EARTHQUAKE MAGNITUDE, INTENSITY, ENERGY, AND ACCELERATION

(Second Paper)

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ABSTRACT

This supersedes Paper 1 (Gutenberg and Richter, 1942). Additional data are presented. Revisions involving intensity and acceleration are minor. The equation $\log a = I/3 - \frac{1}{2}$ is retained. The magnitude-energy relation is revised as follows:

$$\log E = 9.4 + 2.14\,M - 0.054\,M^2 \qquad (20)$$

A numerical equivalent, for M from 1 to 8.6, is

$$\log E = 9.1 + 1.75\,M + \log(9 - M) \qquad (21)$$

Equation (20) is based on

$$\log(A_0/T_0) = -0.76 + 0.91\,M - 0.027M^2 \qquad (7)$$

applying at an assumed point epicenter. Eq. (7) is derived empirically from readings of torsion seismometers and USCGS accelerographs. Amplitudes at the USCGS locations have been divided by an average factor of $2\frac{1}{2}$ to compensate for difference in ground; previously this correction was neglected, and $\log E$ was overestimated by 0.8. The terms M^2 are due partly to the response of the torsion seismometers as affected by increase of ground period with M, partly to the use of surface waves to determine M. If M_S results from surface waves, M_B from body waves, approximately

$$M_S - M_B = 0.4\,(M_S - 7) \qquad (27)$$

It appears that M_B corresponds more closely to the magnitude scale determined for local earthquakes.

A complete revision of the magnitude scale, with appropriate tables and charts, is in preparation. This will probably be based on A/T rather than amplitudes.

Introduction

The present purpose is primarily to revise and extend an earlier investigation (Gutenberg and Richter, 1942; Paper 1) which dealt chiefly with the relations of earthquake magnitude to energy release, and of intensity to acceleration. In this revision we shall not attempt to consider further the effect of variable hypocentral depth, and shall limit the discussion to shocks in the California region. New data, chiefly for earthquakes since 1941, are presented; the data of Paper 1 are used but not repeated.

Notation

A = maximum ground amplitude of the surface (cm., unless otherwise noted)
a = maximum ground acceleration (cm/sec.^{2} = gals)
B = seismographic trace amplitude (mm.)
b = value of B for a shock of magnitude zero
D = hypocentral distance (km.)
Δ = epicentral distance (km.)
Θ = epicentral distance in degrees
E = energy radiated in elastic waves (ergs)
h = hypocentral depth (km.)
I = seismic intensity on the Modified Mercalli Scale of 1931 (Wood and Neumann, 1931)
k = coefficient of absorption
λ = wave length (km.)

M = earthquake magnitude
M_B = magnitude calculated from body waves in teleseisms, M_S from surface waves
n = number of waves in maximum group
N = number of observations
q = $\log (A/T)$ with A in microns
R = radius of the earth (km.)
r = value of Δ at limit of perceptibility
ρ = density (gm/cm.^{3})
T = period of vibration (sec.)
τ = time
t = duration of maximum wave group
USCGS = United States Coast and Geodetic Survey
V = static instrumental magnification
v = wave velocity (km/sec.)
The zero subscript refers to the value of the respective quantity at the epicenter.
log = common logarithm (base 10)

Materials Used

This paper employs data of the same type as those of Paper 1, including later issues (through 1952) of the series "United States Earthquakes" published by the USCGS, and preliminary mimeographed bulletins giving material of the same character.

The strong-motion instruments operated at Pasadena during the later interval have periods of near 8 seconds (not 10 sec.).

Definition and Calculation of Magnitude

The original definition of magnitude (Richter, 1935) may be stated in the form that for two earthquakes at a given epicentral distance Δ, and for the maximum recorded trace amplitude,

$$M_1 - M_2 = \log B_1 - \log B_2 \tag{1}$$

with the following specifications:

1) The definition applies strictly only for $\Delta = 100$ km. That the equation is applicable at other distances is a hypothesis, to be confirmed by observation. It holds at least for shocks of M not over 6 recorded at distances less than 1,000 km. A fuller discussion is given at the end of this paper.

2) The maximum amplitudes denoted by B_1 and B_2 are recorded by a standard horizontal-component torsion seismometer (free period 0.8 sec., $V = 2{,}800$, damping ratio approximately 50 : 1). Theoretically, any well-calibrated seismometer could be used, by first evaluating the true ground motion for the whole seismogram and then computing what the corresponding maximum deflection of the standard torsion seismometer should be. However, it would be erroneous to assume without other verification that the wave which appears as the maximum on the given seismogram of nonstandard type is the wave which would write the maximum amplitude on the standard seismogram. This is very likely to be wrong if the seismometer used has a free period much longer than 1 second.

3) The zero of the scale is fixed by setting $M = 3$ when $B = 1$ mm. at the standard distance of 100 km.

4) The definition refers to local shocks of ordinary character in southern California. This implies hypocenters at the generally prevailing depth (now believed to

be about 16 km.), and postulates that no exceptional structures or materials are involved.

5) In reading the maximum recorded amplitude on a single seismogram, the amplitude is taken as the half range, which is the mean between successive deflections in opposite directions. In combining the two horizontal components the mean of the two maxima is taken. This procedure applies only to local earthquakes, for which any other rule would soon encounter practical difficulties. Exceptionally, the maximum apparent amplitude on the seismogram in one component may be in the P phase. This should be ignored.

These specifications are supplemented by certain procedures which complete the working definition of magnitude.

TABLE 1

LOGARITHMS OF THE AMPLITUDES B (IN MM.) WITH WHICH THE STANDARD TORSION SEISMOMETER SHOULD REGISTER A SHOCK OF MAGNITUDE ZERO

Δ	$-\log B$	Δ	$-\log B$	Δ	$-\log B$
0	1.4	100	3.0	330–340	4.2
10	1.5	110–120	3.1	350–370	4.3
20	1.7	130–140	3.2	380–390	4.4
25	1.9	150–160	3.3	400–420	4.5
30	2.1	170–180	3.4	430–460	4.6
35	2.3	190–200	3.5	470–500	4.7
40	2.4	210	3.6	510–550	4.8
45	2.5	230–240	3.7	560–590	4.9
50	2.6	250–260	3.8	600	5.1
60–70	2.8	270–280	3.9	700	5.2
75–85	2.9	290–300	4.0	800	5.4
90	3.0	310–320	4.1	900	5.5
				1,000	5.7

1) Reduction to $\Delta = 100$ km. employs an empirically determined table for $\log b$ as a function of Δ (table 1). For 600–1,000 km., this includes data collected by Mr. Lomnitz for earthquakes in the Gulf of California recorded at stations in southern California. For earthquakes off the northern California coast, recording in the same range of distance, he finds slightly lower values. Beyond 200 km. the data of table 1 are represented closely by the inverse cube law $\log b = 3.37 - 3 \log \Delta$.

2) For distances less than 25 km., and for large shocks, magnitude assignment frequently requires the use of short-period motion recorded by strong-motion instruments. This is discussed below in connection with amplitudes.

3) In combining readings from different stations, the corrections given in table 3 are applied. These have been revised by Mr. Cinna Lomnitz, using seismograms written in 1953 and 1954. There are no significant differences among the results obtained by three investigators working at three different times from independent sets of data (table 3). The corrections refer to the mean of the entire group of stations. The second decimals are of only statistical significance.

Gutenberg took the means of the amplitudes for two components and worked out station corrections based on these means; Richter and Lomnitz used the data for each instrument individually. The results of the two procedures do not differ significantly.

Magnitudes used in this paper also involve amplitudes recorded at La Jolla, for which the correction +0.1 has been retained; and at Woody (E only) and Barrett (N only), for which Mr. Lomnitz found −0.08 and −0.18 respectively.

This completes the description of the magnitude scale for local earthquakes. The

TABLE 2

LOGARITHMS OF THE RESULTANT HORIZONTAL GROUND AMPLITUDES A (IN MICRONS); MAXIMUM SURFACE WAVES, T = 20 SEC.; FOR A SHOCK OF MAGNITUDE ZERO

θ	$-\log A$	θ	$-\log A$	θ	$-\log A$
Deg.	Microns	Deg.	Microns	Deg.	Microns
20	4.0	60	4.8	110	5.2
25	4.1	70	4.9	120–150	5.3
30	4.3	80	5.0	160	5.35
40	4.5	90	5.05	170	5.3
45–50	4.6	100	5.1	180	5.0

TABLE 3

STATION CORRECTIONS FOR MAGNITUDE

(1932, Richter [1935]; 1943, Gutenberg; 1954, Lomnitz)

Year	Pasadena N	Pasadena E	Riverside N	Riverside E	Sta. Barbara N	Sta. Barbara E	Haiwee N	Haiwee E	Tinemaha N	Tinemaha E
1932	+.23	+.25	+.21	+.20	−.13	−.12	+.08	+.02	−.24	−.40
1943	+0.2		+0.2		−0.1		0.0		−0.2	
1954	+.23	+.14	+.18	+.14	−.21	−.24	−.04	+.01	−.14	−.26

extension to teleseisms was initiated by Gutenberg and Richter (1936), and has been developed by Gutenberg (1945*a*, *b*, *c*). To make use of instruments of all types it was necessary to base the extended scale on the calculated motion of the ground. Present procedure is as follows:

1) Surface waves of periods near 20 seconds (shallow earthquakes only). Combine the horizontal components vectorially and apply table 2 or equivalent charts to the maximum result. Båth (1952) has worked out a corresponding table for the vertical component of surface waves. To correct for depth h add approximately $0.01h - 0.2$.

2) Body waves (especially P, S, PP). For each phase use the calculated maximum of the particle velocity A/T, separately for the vertical component and for the vectorially combined horizontal components. Apply tables and charts as given by Gutenberg (1945*c*). For large M a provisional correction of $+(M - 7)/4$ is being applied in current practice to remove a systematic discrepancy between magnitudes determined from body waves and from surface waves. This point is discussed more fully toward the end of the present paper. Decision respecting which type of wave

provides the better magnitude standard at large distances should be reached as part of a contemplated revision of the magnitude scale. Retaining the present definition based on trace amplitudes of torsion seismometers, it appears that neither ground amplitudes of surface waves nor ground velocity of body waves yield exact magnitudes; for it follows from equation (7) of the present paper that A/T is not simply related to M. The implied change in prevailing period with increasing M must affect the response of the torsion seismometer. This is unfortunate in view of the definition of M.

3) Station corrections similar to those of table 3 are used. That for Pasadena is +0.2 for body waves, +0.1 for surface waves.

4) Allow for the effect of path and of unequal radiation of energy in different azimuths from the source. If the latter effect is large, as for the surface waves of the major Kern County earthquake of 1952 (Benioff, Gutenberg, and Richter, 1954, p. 980), the best practical course is to plot, as a function of azimuth, the amplitudes reduced to a fixed distance (for surface waves, by using table 2). A sine function of the form $w + z \sin \alpha$ (where α is the azimuth) is fitted to the data, and the magnitude is computed for A such that $A^2 = w^2 + z^2/2$. This corresponds to taking the square root of the mean of A^2. Analogous procedure has been applied to the body waves.

Because of the azimuthal effect, when earthquakes are repeated in the same region and on the same structures, recorded amplitudes for these at a given station may be systematically high or low. In addition, special allowance must be made for paths along which there is abnormal loss of energy, so that recorded amplitudes at the end of such paths are systematically low (Gutenberg, 1945*a*, p. 9). These two effects combine to produce the geographical corrections to observed magnitudes which have been worked out for several stations, notably for Pasadena (Gutenberg, 1945*a*; Båth, 1952), Strasbourg (Peterschmitt, 1950), Rome (Di Filippo and Marcelli, 1949, p. 488), Uppsala and Kiruna (Båth, 1954), Vienna and Graz (Trapp, 1954).

Many stations now report magnitudes for teleseisms, but not all of these report the amplitude data on which these results are based.

Magnitudes for local earthquakes are being determined from torsion seismometer records at Prague, Apia, Berkeley (for the group of northern California stations), and at Wellington (for the New Zealand stations); magnitudes given by Wellington previous to 1949 should be increased by 0.3 to allow for a factor 2 in the static magnification.

Di Filippo and Marcelli (1950) developed a magnitude scale for Italian local earthquakes based on Wiechert instruments. Tsuboi (1951) has worked out a magnitude scale for Japan using ground amplitudes A.

Physical Elements of Earthquake Motion

For calculating earthquake energy, and for the physical interpretation of intensity as well as of magnitude, the following quantities are used: t, T, A, a, and the particle velocity A/T.

Duration t.—This will usually refer to the wave train of maximum A/T; which is not necessarily the same as the wave train of maximum a, and nearly always differs from that of maximum A. In reading a seismogram close attention must be given to the response characteristics of the instrument. Short-period instruments are generally preferable for the present purposes. For Δ less than 100 km., effective duration t of maximum trace motion shorter than the duration of the wave group containing the maximum has been read from accelerograms reproduced in "United States Earthquakes" (USCGS); $\log t$ was then plotted as a function of Δ and M. For given M, $\log t$ is nearly independent of Δ up to 50 km. Beyond 50 km. the duration of the short-period motion which is represented by t on these records decreases with distance, descending to half at roughly $\Delta = 80$ km. (The duration of the longer-period

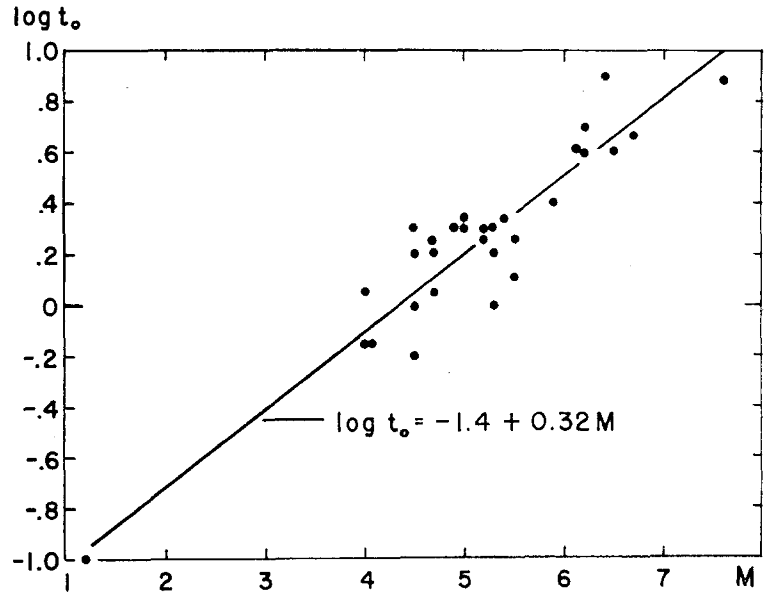

Fig. 1. Duration t_0 of strong motion at short distances, as a function of magnitude M.

motion increases rather rapidly with distance.) Duration t has been extrapolated to give t_0; $\log t_0$ is plotted as a function of M in figure 1.* In addition, $\log t_0 = -1.0$ ($t_0 = 0.1$ sec.) has been plotted at $M = 1.2$, representing a reading by Richter and Nordquist (1948) from standard torsion seismograms. The data are well represented by

$$\log t_0 = -1.4 + 0.32\,M \tag{2}$$

which replaces equation (28) of Paper 1.

Period T.—Data for periods of the waves of maximum amplitude, taken from "United States Earthquakes," appear in table 5. Corresponding readings from standard torsion seismograms are reported in table 6. Both groups of data were plotted in terms of M and Δ. In general there is little difference among the individual stations; however, the dominant recorded periods at Riverside are distinctly shorter than those at the other stations used. This agrees with the results of a former study (Gutenberg, 1936), in which it also appeared that Mount Wilson has nearly the

* All figures have been drafted by Mr. J. M. Nordquist.

same short-period characteristics. Observation during routine measurement suggests that the same is true at Woody, Palomar, and China Lake. All the stations named are on or close to granitic rock. For Δ less than 50 km. there is practically no change of period with distance. This distance range has been used to study the effect of magnitude on period. The following average periods have been found:

M	1–2	2–3	3.5–5.5	6–6.6	7.6
T	0.1	0.20	0.25	0.3	0.5 sec.

These data are taken as representing T_0. A rough representation for $M < 7$ is

$$\log T_0 = -1.1 + 0.1\,M \tag{3}$$

This replaces equation (32) of Paper 1. It is at least partly a consequence of greater extent of faulting with increased magnitude. For the largest shocks T may also increase because of change in elastic constants. For $M > 7$, T_0 seems to increase more rapidly than given by (3); however, there are no data for $\Delta < 80$ km. Combining (2) and (3),

$$\log n = -0.3 + 0.22\,M \tag{4}$$

which gives an increase from a single wave, at $M = 1+$, to 30 waves at $M = 8$. This is consistent with all available observations. On the other hand, a similar combination of equations (28) and (32) of Paper 1 gives an increase from 7 waves for the smallest shocks to only about 11 waves for $M = 8$; this unacceptable result was one reason for undertaking the present revision.

Beyond 50 km. the periods increase. Data are adequate only for magnitudes from 5.5 to 6.5 (table 5 only), for which the means are approximately as follows:

Δ	0–50	51–100	101–150	151–200	201–250
T	0.25	0.3	0.4	0.4	0 6

Amplitudes A.—These are taken from "United States Earthquakes" and from direct readings on seismograms of standard torsion instruments. Tables 5 and 6 give log A after applying the station corrections from tables 7 and 1 respectively. Table 7 was derived by the same general method used in arriving at table 3. Data used for tables 5 and 7 are exclusively from accelerographs in basements or on ground floors. Readings from Weed instruments, from displacement meters, or from accelerographs on upper floors, were not used.

In combining the two sets of data it appeared that the mean of the data of table 5, as reduced for magnitude and distance, to which the corrections in table 7 refer, is about 0.4 unit of the logarithm higher (corresponding to a factor 2.5±) than the corresponding mean of the data of table 6 with corrections from table 3. This has been verified directly by three quantitative methods:

1) Maximum amplitudes reported for the USCGS installation in Pasadena (Pasadena A. in the tables), on the main campus of the California Institute of Technology, have been compared directly with those computed for the same earthquakes from the strong-motion seismograms written at the Seismological Laboratory. The logarithms of these amplitudes have been increased by the appropriate corrections of +0.1 and +0.2 respectively. The mean residual logarithmic difference is 0.4 ± 0.2 (table 8).

2) Values of A_0/T_0 derived from tables 5 and 6 for shocks of magnitude 4.1 to 5.4 inclusive lead to a corresponding logarithmic difference of 0.39 ± 0.08. Details are set forth in a later paragraph.

3) As a further test, a standard torsion seismometer was operated in February and

TABLE 4

OBSERVED DURATION t OF LARGE SHORT-PERIOD MOTION ON USCGS ACCELEROGRAPH RECORDS, AND CORRESPONDING DURATION t_0 AT EPICENTER

(For reduction from t to t_0, see text)

Date	Station	Δ	M	t (sec.)	log t	log t_0
1933, Mar. 10.	Long Beach	27	6.2	5	0.7	0.7
	Vernon	53	6.2	3	0.5	0.6
1933, Oct. 2.	Long Beach	5	5.3	1	0.0	0.0
	Vernon	24	5.3	1½	0.2	0.2
1934, Dec. 30.	El Centro	64	6.5	2½	0.4	0.6
1936, Feb. 23	Colton	10	4.5	2	0.3	0.3
1937, Mar. 8	Berkeley	3	4.5	¾	−0.2	−0.2
1938, Apr. 12	El Centro	13	4.5	1	0.0	0.0
May 31	Colton	48	5.3	1½	0.2	0.3
June 5	El Centro	35	5	2	0.3	0.35
Dec. 3	Bishop	40	5.5	1½	0.2	0.25
1939, Mar. 21	El Centro	38	4	¾	−0.2	−0.15
Mar. 24	El Centro	38	4	¾	−0.2	−0.15
1940, May 18	El Centro	40	6.7	4	0.6	0.65
Oct. 10	Vernon	30	4.7	1±	0.0	0.05
1941, June 30	Santa Barbara	16	5.9	2½	0.4	0.4
Nov. 14	Vernon	24	5.4	2	0.3	0.35
1942, Oct. 21	El Centro	48	6.4	7	0.8	0.9
1943, May 29	Livermore	35	4	1	0.0	0.05
May 31	Bishop	16	4.5	1½	0.2	0.2
Oct. 25	San Jose	22	5.5	1¼	0.1	0.1
1947, Aug. 10	Hollister	5	5.0	2	0.3	0.3
1949, Mar. 9	Hollister	21	5.2	2	0.3	0.3
Mar. 13	Hollister	21	4.7	1½	0.2	0.2
1951, Dec. 27	Bishop	34	5.2	1¾	0.2	0.25
1952, July 21	Taft	47	7.6	7½	0.8	0.9
Aug. 7	Wheeler Ridge	10	4.9	2	0.3	0.3
Nov. 21	San Luis Obispo	77	6.2	2½	0.4	0.6
Aug. 29	Tehachapi	35	4.7	1½	0.2	0.25

March, 1955, on the campus in the same basement room as that housing the USCGS accelerograph. For the maxima in local shocks, this shows amplitudes averaging 0.6± units of the logarithm larger than those recorded for the same motions by the corresponding torsion instrument at the Seismological Laboratory (fig. 2; table 9). The long-period surface waves of teleseisms do not show this effect.

TABLE 5

LOCAL INTENSITY I, GROUND AMPLITUDES A (IN MICRONS), GROUND VELOCITY A/T (MICRONS/SEC.), ACCELERATION a FOR CALIFORNIA EARTHQUAKES, RECORDED BY USCGS ACCELEROGRAPHS

Corrections of table 7 are applied to A, A/T, $(A/T)_0$, and a

For calculated values at epicenter, see text.

Date	M	Station	Δ	I	log a	log A	T	$q =$ log A/T	log a_0	$q_0 =$ log $(A/T)_0$
1942, Oct. 21	6.4	Colton	177	5	2.5	2.1	0.14	2.9	2.4	...
		El Centro	48	6	1.7	2.6	0.3	3.1	2.4	4.0
		Hollywood	241	5	0.5	2.5	0.4	2.9	2.1	...
		Long Beach	225	4	0.3	2.8	0.6	3.0	2.2	...
		San Diego	113	6	1.3	2.8	0.3	3.3	2.7	4.6
		Vernon	225	5	0.5	1.8	0.4	2.2	2.4	...
1943, May 31	4.5	Bishop	20	6	1.3	1.9	0.14	2.8	1.3	3.0
1943, Aug. 28	5.5	Hollywood	153	4	0.7	2.1	0.4	2.5	2.0	...
		Los Angeles	121	4	0.3	2.0	0.4	2.4	1.7	3.7
		Colton	24	5	1.0	2.6	0.15	2.8	1.5	3.2
		San Bernardino	35	5	0.5	...	0.3	...	...	...
		San Diego	180	4	0.3	1.7	0.3	2.2	1.9	...
1943, Oct. 25	5.5	Livermore	29	5	1.5	2.8	0.3	3.3	2.1	3.8
		Martinez	80	5	1.4	...	...	...	2.7	...
		Oakland	69	5	1.0	2.5	0.2	3.2	2.4	3.9
		San Jose	23	6	2.0	3.9	...	...	2.7	...
		San Francisco	48	5	1.0	2.2	0.3	2.7	1.9	3.6
1944, June 12	5.3	Los Angeles	138	5	0.5	1.3	0.15	2.1	2.0	...
1944, June 18 17^h	4.5	Hollywood	26	5	0.9	2.2	0.3	2.7	1.3	3.1
		Los Angeles	19	5	1.1	2.5	0.25	3.1	1.4	3.3
		Long Beach	7	5	1.3	2.9	0.25	3.5	1.6	3.6
		Vernon	14	6	2.0	2.7	0.15	3.5	2.0	3.6
1944, June 18 20^h	4.4	Hollywood	26	5	0.9	2.5	0.5	2.8	1.3	3.2
		Los Angeles	19	5	0.9	2.3	0.3	2.8	1.2	3.0
		Long Beach	7	5	1.0	3.1	0.6	3.5	1.3	3.6
		Vernon	14	6	1.3	2.1	0.2	2.8	1.3	3.0
1945, Apr. 1	5.4	Santa Barbara	58	4	0.7	2.4	0.4	2.8	1.7	3.8
1945, May 17	4.5	Hollister	3	6	1.7	2.6	0.2	3.5	1.5	3.5
		San Francisco	137	2	0.3	1.9	0.4	2.3	1.7	3.6
1945, Aug. 15	5.7	El Centro	68	5	1.1	1.8	0.2	2.5	2.0	3.5
		San Diego	126	5	0.6	2.0	0.3	2.5	2.1	3.8
1945, Aug. 27	4.7	San Jose	12	6	0.2	2.3	0.15	3.1	0.6	3.2
1946, Jan. 8	5.4	El Centro	246	4	0.8	1.7	0.2	2.4	2.2	...

TABLE 5—(*Continued*)

Date	M	Station	Δ	I	log a	log A	T	q = log A/T	log a_0	q_0 = log $(A/T)_0$
1946, Mar. 15	6.3	Los Angeles	188	5	0.9	2.7	0.6	2.9	2.5	...
		Hollywood	184	5	0.7	2.4	0.5	2.7	2.2	...
		San Francisco	449	1	0.3	2.2	0.6	2.4	2.5	...
		Pasadena A.	177	4	0.7	2.6	0.8	2.7	2.4	4.2
		Santa Barbara	211	5	0.9	3.1	0.9	3.2	2.4	...
		Westwood	188	5	0.5	2.4	0.4	2.8	2.2	...
1947, Apr. 2	4.2	El Centro	24	5	1.1	1.7	0.25	2.3	1.2	2.5
1947, Apr. 10	6.4	Bishop	311	?	1.0	...	...	...	2.6	...
		Colton	124	5½	0.9	2.3	0.25	2.9	2.5	4.2
		Hollywood	190	5½	1.0	2.2	0.3	2.7	2.5	4.2
		Los Angeles	185	5½	0.8	2.3	0.3	2.8	2.4	4.3
		Pasadena A.	169	5½	0.0	2.1	0.5	2.4	1.8	3.9
		San Diego	254	4	0.3	2.0	0.5	2.3	2.1	...
		Vernon	184	5½	0.8	2.4	0.3	2.9	2.2	4.3
		Westwood	187	5	0.6	2.7	0.6	2.9	2.3	4.3
1947, June 22	5±	Hollister	37	6	1.3	2.8	0.4	3.2	1.9	3.9
		Oakland	100	5	0.3	1.7	0.2	2.4	1.6	3.7
		San Francisco	105	5	0.8	1.5	0.2	2.2	2.1	3.5
		San Jose	39	5	0.0	1.6	0.2	2.3	1.1	3.0
1947, June 22	5±	Los Angeles	161	4	0.0	1.5	0.3	2.2	1.5	...
1948, Mar. 28	4.6	Hollister	15	5	1.0	2.0	0.2	2.7	1.0	2.9
		San Francisco	128	4	0.6	2.3	0.5	2.6	1.9	3.9
1948, June 7	4.2	Bishop	32	5	1.3	1.7	0.2	2.2	1.6	2.7
1948, July 20	4.5	San Jose	13	4½	0.0	2.0	0.4	2.4	0.5	2.6
1948, Dec. 4	6.5	Colton	94	6	1.2	2.6	0.2	3.3	2.7	4.6
		Hollywood	186	6	1.1	2.9	0.3±	3.4	2.6	4.9
		Long Beach	170	4	0.8	2.2	0.3	2.7	2.6	4.1
		Los Angeles	178	6	1.3	3.4	0.6	3.6	2.9	5.0
		Pasadena A.	167	6	1.2	3.1	0.6	3.3	2.9	4.7
		San Diego	154	6	0.9	2.6	0.4	3.0	2.4	4.4
		Vernon	173	6	1.2	2.7	0.4	3.1	2.5	4.5
		Westwood	183	6	0.8	2.9	0.5	3.2	2.7	4.7
1948, Dec. 31	4.5	Hollister	31	5½	1.1	2.6	0.4	3.0	1.2	3.5
1949, Mar. 9	5.2	Hollister	21	7	2.2	3.3	0.3	3.8	2.3	4.0
		Oakland	115	6	0.0	1.2	0.03	2.7	1.6	4.0
		San Francisco	114	6	0.8	2.2	0.4	2.6	2.1	3.9
		San Jose	48	6	0.5	2.0	0.2	2.7	1.8	3.6
1949, Mar. 13	4.7	Hollister	21	6	1.2	2.2	0.25	2.8	1.3	3.0

TABLE 5—(*Continued*)

Date	M	Station	Δ	I	log a	log A	T	$q =$ log A/T	log a_0	$q_0 =$ log $(A/T)_0$
1949, May 2	5.9	Hollywood	238	2?	0.0	1.9	0.6	2.1	1.6	...
		Colton	143	4	0.0	1.8	0.4	2.2	1.6	3.6
		Los Angeles	230	2?	0.0	2.0	0.6	2.4	1.7	...
		Pasadena A	220	2?	0.0	2.1	0.5	2.4	1.7	...
		Vernon	223	2?	0.3	2.0	0.5	2.3	1.7	...
1949, June 9	4.9	San Francisco	90	4	0.3	1.6	0.3	2.1	1.6	3.3
		San Jose	26	6	0.7	2.2	0.3	2.7	1.5	3.1
1949, Oct. 22	4.7	Hollister	34	5	1.2	2.4	0.3	2.9	1.7	3.4
1949, Nov. 4	5.7	El Centro	118	4	0.9	2.0	0.3	2.5	2.0	3.8
		San Diego	101	6	1.1	2.3	0.15	3.1	2.5	4.4
1949, Dec. 9	4.6	Bishop	14	5	0.7	1.2	0.15	2.0	0.6	2.2
1950, Jan. 11	4.1	Los Angeles	8	5	0.3	1.0	0.1	2.0	0.4	2.1
1950, July 29	5.5	El Centro	28	5	0.2	2.4	0.3	2.9	1.2	3.4
1950, Aug. 21	2.3	Calipatria	0	6	0.5	1.5	0.2	2.2	0.5	2.2
1951, Jan. 23	5.6	El Centro	31	6	1.4	...	...	...	1.7	...
1951, July 29	5.4	Hollister	34	5	1.2	2.4	0.3	2.9	1.7	3.4
1951, Aug. 6	4.6	Hollister	32	5	1.5	2.6	0.3	3.1	1.9	3.6
1951, Oct. 3	4.2	Hollister	11	4	0.7	1.8	0.2	2.5	0.6	2.6
1951, Oct. 30	4.1	Hollister	5	4	1.0	2.1	0.3	2.6	0.9	2.7
1951, Oct. 31	4.7	Hollister	5	6	1.4	2.6	0.3	3.1	1.3	3.2
1951, Dec. 27	5.2	Bishop	34	5	1.3	1.5	0.1	2.5	1.7	3.0
1951, Dec. 25	5.9	Hollywood	140	5	0.4	1.6	0.3	2.1	1.8	3.5
		Long Beach	110	6	0.4	2.8	0.8	2.9	2.0	4.2
		Los Angeles	130	5	0.3	2.5	0.5	2.8	1.8	4.2
		San Diego	110	6	1.1	3.2	0.5	3.5	2.5	4.8
1952, Feb. 17	4.5	Colton	16	5	1.0	2.3	0.2	3.0	1.3	3.2
1952, May 23	4.9	Hoover Dam	16	6	1.5	2.0	0.1	3.0	1.7	3.2
1952, July 21	7.6	Bishop	275	5½	1.2	3.2	0.9	3.3	2.7	...
		Colton	185	6	1.1	3.3	0.6	3.5	2.8	4.9
		El Centro	406	4	0.6	3.2	1.7	3.0	2.5	...
		Hawthorne	390	4	0.5	3.2	1.3	3.1	2.7	...
		Hollister	303	5	0.9	3.3	1.0	3.3	2.6	...

TABLE 5—(*Continued*)

Date	M	Station	Δ	I	log a	log A	T	$q =$ log A/T	log a_0	$q_0 =$ log $(A/T)_0$
1952, July 21	7.6	Hollywood	120	7	1.5	3.4	0.6	3.6	2.8	5.0
		Hoover Dam	405	4	0.4	3.5	2.0	3.2	2.6	...
		Long Beach	163	6½	1.1	3.7	1.0	3.7	2.8	5.1
		Los Angeles	124	7	1.3	4.0	1.0	4.0	2.8	5.4
		Oakland	430	3½	0.0	2.9	1.2	2.8	2.5	...
		Pasadena A.	120	5½	1.5	3.9	0.7	4.1	3.0	5.5
		San Diego	310	4½	0.6	2.5	0.4	2.9	2.5	...
		San Francisco	440	4	0.5	2.9	1.0	2.9	2.7	...
		San Jose	370	4	0.6	4.1	2.5	3.7	3.0	...
		San Luis Obispo	157	5	1.1	3.3	0.8	3.4	2.6	4.8
		Santa Barbara	90	7	1.8	3.1	0.8	3.2	3.1	4.4
		Taft	47	7	2.1	4.0	0.5	4.3	2.8	5.2
		Vernon	130	6	1.5	3.7	0.8	3.8	2.8	5.2
		Westwood	120	6	1.3	3.6	0.7	3.8	2.9	5.2
1952, July 29	6.1	Los Angeles	160	3	−.3?	...	...	...	1.2?	...
		Taft	70	?	1.6	2.8	0.3	3.3	2.6	4.3
1952, July 31	4.6	Taft	80	5½	1.2	2.6	0.13	3.5	2.3	4.6
1952, Aug. 7	3.7	Taft	45	?	1.4	2.4	0.3	2.9	2.1	3.7
		Wheeler Ridge	10	?	1.6	3.5	0.4	3.9	1.7±	4.0
1952, Aug. 13	4.7	Wheeler Ridge	10	?	1.0	2.2	0.2	2.9	1.1±	3.0
1952, Aug. 14	4.2	Wheeler Ridge	5	?	1.7	2.7	0.15	3.5	1.8±	3.6
1952, Aug. 22	5.8	Arvin	16	5	1.5	2.5	0.5	2.8	2.5?	3.0
		Taft	53	4	1.2	2.1	0.25	2.7	1.8	3.7
		Tehachapi	44	4½	1.3	2.9	0.35	3.4	1.8±	4.2
		Wheeler Ridge	35	?	1.2	2.4	0.2	3.1	2.1±	3.7
1952, Oct. 12	4.5	Oakland	13	5	1.4	2.2	0.1	3.2	1.8	3.3
		Berkeley	19	5	1.3	2.0	0.1	3.0	1.6	3.2
		San Francisco	27	4	0.9	1.5	0.15	2.7	1.3	2.8
1952, Oct. 21	4.1	San Francisco	10	5	0.7	...	...	...	0.7	...
1952, Nov. 21	6.2	San Luis Obispo	77	6	1.7	3.0	0.3	3.5	2.9	4.6
		Hollister	105	5	1.0	2.5	0.4	2.9	2.2	4.2
		Pasadena A.	335	2	0.2	2.4	0.7	2.6	2.1	...
		San Francisco	250	4	0.3	2.4	0.7	2.6	2.0	...
		Santa Barbara	205	4	0.6	3.0	1.1	2.9	2.1	...
		Taft	185	4	0.6	2.5	0.7	2.7	2.0	4.1
1953, May 22	4.2	Taft	32	?	1.2	2.4	0.3	2.9	1.6	3.4
1953, June 13	5.5	El Centro	20	6	1.5	2.3	0.2	3.0	1.5	3.2

TABLE 5—(*Concluded*)

Date	M	Station	Δ	I	log a	log A	T	$q =$ log A/T	log a_0	$q_0 =$ log $(A/T)_0$
1953, Dec. 15	3.8	Arvin	0	5	1.7	3.2	0.25	3.8	2.0±	3.8
1954, Jan. 12	5.9	Arvin	29	6	1.3	...	0.3	...	2.3	...
		Bishop	275	1	0.5	...	...	...	2.0	...
		Colton	185	4	0.3	...	0.3	...	2.0	...
		Hollywood	120	5	0.8	2.2	0.3	2.7	2.2	4.1
		Los Angeles	124	5	0.6	2.3	0.4	2.7	2.1	4.1
		Pasadena A.	120	5	0.7	...	0.5	...	2.3	...
		Santa Barbara	90	6	0.8	2.4	0.3	2.9	2.1	4.1
		Taft	47	6	1.8	3.0	0.3	3.5	2.6	4.4
		Vernon	130	4	0.8	2.3	0.4	2.7	2.3	4.1
		Westwood	120	4	0.7	2.4	0.4	2.8	2.2	4.2
1954, Jan. 27	5.0	Arvin	17	6	1.6	3.3	0.2	4.0	2.0±	4.2
1954, Mar. 19	6.2	Colton	130	5	0.9	...	0.3	...	2.5	...
		El Centro	90	5½	1.3	...	0.6	...	2.4	...
		Hollywood	210	4½	0.5	2.5	0.6	2.7	2.0	...
		Long Beach	180	4	0.3	...	1.0	...	2.3	...
		Los Angeles	200	4½	0.5	3.5	0.5	3.8	2.1	...
		Pasadena A.	207	4½	0.3	3.1	0.8	3.2	2.0	...
		San Diego	110	5	1.1	3.5	0.9	3.6	2.5	5.0
		Vernon	190	4½	0.7	2.8	0.8	2.9	2.1	4.4
		Westwood	220	4½	0.3	3.6	0.6	3.8	2.1	...
1954, Apr. 25	5.2	Hollister	20	7	1.7	...	0.7	...	1.8	...
		Oakland	110	4	0.6	3.2	0.3	3.7	2.2	5.0
		San Francisco	125	5	0.8	...	0.4	...	2.2	...
		San Jose	60	4½	0.9	3.0	1.0	3.0	2.3	4.0
1954, May 23	5.1	Bakersfield	43	4	0.8	2.4	0.3	2.9	1.7±	3.6
		Taft	47	?	1.0	2.3	0.3	2.8	1.8	3.7
		Westwood	120	1	0.3	2.1	0.4	2.5	1.9	3.9

TABLE 6

GROUND AMPLITUDES A (IN MICRONS), $q = \log (A/T)$, ACCELERATION a FOR CALIFORNIA EARTHQUAKES, FROM RECORDS OF STANDARD TORSION SEISMOGRAPHS AT STATIONS OF THE CALIFORNIA INSTITUTE OF TECHNOLOGY, PASADENA

(Corrections of table 3 are applied to A, a and q. For calculation of q_0, see text.)

Date	Hour	M	Station	Δ	log A	T	log a	q	q_0
1935, July 13	18	4.7	Pasadena	30	+1.9	0.2	+0.9	2.6	3.1
1938, May 31	0	5.3	Riverside	39	+2.1	0.15	+1.4	2.9	3.6
July 5	10	4.4	Riverside	40	+1.9	0.15	+1.1	2.7	3.4
Aug. 6	14	3.8	Riverside	36	+1.2	0.18	+0.4	2.0	2.6
1939, Jan. 7	12	5.0	Haiwee	25	+2.1	0.5	+0.2	2.4	2.8
Jan. 29	17	4.1	Pasadena	23	+1.9	0.4	+0.3	2.3	2.5
Nov. 7	10	4.7	Riverside	9	+2.0	0.4	+0.5	2.4	2.5
Nov. 7	10	4.7	Mount Wilson	29	+1.7	0.2	+0.7	2.4	2.8
Dec. 27	11	4.7	Pasadena	41	+1.7	0.2	+0.7	2.4	3.1
1940, Feb. 19	4	4.6	Riverside	30	+1.9	0.15	+1.1	2.7	3.2
Apr. 18	10	4.4	Riverside	5	+1.8	0.13	+1.1	2.6	2.7
Oct. 21	22	4.9	Pasadena	37	+2.0	0.4	+0.5	2.4	3.1
Oct. 22	2	3.8	Pasadena	31	+1.4	0.7	+0.4	1.6	2.1
Nov. 14	0	5.4	Pasadena	41	+2.0	0.2	+1.1	2.7	3.4
1950, Sept. 17	20	3.2	Pasadena	81	−0.3	0.15	−1.0	0.5	1.6
Sept. 20	6	2.4	Pasadena	41	−0.2	0.17	−1.1	0.6	1.3
Sept. 21	14	3.4	Pasadena	86	+0.2	0.2	−0.8	0.5	1.7
Nov. 7	7	2.8	Pasadena	51	−0.3	0.3	−1.6	0.2	1.1
Nov. 11	11:14	3.3	Pasadena	85	+0.2	0.6	−1.7	0.4	1.6
Nov. 11	11:32	3.1	Pasadena	85	0.0	0.35	−1.5	0.5	1.7
Nov. 11	21	2.9	Pasadena	33	+0.2	0.12	−0.3	1.1	1.6
Nov. 15	18	3.1	Pasadena	30	+0.9	0.2	−0.1	1.6	2.1
Nov. 16	19	3.8	Pasadena	30	+1.3	0.17	+0.4	2.1	2.6
Dec. 8	16	2.0	Pasadena	26	−0.3	0.3	−1.6	0.2	0.6
Dec. 26	4	3.1	Pasadena	56	0.0	0.2	−1.0	0.7	1.6
1951, Jan. 19	17	3.2	Pasadena	100	−0.1	0.15	−0.9	0.7	2.0
Feb. 4	20	2.6	Pasadena	35	−0.2	0.12	−0.8	0.7	1.3
Feb. 8	5	2.5	Pasadena	37	−0.2	0.15	−0.9	0.6	1.2
Feb. 10	4	3.3	Pasadena	65	+0.4	0.25	−0.8	1.0	2.0
Mar. 14	5	3.3	Pasadena	59	+0.3	0.13	−0.3	1.2	2.2
Apr. 13	16	3.4	Pasadena	44	+0.5	0.18	−0.3	1.3	2.0
Apr. 14	12	2.3	Pasadena	39	−0.3	0.18	−1.2	0.5	1.2
June 11	3	2.6	Pasadena	16	−0.1	0.18	−1.0	0.7	0.9
June 22	4	2.7	Pasadena	41	−0.3	0.25	−1.5	0.3	1.0
July 11	8	3.4	Pasadena	62	−0.1	0.2	−1.3	0.6	1.6
July 11	8	3.4	Santa Barbara	95	0.0	0.2	−0.9	0.7	2.0
July 14	22	2.9	Haiwee	42	+0.1	0.3	−1.2	0.6	1.3
July 17	14	3.2	Riverside	86	−0.1	0.2	−1.1	0.6	1.8
July 27	2	3.5	Haiwee	67	+0.7	0.15	−0.2	1.5	2.5

TABLE 6—(*Continued*)

Date	Hour	M	Station	Δ	log A	T	log a	q	q_0
1951, July 28	3	2.8	Pasadena	14	+0.8	0.15	0.0	1.6	1.7
July 28	11:29	3.5	Haiwee	42	+0.7	0.25	−0.2	1.3	2.0
July 28	11:36	2.9	Haiwee	42	+0.6	0.2	−0.4	1.3	2.0
Aug. 9	17	3.0	Riverside	86	−0.3	0.2	−1.3	0.4	1.6
Aug. 11	17	2.6	Riverside	49	−0.1	0.15	−0.9	0.7	1.6
Aug. 19	1	3.3	La Jolla	76	+0.4	0.15	−0.4	1.2	2.2
Aug. 21	14	2.7	Riverside	36	+0.3	0.15	−0.5	1.1	1.7
Aug. 23	14	3.2	Riverside	49	+0.6	0.15	−0.2	1.4	2.3
Aug. 25	1	2.9	Riverside	52	+0.5	0.13	−0.1	1.4	1.9
Aug. 27	1	3.5	Riverside	90	+0.4	0.15	−0.3	1.2	2.4
Aug. 30	16	2.4	Haiwee	0	+0.5	0.2	−0.4	1.2	1.2
Sept. 2	6	3.2	Riverside	70	−0.1	0.18	−1.0	0.7	1.7
Sept. 2	11	2.9	Haiwee	54	+0.1	0.2	−0.6	0.8	1.8
Sept. 7	3	3.3	Pasadena	21	0.0?	0.2	−1.0?	0.7?	1.1?
Sept. 22	8:22	4.3	Pasadena	76	+1.2	0.35	−0.3	1.7	2.8
Sept. 22	8:22	4.3	Riverside	14	+1.8	0.18	+1.0	2.6	2.8
Sept. 22	8:25	2.7	Riverside	14	+0.4	0.15	−0.3	1.2	1.4
Sept. 28	5	3.1	Pasadena	65	−0.2	0.15	−1.0	0.6	1.6
Sept. 28	5	3.1	Riverside	11	+1.6	0.15	+0.8	2.4	2.5
Oct. 1	23	2.5	Haiwee	16	+0.5	0.2	−0.4	1.2	1.4
Oct. 6	5	2.5	Riverside	11	+0.8	0.1	+0.1	1.6	1.7
Oct. 16	12	4.0	Riverside	44	+1.2	0.15	+0.5	2.0	2.7
Oct. 22	1	3.2	Riverside	17	+1.0	0.15	+0.2	1.8	2.0
Oct. 26	16	3.0	Santa Barbara	10	+0.6	0.15	0.0	1.4	1.5
Nov. 1	18	2.5	Pasadena	23	−0.1	0.15	−0.9	0.7	1.0
Nov. 4	19	2.6	Riverside	33	−0.1	0.18	−0.8	0.6	1.1
Nov. 14	23	4.1	La Jolla	89	+1.3	0.3	0.0	1.8	3.0
Nov. 17	4	3.4	Santa Barbara	87	+0.5	0.25	−0.5	1.1	2.3
Nov. 18	0	3.0	Pasadena	54	0.0	0.2	−1.0	0.7	1.7
Nov. 18	19	3.0	Haiwee	42	+0.2	0.25	−0.9	0.8	1.5
Nov. 20	19	2.3	Riverside	20	+0.2	0.18	−0.6	1.0	1.2
Nov. 24	1	2.7	Riverside	16	+0.3	0.2	−0.6	1.0	1.2
Nov. 26	19	3.0	Santa Barbara	49	−0.1	0.25	−1.2	0.5	1.4
Dec. 3	18:08	2.5	Riverside	10	+0.5	0.15	−0.2	1.3	1.4
Dec. 3	18:18	2.7	Riverside	8	+1.1	0.18	+0.3	1.9	2.0
Dec. 14	11	2.3	Pasadena	44	+0.2	0.18	−0.7	1.0	1.8
Dec. 14	11	3.3	Riverside	90	+0.4	0.18	−0.4	1.2	2.4
Dec. 23	0	3.0	Riverside	37	+1.0	0.18	+0.1	1.8	2.5
Dec. 30	17	2.7	Haiwee	0	+0.6	0.3	−0.7	1.1	1.1

TABLE 6—*(Continued)*

Date	Hour	M	Station	Δ	log A	T	log a	q	q_0
1952, Jan. 4	3	2.7	Riverside	44	+0.2	0.15	−0.6	1.0	1.8
Jan. 8	6	4.4	Riverside	94	+1.1	0.3	−0.3	1.6	2.8
Jan. 27	16	3.2	Riverside	46	+0.3	0.18	−0.4	1.1	1.9
Jan. 28	10	2.9	Riverside	39	+0.3	0.18	−0.4	1.1	1.8
Jan. 31	20	2.6	Santa Barbara	32	0.0	0.25	−0.12	0.6	1.1
Feb. 4	18	3.9	Tinemaha	58	+0.6	0.3	−0.7	1.1	2.1
Feb. 6	23	3.2	Pasadena	87	0.0	0.28	−1.3	0.5	1.7
Feb. 9	8	2.5	Tinemaha	23	+0.4	0.35	−1.1	0.9	1.2
Feb. 9	19	3.7	Haiwee	57	+0.7	0.45	−1.0	1.0	2.0
Feb. 9	22	3.6	Santa Barbara	102	+0.3	0.5	−1.5	0.8	2.1
Feb. 10	13	4.0	Santa Barbara	103	+0.7	0.45	−1.0	1.1	2.4
Feb. 13	1	3.6	Pasadena	70	+0.5	0.15	−0.2	1.3	2.3
Feb. 14	16	3.0	Riverside	31	+0.5	0.28	−0.8	1.0	1.5
Feb. 15	7	2.7	Riverside	13	+0.7	0.23	−0.4	1.3	1.4
Feb. 15	8	2.6	Riverside	12	+0.6	0.23	−0.5	1.3	1.4
Feb. 16	4	2.2	Riverside	12	+0.3	0.23	−0.8	1.0	1.1
Feb. 17	12	4.5	Pasadena	88	+1.2	0.3	−0.1	1.7	2.9
Feb. 19	9	2.6	Riverside	21	+1.0	0.2	+0.1	1.3	1.5
Mar. 3	2	2.5	Riverside	17	+0.7	0.2	−0.2	1.4	1.6
Mar. 3	6	2.8	Riverside	11	+0.9	0.15	+0.1	1.7	1.8
Mar. 3	16	3.5	Riverside	64	+0.7	0.23	−0.4	1.3	2.3
Mar. 10	18	3.9	La Jolla	110	+1.0	0.45	−0.9	1.4	2.7
Apr. 17	21	3.0	Riverside	30	+0.5	0.18	−0.3	1.3	1.8
May 27	13	2.2	Riverside	35	0.0	0.2	−0.9	0.7	1.3
May 27	14	2.4	Riverside	35	+0.2	0.2	−0.7	0.9	1.5
June 29	7	2.6	Riverside	16	+0.8	0.15	0.0	1.6	1.8
July 1	16	3.1	Santa Barbara	16	+0.7	0.2	−0.3	1.4	1.6
July 10	8	3.7	Pasadena	19	+1.3	0.15	+0.5	2.1	2.3
July 10	8	3.7	Riverside	77	+1.2	0.18	+0.3	1.9	3.0
July 21	21	4.3	Santa Barbara	80	+1.4	0.6	−0.5	1.6	2.7
July 21	23	4.5	Santa Barbara	86	+1.5	0.7	−0.5	1.7	2.9
July 22	7	4.1	Santa Barbara	91	+0.9	0.7	−1.1	1.1	2.3
July 22	8	4.7	Santa Barbara	113	+1.3	0.8	−0.9	1.4	2.8
July 22	10	4.1	Santa Barbara	88	+0.9	0.5	−1.0	1.2	2.4
July 22	14:05	4.3	Santa Barbara	98	+0.9	0.5	−1.0	1.2	2.5
July 22	14:30	4.3	Santa Barbara	79	+1.1	0.4	−0.5	1.5	2.6
July 22	17:52	4.1	Santa Barbara	90	+0.8	0.6	−1.1	1.0	2.2
July 23	0	4.4	Santa Barbara	90	+1.3	0.5	−0.5	1.6	2.8
July 23	17	4.1	Santa Barbara	93	+1.0	0.6	−1.0	1.2	2.4
July 24	9	4.3	Santa Barbara	96	+1.1	0.4	−0.5	1.5	2.7
July 27	20	3.5	Santa Barbara	25	+1.6	0.5	−0.7	1.6	2.0

TABLE 6—(*Concluded*)

Date	Hour	M	Station	Δ	log A	T	log a	q	q_0
1952, Aug. 23	10	5.0	Pasadena	39	+2.0	0.2	+1.0	2.7	3.4
Aug. 23	10	5.0	Riverside	96	+1.8	0.15	+1.0	2.6	3.9
Aug. 29	2	3.5	Haiwee	35	+1.1	0.18	+1.1	1.9	2.5
Aug. 30	14	3.3	Santa Barbara	15	+1.2	0.3	−0.2	1.9	2.1
Sept. 26	16	2.8	Riverside	11	+0.8	0.18	0.0	1.6	1.7
Sept. 28	4	3.0	Riverside	11	+1.1	0.18	+0.3	1.9	2.0
Oct. 17	12	3.0	Riverside	13	+0.5	0.15	−0.2	1.3	1.4
Oct. 17	15	2.5	Riverside	13	−0.1	0.15	−0.8	0.7	0.8
Oct. 19	14	3.9	Pasadena	41	+0.9	0.2	−0.1	1.6	2.3
Nov. 16	13	3.8	Riverside	23	+1.7	0.2	+0.7	2.4	2.7
Nov. 17	3	3.9	Haiwee	43	+1.2	0.45	−0.5	1.5	2.2
1953, Jan. 23	6	2.4	Riverside	11	+0.5	0.2	−0.5	1.2	1.3
Feb. 7	0	3.9	Tinemaha	27	+1.0	0.25	−0.2	1.6	2.0
Feb. 7	22	2.9	Riverside	10	+0.9	0.2	−0.1	1.6	1.7
Feb. 17	17	3.0	Riverside	14	+0.7	0.2	−0.3	1.4	1.5
Feb. 26	3	3.5	Haiwee	38	+0.3	0.23	−0.5	1.0	1.7
Mar. 16	5	2.1	Riverside	14	0.0	0.15	−0.8	0.8	1.0
May 2	5	3.9	Riverside	26	+1.8	0.15	+1.0	2.6	3.0
May 2	6	2.5	Riverside	26	+0.6	0.1	0.0	1.6	2.0
May 4	2	3.4	Riverside	17	+1.1	0.2	+0.1	1.8	2.0
June 7	5	2.9	Riverside	26	+0.9	0.15	+0.2	1.7	2.1
Nov. 1	18	2.4	Riverside	8	+0.4	0.2	−0.6	1.1	1.2
Nov. 4	15	2.5	Riverside	10	+0.8	0.4	−0.8	1.2	1.3
1954, Apr. 20	9:11	2.8	Haiwee	5	+1.0	0.2	+0.2	1.7	1.8
Apr. 20	9:18	3.5	Haiwee	5	+1.4	0.2	+0.6	2.1	2.2
June 5	0	2.6	Riverside	14	+0.9	0.15	+0.1	1.7	1.8
June 5	23	2.4	Riverside	14	+0.3	0.15	−0.5	1.1	1.2
July 23	3	2.7	Riverside	15	+0.8	0.15	+0.1	1.6	1.7
Aug. 20	8	4.4	Tinemaha	73	+1.0	0.45	−0.7	1.3	2.3
Sept. 3	7	2.0	Pasadena	23	0.0	0.13	−0.6	0.9	1.2
Sept. 10	21	2.6	Riverside	13	+0.9	0.15	+0.1	1.7	1.9
Oct. 26	8	3.3	Riverside	33	+1.1	0.18	+0.3	1.9	2.5
Oct. 26	16	4.1	Riverside	33	+1.5	0.2	+0.5	2.2	2.7
Nov. 17	7	4.1	Haiwee	33	+1.1	0.4	−0.3	1.5	2.0
Nov. 27	2	3.2	Haiwee	10±	+1.6	0.4	+0.2	2.0	2.1

Test runs at the Laboratory before and after show close agreement between the two instruments.

All this evidence agrees with the repeated observation that shocks generally perceptible in the city of Pasadena, including the California Institute campus, are not felt by the staff at the Seismological Laboratory in spite of an alarm and an ink-writing seismograph which can be seen recording. It confirms the natural interpretation that the difference in recording between the campus, where the ground is alluvial, and the Seismological Laboratory, founded on weathered granitic rock, is of the same nature as the generally observed effect of ground on intensities and recorded amplitudes. Most of the USCGS installations are in centers of population, and consequently in general on less consolidated ground than the average of the seismological stations of the Pasadena group, most of which have been deliberately placed on rock (where possible, on granitic rock) in order to reduce the effect of ground disturbance.

TABLE 7

STATION CORRECTIONS FOR USCGS ACCELEROGRAPH LOCATIONS, EXCLUDING THOSE ABOVE GROUND LEVEL

(The corrections are to be added to the logarithm of the ground amplitudes to find the value corresponding to the mean of these stations. N is the number of observations.)

Station	Corr.	N	Station	Corr.	N
Colton	+0.1	8	Vernon	−0.2	9
Arvin	(+0.4)	2	San Francisco	−0.1	7
Los Angeles	0.0	10	Westwood	+0.1	6
San Diego	0.0	6	Bishop	−0.3	3
San Jose	+0.3	5	Long Beach	+0.2	6
Taft	−0.2	5	Hollister	−0.2	5
El Centro	−0.3	3	Oakland	+0.2	5
Hollywood	−0.1	11	Santa Barbara	−0.1	4
Pasadena A	+0.1	8			

TABLE 8

COMPARISON OF RECORDING AT THE TWO PASADENA INSTALLATIONS: (1), SEISMOLOGICAL LABORATORY; (2), ATHENAEUM, CALIFORNIA INSTITUTE OF TECHNOLOGY CAMPUS, USCGS

(Logarithms of ground amplitudes in microns, with station correction applied from tables 3 and 7.)

Date	Δ	*M*	(1)	(2)	(2) − (1)
1933, Oct. 2	41	5.3	1.0	0.7	−0.3
1946, Mar. 15	175	6.3	1.2	1.6	+0.4
1947, Apr. 10	175	6.4	1.3	1.1	−0.2
1948, Dec. 4	167	6.5	1.4	2.1	+0.7
1952, July 21	120	7.6	2.4	2.9	+0.5
1952, Nov. 21	335	6.2	0.6	1.4	+0.8
1954, Mar. 19	207	6.2	1.1	2.1	1.0

Table 10 extends table 1 of Paper 1 to later dates, giving trace amplitudes B as recorded by the strong-motion instruments at the Seismological Laboratory, with the addition of the corresponding periods, local intensities, and corrected acceleration.

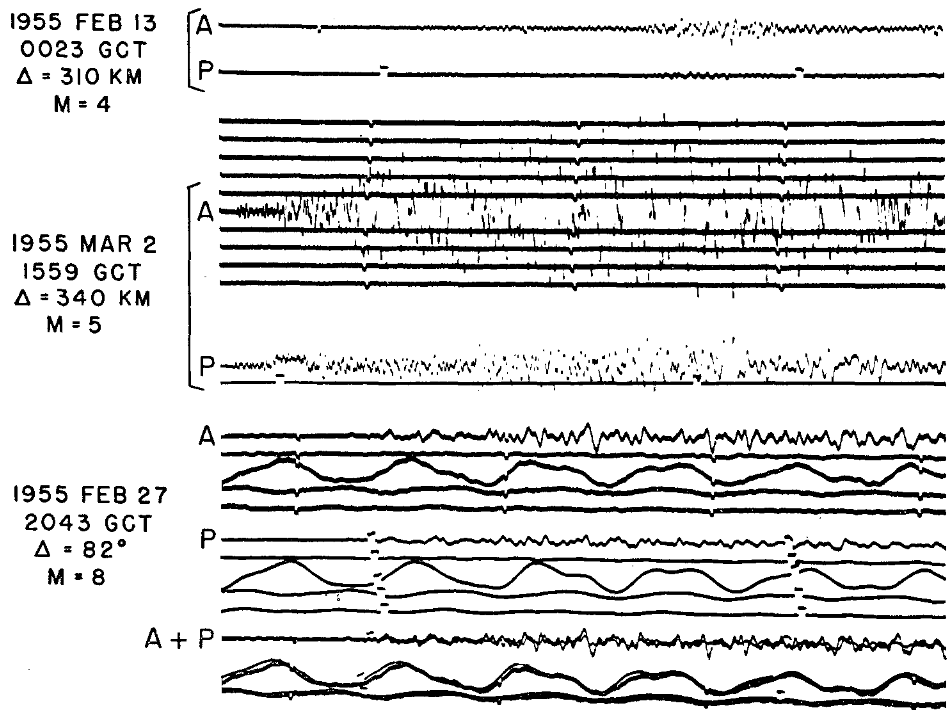

Fig. 2. Simultaneous seismograms written by standard torsion seismometers, oriented N–S. (A) Athenaeum, California Institute of Technology campus, (P) Pasadena Seismological Laboratory. (A + P) superposition of (A) and (P).

TABLE 9

Comparison of Maximum Trace Amplitudes Recorded by Identical Standard Torsion Seismographs, N–S Component, at the Seismological Laboratory (1) and at the Athenaeum (2) on the Campus of the California Institute of Technology, 1955

Date	Hour	Region of epicenter	P group				S group			
			(1)	(2)	T	Log (2)/(1)	(1)	(2)	T	Log (2)/(1)
Feb. 11	19:44	Kern County....	3.5	5	0.6	0.2	16.5	≧28	0.7	≧0.3
Feb. 11	22:13	Kern County....	....	...	...	...	0.1	0.5	0.5	0.7
Feb. 11	22:26	Kern County....	....	...	...	...	0.2	0.8	0.5	0.6
Feb. 12	11:34	Kern County....	....	...	...	...	0.07	0.6	0.3	0.9
Feb. 12	12:03	Nevada?........	....	...	...	...	0.05	0.6	0.2	1.1
Feb. 12	13:53	Kern County....	0.17	0.5	0.5	0.8	0.8	2.6	0.5	0.5
Feb. 18	15:28	Baja California .	0.15	0.8	0.5	0.7	0.3	0.9	0.5	0.5
Feb. 23	19:44	SE of Pasadena.	....	...	...	...	1.6	2.4	0.3	0.2
Mar. 2	15:59	Coast Ranges?..	....	...	...	...	5	15	1	0.5
Mar. 14	15:03	Nr. Wrightwood	....	...	...	...	3	15	0.2	0.7

TABLE 10

DATA FROM STRONG-MOTION SEISMOGRAPHS AT PASADENA (SEISMOLOGICAL LABORATORY), AND CORRESPONDING INTENSITIES

Date	M	Δ	I	B	Log $B-M$	T	Log a + 0.2
1942, Oct. 21, 8h	6.5	240	3½	5	−5.8	2½	+0.1
Oct. 21, 17h	5.7	250	...	0.5	−6.0	1	−0.1
1944, June 10	4.5	128	...	0.05	−5.8	?	?
June 12, 2h	5.1	133	...	0.3	−5.6	?	?
June 12, 3h	5.3	133	...	0.4	−5.7	?	?
June 18, 16h	4.5	33	3	0.6	−4.7	1	0.0
June 18, 19h	4.4	33	3	0.4	−4.8	1	−0.2
July 2	4.7	136	...	0.03	−6.2	?	?
1945, Apr. 1	5.4	171	...	1.1	−5.4	2½	−0.5
1946, Jan. 8	5.4	252	...	0.35	−5.9	?	?
Mar. 15, 5:21	5.2	175	2	0.6	−5.4	1	0.0
Mar. 15, 5:55	6.3	175	4	4.2	−5.7	1.8	+0.3
Sept. 27	5.0	124	...	0.25	−5.6	?	?
1947, Apr. 10, 7h	6.4	175	5	5	−5.7	1	+0.9
Apr. 10, 8h	5.1	175	...	0.2	−5.8	0.8	−0.3
July 24	5.5	155	...	0.6	−5.7	?	?
1948, Dec. 4	6.5	170	5½	7	−5.7	0.8	+1.3
1949, Nov. 4	5.7	266	...	0.3	−6.2	?	?
1952, July 21, 11:52	7.6	120	6½	70	−5.8	1.8	+1.5
July 21, 12:05	6.4	121	4½	2½	−6.0	1.8	0.0
July 23, 0h	6.1	140	4½	1½	−5.9	2	−0.2
July 23, 7h	5.4	140	4	1¼	−5.3	0.8	+0.5
July 23, 13h	5.7	132	4	0.8	−5.8	1.3	−0.1
July 25, 19:10	5.7	133	4	3	−5.2	1.3	+0.4
July 25, 19:43	5.7	133	4	2	−5.4	1.3	+0.3
July 29, 7h	6.1	150	5	(5)	−5.4	3	0.0
July 29, 8h	5.1	150	...	0.5	−5.4	1.3	−0.3
July 31	5.8	136	4	1.6	−5.6	1.3	+0.1
Aug. 1	5.1	110	4	0.3	−5.6	1.5	−0.7
Aug. 22	6.4	121	4½	2.5	−6.0	1.8	0.0
Aug. 23	5.0	39	4½	0.5	−5.3	1.2	−0.3
Nov. 22	6.2	332	2	1	−6.2	0.8	+0.4
1953, June 14	5.5	275	...	0.4	−5.9	?	?
1954, Jan. 12	5.9	120	4½	1.3	−5.8	0.7	+0.6
Jan. 27	5.0	120	2	0.35	−5.5	?	?
Mar. 19	6.2	208	...	3.3	−5.7	1.2	+0.5
July 6, 11h	6.8	593	...	0.8	−6.9	?	?
July 6, 22h	6.4	593	...	0.5	−6.7	?	?
Aug. 24	6.8	593	...	1.2	−6.7	?	?
Nov. 12	6.3	360	2	1.0	−6.1	1.5	−0.1

In Paper 1 the present equation (1) was applied in effect to trace amplitudes recorded by these strong-motion instruments. This disregards the effect of variable T on the magnification of the standard torsion seismometer. Table 11 gives factors by which trace amplitudes B (in mm.) are to be multiplied to obtain ground amplitudes in microns ($= 10^4 A$), for T ranging from 0.1 sec. to 2.0 sec., and for three types of torsion seismometers, all with free period of 0.8 sec., but otherwise as follows:

	Type I	Type II	Type III
Static magnification	2,800	2,800	100 (effective)
Damping ratio	20 : 1	50 : 1	50 : 1

The effective magnification for type III and for some other installations is as viewed on a film projector with eightfold magnification. With minor changes, during the entire interval since the setting up of the magnitude scale in 1932, instruments of type II have been in service at Pasadena, Haiwee, and Tinemaha, and type I instruments at the other stations of the Pasadena group. The additional type III instruments (at Pasadena only) began recording in 1954.

TABLE 11

Conversion Factors, Trace to Ground Amplitudes (mm. to microns) for Torsion Seismometers of Types I, II, III

Type of instrument	T (sec.)									
	0.1	0.3	0.5	0.6	0.8	1.0	1.2	1.4	1.6	2.0
I	0.36	0.36	0.39	0.41	0.50	0.66	0.90	1.14	1.5	2.2
II	0.36	0.37	0.40	0.45	0.57	0.73	0.96	1.25	1.6	2 3
III	10.0	10.4	11.3	12.5	16.0	20.4	27	35	46	65

For the strong-motion instruments providing the data for table 10, magnification is sensibly equal to the static magnification of 4 when T is less than about 4 seconds (this gives a value of 250, corresponding to table 11). Calibration for such an instrument on average ground in terms of magnitude is given in table 12. β_0 is revised. Applying this to recorded amplitudes, as in table 6, the Pasadena station correction of 0.2 must be added to the logarithms. The difference (δ, table 12) between these data and those for the torsion seismometer (table 1) shows a progressive increase with distance, owing to the increase in prevailing periods of the recorded maxima. There is a further effect due to systematic increase of prevailing period with magnitude at fixed distance, as in equation (3). This gives rise to serious difficulty in comparing magnitudes derived from torsion and from strong-motion instruments. Magnitudes determined from the latter using table 12 without considering the effect of period are higher than those from table 11. Moreover, because of the larger amplitudes with which long periods are written by the strong-motion instrument, there is often much trouble in reading the amplitudes of the superposed short-period vibrations, and consequent loss of accuracy.

The dependence of log A on M and Δ has been investigated by plotting the data of tables 5 and 6 in various ways. The decrease of log A with Δ is somewhat irregular, with a definite break near 100 km. (compare the values of δ in table 12). This is probably connected with a shift of the maximum amplitudes from a group of direct transverse waves to various types of channel waves including the former $\bar{S}$. This shift appears also clearly from the data of a previous publication (Gutenberg, 1945*d*, table IX, p. 305).

TABLE 12

EFFECT OF INCREASING DISTANCE Δ ON β AND δ

(β = log b for strong-motion instruments: period 8 seconds, static magnification 4, critical damping. δ = difference between β and corresponding data for the standard torsion seismometer, table 1.)

	Δ							
	0	20	40	70	100	200	300	500
β............	4.3	4.5	5.0	5.4	5.6	5.8	6.1	6.6
δ............	2.8	2.8	2.6	2.6	2.6	2.3	2.0	1.9

In the following paragraphs a relation between log (A_0/T_0) and M will be set up (equation 7). Combining this with equation (3)

$$\log A_0 = -5.9 + M - 0.027\,M^2 \tag{5}$$

Observe that A_0 is here expressed in centimeters. This represents the data in table 5 well if log A from that table is decreased by 0.4 to correct for the ground at the USCGS stations, as discussed above.

The quantity A/T is required for calculating kinetic energy. For convenience we use the notation q = log (A/T), in which A is measured in microns. Data are reported in tables 5 and 6, corrected for the station ground effects from tables 7 and 3 respectively. Dependence of q on Δ and M has been investigated by plotting the tabulated data in various forms. Average values appear in table 13. Note that no data are available for $M > 5.5$; for $M > 4$ with $\Delta < 30$ km. readings are few, and tabulated values are uncertain.

The quantity $q_0 - q$ is nearly independent of M in the range of table 13. Its behavior as Δ increases is shown in table 14.

To compare A/T with trace amplitudes β, table 14 also shows $\beta_0 - \beta$. Theoretically

$$q_0 - q = \beta_0 - \beta + \log (T/T_0) \tag{6}$$

Numerically log (T/T_0) is near zero up to Δ = 60 km., increasing to about +0.1 at 100 km. Table 14 is consistent with this, considering that the tabulated results depend on extrapolated values for q_0 and β_0, both of which are uncertain by at least ± 0.1 in the logarithm.

Values of $q_0 - q$ from table 14 have been used to reduce the individual observations of A/T to Δ = 0 regardless of M. The results for Δ less than 200 km.

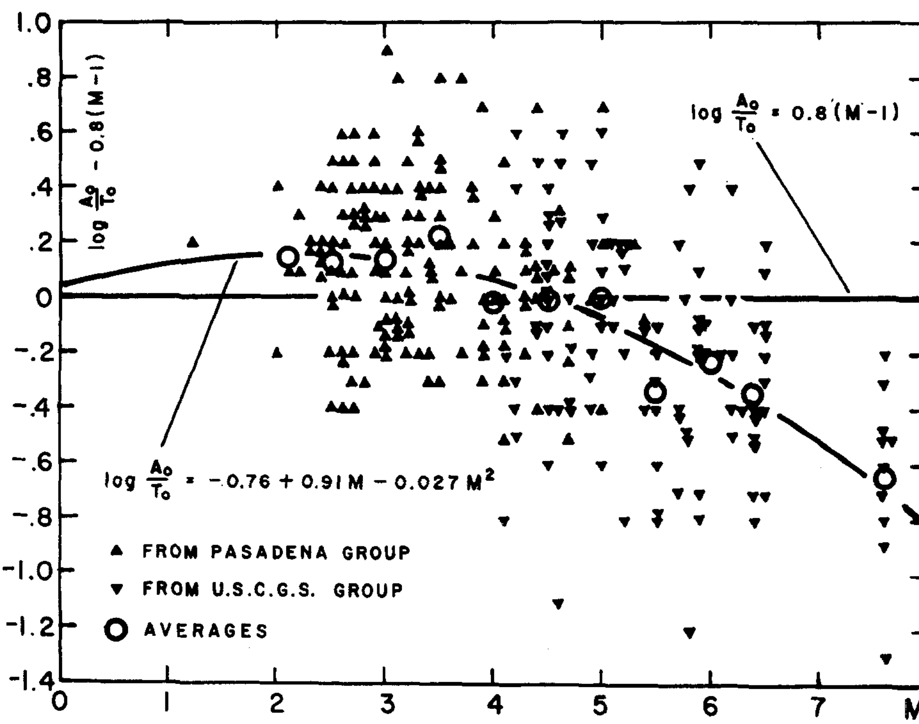

Fig. 3. Logarithm of A/T extrapolated to a point source, as a function of M.

TABLE 13

MEAN VALUES OF q = LOG (A/T), A BEING MEASURED IN MICRONS, FOR GIVEN Δ AND M

A. USCGS data (table 5)

M	Δ								
	0	10	20	30	40	50	60	70	100
4.4	(3.3)	(3.2)	(2.8)	2.5	...	...	...	...	...
4.8	(3.8)	(3.5)	(2.9)	2.8	(2.8)	...	...	...	...
5.2	...	(4.0)	(3.8)	3.1	2.8	2.8	2.8	...	2.6
5.5	...	...	...	(3.5)	3.3	3.2	3.1	3.1	3.1

B. Torsion seismometer data (table 6)

M	Δ								
	0	10	20	30	40	50	60	70	80
2.2	1.1	1.0	0.9	0.7	0.5	...	...	...	...
3.0	1.8	1.7	1.6	1.3	1.1	0.8	0.6	...	...
3.5	2.1	2.1	2.0	1.7	1.4	1.3	1.2	1.1	1.0
4.0	2.4	2.4	2.3	2.2	1.9	1.6	1.4	...	...

appear in tables 5 and 6. As was to be expected from the results for amplitudes, when reduced values of q for the two groups of data are plotted together those from table 5 are seen to be systematically higher than those from table 6. The magnitude ranges of the two tables overlap from 4.1 to 5.4; in this common range $q - q_0$ was calculated, using the data as plotted in figure 3 to correct for the effect of variable M. For 54 values derived from table 5 the mean was -0.37 ± 0.06, and for 33 values derived from table 6 the mean was -0.76 ± 0.05. The mean adjustment between the two sets of data is therefore 0.39 ± 0.08, as already given in discussing the amplitudes.

In figure 3 the quantity $q_0 - 0.8(M - 1)$ is plotted with separate signatures for data from tables 5 and 6. Averages over half-magnitude intervals are also plotted.

TABLE 14

MEAN VALUES OF $q_0 - q$ FOR GIVEN Δ AND ALL MAGNITUDES FROM 2 TO 5. CORRESPONDING VALUES OF $\beta_0 - \beta$ DERIVED FROM TABLE 12

	Δ							
	10	20	30	40	50	60	80	100
$q_0 - q$	0.1	0.2	0.5	0.7	0.9	1.0	1.1	1.3
$\beta_0 - \beta$	0.0	0.2	0.5	0.7	0.9	1.0	1.2	1.3

An additional point representing one of the small earthquakes near Riverside reported by Richter and Nordquist (1948) appears at magnitude 1.2. Data for large M with small Δ are deficient.

The complete body of data cannot be represented by q_0 as a linear function of M. This is ultimately due to the effect of increasing period on the dynamic magnification of the torsion seismometer on which the magnitude depends. A term in M^2 was added for convenience, although this can hardly have physical significance. An attempt might be made, following a suggestion by Dr. Benioff, to draw two separate straight lines. A least-squares solution for the quadratic form yielded:

$$q_0 = -0.76 + 0.91\,M - 0.027\,M^2 \tag{7}$$

For $1 < M < 8.7$ this is closely equivalent numerically to:

$$q_0 = -0.9 + 0.72\,M + \tfrac{1}{2}\log(9 - M) \tag{8}$$

Acceleration a.—This was used in Paper 1 as a basis for calculating energy; in the present paper it is effectively replaced by A/T (or q). However, values of acceleration are of importance, especially for correlation with intensities.

All acceleration data not already presented in Paper 1 are reported in tables 5 and 10. Station corrections as in table 7 are to be applied to table 5. In table 10, $\log a$ is increased by 0.2. For reduction to $\Delta = 0$ the relation

$$\log a_0 - \log a = q_0 - q + \log(T/T_0) \tag{9}$$

is used. When Δ does not exceed 200 km. this leads also to nearly the same values as those for $\beta_0 - \beta$ in table 14. Table 5 includes the value of $\log a_0$, corrected by using (9) for each observation of a. These corrected values can be applied directly for correlation with intensities; but for applications involving M they should all be decreased by 0.4 to pass from the USCGS installations to the California Institute stations.

Figure 4, a shows the mean decrease of acceleration with distance for a shock of given magnitude in the California area.

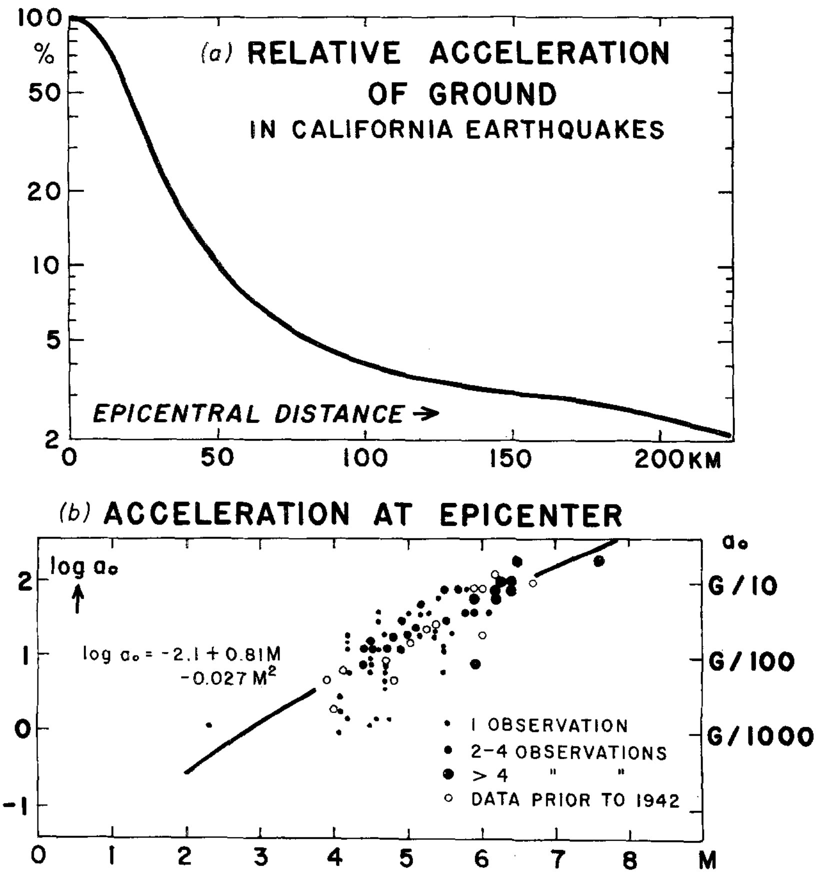

Fig. 4. Acceleration as dependent on distance and magnitude. Values for $\Delta = 0$ extrapolated.

When there are more than two observations for a shock in table 5 the corresponding corrected average of log a_0 is presented in table 15. These, together with the remaining corrected values of log a_0 from table 5, are plotted against the corresponding magnitudes in figure 4, b. In this figure the further correction -0.4 has been applied. Now:

$$a = (A/T)\, 4\pi^2/T \tag{10}$$

Noting that A is expressed in microns in calculating q, and introducing q_0 from (7) and log T_0 from (3)

$$\log a_0 = -2.1 + 0.81\, M - 0.027\, M^2 \tag{11}$$

The corresponding curve is drawn in figure 4.

From minimum perceptibility $a_0 = 1$ gal, or $\log a_0 = 0$. Substituting this in equation (11), one finds $M = 2.8$. This should represent the smallest shock normally perceptible on ground such as that at the Pasadena group of stations. For the USCGS installations the corresponding result is obtained by putting $\log a_0 = -0.4$, whence $M = 2.2$. This corresponds to ordinary centers of population. For relatively unstable ground, as at El Centro or Bishop, the proper value of $\log a_0$ is -0.7, whence $M = 1.7$. These results correspond to general experience with the magnitude scale.

TABLE 15

Elements of California Shocks from Data of USCGS

(Log a_0 is diminished by 0.4)

Date	M	I_0	r (km.)	$\log a_0$
1942, Oct. 21	6.4	8	250	2.0
1943, Aug. 29	5.5	6	150	1.4
1944, June 12	5.3	6	180	1.6
1946, Mar. 15	6.3	8	350	2.0
1947, Apr. 10	6.4	8	280	1.9
1947, July 24	5.5	5	150	1.1
1948, Mar. 1	4.7	6	110	?
Apr. 16	4.7	6	80	0.6
Dec. 4	6.5	8	250	2.3
1949, May 2	5.9	7?	200	1.3
Nov. 4	5.7	?	200	1.9
1950, July 29	5.4	8	170	0.8
Sept. 5	4.8	5½	100	?
1951, Jan. 23	5.6	6½	150	1.3
Dec. 5	4.5	6½	100	?
Dec. 25	5.9	?	200	1.6
1952, July 21	7.6	10	450	2.3
Aug. 22	5.8	7+	200	1.6
Aug. 23	5.0	5	180	?
Nov. 21	6.2	7+	250	1.8
1953, June 13	5.5	7	120	1.1
Sept. 25	5.2	6	110	?
1954, Jan. 12	5.9	8	220	1.8
Mar. 19	6.2	6½	240	1.8
Apr. 25	5.2	7½	140	1.7

For an acceleration of 0.1 g, $\log a_0 = 2$, and (11) gives $M = 6.5$ for the better ground. For $M = 8.6$ (11) yields $\log a_0 = 2.9$, or $a_0 = 0.8\,g$.

Intensities.—The intensities in tables 5 and 10 are taken from USCGS reports, supplemented by local information for the two Pasadena locations. I_0 and r as given in table 15 are mainly from USCGS data. Data reported in Paper 1 have been combined with all these to arrive at table 16.

These are approximately represented by the empirical relation given in Paper 1:

$$\log a = I/3 - \tfrac{1}{2} \tag{12}$$

which fails for higher intensities.

The relation of intensity to acceleration has been investigated in much more detail, including the effect of variable ground conditions, by Housner (1952) and by Neumann (1954). Neumann is attempting to redefine intensity in terms of acceleration, as was attempted by Cancani (1904). However, the destructive effects of earthquake motion also depend on amplitude, especially for the longer-period elements of the disturbance, and on the duration of strong shaking; these factors are particularly significant for great earthquakes.

Radius of perceptibility r.—This is given in table 15 from USCGS data. Figure 5 represents the relation between I_0 and r as derived from these data and those of

TABLE 16

Relation of Intensity to Acceleration (California Data)

	I								
	2	3	4	5	6	7	8	9	10
Log *a*	0.0	0.2	0.5	1.0	1.5	1.9	2.3	2.5	2.6

Paper 1. Radius r varies extremely with ground, as well as with path, extent of faulting, etc. The graph in figure 5 represents the empirical relation given in Paper 1:

$$r = 0.5\, I_0^3 - 1.7 \tag{13}$$

which fits the data fairly well.

In figure 6 the data for I_0 and M are correlated. The resulting empirical equation:

$$M = 1 + 2I_0/3 \tag{14}$$

differs slightly from the corresponding equation in Paper 1.

Figure 7 shows $\log r$ as a function of M. Two empirical equations, which slightly differ from those in Paper 1, are drawn as curves on this figure:

$$M = -3.0 + 3.8 \log r \tag{15a}$$

$$r = 1.4\,(M - 0.614)^3 \tag{15b}$$

Energy and Magnitude

An absolute scale?—The magnitude scale was originally set up for a limited and practical purpose; definitions were closely tied to the observational data in hand. The desirability of relating magnitude to physical quantities, especially to the total energy radiated from the earthquake source in the form of elastic waves, was then clearly recognized; but it was thought that more information was desirable before laying down a definition in absolute physical terms. Failing this, several attempts have been made to establish a relation between magnitude, as originally defined, and energy. The results have been revised repeatedly and drastically. It is hoped that equation (20) may serve as a good approximation for the future.

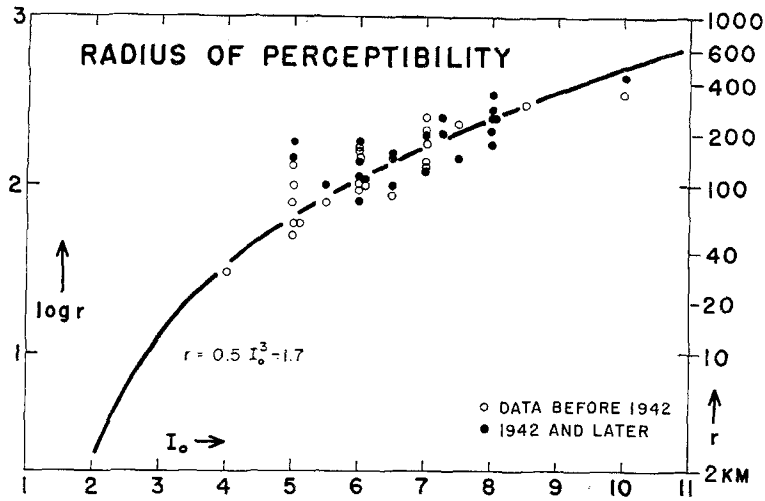

Fig. 5. Radius r of perceptibility as a function of epicentral intensity I_0.

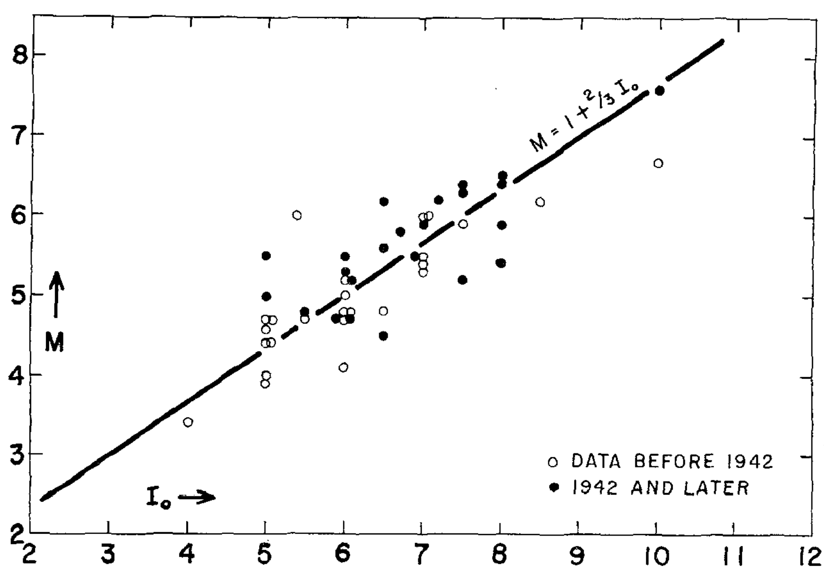

Fig. 6. Relation of magnitude M to epicentral intensity I_0.

It has been proposed to define a new "absolute" magnitude directly in terms of energy. This would be convenient for theoretical calculation; but the use of the term magnitude in two different senses would surely lead to confusion and error. Under present circumstances the authors would rather operate directly with log E. However, a change of the basic definition by substituting A/T for B is being undertaken. Details will be reported in a later paper.

It is not at all certain that there is a one-to-one correspondence between magnitude as now used and the total energy radiation. Especially when there is extended faulting, the relationship probably involves such variables as the linear extent of faulting, the amount of throw, the rate of progression of fracture, and the time required for the throw to take place at any given point. An equation such as (20) can at best represent only some sort of average condition, and is likely to fail especially for large and complex tectonic events.

A clearly desirable objective more nearly within reach is to free the magnitude scale from dependence on the local tectonic circumstances in California and on a

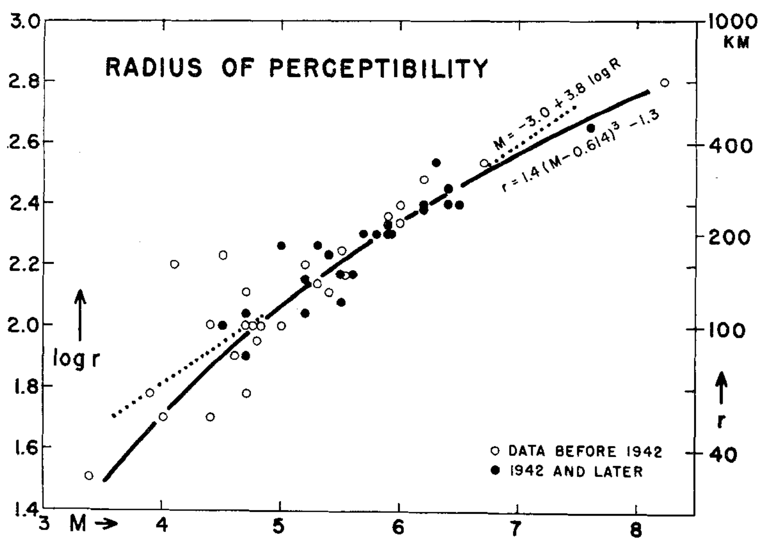

Fig. 7. Radius r of perceptibility as a function of magnitude M.

particular instrument (the standard torsion seismometer). This is approximated in practice by the procedure used to determine magnitude from the data for distant recording stations.

Linear magnitude–energy relations.—Previous publications have employed equations of the general form

$$\log E = P + QM \tag{16}$$

Some of the values published for the constants P and Q are listed in table 17. $P = 11$, $Q = 1.6$ is mentioned as preferable by Gutenberg and Richter (1954), but in that publication $P = 12$, $Q = 1.8$ was retained for calculations to avoid expensive resetting. $P = 9$, $Q = 1.8$ is being used by H. Benioff in current publications.

The linear form (16) cannot be made to fit all the data including both the largest and smallest observed magnitudes. Dr. Benioff has suggested two straight lines intersecting near magnitudes 5 and 6. Equation (20) of the present paper represents all the observations adequately, and is here adopted to replace (16).

Calculation of energy from body waves.—If we assume a point source,

$$E = 3\pi^3 h^2 v\, t_0 \rho\, (A_0/T_0)^2 \tag{17}$$

An exact equation would involve integrating $(A_0/T_0)^2$ over the duration of large motion on the seismogram. This integration is carried out in effect by taking A_0/T_0 as constant at its maximum value over an appropriate interval t_0 which is estimated from and is shorter than the duration t of the recorded wave group in which the maximum occurs. Equation (17) differs from equation (24) of Paper 1 only in the numerical factor 3 instead of 4. This change expresses the following theoretical modifications:

1) In Paper 1 by an oversight only kinetic energy was considered. Since the mean potential and kinetic energies are equal, the result must be doubled. This was pointed out to the authors by Dr. P. G. Gane.

TABLE 17
CONSTANTS P AND Q FOR THE FORMULA $\log E = P + QM$ (16), NOW TO BE REPLACED BY EQUATION (20)

P	Q	Reference
6	2	Richter (1935)
8	2	Gutenberg and Richter (1936)
11.3	1.8	Gutenberg and Richter (1942)
12	1.8	Gutenberg and Richter (1949)
9.154	2.147	Di Filippo and Marcelli (1950)
11	1.6	Gutenberg and Richter (1954, p. 10)
7.2	2.0	Båth (1955); for Rayleigh waves only
9	1.8	Benioff (1955)

2) A factor 1/4 is applied to allow for the doubling of amplitudes at the free surface.

3) In Paper 1 the calculation was applied only to the maximum waves, which at short distances are S waves. The energy in P waves must be added; this is here taken as half that in the S waves.

For numerical calculation we take $h = 16$ km., $r = 3.4$ km/sec., $\rho = 2.7$ g/cm.^{3} To find E in ergs an additional factor 10^{15} is required, whence:

$$\log E = 20.34 + \log t_0 + 2 \log (A_0/T_0) \tag{18}$$

or

$$\log E = 12.34 + \log t_0 + 2\, q_0 \tag{19}$$

In (19) A_0, measured in centimeters, has been replaced by q_0, for which the amplitude is measured in microns.

Introducing t_0 from (2) and q_0 from (7):

$$\log E = 9.4 + 2.14\, M - 0.054\, M^2 \tag{20}$$

For M ranging from 1 to 8.7 this is numerically equivalent to

$$\log E = 9.1 + 7M/4 + \log (9 - M) \tag{21}$$

which similarly follows from (19), (2), and (8).

Table 18 exhibits some results of (20). The new values of log E are significantly lower than those found in Paper 1, and agree better with other determinations (fig. 8), in spite of theoretical inexactitude including the assumption of a point source for large shocks and imperfections in the methods used to reduce the observations.

The present revision removes a group of discrepancies for which the chief causes were as follows:

1) Accelerations at the USCGS installations were used in conjunction with magnitudes determined from seismograms at the California Institute stations. This overlooks the ground factor which relatively increases the motion at the USCGS locations. Since the square of the acceleration is used in computing energy, log E was overestimated by about 0.8.

TABLE 18

Log E as a Function of M from Equation (20)

	M									
	0	1	2	3	4	5	6	7	8	8.6
Log E	9.4	11.5	13.5	15.3	17.1	18.8	20.3	21.7	23.1	23.8

2) Log t_0 and log T_0 were established on plausible but not adequate assumptions, as discussed following equation (4).

3) The linear form (16) cannot be made to fit the data; the chosen values of P and Q led to far too large values of log E for large M.

Di Filippo and Marcelli (1950) set up a magnitude scale for local earthquakes recorded in Italy on Wiechert seismographs of a standard type. In calculating the corresponding energies they closely followed Paper 1 (including omission of the potential energy and of the effect of the free surface). Their values for P and Q (table 17) give log E diverging from table 18 chiefly for large M.

Sagisaka (1954) applied the same general method to calculate the energy of an earthquake under Japan at $h = 360$ km. He allowed for the effect of unequal radiation of energy in azimuth, as well as the other points discussed in setting up equation (18). His results for the energy in the P waves are slightly falsified by using Young's modulus in place of the appropriate elastic constant.

Calculation of energy from surface waves.—A seismogram may be integrated to find the energy flux in the surface waves, which then may be integrated for the entire circular wave front. With allowance for absorption the energy radiated from the source in this form may be calculated. At large distances the surface waves represent most of the arriving energy, so that log E has been estimated by this means. Some assumption as to decrement of ground amplitude with depth is required; this is most simply done by assuming the classical equations for Rayleigh waves, as by Galitzin (1915) and Klotz (1915). Jeffreys (1923) uses:

$$E = 8\pi^3\, \delta R \sin\Theta \int A^2 H v\, d\tau / T^2 \tag{22}$$

where E is the energy in the wave front, R the radius of the earth, and

$$H = 7.06\, TV/2\pi \tag{23}$$

The coefficient 7.06 is a dimensionless number dependent only on Poisson's ratio, here taken as $\frac{1}{4}$. Assuming a constant velocity $V = 3\frac{3}{4}$ km/sec. (actually V decreases along the train of surface waves) and considering absorption, equations (22) and (23) lead to:

$$E = 8 \times 10^{14}\, e^{k\Delta} \sin\Theta \sum A^2 \tag{24}$$

or

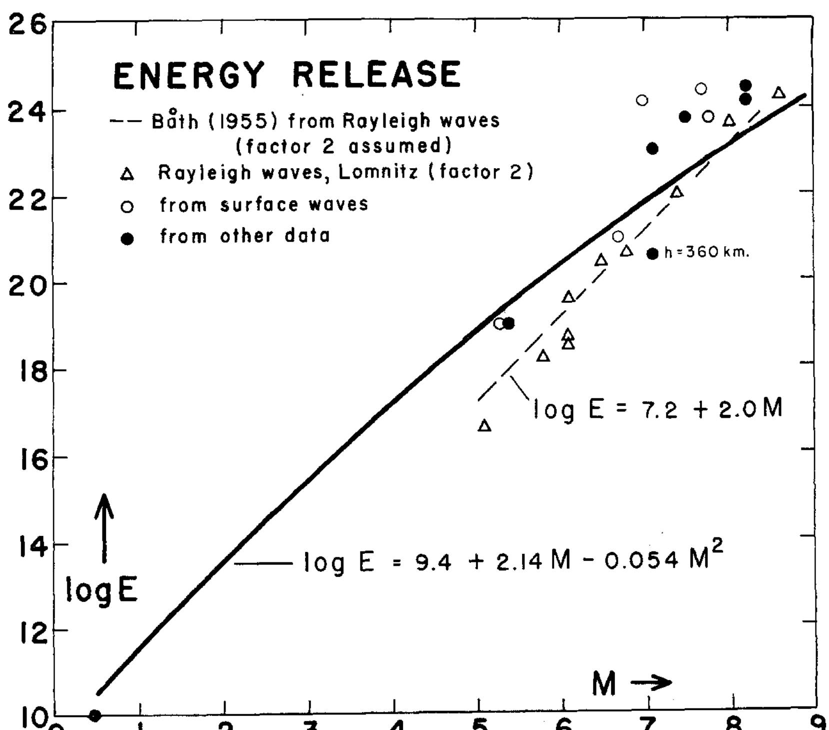

Fig. 8. Energy release E as a function of magnitude M (without distinguishing between M_L, M_B, and M_S).

$$\log E = 14.9 + 0.43 k\Delta + \log \sin\Theta + \log \sum A^2 \tag{25}$$

where A is the amplitude for each wave measured in microns, and the summation extends over all the individual waves.

Mr. Cinna Lomnitz has obtained some of the results reported in table 19 on the assumption that the total energy of the shocks is twice that given by applying (25), derived for Rayleigh waves, to the whole series of surface waves. The coefficient of absorption k is taken as 0.12 per 1,000 km. over continental paths, 0.3 over Pacific paths, and 0.2 over mixed paths; log E was increased by a further 0.2 representing the ground correction of +0.1 for surface-wave amplitudes at Pasadena. Magnitudes are from all available amplitudes of body and surface waves.

The authors are indebted to Dr. Markus Båth for discussion by correspondence and for the privilege of examining a manuscript (Båth, 1955) in which he uses (22) with slightly different numerical assumptions to calculate E by integrating from the beginning of the Rayleigh waves to the end of the surface waves on seismograms written at Kiruna. For 27 shocks of magnitudes 5.3 to 7.8 (as determined from surface-wave amplitudes at Kiruna) he finds:

$$\log E = (7.2 \pm 0.5) + (2.0 \pm 0.07)M + \log (x/2) \tag{26}$$

where x is the ratio of the total energy to that of the Rayleigh waves. The line corresponding to equation (26) with $x = 2$ is drawn in fig. 8. The plotted points representing Lomnitz' results fit this line closely. The curve corresponding to equation (20) is also drawn in figure 8. The systematic divergence of the two curves, corresponding roughly to a magnitude difference of $\frac{1}{2}(8 - M_S)$ for a given energy, is beyond reasonable doubt.

It is already known (Gutenberg, 1945*b*, p. 64) that, if magnitudes for body waves (M_B) and from surface waves (M_S) are adjusted to agree near $M = 7$, they differ for other magnitudes by approximately $\frac{1}{4}(M - 7)$. To refine this, if possible, M_B and M_S as found in determining magnitudes for *Seismicity of the Earth* (Gutenberg and Richter, 1954) are used. In figure 9, $M_S - M_B$ is plotted as function of M_S. A least-squares solution shows that these points are well represented by:

$$M_S - M_B = 0.4\,(M_S - 7) \tag{27}$$

There are very few earthquakes for which one can determine M_S and M_B from amplitudes at distant stations, and, in addition, M_L, the "local magnitude," from torsion seismometers at short distances. Most of our data for these are from the Kern County, California, earthquakes of 1952. These are used for figure 10, which suggests that for M_L near 6, approximately $M_B = M_L$. From all data now in hand, we find very roughly for magnitudes from 5 to 7:

$$M_L - M_B = 0.4(M_B - 6) \tag{28}$$

M_S is appreciably lower than M_L for the smaller magnitudes; a least-squares solution from the data in figure 10 gives:

$$M_S - M_L = 0.32(M_L - 6.6) = 0.47(M_S - 6.7) \tag{29}$$

Apparently, the proportion of energy transferred to surface waves decreases rapidly as the magnitude decreases. The duration of the train of maximum surface waves decreases more than the maximum amplitudes. This, like the variation of prevailing period with magnitude (eq. 3), is intimately connected with the change in all three linear dimensions at the source.

To put this on a theoretical basis, consider a group of body waves arriving at a given distance Δ from a fixed source of energy:

$$E = c\,(A/T)^2\,t \tag{30}$$

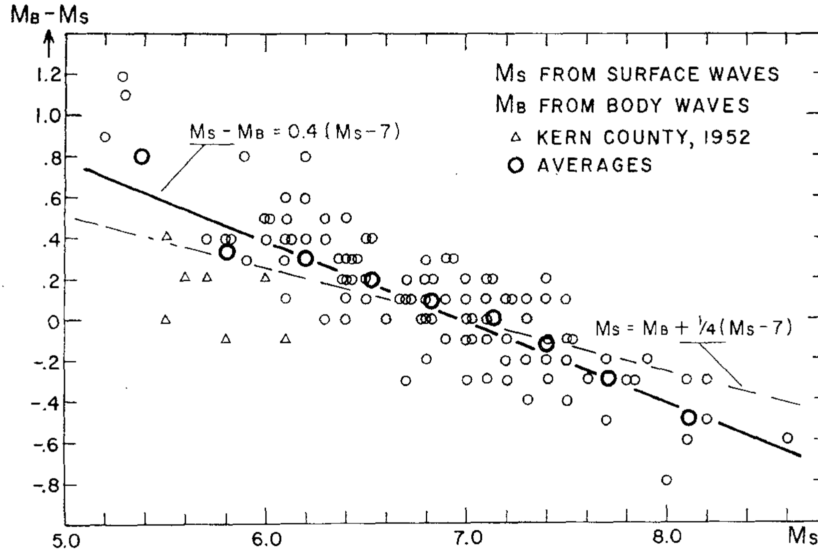

Fig. 9. Comparison of magnitude determined from surface waves and from body waves.

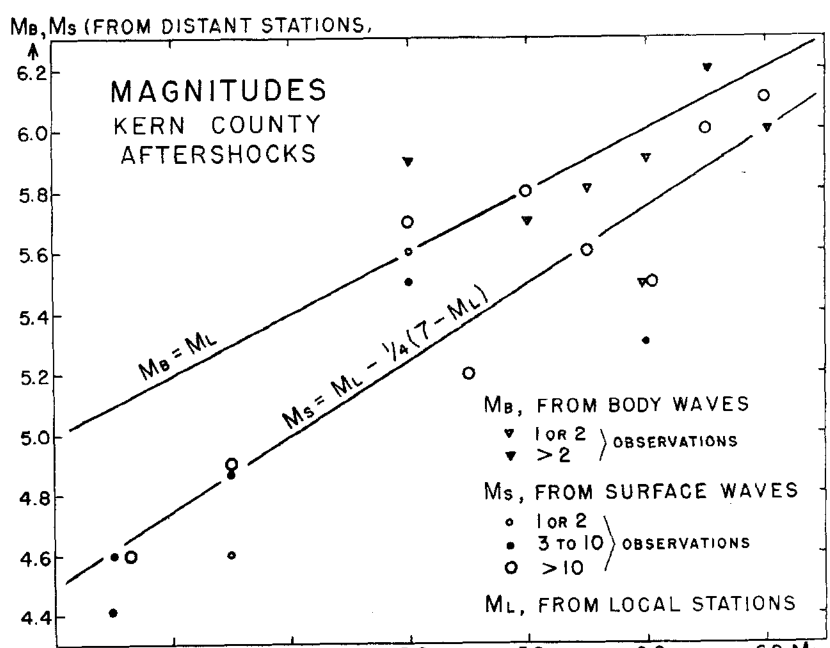

Fig. 10. Comparison of magnitudes determined from distant and from near-by stations. Kern County aftershocks, 1952.

where t is the duration of the given wave group; the constant c includes transmission factors, effects along the fixed path, elastic constants, and density. From this:

$$\log E = \log c + 2q + \log t \tag{31}$$

If the magnitude is determined from body waves, and then corrected to give M_S, from (27):

$$M_S = q - f + 0.4\,(M_S - 7) \tag{32}$$

where f is the tabulated value of q for $M = 0$, whence:

$$\log E = 1.2\,M_S + \log t + \log c + 2f + 5.6 \tag{33}$$

If $\log t = x + y\,M_S$, where x and y are constants, this result is in close agreement with (20) for $y = 0.2$, a reasonable assumption. This shows that the failure to represent the relation between $\log E$ and M by a single linear function is due partly to the use of recorded surface waves to determine M.

Energies estimated by other methods.—Reid (1910, p. 22) estimated the energy involved in the California earthquake of 1906 (table 19) by calculating the work done in displacing the crustal blocks. By considering the extent of isoseismals compared with those of 1906, he arrived at figures for the energy of 12 other earthquakes (Reid, 1912). Similar estimates for several earthquakes are reported by Sagisaka (1954). Various other attempts are discussed by Sieberg (1923, pp. 156–160). From the known strength of rocks Tsuboi (1940) calculated for the maximum possible earthquake $\log E = 24.7$. Data from these sources are included in table 19 and plotted in figure 8.

An official publication (Anon., 1950, p. 14) erroneously gives the energy of the largest earthquake as a million times that of an atomic fission bomb of the Hiroshima type. This depends on $\log E = 26$ from the superseded magnitude-energy relation of Paper 1, together with the officially adopted bomb energy for which $\log E = 20.9$. If this latter figure is used in combination with the revised magnitude-energy relation (20), ten thousand should be read instead of a million. For table 19 it has been assumed that 1 per cent of the energy of the Bikini bomb entered the rock of the ocean bottom.

Elements of California Shocks

Table 20 summarizes average data for earthquakes in southern California, based on the findings of this paper and Paper 1. Individual earthquakes may show large departures from these figures, caused by shallower or deeper hypocentral depth, unusual crustal structure near the source or along the wave path, unstable ground in the heavily shaken area, unequal distribution of energy radiation in azimuth, etc.

In table 20 a_0, A_0, T_0, r represent normal ground conditions in the metropolitan centers where most of the USCGS installations are situated. For locations on solid rock, accelerations and amplitudes may be decreased to half or less and intensities may be lower by one grade or more. Conversely, on unstable ground such as watersoaked alluvium, beach sand, or artificial fill, accelerations and amplitudes may be double or more, and intensities may be higher by one grade or over; on such ground the effect of large amplitudes with long periods is particularly increased.

TABLE 19

ENERGY OF EARTHQUAKES ESTIMATED BY VARIOUS METHODS

Date	Epicenter	Method used	Reference	log E	M
1926, July 30	Jersey	Rayleigh waves	Jeffreys (1952, p. 101)	19	5.3
1925, June 28	Montana	Rayleigh waves	Jeffreys (1952, p. 101)	21	6.7
1906, Apr. 18	San Francisco	Work of faulting	Reid (1910)	24.2	8.2
1929, June 2	Japan, h = 360 km.	Integration	Sagisaka (1954)	20.5	7.1
1923, Sept. 1	Kwanto	Work done	Sagisaka (1954)	24½	8.2
1930, Nov. 25	North Idu	Work done	Sagisaka, calc. Homma	23	7.1
1908, Dec. 28	Messina	Comparison	Reid (1912, p. 270)	23.8	7.5
1915, Jan. 13	Avezzano	Galitzin, surface waves	Sieberg (1923, p. 159)	24.1	7
1911, Feb. 18	Pamir	Surface waves	Klotz (1915)	23.8	7.7
1915, Oct. 3	Nevada	Surface waves	Sieberg (1923)	24.5	7.7
1946, July 24	Bikini, Baker Day	1 per cent of energy	Gutenberg-Richter (1946)	29±	5.3±
1948, Feb. 18	Minimal earthquakes near Riverside	Integration	Richter-Nordquist (1948)	10±	½±

Date	Hr.	Epicenter	Station	Θ	log E[a]	M
1953, Nov. 25	17	Japan	Pasadena	79°	23.7	8.0
Nov. 26	4	Japan	Pasadena	79	20.5	6.5
Nov. 26	8	Japan	Pasadena	79	20.6	6.8
1950, Aug. 15	14	Tibet	Pasadena	110	24.3	8.6
1952, July 23	0	Kern County	Ottawa	34	18.5	6.1
July 23	0	Kern County	Strasbourg	84	18.7	6.1
Aug. 22	22	Kern County	Ottawa	34	18.9	5.8
July 29	7	Kern County	Ottawa	34	19.6	6.1
July 29	8	Kern County	Ottawa	34	16.6	5.1
1953, Nov. 4	3	New Hebrides	Pasadena	85	22.0	7.4

[a] Log E calculated from equation (25) by C. Lomnitz (adding 0.3 for total energy).

LIMITATIONS AND POSSIBILITIES OF THE MAGNITUDE SCALE

Many approximations and extremely simple assumptions are necessarily involved in setting up any form of earthquake magnitude scale. It was surprising that the first attempt worked as well as it did, and it is still more astonishing that the extension to large distances and large earthquakes has continued to provide internally consistent results suitable for many statistical purposes.

Consequences of instrumental peculiarities.—The initial success of the scale appears now to have been due to a number of fortunate circumstances. The torsion seismometers are particularly stable, and write very legible records for small or moderately large local shocks. In this range of magnitude the prevailing periods of maximum waves are such that the magnification of the torsion instrument is nearly constant. For larger shocks, where difficulties might arise, the recorded amplitudes are too large for these instruments; hence the scale was supplemented by using other data. Magnitude determination for large shocks and at large distances from body waves was accordingly based in effect not on the recorded amplitude but on the

TABLE 20

ELEMENTS OF SOUTHERN CALIFORNIA EARTHQUAKES FOR AVERAGE GROUND CONDITIONS IN METROPOLITAN CENTERS

(For effect of ground, see text)

	M							
	2	3	4	5	6	7	8	8½
I_0	1½	3	5	6½	7½	9½	11	12
A_0 (cm.)	....	....	0.01	0.08	0.3	1½	5	20
a_0 (cm/sec.^{2})	1	3	12	50	150	500	800±	≧1000±
a_0/g	....	....	0.01	0.05	0.15	0.5	0.8	≧1±
r(km.)	0	15	80	150	220	400	600	....
log E	13.5	15.3	17.1	18.8	20.3	21.7	23.1	23.7
T_0 (sec.)	0.1	0.15	0.2	0.25	0.3	≧0.4	≧½	1±
T (Δ = 100 km.)	....	0.2	0.3	0.4	½±	¾±	1–2	....
t_0 (sec.)	½	¾	1	3	5	10±	20±	30±
q_0	0.95	1.73	2.43	3.11	3.73	4.29	4.79	5.03

particle velocity. Using this procedure as a general basis for the entire scale would remove this with other inconsistencies and would have many other advantages.

The original reliance on the torsion seismometer was unfortunate so far as this instrument has response characteristics differing from those of most other seismographs. It is a displacement meter for periods below one second and an accelerograph for long periods. This is one of the reasons for the complicated form of the empirical magnitude-energy relation, since the periods of the maximum waves increase with increasing magnitude and distance.

If one attempts to calculate magnitudes for large shocks from accelerograph readings at short distances by computing the response of a torsion seismometer to the recorded accelerogram maximum, the result is usually much too low. A standard torsion seismometer under the same circumstances writes a seismogram the maximum of which is of relatively long period and consequently is not the maximum wave of the accelerogram. An example is the accelerogram at Taft on July 21, 1952 (table 5). To pass from log A to log B we add 0.4 to compensate for ground. From table 1, log b = -2.6, whence $M = 4.0 + 2.6 = 6.6$, to be compared with the well-established value $M = 7.6$ for this earthquake.

For energy calculations, and probably for the projected revision of the magnitude scale, the response to A/T is especially important. The maximum of A/T is more reliably determined from the USCGS accelerograms than from torsion seismometers; strong-motion seismometers of longer period, like those at Pasadena, are still less suitable. The torsion seismometer compares with the USCGS accelerographs (most of which have free periods near 0.07 sec.) about as the usual Wiechert seismograph does with the torsion seismometer.

Shortcomings of the theory.—The energy calculation culminating in equation (20) assumes a point source. This is obviously inappropriate for shocks of large magnitude, considering linear extent of faulting, dimensions of crustal blocks displaced, etc. For the largest shocks it may be necessary to consider change in elastic properties for large displacements.

If the source of even a small shock is in or close to a low-velocity channel, slight change in depth may bring about relatively large change in amplitudes at the surface; this may account for some observed irregularities in determining magnitude.

Increase in period with magnitude.—This increase, which complicates the calculation of energy, is attributable to various causes, some of which also affect the applicability of the point-source assumption. Among these are:

1) Larger linear extent of the origin in all dimensions.

2) Effect of progression of faulting, resulting in a moving source and in shifting direction from which the waves arrive at the nearest station.

3) Approach to the limiting strength of rocks in large earthquakes.

The question of duration.—This question arises in several forms. As it enters into the energy calculation, appropriate data for it are difficult to obtain. The variation of duration of any phase of the seismogram with Δ and M is almost unknown. The effect of ground on the duration of body waves is large (fig. 2). The effect of depth h on t_0 is quite uncertain; this introduces further uncertainty into the magnitude-energy relation for deep shocks. Judging by the generally shorter periods and shorter wave groups recorded, t_0 appears to be shorter for deep shocks than for shallow shocks; this would imply a lower energy for given M in deep shocks.

Determination of magnitude at large distances.—Theoretically, this should be based rather on body waves than on surface waves; unfortunately, readings of amplitude and period for body waves are less generally available. Establishing an accurate magnitude-energy relation would call for integrating significant portions of seismograms with reference to the path of each individual wave, which can be considered as known only for the principal identified phases. On the seismograms of large shocks successive phases overlap; it is impossible to tell how much of the recorded complexity is due to the source and how much to wave propagation. To make the best use of body waves, improvement of the fundamental tables and charts giving magnitude in terms of Δ, h, and A/T is desirable.

The disadvantage of using surface waves to determine magnitude appears from the well-known theorem that for a source of given energy the amplitudes of the surface waves include a factor $e^{-ch/\lambda}$, where c is constant. This causes no serious trouble in working with the surface waves recorded at large distances with T near 20 seconds and wave lengths consequently of the order of 70 km.; uncertainties in magnitude then seldom exceed 0.2 unit unless h is more than 40 km. However, when we consider the generation of surface waves nearer the epicenter, we find that the factor $e^{-ch/\lambda}$ may well account in large part for the effects shown in figure 9. For the smaller shocks λ becomes much shorter, so that h/λ is larger, and there is considerable decrease in the surface waves relative to the body waves; it follows that relatively small surface waves are not necessarily an indication of deep focus when the magnitude is lower than 6, say. In large earthquakes, faulting extends toward the surface; this has the effect of decreasing h/λ and increasing the surface waves.

Conclusion

A revision of the magnitude scale is clearly desirable. Present studies are directed toward a scale based on the quotient (A/T) rather than on amplitudes.

The best means now available for estimating earthquake energies is that afforded by the magnitude-energy equation (20), after the magnitude has been fixed as well as possible by the methods described near the beginning of the present paper.

A question which, in view of the immediate prospect of revision, becomes somewhat academic relates to correctness of the magnitude numbers now assigned to large shocks, in terms of the scale as originally defined. It is believed that magnitudes determined from body waves of teleseisms are more coherent with the original scale than those from surface waves, which have been in use as the general standard. If further investigation confirms this, it will be established that the magnitudes above 7 determined from surface waves have been overestimated, while those below 7 are underestimated.

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References

Anonymous

1950. *The Effects of Atomic Weapons* (Washington, D.C.: U. S. Government Printing Office.) 456 pages.

Båth, M.

1952. "Earthquake Magnitude Determination from the Vertical Component of Surface Waves," *Trans. Am. Geophys. Union*, 33: 81–90.

1954. "Various Methods for Earthquake Magnitude Determination," *Assoc. Internat. Seism.*, meeting at Rome, abstract. See also "The Problem of Earthquake Magnitude Determination," in press.

1955. "The Relation between Magnitude and Energy of Earthquakes," *Trans. Am. Geophys. Union, 36: 861–865.*

Benioff, H.

1955. "Seismic Evidence of Crustal Structure and Tectonic Activity," Geol. Soc. Am., *Special Papers 62*, pp. 61–73.

Benioff, H., B. Gutenberg, and C. F. Richter

1954. "Progress Report, Seismological Laboratory, California Institute of Technology, 1953," *Trans. Am. Geophys. Union*, 35: 979–987.

Cancani, A.

1904. "Sur l'emploi d'une double échelle sismique des intensités, empirique et absolue," *Gerlands Beitr. z. Geophysik*, Ergänzungsband 2, pp. 281–283.

Di Filippo, D., and L. Marcelli

1949. "Sul movimento iniziale delle onde sismiche registrate a Roma durante il periodo 1938–1943," *Ann. di Geofisica*, 2: 589–606.

1950. "Magnitudo ed energia dei terremoti in Italia," *Ann. di Geofisica*, 3: 337–348.

Galitzin, B.

1915. "Earthquake of February 18, 1911," *Comptes Rendus*, 160: 810–813.

Gutenberg, B.

1936. "Periods of the Ground in Southern California Earthquakes," U.S.C.&G.S., *Spec. Publ. 201*, pp. 163–225.

1943. "Earthquakes and Structure in Southern California," *Bull. Geol. Soc. Am.*, 54: 499–526.

1945*a*. "Amplitudes of Surface Waves and Magnitudes of Shallow Earthquakes," *Bull. Seism. Soc. Am.*, 35: 3–12.

1945*b*. "Amplitudes of P, PP, and S and Magnitude of Shallow Earthquakes," *Bull. Seism. Soc. Am.*, 35: 57–69.

1945*c*. "Magnitude Determination of Deep-Focus Earthquakes," *Bull. Seism. Soc. Am.*, 35: 117–130.

1945*d*. "Variations in Physical Properties within the Earth's Crustal Layers," *Am. Jour. Sci.*, Ser. A, 243: 285–312.

Gutenberg, B., and C. F. Richter

1936. "On Seismic Waves (Third Paper)", *Gerlands Beitr. z. Geophysik*, 47: 73–131.

1942. "Earthquake Magnitude, Intensity, Energy, and Acceleration," *Bull. Seism. Soc. Am.*, 32: 163–191.

1946. "Seismic Waves from Atomic Bomb Tests," *Trans. Am. Geophys. Union*, 27: 776.

1949, 1954. *Seismicity of the Earth* (Princeton Univ. Press). 1st ed., 273 pages; 2d ed. 310 pages.

Housner, G. W.

1952. "Intensity of Ground Motion during Strong Earthquakes," Calif. Inst. of Tech., *Techn. Report Project NR-081-095.* 59 pages.

Jeffreys, H.

1923. "The Pamir Earthquake of 1911 February 18, in Relation to the Depths of Earthquake Foci," *Mon. Not. Roy. Astron. Soc., Geophys. Suppl.*, 1: 22–31.

Klotz, O.

1915. "Earthquake of February 18, 1911," *Bull. Seism. Soc. Am.*, 5: 206–213.

Neumann, F.
1954. *Earthquake Intensity and Related Ground Motion* (Univ. of Washington Press). 94 pages.
Peterschmitt, E.
1950. "Etude de la magnitude des séismes," *Ann. de l'Instit. de Physique du Globe, Strasbourg,* 6: 51–58.
Reid, H. F.
1910. *The California Earthquake of April 18, 1906*, Vol. 2: *The Mechanics of the Earthquake* (Carnegie Inst. of Washington, Publ. No. 87). 192 pages.
1912. "The Energy of Earthquakes," *Comptes Rendus, Assemblée Assoc. Internat. Sismol. Manchester*, 1911, pp. 268–274.
Richter, C. F.
1936. "An Instrumental Earthquake Magnitude Scale," *Bull. Seism. Soc. Am.*, 25: 1–32.
1948. "History and Applications of the Magnitude Scale," *Bur. Central Seismol. Internat.*, Ser. A, 17: 217–224.
Richter, C. F., and J. M. Nordquist
1948. "Minimal Recorded Earthquakes," *Bull. Seism. Soc. Am.*, 38: 257–261.
Sagisaka, K.
1954. "On the Energy of Earthquakes," *Geophys. Mag. Tokyo*, 26: 53–82.
Sieberg, A.
1923. *Erdbebenkunde* (Jena). 572 pages.
Trapp, E.
1954. "Ableitung der Magnitudengleichung für die Erdbebenstationen Wien und Graz und allgemeine Bemerkungen zur Magnitudenberechnung," *Archiv f. Meteorol., Geophysik und Bioklim.*, Ser. A, 6: 440–450.
Tsuboi, Ch.
1940. "Isostasy and Maximum Earthquake Energy," *Proc. Tokyo Acad.*, 16: 449–454.
1951. "Determination of the Richter-Gutenberg's Instrumental Magnitudes of Earthquakes Occurring In and Near Japan," *Geophys. Notes, Tokyo Univ.*, Vol. 4, No. 5, pp. 1–10.
Vaněk, J.
1953. "Determination of Earthquake Magnitude from Surface Waves for the Stations Hurbanovo and Skalnaté Pleso," *Travaux Inst. Geophys. Acad. Tschécoslav.*, No. 6, pp. 83–89.
Zátopek, A., and J. Vaněk
1950. "On the Regional Repartition of Magnitude Differences between Pasadena and Praha," *Kartograf. přehled Praha*, 5: 41–55, 123–128.

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