Bulletin of the Seismological Society of America

VOL. 25 JANUARY, 1935 No. 1

AN INSTRUMENTAL EARTHQUAKE MAGNITUDE SCALE*

BY CHARLES F. RICHTER

In the course of historical or statistical study of earthquakes in any given region it is frequently desirable to have a scale for rating these shocks in terms of their original energy, independently of the effects which may be produced at any particular point of observation. On the suggestion of Mr. H. O. Wood, it is here proposed to refer to such a scale as a "magnitude" scale. This terminology is offered in distinction from the name "intensity" scale, now in general use for such scales as the Rossi-Forel and Mercalli-Cancani scales, which refer primarily to the local intensity of shock manifestation.

The writer is not aware of any previous approach to this problem along the course taken in this paper, except for the work of Wadati cited below. Total original energies have been calculated for a number of shocks, using seismometric and other data; but such a procedure is practicable only for a limited number of cases, whereas it is desired to apply a magnitude scale to all or nearly all of the shocks occurring.

Mr. Maxwell W. Allen states that he has for some time employed an arbitrary scale for rating large earthquakes, based on the amplitudes of earth motion calculated from the reports of distant stations. This laborious procedure is not far removed in principle from that adopted in the following discussion. Doubtless it has also occurred to others, but has failed of general application because of its paucity of dependable results.

In the absence of any accepted magnitude scale, earthquakes have occasionally been compared in terms of the intensity on the Rossi-Forel or some similar scale, as manifested near the epicenter. Even when reliable information is obtainable, this method is obviously exposed to uncertainties arising from variations in the character of the ground, the depth of the focus, and other circumstances not easily allowed for. In a region such as Southern California, where a large proportion of the shocks occur

* [Received for publication June 17, 1934.]

in almost unpopulated districts, while still others are submarine in origin, any general procedure of this kind is out of the question.

Despite the evident difficulties the requirements of research, as well as the public interest, call for some estimate of the magnitude, in the sense here used, of each important shock in the California region. This led to an attempt at constructing a magnitude scale based on instrumentally recorded amplitudes at the seven stations of the Southern California group.

Precision in this matter was neither expected nor required. What was looked for was a method of segregating large, moderate, and small shocks, which should be based directly on instrumental indications, and thus might be freed from the uncertainties of personal estimates or the accidental circumstances of reported effects. The method used proved to be much more selective than had been anticipated, assigning observed earthquakes to as many as fifteen well-defined scale numbers, with possibilities of further extension and finer subdivision.

The procedure used was suggested by a device of Wadati,[1] who plotted the calculated earth amplitudes in microns for various Japanese stations against their epicentral distances. He employed the resulting curves to distinguish between shallow and deep earthquakes, to calculate the coefficient of absorption for surface waves, and to make a rough comparison between the magnitudes of several strong shocks.

On certain assumptions, which cannot hold to any high accuracy, it is possible to derive a quantitative magnitude scale from curves plotted in this way. Suppose two shocks of different magnitude were to take place at exactly the same focus, all other circumstances being identical in the two cases. Then any seismograph at a particular station should write two records, one of which should be very closely an enlarged copy of the other. The ratio of this enlargement should be the same for all seismographs, provided of course that the constants remain unaltered between the two events, and that the response of the registering apparatus is linear. This ratio could then be used to measure the relative magnitudes of the two shocks. With the given assumption that the mechanism of shock production is the same in the two cases, the ratio of the seismometric amplitudes is the square root of the ratio of the energies liberated.

In practice we have to compare shocks from different foci, and probably different also in the mechanism of occurrence. Comparison is thus rendered very inexact. However, useful results can be obtained by com-

[1] K. Wadati, *Geophysical Magazine* (Tokyo), **4**, 231, 1931.

parison of the records at several stations. It is necessary to establish empirically a relation between the maximum seismographic amplitudes of a given shock at various distances; this is done by assuming that the ratio of the maximum amplitudes of two given shocks, as registered by similar instruments at equal epicentral distances, is a constant. That is, if shock A is registered with maximum amplitude 5 millimeters at 75 kilometers and 2 millimeters at 200 kilometers, while shock B registers with maximum amplitude 15 millimeters at 75 kilometers, then shock B should register 6 millimeters at 200 kilometers.

The precision of magnitudes based on such an assumption is evidently impaired by a variety of conditions. The most obvious of these are the effects of inhomogeneity in the propagation of elastic waves, of varying depth of focus, of difference in mechanism of shock production, of the ground at the several stations, and of the instrumental constants.

The most serious of these difficulties is the first. In most cases energy appears to be radiated unequally in different azimuths from the point of origin. This may arise from the circumstances of origination of the shock (strike of the fault, nature of displacement on the fault) or from differences in geological structure along the various wave paths. When the records of a number of stations surrounding the epicenter are available, this effect can be allowed for to some extent; but it remains an obstacle in the way of any precise determination of earthquake magnitudes, which can only be overcome with the advent of a more detailed understanding of the dynamics of shock production, and more complete information as to the various local structures.

Variation in depth of focus is less important. The majority of shocks in this region appear to originate at depths not far different from 15 kilometers. The effect of even considerable departures from this level can be reduced, for all but the smallest shocks, by using the records at stations distant 100 kilometers or over.

It is nearly certain that in most, though not all, of the stronger shocks the distribution of energy among various frequencies is not the same as for weaker shocks. Especially when there is evidence of extended movement along a fault, a high proportion of energy appears to go into waves of long period. As the maximum phase on the seismograms usually exhibits longer periods than the beginning of the record, the effect is to exaggerate the maxima. Comparison with the recorded maxima of a smaller shock then leads to an overestimate of the difference in magnitude. Fortunately, this effect does not appear to be larger than the other sources of error; and with long experience, or with more precise theories of shock production, it should be possible to take it into account quantitatively.

In comparing records from different stations, conclusions are affected by differences in ground and in the instrumental constants. If the latter are known with precision, the periods of the seismographic maxima may be measured, and the actual amplitudes of earth displacement calculated and used for estimating magnitudes. The procedure is somewhat laborious; and, as will appear presently, it can be dispensed with if the constants of the various instruments are approximately the same. The short period torsion seismometers installed at the Southern California stations are designed to have identical constants; but, owing to unavoidable irregularities in manufacture, some differences exist. It is not convenient to determine the constants from time to time; however, it is known that the constants of any one instrument remain relatively fixed over periods of years.

Determination of constants would make it possible to separate the purely instrumental effects from those due to ground; but, because of the uncertain elements in the latter, no great access of precision in estimating magnitudes would follow. In practice it is considered that the effect of ground and that of the instrument combine in each case into a fairly uniform deviation from the mean registered amplitudes for all stations and instruments; so that statistical study of a group of shocks will lead to average corrections applicable to the amplitudes registered by each individual instrument. These corrections turn out to be small, and of the same order as fluctuations due to other causes.

For precise purposes, it would be desirable to identify the phases of each seismogram, and to compare amplitudes of the same wave or set of waves at the various distances. Such identification is difficult and questionable for many of the smaller shocks, and is too time-consuming for use in routine work where hundreds or thousands of shocks must be dealt with. Thus the scale has been set up on the basis of measurements of the maximum recorded amplitude. This maximum will of course not always correspond to the same wave-group or phase. It will change especially with distance, coinciding with $\bar{S}$ or Q for very near shocks, at intermediate distances with some member of the complicated S series of phases, and at the larger distances with a slow surface wave. However, if the magnitude scale is set up empirically for the measured maximum amplitudes, these considerations do not directly affect its precision. If it were strictly true that all seismograms written by identical instruments at any one distance were simply enlarged or reduced copies of one another, such an empirical scale would apply perfectly, and magnitudes derived from it would be exact.

The foregoing considerations are preliminary to the actual setting up of a workable empirical scale of magnitudes. To derive such a scale, a representative group of shocks (those of January, 1932) was carefully studied, and the logarithm of the recorded amplitude in each case plotted against the epicentral distance. Curves were drawn through the several points referring to each shock, and were seen to be roughly parallel, as the hypothesis of proportional amplitudes requires. These were then combined into a single curve, parallel to the individual shock curves, and passing through an arbitrarily selected point. From this composite curve were read off the numerical values presented in Table I.

Table I gives the logarithm (to the base 10) of the calculated amplitude, in millimeters, with which the standard short-period torsion seismometer ($T_0 = 0.8$ sec., $V = 2{,}800$, $h = 0.8$) should register at various distances an earthquake of standard magnitude; this is chosen so that the calculated amplitude of registration at an epicentral distance of 100 kilometers is 0.001 millimeters (1 micron).

Note that the logarithms are all negative, as all the amplitudes are less than one millimeter. However, they are given as negative quantities instead of in the usual common-logarithm form of negative characteristic and positive mantissa. Thus, the tabulated logarithm at 65 kilometers is -2.79; in the usual notation this would be given as $7.21 - 10$ or $\bar{3}.21$. The form used in Table I is more convenient in the actual calculation of magnitude.

Table I can be applied to assign a magnitude scale number to any shock for which measured amplitudes at known epicentral distances are available. The following procedure is in routine use: the measured amplitude at any station is expressed in millimeters, and the logarithm of the number is taken. From this is algebraically subtracted the logarithm appearing in Table I opposite the given epicentral distance. The result is taken as the magnitude scale number, and is clearly the logarithm of the ratio of the amplitude of the given shock to that of the standard shock, represented by Table I, at the same epicentral distance.

As a numerical example, suppose a shock recorded with an amplitude of 5 millimeters at 225 kilometers. The logarithm of 5 is 0.70; opposite 225 kilometers we find -3.68; hence the magnitude is $0.70 - (-3.68) = 4.38$.

When this calculation is carried out separately for each station at which the shock is recorded, the magnitude scale numbers found at the

TABLE I

LOGARITHMS OF THE AMPLITUDES (IN MILLIMETERS) WITH WHICH THE STANDARD TORSION SEISMOMETER ($T_0 = 0.8$, $V = 2{,}800$, $h = 0.8$) SHOULD REGISTER A SHOCK REGISTERED AT $\Delta = 100$ KILOMETERS WITH AN AMPLITUDE OF 0.001 MILLIMETERS (1 MICRON)

Δ(km)	Log *A*	Δ(km)	Log *A*	Δ(km)	Log *A*
25	—1.65	205	—3.56	405	—4.48
30	—2.10	210	—3.59	410	—4.50
35	—2.32	215	—3.62	415	—4.51
40	—2.43	220	—3.65	420	—4.52
45	—2.54	225	—3.68	425	—4.54
50	—2.63	230	—3.70	430	—4.56
55	—2.70	235	—3.72	435	—4.57
60	—2.77	240	—3.74	440	—4.59
65	—2.79	245	—3.77	445	—4.61
70	—2.83	250	—3.79	450	—4.62
75	—2.87	255	—3.81	455	—4.63
80	—2.90	260	—3.83	460	—4.64
85	—2.94	265	—3.85	465	—4.66
90	—2.96	270	—3.88	470	—4.68
95	—2.98	275	—3.92	475	—4.69
100	—3.00	280	—3.94	480	—4.70
105	—3.03	285	—3.97	485	—4.71
110	—3.08	290	—3.98	490	—4.72
115	—3.10	295	—4.00	495	—4.73
120	—3.12	300	—4.02	500	—4.74
125	—3.15	305	—4.05	505	—4.75
130	—3.19	310	—4.08	510	—4.76
135	—3.21	315	—4.10	515	—4.77
140	—3.23	320	—4.12	520	—4.78
145	—3.28	325	—4.15	525	—4.79
150	—3.29	330	—4.17	530	—4.80
155	—3.30	335	—4.20	535	—4.81
160	—3.32	340	—4.22	540	—4.82
165	—3.35	345	—4.24	545	—4.83
170	—3.38	350	—4.26	550	—4.84
175	—3.40	355	—4.28	555	—4.85
180	—3.43	360	—4.30	560	—4.86
185	—3.45	365	—4.32	565	—4.87
190	—3.47	370	—4.34	570	—4.88
195	—3.50	375	—4.36	575	—4.89
200	—3.53	380	—4.38	580	—4.90
		385	—4.40	585	—4.91
		390	—4.42	590	—4.92
		395	—4.44	595	—4.93
		400	—4.46	600	—4.94

several stations normally agree within one unit or less, especially when allowance is made for certain instruments which regularly register unusually large or small amplitudes. Accordingly, in published reports the magnitude is stated to the nearest half-unit of the logarithm. This means that the true energy of the shock may be more than three times larger or smaller than that computed from the given magnitude number.

The procedure may be interpreted to give a definition of the magnitude scale number being used, as follows: *The magnitude of any shock is taken as the logarithm of the maximum trace amplitude, expressed in microns, with which the standard short-period torsion seismometer ($T_0 = 0.8$ sec., $V = 2800$, $h = 0.8$) would register that shock at an epicentral distance of 100 kilometers.*

This definition is in part arbitrary; an absolute scale, in which the numbers referred directly to shock energy or intensity measured in physical units, would be preferable. At present the data for correlating the arbitrary scale with an absolute scale are so inadequate that it appears better to preserve the arbitrary scale for its practical convenience. Since the scale is logarithmic, any future reduction to an absolute scale can be accomplished by adding a constant to the scale numbers.

Table I presents, unaltered, the result of studying a comparatively small group of shocks—those of January, 1932. This was immediately applied to the shocks of subsequent months. The magnitude numbers computed from the several stations should agree within reasonable limits; and this is the case. A representative example is the shock of February 15, 1932, recorded at stations distant 39, 100, 107, 255, 260, and 345 kilometers with amplitudes of 6, 3, 1.2, 0.3, 0.3, and 0.2 millimeters. The magnitudes derived from Table I are then 3.20, 3.48, 3.13, 3.29, 3.31, and 3.54. Considering the large range in distance and amplitude, neither of which is determined with precision, the range of 0.41 in the computed magnitude is surprisingly small.

Instances of this kind could easily be multiplied; but a very much better test of the method is available. For his study of the propagation of seismic waves in Southern California[2] Professor Gutenberg employed the records of twenty-one well-registered earthquakes. The epicenters of these shocks are thus determined with unusual accuracy; although, as may be seen from Table I, slight errors in distance will not much affect computed magnitudes, these shocks afford the most reliable test of the

---

[2] B. Gutenberg, "Travel Time Curves at Small Distances, and Wave Velocities in Southern California," *Gerlands Beiträge zur Geophysik,* **35**, 6, 1932.

magnitude scale. Gutenberg's results are given in Table II; the letters in the first column were assigned by him for purposes of identification.

TABLE II

Shock	Date	Epicenter North Latitude	Epicenter West Longitude	Location
A	Aug. 17, 1930	35° 13′	116° 51′	Mojave Desert
B	Sept. 26, 1929	34 50	116 31	Near Newberry, Mojave Desert
C	May 28, 1930	35 30	117 14	Garlock fault near Searles Lake
D	April 20, 1930	34 39	117 04	Northeast of Victorville
E	Feb. 24, 1930	34 57	117 02	Near Barstow
F	Jan. 8, 1931	34 56	117 03	Near Barstow
G	Jan. 15, 1930	34 11	116 55	San Bernardino Mountains
H	Jan. 15, 1930	34 11	116 55	San Bernardino Mountains
J	April 23, 1931	35 25	117 36	Near Trona
K	April 27, 1931	34 21	116 17	Southern Mojave Desert
a	Oct. 31, 1929	33 38	118 12	San Pedro Channel
b	Sept. 13, 1929	33 38	118 12	San Pedro Channel
c	Nov. 8, 1929	35 46	120 28	West of Coalinga
d	Aug. 30, 1930	33 56	118 37	Santa Monica Bay
e	May 12, 1930	33 12	116 43	San Diego County (Elsinore fault)
f	Jan. 17, 1931	37 35	118 03	Southeastern Mono County
g	Aug. 18, 1930	34 26	120 11	Off Point Concepcion
h	Feb. 23, 1931	35 46	120 40	Northeast of Paso Robles
i	April 24, 1931	33 46	118 29	Off Point Vicente
k	April 21, 1931	35 19	118 55	Bakersfield district
l	April 29, 1931	34° 15′	118° 39′	Near Chatsworth

In the following tabulations the letters P, MW, R, SB, LJ, T, H, are used as abbreviations for the names of the stations at Pasadena, Mount Wilson, Riverside, Santa Barbara, La Jolla, Tinemaha, and Haiwee, respectively.

Table III gives the epicentral distances for the several shocks at the recording stations; the distances are either as given by Gutenberg, or as measured from a map with an error not over two kilometers. Table IV gives the maximum seismographic trace amplitudes in each case. Where possible, the amplitudes for both horizontal components are given, that for the N component being entered above that for the E component. The readings in parentheses for the N component at Pasadena are to be doubled in computing magnitudes, since they refer to a period when the optical system on this particular instrument was so arranged as to give only half the usual magnification. Where the reading is followed by a + it may be considerably less than the true maximum, owing to photographic underexposure.

TABLE III

Epicentral Distances (kilometers)

Shock	P	MW	R	SB	LJ	T	H
A	173	160	146	278	263	235	140
B	171	157	122	295	226	291	192
C	176	163	170	257	292	194	95
D	117	103	75	245	197	295	181
E	139	124	111			261	155
F	137		109	253		262	155
G	118	106	48	265	142	345	230
H	118	106	48	265	142	345	230
J	178	166	179	250	302	179	78
K	180	170	110	320	186		249
a	54	63	86	165	123	378	274
b	54	63	86	165	123	378	274
c	276	280	347	168	440	248	235
d	46	59	117	113	176	351	247
e	178	175	108	308	57	454	345
f	381		403	381	530	59	162
g	188	199	267	45	324	343	279
h	288		362	175		265	244
i	49	63	108	132	153	366	260
k	145	144	204	124		205	120
l	44	54	123	100	200	315	216

Table V gives the magnitudes of the shocks as calculated from Table I and the data of Tables III and IV. The arrangement of the data for the two components is the same as for Table IV. At the right of the table is entered the mean of all the magnitudes calculated for each shock.

It is evident that the deviations of the individual determinations of magnitude from the mean for each shock are numerically small. Since the distances range from 44 to 530 kilometers, this is good evidence for the validity of Table I. It will be observed that the mean magnitudes of the several shocks do not differ greatly. This is a reasonable result, as the shocks used in Gutenberg's study were necessarily moderately strong, and no very strong shock occurred in the region during the interval for which records were then available.

Still closer agreement can be obtained if attention is given to the behavior of each individual instrument. Thus, the E component at Tinemaha regularly registers larger amplitudes than the mean. If the excess of the magnitudes calculated from this instrument over the mean magnitude is determined for each shock, and the average taken for all shocks, the result is 0.40. This is a quantity to be subtracted from the magnitude calculated from this instrument, as a correction for the ground conditions and instrumental constants.

TABLE IV

Maximum Seismographic Trace Amplitudes (Millimeters)

Upper Figures Refer to N–S Component, Lower to E–W

Shock	P	MW	R	SB	LJ	T	H
A	(0.5)	2.3	1.9	0.9	1.1	1.1	2.6
	0.8	1.9	2.8	1.1	1.2	1.1	2.2
B		27		24		14	25
	56+			23		17.5	
C	(1.4)	4.4	2.3	4.2	2.5	4.5	
	2.5	3.9	3.1	3.8	2.3	3.9	6.7
D	(0.9)	2.9	1.8	1.4	1.9	0.6	0.8
	1.4	2.9	1.9	0.8	2.3	0.7	
E	(1.8)		3.1			1.1	1.2
	4.6	3.1	3.5			2.4	1.4
F			3	1.9		1.8	2.2
	3.3		2.8	1.9		1.8	3.0
G	(39)	88+			66+	24	39+
	39+				60+	54	58
H	(38)	82			80+	10	23
	30+			26	56+	19.5	31
J		12.2	2.9	4.6	2.2	5.1	22.5
	7.2	13.4	4	4.4	3.6	7.2	
K	3.6	2.5		0.7	3.9		1.6
	2.5	3.3		1.0	5.2		1.7
a	1.5	3.4	12.1	2.0	5.7	0.4	0.3
	1.9	6.3	4.8	2.1	4.5	0.6	0.6
b	5.4		6.5	3.8		0.8	1.1
		11.6	11	3.8	9	1.2	1.3
c		1.9	0.7	14.3	0.3	8.2	7.7
			0.8	12.2	0.3	15.9	5.2
d				65	52+	33	37
	135+			61+	60+	39	31
e	(1.8)	2.9	9.0	2.1	20.2	1.2	1.3
	3.8	5.0	11.1	1.6	15.6	2.0	1.6
f	(0.2)		0.2	1.3	0.2	29	1.9
	0.4		0.3	1.8	0.2	36	2.2
g	(0.5)	1.4	0.4	22.7	0.4	0.7	0.9
	1.4	1.7	0.3	22.1	0.5	1.0	0.7
h	(0.4)		0.6	6.1		5.7	4.2
	1.1		0.3	5.7			
i	30			28	20	1.6	2.6
	19.3	25		29	20.4	4.8	
k	1.7	4.3	0.3	2.9		2.6	1.0
	1.7	2.3	0.3	3.1		3.0	
l		8+	5.0	8.9	2.1	0.6	1.7
	13.1	12.1	4.1	6.6	3.8	1.6	1.6

TABLE V

Calculated Shock Magnitudes

Upper figures refer to N–S component, lower to E–W

Shock	P	MW	R	SB	LJ	T	H	Mean
A	3.39	3.68	3.56	3.88	3.88	3.76	3.64	3.69
	3.29	3.60	3.73	3.97	3.92	3.76	3.57	
B		4.74		5.38		5.13	4.88	5.12
	5.12+			5.36		5.22		
C	3.86	3.98	3.74	4.44	4.39	4.14		4.06
	3.81	3.93	3.87	4.40	4.35	4.08	3.81	
D	3.37	3.48	3.13	3.92	3.79	3.78	3.33	3.54
	3.26	3.48	3.15	3.67	3.87	3.85		
E	3.79		3.57			3.87	3.38	3.71
	3.89	3.63	3.62			4.21	3.45	
F			3.55	4.08		4.10	3.64	3.84
	3.74		3.52	4.08		4.10	3.78	
G	5.00	4.98			5.07+	5.62	5.29+	5.24
	4.70+				5.04+	5.97	5.46	
H	4.99	4.95			5.15+	5.24	5.06	5.10
	4.59+			5.26	5.00+	5.53	5.19	
J		4.45	3.88	4.45	4.37	4.13	4.24	4.30
	4.28	4.49	4.02	4.43	4.59	4.28		
K	3.99	3.78		3.97	4.04		3.99	3.98
	3.83	3.90		4.12	4.17		4.02	
a	2.88	3.31	4.02	3.65	3.90	3.97	3.39	3.61
	2.98	3.58	3.62	3.67	3.79	4.15	3.69	
b	3.42		3.75	3.93		4.27	3.95	3.97
		3.84	3.98	3.93	4.09	4.45	4.02	
c		4.22	4.10	4.53	4.07	4.69	4.61	4.39
			4.15	4.46	4.07	4.98	4.44	
d				4.90	5.12+	5.78	5.35	5.22
	4.67+			4.88+	5.18+	5.85	5.27	
e	3.98	3.86	4.01	4.38	4.03	4.71	4.35	4.22
	4.00	4.10	4.11	4.26	3.92	4.93	4.44	
f	3.98		3.77	4.49	4.10	4.23	3.61	4.07
	3.98		3.95	4.64	4.10	4.32	3.67	
g	3.46	3.68	3.46	3.90	3.74	4.08	3.89	3.76
	3.61	3.76	3.34	3.88	3.84	4.23	3.79	
h	3.88		4.09	4.19		4.61	4.38	4.14
	4.02		3.79	4.16				
i	4.09			4.51	4.50	4.52	4.24	4.40
	3.90	4.18		4.52	4.51	5.00		
k	3.51	3.90	3.03	3.60		3.97	3.12	3.54
	3.51	3.64	3.03	3.62		4.04		
l		3.59+	3.84	3.95	3.85	3.88	3.86	3.86
	3.64	3.79	3.75	3.82	4.11	4.31	3.83	

TABLE V—*Continued*

Shock	P	MW	R	SB	LJ	T	H	Mean
Mean correction	+0.23	+0.13	+0.21	−0.13	−0.04	−0.24	+0.08	
	+0.25	+0.06	+0.20	−0.12	−0.07	−0.40	+0.02	

Corrections of this type were determined for each instrument, as given in Table V, and duly applied. It was expected that the corrected values could be used to improve Table I by plotting the deviations from the mean magnitude against the distance. The result was disappointing, in that the deviations showed no clear dependence on distance, but rather a uniform scattering on both sides of the zero line. As these observed deviations nearly all were less than half a unit, it follows that the values of Table I are not in error by more than half a unit; the error in taking these values as representative of the normal conditions is smaller. However, it also follows that Table I cannot be improved from the data in hand. The conclusion is that the sources of irregularity mentioned previously are such that a tabulation of greater precision is impossible until better data are available for eliminating such effects.

The curve whose co-ordinates are given in Table I is evidently not capable of being represented by a simple continuous function. The amplitude falls off much more rapidly at the smaller distances than it does beyond 150 kilometers. In all probability the maximum amplitudes at the larger distances are determined by surface waves, while at the smaller distances these amplitudes represent waves such as $\bar{\bar{S}}$, S_x, S_m. The curve is not very well determined between 25 and 50 kilometers, as the amplitude varies so rapidly with distance that only exceptionally well-located shocks can be used for revision, while the uncertainties due to varying depth and to effects of local structure are much intensified.

There is no immediate prospect of extending Table I to distances less than 25 kilometers. Shocks occurring so close to any station, if sufficiently large to be clearly registered at other stations, usually record at the small distance with unmanageably large amplitudes. The only good prospect of surmounting this difficulty is in the occurrence of a group of shocks of various magnitudes at a focus within 25 kilometers of one of the stations. The larger shocks can then be precisely located, and their magnitudes determined, from the records of the other stations; magnitudes of smaller shocks can be determined from the ratios of their amplitudes to those of the larger ones, and in this way a shock may be found with legible amplitude at the near station to which both magnitude and distance can be assigned with some confidence. On a later page will be found the data for such a group of shocks occurring about 13 kilometers from the station at Haiwee; but in this case the epicenter is not known with sufficient precision.

For extending Table I to greater distances we require a large shock, favorably placed with respect to the stations, and with an epicenter well located on the basis of seismometric or field data. This combination of circumstances thus far has not taken place; but a rather puzzling numerical relation exists which may provide a substitute. It should be clearly understood that the numerical values of Table I were arrived at on a purely empirical basis; they were read from a smooth curve which was drawn to fit the observations. It is consequently a little surprising to find that from 200 to 600 kilometers Table I is represented with considerable accuracy by the formula

$$A = 2{,}350\Delta^{-3} \quad \text{or} \quad \log A = 3.37 - 3 \log \Delta$$

This formula is in all probability a fortuitous result without much physical significance, as theory indicates more complicated conditions. However, it may prove to be of considerable practical value in extending the method to larger distances, and in comparing results of the same character obtained in other regions or with other instruments. For distances less than 200 kilometers the observations show amplitudes smaller than those given by the formula above, the discrepancy becoming very large at the shortest distances used.

Table I has been in routine use for the study of shock magnitudes for more than two years, during which period no shock has been recorded for which the difference between the magnitudes calculated from the records at the several stations was exceptionally large.[3] It is always possible to assign a magnitude to half a unit, and for well-observed shocks a magnitude to a tenth of a unit is easily given as the value which best fits the observations.

Lest the impression should be created that great precision is being claimed for this method, it is desirable here to emphasize its actual crudity. In ordinary routine, as has just been mentioned, shock magni-

[3] Certain shocks southwest of Pasadena, of which *a*, *b*, and *i* of Tables II–V are examples, record at that station and at Mount Wilson with abnormally small amplitudes. This appears to be an effect of local structure, although in some cases there is evidence that the shocks originate at unusual depth (about 25 kilometers instead of the normal 15 kilometers).

tudes are assigned to half a unit. This means that the shocks are distributed into groups, each consisting of shocks which have roughly ten times the initial energy of those assigned to the next lower group. Obviously no great accuracy is required to distinguish instrumentally between such widely separated levels of magnitude; and the matter would be of no practical importance, were it not for the fact that the actual range of magnitude in observed shocks is so immense that even so rough a division separates it into a convenient number of levels.

In choosing the arbitrary zero of the magnitude scale, the intention was to make it coincide with the smallest shocks registered. The choice made has been justified by further experience. The very smallest observed shocks, which are recorded only at very short distances, appear to be of magnitude 0; there is a slight uncertainty in this, on account of the difficulty in extending Table I to shorter distances.[4]

The largest shock to which the present method can be applied is the major earthquake in Nevada on December 20, 1932. As will be shown in detail below, the best evidence indicates a magnitude of about 7.5. The difference in the extreme observed magnitudes is thus 7.5 — 0, or 7.5. The logarithm of the ratio of the energies involved should be double this, or 15. That is, we are recording shocks with an extreme range in energy of at least 10^{15} to 1. From this it is conspicuously evident that adequate study of all earthquakes in an active region involves the distribution of instruments of several different grades of sensitiveness.

In order to demonstrate the reality of the computed grades of magnitude, there now follows a detailed discussion of the observed effects of a number of representative shocks. It may assist the reader to anticipate the general conclusions at this point; detailed results will be found on a later page. In general, shocks of magnitudes 0, 1, 2, are not reported as felt; shocks of magnitudes 3 and 4 are felt, but cause no damage; magnitude 5 may cause considerable minor damage; magnitude 6 is usually destructive over a limited area; and magnitudes 7 and 8 transgress the lower limit of major earthquakes.

The discussion of observed effects is best begun with the earthquakes of Tables II–V. Unfortunately, most of the group indicated by capital

---

[4] To forestall a possible misunderstanding, it should be pointed out that 0 of the magnitude scale does not refer to "zero shock," or "no shock." The scale is logarithmic; and as 0 is the logarithm of 1, 0 on the scale refers to the unit or standard shock, the logarithms of whose recorded amplitudes are given in Table I. A shock of magnitude 1 has ten times the recorded amplitude of the standard shock, magnitude 2 has 100 times the amplitude, etc.

letters are not very satisfactory for this purpose, as their location in a desert region renders reports scanty. Thus, for *A, C, D, E,* there is no information at hand, while for *B, F, K,* there are only one or two reports each. *G, H,* and to a less extent *J,* are well observed. Of the remaining shocks, only *d, g, h,* and *i* are represented by considerable groups of reports.

Table VI lists, in order of magnitude, the twelve shocks of Tables II–V for which data are sufficient to assign numerical values to the mean outer radius of the area of perceptibility (outer limit of II), and to the

TABLE VI

Shock	Magnitude	Outer Limit of II (km)	Outer Limit of IV (km)
d	5.2	120	90
G	5.2	240	150
B	5.1	170	
i	4.4	50	
c	4.4	60	
J	4.3	150	100
h	4.1	60	30
b	4.0	40	
l	3.9	25	
g	3.8	30	
F	3.8	>75	
a	3.6	20	

corresponding mean outer radius of intensity IV. These intensities, and all others given in this paper, refer to the modified Mercalli scale of 1931.[5]

Shock *d* is a very important case; this is the Santa Monica Bay earthquake, which has been discussed in a detailed publication.[6] The shaking seems to have been of unusually short duration at all points; this may account for the rather small radii given in the table. The distribution of higher intensities was so irregular that no sound conclusions can be drawn.

Shocks *G* and *H* occurred only ten minutes apart; as their epicenters were nearly identical, it is not possible to discuss them separately. The epicentral intensity was VII, or perhaps slightly less.

Shocks *a* and *b* also had nearly identical epicenters; but as they took

[5] Harry O. Wood and Frank Neumann, *Bulletin of the Seismological Society of America,* **21,** 277, 1931.

[6] B. Gutenberg, C. F. Richter, and H. O. Wood, "The Earthquake in Santa Monica Bay, California," *Bulletin of the Seismological Society of America,* **22,** 138, 1932.

place on different dates, it is possible to compare them; the shock of larger magnitude clearly had the greater radius of perceptibility.

In a number of cases some of the stations were located within the area of perceptibility; these instances are of interest for their bearing on the problem of earth motion during felt shocks. Thus, shock *d* was of intensity V in the city of Pasadena; it is quite clear, however, that the amplitude on the Pasadena records should have been in excess of the 135 millimeters given in Tables II–V. Calculation from Table I with magnitude 5.2 and distance 45 kilometers gives 460 millimeters.

Shock *G* was reported in the city of Pasadena as of various degrees of IV; but at the Seismological Laboratory it was not noticed. This type of observation is very frequent; it is undoubtedly due to the solid construction of the Laboratory building, and its foundation directly on granitic rock. The recorded amplitudes, as noted in the table, were about 40 millimeters. Note that the amplitudes at La Jolla are clearly larger, in spite of the greater distance. The shock was actually stronger in the near-by city of San Diego than might be expected, being high in intensity IV at most points. Thus both instrumental and macroseismic data agree in showing an uneven distribution of intensity in different azimuths; this is of course a general phenomenon, and, as already noted, is one of the principal obstacles to improving Table I.

Shock *i* is a very critical case for determining the limit of perceptibility. In the city of Pasadena the shock was positively observed, but by a few persons only; at the Seismological Laboratory it was quite imperceptible, this observation being very definite, as the shock was seen recording on an ink-writing instrument. As the amplitudes on the two components were 30 millimeters and 19 millimeters, this shock suggests 30 millimeters as the recorded amplitude at the lower limit of perceptibility in ordinary structures on ordinary ground. However, this should still be increased by the correction for the ground and instruments at Pasadena, as given in Table IV. The logarithm of this factor is slightly in excess of 0.2, which corresponds to the number 1.7. The corrected lower limit of registration for a felt shock thus is about 50 millimeters.

A somewhat smaller limiting amplitude would be derived from the seismograms of a shock originating near Santa Barbara on August 5, 1930, which was barely perceptible to a few persons in Pasadena, and which registered on the two components with amplitudes of 18.5 and 13.6 millimeters.

The distances given in Table VI appear at first rather irregular, and seem to bear no positive relation to the shock magnitudes. On the logarithmic basis these irregularities are less evident; from Table I and Table VI we can easily calculate the logarithm of the seismographic amplitude, in millimeters, at the outer limit of II, and find, for the several shocks:

Shock	*d*	*G*	*B*	*i*	*c*	*J*	*h*	*b*	*l*	*g*	*F*	*a*
Log amplitude	2.1	1.5	1.7	1.8	1.6	1.0	1.3	1.6	2.2	1.7	>1.0	2.0

The mean of these values is 1.7, and it will be observed that few of the individual values diverge more than 0.5 from this. The number whose logarithm is 1.7 is 50; that is, there is an exact coincidence, probably fortuitous, between this result and that derived above from the Pasadena record of shock *i*.

It is important to determine the acceleration to which this minimum recorded amplitude corresponds. For the shock of August 5, 1930, mentioned above, the period of the maximum waves recorded at Pasadena was about 0.6 of a second. With the standard constants of the short-period torsion instruments, calculation of the magnification for simple harmonic motion gives the result that the acceleration in milligals is 5.01 times the trace amplitude in millimeters. If we assume an amplitude of 50 millimeters, instead of the somewhat smaller amplitude recorded for this particular shock at the Pasadena station, we find an acceleration of 250 milligals. This is the value given by Cancani as the boundary between I and II of his intensity scale; as a lower limit of perceptibility this is questionable. H. O. Wood, from observations in Hawaii, placed the limit in question at 1,000 milligals.[7] Even this has been objected to as too low for ordinary use, on the assumption that it refers to the observations of skilled observers especially on the lookout.[8] Mr. Wood informs the writer that this assumption is hardly justified, as the shocks considered as perceptible in the study referred to were not observed by a scientific staff, but were those noticed by other persons employed or residing in the vicinity. The fact that a still lower value is indicated by the methods of the present paper suggests that 1,000 milligals is not too low. The value of 250 milligals may be taken as referring to the occasional instances, likely to be reported by at least a few persons in a community as large as Pasadena, of especially unstable ground or structure, or of the

[7] Harry O. Wood, "Concerning the Perceptibility of Weak Earthquakes and Their Dynamical Measurement," *Bulletin of the Seismological Society of America*, **4**, 29, 1914.

[8] R. D. Oldham, "The Depth of Origin of Earthquakes," *Quarterly Journal of the Geological Society of London*, **82**, Part I, 73 ff., 1926.

action of loose or suspended objects as seismoscopes. It should correspond, therefore, to the condition at the outer margin of perceptibility in a settled district.

This suggests a method of calculating the magnitude of a shock for which no instrumental data are available, by determining the mean radius of the area of perceptibility and assuming a recorded amplitude of 50 millimeters at that distance. However, shock intensity is often so irregular in distribution that such a procedure can give trustworthy results only for shocks which have been reported by numerous observers in all azimuths; and the data now in hand suggest that even in such a case the result may differ by half a unit from the instrumentally determined magnitude.

With the same assumption of 50 millimeters at the margin of perceptibility, the mean radius of the felt area, as calculated from Table I, is as follows:

Magnitude	3.0	3.5	4.0	4.5	5.0	5.5	6.0	6.5	7.0	7.5
Radius of II (km)	<25	27	35	65	150	250	360	530	770	1,060

The last three values are extrapolated beyond the range of Table I, using the result that the trace amplitude varies with the inverse cube of the distance. It is almost certain that these three values are too large. Since the period of the maximum waves increases notably at the larger distances, the perceptibility to persons falls off; for this perceptibility is not simply a function of the acceleration. On the other hand, the torsion seismometers, with their short periods, function well as accelerometers for the longer-period waves, so that their registration will be more nearly in accord with Table I and its extension by the inverse cube law. Thus, with a shock of magnitude 7.5 the perceptibility to persons will probably not extend to 1,060 kilometers, though the shock may be brought to notice by the behavior of loose or suspended objects; while in all probability the torsion instruments will register trace amplitudes of the order of 50 millimeters.

The lower limit of Cancani's grade IV is at 1,000 milligals. Oldham (*loc. cit.*) indicates 600 milligals as the lower limit of IV on the Rossi-Forel or Mercalli scale. With a period of 0.6 of a second, these limits would correspond to registered amplitudes of 200 or 120 millimeters on the torsion seismometer. This is not an unreasonable result, and agrees roughly with the distances at which IV is observed; but data on this point are quite too scanty for any sound conclusions. The matter can only be handled statistically when a large mass of observations has accumulated. The higher grades of intensity are obviously so very much affected by ground and other factors that their irregular distribution offers few points for the application of the methods of the present study.

Thus far we have been discussing chiefly the shocks studied by Gutenberg, since they offer unusually good data on magnitudes. The next procedure is the investigation of the magnitudes of other earthquakes, with a view to presenting a more complete picture of the behavior of the different types. We have considered shocks only from magnitude 3.5 to 5. Instances of shocks of smaller magnitude are required, partly as a demonstration that such shocks exist, and that Table I does not result in lumping all observations together.

The data for a few selected shocks appear in the following table:

TABLE VII

DISTANCES

	Date	P	MW	R	SB	LJ	T	H
(1)	1934, Jan. 16	44	36	30	190	144	—	226
	1932, Mar. 30							
(2)	21:00	240	230	240	—	—	97	13
(3)	21:19	—	—	—	—	—	97	13
(4)	21:33	—	—	—	—	—	—	13

AMPLITUDES

	P	MW	R	SB	LJ	T	H
(1)	1.5	10	9	0.2	0.4	—	0.1
	1.5	8	11	0.2	0.5	—	0.1
(2)	0.2	0.3	(0)	—	—	1.1	37
	0.2	0.1	(0)	—	—	1.1	23
(3)	—	—	—	—	—	0.05	1.9
	—	—	—	—	—	0	1.1
(4)	—	—	—	—	—	—	0.1
	—	—	—	—	—	—	0.1

CALCULATED MAGNITUDES

	P	MW	R	SB	LJ	T	H	Mean
(1)	2.72	3.33	3.06	2.77	2.88	—	2.68	2.92
	2.72	3.23	3.17	2.77	2.98	—	2.68	
(2)	3.04	3.18	<2.74	—	—	3.03	2.8?	2.89
	3.04	2.70	<2.74	—	—	3.03	2.6?	
(3)	—	—	—	—	—	1.69	1.5?	1.50
	—	—	—	—	—	—	1.3?	
(4)	—	—	—	—	—	—	0.2?	0.2
	—	—	—	—	—	—	0.2?	

The shock of January 16, 1934, is a good recent instance of magnitude 3 (slightly under). The difference between the extreme computed magnitudes is 0.65, and the mean value of 2.9 is undoubtedly well representative of the shock. Only two reports of this shock as felt are at hand: one, at a point only a few kilometers distant from the epicenter, indicates intensity low in grade IV, while the other indicates that the shock was barely perceptible to a sensitive observer in the environs of Pasadena. Thus this shock passed with very little notice, while larger shocks from the same epicenter were reported by numerous observers.

For the three shocks of March 30, 1932, no data of observed effects are available. They are given to illustrate the existence of small magnitudes. The magnitudes for Haiwee are questioned, since the distance of 13 kilometers falls below the lower limit of Table I. They are derived from a preliminary estimate of —1.2 for the value of the logarithm in Table I at 13 kilometers. Comparison with the other stations suggests —1.4 as a better value; this change has not been made, as data of this kind are too scanty to justify any positive conclusion.

Direct comparison of the amplitudes at Haiwee of the three shocks makes it evident that the magnitudes of the two latter must be close to 1.7 and 0.4 if that of the first is given the suggested value of 3.0.

A similar comparison of a group of small shocks originating near Pasadena in 1933 gave magnitudes down to 0.0.[9]

To round off the discussion of the effects of shocks of various magnitudes, there now follows a series of magnitude determinations for those larger earthquakes, not already considered, for which instrumental data are available.

Prior to 1930 the maxima of important shocks were usually photographically underexposed, so that magnitudes can only be derived by an indirect process. In that year a relay was installed at Pasadena which increases the brilliancy of the recording lamp during the larger shocks, so that the data of recent years are much more reliable.

1. The Whittier-Norwalk earthquake of July 8, 1929, is of interest as the subject of a special study.[10] At the four stations for which useful records are available the maximum amplitudes are off-scale and underexposed; but on each seismogram the amplitudes of certain smaller parts of the main shock record have been compared with those of certain aftershocks at corresponding intervals following P; then, if it be assumed that the maximum of the main shock is in the same ratio to the maximum

---

[9] See footnote 4 on page 14.

[10] Harry O. Wood and Charles F. Richter, "Recent Earthquakes near Whittier, California," *Bulletin of the Seismological Society of America,* 21, 183, 1931.

of the smaller shock, the former quantity can be estimated. Such determinations are uncertain; but the results are as follows:

	P	MW	R	LJ
Estimated amplitude in millimeters	500	300	100	20
Distance in kilometers	29	35	62	137
Computed magnitude	4.7	4.8	4.8	4.5

These data clearly indicate magnitude 4.7. The distribution of outer isoseismals in this shock was fairly regular, and the following data can be given:

Mean outer limit of II	60 kilometers
Mean outer limit of IV	40 kilometers
Mean outer limit of V	11 kilometers

The amplitudes computed from Table I are then 80 millimeters, 200 millimeters, and 4,000 millimeters.

2 and 3. Clear cases are provided by the Imperial Valley earthquakes of February 25 and March 1, 1930. The magnitudes are consistently given by the instrumental amplitudes, the former shock as 5.0 and the latter as 4.5. The radii of the areas of perceptibility were about 140 and 90 kilometers, respectively, which yield calculated instrumental amplitudes of 60 and 35 millimeters. For the earlier shock the outer limit of IV can be given as about 60 kilometers, for which the computed trace amplitude is 170 millimeters.

4. In a sense the most important shock for the present study is the major Nevada earthquake of December 20, 1932. The estimate of magnitude is quite difficult; the lamp relays functioned, but the amplitudes are off-scale at all stations. At the most distant station, La Jolla, the seismograms plainly show that the maximum trace amplitude exceeded 80 millimeters; this in spite of the fact that the station (distant 670 kilometers) lies outside the reported area of perceptible shaking, which confirms what was said above with regard to the decrease of perceptibility at larger distances, presumably connected with increasing period. Extrapolating Table I by the inverse cube rule, we find that an amplitude of 80 millimeters at 670 kilometers gives a magnitude of 7.0. That the actual amplitude was significantly larger follows from the fact that 30 seconds after the beginning the amplitude of the main shock is 26 millimeters, while at the corresponding time the amplitude of an aftershock with maximum 6.5 millimeters is only 1.1 millimeters. The proportion indicates 156 millimeters for the main shock, and this yields a magnitude of 7.3.

At Pasadena, distant 525 kilometers, a similar proportion with the aftershocks gives 550 millimeters (in the N component). A further conclusion is possible from the record of the strong-motion instrument (E component) which had a maximum of 77 millimeters; for on January 30, 1934, occurred another strong shock in the same region (see below), and on that occasion that maximum amplitude on the E component standard torsion seismometer was 54 millimeters, while that of the strong-motion instrument was 2.6 millimeters. This proportion gives 1,600 millimeters for the amplitude, on the standard instrument, of the main shock. The two indications, while differing widely, thus suggest amplitudes of the order of 1,000 millimeters, which at 525 kilometers would correspond to magnitude 7.8.

From these data emerges the conclusion that the magnitude of the shock in question must have been about 7.5. This seems rather large, but it is not likely that better data would appreciably reduce the figure. We thus have an estimate of the magnitude on our scale of a shock which, although properly considered a major earthquake, is one of the lesser members of that class.

5. Another important question is that as to the magnitude of the Long Beach earthquake of March 10, 1933. The difficulties are similar to those for the Nevada shock; the best data are obtained by comparison with the largest aftershock, which occurred five hours later, and had nearly the same epicenter as the main shock. This aftershock is assigned magnitude 5.5. Comparison of the preliminary portions of the main-shock records with those of this aftershock, on the Tinemaha and Haiwee seismograms, indicates an amplitude ratio of about 6 to 1. The amplitudes on the strong-motion instrument at Pasadena are in the ratio of about 25 to 1. However, it is very probable that the long-period waves, which constitute the maximum on the strong-motion seismogram, are proportionately larger in the main shock than in the aftershock. If we assume the ratio 6 to 1 as nearer correct, the magnitude of the main shock is about 6.2; it may exceed this somewhat, but can hardly be less.

In this connection it is worth repeating that comparison with the Santa Barbara earthquake of 1925 shows so close a correspondence in area of perceptibility, distribution of effects (allowing for differing character of ground and structures), and seismographic registration that the two disturbances are to be considered of very nearly the same magnitude. If one of these is about 6.2, so is the other.

The macroseismic effects of the Long Beach earthquake are of considerable importance, as affording a typical example of magnitude 6. The mean radius of the area of perceptibility was about 300 kilometers. If an amplitude of 50 millimeters be assumed at this distance, the calculated magnitude is 5.7, which is almost certainly too small; this again confirms the observation that perceptibility in the marginal areas of larger shocks falls off more rapidly than the seismographic amplitude. Thus, at Tinemaha, distant over 390 kilometers, the maximum recorded amplitude was over 80 millimeters, which would indicate a magnitude in excess of 6.3. There exists a report from a distance of about 470 kilometers; and in this case it is specifically stated that no shaking was felt, but that a chandelier was observed to sway.

The mean outer radius of IV was about 250 kilometers. Much interest attaches to the boundary of the region affected by intensity VII, as this is the outer limit of damage. The apparent limit coincides at many points with the margin of the alluvial Los Angeles Basin; but by considering the effect of ground carefully, and making allowance for the extension of activity along the line of the Inglewood fault, an estimate of 25 kilometers for this important radius is arrived at.

The strong aftershock of October 2, 1933, deserves special discussion. The epicenter is very well determined, at 33° 47′ N., 118° 08′ W. (near Signal Hill), and the amplitudes, especially at the more distant stations, are legibly recorded, so that the magnitude 5.0 can be assigned with confidence. Although this shock resulted in considerable minor damage, such effects were confined to inferior structures, most of them having been weakened by the March earthquake. Nearly all such cases were also located on unstable ground; so that the outer radius of VII, as applied to ordinary structures on average ground, could hardly have exceeded a few kilometers. The mean outer radii of IV and II were about 80 and 140 kilometers, respectively.

6. On May 16, 1933, there was an earthquake which was felt sharply in the San Francisco Bay district; the maximum intensity was barely VII in the vicinity of Niles, which suggests origin on the Hayward fault or a neighboring fracture. The amplitudes at the Southern California stations, all at distances over 300 kilometers, indicate magnitude 5.1. Allowing for ground, the mean outer radii of IV and II were about 90 and 130 kilometers.

7. On June 25, 1933, an earthquake took place in west central Nevada. Although uncertainty attaches to nearly every important circumstance about this shock, the records are such that it seems highly desirable to include the case.

An epicenter has been assumed at 39° 15′ N., 119° 20′ W. This lies in the region of reported highest intensity, and is near the preliminary epicenter determined from instrumental data by the United States Coast and Geodetic Survey (39° N., 119° W.). The following are the data and results from the torsion seismometers:

	P	MW	R	LJ	T	H
Distance, kilometers	570	565	605	725	250	360
A_N, millimeters....	15	10	10	11	> 90	71
A_E, millimeters....	16	26	25	14	> 90	> 80
Magnitudes	6.1	5.9	6.0	6.2	>5.8	6.2
	6.1	6.3	6.4	6.4	>5.8	>6.2

The distances given are not accurate, having been measured roughly on a small-scale map. The calculated magnitudes are so very consistent that 6.2 can be assigned to this shock with confidence. This is identical with the magnitude assigned, with less certainty, to the Long Beach earthquake; accordingly, comparison of effects is of much interest.

Intensity VII was manifested in the epicentral region, at distances up to about 25 kilometers, just as for the Long Beach shock. As the area is sparsely settled, no similarly spectacular consequences ensued.

Excellent data are available for the outer limits of IV and II in a westerly direction from the epicenter, as these limits are within the settled area of Central California. In this direction the outer radii of IV and II are about 270 and 300 kilometers, respectively. There is evidence that these numbers are higher than the means for all azimuths, so that more representative values might be 240 and 270 kilometers.

The marked similarity between the effects of this shock and those of the Long Beach earthquake leaves little room for doubt that the magnitude of the latter is correctly determined as about 6.2.

8. The Nevada earthquake of January 30, 1934 (at 12:17 P.S.T.) recorded at most of our stations with legible maximum amplitudes. At Pasadena, Mount Wilson, Riverside, Santa Barbara, and La Jolla, the amplitudes are about 50 millimeters. These stations being distant over 500 kilometers, the magnitude is 6.5 to the nearest half-unit. This shock thus is of slightly greater magnitude than the Long Beach earthquake. It is probable that the intensity reached VIII on firm ground near the epicenter; the radius of perceptibility was about 350 kilometers.

9. The Utah earthquake of March 12, 1934, offers a possibility of extending the method to larger distances. The data are tabulated below:

	P	MW	R	SB	LJ	T	H
Distance, kilometers	980	970	970	1030	1080	720	800
A_N, millimeters	21	14	22	25	20	>98	23
A_E, millimeters	22	17	44	24	24	75	41
Magnitudes	6.9	6.8	6.9	7.1	7.0	>7.1	6.7
	7.0	6.8	7.2	7.0	7.1	7.1	7.0

As in previous tabulations, the upper row of computed magnitudes is derived from the N component, the other from the E component. The magnitudes are given only to the tenth, as they depend on extrapolating Table I by the inverse cube law. Since the distances extend from just beyond the end of Table I to a considerably larger value, the excellent agreement of the results confirms the validity of that extrapolation. However, further instances of this kind will be required before the inverse cube law can be applied with confidence.

The magnitude 7.0 for this shock is very reasonable in view of the intensity manifested, the extent of the area of perceptibility, and the judgment of various seismologists on comparison of their records. The general opinion has been that this shock falls between the Long Beach earthquake and the major Nevada shock in point of magnitude, which is precisely the relation found. The shock is a good instance of the decrease in perceptibility at large distances; the radius of perceptibility is very much less than the 770 kilometers calculated for a marginal amplitude of 50 millimeters.

Until more instrumental data on larger shocks are available, it does not seem prudent to attempt an application of the magnitude scale to shocks which occurred in years when no instrumental data of the type here used were available. In the course of time it may become possible to assign magnitudes to the larger shocks on the basis of the extent of the area of perceptibility; but at present such estimates must necessarily be so tentative that it seems inadvisable to give figures which might readily lend themselves to misinterpretation. Something can be done in the way of comparing shocks occurring in the same general region; thus all the phenomena indicate that the major earthquake in Nevada on October 2, 1915, was of somewhat higher magnitude than the Nevada shock of December 20, 1932, studied above, which has been assigned a magnitude of 7.5 with some uncertainty. It would be unwise to go on to estimate by how much the magnitude of the two shocks differs. Opinion based partly on comparison of seismograms has classified the 1915 shock as of about the magnitude of the San Francisco earthquake of 1906; it appears safe to conclude that both of those shocks exceeded magnitude 7.0, and may have been of magnitude 8 or perhaps larger. In the case of the 1906 shock, the extended motion on the fault must have required a relatively long time for its completion; the seismograms in such a case would presumably not be representative of the total energy liberated.

Another case is that of the Imperial Valley earthquake of June 22, 1915, which very obviously exceeded the shock of February 25, 1930, in the same region, and therefore is known to have been of magnitude greater than 5.0. As shocks continue to occur, and our knowledge of the seismographic and other effects in the marginal areas of perceptibility increases, we may eventually be able to decide upon magnitudes for all the important earthquakes which have occurred in this region.

It is of some interest to evaluate the order of magnitude of the energies released in shocks of various magnitudes. Jeffreys[11] has assigned to the Montana shock of 1925 an energy of the order of 10^{21} ergs. Comparison of this shock with the Nevada and Utah earthquakes discussed above suggests a magnitude somewhat in excess of 7. If this magnitude be taken as 7.5, then the smallest shocks recorded (magnitude 0.0) have energies of only 10^{6} ergs. This is not unreasonable, as such small shocks record with amplitudes less than those known to have been produced by the detonation of a few sticks of dynamite; they have been identified as earthquakes only by their identity in origin with larger shocks, by evidence of depth, and by their occurrence at irregular hours.

Returning now to the discussion of the effects of various magnitudes, it appears that shocks of magnitude 1.5 (10^{9} ergs) are the smallest definitely reported as perceptible. Such reports usually refer to aftershocks of larger earthquakes, when persons are in a specially sensitive frame of mind, or they come from observers in possession of instruments or mechanical indicators which they can use to test their impressions. The smallest shocks which are likely to be noticed in the immediate vicinity of the epicenter are of about magnitude 2.5 (10^{11} ergs). A few reports are usually received when such shocks occur in settled areas. Magnitude 3 is almost always reported; while magnitude 3.5 (10^{13} ergs) attracts general attention, is reported felt to distances of the order of 30 kilometers, and reaches intensity IV on average ground near the epicenter.

The lower limit of damaging shocks is about magnitude 4.5 (10^{15} ergs). The most serious results at this stage are broken chimneys and injured brick walls, when constructed poorly and situated on bad ground. Examples are: the Brawley earthquake (4.5) of March 1, 1930 (in this case damage was probably increased by the weakening of inferior structures in the shock of magnitude 5.0 four days earlier), and the Whittier-Norwalk shock of July 8, 1929 (4.7).

The discussion has supplied several instances of shocks of magnitude near 5.0. Their effects are closely similar, and conform to the following description: On good ground apparent intensity VII is manifested only

---

[11] H. Jeffreys, "The Earth," p. 109, 1928.

within a few kilometers of the epicenter; but on soft ground it may occur at considerably larger distances, and some instances of apparent intensity VIII may be observed. The mean radius of the outer limit of IV is about 90 kilometers, and perceptibility extends to about 130 kilometers.

Two shocks of magnitude slightly exceeding 6 have been studied. In both cases, VII is manifested on good ground to about 25 kilometers from the epicenter. It is probable that the maximum intensity on good ground would be nearly VIII; in one case no data are available, and in the other the epicentral area lay in an alluvial basin, where VIII was manifested in many places, and possibly IX in a very few instances. The outer limit of IV is near 250 kilometers, and perceptibility extends to about 300 kilometers.

One shock has been assigned magnitude 6.5. Its effects definitely exceed those of the two shocks just mentioned. The Utah earthquake of magnitude 7.0 is the largest shock to which a magnitude can be assigned with the same precision as for smaller shocks. This and the Nevada shock of 1932 clearly manifested higher intensities at all epicentral distances than any of the other shocks here studied; and it is equally evident that the Nevada shock was much the larger of the two, which supports the magnitude 7.5 assigned to it.

In view of the foregoing facts, it seems assured that earthquakes destructive over even a moderately extended area are of magnitude 6 (10^{18} ergs) and over, except in cases where very bad ground and construction are involved. The lower margin of major earthquakes, in which phenomena of faulting, etc., are to be expected to a significant extent, appears to be about magnitude 7.5 (10^{21} ergs). How far above this the magnitudes of actual earthquakes may extend is a difficult, and in one sense an unanswerable, question. Judging by the relative amplitudes of distant recorded shocks, there must be cases of at least magnitude 9, and very probably 10.

A much more definite answer can be given as to what may be termed the range of a station for various magnitudes—that is, the distances to which various shocks are recorded. This obviously has important bearing on statistical conclusions as to the number of earthquakes occurring in a given region.

On the standard torsion seismometer shocks may be recorded with maximum amplitudes as low as 0.1 millimeters; but such records are likely to be overlooked in routine and only found on careful search, besides being too small to allow of accurate reading of phase times. The lower limit of useful registration is better taken at 0.2 millimeters. This occurs at the distances given below, calculated from Table I for the magnitudes shown, extrapolated by the inverse cube rule when necessary.

Magnitude	Limiting Δ
0	<25 km.
0.5	27
1	35
1.5	66
2	155
2.5	252
3	360
3.5	530
4	770
4.5	1160
5	1695

Except for shocks at large distances, the amplitudes recorded by the Benioff short-period vertical seismometer are of the order of ten times those recorded by the short-period torsion instruments, at stations such as Pasadena where it is possible to operate the Benioff instrument with high magnification. The minimum amplitude of legible registration is also about 0.2 millimeters in this case, so that the ranges for the Benioff instrument can be found by subtracting a unit from each magnitude as given in the table above.

No point of the Southern California region which is being investigated in the program centered at Pasadena is distant more than 150 kilometers from the nearest station of the group; so that a thorough survey of the records of the torsion seismometers should at least indicate the occurrence in the region of any shock of magnitude 2 or more; that is, of any shock likely to be felt except under unusual circumstances. The Benioff instruments improve this situation still further, but not as much as might be anticipated; for at two of the outer stations, Santa Barbara and La Jolla, local disturbance is so strong that the Benioff instruments are operated with constants which make the recorded amplitudes of most shocks only two or three times those written by the torsion seismometers. The practice at present is to go over the records of the vertical instruments at Pasadena with extreme care, so that shocks of magnitude 2.5 should be detected within a radius of 500 kilometers; the instruments at the other stations then serve to prevent any overlooking of small shocks in the marginal area, and also indicate larger shocks at greater distance.

The following is a tally of shocks included in the report on local earthquakes of the region which was issued for the period from January to August, 1932, inclusive, with the addition of unpublished data for September, 1932:

	NUMBER OF SHOCKS Magnitude										
1932	0	0.5	1	1.5	2	2.5	3	3.5	4	4.5	Total
January			2	5	12	6	20	2	1	1	49
February			3	2	10	10	10	5	2		42
March		4	13	10	6	9	12	2	1		57
April		3	5	9	7	15	11	3	2		55
May			6	9	11	15	9	2			52
June		2	15	8	11	9	1	2	2		50
July		1	10	11	13	7	2	3		2	49
August	8	9	26	22	20	22	4	2	1		114
September	5	14	16	10	8	12	9	4		1	79
Totals	13	33	96	86	98	105	78	25	9	4	547

The shocks in the tally are only those whose epicenters were located in the irregular area bounded by the parallel of 32°, the meridian of 115°, the California-Nevada state line, the parallel of 38°, the meridian of 120°, the parallel of 36°, and the continental margin. The numbers might be changed slightly with differences in judgment as to the inclusion of doubtful cases.

For August the totals in the smaller magnitudes are swelled by a swarm of thirty-five small shocks near Haiwee. It will be noticed that the number of shocks falls off very rapidly for the higher magnitudes; while the decrease in numbers for the lower magnitudes is obviously due to the decrease in the range of the station. Without the station at Haiwee, which is located close to a particularly active source of small shocks, there would be notably fewer shocks in the first four columns.

For the remainder of 1932 complete statistics are not at present available. Following December 20, 1932, the records were complicated by the very numerous aftershocks of the major Nevada earthquake of that date; and these had not subsided when the Long Beach earthquake occurred on March 10, 1933. For nearly the whole of the remainder of the year 1933 the chief body of recorded shocks consisted of the aftershocks of these two disturbances. Beginning with January, 1934, it was again found possible to establish routine measurement. The tally below is not as inclusive for the small shocks as that for 1932. The Nevada group of shocks is excluded by the geographical limits; it has been thought advisable also to list separately aftershocks of the Long Beach earthquake —that is, small shocks occurring along the line of the Inglewood fault.

NUMBER OF SHOCKS

1934	Magnitude 0	0.5	1	1.5	2	2.5	3	3.5	4	4.5	Total
		Exclusive of Long Beach Aftershocks									
January			4	4	18	10	6	2	3	3	50
February			3	7	8	10	6	6	2		42
March			1		4	8	9	2		1	25
April				3	2	11	8	5	2		31
Totals			8	14	32	39	29	15	7	4	148
		Long Beach Aftershocks									
January					10	4	2				16
February				3	5	3		1			12
March			2	1	8	4		1			16
April				2	13	5	2		1		23
Totals			2	6	36	16	4	2	1		67

The three groups of data clearly show analogous conditions. In every case, the number of shocks increases rapidly in passing from the higher to intermediate magnitudes; this undoubtedly represents a real increase in frequency of occurrence. The falling off in the numbers for the lowest magnitudes is almost as certainly due to the decrease in the range of the stations. It is noteworthy that the Long Beach aftershocks definitely show an excess of shocks of magnitude 2 over those of magnitude 2.5; this is not true of the other lists, and probably arises, not from a larger proportion of small shocks, but from more positive identification and listing than for other earthquakes.

It is also evident that the number of shocks of any given magnitude does not vary widely from month to month, and that the total fluctuates still less. Judging by tabulations for periods in which the magnitude scale was not applied, this will hold in general, except when the number of shocks is increased abnormally by a series of aftershocks, or by one of the occasional shock swarms not associated with a strong earthquake.

For the region studied, the evidence is that the statistics are fairly complete for shocks of magnitude 2.5 or over, and complete for all practical purposes for magnitude 3 or over. For at least these higher magnitudes, it is possible to conclude that the release of strain takes place principally in the larger shocks. For a shock of magnitude 4.5 represents energy of the order of one thousand times that of a shock of magnitude 3; if the latter were to account for the same release of strain in a given time, such shocks would have to be of the order of one thousand times as frequent, which is far from being the case. The argument can be extended with only a little less certainty to larger and smaller magnitudes. In the region studied we quite positively do not have one thousand times as many 4.5 shocks as shocks of magnitude 6 in any given period; and the shocks of small magnitude cannot possibly have the very great frequency which similar calculations would require from them, except on the unlikely hypothesis that nearly all of them occur too far from our stations to be picked up instrumentally.

This argument does not apply quite so strongly to aftershocks; thus in the three weeks between the Long Beach earthquake of March 10, 1933, and the end of the month, there occurred more than seventy aftershocks of magnitude 4 and over, among which were at least six of magnitude 5 to 5.5. Even so, the total energy of these shocks was not a large fraction of that released in the main shock of magnitude 6 or over. Thus the conclusion is warranted that seismic energy is released principally in the larger shocks and their trains of aftershocks; while smaller shocks, occurring from time to time, do not appreciably contribute to the adjustment of regional strain, but are rather to be looked upon as minor symptoms of its existence.

Summary

1. Comparison of the maximum amplitudes recorded at different epicentral distances by the torsion seismometers of the Southern California group makes it possible to rate earthquakes in this region in terms of a magnitude scale. The magnitude assigned is characteristic of the shock as a whole; it thus differs from the intensity, which varies from point to point of the affected area.

2. The magnitude of a shock is defined as the logarithm of the calculated maximum trace amplitude, expressed in microns, with which the standard short-period torsion seismometer ($T_0 = 0.8$, $V = 2{,}800$, $h = 0.8$) would register that shock at an epicentral distance of 100 kilometers.

3. Shock magnitudes can be assigned in routine to the nearest half-unit, and in cases specially studied to the nearest tenth. The smallest shocks recorded are of magnitude 0; the smallest reported felt are of magnitude 1.5. Shocks of magnitude 3 are perceptible over an area some 20 kilometers in radius; those of magnitude 4.5 are capable of causing slight damage near the epicenter; those of magnitude 6 are destructive over a restricted area; those of magnitude 7.5 are at the lower limit of major earthquakes.

4. Representative shocks of various magnitudes are discussed. The Santa Monica Bay earthquake of August 30, 1930, is assigned magnitude 5.2; the Long Beach earthquake of March 10, 1933, 6.2; the Utah earthquake of March 12, 1934, 7.0; the Nevada earthquake of December 20, 1932, 7.5.

5. The amplitude registered by the torsion seismometer at the lower limit of perceptibility to persons is found to be about 50 millimeters. With a period of 0.6 of a second, which applies at moderate distances, this corresponds to an acceleration of 250 milligals. At larger distances the period of the maximum waves increases, and perceptibility decreases more rapidly than amplitude of registration.

6. Beyond 200 kilometers the maximum registered amplitude for a given shock appears to vary nearly as the inverse cube of the epicentral distance.

7. The maximum range to which the torsion seismometer usefully records shocks of various magnitudes is discussed; shocks of magnitude 2.5 and over should be detected within 500 kilometers of Pasadena.

8. Statistics of the frequency of shocks of various magnitudes are given. It is found that the seismic energy liberated in a given region during a given period is almost wholly accounted for by the larger shocks; the smaller shocks are not sufficiently frequent to contribute more than a small fraction of this energy. It follows that the smaller shocks do not appreciably mitigate the strains which are released in the larger earthquakes, but must be regarded as minor incidents in and symptoms of the accumulation of such strains.

CARNEGIE INSTITUTION OF WASHINGTON
SEISMOLOGICAL RESEARCH
PASADENA, CALIFORNIA
June 8, 1934