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DURATION MAGNITUDES FOR SWEDISH EARTHQUAKES

by

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Abstract

By calibration with the M_L-scale for Fennoscandian earthquakes, a regional magnitude formula for Swedish earthquakes based on signal duration is developed for each of the six stations of the Swedish Seismograph Station Network. Derived formulae are of the type $M_\tau = C_1 + C_2 (\log \tau)^2$ or $M_\tau = C_1 + C_2 (\log \tau)^2 + C_3\Delta$, where C_1, C_2 and C_3 are constants, specific for each station, τ is the recorded signal duration after the Sg-wave onset and Δ is the epicentral distance.

1. *Introduction*

The use of the duration of seismic signals recorded from near earthquakes for magnitude determination has become increasingly frequent in the last years. In the childhood of instrumental seismology, the duration of a recorded signal was used as an indication of the size of the earthquake. Later on, starting with RICHTER [21], magnitude scales were developed based on maximum trace or ground amplitudes of instrumentally recorded seismic waves. The application of signal length as an event-size indicator thereafter came out of fashion. BISZTRICSÁNY [6] developed duration-magnitude formulae for some European stations by relating coda lengths to amplitude-based magnitudes. The formulae concern surface waves from epicentral distances 4°–160°. For teleseismic events, amplitude-based magnitudes (m and M) are today still prevailing. SOLOV'EV [22] and TSUMURA [26] were the first to use Bisztricsány's method for near earthquakes. At present, the duration-magnitude method has its main field of application in the size determination of local (distances less than about 1°) or regional (distances less than about 10°) earthquakes. According to LEE and WETMILLER [18] and ADAMS [1], duration-based and amplitude-based methods to calculate the magnitude of near earthquakes are today approximately equally frequently applied.

2. *Present instrumental magnitude determination of Swedish earthquakes*

Swedish earthquakes are small and infrequent. They are rarely recorded at teleseismic distances. Therefore, an instrumental magnitude scale for regional earthquakes can not be established by reference to teleseismic magnitudes. BÅTH *et al.* [5] and WAHLSTRÖM [27] avoided this obstacle by calculating amplitude-distance relations for Sg, the wave that yields the largest amplitudes for Swedish earthquakes, and then applying the zero-magnitude definition of RICHTER's [21] concept, M_L.

3. *Observational data*

The time interval starting at the onset of Sg and ending at the disappearance of the coda in the background noise was measured at records from short-period, vertical-component instruments of the Swedish Seismograph Station Network (SSSN) for Swedish earthquakes 1970–1976. The Sg-duration is denoted by τ in this paper. Locations and instrument data of the stations are in Table I; locations are also plotted in Fig. 1. Measurements were made independently by three persons, Conny Holmqvist (C), Klaus Meyer (K) and the author (R). Only readings where the duration was measured as at least 10 s entered the set of data to be treated. Another criterion for admittance was, that M_L according to WAHLSTRÖM [27] could be calculated for the event, *i.e.* that at least one station at an epicentral distance greater than 100 km has recorded the event with a maximum trace amplitude of Sg of 0.20 mm or more.

Altogether 736 duration readings from 256 records from 66 earthquakes remained as the observational material. Earthquake epicentre locations are given in Table II and

Table I. Seismograph station data.

Station	Station coordinates		Seismograph	Seismometer period	Galvanometer period	Maximum magnification
Uppsala (UPP)	59.858°N	17.627°E	Benioff	1.0 s	0.7 s	40000
Kiruna (KIR)	67.840	20.417	Grenet	1.3	0.7	13310
Skalstugan (SKA)	63.580	12.280	Grenet	1.4	0.8	12040
Umeå (UME)	63.815	20.237	Benioff	1.0	0.7	75000
Uddeholm (UDD)	60.090	13.607	Benioff	1.0	0.7	75000
Delary (DEL)	56.472	13.868	Grenet	1.4	0.7	12990

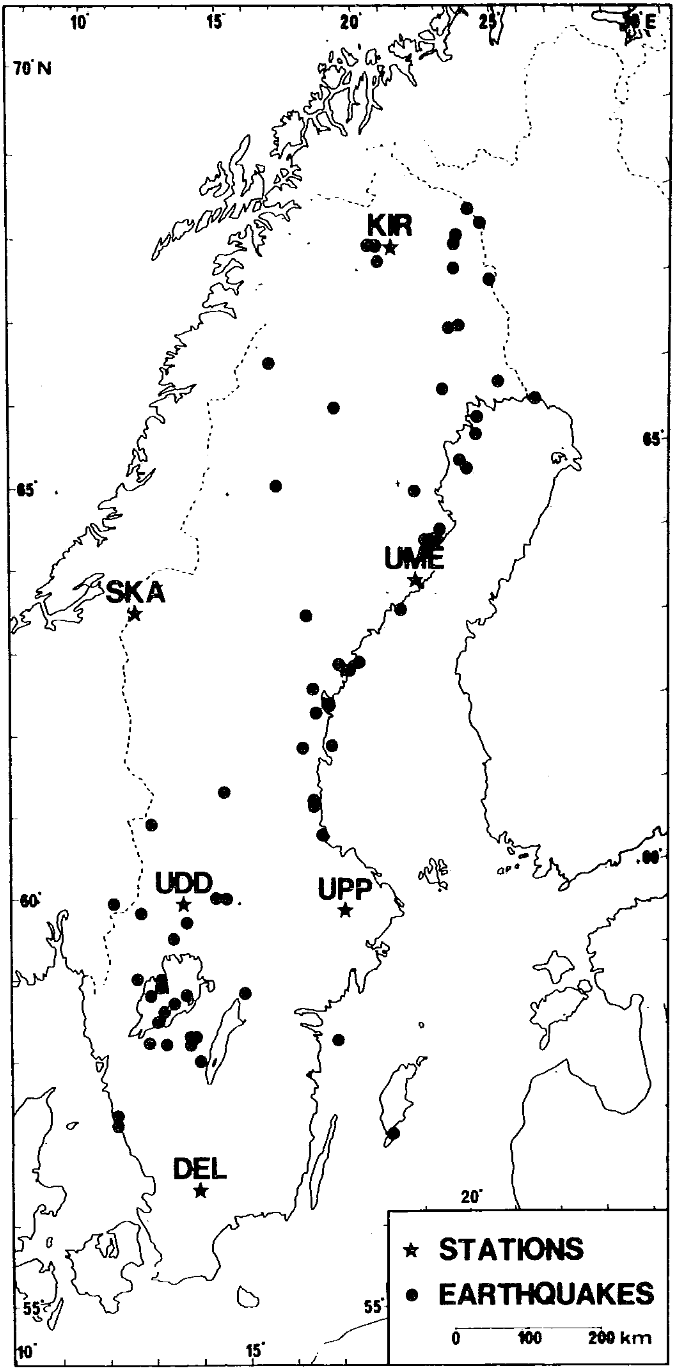

Fig. 1. Epicentre locations and station sites.

Table II. Earthquake locations and magnitudes.

No	Date	Origin time GMT	Epicentral coordinates		Magnitude, standard deviation and number of stations used M_L (Wahlström 1978)			M_T (present study)		
1	70 03 24	14 04 29	59.0°N	13.1°E	2.67	±0.17	3	2.62	±0.22	3
2	70 03 24	15 58 36	59.0	13.1	2.14		1	2.08		1
3	70 03 28	07 28 09	67.3	23.6	2.32	±0.04	2	2.52	±0.10	3
4	70 05 12	14 14 13	61.0	12.8	3.10	±0.17	6	3.09	±0.13	6
5	70 05 24	00 22 02	59.8	13.7	2.69	±0.14	4	2.64	±0.23	6
6	70 06 14	16 24 05	65.2	21.9	2.03		1	2.14	±0.09	3
7	70 06 14	17 33 52	65.1	22.1	2.92	±0.22	4	2.74	±0.16	5
8	70 08 12	19 28 46	65.1	16.4	2.90	±0.20	5	2.87	±0.11	4
9	70 09 28	04 17 59	60.1	14.7	1.62		1	2.10		1
10	70 10 02	15 16 22	58.3	17.3	2.34	±0.18	3	2.32		1
11	71 04 17	08 05 03	67.8	22.6	3.03	±0.17	4	2.95	±0.17	5
12	71 04 20	23 33 37	64.3	20.8	2.96	±0.15	4	2.86	±0.17	5
13	71 07 10	04 12 13	58.3	13.2	2.45	±0.42	4	2.58	±0.21	4
14	71 08 27	07 16 29	62.9	17.9	2.04	±0.21	2	2.21	±0.16	3
15	71 09 07	02 41 37	61.2	17.0	2.61	±0.16	2	2.58	±0.15	3
16	71 09 10	11 04 06	66.6	16.4	2.22	±0.11	2	2.45	±0.06	3
17	71 09 11	14 16 25	58.9	12.8	2.61	±0.35	4	2.60	±0.16	5
18	71 09 14	16 54 32	65.5	22.5	2.17		1	2.20	±0.13	2
19	71 10 07	09 33 08	58.6	13.0	2.77	±0.37	3	2.70	±0.19	2
20	71 11 25	13 46 50	66.8	22.1	2.71	±0.23	2	2.41	±0.12	2
21	72 03 06	16 03 04	64.9	20.5	1.75		1	2.33	±0.01	3
22	72 04 19	00 18 30	62.6	17.2	2.20	±0.10	2	2.13		1
23	72 06 12	04 31 33	60.1	14.5	2.72	±0.37	4	2.80	±0.14	5
24	72 08 20	02 52 35	61.9	16.8	2.89	±0.00	2	2.84	±0.25	4
25	72 09 04	00 26 33	57.1	18.4	2.31	±0.19	2	2.81	±0.13	2
26	72 09 25	02 55 34	58.9	13.7	2.70	±0.06	2	2.55		1
27	72 12 16	10 09 27	63.5	19.7	2.79	±0.22	4	2.83	±0.19	4
28	73 02 13	00 05 17	66.0	18.3	3.07	±0.19	4	2.96	±0.12	4
29	73 04 11	05 01 37	58.8	13.4	3.89	±0.26	5	3.69	±0.20	6
30	73 04 17	06 17 59	67.9	20.0	3.30	±0.09	4	3.07	±0.30	6
31	73 07 22	04 02 56	58.3	13.8	2.89	±0.56	4	2.96	±0.11	5
32	73 08 17	08 09 35	62.4	17.6	2.02	±0.07	2	2.14		1
33	73 10 01	16 44 19	60.0	11.9	2.92	±0.33	3	2.84	±0.18	5
34	73 11 26	21 45 38	62.9	18.5	3.17	±0.16	5	3.00	±0.12	5
35	74 01 13	18 27 08	63.5	17.1	2.56	±0.24	2	2.31	±0.05	4
36	74 02 05	22 33 59	58.1	14.0	2.36	±0.35	4	2.37	±0.25	2
37	74 03 02	17 16 23	60.8	17.2	2.15	±0.31	3	2.43		1
38	74 03 26	12 19 16	61.4	14.7	2.38	±0.14	4	2.34	±0.20	2
39	74 05 21	16 51 21	58.3	12.8	3.35	±0.27	6	3.46	±0.10	6
40	74 06 04	23 13 51	62.3	17.2	3.68	±0.10	5	3.39	±0.24	5
41	74 06 08	14 33 55	59.1	12.5	2.43	±0.20	4	2.55	±0.10	3
42	74 06 30	14 38 52	67.9	19.8	2.34		1	2.55	±0.20	3
43	74 07 05	16 19 50	57.3	12.1	3.01	±0.22	4	3.01	±0.10	5
44	74 09 12	00 32 06	66.1	21.7	2.55	±0.22	2	2.65	±0.12	2
45	74 09 27	17 36 36	64.3	20.6	2.76	±0.17	4	2.84	±0.14	5
46	74 10 28	21 56 29	57.4	12.1	2.70	±0.15	3	2.71	±0.30	3
47	74 12 01	19 35 58	67.8	20.1	3.00	±0.31	4	2.97	±0.19	5
48	74 12 09	19 08 24	66.1	23.4	2.16		1	2.22		1

cont...

Table II. Cont.

No	Date	Origin time GMT	Epicentral coordinates		Magnitude, standard deviation and number of stations used M_L (Wahlström 1978)			M_T (present study)		
49	75 08 11	18 28 09	67.5 °N	22.5 °E	3.80	±0.11	5	3.77	±0.26	5
50	75 08 29	04 42 24	65.8	24.4	2.34	±0.14	2	2.40		1
51	75 10 08	18 21 23	61.9	17.6	2.21	±0.32	3	2.46	±0.12	3
52	75 12 13	15 29 55	66.8	22.4	2.82	±0.07	4	2.65	±0.10	3
53	76 01 04	15 08 33	64.3	20.9	2.77	±0.02	2	2.34	±0.08	2
54	76 03 12	23 22 21	61.2	17.0	3.23	±0.23	6	3.24	±0.16	6
55	76 03 16	06 27 08	58.7	13.2	2.54	±0.20	3	2.63	±0.28	2
56	76 03 29	14 14 39	59.9	12.6	3.17	±0.40	4	3.13	±0.11	5
57	76 04 03	21 39 34	68.0	23.5	1.91		1	2.15		1
58	76 04 30	12 53 53	64.4	21.1	2.98	±0.42	4	2.90	±0.12	5
59	76 05 13	00 13 53	68.2	23.3	2.07	±0.47	2	2.05		1
60	76 07 03	07 26 01	58.4	13.8	2.85	±0.17	5	3.03	±0.13	5
61	76 07 12	03 29 50	62.8	18.2	2.27	±0.05	2	2.28		1
62	76 08 17	22 32 48	65.7	22.6	2.17	±0.11	2	2.38	±0.10	3
63	76 08 22	15 15 17	67.9	22.7	2.34		1	2.47		1
64	76 08 25	21 24 28	58.9	15.1	2.87	±0.34	5	3.06	±0.11	5
65	76 09 03	04 28 00	58.4	13.9	3.66	±0.20	6	3.56	±0.05	4
66	76 09 07	15 21 56	59.6	13.4	2.74	±0.22	4	2.83	±0.16	6

are plotted together with station locations in Fig. 1. Table II also gives origin times and M_L-values, with standard deviations, of the earthquakes. To estimate the accuracy of measured durations, which is due to the accuracy of measured coda ends, is difficult and, very likely, considerably varying from reading to reading. The estimated accuracy of calculated epicentral distances is ±10 km and the standard deviation of M_L, calculated from a single reading, is ±0.26 (WAHLSTRÖM, [27]).

4. *Factors of influence on recorded signal length*

There has been some discussion about the degree of influence that various factors have on the duration of recorded signals. The following résumé is based on the studies by LEE *et al.* [17], REAL and TENG [20], ELLIS and DAVIES [9], JACOB [12], LAHR *et al.* [14], LANGENKAMP and COMBS [15], AKI and CHOUET [3], HERRMANN [10], BAKUN and LINDH [4], PENNINGTON [19], SUHADOLC [23] and SUTEAU and WHITCOMB [24].

The instrument characteristics and the magnitude of the chosen cut-off level defining the coda end are obvious factors of great influence.

The station-site geology and the tectonic feature of the area are factors of relevance to signal duration. Other relevant factors are the degree of attenuation and the degree of scattering of the waves, and the thickness of the crust (longer »ringing» of thin crust at high frequencies – see JACOB, [12]). For large events, the local geology does not have severe influence on duration since, according to scattering theory, the last part of the coda consists of waves scattered from a large volume extending down into the upper mantle (BAKUN and LINDH, [4]).

The asymmetric distribution of energy from the source seems to be of minor importance and the only manifested source factor affecting signal length is the earthquake size, which is what we want to measure. The connection of the non-linear behaviour of the magnitude-duration relation, established for a given instrument, with source-spectrum variations due to the event size, will be discussed in the next chapter.

The epicentral distance has little effect on duration. Hence, for a station or a network covering events from a small area (epicentral distances up to 100–200 km), the distance-term of the magnitude-duration relation is often neglected.

LAHR *et al.* [14] found a systematic linear decrease of duration-based magnitude compared with amplitude-based magnitude with increasing focal depth. The difference is one magnitude unit over the depth range of the uppermost 150 km.

5. *Theoretical models for the magnitude-duration relation*

According to AKI [2] and AKI and CHOUET [3], the protracted coda recorded on seismograms consists of waves back-scattered from lateral heterogeneities (scatterers). In AKI's [2] theory, the seismic moment is expressed in terms of the density of the scatterers, the dominating wave frequency and the wave amplitude as functions of the time elapse from the earthquake origin, and other factors. By using Aki's expression and the relation between seismic moment and local Richter magnitude, M_L, SUTEAU and WHITCOMB [24] obtained a coda-magnitude formula. A particular of this, a duration-magnitude formula, is obtained by substituting for the time elapse from the earthquake origin an expression containing the signal duration, the hypocentral distance and the average P velocity along the travel path.

A theory describing the signal coda envelope as a function of exponential decrease with time was presented by HERRMANN [10]. His magnitude formula is of the type

$$\text{Magnitude} = C_1 + C_2 \log t + C_3 \Delta \qquad (1)$$

where t is the total signal length from the P onset, Δ is the epicentral distance, and C_1, C_2 and C_3 are constants. Empirical relations of this form are usual – see *e.g.* TSUMURA [26], LEE *et al.* [16], LEE *et al.* [17], REAL and TENG [20] and TENG *et al.* [25]. Regression analysis yields values of C_1, C_2 and C_3 for a given station or network of stations and a given seismic region. C_3 is generally small, and where short distances are considered, the Δ-term is usually neglected – see *e.g.* SOLOV'EV [22], CROSSON [7], HORI [11], ELLIS and DAVIES [9], JACOB [12], LANGENKAMP and COMBS [15], KORHONEN [13], BAKUN and LINDH [4], PENNINGTON [19] and SUHADOLC [23].

In the present study, the approach is similar to that of HERRMANN's [10]. For almost all regional earthquakes recorded by the SSSN, only Sg is recorded at most stations. The recorded signal may continue for many tens of seconds, sometimes several minutes. At stations at small epicentral distances and for large events also at large distances, Pg and sometimes other types of waves are recorded in addition to Sg. These records usually show a rapid decrease of Pg amplitude with time, and often the record is almost quiet during the time interval just before the onset of Sg. Again, the motion following the Sg-onset may last for minutes. These observations make clear, that what is recorded at the end of the total coda is evidently in some way connected with Sg, *e.g.* as scattered waves.

From these observations, a model of exponential decrease with time of the envelope of the coda following each recorded wave onset is plausible. Thus, each envelope can be expressed

$$E[t_w] = E[t_w = 0] \cdot (t_w + 1)^{-k_w} \tag{2}$$

where t_w is the time elapse after the onset of wave w, $E[t_w]$ is the envelope amplitude at time t_w and k_w is the exponent of decrease of the envelope starting at the onset of wave w. For all Swedish earthquakes, we further assume k_w to be constant for each station and each wave type, *i.e.* each k_w to be independent of source conditions, epicentral distance, station-to-source azimuth, path properties, etc.

The end of the total signal coda is assumed to be the end of the Sg-coda. This is the time instant, $t_{Sg} = \tau$, when the Sg-envelope has decreased to a certain level, LEV. We can write

$$\mathrm{LEV} = E[\tau] \tag{3}$$

τ is the duration, according to the notation convention in chapter 3. The assumption that the end of the Sg-coda is also the end of the total coda is valid for all Swedish earthquakes, except for near-surface events with recorded Rg-phase. These cases are excluded in this study, and later derived formulae should not be applied to these events.

WAHLSTRÖM [27] developed for Fennoscandian earthquakes, recorded by Finnish and Swedish stations, a magnitude formula that can be written as

$$M_L = \log A_{\max} + f_1(\mathrm{T}) + f_2(\Delta) \tag{4}$$

where A_{max} is the maximum recorded trace amplitude of Sg, T is the corresponding period and Δ is the epicentral distance. $f_1(T)$ reflects the response characteristics of the actual seismograph and of the Wood-Anderson seismograph.

The maximum Sg trace amplitude is usually within the very first swings of the wave train, as is also prescribed by the envelope hypothesis, *i.e.*

$$E[t_{Sg} = 0] = A_{max} \tag{5}$$

is a fair approximation.

From (2), (3), (4) and (5) it follows that, for $t_{Sg} = \tau$,

$$M_L = \log \mathrm{LEV} + k_{Sg} \log(\tau + 1) + f_1(T) + f_2(\Delta) \tag{6}$$

If we make the rough approximation, that $f_1(T)$ is constant for the period range of interest and if we further put, in conformity with the design of most duration-magnitude scales,

$$f_2(\Delta) = K_1\Delta + K_2 \tag{7}$$

where K_1 and K_2 are constants, then for large τ-values ($\log(\tau + 1) \approx \log \tau$), (6) is written as

$$M_L = k_1 + k_2 \log \tau + k_3\Delta \tag{8}$$

where k_1, k_2 and k_3 are constants. It should be mentioned here, that Swedish earthquakes are crustal phenomena, and there is no need for a term to normalize for focal depth fluctuations.

A formula analogous to (8), namely

$$M_L = k_1 + k_2 \log t + k_3\Delta \tag{8a}$$

where t denotes the total signal length, starting at the first recorded wave onset, has been used in many studies. This type of formula has been found to yield too small M_L for large events. REAL and TENG [20] besides a model of the type (8a) used a non-linear model of the type

$$M_L = l_1 + l_2 \log t + l_3\Delta + l_4 (\log t)^2 \tag{9}$$

suitable for application to earthquakes with $M_L > 3.8$. They explained the non-linearity in terms of the intrinsic behaviour of the M_L concept. In the present study, the approximation $f_1(T)$ = constant may be a corresponding cause of non-linearity.

HERRMANN [10] proposed models where k_2 of (8a) increases continuously or stepwise with increasing signal length, which again implies non-linearity of the magnitude-duration relation. Herrmann's explanation is in terms of the source-spectrum fluctuation with earthquake size combined with instrument response characteristics.

BAKUN and LINDH [4] tried various models for the magnitude-duration relation. They found, that for their set of data, a stepwise increase of k_2 of (8a) with duration makes the best fit.

From AKI's [2] scattering theory, SUTEAU and WHITCOMB [24] derived a term proportional to $\sqrt[3]{t}$ (in addition to a term proportional to t) to account for the non-linearity.

6. M_T-*scale for Sweden*

By (3) the coda end is defined as the time instant when the coda envelope falls below a specified level, LEV. In some respects this is an inconvenient criterion. Small events with no amplitude of the wave train exceeding the level must be disregarded. Events occurring at times of large background noise (above the level) must also be disregarded. Moreover, the level is defined with reference to a particular instrument type and sensitivity.

In many studies, *e.g.* SOLOV'EV [22], TSUMURA [26], CROSSON [7], REAL and TENG [20], KORHONEN [13] and SUHADOLC [23], the coda end is defined as the time instant when the signal-to-noise amplitude ratio equals one or when the signal disappears in the noise. The signals from regional events and the seismic noise usually have separated dominating periods at the instrument records used in the present work. Therefore, the seasonal or occasional noise level variations do not greatly influence coda length measurements, if the disappearance of the signal is used as coda-end criterion. With this criterion, (3) used in the derivation of (8) is in practice still valid. LEV then denotes the fairly constant level where the signal ceases to be detectable on the records. The ground amplitude corresponding to this level is different for different instruments. Considering this and the effect of local geology on signal duration, agreed upon in many studies, *e.g.* LEE *et al.* [17], REAL and TENG [20], ELLIS and DAVIES [9] and HERRMANN [10], it is adequate to treat each station separately.

According to scattering theory, it is dangerous to apply a duration magnitude for very small events. For these events, the end of the coda might not consist of scattered waves or the dominating recorded frequency might be greater than the source-spectrum corner frequency (see SUTEAU and WHITCOMB, [24]). In the present study, very small events are excluded by permitting only duration readings equal to or greater than 10 s (chapter 3).

For our case, (9) is rewritten so that

$$M_L = l_1 + l_2 \log \tau + l_3 \Delta + l_4 (\log \tau)^2 \tag{9a}$$

For each of the 6 stations (Table I), 8 various populations of coefficients of (8) and (9a) were obtained. Each element of Set 1 below has been combined with each element of Set 2 and stepwise multiple regression analysis according to DRAPER and SMITH [8] (IMSL-subroutine RLSEP) has been applied. The significance level α = 0.05 was used in all regressions.

Set 1 a: M_L calculated from the actual station [M_{LIND}]
b: M_L calculated from all available Swedish station readings (*i.e.* M_L in WAHLSTRÖM, [27]) [M_{LSUM}]

Set 2 a: τ measured by person C
b: τ measured by person K
c: τ measured by person R
d: $\tau = \frac{1}{3}(\tau_C + \tau_K + \tau_R)$ [τ_{MEAN}]

Notation in brackets refer to later text. τ is in s and Δ in km.

There is, for most cases, little difference in values obtained from duration measurements done by various persons, although considerable variations occurred in individual cases. Coefficient standard deviations are generally smallest when τ_{MEAN}-data are used. Furthermore, it is least subjective to base a magnitude-duration scale on τ_{MEAN}, where the personal bias of persons C, K and R is averaged.

The differences in values of coefficients obtained for various stations are usually small. No major division into a Grenet and a Benioff group can be legalized from the results.

(6) and thus (8) were derived from the assumption, that the exponential decrease of the Sg-coda envelope with time starts from the maximum recorded amplitude. Coefficient standard deviations are generally smaller when M_{LSUM} data are used than when M_{LIND} data are used. It therefore seems appropriate to assume, that the exponential decrease starts rather from a directionally averaged maximum amplitude, *i.e.* the amplitude that would have been recorded for spatially uniform source energy distribution. $E[t_w = 0]$ in (2) would then be related to this averaged amplitude.

The regression solutions show, that k_2 and k_3 are significant for around 80 % of the 48 solutions of (8). For the remaining cases, the Δ-term is insignificant. Around 55 % of the 48 solutions of (9a) show significance for l_3 and l_4 but not for l_2, and around 30 % of the cases show significance for l_4 alone. Thus, the non-linearity of the magnitude-duration relation found in many studies (see chapter 5) is indicated. The most common forms of the obtained relations are (M_L is now replaced by M_τ)

$$M_\tau = k_1 + k_2 \log \tau + k_3 \Delta \tag{10.1}$$

$$M_\tau = l_1 + l_3 \Delta + l_4 (\log \tau)^2 \tag{10.2}$$

$$M_\tau = l_1 + l_4 (\log \tau)^2 \tag{10.3}$$

That Δ-terms are in most cases significant could be expected for the large distance ranges of several hundred km. (As distinguished from amplitude-distance relations, we need here not fear to include very small distances.)

Summarizing, preferred solutions are those where regressions are performed using (9a) and the variables M_{LSUM} and τ_{MEAN}. Final duration-magnitude formulae obtained for the six stations read

$$\text{UPP:} \quad M_\tau = 2.20 + 0.22 (\log \tau)^2 \tag{11.1}$$

$$\text{KIR:} \quad M_\tau = 1.42 + 0.28 (\log \tau)^2 + 0.84 \cdot 10^{-3} \Delta \tag{11.2}$$

$$\text{SKA:} \quad M_\tau = 1.56 + 0.29 (\log \tau)^2 + 0.73 \cdot 10^{-3} \Delta \tag{11.3}$$

$$\text{UME:} \quad M_\tau = 1.49 + 0.27 (\log \tau)^2 + 0.90 \cdot 10^{-3} \Delta \tag{11.4}$$

$$\text{UDD:} \quad M_\tau = 1.43 + 0.27 (\log \tau)^2 + 0.89 \cdot 10^{-3} \Delta \tag{11.5}$$

$$\text{DEL:} \quad M_\tau = 2.22 + 0.22 (\log \tau)^2 \tag{11.6}$$

The event magnitude, M_τ, is calculated as the mean of obtained M_τ-values for various stations. Event magnitudes, with standard deviations calculated analogically to (18) in WAHLSTRÖM [27], are given in Table II.

7. *Amplitude-based or duration-based magnitude?*

To base a magnitude concept on signal duration instead of maximum amplitude has many advantages. The limited dynamic range of an instrument puts restrictions on the application of an amplitude-based magnitude. For large events, trace amplitudes may be oversaturated and thus impossible to measure. The same may be the case for small events recorded at small distances. It may further be difficult to resolve the high-frequency coda, and thus to attain the correct magnification for instruments with high-gradient amplitude response in the period domain of interest.

Not only the earthquake size but also the directional variation in attenuation and the fault dip orientation relative to the station effect the wave amplitudes. The signal length is, however, almost insensitive to the two last factors.

If the coda end is defined by means of cut-off level, then difficulties with small events and varying noise level arise, as mentioned in chapter 6. If cut-off level is not used, then other difficulties arise due to the subjectivity in the assignment of the signal end (see REAL and TENG, [20]), which includes the personal bias. Measurements by various persons are in the present study found to vary considerably for individual duration readings – see also TSUMURA [26] – but obtained magnitude-duration relations for various persons agree fairly well.

Duration-magnitude scales also have the drawback, that for multiple events or an aftershock sequence, codas from more than one earthquake may interfere. Often the maximum amplitude of each shock can be separated and measured in these cases.

The utilization of an amplitude-based magnitude scale should not exclude the utilization of a duration-based scale, and vice versa. It is assumed that both scales are calibrated from one and the same reference scale or either of the scales is calibrated from the other, as is the case for the two Swedish scales. Instead, the scales could be used as complements.

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