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2. *Construction of Amplitude Curves and Determination of Station Corrections for a Seismic Network Located in the Epicentral Area.*

(Based on the data of Wakayama Micro-earthquake Observatory and its Sub-stations)

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1. Introduction

At present the investigation of seismic wave amplitudes near the source is of great practical importance. The normalized amplitude-distance curve is the basis for a magnitude determination of local micro-earthquakes as well as for studying the attenuation of seismic body waves in the epicentral area. The construction of an amplitude curve for short epicentral distances is usually accompanied with the same difficulties connected with determination of the station corrections and normalization of the curve itself.[1] Hereafter, an attempt is made for the amplitude curve of the maximum recorded velocity amplitude for Kii Peninsula, southwestern Honshu, to be plotted by the records from Wakayama Micro-earthquake Observatory and its sub-stations.

The method used does not require any additional information, as for instance, data for the magnitude of earthquakes, no knowledge in advance being required as far as the station correction is concerned.

2. Method for Determination of the Station Corrections

It is clear that, not having the determination of the true values of the station corrections, one cannot successfully construct the amplitude curve, especially for short epicentral distances. As for Kii Peninsula no calibrating function for the body waves exists, we applied a special method for the determinations of the station corrections. The principal idea of this method consists in the following. Let us consider,

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1) L. Christoskov, "Calibrating Functions for PH- and SH-waves at epicentral distances up to 21°", *Bull. Geophys. Inst. Bulg. Acad. Sci.*, **10** (1967), 85-98 (in Bulgarian).

for example, the case when two seismic stations, denoted as i-th and k-th station, are located at one and the same distance $\Delta=\Delta_i=\Delta_k$ from an epicenter E (Fig. 1). If we accept station i as a basic one, for a given type of seismic wave, the difference in the logarithm of the maximum ground velocity amplitude measured at station i and k respectively will be equal to the station correction S_k for station k with respect to the basic station for which the correction is $S_i=0$, *i.e.*,

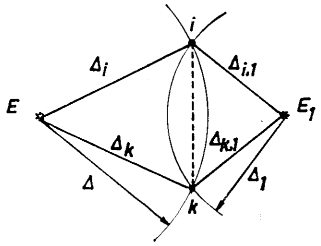

Fig. 1. Location of the epicenter (E) and stations i and k for determination of $\delta A_{i,k}$ in equation (1).

$$S_k=\delta A_{i,k}=\log\left(\frac{A}{T}\right)_i-\log\left(\frac{A}{T}\right)_k . \tag{1}$$

The value S_k should not be affected by the shape and the gradient of the amplitude-distance curve as the determination is made for a certain constant distance Δ, but it will depend on such factors as the mechanism of the source and the different path of the seismic waves to these stations. Having a set of epicenters on condition that Δ_i is kept equal to Δ_k (for example E and E_1 in Fig. 1), one can determine $\overline{\delta A}_{i,k}=\overline{S}_k$ as a mean value from all the observations eliminating the influence of these factors.

In the case of the observation network of n stations it is reasonable to calculate the relative correction between every two stations i and k as a mean value from all possible and independently calculated differences $\overline{\delta A}_{i,j}$ and $\overline{\delta A}_{k,j}$ by the formula

$$\delta S_{i,k}=(n-1)^{-1}\left(\sum_{j=1}^{n}(\overline{\delta A}_{i,j}+\overline{\delta A}_{j,k})-\overline{\delta A}_{i,k}\right) , \tag{2}$$

where $k>i$; $\delta A_{i,i}=\delta A_{k,k}=0$ and n is the total number of the stations. If again the i-th station is the basic one, the station correction for k-th station will be

$$S_k=(n-1)^{-1}\left(\sum_{j=1}^{n}(\delta S_{i,j}+\delta S_{j,k})-\delta S_{i,k}\right) , \tag{3}$$

$$\left.\begin{array}{ll} & \delta S_{1,2}+\delta S_{2,3}+\cdots+\delta S_{n-2,n-1}+\delta S_{n-1,n}=\delta S_{1,n} , \\ \text{or} & \delta S_{1,2}+\delta S_{2,3}+\cdots+\delta S_{n-1,n}+\delta S_{n,1}=0 . \end{array}\right\} \tag{4}$$

When equation (4) is not satisfied, as for the usual geodetic surveying we can recalculate $\delta S_{i,k}$ and S_k from (2) and (3) getting a new approximation for these quantities aiming to reduce as much as possible the error of closure.

This procedure was applied for all the stations of the Wakayama Observatory network, except Kinomoto station, using the data for the maximum recorded velocity amplitude in μ kine published in the Bulletin[2] as A/T. Probably for the first hundred kilometres from the focus this maximum amplitude is typical for S_g wave.

It was accepted that two stations are at one and the same distance from a certain epicenter if $\delta t = t_s - t_p$ differs less than 3%. In the calculation we deal with 39 independent means of values $\overline{\delta A_{i,k}}$ (i and k correspond to two different station numbers—see Table 1) for the result of about 500 single observations of the value $\delta A_{i,k}$. The standard deviation in the mean values $\delta A_{i,k}$ has been varying around 0.05. The final corrections (the third approximation) with respect to the basic station Wakaura as well as the first approximation, are presented in Table 1. As it is to be seen, there is a correlation tendency between

Table 1

No.	Station Name	Foundation	Station Correction Approximation		
			First	Third-Fourth	Final
1	Wakaura (Wk)	Crystalline schist	0.000	—	0.00
2	Todoroki (Td)	〃 〃	−0.223	−0.239	−0.24
3	Arida (Ar)	〃 〃	0.000	−0.010	−0.01
4	Hidaka (Hd)	Mesozoic shale	−0.327	−0.321	−0.32
5	Kainokawa (Kk)	〃 〃	−0.500	−0.487	−0.49
6	Shichikawa (Sk)	Paleogene shale	−0.396	−0.389	−0.39
7	Sarutani (St)	Mesozoic shale	−0.379	−0.379	−0.38
8	Haibara (Hb)	Granitic rock	−0.141	−0.146	−0.15
9	Iinan (In)	River terrace	−0.405	−0.368	−0.37
10	Ise (Is)	Serpentine	+0.075	+0.004	0.00
11	Gozaisho (Gz)	Granite	+0.016	+0.058	+0.06

the obtained station corrections and the particular rock type on which the stations are situated.

This result is necessary to be considered as a first step in the determination of the station corrections for Wakayama Observatory network, the revision in the future based on the large number of data may be advisable.

3. Amplitude-Distance Curve

For constructing the amplitude-distance curve, if a normalizing set

2) Seismological Bulletins of Wakayama Micro-earthquake Observatory and its Substations, Jan.-June, July-Sept., Oct.-Dec. 1965, (1966, 1968, 1969 .

of earthquakes with known magnitudes defined by an external source does not exist, one can use the gradient of the amplitude curve itself. For a network of near stations we have simultaneous records of the maximum ground velocity amplitude at several stations in a sufficiently small distance interval $d\Delta$ for every earthquake. Knowing the station correction and correcting the individual observed values $\log(A/T)$ at every station for every earthquake, it is possible to calculate the gradient α of the amplitude curve (see for instance[3]):

$$\alpha = da/d\Delta \ ,$$

where $a=\log(A/T)$ and its values coordinated to the mean Δ in the interval $d\Delta$. For two stations i and k, practically $da = \log(A/T)_i - \log(A/T)_k$, $d\Delta = \Delta_i - \Delta_k$ and $\Delta = (\Delta_i + \Delta_k)/2$. Having continuous data for α with the distance, we can construct the gradient curve $\alpha = \alpha(\Delta)$. Such a kind of gradient curve for the maximum ground velocity amplitude for S_g wave (vertical component) is represented in Fig. 2, together with the observed individual values for α. The final curve is the result of connecting the mean values for α within distance intervals of 2.5 km ($\Delta \leqq 75$ km) and 5.0 km ($\Delta > 75$ km). As can be seen up to 50 km the gradi-

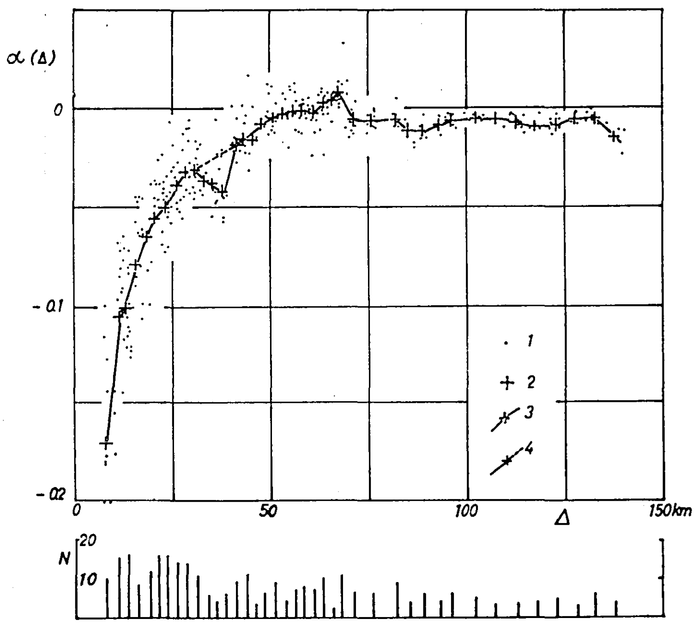

Fig. 2. The gradient curve for Kii Peninsula.
1-observed values for α; 2-mean gradient for certain distance range; 3-gradient curve (variant a); 4-gradient curve (variant b); N-number of data.

3) J. VANĚK, "Amplitude Curves of Seismic Body Waves and the Structure of the Upper Mantle in Europe", *Tectonophysics*, **5** (1968), 235-243.

ent is very big in comparison with its value for bigger distances.

By integrating (graphically) the gradient curve we can obtain an approximation of the amplitude-distance curve

$$A^*(\Delta)=\int\alpha(\Delta)d\Delta\ , \tag{6}$$

In this way the amplitude curve as a result of data from the given observation network only can be plotted. The integrated curve for Kii Peninsula by step of 2 km and initial condition that $A^*=3.0$ for $\Delta=7.5$ km is shown in Fig. 3. The variants a and b correspond to the two possibilities for interpreting the gradient curve at distance range 30—40 km, besides, the variant b is more probable because we cannot expect a big oscillation of the gradient near the source within a very small epicentral distance range. As can be seen, for distance $\Delta<50$ km the amplitude curve decreases with Δ rather quickly as an exponential function (dotted curve in Fig. 3)

$$A^*(\Delta)=A_\beta{}^*e^{-\beta\Delta}\ , \quad (7.5<\Delta<50) \tag{7}$$

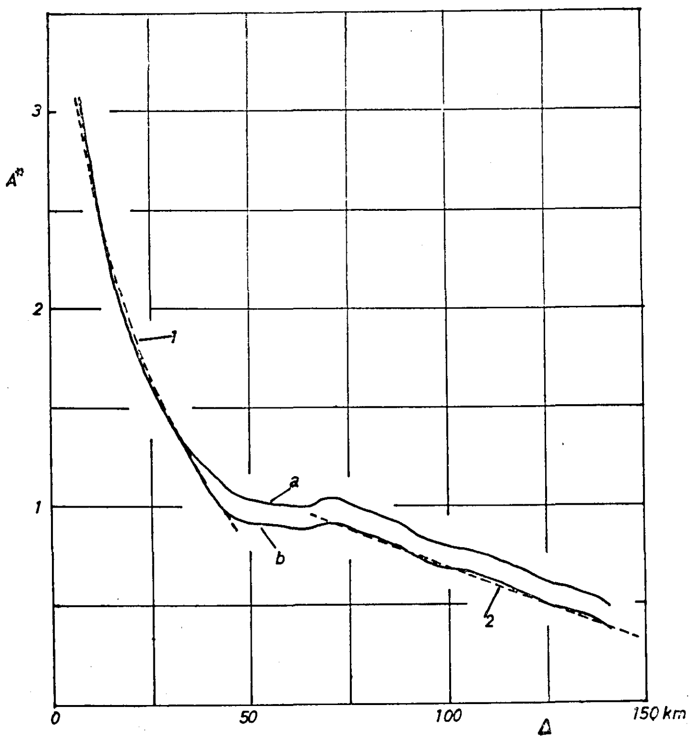

Fig. 3. Amplitude-distance curve for Kii Peninsula (integrated gradient curve) a-variant a; b-variant b; 1-approximation according to equation (7); 2-approximation according to equation (8).

where A_β^* and β were found to be 3.58 and 0.0306 respectively.
Between 50 and 70 km the amplitude curve is changing very slowly, and for distance $\Delta \geqq 70$ km it decreases as a quasi-linear function of the distance (dotted line in Fig. 3.)

$$A^*(\Delta) = A_b^* - b\Delta\ , \quad (\Delta \geqq 70) \tag{8}$$

where $A_b^* = 1.45$ and $b = 0.757 \times 10^{-2}$ are determined also by the method of least squares.

The observed difference in the attenuation of the maximum ground amplitude with the distance is an indication for the probable change of the seismic wave. After 70 km it is possible that the maximum amplitude is connected with some types of short-period surface waves.

The problem of the absolute normalization of $A^*(\Delta)$ will be discussed hereafter.

4. Calibrating Function and its Normalization

As already known, the calibrating function $\sigma(\Delta)$ is

$$\sigma(\Delta) = -(A^*(\Delta) + \delta A^*)\ , \tag{9}$$

where δA^* is a constant reducing the level of $A^*(\Delta)$ to the amplitude curve $A_0^*(\Delta)$ for an earthquake with magnitude zero, *i. e.*

$$A_0^*(\Delta) = A^*(\Delta) + \delta A^*\ . \tag{10}$$

To find the constant δA^* we used the data obtained in (4) for the relation between the maximum velocity amplitudes and magnitude M_{F-P} determined by the total duration of oscillation for Shichikawa and Ise stations.

From the basic magnitude equation the calibrating function $\sigma(\Delta)$ is defined as

$$\sigma(\Delta) = M - \log\left(\frac{A}{T}\right) - S\ , \tag{11}$$

where again S is the station correction, for station Shichikawa ($M=2.0$) at $\Delta = 80$ km we have $\sigma_{80} = 0.37$ and for Ise ($M=2.0$) at $\Delta = 150$ km, $\sigma_{150} = 0.88$ or as observed one relative difference of -0.51 of the magnitude unit. The corresponding value for the obtained amplitude curve is 0.53, which is a good coincidence. Thus, we know the absolute level for $\sigma(\Delta)$ or $A_0^*(\Delta)$ in two points (even one point will be sufficient, be-

4) K. TSUMURA, "Determination of Earthquake Magnitude from the Total Duration of Oscillation", *Bull. Earthq. Res. Inst.*, **15** (1967), 7-18.

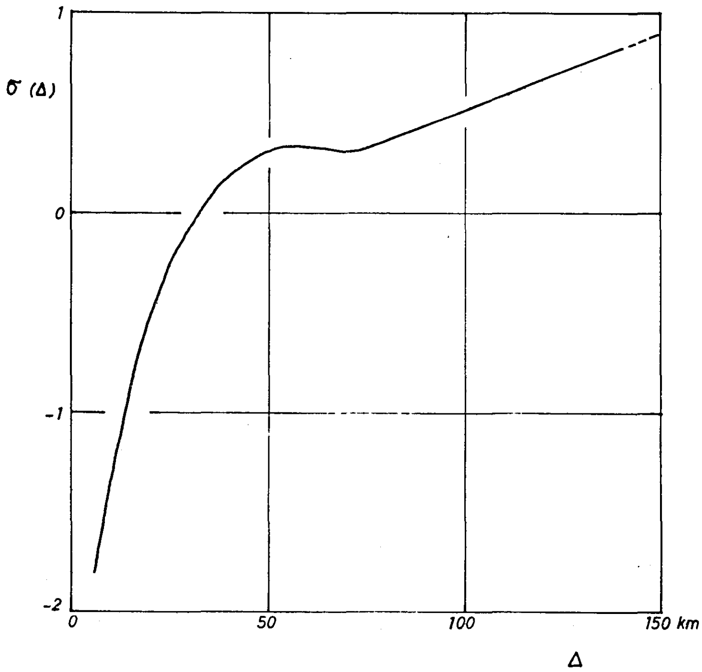

Fig. 4. Smooth calibrating function (variant b).

Table 2

Δ km	$\sigma(\Delta)$	Δ km	$\sigma(\Delta)$	Δ km	$\sigma(\Delta)$
8	−1.67	42	0.21	76	0.33
10	−1.39	44	0.24	78	0.35
12	−1.16	46	0.26	80	0.36
14	−0.96	48	0.28	85	0.40
16	−0.79	50	0.30	90	0.44
18	−0.64	52	0.31	95	0.48
20	−0.52	54	0.31	100	0.52
22	−0.41	56	0.32	105	0.55
24	−0.31	58	0.32	110	0.59
26	−0.22	60	0.32	115	0.63
28	−0.15	62	0.33	120	0.67
30	−0.09	64	0.33	125	0.71
32	−0.03	66	0.32	130	0.74
34	0.03	68	0.30	135	0.78
36	0.09	70	0.29	140	0.85
38	0.13	72	0.30	145	(0.85)
40	0.17	74	0.32	150	(0.89)

cause we are not in doubt about the actual shape of the curve $A^*(\Delta)$) and the constant δA^* was defined as -1.33 for variant b (Fig. 3).

In Fig. 4 is represented the smoothed calibrating function $\sigma(\Delta)$ (variant b) and its numerical values are given in Table 2.

A simple check-up of the station corrections S and the calibrating function $\sigma(\Delta)$ could be carried out by one comparative study of the magnitudes M_{F-P}[4] and the magnitudes determined by the formula

$$M=\log\left(\frac{A}{T}\right)+\sigma(\Delta)+S\,, \tag{12}$$

where (A/T) is the maximum recorded velocity amplitude measured in μ kine (10^{-3} cm.sec.$^{-1}$). For several earthquakes the values M_{F-P} and M (with and without station corrections) are given together in Fig. 5; Δ corresponds to the epicentral distance of the station for which the determination has been made.

This simple comparison shows that no principal difference between

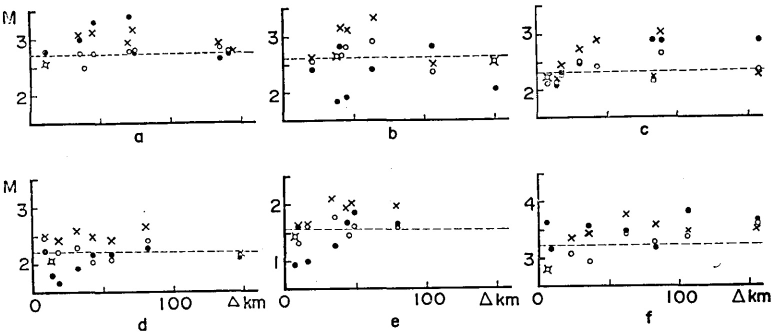

Fig. 5. Comparison of M_{F-P} (full circle), M according to equation (12) (open circle) and $M^*=\log A/T+\delta(\Delta)$ (cross) for several earthquakes recorded at Wakayama Observatory network during 1965 (see also Table 3).
a-July 6, 08:15:36.53; b-July 15, 16:56:06.02; c-August 4, 15:22:01.45; d-August 5, 02:38:57.72; e-August 21, 16:44:68.83; f-August 31, 17:31:12.55.

Table 3

Earthquake	M			M_{F-P}			M^*		
	mean value	number of data	square error	mean value	number of data	square error	mean value	number of data	square error
a	2.75	8	0.12_4	2.98	7	0.27_0	2.95	8	0.21_1
b	2.66	7	0.18_8	2.33	7	0.38_5	2.85	7	0.31_5
c	2.32	8	0.17_0	2.45	7	0.30_6	2.52	8	0.32_4
d	2.22	8	0.15_4	2.06	8	0.21_5	2.43	8	0.22_3
e	1.56	7	0.13_4	1.42	7	0.34_1	1.82	7	0.23_1
f	3.21	7	0.26_8	3.50	7	0.22_6	3.45	7	0.30_0

M_{F-P} and M with the distance and the magnitude itself is observed. The scattering of the data around the mean value does not exceed 1/4 of the magnitude unit and it is of the same order as the deviations typical for the magnitude determination in general. (*cf.* Table 3)

5. Summary and Conclusion

In this study an attempt is made to determine the station corrections and to plot the amplitude-distance curve for maximum velocity amplitudes at short epicentral distances, besides, the method used does not require information from external sources and is based only on the data obtained at these stations. A mutual correction and checkup of the station corrections is possible, if we draw into the calculation all observed individual corrections between every two stations of a certain observation network. This procedure increases the accuracy of the station correction determinations, which is of great importance for constructing the gradient curve at small epicentral distances. The amplitude-distance curve is obtained by integration of the gradient amplitude curve. Its absolute normalization is made on the grounds of the relation between maximum ground velocity amplitude and the magnitude determined by the total duration of oscillation.

The numerical results obtained for the Wakayama Observatory network have to be assumed as a first step or approximation which should be followed up by a more detailed study. Nevertheless, the magnitude determined by the maximum ground velocity amplitude (equation (12)) could be used in practice, for statistical study of the energy distribution in the epicentral area. A future construction of the gradient and amplitude curves for *P*-wave and an investigation of the azimuth influence (or effect) on the shape of the amplitude curves will be very useful in the practice.

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2. 震央域にある地震観測網に対する振幅減衰曲線と観測点補正値
(和歌山微小地震観測所とその衛星観測点の資料にもとづく)

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和歌山微小地震観測所およびその衛星点で観測されその観測報告[2]に発表されている局地地震の最大地動速度振幅値の資料をもちい，まず各観測点に対する観測点補正値を和歌浦観測点を基準としてもとめた（第1表）．これは動木 Td を例外として，ほぼ観測点のある地盤の地質のかたさの順になつている．

つぎにこの補正値を採用し，振幅減衰の勾配をもとめ (Fig. 2) それを数値積分して振幅減衰曲線をもとめた．(Fig. 3) これによつてマグニチュードを計算するための calibrating function $\sigma(\Delta)$ がえられ（第 2 表），紀伊半島の局地地震のマグニチュードをもとめることができるようになつた．そこで例として若干の地震についてこの方法でえたマグニュチードと観測点補正値をくわえないでえたマグニチュードを計算し，これらと津村[4]によりだされた全振動継続時間 $F-P$ によりもとめたマグニチュードとを比較し，観測点補正値をいれたものは $F-P$ によるものとまず一致しているといえることがわかつた．(第 3 表および第 5 図)

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