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Teleseismic Magnitudes

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TELESEISMIC MAGNITUDES - A Review

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INTRODUCTION

Generally speaking, all modern science is essentially empirical and final tests are made through observations and their interpretations. We see this very clearly also in seismology where research without observational support is practically nonexistent and where new directions of research have often been stimulated by observational data. In many fields of science, rather complicated instrumentation is prerequisite even to be able to recognize the studied phenomenon. This is not the case of seismology. Earthquakes were always felt, recognized and respected as manifestation of God's displeasure (some time ago) or as release of strain energy accumulated in the upper part of the Earth (more recently). On the other hand, the object of our studies, the earthquake itself, usually takes place at inaccessible depth, say, down to 800 km beneath the Earth surface, which makes direct observations impossible. Likewise, it is impossible to bring the earthquake into the laboratory for detailed investigation. We are simply left with the option to carry out our observations on the Earth surface (or in mines, bore holes, etc), often far away from the earthquake hypocentre, and to interpret the measured data (seismograms) in terms of the properties of the source (earthquake), the propagation path and the receiving instrumentation. It is obvious that the character of seismological data (indirect observations) requires special interpretation techniques and influences the accuracy of results achieved.

First questions usually asked about an earthquake are its location, time of occurrence and size. For historic, i.e. pre-instrumental events, these, so-called source parameters, can only be estimated from intensity data. Macroseismic data are to some degree subjective and can not be collected for earthquakes occurring beneath the oceans or in unpopulated areas difficult to reach. With the current global network of seismographic stations, comprising several thousand units, the time of occurrence and the epicentral location of a globally recorded earthquake are usually determined with an amazing accuracy of the order of a second and of less than ten kilometers, respectively. Estimates of focal depths, made by standard travel-time routines, still remain a problem and provide accuracies of the order of 10 km.

Magnitude determinations on a global scale are available only for the larger events. In many cases, it is obvious that one and the same earthquake has been assigned different magnitudes in different bulletins and/or catalogs. Also quite usual is the case where various magnitudes are listed for the same event even in the same catalog. To explain these mishaps and to brush up the catalogs is not always easy. An important role certainly play the different magnitude scales and often different methods applied in a rather confused way in magnitude estimations. In any case, by making use of such catalogs, it is difficult, if not impossible, to discuss the spatial and temporal pattern of seismicity on a uniform basis.

Recent advances in seismometry (broad-band seismometers) and access to fast and powerful computers make it possible to introduce a new source-size parameter, namely the seismic moment scalar, M_0. This parameter defined as a product of rigidity, faulted area and average slip, $M_0=\mu AD$, exhibits a number of advantages when compared with the concept of magnitude. Since M_0 can nowadays also be determined from seismograms in a routine manner, it is likely that in several years, seismic moment will replace the earthquake magnitude. For the time being, however, magnitude is still a popular measure of earthquake size. It does not have a physical meaning, nevertheless it is emazing that a single dimensionless number, derived from simple seismogram measurements, can represent the overall size of a wide range of earthquakes, from rather weak events to disastrous shocks, well in agreement with Charles Richter's "It is interesting that it works at all". Magnitudes can easily be determined from available seismograms even at simple observational sites with no access to computer facilities. In this context, also important is the fact that the routine seismic moment determinations entered the scene first in the late 1970's. This means that about three quarters of the larger earthquakes recorded during this century lack the moment estimates. Thus, for the time being, magnitude is the only objective instrumental measure to compare the size of historical earthquakes with more recent ones.

In the present notes, I shall address the issue of magnitude, in particular definitions of various magnitude scales and techniques of magnitude estimations for teleseismic events. I shall also discuss the magnitudes listed in different global magnitude catalogs. As we shall see during the course of this lecture, a systematic and unique approach in the magnitude determination does not exist. To penetrate and examine the various techniques, methods and scales is an important task not only for the everyday seismological practice but also for research and applications e.g. for tectonophysical studies, for comparison of the seismic activity of different regions, for studying the variation of activity with time, for investigations of attenuation laws or for seismic hazard evaluations.

Part 1. MAGNITUDE SCALES

History

In spite of its many drawbacks, magnitude is still frequently used by both scientists and the public to identify the size of an earthquake. It is interesting to mention that the magnitude scales, which all are using seismogram measurements, were developed first about 50 years after the first seismogram from a teleseismic event was made at Potsdam, Germany on April 17, 1889. Note that other source parameters, i.e. the location and origin time were routinely determined, with reasonable accuracy, already around the turn of the century.

In 1935, Charles Richter set up, at the California Institute of Technology, the local magnitude scale M_L to provide an instrumental measure of the size of earthquakes in southern California. The applied technique, to measure seismic wave amplitudes on seismograms, was similar to that introduced earlier by Kiyoo Wadati to evaluate Japanese earthquakes. According to Richter, the local magnitude is the logarithm (all logarithms here refer to the base 10) of the maximum seismic wave amplitude, A_0 (in μm), measured on a seismogram made by a standard torsion horizontal-component Wood-Anderson seismograph (free period 0.8 sec, maximum magnification 2800, damping factor 0.8) located 100 km from the epicentre. The M_L magnitude is basically a relative scale. It makes use of a standard size earthquake and rates the other shocks in a relative manner by their maximum amplitude under identical observational conditions. This follows from Richter's (1958) definition

$$M_L = \log[A(\Delta)/A_o(\Delta)] = \log A(\Delta) - \log A_o(\Delta) \qquad (1)$$

where Δ is epicentral distance and A_o and A are respectively, the maximum trace (recorded) amplitudes, written by a standard seismograph, of the standard event and of a given earthquake which occurred at a known distance. The standard earthquake, i.e. $M_L = 0$, by Richter also called "the zero shock" (not to be mixed with "no earthquake"), is an event which is recorded at a distance of 100 km by a Wood-Anderson instrument with a maximum trace amplitude of 1 μm. Analogously, if the Wood-Anderson instrument gives a peak amplitude of, say, 1 mm for an earthquake 100 km away, the magnitude M_L=3. It should be emphasized that when Charles Richter introduced the concept of earthquake magnitude, all the seismographic stations in southern California were equipped with Wood-Anderson seismographs with the same instrumental constants (Fig 1). The "zero level" was intentionally chosen low enough to make the magnitudes of the smallest events positive. It is assumed that the period of maximum amplitudes remains practically constant. Magnitudes of shocks at other distances can be calculated by making use of the variation of the maximum amplitude with distance. No particular wave type has been specified, so this maximum amplitude can be measured from whichever part of the seismogram showing the largest swing. Most likely, crustal and uppermost mantle waves of *S* type (*Sg, Sn*) have been used.

Richter's empirical attenuation formula for southern California reads (Kasahara, 1981)

$$\log A_o = 6.37 - 3 \log(\Delta) \qquad (2)$$

where A_o is measured in μm, Δ in km and $\Delta \leq 600$ km. Bullen and Bolt (1985) refer to a slightly different amplitude-distance dependence

$$\log A_o = 5.12 - 2.56 \log(\Delta) \qquad (3)$$

for $100 \leq \Delta \leq 600$ km. Empirical tables can be found in Richter (1958). Richter calculated the M_L magnitudes separately from the N-S and E-W component seismograms made at a number of stations and took the mean of all magnitude determinations.

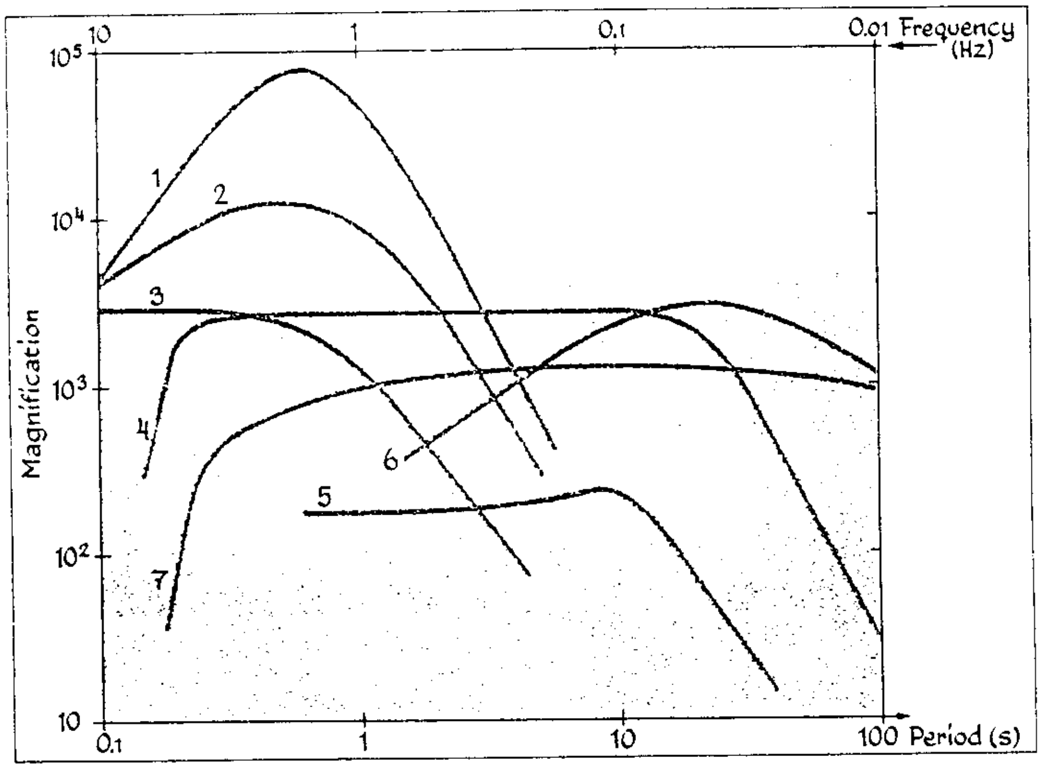

Fig. 1. Period-dependent response characteristics for several seismograph systems: 1) Benioff (SP-WWSSN); 2) Grenet-Coulomb; 3) Wood-Anderson; 4) Kirnos; 5) Wiechert; 6) Press-Ewing (LP-WWSSN); 7) Broad-band. SP=short period, LP=long period.

As follows from the above description, the M_L scale can be applied only to shallow earthquakes, with depths less than approximately 60 km, in southern California and to measurements made by the Wood-Anderson seismometer at distances up to 600 km. However, taking into account the magnification of 2800 for the instrument, we can replace the maximum trace amplitude, A, by the ground amplitude, a. We write

$$\log A = \log (2800\, a) \qquad (4)$$

Introducing (2) and (4) into (1), we obtain (Kasahara, 1981)

$$M_L = \log a + 3 \log(\Delta) - 2.92 \qquad (5a)$$

whereas application of the correction function (3) leads to

$$M_L = \log a + 2.56 \log\Delta - 1.67 \qquad (5b)$$

These formulae have a more general application, as they may be used for any type of instrument, provided that the ground amplitude, a, is known. Outside California, the scaling functions (2) or (3) are of provisional utility only. It should also be emphasized that Californian earthquakes are usually confined to a rather narrow focal-depth range, say, between 10 and 20 km. Hence, considerably different amplitude-distance curves than (2) or (3) are expected for deeper shocks and other regions. To present a review of regional magnitude scales applied at various local or regional observatories exceeds the scope of this lecture. In the following discussion, we shall limit ourselves to teleseismic magnitudes only.

Further developments of magnitude scales led to utilization of various specified seismic waves and/or phases and well defined calculation methods/techniques. Magnitude scales were extended to all epicentral distances, to shocks deeper than normal and to the use of particle ground velocity instead of trace amplitude. Large efforts have been devoted firstly, to unify the results achieved by different techniques into a common measure of the earthquake size and secondly, to make use of observed differences between individual magnitude estimates for a given event to deduce a picture of the character of the source.

It was difficult to extend the amplitude-distance dependence to distances much beyond 600 km for the area of California. Hence, the M_L magnitude was designed to measure only local or regional events and used nonspecified wave types. Then, it was quite natural that Gutenberg (1945a) extended the magnitude concept to teleseismic distances and special wave types. He developed the magnitude M_S using 20-sec surface waves from shallow earthquakes measured within the epicentral distance from 15° to 130°. The amplitude-distance correction which was determined theoretically and empirically includes corrections for geometrical spreading, absorption and dispersion. The M_S scale was adjusted to give roughly a continuation of M_L for large-distance events of magnitude from 6 to 7. The final formula reads

$$M_S = \log A + 1.656 \log(\Delta) + 1.818 \qquad (6)$$

where A is the combined maximum amplitude of the ground motion on horizontal components, measured in μm, for surface waves having 20 s period. Obviously, the M_S scale cannot be applied to intermediate-focus and deep-focus earthquakes because they do not generate large surface waves. M_S of major shallow earthquakes in the world have been determined since the end of the last century.

Gutenberg (1945b) also introduced the concept of body-wave magnitude, m_b, based on recorded P, PP and S waves from shallow earthquakes. He observed that the relative dependence of the ratio (A/T) for the three phases remains roughly constant for a relatively broad range of periods. The m_b scale was calibrated to agree with M_S defined in (6). The formula introduced by Gutenberg has the form

$$m_b = \log(A/T) + q(\Delta) + 0.1(m_b - 7) + C_r \qquad (7)$$

where the amplitude-distance correction term $q(\Delta)$ was constructed by theory and observations and comprises correction for both the geometrical spreading and anelastic absorption. $q(\Delta)$ is tabulated for PH, PZ, PPH, PPZ and SH wave types. C_r is an empirically determined station correction. The term $0.1(m_b\text{-}7)$ was included in order to achieve agreement between m_b and M_S. It shows that magnitude scales using body-wave and surface-wave measurements are not compatible with each other without corrections. Note also that ground motion amplitude, A, has ben replaced in eq (7) by a ratio A/T, which means that displacement has been substituted by velocity.

Further modifications of the body-wave magnitude definition (7) were introduced by Gutenberg (1945c) to be able to quantify also deep distant earthquakes. The philosophy followed during this work was that two earthquakes at different focal depths should have the same magnitude provided that they release the same seismic energy. In other words, seismic energy becomes an important parameter in quantifying earthquakes. Calibrating functions $q(\Delta,h)$ are given again for PH, PZ, PPH, PPZ, and SH in tabular and graphical form.

In 1955, Gutenberg and Richter presented improved empirical calibration functions and the body-wave magnitude, m_b, is now evaluated through a simplified formula

$$m_b = \log(A/T) + Q(\Delta,h) \tag{8}$$

The distance-depth correction factors are available in tabular form for shallow shocks and *PZ, PH, PPZ, PPH, SH* in the distance range 16°-170° (Table 1). For earthquakes with focal depths down to 700 km, $Q(\Delta,h)$ are available in diagrams covering epicentral distances from 5° to 110° for *PZ*, 10° to 110° for *SH* and 20° to 170° for *PPZ*. As an example, the diagram for vertical-component *P* waves is displayed in Fig. 2. One important conclusion can be drawn at this moment, namely that the two magnitudes, m_b and M_S, are not compatible, which means that they cannot be made to agree in their entire extent. Rather a relation between them has to be specified. For shallow events, the following formulae have been found empirically

$$\begin{aligned} m_b &= 0.63M_S + 2.5 \\ M_S &= 1.59m_b - 3.97 \end{aligned} \tag{9}$$

Note that the two values agree at $m_b = M_S \doteq 6.75$; above this $M_S > m_b$, below it $M_S < m_b$.

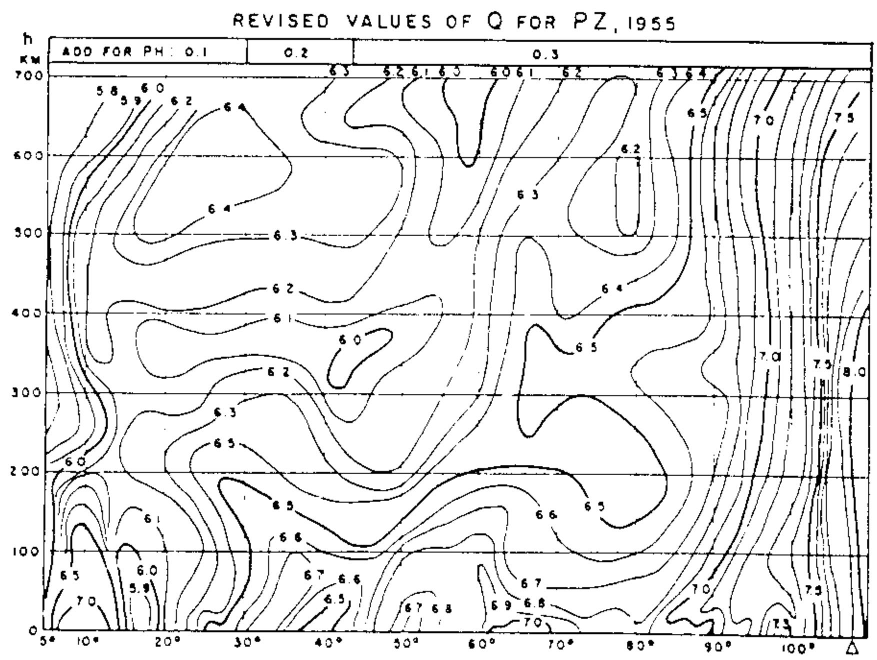

Fig 2. Values of $Q(\Delta,h)$ for vertical-component *P* waves (Gutenberg and Richter, 1956).

TABLE 1

Calibrating function for body-wave magnitudes and shallow shocks (Gutenberg and Richter, 1956).

VALUES OF 10 Q FOR SHALLOW SHOCKS

Δ	PZ	PH	PPZ	PPH	SH	Δ	PZ	PH	PPZ	PPH	SH	Δ	PZ	PH	PPZ	PPH	SH
16	59	60			72	56	68	71	69	70	66	96	73	76	72	74	71
17	59	60			65	57	68	71	69	70	66	97	74	78	72	74	72
18	59	60			62	58	68	71	70	71	66	98	75	78	72	74	73
19	60	61			58	59	68	71	70	72	66	99	75	78	72	74	73
20	60	61			58	60	68	71	71	73	66	100	74	77	72	74	74
21	61	62			60	61	69	72	72	74	67	101	73	76	72	74	74
22	62	63			62	62	70	73	73	74	67	102	74	77	72	74	74
23	63	64			62	63	69	73	73	74	67	103	75	79	72	74	73
24	63	65			62	64	70	73	73	75	68	104	76	79	73	75	73
25	65	66			62	65	70	74	73	75	69	105	77	81	73	75	72
26	64	66			62	66	70	74	73	74	69	106	78	82	74	76	72
27	65	67			63	67	70	74	72	74	69	107	79	83	74	76	72
28	66	67			63	68	70	74	71	73	69	108	79	83	74	76	72
29	66	67			63	69	70	74	70	72	69	109	80	84	74	76	72
30	66	68	67	68	63	70	69	73	70	72	69	110	81	85	74	76	72
31	67	69	67	68	63	71	69	73	71	73	70	112	82	86	74	76	
32	67	69	68	69	64	72	69	73	71	73	70	114	86	90	75	77	
33	67	69	68	69	64	73	69	72	71	73	69	116	88		75	77	
34	67	69	68	69	65	74	68	71	70	72	68	118	90		75	77	
35	67	69	68	69	66	75	68	71	69	71	68	120			75	77	
36	66	68	67	68	66	76	69	72	69	71	68	122			74	76	
37	65	67	67	68	66	77	69	72	69	71	68	124			73	75	
38	65	67	67	68	66	78	69	73	69	71	69	126			72	74	
39	64	66	66	67	67	79	68	72	69	71	68	128			71	74	
40	64	66	66	67	67	80	67	71	69	71	67	130			70	73	
41	65	67	65	66	66	81	68	72	70	72	68	132			70	73	
42	65	67	65	66	65	82	69	72	71	73	69	134			69	72	
43	65	67	66	67	65	83	70	74	72	74	69	136			69	72	
44	65	67	67	68	65	84	70	74	73	75	69	138			70	73	
45	67	69	67	68	65	85	70	74	73	75	68	140			71	74	
46	68	71	67	68	66	86	69	73	73	75	67	142			71	74	
47	69	72	67	68	66	87	70	73	72	74	68	144			70	73	
48	69	72	67	68	67	88	71	75	72	74	68	146			69	72	
49	68	71	67	68	67	89	70	74	72	74	68	148			69	72	
50	67	70	67	68	66	90	70	73	72	74	68	150			69	72	
51	67	70	67	68	65	91	71	75	72	74	69	152			69	72	
52	67	70	67	68	65	92	71	74	72	74	69	154			69	72	
53	67	70	67	68	66	93	72	75	72	74	69	156			69	72	
54	68	71	68	69	66	94	71	74	72	74	70	158			69	72	
55	68	71	69	70	66	95	72	76	72	74	70	160			69	72	
												170			69	72	

The view of "one earthquake - one magnitude" led Gutenberg and Richter (1956) to the concept of the unified magnitude, defined as a weighted mean of the body-wave and surface-wave magnitude. They never published the details of their method to determine the unified magnitude. A comprehensive description of the unified magnitude by Gutenberg and Richter my be found in Geller and Kanamori (1977). According to Båth (1981) the formula reads

$$m_b(\text{unif}) = (3/4)m_b + (1/4).(0.63M_S + 2.5) \qquad (10)$$

which follows from eq.(9a). Equation (10) shows that the unified magnitude is determined primarily from body-wave magnitudes, with only supplemental contribution from surface-wave magnitudes. Note that the two magnitudes are measured at different periods. Two years later, Richter (1958) modified the concept of the unified magnitude as given in eq.(10) and defined it on the basis of surface waves as (Geller and Kanamori, 1977)

$$\begin{aligned} M_S(\text{unif}) &= (1/4)M_S + (3/4).(1.59m_b - 3.97) \quad \text{for } h < 40 \text{ km} \\ M_S(\text{unif}) &= 1.59m_b - 3.97 \qquad \text{for } h = 40\text{-}60 \text{ km} \end{aligned} \qquad (11)$$

Equations (11) are using the conversion formula (9b) and so it seems that the unified magnitude based on surface waves is also emphasizing m_b. In spite of the fact that M_L, m_b and M_S, or their ample modifications, quickly became the standard fundamental earthquake source parameters, m_b(unif) and M_S(unif) did not receive a world-wide endorcement for routine use.

In the 1950's and 1960's, magnitude formulae of Gutenberg and Richter have been frequently used at individual seismograph stations, seismological laboratories, networks and agencies around the world and one could feel that the problem of quantifying earthquakes has been solved. However, it was recognized that the technique of magnitude determination employed during those years was unsatisfactory and cannot be used as an objective standard treatment to estimate the size of earthquakes throughout the world. Different and often poorly defined magnitude scales came into use. Often it became impossible to compare the magnitude values obtained from different units for the same earthquake. The divergencies frequently exceeded the expected measuring scatter. Therefore, the standardization of magnitude scales was one of the issues on the agenda of the XXII General Assembly of the IUGG held in Helsinki, Finland, in 1960. The problem was examined by a special commission consisting of workers from former USSR and Czechoslovakia. In 1962, the commission made a proposal (Vanek et al., 1962) of a standard magnitude formula and of generally acceptable calibrating functions. They suggested that the magnitude from all types of waves should be determined as

$$M = \log(A/T)_{max} + \sigma(\Delta) \qquad (12)$$

where A is the displacement amplitude of the ground motion in μm, T is the corresponding period in seconds and $\sigma(\Delta)$ is the calibrating function that expresses the decay of A/T with epicentral distance and is different for different wave types. The ratio A/T in eq. (12) has been used for two reasons. Firstly, for body-wave magnitudes, one can obtain comparable values only when using the ratio A/T rather than A. Secondly, for surface-wave magnitudes determined for a fixed period, e.g. 20 s, the calculation will be restricted to epicentral distances of, say 50°-180°, at which 20 s surface waves show their maximum amplitudes (Table 2). Only stations equipped with instruments of sufficiently long periods could be used. Vanek et al.(1962) claim that the maximum value of A/T is stable for any period of the maximum of surface waves in the entire distance range from 2° to 180°.

TABLE 2

Mean periods corresponding to the maximum amplitudes of surface waves (Vanek et al.,1962)

Δ°	T[s]	Δ°	T[s]	Δ°	T[s]
1	3-5	20	9-14	80	16-22
2	4-6	25	9-16	90	16-22
4	5-7	30	10-16	100	16-25
6	5-8	40	12-18	120	16-25
8	6-9	50	12-20	140	18-25
10	7-10	60	14-20	160	18-25
15	8-12	70	14-22	180	18-25

Vanek et al (1962) present a calibrating function for body-wave magnitudes, $\sigma(\Delta)$, which differs only slightly from $Q(\Delta,h)$ in eq.(8) for shallow shocks. Note that Vanek et al. (1962) do not present $\sigma(\Delta)$ for deeper shocks.

To construct a standard calibrating function for surface-wave magnitudes, Vanek et al. (1962) examined fourteen calibrating functions developed by other authors. A statistical averaging procedure yielded the standardized calibrating function of the form

$$\sigma(\Delta) = 1.66 \log\Delta + 3.3 \quad (13)$$

for horizontal-component surface waves and epicentral distances between 2° and 160°.

The IASPEI Committee on Magnitudes, met in Zurich, Switzerland, in 1967 and recommended (see Appendix) the use of the standardized calibration formula (13) and the determination of magnitudes according to formula (12). Stations should report magnitudes for all waves for which calibration functions are available. Surface-wave magnitude, M_S, and body-wave magnitude, m_b, are determined by the following equations and reported separately. Historically, the surface wave magnitude was first calculated through eg.(6) as an extension of the local magnitude.

The Zurich recommendation for surface-wave magnitude is

$$M_S = \log(A/T)_{max} + 1.66 \log\Delta + 3.3 \quad (14)$$

where $(A/T)_{max}$ refers to the horizontal-component Rayleigh waves, A is trace amplitude in μm, T = 17-23 s (some authors give the period range 18-22 s, other recommend the range of 10-30 s or even 10-60s), Δ = 20° - 160°. Formula (14), which is commonly known as Moscow-Prague formula also referred to as Prague formula, is applicable only to shallow shocks with focal depth $h \leq 50$ km. It is possible to correct M_S for focal depth using e.g. the following depth corrections suggested by Båth (1981).

h (km)	0-50	60	70	80	90	100	>100
M_S	0	+0.1	+0.2	+0.3	+0.4	+0.4	+0.4

As far as body-wave magnitude is concerned, the Zurich recommendation reads

$$m_b = \log(A/T)_{max} + \sigma(\Delta,h) \quad (15)$$

where $(A/T)_{max}$ is determined for all wave types (*PZ, PH, PPZ, PPH, SH*) for which the calibrating functions, $\sigma(\Delta,h)$, by Gutenberg and Richter (1956) are available. Observe that for shallow earthquakes at short distances, say, less than 20°, the amplitude-distance functions vary substantially from area to area and regional calibration functions must be invoked. A conversion from M_S to m_b can be made through

$$m_b = 0.56M_S + 2.9 \tag{16}$$

which differs from that given in eq.(9a). One important aspect of the 1967 Zurich recommendation is the recognition that m_b and M_S are two different scales which should be calculated and reported separately and not be combined into a unified magnitude of any type.

The preceding recapitulation should be viewed as a brief description of the development of magnitude scales up to the Zurich recommendations. At least three issues are of special importance. Firstly, the definition of magnitude which grew up gradually from the work of Richter in 1935 to conclusions accepted in Zurich in 1967. Secondly, the use of different kinds of seismic waves, originally introduced to include also deeper shocks. These types of waves also cover the long-period portion of the spectrum which is required in modern seismogram analysis. Thirdly, separate calculation and reporting of body-wave and surface-wave magnitudes.

Recent Developments

After the Zurich meeting, further significant developments in earthquake quantifications is mainly due to wast installations of modern instruments and to introduction of new processing techniques.

In the middle of the 1940's, when Gutenberg introduced the teleseismic body-wave magnitude scale, most of the operating seismographs were then the so-called medium-period instruments. Measured P waves showed periods around 5 s, whereas S waves exhibited periods of the order of 10 s. In the early 1960's, the World-Wide Standardized Seismograph Network, WWSSN, was installed (about 120 stations in 60 countries). The instrumentation used for the WWSSN extended considerably the spectrum of recorded seismic signals both towards longer as well as towards shorter periods. Short-period instruments record P waves with dominating periods around 1 s. These records are of great importance of detecting weak seismic events. Long-period instruments, with maximum magnification around 20 s, have proved to be particularly useful for recording large events (Fig. 1). Seismograms from the largest earthquakes often display ample surface waves with periods of hundreds of seconds. The access to a wider signal spectrum contributed to an understanding of the magnitude problem and accelerated its further development.

Saturation of magnitude scales

According to Brune (1970), the far-field body-wave displacement spectrum can be approximated by a constant long-period level, Ω_0, and a high-frequency decay for frequencies above the corner frequency, f_0. Ω_0 and f_0 are related, in a relatively simple manner, to the seismic moment scalar, M_0, fault length, L, and stress drop, $\Delta\sigma$. M_0 is proportional to Ω_0, while L is inversely proportional to f_0. As the first approximation, we can also assume log M_0 be linearly related to any magnitude (Fig. 3)

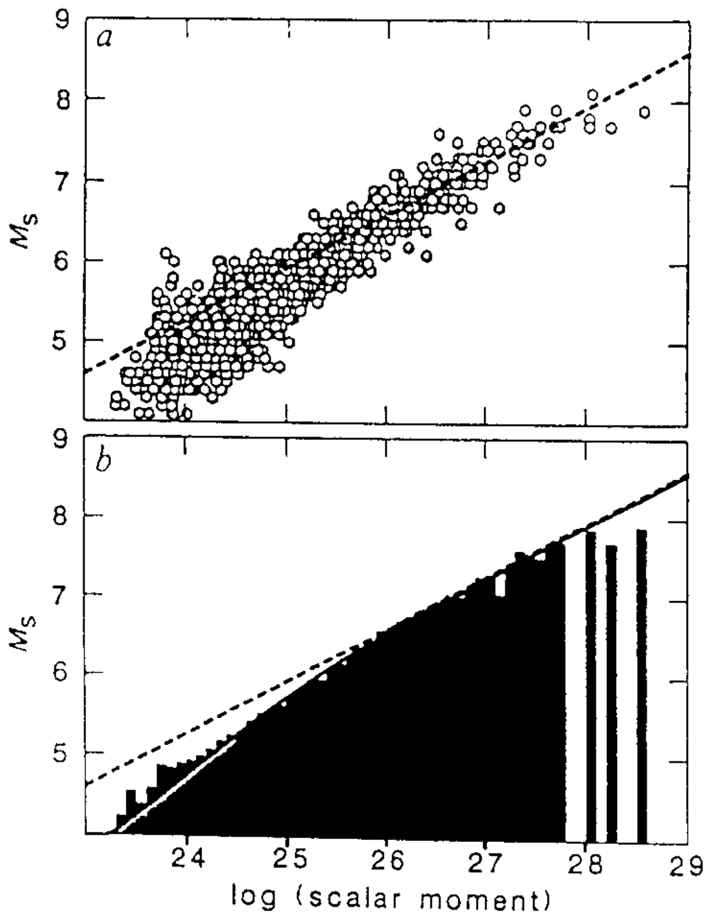

Fig. 3. Surface-wave magnitude, M_S, and seismic moment, M_0 (dyne-cm), for recent major earthquakes (from Koyama, 1997, after Ekström and Dziewonski, 1988, modified)

By making use of the so-called ω-square model (flat spectrum for frequencies lower than f_0 and a fall off with a gradient of f_0^{-2} for higher frequencies) we obtain the family of curves displayed in Fig. 4. Note that the spacing between the curves is uniform at the period of 20 s to assure the linearity between the surface-wave (20 s) magnitude and log M_0. The broken line in Fig. 4 has a gradient of T^3. It follows from the figure that the linearity between M_S and log M_0 is preserved as long as the measured signals show dominant periods in the flat portion of the assumed spectrum (Fig. 4). This will be the case for events of magnitude up to roughly 7, for dominating 20 s surface waves. For larger events, a contraction occurs, leading to a saturation (underestimation) of M_S. It follows from this brief descriptions that one of the major advantages of the seismic moment, M_0, is that it never shows the saturation effect as it is always deduced from the flat portion of the spectrum.

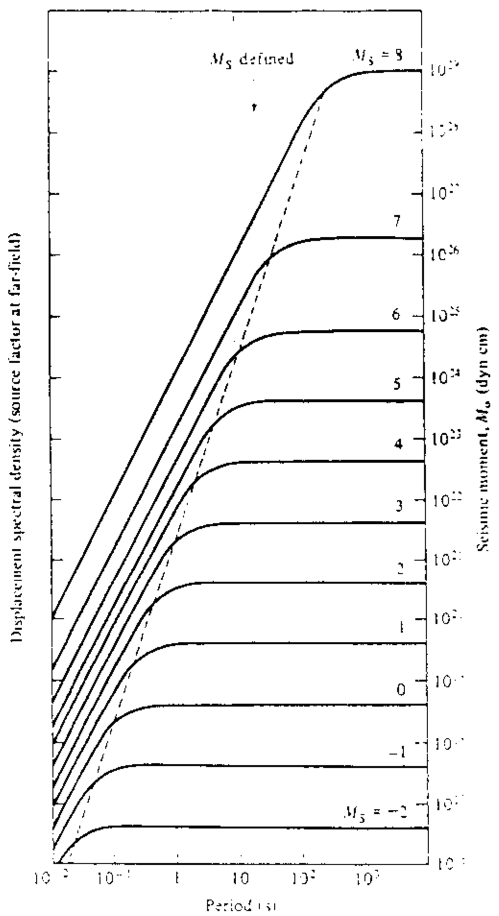

Fig. 4. Far-field displacement spectrum from earthquakes with various magnitudes, M_S, and moments, M_0 (dyne-cm), calculated by making use of the ω-square model (from Kasahara, 1981, after Aki, 1967).

It is possible to explain the saturation effect in terms of the source dimensions and mechanism. Recorded short-period P waves represent a sequence of ruptures (in the source volume), where each single onset in the seismogram corresponds to just one break (Båth,1981). Extremely large sources generate extremely long-period seismic waves. The long-period surface waves do not respond to each single break with a new onset in seismograms but integrate the source processes and represents combined effect of the total phenomenon. Surface waves of 20 s period, i.e. wavelengths of 50-80 km, "do not see" source motions along faults which are significantly longer than 50-80 km. In other words, the 20-s surface waves are inefficient in integrating the source motions on faults with dimensions larger than the wavelength. Since a fault length of 50-80 km corresponds approximately to en earthquake of magnitude 7.0, surface-wave magnitudes (20 s) of large and great earthquakes are most likely heavily underestimated. M_S scale loses its validity for events beyond magnitude 8 (Kasahara, 1981).

Saturation effect influences also m_b magnitudes. Since these are measured at 1 s periods, saturation starts at lower magnitudes, probably at m_b=6. We may conclude here that in spite of the fact that m_b or M_S magnitudes are, by definition, unbounded from above (although earthquake size certainly has an upper limit and news media erroneously often talk about "12-degree Richter scale"), in fact, they are so bounded due to the finite bandwidth of instrumentation used at current seismographic stations. With regard to the source spectrum, magnitudes would be more realistic if categorized as long-, medium- and short-period rather than surface- and body-wave magnitudes.

Seismologists are well aware of the saturation effect and several procedures have been suggested to circumvent this drawback. The dominating idea is of course to determine the magnitude from long- or very long-period seismograms. For instance, the seismographic station Obninsk, Russia, determines, for selected major earthquakes, up to eight P-wave magnitudes measured on records from eight seismographs with dominating periods from 1 s to approximately 100 s. The maximum P-wave magnitude and its period is published together with the conventional body-wave and surface-wave magnitudes. Nortmann and Duda (1983) are using band-pass filtered broad-band records to determine so called spectral magnitudes. They developed a calibration function which is a function of distance, focal depth and period. Okal and Talandier (1989) introduced the concept of mantle magnitude, M_m. Measurements of M_m is taken on the Fourier spectrum of mantle Rayleigh waves and corrections for the source and distance are added. The largest value of M_m over the available Fourier spectrum is retained and is assumed to be proportional to log M_0. Shallow as well as deep shocks are quantified and the periods considered are 50 s and longer.

The three method examples mentioned, in spite of their justification by research and specific requests, have not attained global acceptance in seismological practice. The situation was, however, different when Kanamori (1977) and Hanks and Kanamori (1979), in seeking a physically meaningful measure of earthquake size, presented their moment magnitude scale. The basic idea was to determine the magnitude from an estimate of the radiated energy obtained from a magnitude independent relation. Kanamori (1977) shows that the radiated energy

$$E_s = (\Delta\sigma/2\mu)M_o \tag{17}$$

where $\Delta\sigma$, is the earthquake stress drop and μ is the rigidity or shear modulus. Taking, e.g. the rigidity to be $5\text{x}10^7$ dyne/cm^2 and assuming the constancy of the stress drop for crust-upper mantle events, say 50 bars, i.e. $5\text{x}10^{11}$ dyne/cm^2, $(\Delta\sigma/\mu)=10^4$ and eq.(17) reduces to

$$E_s = (1/2\text{x}10^4)M_o \tag{18}$$

Gutenberg-Richter relation between E_S, in ergs, and M_S reads

$$\begin{aligned} \log E_S &= 1.5M_S + 11.8 \text{ (erg)} \\ &= 1.5M_S + 4.8 \text{ (J)} \end{aligned} \tag{19}$$

Since M_S is bounded, so too is E_S determined from (19). However, if E_S is evaluated independently, e.g. from eq.(17), it can be introduced into (19) to determine a moment magnitude, M_W, which will not saturate. An important feature of M_W, as defined here, is that it is in good agreement with M_S for a number of earthquake with $M_S \simeq 8$. Introducing (18) into (19), the moment magnitude M_W is defined as

$$\log M_o = 1.5M_W + 16.1 \tag{20}$$

which is remarkably coincident with several relationships defined empirically by other workers. Then, a single moment magnitude, M, may be written as

$$M = (2/3) \log M_o - 10.7 \tag{21}$$

where M_o is in dyne-cm. M defined in (21) is uniformly valid for $3 \leq M_L \leq 7$, $5 \leq M_S \leq 7.5$, and M_W for larger magnitudes. In contrast to various spectral magnitudes mentioned above, the moment magnitude of Hanks and Kanamori is frequently used by the seismological community to evaluate especially large earthquakes.

The largest surface-wave magnitudes of 8.9, measured so far, are associated with two earthquakes. One in 1906, off coast of Colombia-Ecuador and the second in 1933 off Pacific coast of Japan (the great Sanriku earthquake). This observation suggests that an upper limit of surface-wave magnitudes, globally speaking, is just below 9. However, if we instead take the moment magnitudes, we reveal that during the second half of this century there were at least four shocks with $M_W \geq 9.0$ (Table 3).

TABLE 3

M_S, M_o, M_W for four great earthquakes (Bullen and Bolt, 1985)

Date	Region	M_S	M_o x10^{27} dyne cm	M_W
1952, Nov 4	Kamchatka	8.25	350	9.0
1957, Mar 9	Aleutian Is.	8.25	585	9.1
1960, May 22	Chile	8.3	2000	9.5
1964, Mar 28	Alaska	8.4	820	9.2

Distance bias

Since the adoption of the Prague formula in 1967, there has been a debate with regard to the edequacy of the amplitude-distance function in eg.(13). Modern research reveals that the Prague formula needs modifying since for short distances, it underestimates the magnitude, while for large distances, the magnitudes are slightly overestimated. Herak and Herak (1993) propose that formula (14) should be replaced by

$$M_S = \log (A/T)_{max} + 1.094 \log(\Delta) + 4.429 \qquad (22)$$

More recently, Rezapour and Pearce (1997) suggested to use

$$M_S = \log (A/T)_{max} + 1.155 \log(\Delta) + 4.269 \qquad (23)$$

Both formulae were determined empirically using global data. A simple numerical test shows that the difference between M_S calculated through the Prague formula (14) and M_S specified in (22) or (23) is of the order of 0.2 magnitude units for distances around 40°, it is less than 0.1 for distances around 80° and becomes less than 0.05 at distances around 100°. In general, eg.(23) gives magnitude estimates which are closer to Prague formula magnitudes, when compared with those deduced from eq.(22). The issue of surface-wave magnitude distance bias is still a subject of research and, as yet, has not influenced the seismological practice.

Magnitude Measurements

The Zurich recommendations established complete definitions of body- and surface-wave magnitudes and basic rules for their measurements on seismic records. Nevertheless, the practice performed at seismographic stations and larger centres is far from uniform. The different procedures, together with radical simplification of the complex physical process at the seismic source, contribute to the scatter observed between different magnitude calculation. Nowadays, it is generally understood that magnitude determinations involve uncertainties of 0.3 even under most favourable conditions. Below, I summarize the rules to be followed in calculating the magnitudes from seismogram measurements. Body- and surface-wave magnitudes are treated separately.

Body-wave magnitudes

Even though distance-depth correction factors to determine m_b are available for *PZ, PH, PPZ, PPH* and *SH*, the current practice is making use, almost exclusively, only of *PZ* and waves with periods around 1 s. One of the factors contributing to the relatively large scatter of m_b is the variation of periods incorporated into the magnitude determination. For example, Shapira and Kulhanek (1978) showed that averaging of magnitudes over a range of approximately 1-2 s, reduced the scatter of body-wave magnitudes, determined at five Swedish seismographic stations, by about 50%.

Body-wave magnitudes are consistently calculated from the ratio $(A/T)_{max}$ which refers to the ground motion. General practice, however is that rather than $(A/T)_{max}$, A_{max} and corresponding period are measured on the seismogram.

Also abandoned is today the habit to measure the first swing of the P-wave. The first swing is often small and will result in too low a magnitude (Fig. 5). Most station analysts nowadays adopted the rule to measure the maximum P-wave amplitude within an interval of, say, 10 s after the first onset. This maximum is considered to be more representative of the energy carried by the P wave (Koyoma, 1997) and is less influenced by the source radiation pattern. The interference of secondary phases like pP or PcP, even though probably of minor importance, should be minimized. On the other hand, for multiple shocks, usually with increasing size, it is worth to measure each successive onset in the P wave train. In doing this, we may obtain important information on the development of the shock sequence, most likely not available from surface-wave measurements.

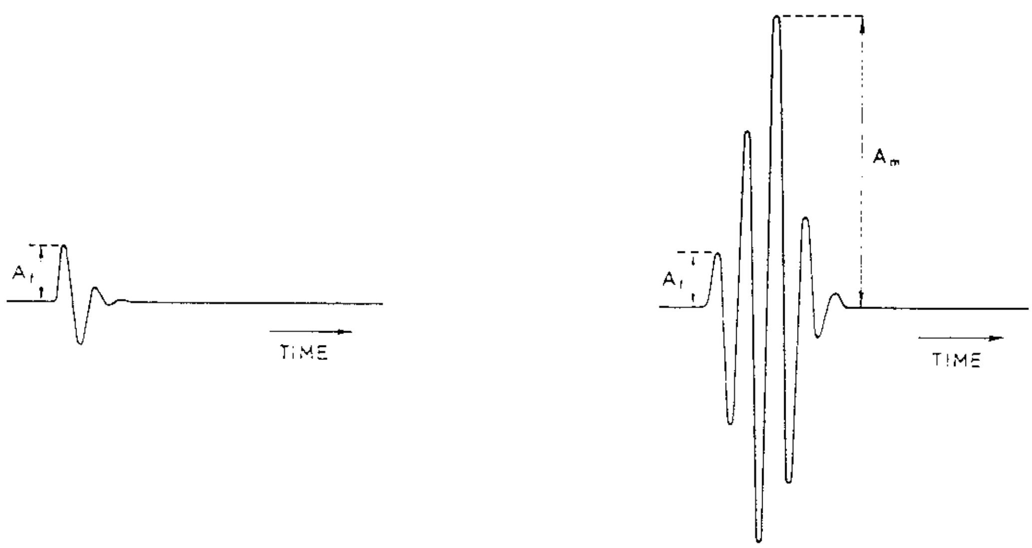

Fig. 5. Different ways of measuring trace amplitudes for body waves (after Båth, 1979).

Magnitudes reported from Uppsala and some other stations are systematically higher than those given by NEIC and ISC. The differences are, on average, as large as 0.7 and 0.8, respectively

(Båth, 1975). This is, most likely, due to the fact that these agencies use different portions of the recorded P wave to determine m_b.

Surface-wave magnitudes

To measure on seismograms the maximum ground particle velocity $(A/T)_{max}$ seems to be a trivial task. However, there are several precautions to be taken, some of them applicable also to other magnitudes.

The recommendation is to use $(A/T)_{max}$. In order to find the maximum ratio, one should calculate A/T for several trace maxima and select the largest value of A/T among them. To the best of my knowledge, nobody nowadays follows this procedure. Instead, A_{max}/T is measured as in the case of m_b. Due to the limited period range, the error introduced by this simplification is generally small. In accordance with the Zurich recommendation, $(A/T)_{max}$ refers to ground particle motion and falls within the period range 17-23 s (Båth, 1981). There is certain discrepancy concerning the recommended period range. For example, in the original proposal, Vanek et al. (1962) assume 20 s periods only, Willmore (1979) and Båth (1979) give the interval 18-22 s, Uppsala bulletins use 10-30 s, while ISC considers periods from 10 to 60 s.

Horizontal-component Rayleigh waves should be used. The maximum horizontal amplitude is obtained by vectorial summation, irrespective of the arrival times. This means that simultaneous amplitudes on the two components, or the maximum amplitudes on the two components are measured. The letter value may considerably exceed the former and differences in arrival times may be of the order of minutes. If one of the two horizontal components is absent, the available maximum amplitude multiplied by $\sqrt{2}$ is used.

The recommendation to use horizontal components is due to the historical development. First instruments deployed at seismographic stations were principally horizontal seismometers. First in the 1960's, the stable operation of medium- and long-period vertical seismometers increased considerably in number. Also theoretically, the use of vertical instruments is more desirable since they record exclusively Rayleigh type waves and only one record needs to be measured. Horizontal instruments, on the other hand, record a superposition of Love and Rayleigh waves and the separation may be difficult. Note that due to the Zurich recommendation, Love waves should strictly not be involved. In many cases, the Prague formula is applied to the vertical component without any change. However, the correct procedure would be to examine at least the constant term, provided that the calibrating function (13) can be accepted as valid also for the vertical component. For a homogeneous structure and Rayleigh waves, the ratio $A_z/A_h = 1.48$, which implies a reduction of the constant term by 0.2.

Concluding remarks

It follows from the above discussion that the advantage of the magnitude is twofold. Firstly, it can be easily measured on records and quickly evaluated through simple formulae. Secondly, it offers, at least, an approximate estimate of earthquake size, and other source parameters, for a large range of events, from very small to great shocks. On the other hand, the concept of the magnitude has been criticized from several aspects and some of them are briefly summarized below.

As already mentioned, the magnitude lacks a physical dimension, it is poorly defined, it has no direct relation to the complicated physical processes at the earthquake source, it suffers from saturation and distance bias. It seems that most, if not all, of these drawbacks will disappear when we replace the magnitude by seismic moment. The only difficulty with M_0 may arise when analysing older, usually photographic-paper, analog records which have to be digitized first.

Any magnitude determination is essentially "monochromatic", i.e. it employs only a single period T (exceptions being spectral magnitudes and coda magnitudes). The wave spectrum generated by the earthquake source is, however, by no means, monochromatic. Therefore, it is too simple and inadequate to replace the whole frequency band emitted by the source by one specific frequency (period). Note also that the correction factors in (8) and in (12) are considered as frequency invariant which certainly is not the case.

In my opinion, the strongest argument against magnitudes as yardsticks of earthquake sizes is that in magnitude calculations we ignore all information, except three points, provided by the seismogram. Therefore, magnitude determination involves the loss of prevailing amount of information which is contained in available seismic records. It is hard to accept that a complicated phenomenon such as faulting of rocks can be reasonably well described by two (amplitude, period) simple measurements on the record and the rest of the seismogram is not considered. Note that, in contrast, M_o is evaluated from the displacement ground motion spectrum, i.e. an integration of the record, or a selected section of it, is part of the procedure.

Part 2. MAGNITUDE CATALOGS AND BULLETINS

In spite of the imperfection of m_b and M_S, these magnitudes, at this writing, are still the most common measure of earthquake size. Together with time of occurrence and location, magnitudes are the essential source parameters frequently listed in many global catalogs or bulletins. However, there are significant differences between magnitudes listed in these publications. This situation did not improve dramatically with time and so even listings published currently by various agencies (apparently using different selection criteria) vary more than what would be justified by measuring errors and the expected satter. Because of the fundamental importance of the earthquake size in many seismological studies, below, I review the major differences of magnitude scales adopted in some of the frequently referred catalogs/bulletins. The review is innevitably far from complete mainly because of several reasons. Firstly, international centers have practically no control of changes in reporting practice made at individual stations, especially if the resulting magnitude difference is well within the scatter of reported values. To improve the situation assistance of data contributers would be required. Secondly, some authors do not specify explicitly the technique used (scales, measurements, components, periods, etc) in magnitude evaluations. Thirdly, it is usually very difficult to collect information on procedures applied earlier, say, prior to 1960.

The International Seismological Centre, ISC

The ISC was established at the XIIIth General Assembly of the IASPEI, replacing the earlier International Seismological Summary, ISS. The main objective of the ISC is to collect and process data from seismographic stations in all parts of the world to determine origin time, location and magnitude of reported earthquakes. Information provided by the ISC monthly bulletins today is probably the most comprehensive publication on global seismicity.

ISS did not report any magnitudes. The ISC, from its very beginning in 1964, reports the body-wave magnitudes, m_b, following the procedure outlined by Gutenberg and Richter (1956). Station magnitudes are determined according to eq.(12) for all stations reporting amplitude and period readings. m_b given in the ISC Bulletin is the average over all station magnitudes. It is assumed that the amplitude and period correspond to a P wave of period less than or equal to 3 s and mesurements are made on short-period, vertical-component records. The maximum amplitude is read out to 5 s, or five cycles, after the first onset. At the beginning, distances from 5° to 160° were considered, however from 1968, values from stations in the range 5° to 20° are ignored.

For earthquakes which provide amplitude and periods for surface waves in the range from 10 to 60 s, M_S is calculated for that observation by employing the Prague formula. Vertical or resultant horizontal components (where readings on each are within 5 s; very few nowadays) are considered. Since May 1983, M_S is calculated only for events with focal depth less or equal to 60 km in the distance range 5° - 160°. An average M_S is calculated from observations received from stations in the distance range 20° - 160°.

The ISC started calculating M_S in January 1971 employing the Prague formula, amplitudes and periods of 20 s surface waves and distance range 20° - 160°. Up to 1976, the assessment of M_S was made by vectorially combining the maximum amplitudes of the horizontal components. Even though individual observations of M_S were calculated, no surface-wave magnitudes were adopted by the ISC prior to January 1979. Starting November 1979, the distance range of stations reporting readings has been increased from 20° - 160° to 5° - 160°.

Preliminary Determination of Epicenter, PDE

The PDE gives rapid (within a month) estimates of earthquake source parameters and publishes global earthquake monthly bulletins. As was the case of the ISC, the PDE bulletins represent one of the most consistent and reliable data sets for global seismology over the last three decades. This service is carried on by the United States Geological Survey through its National Earthquake Information Service, NEIS.

Starting August 1983, the reported m_b and M_S magnitudes are the 25-percent trimmed mean of magnitudes calculated from individual station observations. Prior to that date, the average magnitudes were computed using the straight arithmetic mean. The change introduced in 1983 was motivated by the observation that the magnitude distribution was non-Gaussian. However, with both procedures, individual station magnitudes will be discarded automatically by the computer if the magnitude value is more than one order of magnitude different from the average value. The body-wave magnitude is calculated through formula (8). The largest vertical-component amplitude in the *P*-wave group is measured, corresponding period is restricted to the interval from 0.1 to 3.0 s and only distances equal to, or larger than 5° are accepted.

In the 1960's or early 1970's, NEIC requested amplitude measurements for m_b to be done within the first three cycles. It is possible that some stations may still follow that practice, although it seems that most have now ignored that request and in fact report the largest amplitude in the P-wave group. Formula (8) and the $Q(\Delta,h)$ correction factors have been used since the middle 1960's. While the formula is unchanged, there has been a change in analyst practice over the years. Between the late 1970's and 1983, m_b for shallow events at distance range between 5° and 15° were removed from the average magnitude computation. This was because these magnitudes were too high compared to the average.

Surface-wave magnitudes are calculated from the Prague formula (14). Prior to 1975, NEIS has published estimates of M_S from horizontal-component measurements, but from May 1975, vertical-component surface waves within the period range 18-22 s and distance range from 20° to 160° are utilized. No depth corrections are applied and M_S are not generally calculated for focal depths greater than 50 km. If the uncertainty of the estimated focal depth is such that the depth could be less than 50 km, an M_S value may still be published in the PDE. In general, the M_S magnitude is more reliable than the m_b magnitude as a means of yielding the relative size of shallow earthquakes.

Some stations report amplitudes for various periods and phases. If this is the case, the PDE attempts to pick the correct values, with preference toward the values that are used in the magnitude determination. For example, if a station reports on two phases with periods 19 s and 25 s, the 19 s period will enter the magnitude calculation even though it may have a smaller magnitude, since 25 s is outside the accepted period range of the station calculation for M_S. If both periods are within the accepted range, then obviously the one with the largest amplitude will be chosen.

Seismicity of the Earth by Gutenberg and Richter (1954)

This magnitude catalog covers the time period 1904-1952. Gutenberg and Richter, hereafter G-R, did not explain the magnitude scale or the technique they used when determining the magnitudes. Here, I shall briefly summarize results achieved by Geller and Kanamori (1977) who carefully re-examined the magnitudes in Seismicity of the Earth from original worksheets of G-R. It follows from the available worksheets and original notes of G-R that the single station values M_S were averaged for each event. Surface-wave magnitudes deduced from the notes are very close to those listed in the G-R catalog. Magnitudes which G-R considered as accurate are given to the nearest tenth, values which they considered to be less accurate are given only to the nearest quarter. Geller and Kanamori (1977) conclude that the magnitudes in Seismicity of the Earth for nearly all shallow events are essentially equivalent to M_S. On the other hand, for shocks at depths of 40-60 km, G-R magnitudes resemble m_b.

Catalog of Great Shallow Earthquakes, In: Richter (1958)

Magnitudes presented in Seismicity of the Earth (Gutenberg and Richter, 1954) have been revised several times by Gutenberg and by Richter. Special interest obviously generate magnitudes of the largest shocks. In 1957, Gutenberg revised the magnitudes listed in Seismicity of the Earth for shocks with m_b=7.9 and larger that occurred between 1904 and 1952. His results, comprising 16 events, are given in Richter (1958, Table 22-5). Richter (1958) extended the period of observation by the interval from 1896 to 1903 and lists revised magnitudes for great (M=7.9 and greater) shallow shocks between 1904 and 1952. Richter did not publish any details of his method of determining the revised magnitudes. Geller and Kanamori (1977) performed an extensive work to reveal the scale employed by Richter. They conclude that the unified magnitudes, defined in eqs. (11), give the revised magnitude, M, of Richter (1958). Hence, the revised magnitude, M, in Richter's catalog is distinctly different from M_S used in Seismicity of the Earth. The magnitudes listed in these catalogs are determined on different scales and errors will result from treating the revised magnitudes, M, as M_S.

Survey of Earthquakes in period 1897-1964, In: Duda (1965)

This catalog comprises large earthquakes, $M \geq 7$, where M corresponds to the magnitude used by Richter (1958). Magnitude information in the Survey was adopted from a number of sources: for events 1897-1903 from Gutenberg as presented in Richter (1958), for 1904-1952 from Seismicity of the Earth with magnitude revisions by Richter (1958), for 1953-1957 from Gutenberg (Seismol. Lab. Bull., Pasadena), for 1957-1963 from the Provisional Readings at Pasadena and for 1964 from the Seismological Bulletin, Uppsala. Duda also examined Wiechert seismograms from the Uppsala station and period 1904-1917. In doing this he could add 146 earthquakes with magnitude 7 and larger to the 138 event reported in Seismicity of the Earth. Duda does not give any details about the magnitude evaluation from Uppsala records. It is likely (see also Abe, 1981) that these are M_S according to eq.(6) or the modified unified magnitudes determined through formulae (11). In any case, it seems that the catalog of Duda (1965) is not homogeneous with respect to the magnitude. Note e.g., that magnitudes given in the Richter's catalog are clearly different from those listed in Seismicity of the Earth.

Major earthquakes (M ≥ 7.0) during the period 1965-1977, In:Båth and Duda (1979)

This is a continuation of the catalog of Duda (1965). Magnitudes listed are averages of magnitude determination for the Swedish stations Uppsala and Kiruna with applications of the Zurich recommendations for body-wave magnitudes, eq.(12) as well as for surface-wave magnitudes, eq.(14). m_b are measured on short-period, vertical-component P waves ($T \sim 1$ s) and M_S on horizontal-component Rayleigh waves (T = 17-23 s). M_S values from Kiruna and Uppsala generally agree within 0.1 unit, whereas for m_b the difference is usually about 0.2 unit. The authors consider the surface-wave magnitude to be more reliable of the two. Depths corrections suggested by Båth (1981) are applied.

Magnitudes of large shallow earthquakes from 1904 to 1980 (Abe, 1981)

Another trial to compile a uniform magnitude catalog was carried out by Abe who determined surface-wave magnitudes, M_S (eq.6), and broad-band body-wave magnitudes m_B for large ($m_B \geq 7$) shallow shocks during the period 1904-1980. More than 900 events enter the catalog. He calculated the magnitudes on the basis of amplitude and period data from various basic (original) materials. The M_S magnitude is the magnitude defined by Gutenberg, see eq.(6). Amplitudes of surface waves with periods 17-23 s measured on horizontal seismograms are employed. Vectorial summation of the two components or multiplication by $\sqrt{2}$, when only one horizontal component is available, is performed. Similarly, for m_B, formula (12) is applied. The difference, with respect to other m_b catalogs, here is that the average period of the body waves used in the magnitude m_B determination is about 9 s. To emphasize this difference a capital "B", instead of "b", in the subscript position is introduced. For the period 1904-1952 the best source of information are the unpublished worksheets of Gutenberg and Richter which lists amplitudes and periods at many stations throughout the world. The worksheets were carefully examined and M_S and m_B were redermined by using formulae (6) and (12), respectively. For the period 1953-1958, mainly Gutenberg's unpublished notes of amplitude and period data were used. For the years after 1958, bulletin data (amplitudes and periods) from more than 20 seismological bulletins were included in magnitude evaluations. From mid-1968, Earthquake Data Reports, published by the U.S. Department of the Interior, give surface-wave magnitudes on routine basis calculated from the Prague formula. Comparing surface-wave magnitude definitions by eq. (6) and eq.(14), at T=20 s, we see that M_S of EDR - 0.18 can be treated as being M_S of Abe's catalog. Determination of m_B ceased in 1974 since most of broad-band instruments were replaced by modern short-period, narrow-band seismographs.

Abe also performed comparison of his magnitudes M_S and m_B with magnitudes adopted in Seismicity of Earth, Earthquake Data Reports and the ISC Bulletin. He concludes that these magnitudes should not be compared directly due to the apparent inhomogeneity of the scales.

Seismological Bulletin, Uppsala

From January 1968, the so-called Uppsala magnitudes, m_b and M_S, are determined in full agreement with the Zurich recommendations. Both m_b and M_S listed in the bulletin are arithmetic means of station magnitudes measured at stations Uppsala and Kiruna. Starting April 1980, surface-wave magnitudes were determined from vertical-component readings, while horizontal-components (vectorial summation) were used prior to this date. Body-wave magnitudes are determined from short-period (~ 1 s) vertical-component *P*-wave measurements.

Uppsala Bulletins report systematically magnitudes since 1952. From 1952 to 1958, magnitudes were evaluated separately from Uppsala and from Kiruna measurements and from 1959 as a mean of the two measurements. Reference is made to formulae of Gutenberg and to Gutenberg and Richter but no details are given.

As a rule, the following waves were used prior to 1970: *P(H,Z,Z')*, *PP(H,Z,Z')*, *S(H)*, *R(H)*, where H = long-period horizontal component, Z = long-period vertical component, Z'= short-period vertical component. For larger earthquakes, m_b may thus be an average of 10-15 individual evaluations. From January 1970, the procedure has been simplified such that m_b is always an average of two determinations (*PZ'*), i.e. from Uppsala and Kiruna, just as M_S.

Acknowledgements

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APPENDIX (Båth, 1969)

I A S P E I

COMMITTEE ON MAGNITUDES

Meeting in Zurich on October 3, 1967

RECOMMENDATIONS

concerning magnitude determination for teleseisms ($\Delta \geq 20^{\circ}$):

1. Magnitudes should be determined from $(A/T)_{max}$ for all waves for which calibrating functions are available: PZ, PH, PPZ, PPH, SH, LH, (LZ).

2. Amplitudes and periods used should be published. Two magnitudes (m = body-wave magnitude, M = surface-wave magnitude) should be used. For statistical studies, M is favoured. The conversion formula $m = 0.56\ M + 2.9$ is recommended.

3. For body waves the Q-values of Gutenberg and Richter (1956) should be used. For surface waves, the Moscow-Prague 1962 formula $\sigma(\Delta) = \log (B/T) = 1.66 \log \Delta^{\circ} + 3.3$ should be used. Determinations of station and epicenter corrections are encouraged.

4. If short-period records are used exclusively, too low magnitudes result. In order to eliminate this error, it is strongly recommended that for short-period readings either A/T or Q be adjusted such that agreement with long-period instruments is achieved.