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CURRENT PROBLEMS IN RESEARCH

New Dimensions in Seismology

Earthquakes are characterized by geographical position, instant of occurrence, depth, and magnitude.

C. F. Richter

The occurrence of an earthquake is a good example of a point event in space-time. In the 1850's, when the foundations of seismology were laid, an earthquake was held to originate in a sudden and probably explosive change affecting a subterranean volume small enough to be taken as a point, for most purposes. This point was termed the *hypocenter*; the spot vertically above it on the surface of the earth was termed the *epicenter*.

On this basis, there are four quantities to be determined; two give the geographical position of the epicenter, a third is the depth of the hypocenter, and the fourth is the instant of occurrence. Various means were tried to eliminate the last two and concentrate on locating the epicenter.

When it became known that such earthquakes as took place in Japan in 1891, and in California in 1906, were accompanied by fracturing of the surface rocks along lines extending for hundreds of miles, it was thought that the ideas of hypocenter and epicenter would have to be abandoned. Instead, the newly developed methods of instrumental seismology furnished revised definitions. They also compelled seismologists to consider the time dimension, for the new definition of the hypocenter, as we shall see, was intimately connected with the idea of a specific origin time.

At first, progress was assisted by the fact that most earthquakes originate at relatively shallow depths, of the order of 15 to 30 kilometers. After 1922, discovery of earthquakes at depths as great as 700 kilometers made it necessary to work with all three spatial dimensions.

A still later development assigns to each earthquake a numerical measure called its magnitude; here, in effect, we are dealing with a fifth dimension.

Duality

An earthquake is a dual event, and consequently seismology is a dual science. On the one hand, an earthquake is a sudden subterranean occurrence. In most earthquakes, including all large ones, this occurrence is not an explosion but a shearing fracture, which may extend to the surface, there to present itself for inspection and study by field geologists. On the other hand, sudden fracturing sends out through the solid material of the earth two principal types of elastic waves, with velocities roughly in the ratio 1.7 to 1. Seismologists designate the faster waves *P* (from *primae*) and the slower, *S* (from *secundae*).

Except for their generally low frequency of oscillation (long period, great wavelength), the *P* waves are physically identical with ordinary sound waves. They are waves of alternating compression and dilatation, with particle displacements longitudinal, along a ray. Waves of the *S* type exist only in solids; they represent shearing deformation, with particle displacements transverse, at right angles to the ray. *S* waves consequently are capable of polarization.

Near the epicenter of a large earthquake these waves arrive at the surface with sufficient energy to occasion perceptible and even violent vibration. Shaking due to these elastic vibrations is the immediate cause of most earthquake damage and of many other common effects of earthquakes.

Instruments which write a continuous magnified record of such earth motion are seismographs. The first effective ones were built about 1880, but some years passed before it was realized that seismographs might be capable of recording earthquakes originating at great distances from the point of observation—teleseisms, as they are now termed. The first identified teleseismic record was written on 17 April 1889, when an instrument set up by von Rebeur-Paschwitz at Potsdam registered a disturbance due to an earthquake in Japan. This date is a milestone, as important in the history of seismology as the night when Galileo turned his first telescope on Jupiter is important in the history of astronomy.

With the discovery of teleseismology, the study of earthquakes was divided into two cooperative disciplines. Instrumental recording, and the interpretation of the resulting seismograms, called for all the normal techniques of laboratory physics, while interpretation of the phenomena of great earthquakes continued to challenge the skill of the field geologist. Teleseismology and field seismology have advanced together.

This article is concerned with a few recent developments only, and with instrumental seismology almost exclusively; at least as much space would be needed to cover the results of field work.

"Instrumental Epicenter"

Earthquakes first studied were those for which seismologists could locate an epicenter from field observations—using the center of the most violent shaken area or the vicinity of surface fractures. Epicentral distance was usually given as the angle Δ at the center of the earth

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between the radii to the epicenter and to the recording station. It developed that the times of arrival of a given type of elastic waves at the stations for any one earthquake depended primarily on Δ:

$$t = t_0 + f(\Delta)$$

where t_0 is an instant taken as that of occurrence of the earthquake, and $f(\Delta)$ is a universal function, the same for all epicenters and stations, no matter where they are located on the earth's surface.

This surprising result was confirmed with progressively greater accuracy as the sensitivity of seismographs and the precision of time-keeping improved. For normal large earthquakes, the recorded times at hundreds of stations, distributed all over the world, can be fitted to a standard time-distance table or plotted curve. The individual deviations from the standard are not usually over 2 seconds, although even for the fastest waves the elapsed times, represented by $f(\Delta)$, range up to more than 20 minutes. This, of course, depends on making the right selection of the epicenter from which the distances Δ are measured; various graphical and analytical methods were developed for the purpose and applied systematically in a huge synoptic publication, the *International Seismological Summary*.

The fact that it is possible to find such a satisfactory, instrumentally determined epicenter has two important implications. First, it shows that the earth has a high degree of spherical symmetry. Indeed, it is possible to detect and compensate for the slight flattening of the terrestrial spheroid, which amounts to only 1 part in 297. Second, there is good reason to suppose that in large earthquakes the actual process of fracturing goes on for some minutes; yet, an origin time t_0 can be fixed within 1 or 2 seconds. Further physical explanation is needed. This is found in the reasonable assumption that fracturing extends itself with a speed less than that of elastic waves through unfractured rock. This being so, the first elastic waves to arrive at a distant station must be those which originated at the point of initial rupture. The "instrumental" epicenter and hypocenter represent the position of that point; the origin time t_0 represents the instant at which rupture began.

A further factor of significance in the successful use of a single set of time-distance tables is the fact that the effect of a slight increase in depth of the hypocenter is nearly equivalent, so far as distant points are concerned, to a slightly earlier origin time. Close distinction is possible only when recorded times are available for both near and distant stations.

Highway offset horizontally by faulting, as indicated by the painted center strip, in Imperial Valley, near Meloland, California, May 1940. [Photograph by J. P. Buwalda. From *Elementary Seismology*]

Deep Earthquakes

In 1922, H. H. Turner published a discovery resulting from his work in editing the *International Seismological Summary*. Certain earthquakes, some quite large, showed recorded times differing systematically from those of the standard time-distance tables. In particular, the times of arrival at very distant stations, in the region of the antipodes of the epicenter, were early relative to the times of arrival over short epicentral distances. These time discrepancies sometimes amounted to as much as 2 minutes, and the whole phenomenon was too systematic to be explained as due to accidental errors. Moreover, careful computation showed that these data were in good agreement with the expected findings for earthquakes with hypocentral depths of as much as 600 or even 700 kilometers.

This result was not generally accepted at first and began to receive serious attention only after K. Wadati, in Tokyo, published his results, derived independently from seismograms written at numerous stations in Japan, confirming the occurrence of deep earthquakes in that region. Beginning in 1931, presentation of a series of papers by many investigators settled the question, and standard time-distance tables were set up for all hypocentral depths (h) down to 700 kilometers. We thus have:

$$t = t_0 + f(\Delta, h)$$

Meanwhile, observations for a large majority of recorded earthquakes continued to fit the established data for shallow or "normal" depth.

The exploration of this new dimensionality raised many new questions bearing on the nature and mechanism of earthquakes. Because of the great pressures at depths of several hundred kilometers, theory at first tended to regard deep earthquakes as due to some explosive process rather than to shearing fracture, which is generally supposed to be the cause of shallow earthquakes. This theory, however, is almost completely ruled out by the character of the recorded seismograms, which show a form of displacement which can originate only in shearing fracture. A further fascinating question, to which at present there is no completely satisfactory answer, is that of why earthquakes occur (in certain regions only) down to depths of about 700 kilometers and no deeper. The problem is sharpened by the fact that, although frequency of occurrence generally decreases with increasing depth, there is a fairly abrupt pinching off at the lowest levels, which vary from about 600 to about 700 kilometers in depth, according to the region. Some of the deepest known shocks were major earthquakes.

Fifth Dimension

In 1931 the newly established seismological laboratory at Pasadena was preparing to issue regular bulletins on earthquakes in southern California. Since the list would include 200 or more earthquakes a year, it was felt that some indication was required to distinguish large from small disturbances; otherwise misinterpretation might follow, and undue alarm be raised. It was hardly practicable to characterize these earthquakes on the basis of perceptibility to persons or of other gross effects. Many of the epicenters were in almost totally unpopulated mountain or desert areas, and some were off the coast. Often seismographs at several stations wrote large records of an earthquake which otherwise would have passed unnoticed.

The way out was found in a direction already taken by K. Wadati, whose work with Japanese deep shocks has been cited. For several earthquakes he plotted the recorded amplitude of ground motion against the corresponding epicentral distance Δ of the recording station. The resulting curves were crudely parallel; the larger the earthquake, the higher the curve appeared on the plot.

A similar attempt was made at Pasadena, but the range in recorded amplitudes was awkwardly large. To get a more manageable plot, the logarithms of the amplitudes were used. This improved the parallelism of the several plotted curves and suggested the drawing of a representative standard curve, parallel in general to any individual curve, which therefore would differ in level by a constant. This constant, characteristic of the individual shock, was termed its *magnitude,* lettered M, and defined in effect by

$$\log A\ (\Delta) - \log A_0\ (\Delta) = M$$

Here A_0 as a function of Δ represents the standard shock, whose amplitudes accordingly correspond to magnitude zero. A shock smaller than the standard would have a negative magnitude; to obviate this, the zero shock was arbitrarily chosen as very small, corresponding to that from an earthquake at the lowest level, recorded with high magnification at very short distances.

For convenience the work was done, not with the true amplitudes of earth displacement, but with the recorded trace amplitudes read directly in millimeters on seismograms of the Anderson-Wood torsion instruments, then in operation at all the stations in southern California. These instruments were all designed to have the same operating parameters: free period of the pendulum, 0.8 seconds; static magnification, 2800; damping, slightly less than critical. The dynamic magnification should therefore be the same for waves of the same period and should not greatly affect the determination of M. To fix the zero level, a shock recorded on the torsion seismogram with amplitude of 1 millimeter at $\Delta = 100$ kilometers was defined as having magnitude 3.

The whole procedure involves very crude assumptions, partially contrary to observation. In effect, it is supposed that two seismograms recorded at the same distance can be derived one from the other by merely multiplying all displacements by a constant. This is not so. Experience shows that the frequency distribution varies with magnitude; large and small earthquakes have appreciably different elastic-wave spectra.

The name *magnitude* was purposely chosen because of analogous use of the term in astronomy. The scale of stellar magnitudes is also logarithmic, though less simply so. In a sense, the astronomical scale is reversed: the brighter the star, the smaller its magnitude. Other familiar logarithmic scales are the decibel and the *p*H scale.

Since only approximate and local use of the magnitude scale was expected, there was little concern about the roughness of the assumptions. There was some astonishment when it proved possible to assign magnitudes to the nearest half unit over the range from 0 to at least 6; in other words, instead of crude separation into large and small, twelve definite levels of magnitude could be distinguished. Later work has extended observed magnitudes above 8 and has made it reasonable to cite them to the tenth of a unit. Magnitude thus becomes truly quantitative and establishes itself as the fifth seismological dimension.

Magnitudes for Distant Earthquakes

Success of the magnitude scale in California does not guarantee its applicability to other areas. The variation of amplitude with distance can be represented by a single function only if the hypocentral depth and the constitution and structure of the upper part of the earth's crust are nearly constant. In principle, there would be no problem in extending the scale to earthquakes in California as recorded at great distances, but those sufficiently large to write decipherable seismograms in Europe, say, are relatively rare. Moreover, the original definition ties the scale to the perform-

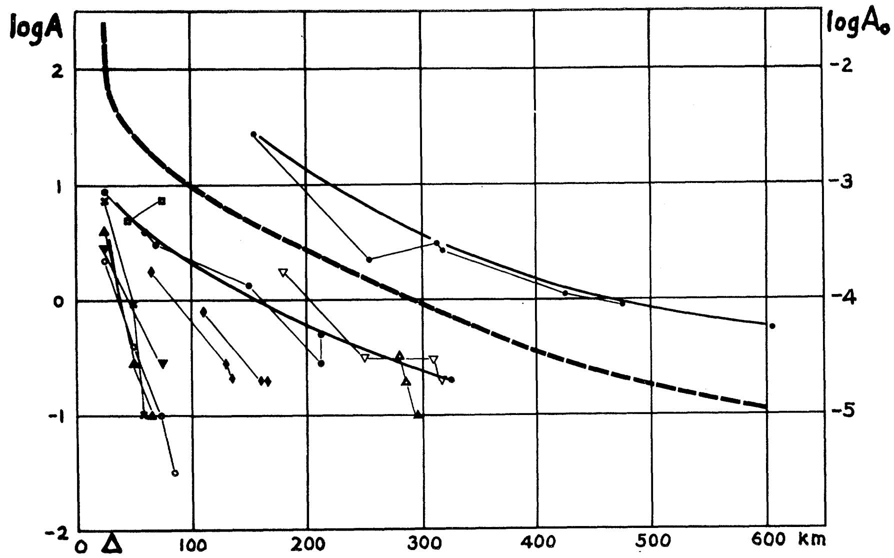
Damage due to shaking, near Meloland, California, May 1940. [Photograph by J. P. Buwalda. From *Elementary Seismology*] ance of a particular type of instrument, differing (especially in its short pendulum period) from most of those in use abroad.

Accordingly, the principle of the magnitude scale was applied to different material. Seismographs of the types most in use record the maximum of a normal shallow teleseism as a train of fairly regular waves with period near 20 seconds. These are surface waves, guided along the surface of the earth, with prevailing period determined by the average crustal structure, which acts selectively to transmit frequencies in the range near 3 cycles per minute. Many stations report the recorded maximum amplitude of earth motion in these waves, in microns, as calculated from the known parameters of the seismographs used. Accordingly, it was possible to plot the logarithms of these maxima against distance over the earth's surface and to define the magnitude in terms of a mean curve. Moreover, it is possible to calculate the expected curve theoretically; it should approximate closely the simple type

$$\log A = a + b \log \Delta$$

This proves to fit the observations well at many stations, provided that a proper choice is made for the constant characteristic of the individual station (which is an expression of the effect of local ground conditions).

The zero for M could be selected arbitrarily; it was originally set by using various means of correlation with the California local scale, with the result that a shock of magnitude $M = 7.0$ records with a maximum amplitude of 100 microns at a distance of 80° of a great circle.



Origin of the magnitude scale. Data for earthquakes in southern California, January 1932. The dashed curve represents adapted data for a typical shock. [From *Elementary Seismology*]

Magnitudes for Deep Earthquakes

Theoretically, the surface waves generated by a given earthquake should decrease exponentially with increasing hypocentral depth. Observation shows that deep earthquakes record with very small surface waves, even when the rest of the seismogram, due to elastic waves through the body of the earth, is large. Accordingly, any magnitude scale for deep shocks must be based on the amplitudes of body waves. Development of a scale of this type was accomplished by Gutenberg; he first set up such a scale for shallow earthquakes whose magnitude M had been determined from surface waves and then adjusted it to deep shocks. Because of the wide range in amplitude and period within the body-wave groups, it developed that meaningful results could be obtained only by working, not with the recorded ground amplitude A, but with the quotient of A and the period T, modifying the definition of magnitude so that

$$\log (A/T) - \log (A_0/T_0) = m$$

where m is a new variety of magnitude. The quotient A/T is a measure of the velocity of the vibrating particle within the wave and is consequently related to its kinetic energy; this is even more satisfactory physically than dealing with the amplitude alone.

For shallow shocks, the quotient A_0/T_0 was chosen to make m coincide with M near the value 7, in the middle of the best observed range of magnitudes. For deep shocks, the effect of varying depth on the amplitudes of body waves was calculated theoretically; the uncertainties of this calculation were on the whole less than those generally associated with the determination of magnitude. The only critical assumption here was that the transition is continuous; if there is any level at which there is a sudden absorption or other loss of energy, then correlation between shocks originating above and below that level is impaired. Shebalin has presented evidence for such absorption at a depth of about 60 kilometers; if substantiated, this would imply that magnitudes of deep shocks have been somewhat overestimated relative to those of shallow ones.

The magnitudes m proved internally consistent and could be assigned with as much assurance as the magnitudes M from surface waves. However, the two scales diverge; between m and M, as determined for teleseisms, a linear relation holds with considerable accuracy:

$$M = 1.59m - 3.97, \quad m = 2.5 + 0.63M$$

This gives $m = M$ near 6¾.

The two magnitude scales are thus related much as the two familiar scales of temperature are related; any choice between them is obviously a matter of convenience and of experience in use. Initially Gutenberg invariably reduced m to M, and results were published in that form. However, during work on a collaborative revision of magnitudes and of the magnitude-energy relation, undertaken with me, he abruptly reversed his position. The scale m has certain advantages; in theory, it is more simply related to earthquake energy than the scale M. Accordingly, in joint publication it was suggested that m be made the basis where energy calculations are involved—or, still better, that results be stated directly in terms of energy in ergs. To reduce the use of exponents, Shebalin has suggested employing a quantity E', which is the logarithm of the calculated energy when expressed in megajoules (units of 10^{13} ergs).

Since much work had been done and since many workers continued to determine and report magnitudes in terms of the M scale, there was objection to what appears to be an unnecessary change. With this I must agree, for the present at least. There is nothing wrong with either the m or the M scale for teleseisms; both are solidly established, and their mutual relation is well determined. However, neither can now be related definitively to the values of magnitude as determined by the original method for California local shocks. Such magnitudes necessarily are still determined and reported on the old basis. Until such time (probably not more than a few years in the future) as the local scale can either be replaced by, or used to adjust, the teleseism scales, I feel that any general use of the m scale can only lead to errors, and make it more difficult to deal with the objections raised to the magnitude scale as such, which usually are based on misinformation.

Kárník at Prague has lately made an earnest effort to base a scale for earthquakes at short distances on the quotient A/T, thus making such a scale continuous with the m scale. The test of this method will require further observations in the future, in order to establish that it does not constitute a fourth new scale, divergent from the other three.

Applications of the Magnitude Scale

It is often taken for granted that the only purpose of the magnitude scale is the calculation of earthquake energy, and this leads repeatedly to proposals to replace it by an absolutely defined scale like Shebalin's E'. This is a misunderstanding; the original purpose, both simpler and more easily fulfilled, was to combat nonsensical errors in compiling statistics of earthquakes and in studying the geography of earthquakes and earthquake risk.

A quantitative result, independent of energy calculations, is the finding of rather regular decrease in the frequency of earthquake occurrence with increasing magnitude. If we define $N(M)$ as the number of earthquakes of magnitude M or greater per unit time (usually per year) in a specified region, or even for the whole world, then to a fair approximation

$$\log N = A - BM$$

Table 1. Approximate annual frequency of earthquakes with respect to magnitude, throughout the world.

Magnitude	Number
9.0 or over	0
8.0 – 8.9	2 +
7.0 – 7.9	18
6.0 – 6.9	150
5.0 – 5.9	800
4.0 – 4.9	6,200
3.0 – 3.9	49,000

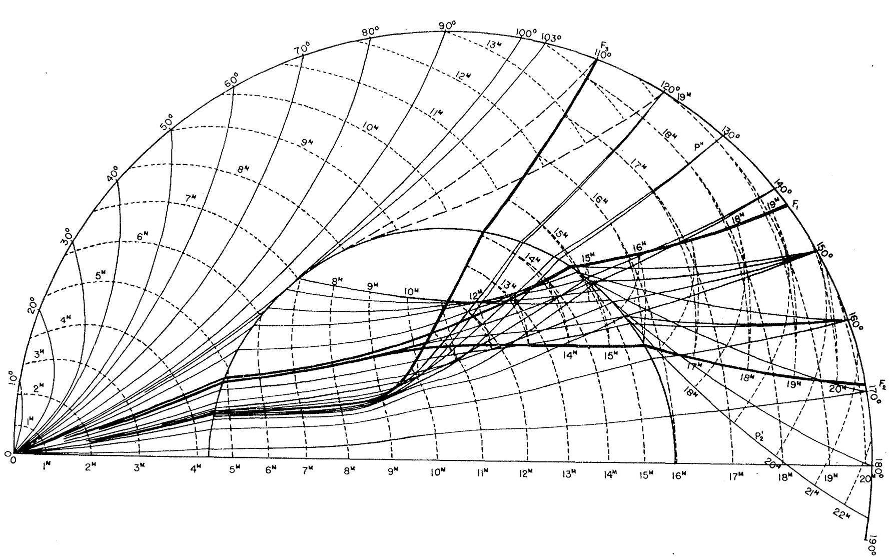
For a large area in southern California, $A = 4.8$, $B = 0.9$; these data are derived from statistics of shocks with M ranging from 2.5 to 6. For the world, we may use $A = 8$, $B = 1$. However, earthquakes of the larger magnitudes are relatively less frequent than the simple relationship would indicate. Gutenberg has lately found results equivalent to $A = 9.1$, $B = 1.1$, for magnitudes of 7 and over. Approximate annual frequencies for the world are given in Table 1. The data for lower magnitudes are extrapolated from statistics for small areas. The total number of shocks in excess of magnitude 2.5 must be well over 150,000, and if very small shocks are counted, a million a year is a reasonable estimate of the total number.

In the preinstrumental period of seismology, the supposed facts about the geographical occurrence of earthquakes were distorted, to a degree then unsuspected, by the dependence of information on population. Many early maps of seismicity showed concentration in isolated spots surrounding the oldest established centers of culture. There was astonishment when the first results of world-wide instrumental recording showed a majority of large earthquakes occurring under the sea, and many in remote regions of the continents. Even data on instrumentally determined epicenters can be misleading if no attention is paid to magnitude; thus, maps showing epicenters taken from the *International Summary* gave an utterly false impression of high seismicity in Europe. The apparent concentration of earthquake locations in that region is due merely to a concentration of recording stations.

When deep-focus earthquakes were first investigated, it appeared that even in the *International Summary* some gross errors in location had resulted from mistaking deep shocks for shallow. In order to derive reasonably acceptable geographical information, it proved necessary to revise the entire corpus of located epicenters, determine magnitudes wherever possible, and construct maps accordingly.

Eighty percent of large earthquakes, and at least as large a proportion of small ones, occur in a wide belt surrounding the Pacific, with branches extending into the East and West Indies. Most of the remainder are included in another belt extending across Asia and through the Mediterranean region of Europe. Other active belts extend through the seas; an important one follows the Mid-Atlantic Ridge.

Earthquakes are classified by hypocentral depth into shallow (down to 60 kilometers below the surface), intermediate (70 to 300 kilometers), and deep (300 to 700 kilometers). No shocks are known to occur at depths greater than about 720 kilometers. Intermediate earthquakes occur in belts that follow in general the course of the shallow earthquake zones, but with deviation or displacement of 100 or 200 kilometers horizontally. They tend to follow lines of active volcanoes. Though most of them occur around the Pacific, they are common in the trans-Asiatic belt. Under the Hindu Kush in central Asia there is the most remarkably persistent earthquake source in the world; at a depth near 220 kilometers, in a volume not over 20 kilometers in diameter,



Rays and wave fronts for longitudinal waves within the earth from a source at the left. Time in minutes. (Gutenberg and Richter, 1939.) [From *Elementary Seismology*] nearly 100 earthquakes with magnitudes from 5 to 7 have occurred in 50 years.

Until 1954, shocks at depths of more than 300 kilometers were known only from the Pacific region; but on 29 March 1954 an earthquake of magnitude 7 took place at a depth of 640 kilometers under southern Spain.

Applicability of the magnitude scale to estimation of earthquake risk is evident. Considerations of magnitude are highly important in regions, such as California and parts of the U.S.S.R., where population and industrialization are now expanding into areas previously unoccupied, some of them near known earthquake sources. Past records of shaking in the older, populated centers are here insufficient. Instrumental records, as well as geological field evidence, must be employed in estimating risk. Thus, in the California region an earthquake of magnitude between 6 and 6½ occurs on the average every year or two. Most of these have hitherto affected thinly populated districts. The few originating close to the towns have been severely damaging, like the earthquake at Santa Barbara in 1925, or locally disastrous, like the Long Beach earthquake of 1933. Every California community has to consider the possibility of such an event in its immediate area. This is the more constant and easily evaluated risk; there is also the continually impending probability of a really great earthquake, of magnitude over 8, like the "San Francisco" earthquake of 1906 (equally great shocks occurred in California in 1857 and 1872).

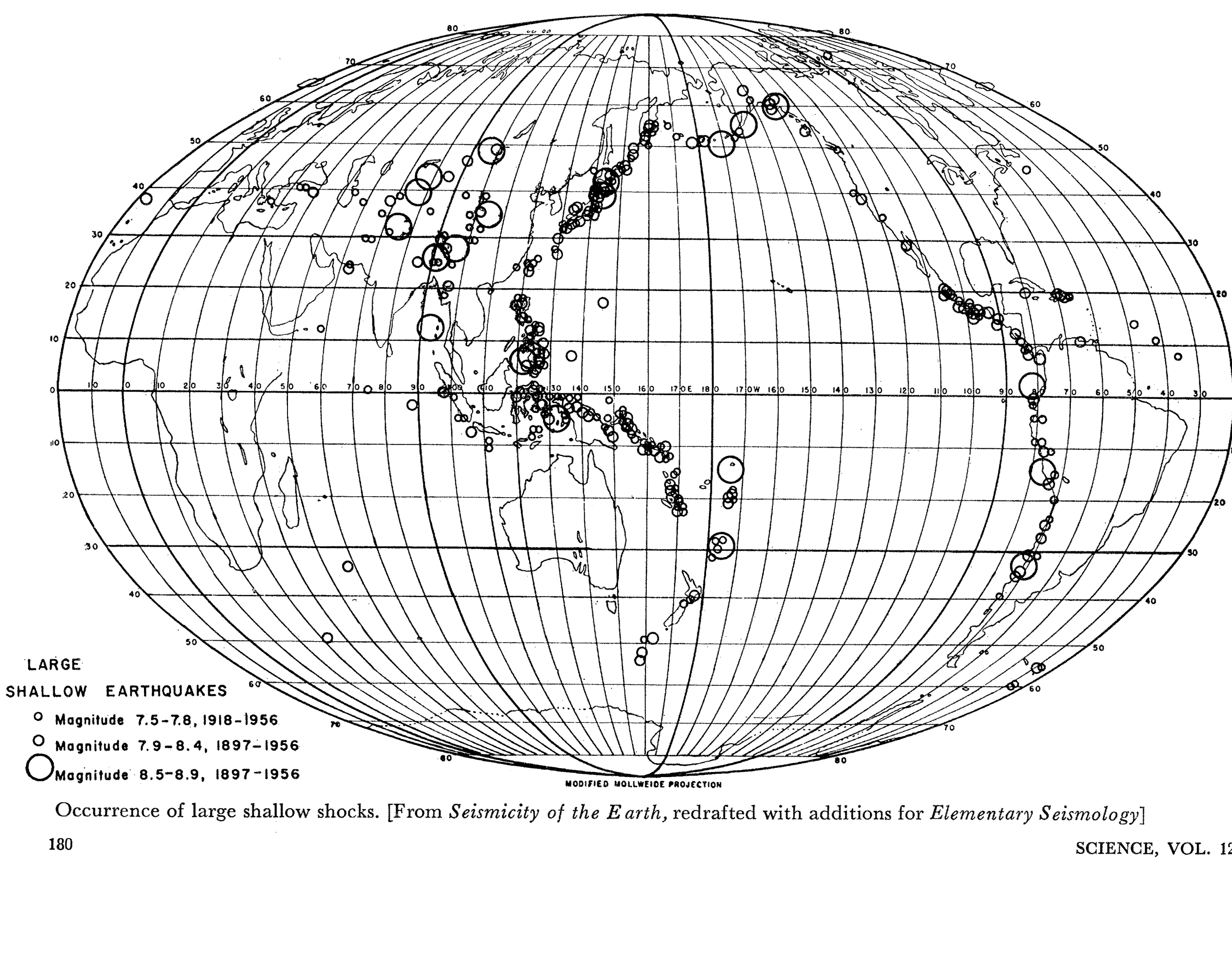
Present uncertainties in relating energy to magnitude are occasionally misapplied to make the magnitude scale appear somehow speculative. It is not speculation but cold fact that seismograms written in Europe for the 1906 earthquake were about 100 times as large as those for the 1933 Long Beach shock; it is also fact that they were over three times as large as those for the July 1952 earthquake in Kern County, while the earthquake that did damage at Bakersfield in the following month (of magnitude 5.8) was barely recorded in Europe.

Magnitude and Intensity

The magnitude scale supplements, but by no means replaces, the older intensity scales which apply to effects observed in the field. The Rossi-Forel scale of ten grades, long in use as an international standard, has now been replaced by several approximately equivalent versions of Mercalli's intensity scale, with twelve grades.

The grades of the intensity scale do not represent measurement and are not primarily defined in physical terms. They are purely descriptive and are ordinarily assigned Roman numerals to discourage treating the scale degrees as measured quantities. For example, intensity I means that the shaking was generally not perceptible to persons; II, that it was barely felt; IV, that it was strong enough to rattle windows and dishes; VII, that it was sufficient to cause general damage to weak construction; XII (the highest grade), that artificial structures were almost totally destroyed, with violence sufficient to hurl stones out of their seats and posts from their holes. The various levels are described in considerable detail.

The relationship between earthquake intensity and magnitude is one familiar in all branches of physics where radiations are observed. It is the same as that between the illumination seen on a surface and the candle power of the light source, or as that between the signal strength on a particular radio receiver and the power of the transmitting station.



Occurrence of large shallow shocks. [From *Seismicity of the Earth,* redrafted with additions for *Elementary Seismology*]

This extremely simple matter has been received with a surprising degree of misunderstanding and objection. There seems to be gross popular ignorance about the nature of an earthquake; its relation to radiation is not grasped, and there appears to be a subconscious idea on the part of many people that an earthquake shakes a whole area with equal violence. Work preliminary to formulation of the magnitude scale showed that actual ground motion drops to about 10 percent at a distance of 40 kilometers (or 25 miles) from the epicenter; the statement of this fact is often received with astonishment. Even seismologists often overlook the rapid decrease with distance and overestimate the magnitudes of large earthquakes whose epicenters are nearer to the seismological stations than is usually the case.

Engineers, accustomed to using the intensity scale, have been puzzled by the use of magnitudes. We are often asked why we use both magnitude and intensity scales, and whether use of a single scale would not be better; such a question ignores the fact that two different objectives are involved. Objections are sometimes based on the idea that magnitude represents a calculated energy; this is the reverse of the actual procedure. Complaints of "confusion" often are heard, sometimes coming from those who have been using intensity ratings as if they were magnitudes, thus really confusing their own thinking.

Magnitude and Energy

It is evident that magnitude should be at least a rough indication of the total energy radiated from the earthquake source in the form of elastic waves. Evidently this is less than the potential energy of elastic strain stored in the rock prior to the earthquake; some energy must be dissipated as heat, some must go into crushing and deformation, and some, in many earthquakes, is expended in raising crustal blocks against gravity.

If the crude assumptions implicit in the magnitude scale were valid, so that all seismograms written at a specified distance could be derived from each other by applying constant multipliers to the whole motion, then the radiated energy should vary as the square of the recorded amplitude, and we should find that

$$\log E = A + 2M$$

where E is the radiated energy and A is a constant.

Actually, the seismic spectrum changes with increasing magnitude; the larger the earthquake, the more likely it is that the largest deflection on the seismogram is due to a long-period wave carrying a disproportionately small fraction of the energy. Such a recorded wave will raise the magnitude; hence, if the energy-magnitude relation is in the general linear form

$$\log E = A + BM$$

the coefficient B should be less than 2.

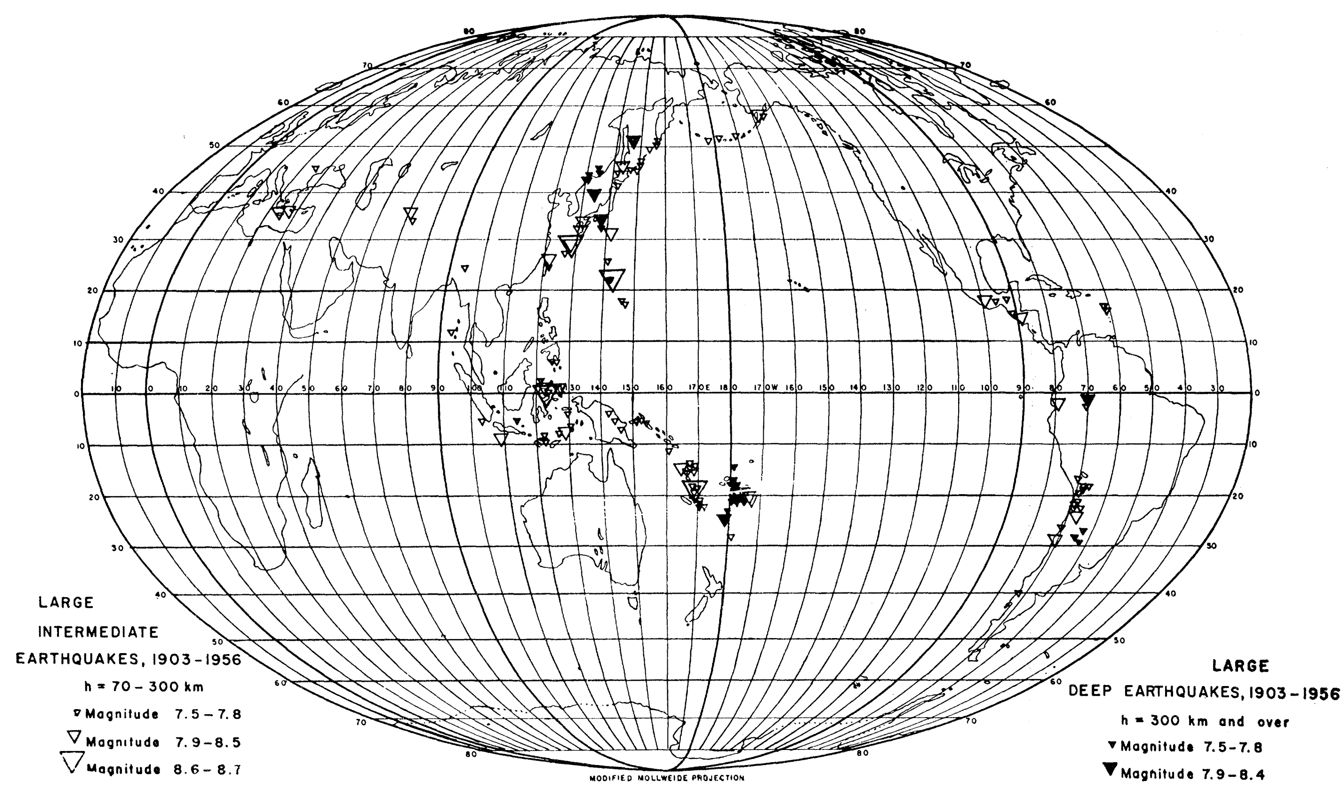
In working with teleseisms, comparison of body waves and surface waves shows that, with increasing magnitude, the surface waves of 20-second period increase relatively to the body waves, as is shown by the relation between m and M. Again the coefficient B must be less than 2. Gutenberg has set up the relation

$$\log E = 5.8 + 2.4m$$

which is equivalent to

$$\log E = 11.4 + 1.5M$$

This is an extremely rough relationship, but good enough for many approximate calculations. It shows, for example, that the energy radiated in the largest earthquakes is of the order of 10^{24} ergs. Combined with the statistical magnitude-frequency relation, it shows that nearly all the annual release of seismic energy occurs in earthquakes of magnitude 7 and over; smaller shocks are not sufficiently frequent to contribute more than an insignificant fraction.



Occurrence of large intermediate and deep shocks. [From *Seismicity of the Earth,* redrafted with additions for *Elementary Seismology*]

Magnitude, Strain Release, and Earthquake Mechanics

Hugo Benioff has applied the magnitude scale to a systematic investigation of the succession of aftershocks which follow almost every important earthquake. A great earthquake represents a readjustment of local strains, involving release of potential energy over distances of the order of a hundred miles or even more. The principal fracturing runs its course in a minute or two, but it is an obvious impossibility for large blocks of the earth's crust to readjust completely in so short a time. The readjustment continues over a period of hours, days, or months, by a process closely allied to what is known in testing laboratories as elastic afterworking. The process is not continuous, because of the frictional resistance to displacement along the original line of fracture; local strains due to afterworking must rise to a certain threshold level before the original displacements can continue. The magnitude scale makes it possible to estimate the strain released in each minor event and to plot the strain release against time. In this way curves are obtained analogous to those observed in laboratory specimens. It is possible to estimate the maximum magnitude of an expectable aftershock during the process of any given sequence, and to decide whether a large following shock is a true aftershock or represents the beginning of a new event.

Markus Båth has found that in many aftershock sequences the largest aftershock has a magnitude about 1.2 units lower than that of the principal shock. Deviations from this occur in identifiably abnormal sequences, usually characteristic of particular regions. Such sequences include earthquake "swarms" such as are common in the region of Imperial Valley, California—sequences of earthquakes, small and large, with no one outstanding principal shock. Such swarms appear to be commonest in areas of structural complexity, where the crust is broken up into blocks of comparatively small size. Earthquake swarms are common in volcanic regions, especially in association with eruptions; here the fundamental mechanism is of another sort, attributable to subterranean movements of molten rock.

It was noted that small shocks contribute relatively little to energy release. In terms of strain release, the relationship is less definite; thus, the strain released in a series of aftershocks may be comparable to that occasioned by the main shock. This is due to the fact that strain release increases with the square root of the energy. Benioff has pointed out that, on this basis, small earthquakes may operate to delay or inhibit the accumulation of a major strain and so act to postpone a large earthquake. However, once a large strain has actually accumulated, the "square" factor comes into play; a major strain can be released only in a major earthquake, or in a great number of small shocks that greatly exceed in frequency those normally observed in a given seismic region. The idea, born chiefly of wishful thinking, that small earthquakes may act as a "safety valve" is thus justifiable only in a restricted way and should not be allowed to affect deliberations in connection with precautionary measures and regulations.

Notes and Bibliography

For additional discussion and references, see C. F. Richter, *Elementary Seismology* (Freeman, San Francisco, Calif., in press).

All illustrations are reproduced here, by agreement with the publisher, from *Elementary Seismology* by Charles F. Richter. San Francisco: W. H. Freeman and Company, in press.

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On "The New Cell Theory"

Two Soviet authors critically review recent Soviet work on the origin of the cell.

L. N. Zhinkin and V. P. Mikhaĭlov

The discovery of the cell and the formulation of the cell theory were considered by F. Engels to constitute one of the most important stages in the development of natural science. Since the time of the creation of the cell theory, more than a century has passed. During that period the cell theory, constituting one of the fundamental elements in contemporary biology, has had its meaning intensified and enriched by a mass of new facts.

Quite recently the cell doctrine has been subjected to radical revision by O. B. Lepeshinskaya, who, on the basis of her own experimental findings, has built up a "new dialectic-materialist cell theory," in which it is asserted that every cell develops from living substance.

The views of O. B. Lepeshinskaya received complete approval at the conference on the problem of living substance and cell development of the Division of Biological Sciences of the Academy of Sciences of the U.S.S.R., held 22–24 May 1950. The conference carried a resolution in which it was affirmed that "the works of O. B. Lepeshinskaya and her co-workers opened a wide avenue for the investigation of acellular forms of life and the most delicate processes of cell development within and outside the organism. . . . The conceptions developed by O. B. Lepeshinskaya must be widely publicized and must be used to the utmost advantage in the practice of medicine and agriculture" (*1*). By a decision of the Presidium of the Academy of Medical Sciences of the U.S.S.R. of 14