10}(D))^{A_2} + A_3 R$$

When the parameters A_o to A_3 have been determined for each seismograph, duration magnitude M_D is found by:

$$M_D = A_o + A_1 (\log_{10}(D))^{A_2} + A_3 R$$

The duration magnitude for an earthquake is the average of the values from all of the seismographs used. Duration magnitudes are only valid over the range of magnitudes used in the derivation of the parameters, currently from 0.5 to 3.5. This is particularly important when A_2 values vary much from about 1.0.

PIT duration magnitudes are currently based on work done by Cuthbertson (1977). Considerable data have been accumulated since then, and parameters will soon be revised.

Some seismographs give considerably longer durations than others. In Victoria, these correspond to magnitude variations of perhaps 0.5 for a given duration, but it seems that a common set of parameter values together with a site correction may be feasible within this area. Recent application to seismograms from New Guinea has shown that typical Victorian parameter values yield duration magnitudes that are up to 2.0 or 3.0 smaller than ML magnitudes, clearly illustrating the need for duration magnitude calibration at each seismograph. Duration magnitudes have the following advantages:

1. The variation of duration with hypocentral distance is quite low, and reasonable estimates of magnitude can be made in seconds using duration alone, with reference to a graph or table. The distance term may be related to S-P interval for rapid evaluation.

2. MD can be calculated even when strong ground motion causes the recorder to clip amplitudes.

3. It is not necessary to calibrate the response of a seismograph in order to determine duration magnitude parameters, provided that ML values are known from other seismographs in the area.

4. There appears to be little or no azimuthal variation, so MD values for a particular earthquake seem to vary less than ML values.

5. Digital seismographs may easily maintain an average signal level using a particular time constant. This may be related to either the dominant period, or transducer natural period. At present, PIT digital recorders use a time constant of about 1.5 seconds, but longer values would be required for more distant earthquakes. The finish of the event may then be automatically determined, and if

distance is neglected, a magnitude estimate made.

It must be emphasised that the duration magnitude function is only as good as the data used in determination of the parameters, particularly the accuracy and range of ML values used.

PERCEPTIBILITY MAGNITUDE, M_p

The best available measure of the size of an historic earthquake is the area over which it was felt. Maximum intensity is a poor indicator of earthquake size as it depends strongly on the earthquake depth, and this is usually not known. Following the work of McCue (1980), we define the radius of perceptibility of an earthquake to be the radius of a circle equal in area to that enclosed by the Modified Mercalli intensity III isoseismal.

For a given area, a perceptibility magnitude scale may be defined as follows. For those felt earthquakes which have computed M_L values, the radius of perceptibility (Rp) and depth (Z) are determined. Parameters B_1 and B_2 for the following function (or a similar function) are determined. Note that Rp is usually much greater than Z, which may often be disregarded.

$$M_L = B_1 \log_e(\sqrt{Rp^2 + Z^2}) + B_2$$

Then, for historic earthquakes, the radius of perceptibility may be estimated, depth may be estimated using area of maximum intensity relative to area of perceptibility or assumed, and M_p calculated using -

$$M_p = B_1 \log_e(\sqrt{Rp^2 + Z^2}) + B_2$$

Current estimates for parameter values in Victoria, for distances in kilometres, are B_1 = 1.01 and B_2 = 0.43.

9. CURRENT MAGNITUDE DETERMINATION PRACTICE AT THE UNIVERSITY OF QUEENSLAND

by

J.P. Webb

TELESEISMIC EARTHQUAKES (OBSERVATORY)

(a) Routinely, only m_b is computed from the CTA short-period vertical WWNSS seismograms. Earlier practice in selecting maximum amplitudes of the first three cycles followed USERL recommendations, but maximum of entire P-wave train is now used to calculate A. Values are derived from 1967 IASPEI formula with no station correction applied.

(b) For some large, shallow earthquakes ($\Delta > 25°$) M_s is computed using the 1967 IASPEI recommendations and data from long-period WWNSS instruments ($T = 20 \pm 2s$).

REGIONAL AND LOCAL EARTHQUAKES (RESEARCH)

Regional ($\sim 4° < \Delta < 10°$): m_b and M_s as above

$$M_L = \log A - \mathrm{Log}\, A_o - \log\left(\frac{\mathrm{MAG}\ (\mathrm{CTA}_T)}{\mathrm{MAG}\ (\mathrm{WA}_T)}\right)$$

(BMR - McGregor & Ripper, 1976)

Local ($\Delta < \sim 4°$):

$$M_L = \log A - \log A_o - \log\left(\frac{\mathrm{MAG}\ (\mathrm{CTA}_T)}{\mathrm{MAG}\ (\mathrm{WA}_T)}\right) \quad \text{(BMR)}$$

$$M_D = -1.06 + 1.04\ (\log T)^{1.612} + 0.003\ D$$

(PIT - G. Gibson, pers. comm., 1980)

Wivenhoe Dam Project (essentially 'Local'):

Wivenhoe network - MEQ system fully calibrated

$$M_L = \log A - \log A_o - \log \left(\frac{MAG\ (WDN_T)}{MAG\ (WA_T)} \right) \quad (BMR)$$

$$M_D = -1.06 + 1.04\ (\log T)^{1.612} + 0.003D \quad (PIT)$$

$$M_L = \log A = \log \left(\frac{G_{WA}}{G_{20}} \right) - \log A_o + \log A_{20} + STAC$$

(Brune & Allen, 1967)

At the present time, if earthquakes are large enough (say magnitudes <~3) then M_L (BMR) is used. For microearthquakes, M_D (PIT) is used. The final value for the magnitude of an event is the average of either M_L (BMR) or M_D (PIT) for all stations.

Analyses are in progress to determine:

(a) our own local M_D formula;

(b) the station corrections for the M_L (Brune & Allen, 1967) formula;

(c) some relationship to convert all magnitude scales used back to M_L.

Historic Earthquakes

These are defined as past events that occurred before instrumentation (pre-1909) and/or events that were recorded at only one station, viz. RIV (1909 to about 1940).

$$M(Rp) = 1.01 \ln Rp + 0.13 \qquad \text{(McCue, 1980)}$$

based on isoseismal maps; Rp is radius of perceptibility enclosing the MM = III isoseismal.

10. RECOMMENDATIONS

Participants at the Workshop decided to set up a database using 1980/81 earthquakes that occurred within the continent of Australia and which were recorded at regional stations.

The format for the data acquisition process is given below. Only earthquakes closer than 1000 km to a station are to be studied.

Column	Data
1-10	Data and origin time of the earthquake in years (80 or 81), months, days, hours, and minutes.
11-15	Station code.
16-20	Epicentral distance in kilometres (optional).
21-25	Hypocentral distance in kilometres - this will usually be the same as the epicentral distance.
25-30	P-time in minutes and seconds (optional) - when only the S or Lg phase is visible, the P-time can be calculated and inserted here so that an M_D can be calculated.
31-35	End time in minutes and seconds (optional) - if the minutes are close to hours, assume that there are 100 minutes in one hour to avoid awkward conversions.
36-40	Duration in seconds from the P arrival-time to the time when the trace amplitude returns to twice the background noise level before the earthquake.
41-55	These columns deal with the S-wave maximum and are optional, most agencies will complete only the trace maximum columns (56-70).
41-44	Period of the S-wave maximum in seconds.
45	Component (Z, N, or E).
46-50	Maximum S-wave peak-to-peak trace amplitude in mm.
51-55	Half maximum S-wave peak-to-peak ground motion in nanometres (10^{-9}m).
56-70	Maximum peak-to-peak trace amplitude in mm, nm etc as per the S-wave split up in columns 51-55 and 46-50.

71-80 Comments

The database can hopefully be used to obtain regional M_D scales, and attenuation factors for M_L determinations.

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