PRINCIPLES UNDERLYING THE INTERPRETATION OF SEISMOGRAMS

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PRINCIPLES UNDERLYING THE INTERPRETATION OF SEISMOGRAMS

I. INTRODUCTION

The first printed edition of *Principles Underlying the Interpretation of Seismograms* was prompted by a steady demand for the edition first issued in mimeographed form in 1930. Many excellent textbooks have been issued on seismology since then, but few, if any, have covered the interpretation of earthquake seismograms in sufficient detail to satisfy the requirements of a seismograph station director, the student who is beginning his career as a seismologist, or the amateur seismologist. This publication is designed to fulfill these needs.

Although years of experience have shown that adequate initial interpretation of seismograms is not often achieved at stations where seismology is a subordinate activity, such stations, nevertheless, perform an invaluable service in obtaining seismographic records and transmitting preliminary data to established processing centers. There, combined with data from other stations, they are used to furnish authentic information on the locations of all important earthquakes. Not only is the scientific world and the public thus furnished immediate information on current earthquakes, but the program provides basic data from which correct interpretations can be made by station directors who furnish the preliminary data. *Principles Underlying the Interpretation of Seismograms* is designed primarily for seismologists working in this level of competence.

In April 1963, a Seismological Data Analysis Center was dedicated by the Coast and Geodetic Survey. At this Center, seismograms from the World-Wide Network of Standardized Seismographs (which consists of over 100 stations located throughout the world) are received, microfilmed, and filed. These seismogram copies, on 35 mm. and 70 mm. film and on photographic paper of original record size, are available at a nominal fee to anyone working in the field of seismology.

The student can become proficient in analyzing seismograms only by first absorbing a certain amount of fundamental background information pertaining to the origin and transmission of seismic waves and the recording characteristics of the principal types of seismographs now used in the long-range detection of earthquakes. This necessarily includes the fundamentals of earth structure since seismic wave propagation is dependent on the nature of the transmitting medium. In following this broad plan only fundamentals are covered and an effort is made to draw a distinction between routine interpretation and research.

For much of the text on the interpretation of short-period records and for many helpful suggestions in the preparation of this publication, the author is indebted to the geophysicists of the Seismology Division, Coast and Geodetic Survey, who have annually supervised the interpretation of thousands of seismograms.

II. SEISMIC WAVES AND EARTH STRUCTURE

NATURE OF SEISMIC WAVES

The great majority of the world's strong earthquakes occur in basement rock roughly within 25 miles of the earth's surface, but some have definitely originated as deep as 450 miles. If one adds to these stronger shocks the thousands of disturbances that represent minor adjustments at all depth levels, the grand total exceeds 1,000,000 shocks annually strong enough to be registered on seismographs in or near the various seismic zones. Using geological time as a yardstick and picturing the earth as a semi-plastic globe, we can only conjecture that stresses are set up beneath the sedimentary layers by the creeping or flowing of great masses of rock in a complex kind of pattern. It is beyond the scope of this publication to speculate on the causes of such creep or flow. If the structure is sufficiently plastic to adjust itself to the changing stresses and can gradually revert to a condition of no strain, there will be no earthquakes; but if the structure is rigid enough to resist this slow deformation, the stresses will accumulate until the elastic limit of the rock is reached, and then somewhere the structure will snap. This is a simple description of the mechanics of an earthquake. In great earthquakes abrupt displacements as great as 50 feet are indicated by changes observed at the surface.

When the initial displacement takes place what happens in the surrounding structure? In the answer to this question lies the first principle of seismic wave propagation. While rock is not ordinarily thought of as an elastic substance it actually is from the geophysical viewpoint. Consequently, when an abrupt displacement of great rock masses occurs, the earth reacts as an elastic solid and seismic waves are propagated to all parts of the earth following paths through the body of the earth itself and around its surface. Everyone has seen how waves radiate from the point of immersion when a stone is dropped into a pond of still water. At the center the waves are high and rough but as they spread out and diminish in amplitude they become smooth, regular, and elongated. This is much like the radiation of seismic waves.

In the case of a water disturbance there is a single train of waves, all apparently of the same general type. In the study of seismology one must deal with three or four elementary wave types, each of which has its own characteristics in regard to velocity and type of motion. This may be considered a second fundamental principle of seismic wave propagation.

The three characteristics of seismic waves that deserve the special attention of seismologists are: (1) the velocity of the wave, (2) the motion of the earth particle, and (3) the appearance of the various wave types on seismographic records. The velocities, or their equivalent travel times, are important because they form the basis of seismological tables from which the distance between focus and station is determined. The motion of the earth particle is important because it enables one to compute the azimuth of the epicenter from the station, and also furnishes a means of identifying wave types. The appearance of the various wave types on the seismogram—their period and amplitude—is of primary importance to the practical seismologist because with experience, he can recognize the more important types or limit his interpretation to possibly two choices if they are not well defined. The correct determination of epicentral distance from a single station depends on the correct interpretation of wave types.

In seismology the various wave groups representing different wave types, or wave groups of the same type arriving at a station over different paths, are commonly called "phases." In fact, any new type of wave activity, whether it can be identified or not, is by custom called a phase. The standard interpretation of a seismogram involves the measurement of the time of arrival, or onset, of the major wave groups or phases and their identification by (1) wave type, and (2) path through the earth, the latter being indicated by standard symbols.

EFFECT OF EARTH STRUCTURE ON SEISMIC WAVES

The elementary wave types registered on a seismograph cannot be discussed intelligently without first presenting a simple picture of the earth structure through which the waves are propagated. The earth is an elastic sphere consisting of three principal parts as follows: (1) An outer crust of somewhat heterogeneous rock having thicknesses varying from about 30 to 40 km. beneath the continents (with greater thickness in some mountain chains) to nearly zero beneath parts of the Atlantic and Pacific Ocean basins. In this crustal structure seismic wave speeds vary accordingly to the kind of rock traversed, but the compressional or "sound wave" velocity is usually less than 6.5 km/sec. (2) The rock mantle—a thick massive shell of basaltic rock—lying between the crust and the earth's core. Its thickness is about 0.45 the earth's radius and for all practical purposes may be considered concentrically homogeneous in structure. Compressional or "sound wave" speeds increase from 8.1 km/sec. just beneath the crust (known as the Mohorovicic discontinuity after its discoverer) to 13.6 km/sec. at the earth's core. (3) The core of the earth, having a radius about 0.55 that of the earth, is believed to be composed primarily of iron and nickel. Current theory suggests an outer core of liquid form and a solid inner core. The compressional wave speed is approximately 8.1 km/sec. just below the mantle-outer core boundary and increases to 11.5 km/sec. at the center.

The above structure has such a pronounced effect on the observed travel times of seismic waves as measured when they reach the surface of the earth that the subdivision of travel-time tables or charts into zones for (1) crustal waves, (2) rock mantle or normal waves, and (3) core waves, was obvious as soon as a relatively small amount of authentic instrumental data was obtained many years ago. (See travel-time chart, fig. 1.) Although seismological literature abounds with evidence of additional discontinuities and various refinements within these three principal parts of the earth's structure, their study lies within the sphere of the research student, not with the seismologist involved in routine analytical work.

A clear understanding of elementary waves requires a thorough understanding of what is meant by the travel time of a wave. The picture of how seismic waves travel from focus to station would be simple if the earth were homogeneous throughout, for then the waves would radiate from every earthquake focus like sound waves, all the

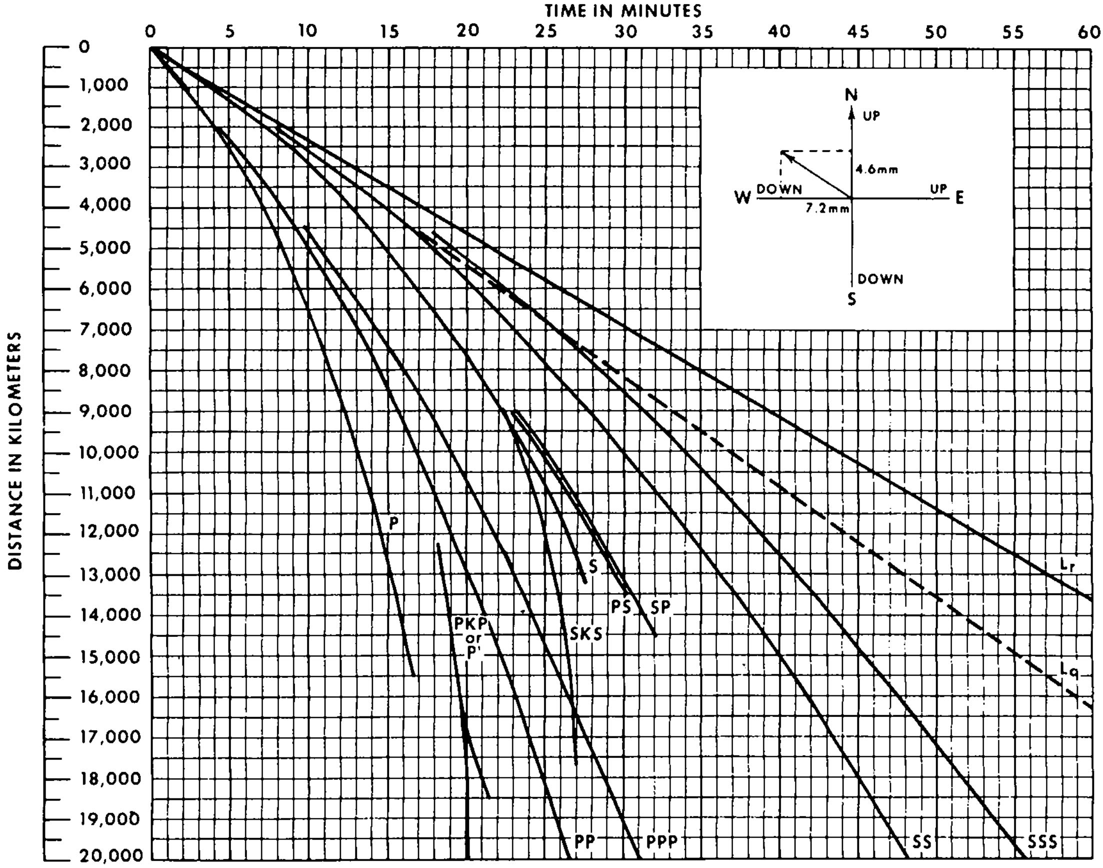

FIGURE 1.—A simplified form of travel-time chart showing approximate travel times of principal phases for shallow-focus earthquakes. Insert illustrates method of computing direction of ground motion from impulsive *iP* phase.

rays would be straight lines (practically chords) between foci and stations, and the velocities of the principal wave types inside this hypothetical earth would be constant. (See fig. 2.) But even in this simple case it should be noted that time data obtained at surface stations would not indicate a constant surface speed, that is, a linear travel time over the surface, even though the speed of the wave in the interior were constant. Near the epicenter—the point on the earth's surface above the focus—the surface speed would correspond to the interior speed but would increase with epicentral distance until it reached a very high speed at the antipode. This is obvious from the fact that the surface trace of an interior wave must traverse the semicircumference of the earth in the same time that a ray to the antipode traverses only the diameter.

These phenomena are also substantially true for bonafide seismic waves, except that seismic rays inside the earth are not chords, but are curves bent concave toward the center of the earth, as viewed from the surface, as a result of the sharp increases in seismic wave speeds with depth, except at the surface of the core. Moreover, at various discontinuities and at the surface, seismic waves undergo all the reflection, refraction, and diffraction phenomena found in optics and are subject to

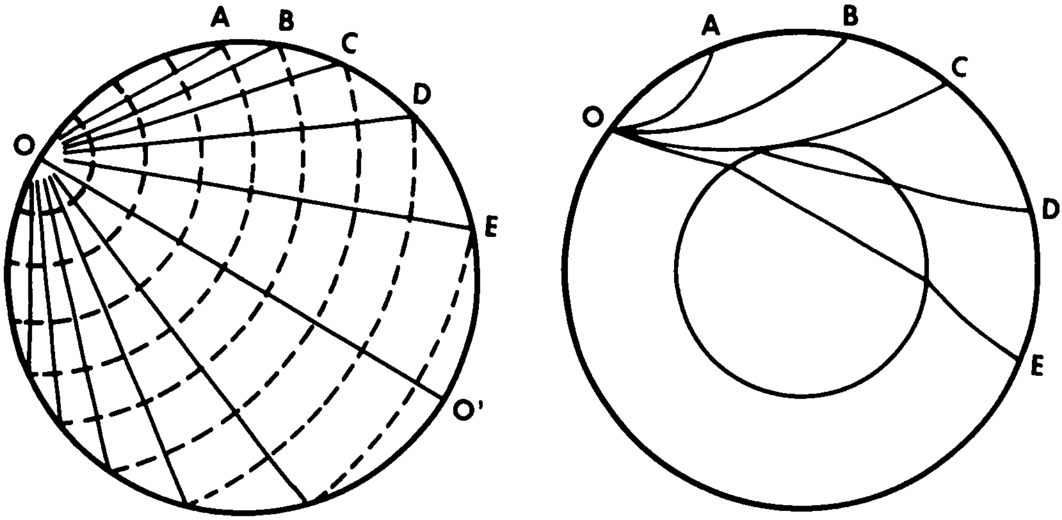

FIGURE 2.—Radiation of waves in a uniformly elastic sphere and in the earth. The diagram on the left illustrates the straight rays and spherical wave fronts that would result from a surface disturbance in an elastic, homogeneous, and isotropic sphere. The diagram on the right shows how seismic rays are bent toward the center of the earth because of increase of velocity with depth. The effect of the earth's core in deflecting the rays is also illustrated.

the same laws. These are all basic principles of seismic wave propagation. It is important to note that in the routine analysis of seismograms, one deals only with surface speeds and surface travel times. The computation of wave speeds and paths inside the earth must be considered seismological research.

ELEMENTARY WAVE TYPES

The four elementary types of seismic waves will be discussed with emphasis on actual earth motion and only secondary reference to their appearance on the seismogram. The latter will be discussed in more detail later. Two wave types, the compressional and transverse, penetrate the earth's interior. Compressional waves are propagated through either solids, liquids, or gases. Transverse waves, requiring rigidity, are transmitted only through solids. Love and Rayleigh waves are two types restricted to surface propagation only. Both require rigidity in the transmitting media. They are often called "long waves" because of their long periods and wave lengths as compared with those of the compressional and transverse waves.

With regard to the seismograph recordings of these waves and the frequent discussions of them to follow, two basic principles of seismometry may be stated at this time. It may be assumed that the period of a wave as measured on a seismogram is also the actual period of the ground wave, because the seismograph pendulum (adequately damped) will be forced to oscillate in unison with any sustained ground vibration regardless of the seismograph pendulum period. Secondly, there is usually a great difference between the relative amplitudes of waves as they are registered on a seismogram and as they actually occur in the ground. This is a complex subject which will be discussed later in the section on the response of a seismograph to earthquake motion.

For a preview of the order in which elementary wave types are registered, the reader should examine the travel-time chart, figure 1.

The compressional-rarefactional wave, P.—This is analogous to a sound wave and is sometimes called a longitudinal wave. When the compressional phase of such a wave passes a seismograph station, the ground in the immediate area is compressed and the seismograph pier moves slightly in the direction in which the wave is traveling, or away from the epicenter. Conversely, when the rarefactional part of such a wave passes a station, the ground is dilated and the pier moves toward the epicenter. These directions are registered on seismographs.

The compressional wave is the faster of the two interior wave types and is therefore designated the first preliminary tremor, *P*, since both types of interior waves are registered on the seismogram long before either of the surface waves previously mentioned. In the epicentral region *P* waves have periods of less than 1 second, but waves of approximately 1-second periods are actually transmitted over great distances—even to the antipode. The amplitudes eventually become so small that only very high-magnification seismographs with operating characteristics favorable to the registration of short-period waves will pick them up. Standard seismographs register these waves only in the epicentral region; with increasing distance they clearly register other periods as long as 5 seconds. The shorter period waves beyond the epicentral area are of very small displacement and are superposed on longer period waves of larger displacement. So-called short-period pendulum seismographs (1-second period) register the short-period part of a compressional-rarefactional wave train best; intermediate- or long-period pendulums (6- to 15- to 30-second period and up) register the longer periods best.

For both shallow- and deep-focus earthquakes the compressional waves registered on a short-period seismograph are always the largest amplitudes on the record. On a standard type of long-period seismograph they are registered in the case of shallow (normal depth) earthquakes with trace amplitudes smaller than those of any of the other types. When a very deep-focus earthquake is registered on a long-period seismograph the surface waves become so small in ground and trace amplitude that the *P* and *S* waves dominate the record. For depths between these two extremes, one finds a long-period seismograph record varying gradually from one extreme to the other.

The transverse, distortional, or shear wave, S.—This is analogous to a light wave or the transverse vibration of a string. The earth particle is always displaced in a direction normal or transverse to the direction in which the wave is traveling. Such a wave can obviously be polarized but knowledge of polarization phenomena is very limited and its investigation falls in the category of research. The directional characteristics play little or no part in the routine analysis of seismograms.

Interior transverse waves travel at about 0.6 the speed of compressional *P* waves and appear as the second most conspicuous wave group on a normal (shallow focus) record of a standard long-period seismograph. They have therefore been designated the second preliminary tremor, *S*. *S* waves follow paths through the earth closely similar to *P*-wave paths, except that no consistent and definite evidence has yet been found that S waves penetrate the core of the earth. This gives rise to the belief that part of the core is a liquid, but it is possible that other types of phenomena may account for the failure of the S wave to penetrate the core.

In terms of actual earth motion it may be assumed that the periods of S waves may be roughly double those of the P waves preceding them and that the amplitudes are also roughly doubled. On standard long-period seismograph records these estimates also apply to the actual trace amplitudes, remembering that the comparison refers to the long-period P waves and not the short-period ones. Because of the greater energy represented in the S-wave group many seismograms will show clear but weak S waves for weak shocks while the P wave will be missing entirely. On records of short-period pendulums that register the short-period P waves so well the S wave will be found rather inconspicuous or absent altogether; however, regardless of the focal depth of an earthquake a search should always be made for an S wave on the record if long-period surface waves are discernible.

Surface waves, L.—These represent by far the greatest amount of wave energy recorded in shallow- and intermediate-depth earthquakes as seen so clearly on the records of long-period seismographs. It is readily evident from the large amplitudes recorded on thousands of seismograms. In unusually deep earthquakes, however, so little energy gets into the crustal layers that they may be missing entirely from the record. This happens so seldom, however, that the analyst can safely forget about it until he obtains a record, or a series of records, that cannot be interpreted on any other basis.

There are two types of surface waves. The faster is a shear wave designated either a Love wave, Lq, after the physicist who developed the theoretical concept, or a G wave, after Gutenberg, the seismologist who discovered and discussed its presence on seismographic records. The motion of the earth particle is transverse to the direction of propagation and takes place in the horizontal plane only (fig. 3). It has no vertical component. Its wave length at long epicentral distances is a function of the thickness of the layered structure traversed and has been used by investigators to determine crustal and mantle thicknesses. Its onset at long epicentral distances is generally marked by the emergence of a low-amplitude wave of about 30-second period or more (2- and 3-minute period waves have been registered in great shocks) with successive waves of the group decreasing in period and increasing in amplitude.

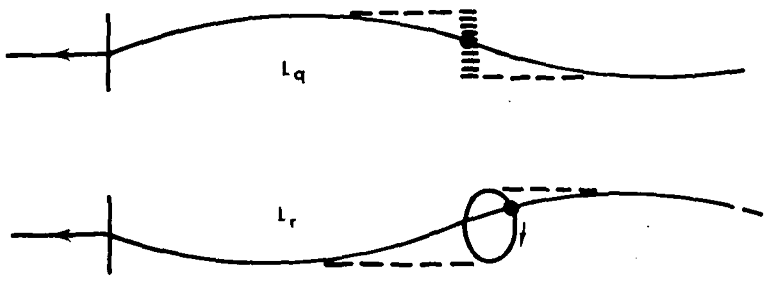

FIGURE 3.—Earth particle motions in surface waves. The upper diagram shows the horizontal earth particle motion transverse to the direction of propagation in the case of Lq waves. The lower diagram shows the retrograde character of the earth particle motion (in the vertical plane) in the case of Lr waves.

If the seismograph pendulums happen to be oriented in such a way as to register the entire wave group on one component the characteristics of the entire group are available for study; but in the great majority of cases, when the pendulum orientation is unfavorable, its continuity is disturbed by the onset of the second type of surface wave, the Rayleigh wave. The speed of the *Lq* wave, 4.5 km/sec., is similar to that of the interior shear wave, *S*, when it is propagated over short epicentral distances. There is an appreciable variation in the velocity of the surface shear wave with path traversed. At short epicentral distances it is very difficult to identify true surface waves because they are obscured by shorter period waves of great trace amplitude associated principally with the *S*-wave group. These short-period waves, however, could also be true surface waves propagated through sedimentary layers.

The Rayleigh wave, *Lr*, also named after the physicist who developed the theoretical concept, arrives a short time after the surface shear wave since its speed is about 0.92 that of the shear wave. In a Rayleigh wave the earth particle follows a retrograde elliptical orbit in a vertical plane through the direction of propagation (fig. 3). Theoretically, the longer axis of the orbit is vertical but this is not always substantiated by observation. There is no motion transverse to the direction of propagation. On most records of normal shocks registered on standard seismographs the Rayleigh wave is usually the surface wave of maximum trace amplitude. This does not necessarily mean that it represents the maximum ground displacement since that can also be associated with the surface shear wave, *Lq*. At short epicentral distances, true Rayleigh waves are obscured by *S* waves on practically all seismograph records in the same manner as *Lq* waves are obscured.

In the epicentral region the entire ground motion, including surface wave activity, may register for only a few minutes if the instrument is insensitive enough to record such strong motion, but at great distances the surface waves may be spread from 2 to even 24 hours under favorable conditions. This elongation is further emphasized by the fact that the surface waves travel over the major arc between station and earthquake as well as over the minor arc. It is not uncommon in great earthquakes to find surface waves that have made several complete circuits of the globe plus the arc between station and epicenter.

DEPENDENT WAVE TYPES

Of the four types of elementary waves just described only the interior *P* and *S* waves generate other wave types or combinations of wave types. As a wave radiates from the origin of an earthquake it encounters reflecting surfaces, not only at the surface of the earth, but also at discontinuities, or interfaces, within the earth itself such as the Mohorovicic discontinuity and the surface of the core (fig. 4). Not only are the elementary waves reflected at these surfaces, and at possibly others not too well established, but they may be transposed into waves of opposite type; thus part of the energy of a *P* wave may be transformed into an *S* wave and vice versa. When a wave strikes an interface at a certain critical angle it may be propagated horizontally in the lower media and be continuously diffracted to the surface of the earth where it will register on seismographs. It is thus possible for five different wave groups or phases to emerge when a wave strikes an interface—a reflected and refracted wave of the same type, a reflected and refracted wave of opposite type, and a diffracted wave. Their magnitude depends upon the distribution of energy in the newly formed waves.

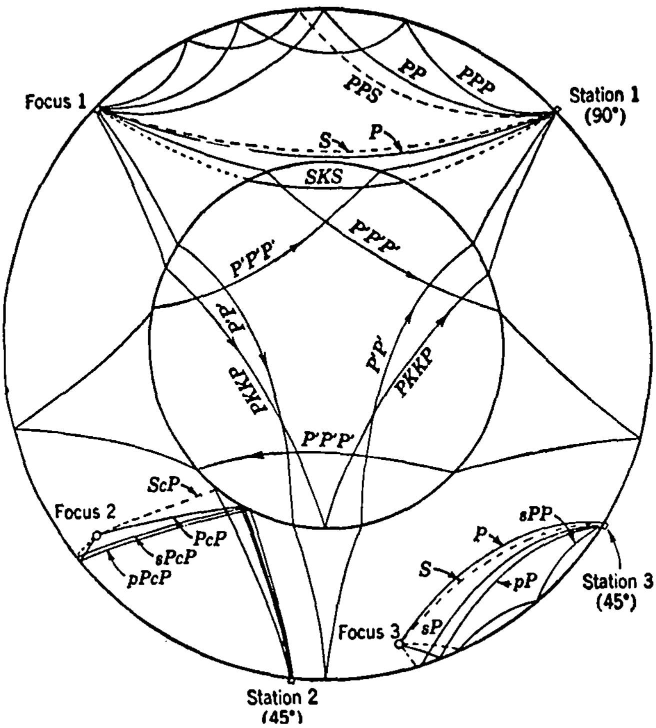

FIGURE 4.—Paths of some of the more important body waves for shallow- and deep-focus earthquakes. Three foci are indicated. (Leet, 1950).

Energy distribution is a complex phenomenon being a function of the angle of incidence of the wave on the interface and the physical constants of the media above and below the interface. Ordinarily the analyst does not take the details of this into consideration but learns by experience the particular phases that will normally be registered on a seismogram.

It is the presence of so many reflected and transposed waves that makes a seismograph record complex and difficult to analyze, but fortunately, the more important elementary wave types dominate the record as a rule. The travel times of many dependent wave types can be computed from a knowledge of the travel times of the elementary waves, making it possible to identify all major dependent wave activities that appear on a seismogram, but some recorded activities still remain unidentified as to wave type and path.

The following paragraphs outline the more important dependent wave types as they appear in four roughly defined distance zones: (1) 0 to 1,000 km., (2) 1,000 to 9,000 km., (3) 9,000 to 15,000 km., and (4) 15,000 to 20,000 km., the last figure being the semicircumference of the earth. In these zones the chart showing the travel times of the various wave types or phases takes on a different character because of the effect of the various reflecting surfaces in generating new phases or eliminating old ones as the epicentral distance increases. This will be evident from an inspection of the travel-time chart, figure 1. Special phases resulting from deep foci will be discussed later without regard to these four zones.

Wave types in distance zone 0 to 1,000 km.—The character of the surface layers plays a dominant part at short epicentral distances. Since the thickness of the surface layers and the surface velocities vary in different parts of the world, it is not possible to develop a single travel-time chart that will fulfill all conditions. For this reason when an intensive seismographic study of a local area is contemplated, the determination of local velocities from observational data is always a prerequisite. In continental areas the first *P* wave recorded at short distances is a wave traveling entirely in the slow-speed layers of the crust, following practically a direct linear course from focus to station. Other *P*-wave rays dip downward toward the high speed, 8.2 km/sec. zone, penetrating the Mohorovicic interface. Some of the energy is transmitted along the high velocity (lower) side of this interface and generates a diffracted wave having a surface speed of about 8.1 km/sec. Other rays dip deeper into the high velocity rocks and become the normal *P* waves recorded at distant stations (fig. 5). The result of this is that up to about 120 km. (for 18 km. focal depths), the first wave to arrive travels at velocities ranging from 6.0 to 6.5 km/sec. Beyond that, up to approximately 1,000 km., a surface velocity of about 8.2 km/sec. represents a combination of *P* waves diffracted from the bottom side of the Mohorovicic discontinuity, and normal *P* waves traveling through the upper portion of the rock mantle where the velocity does not differ materially from that of the diffracted wave.

S waves generally duplicate the *P*-wave phenomena except at a slower speed, the velocity of the *S* wave being roughly 0.6 times that of *P*. At short epicentral distances one cannot distinguish between *S* waves and *Lq* waves since they have nearly the same velocity. Interior (crustal) *S*, *Lq*, and *Lr* waves are all crowded so closely together at short epicentral distances that they generally defy separation. Toward the end of the 1,000 km. zone, however, they separate enough to discern the long periods that identify true surface waves.

In this zone it is standard practice to read only the first *P* and *S* waves and the phases following closely thereafter to distinguish, if possible, between crustal and normal waves and other phases that may be significant. Surface waves are reported when they can be distinguished as indicated by their long periods and travel times. While waves reflected from interfaces may be prominent on records obtained at short epicentral distances, they are usually of value only where local station networks are in operation on special projects. They are generally associated with the records of short-period pendulums.

While 1,000 km. is given as the limit of this first distance zone, no sharp line of demarcation actually exists between any of the four zones described here. The transition in the wave-type and travel-time characteristics from one zone to another is very gradual, with the border-line areas showing some of the characteristics of both adjoining zones.

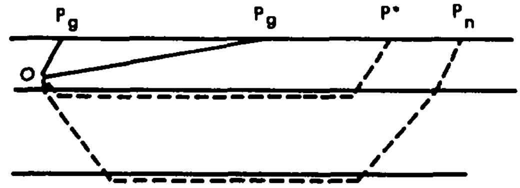

FIGURE 5.—Paths of principal local shock waves in crustal layers according to Jeffreys.

Wave types in second distance zone, 1,000 to 9,000 km.—Here is found the ideal condition for recording both elementary and dependent wave types. In addition to the elementary *P*, *S*, and *L* waves, there are often recorded very clear *P* and *S* waves that have been reflected one or two times from the surface of the earth. When a reflected wave, *PP*, is recorded it means that an elementary *P* wave has hit the surface of the earth at a point halfway between focus and station and has continued from there to the station as a reflected wave. It is obvious that the time it takes a once-reflected wave to travel from a shallow focus to a station is double the time it takes an unreflected direct wave to travel half the distance. The student can check this on travel-time charts. Among the less prominent phases registered in this zone are *P* and *S* waves reflected from the core of the earth, *PcP* and *ScS*; also *P′P′*, transmitted through the core and reflected from the earth's surface through the core a second time. *P′P′* is well registered only on short-period, high-magnification seismographs.

An important wave transformation is also clearly evident in this zone. When elementary *P* and *S* waves strike the surface of the earth they are partly transformed into waves of opposite type, but the angles of incidence and reflection are not equal because of the difference in the velocities of the incident *P* wave and the reflected *S* wave, and vice versa. This is in accord with the theory of optics. It thus happens that while the travel times of a *PS* combination and an *SP* combination are approximately equal for a given epicentral distance, there are two different points of reflection. This phase is registered shortly after the arrival of the elementary *S* wave and is generally clearly defined at the greater distances in this zone. It becomes increasingly important at the greater distances.

Surface waves in this zone and the two following zones show considerable variations in speed due to variations in the crustal layering. Sometimes the records show evidence of more than one *Lq* or *Lr* phase, indicating the transmission of energy in two or more layers having distinctive velocities. Outstanding cases of this kind are reported in routine analyses but study of these phenomena on a large scale is considered research. There is no fixed ratio in the relative amount of *Lq* and *Lr* energy to expect; it is probably a function of the direction of the focal displacement that generates them.

Wave types recorded in the third distance zone, 9,000 to 16,000 km.—This is of special significance in featuring the effect of the earth's core on the transmission of the interior *P* and *S* waves. When *P* and *S* waves are recorded at epicentral distances greater than 11,500 km., their rays graze the surface of the core (fig. 4). The *P* wave is subsequently sharply refracted into the core because of the sudden decrease in velocity. After penetrating the core it emerges and, on experiencing a reversal of the refraction phenomenon at the surface of the core, continues on to the earth's surface. An inspection of figure 1 shows that the travel-time curve of this *P′* phase is not just an extension of the normal *P*-wave travel time, but there is an offset of several minutes. This most distinctive feature of a travel-time chart is the seismologist's evidence for the existence of a core. The normal *P* wave, between 11,500 km. and 15,700 km., is in a shadow zone, the activity from there on being attributed in part to a diffracted wave originating on the core interface when it is grazed by the last of the normal *P* rays to reach the earth's surface.

The distribution of *P′* rays penetrating the core is such that there is a concentration of energy at an epicentral distance of 15,600 km. This is revealed on all seismograph records, the amplitudes being comparable with those of *P* waves registered in the second distance zone. The reason for this is the presence of a very unusual type of phase designated P'_2. A P'_2 wave is generated by the first group of normal P rays that penetrate the core after grazing its surface. As these rays strike the core's surface at increasingly greater angles, the rays ultimately emerging at the surface of the earth do not emerge at epicentral distances that become greater and greater as they normally do, but at distances that actually become less and less. This is the P'_2 phase of core-wave phenomena. At about 15,600 km., this decreasing of epicentral distance stops and the points of emergence again increase with increasing dip of the initial P rays and the phase becomes a normal P'_1. This concentration of so many rays at approximately the same time and epicentral distance is responsible for an apparent amplification of the P' wave. While this simple explanation serves the purpose of routine interpretation, the research student finds much that still remains to be explained in the present concept of core structure and core waves. Recent postulation of an inner core of about 0.2 the earth's radius has complicated the picture and tended to confuse the symbolism. Another important core phase frequently found on sensitive short-period seismographs is $PKKP$, a P wave reflected within the core.

When elementary S waves strike the core they are partly transformed into P waves that penetrate the core and emerge at the earth's surface either in the form of P or S waves. As the normal S wave goes out of the picture it is thus replaced by SKS waves which have shorter travel times than the normal S wave would be expected to have. In this distance zone many different combinations of P and S waves are theoretically possible, as indicated on the Gutenberg travel-time charts, in the explanation of the nomenclature on page 41, and in figure 4. See notes on "second distance zone" for comment on surface waves in this zone.

Wave types recorded in fourth distance zone, 15,000 to 20,000 km.—In this zone there are no normal P and S waves since they are deflected into other paths by the core. Only reflected and core waves are registered. See note in "second distance zone" on surface wave characteristics.

Deep-focus phases recorded in all distance zones.—When earthquakes originate deep in the rock mantle beneath the crust, the earth's surface in the epicentral area serves as a nearby reflecting surface that produces well-defined waves on seismograms at distances of approximately 2,000 km. and over. While the great majority of earthquakes originate in the crustal layers or a very short distance below them, a small number occur at various levels down to 700 km. It thus becomes possible for a deep-focus P wave to be reflected, with angle of reflection equal to angle of incidence, at a point near the epicenter. The short leg of the ray from focus to reflecting point is designated p; the long leg from reflecting point to station, P. The combination, pP, represents perhaps the most important phase that is indicative of the deep-focus character of an earthquake. The phenomenon is repeated in the case of the elementary S wave. Figure 4 and the Gutenberg travel-time charts in figures 13, 14, and 15 show how the travel times of all interior waves are influenced by intermediate and deep foci.

III. RESPONSE OF SEISMOGRAPHS TO SEISMIC WAVES

All seismograms discussed in this publication are records obtained from pendulum- or inertia-type seismographs. This means that the motion of the ground is measured by observing, either directly or indirectly, the relative motion between the moving ground and the moving center of oscillation of some form of pendulum. It is the inertia reaction that makes the mass of a pendulum oscillate in a manner different from that of the ground that supports the pendulum.

OPERATING CHARACTERISTICS OF SEISMOGRAPHS

Direct recorders and electrical recorders.—Seismographs are divided into two main classes depending upon whether the motion of the pendulum (relative to the moving ground) is registered directly on the recording paper through a simple mechanical or optical lever system, or whether it is measured indirectly by generating an electrical current that depends primarily on the motion of the pendulum. In the case of the so-called "direct" recorder, the seismogram gives a uniformly magnified picture of the relative motion between pendulum and ground and consequently offers the simplest possible form of mathematical problem in computing actual ground motion from the seismograph record. The problem reduces to that of a forced vibration of a simple oscillator discussed in many advanced textbooks on physics and theoretical mechanics.

The electrical or galvanometric recorder generally utilizes a coil of wire fixed to the pendulum in such a way that, when the pendulum oscillates, the coil moves between the poles of a strong permanent magnet. The current thus generated is proportional to the velocity of the relative motion between pendulum and ground. The current is not registered as such on the seismogram because the recording galvanometer has oscillatory characteristics of its own that can control the entire character of the seismographic record. In recent years electronic circuits have been successfully employed to overcome the friction of pen-and-ink-recording seismographs while still retaining magnifications as high as, or greater than, optical recorders. The principles of electrical induction, reluctance, and variable capacity have all been employed in both types of electrical recorders. The problem of theoretically determining ground motions from such recorders, however, is so difficult that it falls within the research category. In practice, shaking tables are frequently used to calibrate them before permanent installation; thereafter, the electrical conditions are frequently checked to assure constancy of calibration.

In appraising the two types it may safely be concluded that the electrical recorder has many features that recommend it for obtaining the arrival times of seismic waves, the determination of epicentral distances, and the location of epicenters. On the other hand, the direct recorder which is less adaptable to extremely high magnifications is much better suited to the measurement of actual ground motion.

Pendulum periods.—Seismograph pendulum periods range all the way from 0.1 second, used in measuring the ground accelerations of destructive earthquake motions, to several minutes' period used to record the ultra long-period surface waves of violent distant earthquakes. The records discussed in this publication will in general be obtained from 1.0 second pendulums (short-period pendulums) and from 6- to 20-second pendulums (moderately long-period pendulums). Short-period pendulums and galvanometers are employed to bring the short-period ground waves in sharp relief on the seismogram; long-period pendulums produce better records of the long-period waves at the sacrifice of clearness in the lower amplitude, short-period waves.

Magnification of seismographs.—The lever magnification of direct-recording seismographs with optical levers is limited to about 25,000 or less. This applies only to short-period waves (1-second period or less) which are always much smaller in amplitude than the longer period waves so often associated with them on the same record. Practically all seismic records are a series of short-period, low-amplitude waves combined with longer period waves of much greater amplitude. In such a series the energy in the various period groups could be of the same order. This means that a long-period pendulum of either direct- or electrical-recording type would operate at a magnification of 500 or 1,500 under conditions paralleling the operation of a short-period pendulum magnifying 10,000 or 25,000 times. All direct recorders register with maximum magnification the ground periods that are less than about half the pendulum period as will be seen on the magnification curve in figure 8.

There is almost no limit to the magnifications obtainable with galvanometric and electronic installations. Stations at excellent recording sites can operate efficiently at magnifications of 400,000 or above. However, 50,000–100,000 is a more realistic value at a site of average quality, where ground disturbances due to wind, traffic, surf, etc., limit the gain. Most instruments of this kind are highly selective, meaning that this high magnification is effective over a ground period range of only a few tenths of a second. Most long-period galvanometric recorders operate with magnification varying from 500 to 20,000, depending largely upon the stability of the site. (See the magnification curves in figs. 8 and 9 for further details on the magnification characteristics of various electrical-recording instruments.)

Damping.—The response of a seismograph to earthquake waves is governed largely by the magnitude of the damping, especially in the period zone where resonance would occur if the pendulum were undamped. Seismograph pendulums are damped to prevent them from swinging in their own natural periods and thereby complicating the record. Figure 6 shows how the harmonic magnification factor becomes indeterminable for undamped pendulums when recording earthquake waves having periods near the pendulum period. This is the resonance zone.

The magnitude of a damping force in direct-recording seismographs may be measured in terms of (1) a damping ratio, ϵ, or (2) a fractional part of critical damping, h. (See page 18 for the mathematical relation between ϵ and h.) If a light spot is deflected 10 mm. from zero position and, on returning, overswings the zero position 1 mm., the damping ratio is 10:1. It is a measure of the overswing of the pendulum. 20:1 is considered a good operating value. Damping is critical when it is just sufficient to prevent a pendulum from overswinging the zero position. Under this condition, h equals unity. When greater than this a pendulum is overdamped, a condition to be

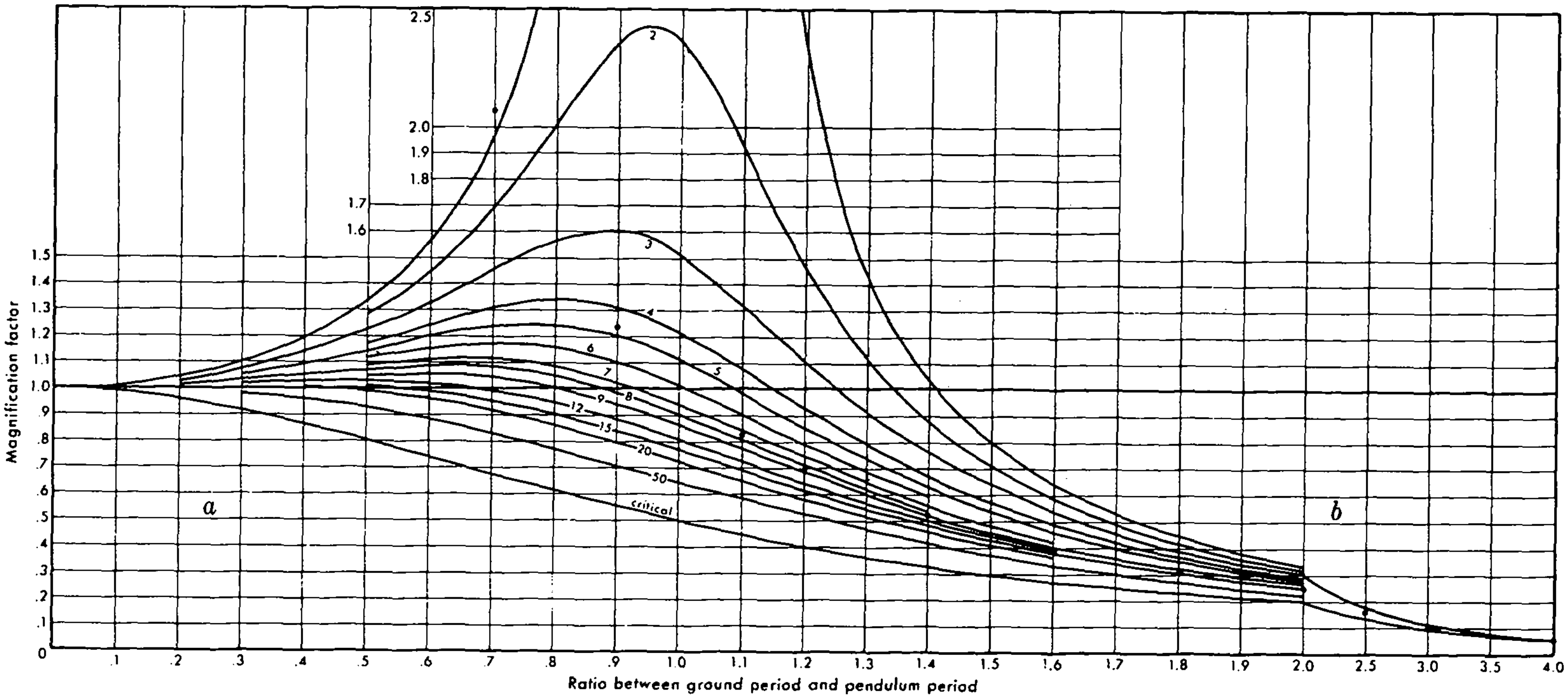

FIGURE 6.—Magnification curves for direct-recording pendulums with varying damping ratios, applicable only when the motion of the seismograph pendulum is registered through a simple lever system on the seismogram, and when the ground motion is of sustained simple harmonic character, or nearly so. This eliminates galvanometric instruments. The figures on the curves are the damping ratios of the seismograph pendulum. The damping ratio is the amplitude ratio of any two consecutive half-cycles recorded in the damping test. To calculate the magnification of the ground motion, the lever (static) magnification of the seismograph is multiplied by the magnification factor obtained from this chart. The line of dots indicates the acceleration function, or the condition to be satisfied when the amplitudes on a seismogram are directly proportional to the impressed ground acceleration.

carefully avoided in direct-recording pendulums. Underdamping is preferable to overdamping as it increases rather than decreases the trace amplitudes. A type of damping in which the damping force is proportional to velocity is ideal because it simplifies the mathematics of reducing trace amplitudes to ground motion. Damping proportional to velocity should not be confused with friction such as exists between a stylus and smoked paper. The latter is a force which is independent of velocity and results in obstructing the free motion of the pendulum for very small displacements. Friction in the pivots of various lever systems is also of this type.

In the case of galvanometric recorders, both the pendulum and galvanometer are critically damped when oscillating as independent units, but when connected for operation, the galvanometer is generally overdamped, especially in the case of long-period pendulums. In this instance greater magnification is obtained by overdamping. The period and damping characteristics of pendulums and galvanometers can be adjusted to obtain an extremely wide variety of responses.

Components.—As earthquake ground motion is a three-dimensional phenomenon, seismographs are usually mounted to record the north-south and east-west components of the true ground motion, but this is sometimes changed so that waves of longitudinal and transverse character coming from a predetermined direction will be automatically separated on the two components. P waves and surface Lq and Lr waves are particularly susceptible to this kind of treatment. A vertical component completes the three-dimensional setup. The short-period vertical component with high magnification is almost indispensable in the satisfactory recording of the onset of the P wave which is so important in epicenter location. The long-period horizonal component seismographs (½ minute and greater) are used not only for detecting S waves, but also very extensively for detecting surface waves to study crustal and mantle structure.

MOTION OF A DIRECT-RECORDING SEISMOGRAPH PENDULUM UNDER THE INFLUENCE OF STEADY STATE VIBRATIONS

The problem of computing the magnitude of the ground displacement from data on the seismogram is a study of forced vibrations. The earth particle vibrates in what may, for our present purpose, be regarded as sustained simple harmonic motion and forces the seismometer pendulum to swing in the same period but not the same phase. The mass of a seismometer may be considered as (1) remaining practically steady in space for extremely rapid earth motions, (2) lagging behind the earth particle for earth motions of shorter period than the natural pendulum period, or (3) leading the earth particle for earth motions of longer period than the natural period of the pendulum. (See fig. 7.) These are "phase differences" which can be calculated precisely from the principles of theoretical mechanics. The complete conversion of a complex seismographic record to true ground motion is in practice extremely difficult and must be considered definitely in the field of research.

Determination of ground displacement.—The amplitude on the seismogram for direct recorders depends upon the amplitude and period of the earth wave, the lever or static magnification of the seismograph, the natural period of the pendalum, and the

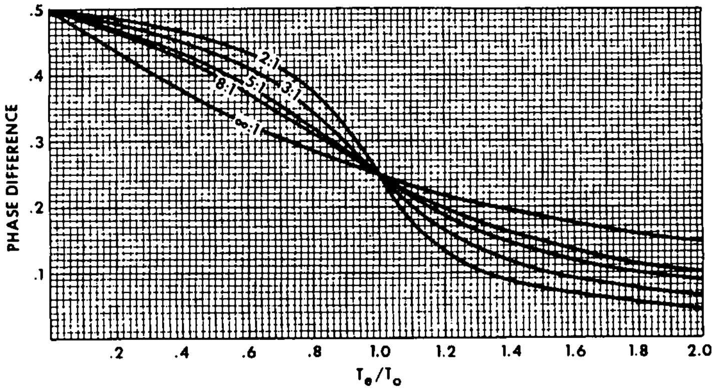

FIGURE 7.—Difference of phase for linear displacements for various values of the damping ratio and for different ratios of the periods of the disturbance and the free period of the pendulum—applicable to direct-recording seismograms. (Reid, 1910).

damping. The following equation shows the relation between trace displacement and actual ground displacement in the case of direct-recording seismographs and sustained (steady state) simple harmonic motion of the ground:

$$A_e = \frac{A_t}{V \cdot M_h}.$$

in which

A_e=amplitude (½ range) of earth movement.
A_t=trace amplitude (½ range).
V=static or lever magnification of seismograph.
M_h=harmonic magnification. See ordinates of curves in figure 6. M_h is the reciprocal of the term used for harmonic magnification by Anderson and Wood (1925, pp. 1-72), in their paper on the torsion seismometer.

The term M_h is developed from the differential equation expressing the motion of a damped oscillator when acted upon by an external force of simple harmonic form.

$$\frac{d^2x}{dt^2} + 2a\frac{dx}{dt} + b^2 = A'_e \sin ct$$

where

x=instantaneous displacement of the oscillator relative to the ground.
t=time.
$2a$=damping factor in which $a=hb$. h is the coefficient of damping.
$b=2\pi/T_o$, T_o being the natural pendulum period.
A'_e=amplitude (maximum) of external displacement due to earthquake wave.
$c=2\pi/T_e$, T_e being the period of the earth wave.

Using the expression for harmonic magnification as developed by Anderson and Wood, this equation becomes

$$A_e=\frac{At}{V}\sqrt{\left(\frac{T_e^2}{T_o^2}+1\right)^2+4\frac{T_e^2}{T_o^2}(h^2-1)}$$

in which

T_e=period of the earth wave.
T_o=natural period of the pendulum.
h=damping coefficient related to the damping ratio ϵ by

$$h=\frac{\log_{10}\epsilon}{\sqrt{1.862+(\log_{10}\epsilon)^2}}$$

where ϵ is the ratio between successive amplitudes of a damped pendulum when displaced and allowed to come to rest under the influence of damping alone.

For any type of seismograph a magnification curve usually can be constructed from theoretical considerations alone and used to convert observed trace amplitudes to actual ground displacements, generally expressed in units of 0.001 millimeter. It also can be determined through shaking table tests in the laboratory.

The direct recorder has a great advantage where it is desired to know the amplitude of the ground motion because it requires no calibration. Its magnification can be determined geometrically from the dimensions of the optical or mechanical levers attached to the pendulum, provided the center of oscillation of the system is known. If this is not known the magnification of any direct-recording pendulum having its period controlled by a spring and/or any component of gravity can be determined from the following equation:

$$V=\frac{4\pi^2S}{T_o^2}$$

where

V=the magnification of the motion of the center of oscillation of the pendulum.
S=the deflection of the seismograph trace when the pendulum is subjected to a side tilt of 3 minutes 30 seconds (equivalent to a static force of 1 cm/sec.^{2}).
T_o=the natural undamped period of the pendulum.

Some seismographs, such as the Milne-Shaw, and some sensitive vibration meters are provided with special screw devices for side-tilting the instrument and measuring the trace deflection so that S can be determined.

The direct-recording seismograph acts as a displacement meter when recording earth waves that have a short period of vibration in comparison with the natural period of the pendulum—one-half or less (fig. 6). In such cases the amplitudes on the seismogram are proportional to displacements of the ground. The seismometer acts as an accelerometer when recording waves that have periods greater than about double the natural pendulum period. In such cases amplitudes on the gram are proportional to acceleration of the ground. When earth wave periods are in the zone of the natural pendulum period, a transition stage exists so far as any fixed relation to displacement or acceleration is concerned.

The physical significance of harmonic magnification as applied to direct-recording pendulums.—The amplitudes on a seismogram depend not only on the amplitudes of the ground displacement, but also on magnification factors that vary with the period of the earth wave as just explained. This relation between the pendulum displacement and the actual ground displacement can be illustrated by a simple experiment, the elements of which are shown in the insert in figure 8. Let a heavy weight E of 5 or 10 pounds represent the oscillating earth particle at the station and let it oscillate through a small arc after suspending it at a point O. The weight E is then a point executing simple harmonic motion in a manner similar to that of a seismograph pier when earthquake waves are passing by. To the weight E attach a light plumb bob S to represent the seismograph pendulum and let it swing in a pail of water as a crude means of damping its motion. When the mass E is oscillating in its own natural period it forces the

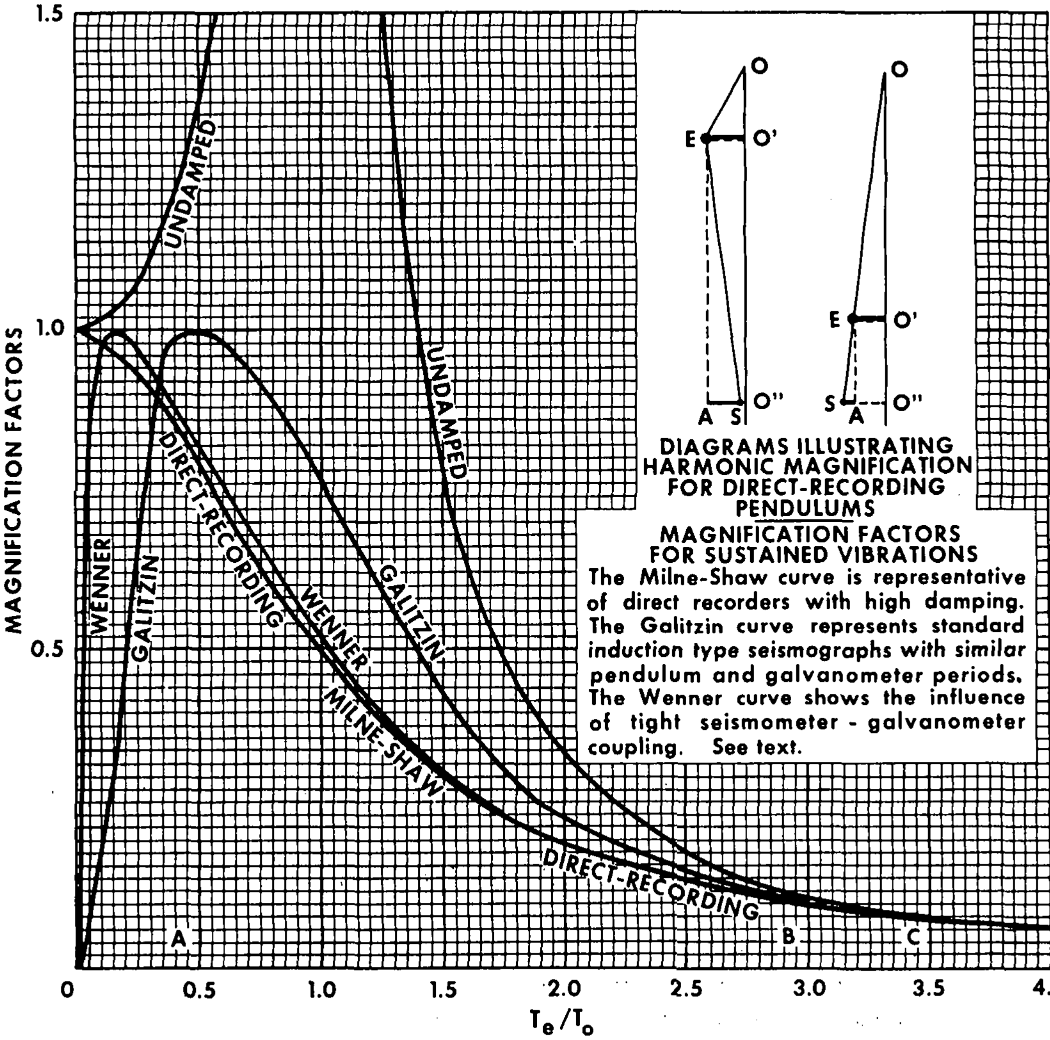

FIGURE 8.—Magnification curves for various types of seismographs, with insert illustrating harmonic magnification.

bob S to swing in the same period just as an earthquake wave forces a seismograph pendulum to swing in the period of the earth wave.

In the first illustration of the insert the period of the earth wave is short compared with that of the pendulum, their periods being governed by the lengths of the strings. It will be observed that after the motion is set up and becomes sustained and when the mass E reaches an extreme position, the bob S will move only a relatively short distance to S from its zero position. Thus, while the earth moves from O′ to E the seismograph pendulum mass moves only from O″ to S. The relative displacement between the ground E and the pendulum S is AS at approximately the extreme position of the swing. The magnification curve shows just what the ratio AS/EO′ is under varying conditions. For the periods indicated in the first illustration the ratio is indicated by an ordinate in the neighborhood of A, figure 8. When the relative magnitudes of earth wave period and pendulum period are reversed, as shown in the second illustration, the ratio SA/EO′ will be indicated by an ordinate in the neighborhood of B. The actual amount which the seismograph pendulum magnifies the sustained simple harmonic ground movement is the ratio SA/EO′ (ordinate of the harmonic magnification curve) multiplied by the lever magnification (mechanical or optical) of the seismograph.

Acceleration.—In figures 6 and 8 it is seen that the magnification curves for various degrees of damping all approach a common curve toward the right-hand side of the graph. It is not desirable to operate ordinary seismographs in this zone (at relatively short pendulum periods) because the harmonic magnification factor for linear displacements is low and the amplitudes on the seismograms would be correspondingly low. This is the zone, however, in which direct-recording pendulums function with nearly constant sensitivity as accelerometers.

From figure 6 it will be seen by inspection of the various magnification curves that the effect of damping becomes almost negligible when T_e/T_o is greater than 2. Beyond this the equation for M_h on page 17 may be written with sufficient accuracy as follows:

$$M_h = \frac{T_o^2}{T_e^2} \text{ for } T_e/T_o > 2$$

The equation for A_e, page 17, then becomes

$$A_e = \frac{A_t}{V} \cdot \frac{T_e^2}{T_o^2} \text{ for } T_e/T_o > 2$$

Substituting this expression in the usual equation for the acceleration of simple harmonic motion

$$\text{a (acceleration)} = \frac{4\pi^2 A_e}{T_e^2} = \frac{4\pi^2 A_t}{V T_o^2} \text{ for } T_e/T_o > 2$$

From this equation, it is seen that the amplitudes on the seismograms are proportional to ground acceleration, regardless of the earth wave period, when T_e/T_o is greater than 2. Under these conditions the trace amplitudes are directly proportional to ground acceleration and inversely proportional to the lever magnification and the square of the natural pendulum period.

MOTION OF GALVANOMETRIC RECORDING SYSTEMS UNDER THE INFLUENCE OF STEADY STATE VIBRATIONS

The amplitudes on galvanometric recorders depend upon the electrical characteristics of the entire system, in addition to those factors already mentioned for direct recorders. Sensitivity is obtained by substituting a very sensitive galvanometer for a direct optical system. Recording can be either optical or, as in recent years, pen-and-ink systems employing electronic circuits to overcome the friction of the writing pen.

The theory of galvanometric recorders is complex and will not be discussed here. While galvanometric recorders are usually more practical for high sensitivity, the harmonic magnification curves in figures 8 and 9 show that they possess one characteristic quite different from those of direct-recording seismographs. For very rapid ground movements the magnification curves begin at zero whereas for direct-recording seismographs, the magnification is the normal lever magnification of the instrument and is constant over a limited period range. The reason for this is that for extremely rapid ground movements, the current generated in the pendulum coils cannot overcome the inertia of the galvanometer suspension system. In the case of the direct recorder, the inertia effect of the optical system is negligible for the period range in which the seismologist is ordinarily interested.

In galvanometric recorders the circuits can be designed in such a way that a combination of short-period pendulum and long-period galvanometer (of the order of 50 seconds) will give the broad performance characteristics of a standard long-period pendulum of the Galitzin type.

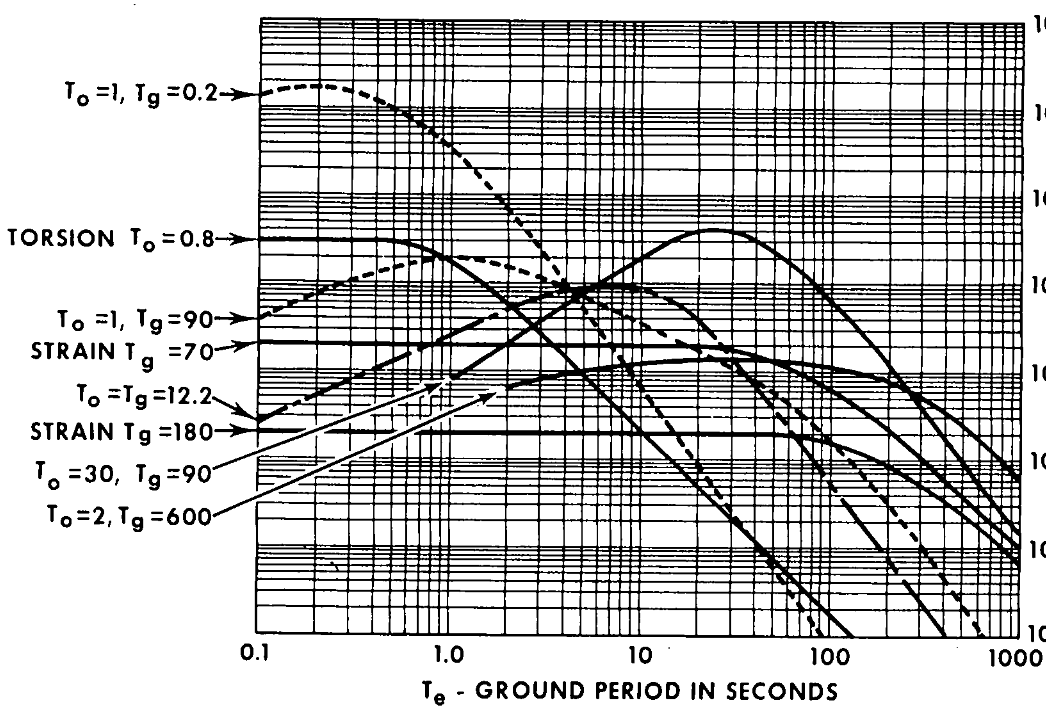

FIGURE 9.—Calibration curves of various seismograph systems.

IV. TRAVEL-TIME TABLES AND CHARTS

Travel-time charts or tables show the times required for the various types of earthquake waves to travel from an earthquake focus (hypocenter) to any point on the earth's surface. They are used principally to determine distance between epicenter and station after certain time intervals (usually S-P) have been measured on a seismogram. These intervals increase with epicentral distance and therefore become measures of the epicentral distance. The time of origin, H, is also ascertained from travel-time tables.

Aside from its value in determining epicentral distance and origin time, the travel-time chart or table supplies the seismologist with the basic data used in postulating the interior structure of the earth. For every change in his concept of the interior structure the research seismologist must be able to show his evidence for such change on a travel-time chart. It is very evident, therefore, that as more detailed and authentic seismographic data are obtained the seismologist will be able to delineate the interior structure more accurately. The great need is to obtain enough evenly distributed instrumental data to enable the investigator to adequately trace the time-distance history of every activity or phase from the time it first appears on the surface of the earth until it disappears. Aside from the variable quality and distribution of seismographs over the earth, one of the handicaps preventing a more satisfactory solution of the earth structure problem is the heterogeneity of the crustal structure which tends to distort the symmetry that would otherwise be obtained in seismographic data. These are problems calling not only for competent seismologists, but for international cooperation on a broad scale, and obviously lie outside the immediate field of interest of the student of seismology. He should know, however, that in his work he is contributing toward this objective.

Determination of distance and time of origin from travel-time tables or charts.—This is based on the fact that any measured time interval between two known phases, such as P and S, must correspond to a definite distance as indicated in the table or on the chart. The time interval increases or decreases as the distance between epicenter and station increases or decreases. Thus, if the interval between P and S is 6m 48s the distance according to the seismological tables is 46.1°, corresponding to approximately 5,115 km. or 3,180 miles. As the time (P–H) required for the P wave to travel this distance is found in the same table, the time of origin may be calculated by subtracting the interval (in this case 8m 40s) from the Greenwich time of P measured on the seismogram. Distance and time of origin can be computed from other phases also, but P and S are generally the best. An exception to this, when sensitive short-period records are available, is the $P'P'$–P interval which has the important advantage of being independent of focal depth. L waves can be used for this purpose only in very special cases because of variations in speed with the type of crustal structure traversed.

The determination of epicentral distance and time of origin is discussed in more detail later under instructions on the interpretation of seismograms, page 33.

Some salient facts about travel-time tables and charts.—In these directions it is presumed that the seismologist will interpret his seismogram with the aid of a travel-time chart or can recognize the various phases from descriptive matter and examples given elsewhere in this publication. Proficiency in this, of course, requires considerable practice. Some seismograms on which there are well-defined activities are nevertheless difficult to interpret, even by experienced seismologists, without knowledge of the approximate distance and time of origin. Sometimes, however, when a solution is problematical and no information is available on the epicenter and origin time, it is possible to limit the interpretation to two choices of distance with different origin times. Such solutions should be used cautiously until a verification of one or the other becomes possible.

The United States Coast and Geodetic Survey distributes preliminary information concerning the locations and times of origin of all strong shocks. This information, published biweekly on the *Preliminary Determination of Epicenter* cards, is available upon request. A primary objective of this service is to assist the seismologist in the proper interpretation of his record. If the distance to an epicenter and the time of origin are known, one only needs to refer to his travel-time table or chart to closely estimate the time of arrival of every wave group at his station. This will be explained in further detail later.

Travel-time charts are available to anyone doing responsible work in seismology. They may be requested from the Environmental Science Services Administration, Coast and Geodetic Survey, Rockville, Md. 20852. The charts are available in two sizes—one similar to those reproduced in this report, and a somewhat larger size for more detailed study—and they can be obtained for depths of 25–700 km.

TRAVEL-TIME TABLES AND CHARTS FOR LONG EPICENTRAL DISTANCES

Travel-time tables are based entirely on observational data. Since a certain degree of consistency has been found in tables developed by various investigators for distances over 1,000 km., there is definite hope that a single table can be developed in the future that will be acceptable to seismologists everywhere. For distances under 1,000 km., the effect of crustal structure is so diversified that it is made the subject of a separate section. The principal factors responsible for the lack of unanimity in the travel times of most of the tables used in long-range work have been (1) time errors in the data due to poor time control and inadequate equipment, (2) use of different criteria in taking the low crustal velocities into account, (3) the assumption of different "shallow" focal depths for use with the standard tables, and (4) the use of different earthquake data by different investigators. Errors of observation, item (1), are being practically eliminated, and items (3) and (4) would seem to require only general agreement as to the most desirable criteria to adopt. This would eliminate most of the differences apparent in the more recent travel-time tables and leave variations in crustal structure as the principal reason for differences between travel times actually observed and those shown in the tables.

Normal depth travel-time tables.—The following tabulation of *P*-wave travel times illustrates the order of differences to be expected in normal depth travel-time tables as developed to date. The Gutenberg-Richter table for long distances assumes an average focal depth of 25 km. in a crustal structure similar to that of southern California; the Macelwane table assumes a 10 or 15 km. depth in continental structure; and the Jeffreys and Bullen international table assumes a focus 33 km. deep at the bottom of a typical continental structure. It will be noted that, in this last table, the travel times are appropriately less than for the more shallow foci assumed in the other two tables. Distances are expressed in degrees, 1° being equal to 111.1 km. of great circle distance. *H* refers to the time of the earthquake at the focus or hypocenter, not the epicenter.

Comparison of travel-time tables

	P–H for normal depths		
Distance in degrees	Gutenberg and Richter 1953	J. B. Macelwane 1933	Jeffreys and Bullen 1940
	m. s.	*m. s.*	*m. s.*
0	0 04	-----	0 05
10	2 28	2 26	2 24
20	4 34	4 33	4 32
30	6 11	6 12	6 08
40	7 36	7 35	7 33
50	8 55	8 53	8 53
60	10 06	10 02	10 06
70	11 12	11 08	11 10
80	12 11	12 07	12 08
90	13 00	12 58	12 57
100	13 46	13 46	13 43
110	14 31	14 33	
120	15 16	15 20	
130	16 01	16 07	

The working seismologist needs this background information to understand why his determination of epicentral distance may not coincide with the epicenter reported by agencies that supply authentic epicenter information. It becomes clear why the final epicenter location is largely an "averaging" process. This means that since the epicentral distance arcs swung from a widespread network of stations do not intersect at a point, the analyst is forced to select an epicenter that is most representative of all the distances used. It should be noted too that an error of one second in determining the arrival time of *P* will result in epicentral distance errors ranging from 8 km. at a distance of 1,000 to 22 km. at a distance of 10,000 km. This applies when epicenters are based on the practice of reconciling all P–H data with a common epicenter and origin time.

Deep-focus travel-time tables.—In the early thirties it was definitely established that, while the great majority of earthquakes originated within or near the crustal layers at depths varying from approximately 10 to 40 km., a few of them originated at varying distances beneath the crust, the deepest reaching down to 700 km. Deep-focus shocks have been accepted as bonafide because the observational data can be reconciled only with theoretical travel times based on such focal depths, the theoretical travel times being based in turn on the interior wave paths and velocities determined from normal depth travel-time tables. Not only are the arrival times of the elementary P and S waves considerably reduced with increase of focal depth, but new phases appear on the records and become a very significant part of deep-focus travel-time curves. Within 2.5 minutes after P and S, new phases are registered that have been reflected one time from the surface of the earth near the epicenter. As previously explained they are designated pP and sS waves, the lower case letter being symbolic of the short ray from the focus to the nearby point of reflection, and the upper-case letter indicative of the long ray from the surface reflecting point to the station.

The travel-time curves in figures 13, 14, and 15 show those phases peculiar to deep-focus shocks. Their arrival times, especially the short interval between P and pP, and S and sS, are the best measures obtainable of focal depth from the records of a single station. When the records and reports of a large number of stations are being processed for epicenter location, the seismologist can utilize both the pP–P intervals and the earlier arrival times of P as measures of focal depth. As stated elsewhere the surface waves for very deep shocks are so weak on the record that their travel time has little significance.

TRAVEL TIMES FOR SHORT EPICENTRAL DISTANCES

For epicentral distances less than 1,000 km. the travel-time problem is difficult because many of the seismic rays traverse the crust over large portions of their paths, and crustal structure can vary greatly insofar as the wave velocity is concerned. It has been stated in preceding sections that in some continental areas, the crustal layering is as much as 40 km. thick and transmits compressional waves through various types of rock at speeds varying from about 6.0 to about 7.5 km/sec. Over some oceanic areas, where crustal layering all but disappears, speeds of 8.0 km/sec., found only in ultra basic rock beneath the crust, may be indicated for local shocks. It is for this reason that seismologists in southern California and in the central Mississippi Valley and the New England areas have all developed travel-time curves applicable to their particular areas. The same is true for other parts of the world. Such local travel-time curves are developed through first accumulating a considerable quantity of nearby seismographic data on local shocks and then, through repeated trials, postulating a structure that yields computed travel-time curves close to those actually observed. Data from large controlled explosions are useful in determining local velocities.

An explanation of local structure and seismic wave velocities in a selected area of southern California will serve to illustrate the problem. The results are based on about 35 years of instrumental data obtained from a network of southern California stations, especially designed and located to solve the travel-time problem, the ultimate objective being the highest attainable accuracy in locating local epicenters. This program is directed by the Pasadena Seismological Laboratory of the California Institute of Technology. Immediately beneath the sedimentary layers a 10 km. thick layer is postulated in which the speed of compressional waves increases from about 6 km/sec. at the top to about 6.5 km/sec. at the bottom. Beneath it there is a possible decrease in velocity with a subsequent increase to 7.0 or 7.5 km/sec. at the Mohorovicic layer which is from 30 to 40 km. deep. Beneath the Mohorovicic layer the compressional wave velocity is 8.2 km/sec. This is substantially the structure contemplated in the local travel-time table for an average focal depth of 18 km. Beyond an epicentral distance of 700 km. the effect of crustal variations becomes negligible. As a result of these studies it is one of the very few regions of the world in which the origin times and epicentral distances obtained from the records of distant stations can be adequately reconciled with those obtained from the records of local stations.

The study of local travel times in other areas can be facilitated by the use of Omori's formula—a formula that assumes simply that all travel-time curves are linear up to somewhat less than 1,000 km. It furnishes a simple relationship between epicentral distance, the $(S-P)$ interval, and P- and S-wave velocities, remembering that the velocity of S is about 0.6 that of P.

If P and S are the phases under consideration, H the origin time, and D the epicentral distance, then

$$\text{the velocity of } P \text{ (slope of the travel-time line)} = D/(P-H) = Vp;$$

$$\text{the velocity of } S \text{ (slope of the travel-time line)} = D/(S-H) = Vs$$

$$(S-P) = (S-H) - (P-H) = D/Vs - D/Vp$$

$$D = \frac{VpVs}{(Vp-Vs)}\,(S-P) = C(S-P)$$

Thus it is seen that when the wave velocities in any locality are known or postulated, the constant C may be computed and the computation of distance simplified. Similar constants may be computed for any pair of phases, such as Pg and Sg ("granitic" layer phases) and P^* and S^*, which are sometimes used in designating speeds in an intermediate zone just above the Mohorovicic discontinuity. $\overline{P}$ and $\overline{S}$ are sometimes used to indicate paths through the entire crust.

V. INTERPRETATION OF SEISMOGRAMS

The preceding discussion of the nature of seismic wave propagation, the recording characteristics of seismographs, and travel-time tables have furnished the background necessary to proceed with the interpretation of seismograms and the preparation of instrumental reports. The outstanding characteristics of the more important ground waves have been outlined and it was made clear that no seismograph has a recording range broad enough to register all seismic ground periods and amplitudes in their true perspective. The remaining discussion will therefore be concerned primarily with what the seismologist actually sees on the seismogram.

ARTIFICIAL AND SEISMIC PHENOMENA COMMON TO ALL TYPES OF PENDULUM SEISMOGRAPHS

Although standard seismographs are designed to register minute ground vibrations due to distant earthquakes, a sensitive pendulum system will not only pick up other types of ground vibration, but will also undergo some peculiar and unexpected motions of its own that will register on the record and sometimes ruin it.

Microseisms.—These are minute vibrations of the ground generated (1) when low-pressure areas pass over large bodies of water, (2) by surf, (3) by the passing of cold fronts over large water areas, (4) by frost, and (5) by miscellaneous types of meteorological disturbances.

The amplitudes of these vibrations vary from hour to hour, depending upon meteorological conditions, and may range in displacement from 0.1 μ (0.0001 mm.) to 25μ and over, depending upon the magnitude and proximity of the disturbance. The smaller amplitudes are generally associated with periods of approximately 0.1 second and over, largely local in character; the larger ones are associated with microseisms of 4- to 7-second period which may travel thousands of miles from their sources over either continental or oceanic paths. There are also microseisms lying between these two period and amplitude extremes. High-magnification and short-period (1 second) pendulums register the short-period microseisms best; approximately 6-second pendulum systems are best for registering the long-period type. The latter group, especially, has been the subject of exhaustive investigation.

Microseisms are generally smooth wave forms that are propagated in groups that alternately vary in trace amplitude from practically zero to the prevailing maxima. They become quite irregular, however, when there is more than one source of disturbance. While this smoothness often distinguishes them from earthquake phases of the same trace amplitude, there are times when microseisms having both long and short periods will be registered and the detection of earthquake phases becomes more difficult. Short-period microseisms can be especially troublesome since they so often lack the smoothness found in those with longer periods.

The magnitude of microseisms and other background activity controls the magnification with which a seismograph may be operated at a given site. Where such background is heavy the weaker phases of distant earthquakes are often totally obscured. Microseisms and other disturbances are generally at a minimum in mid-continental areas and worst on islands in mid-ocean areas. Soft alluvial formations undergo larger amplitudes of motion in the short-period range than rock formations, all other conditions being equal, and are therefore often unsatisfactory as station sites.

Artificial vibrations.—Disturbances due to traffic, machinery, closing doors, the moving about of people on stairs and elsewhere, may be temporary or practically continuous, but they usually quiet down during the night hours. They generally appear on seismograms as short-period vibration on the order of a few tenths of a second or more so that on seismographs that have paper speeds less than 30 mm/min., they may appear as little more than blurs; otherwise the individual waves may be discernible. Weak local shocks sometimes produce brief, short-period records that could be difficult to distinguish from some artificial disturbances. Beginners often have difficulty in distinguishing between real earthquake activity and artificial disturbances of this kind. An earthquake record has a definite pattern depending upon the recording characteristics of the instrument, focal depth, etc.; in other words the experienced seismologist subconsciously looks for a sequence of *P*, *S*, and *L* waves on the seismogram. Through experience alone, the seismologist must learn to recognize the particular types of artificial disturbances to which his station is subject. As in the case of short-period microseisms, the amplitudes of artificial disturbances are always less on rock outcrops than on filled ground or alluvium.

Effects of wind, sun, and convection currents.—Pendulum movements due to these causes can be responsible for much loss of record. In strong winds, building motion may be enough to cause horizontal movement or tilting of the basement floor or pier on which the seismograph is mounted. Such disturbances are generally irregular in character. The alternate heating of the east and west sides of a building may be the cause of an appreciable diurnal tilt, generally effected through the seismometer pendulum, but galvanometers also may be the source of such trouble. This causes a crowding together of the lines on the seismogram at one time of the day and a spreading apart at another. The effect of horizontal tilting of this kind is less on vertical components than on horizontal components because the sine function of a small angle changes faster than the cosine function, and these functions control the amount of the pendulum deflection.

Convection currents in a seismograph room set up pendulum oscillations of a rhythmic character that are often serious and difficult to overcome. Variations in the amount of heat present seem to be the primary cause; such heating comes from the sun, steam pipes, electronic equipment, or other sources. Air-conditioning in a room may generate convection currents. Drying machines have helped to alleviate the condition, but the ultimate solution is most often found in encasing the instrument in an additional outside case and/or covering both instrument and pier with a heavy blanket.

TIMING AND IDENTIFICATION OF PHASES

Definition of "activity" or "phase."—The first step in the interpretation of a record, regardless of its type, is to pick out the times of beginning of the activities or wave groups to be reported before attempting to designate or interpret them. The preliminary activities preceding the long-period surface waves constitute the most important part of the seismogram for determining distance to epicenter and time of origin, and the very first activity regardless of how minor it may seem is nearly always the most important.

If the trace is quiet, no difficulty is usually experienced in recognizing the first activity, but more often than not there will be a certain amount of microseismic activity, and possibly some artificial disturbance, present. If there is any doubt as to whether the activity is really seismic, other parts of the gram should be examined to see if any similar types of activity are present. If so, the questionable activity should be rejected as not of seismic origin. This is usually a safe criterion to follow. More often than not, however, the beginner is inclined to overlook weak preliminary activities and must be frequently reminded of their importance.

There are two ways in which obscure activities may often be made more discernible, especially on low-magnification seismographs. First, procure a long cylindrical lens, similar to those shown in optical shops, which magnify two or three times in the transverse direction. Sighting on the trace through such a lens will be equivalent to doubling or tripling the magnification of the intrument. Another trick, especially useful in revealing the longer period preliminary activities of low amplitude, is to raise one end of the gram close to one eye, or bend down to it, and sight down the trace. It is surprising how otherwise imperceptible activities will stand out in contrast with activities of slightly different character.

After the beginning of the first activity has been selected, other activities should be chosen that are markedly different, either in amplitude or period, from preceding activities. Sometimes only an obvious interference in a prevailing activity will mark the entry of a new wave group. The phases most likely to appear prominently on the seismogram are indicated by the travel-time curves in figure 1.

Time accuracy in reading phases.—The time of beginning of an activity may be measured to the nearest tenth of a second, or second, depending upon the sharpness of the onset. In regional investigations with modern seismographs employing 60 mm/min. paper speeds, important impulsive phases are read to the nearest 0.1 second. Impulsive activities, usually *iP* or *iS*, often appear in standard-type records of strong shocks and the greatest care should be exercised in measuring the Greenwich times of the onsets to the nearest second. If the measured times of the first impulse on two components differ by 1 or 2 seconds the earliest arrival time should be used. It often happens that times are recorded to the nearest second merely out of habit. Such procedure is not justified when the time of emergence is doubtful. The questionable "seconds" readings in such cases should be enclosed in parentheses in the report. It is extremely difficult to select the exact time of beginning of an emergent phase and two persons will often differ in their conclusions concerning the same activity. It is reasonable to presume that anyone using the data may expect the reported time of beginning of a phase to be in error as much as indicated by the last significant figure reported—0.1 second if the time is given to the nearest tenth of a second, and 1 second if it is given to the nearest second. The onsets of surface waves are often so questionable that it may be sufficient to record them to the nearest minute.

The "time correction" is considered positive when the clock which marks time on the seismogram is slow, and negative when the clock is fast.

CHARACTERISTICS OF VARIOUS PHASES ON THE SEISMOGRAM

In the preceding sections describing elementary wave types, *P*, *S*, *Lq* and *Lr* and a series of dependent wave types, emphasis was placed on their origin and nature considered as natural phenomena. Only in a secondary way was reference made to their appearance on the seismogram. With the seismogram now in front of the student, the primary problem is to know how to identify the various wave types and reconcile them with the information found on a travel-time chart or in a table. It will be remembered that periods measured on the seismogram are actual ground periods of the earth wave. The amplitude of the recorded motion, however, varies with the period of the pendulum, the period of the ground wave, and many other factors, and therefore differs greatly for different instruments and ground periods.

Phases at epicentral distances under 1,000 km.—Short-period preliminary and surface waves that last only a matter of some minutes are the mark of a nearby earthquake. Periods are of the order of one-half to several seconds. There will be no great variation in the periods of any of the wave types. The *S* phase is usually distinguishable because of its larger amplitude as compared with *P*, but it is often hopeless to differentiate between *S* waves and *L* waves. This is generally the picture on both long-period and short-period pendulum records, except that short-period pendulum records take on a characteristic form between 700 and 1,200 km. Because this is a zone of weakness for *P* and *S* onsets the record assumes the form of two roughly symmetrical envelopes, one for the *P* group and one for the *S* and *L* groups. Recognition of this characteristic often enables the analyst to make a correct interpretation.

On short-period instruments and sometimes on long-period ones, the concentration of *P′* energy at 15,600 km. may take on the appearance of a local shock and confuse the interpreter. This is also true for *PKKP*, *P′P′*, and *SKPP′* which arrive approximately 18, 26, and 29 minutes after *P*. All such phases have periods varying from 1 to 2 seconds and durations of 1 to 3 minutes.

On long-period pendulum recorders, increasing epicentral distance in this zone is accompanied by the appearance of definitely longer periods that help to identify surface waves. *P* waves are, as always, the shortest periods, but may begin to show underlying periods of perhaps 5 seconds with 1-second periods superposed on them. The *S* wave, too, begins to show great complexity and a tendency toward longer periods. The analyst should definitely be finding other surface layer phases following *P* and *S*. They are characterized by sudden increases in trace amplitude rather than by changes of wave pattern. On the records of short-period pendulums, the *P* and *S* groups are distinguishable more because of amplitude differences than because of the emergence of longer period waves, because such waves are always more subdued on short-period pendulum recorders.

This is a distance zone in which local travel-times must be determined to compute epicentral distances by Omori's formula as previously explained. The analyst should expect to find crustal layer phases, especially in the *P*-wave portion of the record, with travel times depending on these predetermined velocities.

P waves at epicentral distances over 1,000 km.—From roughly 1,000 km. up to the antipodal distance of 20,000 km., the wave-type characteristics of *P* waves do not change materially, although some increase in period with distance is to be expected.

They seldom exceed 5 seconds in period. With increasing distance the short-period, 1-second waves tend to disappear from all groups except *P* and its various combinations, as proved by the records of short-period pendulums that register only these waves at the greater distances. It is the impulsive *P* wave from such instruments, especially the vertical, that today furnishes the best epicenter data. Core reflections are also well registered on vertical components.

A special property of the short-period, high-magnification seismograph is its ability to record elastic sub-sonic waves, called *T* waves, that are sometimes generated by submarine earthquakes and travel thousands of miles across oceanic sound channels. They are registered long after the seismic surface waves are registered, their velocity being only about 1.5 km/sec. Such waves may be transmitted from water to land and back again to water whenever oceanic islands lie in their paths.

On long-period instruments *P* waves and their dependent types are characterized by a combination of long- and short-period waves, the former of the order of 5 seconds' period and the latter around 1 second. While the ground amplitudes of the short-period waves are always smaller than those of longer period waves, they may or may not seriously mask the longer period waves, depending upon the character of the magnification curve. These curves, figures 8 and 9, show how much the displacements for the different ground periods are magnified for various types of seismographs. The analyst must learn the characteristics of different instruments to become proficient in reading seismograms.

In the case of strong distant shocks the first longer period wave of the *P* group may appear very prominently at the start of the record. It may appear coincident with the first impulsive short-period (1-second) wave or may vary from it slightly in arrival time. The longer period onset is designated *eP;* the shorter impulsive onset, *iP*. All normal *P* waves begin to weaken considerably around 11,500 km. because the rays are entering the core. Beyond 15,700 km. a weak phase associated with *P* persists up to about 18,000 km.

Within the first 2½ minutes of *P*-wave activity, the analyst must search carefully for deep-focus phases that will usually appear with sharp onsets, that is, marked increases in amplitude over prevailing activity, or an obvious interference in prevailing activity that markedly increases the complexity of the waves that follow. *PP* waves and *PPP* waves are the most outstanding phases between *P* and *S*. They deviate from the general pattern of *P* waves only in complexity, being superposed on the prolongations of preceding phases. This complexity is most pronounced in moderate to weak signals.

At great distances *P*-wave groups become elongated and seem to lose some of their complexity, except when the so-called focal area of the *P′* core waves is reached at 15,600 km. The trace then may take on the character of a local shock. This is also true of *PKKP*, *P′P′*, *P′P′P′*, and *SKPP′* phases on short-period records. Between 14,500 km. and 15,500 km., *PKS* is the largest of all preliminary activities, arriving about one minute after *PP*. When *P*-type waves arrive almost simultaneously with *S*-type waves, as in the case of the *PS* and *SP* phases, the activity is usually, though not necessarily, complex because of the combining of the longer period *S* waves with the shorter period *P* waves.

S waves at epicentral distances over 1,000 km.—On the average type of long-period seismograph record *S* waves may be pictured as having possibly double the period and amplitude of *P* waves. This may be modified considerably by the magnification curves of the individual seismograph, but for pendulum periods over about 8 seconds, one may always expect to find *S* waves with greater amplitudes than *P* waves. It is for this reason that, for weak recordings, *P* waves may be missing altogether while *S* waves may be fairly well registered. Since the records of 1-second pendulums show these longer period waves in such an obscure way, except in the case of exceptionally strong shocks, the following explanation will refer to the records of long-period or intermediate-period pendulums only.

Although the *S* wave appears in its purest form between 1,000 and 9,000 km., it is still very complex. Some investigators believe that it consists of several different wave types. Because of this and the fact that *P* waves are being registered when the first *S* wave arrives, the arrival time of *S* is never timed as accurately as the first *P*-wave arrival. This explains why more consistent epicentral distances are obtained by using *P–H* intervals (obtaining the correct origin time by trial and error) than by using the *S–P* intervals alone. Because of the uncertainty attached to the onset of the normal *S* wave, the analyst should, by all means, read the two or more outstanding activities present unless there seems little doubt that one reading will give an adequate timing of the phase.

In this zone one finds the *PS* and *SP* phases arriving almost simultaneously, shortly after the normal *S*, and there will more than likely be found some short-period waves attributable to the last *P*-wave leg of the *SP* combination. *SS* and *SSS* phases show waves of longer period with increasing epicentral distance so that at great distances, there is sometimes danger of confusing them with surface waves. *P*-wave and *S*-wave surface reflections are distinctive in that they are registered at all distances from the epicentral area to the antipode.

At 10,000 km. the normal *S* wave begins to fade out of the picture because the rays are being partly refracted into the core as *P* waves and refracted out again to the surface as *S* or *P* waves. The phase, *SKS*, is readily identified on the record, coming in somewhat ahead of an imaginary extension of the normal *S*-wave curve, and followed after some minutes by the *PS* and *SP* combination that is not affected by the core. As the normal *S* phase fades out, *SKS* builds up in trace amplitude and, together with *PS* and the surface-reflected phases, comprises the most important *S*-type phases on the record at great distances. The Gutenberg travel-time charts should be consulted for a large number of other phases that may be found on the seismogram.

Surface waves at all distances.—Surface waves are always the largest trace amplitude waves on the records of all long- or intermediate-period pendulums. On short-period pendulum records they are out of the picture, except in the stronger shocks. Because of their irregular travel times due to variations in crustal structure, they have little use in helping to determine epicentral distances, except that their approximate arrival time is often sufficient to make a choice between two possible distance determinations based on preliminary phases alone. The study of surface waves as a means of analyzing surface structure is in the research category. Surface waves increase on most records from a few seconds' period in the epicentral area to several hundred seconds at great distances for the waves of maximum trace amplitude. Investigations of destructive earthquake motion with special instrumentation show that long-period surface waves actually are present in epicentral areas, but on most nearby seismographs, they are obscured by the shorter period waves that are generally magnified many times more than the long-period waves.

On many records the arrival time of the *Lq* wave is frequently obscured by *S*-wave surface reflections, but it can usually be identified because the first few *Lq* waves are often the longest on the record, sometimes reaching a period of a minute or more. If the pendulums happen to be oriented end-on with respect to the epicenter and all the *Lq* waves are on one component, successive waves of the train will be seen to increase in amplitude as they decrease in period until a maximum amplitude is reached. With such pendulum orientation, the *Lr* waves will be seen to arrive later on the other horizontal and the vertical components and behave in much the same way, except that ultra long periods are not found at the beginning of the wave train. If the pendulums are not oriented end-on, which is far more likely, the *Lq* waves will arrive first with the *Lr* waves piling over them some minutes later (at the greater distances) to make a tangle of wave motion that can be separated only by a laborious type of analysis that is never done in routine work and seldom in special investigations.

Recent developments of seismographs with very long-period characteristics have furnished a great impetus to a more complete understanding of the propagation of long-period waves. The waves in this particular group or family are known as guided waves because they are transmitted through layers or channels in the earth's crust and mantle. The shorter period waves of the Love and Rayleigh group, described above, are channel waves in the crust (continental and oceanic). The *Lq* or *G* wave with its long period and great wave length is essentially a channel wave in the mantle. The *Lq* wave, a short-period wave with transverse characteristics similar to *L* and usually superposed on a long-period surface wave, travels over long continental paths with relatively little loss of energy, but it is cut off abruptly when its path has even a small oceanic segment. It is a channel wave in the crust. A low-velocity channel in the ocean, the Sofar channel, transmits energy horizontally with slight energy loss and has been used for long distance signaling. The *T* wave is a short-period wave group that travels from its source with the speed of sound in water and is propagated from shore lines into the continental interior with ordinary seismic velocities.

IDENTIFICATION OF PHASES AND DETERMINATION OF EPICENTRAL DISTANCE AND ORIGIN TIME

An experienced seismologist will, more often than not, be able to assign correct interpretations to the activities registered on a seismogram merely by inspection. It is always desirable, however, to check the interpretation on a travel-time chart. Presuming that the times of beginning of the various uninterpreted activities have been tabulated on forms provided for that purpose, or on a scratch pad, the following procedure should be observed:

(1) Take a narrow strip of paper (such as adding-machine tape) nearly twice as long as the travel-time chart, and lay it on the chart below a graduated time scale such as that at 15,500 km. in figure 1. Along the top edge of the strip make two marks (such as inverted V's) to coincide with the 0- and 60-minute time ordinates. These are merely references that make it easy to replace the scale in the same position later on. The zero time ordinate indicates the hour of the earthquake, such as 3h, 10h, 23h, etc.

(2) With the strip lying along the time scale in the position indicated under (1) make marks, short lines preferably, along the upper edge at points corresponding to

the times of the activities previously measured on the seismogram. Thus, if the time of the first activity on the seismogram has been measured and reads 10h. 13m. 35s., make a mark on the tape at a point about 0.6 or 35/60 the distance between the 13- and 14-minute graduations. It is sufficient to use tenths of minutes in this exploratory procedure. Make similar marks on the strip for all the other recorded phases, marking the last one *L* as there is usually not much difficulty in recognizing surface waves on records of distant earthquakes. Marking the outstanding activities with a heavier and longer line on the slip will be a help in the steps to follow. It will also help to indicate the *P*- or *S*-wave character of all the phases on the strip of paper as well as they can be judged by inspection of the seismogram.

(3) Now move the strip from its original position on the time scale and, keeping it horizontal, slide it up and down and laterally over the curves. If *P*-wave types are indicated on the seismograms, then the first marks ought to coincide either with the *P*, *PP*, or *P′* curves. At the same time the mark corresponding to the first *S* wave (usually having greater periods and amplitudes than *P* waves) should coincide with either *S*, *SS*, or *SPS* (refer again, if necessary, to the characteristics of the various types as described in other parts of these notes). If the correct distance has been found, the mark for the *L* waves will lie reasonably close to one of the *L*-wave curves, within possibly a minute or two if the distance is great. Make a note on the tape of the distance at which the final setting is made. This is the approximate distance from station to epicenter. With the *P*, *S*, or other phases thus identified, a more accurate value of distance can be obtained from the seismological tables.

(4) With the strip in the last position on the curves, and practically all of the important measured activities coinciding with the travel-time curves, write beneath each tape mark the symbol of the phase with which it coincides such as *P*, *PP*, *S*, *SS*, *Lq*, etc. When this is done, make a new mark on the tape (with the strip in the same position) to indicate the point of coincidence with the zero time ordinate and mark it *H* for time origin.

(5) Move the tape from its position on the curves back to its original position on the time scale as described in (1) being sure to make the even hour marks on the slip coincide with the 0- and 60-minute ordinates on the chart. As the slip marks have all been interpreted in terms of *P*, *S*, and *L*, etc., it is then a simple matter to write these interpretations of the phases or activities next to the times of the onsets which have already been tabulated. With the strip on the time scale, note the time at which the origin mark *H* falls on the scale. This reading will be the time of origin of the earthquake at the hypocenter or focus. For a more accurate origin time, read the *P–H* interval in the seismological table that corresponds to the *S–P* interval determined in (3) and subtract it from the time of *P* as read on the seismogram.

The above directions seem simple, but many difficult situations arise in practice. Phases indicated on the travel-time charts are not always discernible on the records, and activities sometimes appear on the records that are not accounted for on the travel-time curves. The key to successful and worthwhile interpretation lies in the proper selection of the really important preliminary activities; not in trying to make poor readings fit the curves. The analyst must memorize the characteristics of the approximately half-dozen outstanding phases and this can be done thoroughly only by experience in reading seismograms. Word descriptions are inadequate because of the complexities and variations involved.

INTERPRETATION OF SEISMOGRAMS WHEN DISTANCE AND TIME OF ORIGIN ARE KNOWN

Within a week or less after important shocks occur, seismological station directors are informed through seismological centers such as the United States Coast and Geodetic Survey of the locations of the epicenters and the times of origin. The seismologist then has every requirement for making a correct analysis of all the activities recorded on his seismograph.

A graphical method of determining the times of arrival of the various phases is, in effect, a reversal of the process just described. Take the paper tape and place it on the time scale as in paragraph (1) of the preceding section. Make a mark on it corresponding to the time of origin as reported, indicating at the same time the even-hour positions. Then move the tape from the time scale to the distance scale graduation corresponding to the correct distance and make the time of origin mark coincide with the zero time ordinate of the graph. Draw lines on the tape corresponding to each of the travel-time curves and write the corresponding phase symbols below each of the lines. Move the tape back to its original position on the time scale. The times corresponding to the various phase marks are the times at which these phases should appear on the grams.

Another graphical method consists in first making a time scale on a long tape exactly to scale with that on the travel-time chart but covering 2 hours of time instead of 1 hour. Lay this tape over the curves in such a position that the graduated edge will cross the distance scale at the correct epicentral distance. The time on the graduated tape corresponding to the time of origin of the earthquake should coincide with zero time on the graph. The arrival times of the various phases will then be indicated on the graduated tape by the points where the various travel-time curves intersect the edge of the tape. This is probably quicker than any other method for estimating arrival times of a group of phases.

A more accurate way to compute arrival times when epicenter and origin time are known is to make use of the *P–H*, *S–H*, and similar intervals usually found in seismological tables.

COMPUTATION OF GROUND AMPLITUDE

When ground amplitudes are desired for seismological bulletins or a wide variety of other purposes, they can be computed as explained in the section on response of the seismograph to seismic waves. The formula used in routine analytical work is

$$A_e = \frac{A_t}{VM_h}$$

where

A_e = amplitude of the earth movement.
A_t = trace amplitude.
V = static or lever magnification of direct-recording pendulums.
M_h = harmonic magnification factor.

In the case of electrical recorders a magnification curve peculiar to each instrument must be substituted for VM_h. The magnification is always a function of the ground period, the pendulum period, and, for electrical recorders, the galvanometer period and other factors.

The trace amplitude of a wave is one-half the distance between its peak and trough (generally the maximum amplitude in the group) and the period is the time interval from crest to crest or trough to trough. On the record of a direct-recording pendulum, let

$A_t=12.5$ mm.	$V=150$
$T_e=20$ sec.	$\epsilon=20:1$
$T_o=12$ sec.	$T_e/T_o=20/12=1.67$

From figure 8 the factor M_h when $T_e/T_o=1.67$ is 0.26. Then

$$A_e \text{ (in 0.001 mm.)}=\frac{1000\times 12.5}{150\times 0.26}=320\mu$$

In the case of electrical recorders, the entire magnification factor must be taken from calibration curves such as shown in figure 8 for Galitzin and Wenner seismographs, or as in figure 9 for various other seismographs.

In special cases where seismographs are serving as accelerographs to record destructive ground motions, the acceleration and period must first be measured directly from the record and the amplitude, or displacement, computed from the formula

$$A_e=\frac{aT_e^2}{4\pi^2}$$

where a is the acceleration in cm/sec.², A_e is expressed in centimeters, and T_e is the period of the wave in seconds.

DIRECTION OF GROUND VIBRATION

Being able to determine the direction of ground motion of the simpler types of wave motion enables the seismologist to determine the direction of the epicenter from the first impulsive P wave when it is clearly registered on the seismogram, and also enables him to study the directional characteristics of the smoother types of surface waves thus aiding in their identification. To do this the directional constants of the two horizontal and the vertical component pendulums must be known. The seismologist must know, for example, whether sudden upward motions of the traces correspond to sudden north or south, east or west, and upward or downward ground motions. Since for a sudden northward movement of the ground the relative motion of the pendulum is to the south, the directional constant for a northward ground motion can be found by giving the pendulum a sudden movement to the south and noting whether the trace moves upward or downward on the record. An impulsive ground motion is always opposite to the relative pendulum motion. The same reasoning applies to the other two components. The results are given in their simplest form when the station director simply states in his bulletin that sudden upward movements of the traces correspond to N (or S), E (or W), and U (or D) ground movements. In galvanometric recorders these directional constants will be reversed if the galvanometer or seismometer terminals are reversed.

Azimuth of epicenter.—The direction of ground motion on the arrival of the first impulsive compressional (longitudinal) wave will be away from the epicenter if it is compressional and toward the epicenter if it is rarefactional. By measuring the amplitudes of the first N-S and E-W impulses and knowing the directional constants of the pendulums, the direction of the resultant motion can be determined by a simple vector analysis such as shown in the insert of figure 1. Thus, if rapid upward movements of the traces correspond to northward and eastward ground movements and the respective trace amplitudes are 4.6 mm. up and 7.2 mm. down, the resultant direction of the first impulsive ground movement will be north 57° west. Since a compressional wave is always indicated by a movement of the ground away from the epicenter and a rarefactional wave by a movement toward the epicenter, it becomes obvious that the epicenter could be either to the northwest or southeast. This ambiguity is overcome when a vertical-motion record becomes available because upward ground motions are always compressional and downward motions rarefactional.

VI. MISCELLANEOUS

OPERATIONAL PROBLEMS AFFECTING THE INTERPRETATION OF SEISMOGRAMS

Previous discussion, under artificial disturbances, has already covered some of the most troublesome features of seismograph operation, but there are a few additional ones that do not fall in that category of artificial disturbances. Those most often affecting the efficient analysis of records are the quality of photographic traces and time control. These two phases of seismograph operation should be uppermost in a station director's mind because deficiencies in either one could largely nullify other excellent features of his work.

Weak traces and poor focusing frequently go hand-in-hand to vitiate what might otherwise be satisfactory records. It is not sufficient that a trace be heavy enough to record just the normal microseismic activity. The function of the instruments calls for the recording of minute tremors of short period, and also amplitudes that approach the edges of the seismogram. But too often a bad focus obliterates the feeble but especially important *P*-wave onset, or, combined with a weak light source, may be responsible for the loss of other important activities in the case of strong shocks. The normal trace should be slightly over one-half millimeter thick, solid black and well focused; that is, the beginning and end of the trace at the minute break should appear as a sharp vertical line.

Good time control is a prime requisite. This does not mean that it is sufficient simply to have a clock somewhere in the building on which the correction is known. Whenever possible, time signals should be placed automatically on each record daily, and more often if local station networks are working on special projects. Crystal clocks with rates of a few thousandths of a second per month are currently being employed in stations for time control and for driving the recording drums. For efficient work the seismologist requires minute marks on the seismogram with preferably a zero correction. On such a record the times of many emergent activities can be read to the nearest tenth minute simply by inspection. This cannot be done when there is a large correction or when only the times of beginning and ending of a day's run are given.

Even at best, a disconcerting problem is always present in measuring time on a gram because it has been demonstrated that even on high-grade drums, the errors due to irregular rotation from one minute mark to another sometimes exceed 1.5 seconds, although 0.3 second is a more probable value. When it is realized that the waves being measured travel with surface velocities from 5 km/sec. to as high as 20 km/sec., in certain zones, the necessity for precision in time becomes readily apparent.

In those cases where seismograms are sent to central offices for final interpretation, a practice that is growing each year, station directors are urged to adhere faithfully to the details of marking the seismograms as suggested by the cooperating central office. As such agencies are today interpreting thousands of records, the speed and effectiveness of their work is greatly enhanced by the correct marking of incoming records. All seismograms should show (1) station name, (2) type of instrument, if more than one type is in operation, (3) component, (4) direction of ground motion, (5) first and last hour of G.C.T., and (6) time corrections at beginning and end of record, indicating whether correction is made to the beginning, end, or middle of the time break.

EARTHQUAKE INTENSITY AND EARTHQUAKE MAGNITUDE

Since the introduction of the Gutenberg-Richter magnitude scale into seismological literature about 25 years ago, much confusion has resulted in the mind of the interested layman and in the press concerning the definition of earthquake intensity and earthquake magnitude.

Earthquake intensity refers to the violence of earthquake motion in any part of the perceptible area of an earthquake and is based solely on the effects observed on people and inanimate objects, such as buildings and their contents. In the United States the scale used to classify intensity is the *Modified Mercalli Intensity Scale of 1931*, described in detail by Wood and Neumann (1931, pp. 277–283). Intensity I is rarely felt; VI causes slight damage; XII is catastrophic. Maps showing the distribution of intensity (isoseismal maps) are based on such gradations.

Earthquake magnitude refers, in effect, to the total energy released at the source of the disturbance. There can be only one magnitude associated with any earthquake. It can be determined from one or more instrumental records obtained at any epicentral distance, but each station record should indicate approximately the same magnitude at the source.

As there should obviously be some fixed relationship between the magnitude of an earthquake and the maximum intensity observed close to the source of the earthquake, the following comparison has been made by Gutenberg and Richter for earthquakes in southern California when the normal depth of focus is considered to be 18 km.

Magnitude	2.2	3	4	5	6	7	8	8.5
M-M Intensity	1.5	2.8	4.5	6.2	7.8	9.5	11.2	12.0

According to the latest estimates of Gutenberg and Richter, a magnitude 5 earthquake would represent the release of 10^{21} ergs of energy which is equivalent to the energy released by an atomic bomb or 20,000 tons of TNT. Shallow-depth earthquakes just strong enough to be felt represent an expenditure of about 10^{11} ergs; the greatest earthquakes release about 10^{25} ergs. The magnitude and intensity scales are both exponential in character. An increase of 1 unit in magnitude signifies a 63-fold increase in energy; an increase of 0.1 unit signifies an approximately 50 percent increase.

DETERMINATION OF INSTRUMENTAL MAGNITUDE

The magnitude scale is derived from an empirical formula based on instrumental results and is used extensively as a measure of earthquake energy released at the focus.

The magnitude of an earthquake is the logarithm of the maximum trace amplitude, expressed in microns, with which the standard short-period torsion seismometer (T_o=0.8 sec., V=2,800, h=0.8) would register that shock at an epicentral distance of 100 km. Richter (1935) published a table for determining the magnitudes according to the above definition for distances from 25 to 600 km. However, it should be cautioned that the data are for southern California earthquakes and as such, may require a correction factor for use in other areas.

In 1945, Gutenberg expanded the magnitude scale for earthquakes recorded by all seismometers with known response characteristics, for all distances, and for all depths.

For the surface waves the magnitude may be obtained from the equation

$$M=\log A-\log B+C+D$$

where

A=the horizontal component of the maximum ground movement, in microns, of surface waves of 20 seconds' period.

B=the same quantity for a magnitude zero shock. Gutenberg (1945, pp. 3–12) has tabulated the values for each degree of distance.

C=a station constant depending on instrumental characteristics and geological foundation (seldom more than ± 0.2).

D=factor depending on depth of focus, original distribution of energy in the azimuth, absorption of wave energy, and path of transmission. The corrections for this term are appreciable and should be evaluated since they frequently are ± 0.5. The large variations are caused by the waves crossing alternately the continental and oceanic structures where the absorption is approximately $-24\ k\ \Delta$, k being a factor contingent on the path and Δ, the epicentral distance measured in degrees.

If well-defined P, PP, and S waves are measurable on the seismograms, the determination of magnitude is greatly simplified by employing the equation

$$M=A+0.1\ (M-7)-\log T+\log u \text{ (or } \log w)$$

where A is a function of distance, as tabulated by Gutenberg (1945, p. 65); T is the period of P, PP, or S in seconds; u is the horizontal ground motion of the same wave in microns; and w is the vertical ground motion. Although this equation has fewer variables than that for the surface wave, the results are not as reliable because of the frequently poor definition of the waves and the directional effect of the P waves, as described by Byerly (1949). The same body waves and equation may be used for the magnitude of deep-focus earthquakes. However, the values of A to be substituted in the equation will differ from the normal depth values as reported by Gutenberg (1945, pp. 117–130).

Those seriously interested in magnitude determinations should carefully study the references cited as this brief description serves only to explain the nature of the computation. Magnitude determination is considered outside the scope of routine work, except at a few centers where full-time seismologists are employed.

SEISMOLOGICAL BULLETINS

A monthly or quarterly seismological bulletin is the usual medium through which a station publishes its seismographic results. In many cases such data are published in composite form by a central agency rather than by individual stations. For many years, international standard practice called for the following information: (1) character of earthquake in terms of strength and distance zone, as explained in the nomenclature in the next section, (2) station and instrumental constants, (3) the phase interpretations and their arrival times on all components, (4) the ground amplitudes of outstanding phases, (5) computed epicentral distance and time of origin, and (6) the compressional or rarefactional character of P.

Because of the great increase in the number of instruments, the centralization of a large part of the analytical work, and the great amount of labor involved in the standard international program, some seismological agencies have been forced to curtail their reports to the general form shown in figure 10 in which no distinction is made between components, and amplitudes are not reported. In cases where the instrumental constants are known the maximum amplitudes for strong shocks are sometimes furnished to supply magnitude data. It is mandatory, however, that the instrumental constants be reported at least annually in all station bulletins.

In figure 10, all coordinates of epicenters, origin times, and focal depths have been calculated with the use of an electronic computer. The epicenters quoted were previously reported on *Preliminary Determination of Epicenter* cards and are published in the *Seismological Bulletin* with some refinement and minor additions.

Station directors who conduct their programs independently or with only limited aid from another organization are urged to become sufficiently proficient in seismogram interpretation to prepare their own bulletins. This adds materially to the prestige of the organization sponsoring the station; it results in more thorough interpretations than are generally possible at a central station; and it alleviates the workload at central stations that are practically all obliged to cut their interpretation programs below a desirable minimum. Station sponsors should look upon such programs as long-range projects, for it is only through years of experience that station directors can become proficient enough to produce bulletins of real value and ultimately conduct research projects of their own.

NOMENCLATURE USED IN INSTRUMENTAL REPORTS

1. *Character of earthquake recorded—(international usage)*:

    I. Slight. II. Moderately strong. III. Violent.

    - *d* — Local shock (origin less than 100 km. distant).
    - *v* — Near shock (origin from 100 to 1,000 km. distant).
    - *r* — Distant shock (origin from 1,000 to 5,000 km. distant).
    - *u* — Very distant shock or teleseism (origin more than 5,000 km. distant).

2. *Body wave phases for normal depth earthquakes:*

    No mark between the symbols P, P' and S indicates reflection at the surface of the earth; K indicates a compressional wave in the core (S is not transmitted); no mark between two K's indicates reflection at the inner surface of the core; c indicates reflection at the outer surface of the core.

    - P or Pn — Direct longitudinal waves that have passed below the continental layers.
    - P' or PKP — Direct longitudinal waves that have traversed the core.
    - S — Direct transverse waves that have passed below the continental layers.
    - PP or PR_1 — Longitudinal waves reflected one time at the earth's surface.

SEISMOLOGICAL BULLETIN

Date and Station	Phase (GCT)
	h m s
JUL 5	
H-21 03 41.4	
51.5N 171.9E	
h ABOUT 26KM	
COL IP	21 09 01.3
HHM E(P)	21 12 03.2
BOZ EP	21 12 26.0
I	12 31.7
EUR IP	21 12 35.9
JUL 5	
H-23 41 46.9	
21.3N 094.0E	
h ABOUT 65KM	
COL EP	23 54 01.2
JUL 6	
H-01 12 47.6	
59.9N 149.3W	
h ABOUT 44KM	
COL IP	01 14 02.3
HHM EP	01 17 52.8
BOZ EP	01 18 21.0
EUR EP	01 18 47.9
ALQ EP	01 20 03.0
JUL 6	
EUR EP	01 40 49.2
JUL 6	
EUR EP	01 44 22.8
JUL 6	
H-02 08 17.3	
17.8S 174.6W	
h ABOUT 26KM	
COL EP	02 20 51.0
JUL 6	
H-03 04 21.2	
22.5S 172.9E	
h ABOUT 54KM	
GUA EP	03 12 35.0
HON EP	03 13 27.0
EL	28 28.0
KIP EP	03 13 28.0
BCN EP	03 17 15.6
EUR IP	03 17 19.2
IpP	17 35.7
EPP	21 22.8
EL	44 28.0
TUC EP	03 17 21.6
COL IP	03 17 24.9
GCA E(P)	03 17 28.8
ALQ IP	03 17 40.6
JUL 6	
H-03 18 42.7	
38.4N 022.4E	
h ABOUT 20KM	
BRW EP	03 29 58.5
WAS IP	03 30 13.0
E(S)	39 43.0
WSC EP	03 30 15.0
COL IP	03 30 33.4
IS	40 28.0
SJG EP	03 30 40.5
ES	40 34.2
CSC IP D	03 30 46.5
E	31 00.0
ES	40 48.0
SIT E(P)	03 31 08.8
HHM IP	03 31 20.1
ES	41 46.0
EP'P'	57 27.3
BOZ EP	03 31 27.7
E(PPP)	36 44.0
ES	42 00.0
ESKS	43 02.0
EPKKP	49 16.0
EP'P'	57 24.0
BUT EP D	03 31 27.0
ESKS	42 07.8
SLC IP	03 31 45.8
ALQ IP	03 31 55.8
EPKKP	49 03.8
EUR IP	03 31 59.1
E	32 35.7
I(PP)	35 47.8
ESKS	42 36.0
IPKKP	49 04.5
IP'P'	57 15.0
BHP EP	03 32 00.0
GCA EP	03 32 00.0
BCN EP	03 32 12.0
TUC EP	03 32 15.0
EPP	36 18.0
KIP EP'	03 37 36.4
JUL 6	
COL EP	03 49 46.0
JUL 6	
COL E(P)	03 54 05.0
JUL 6	
H-04 08 47.4	
46.8N 152.5E	
h ABOUT 39KM	
BRW EP	04 15 31.0
COL IP	04 15 50.4
HHM IP	04 18 46.1
BUT EP	04 19 01.3
BOZ IP C	04 19 07.0
EsP	19 21.0
EUR IP	04 19 15.8
IpP	19 26.4
IsP	19 31.0
BCN EP	04 19 35.2
EsP	19 49.6
TUC EP	04 20 05.2
IsP	20 21.1
JUL 6	
H-04 09 00.8	
6.7 N 127.0 E	
h ABOUT 44KM	
COL EP	04 21 16.2
CSC EP	04 21 26.0
JUL 6	
H-04 34 22.4	
07.3N 127.0E	
h ABOUT 45KM	
COL EP	04 46 36.0
JUL 6	
EUR IP	04 45 25.0
JUL 6	
H-04 48 33.6	
04.0N 113.7E	
h ABOUT 65KM	
COL E(P)	05 01 27.5
BOZ EP'	05 07 13.0
EUR EP'	05 07 14.0
ALQ EP'	05 07 30.4
JUL 6	
H-04 58 55.6	
55.0N 162.0E	
h ABOUT 34KM	
BRW EP	05 04 09.4
COL IP	05 04 28.8
BOZ E(P)	05 07 13.0
HHM E(P)	05 07 44.7
BOZ IP	05 08 07.5
EUR IP	05 08 22.9
I	08 35.9
BCN EP	05 08 46.0
E	08 54.0
TUC EP	05 09 19.0
E	09 27.0
ALQ EP	05 09 21.2
JUL 6	
H-05 15 12.7	
7.3 N 126.8 E	
h ABOUT 84KM	
COL E(P)	05 27 21.2
EUR E(P)	05 15 42.7
JUL 6	
EUR E(P)	05 31 51.0
JUL 6	
EUR E(P)	05 47 28.0
JUL 6	
BOZ EP	05 50 42.0
JUL 6	
KIP E(P)	05 54 47.6
JUL 6	
H-06 12 41.4	
54.3N 160.1E	
h ABOUT 33KM	
COL E(P)	06 18 30.4
EUR EP	06 22 18.3
JUL 6	
BOZ EP	06 30 33.0
EUR EP	06 30 42.7
JUL 6	
HHM EP	06 35 16.6
JUL 6	
COL EP	06 39 22.0
JUL 6	
H-06 39 25.0	
44.5N 148.3E	
h ABOUT 56KM	
COL IP	06 46 55.4
HHM EP	06 49 46.8
BOZ IP	06 50 07.0
EUR EP	06 50 15.5
IpP	50 30.5
JUL 6	
BOZ IP	07 14 26.6
IL	14 35.0
HHM EP	07 15 00.4
EL	15 35.2
JUL 6	
BOZ IP	07 34 13.1
IS	34 24.5
JUL 6	
HHM EP	07 35 11.2
E	36 08.2

Figure 10.—Sample page from a seismological bulletin of the Coast and Geodetic Survey for July 1965.

PPP or PR_2	Longitudinal waves reflected two times at the earth's surface.
PcP	Longitudinal waves reflected from the outer surface of the core.
PcP'	Longitudinal waves reflected from the outer surface of the core and then reflected from the earth's surface as *P'* waves.
PcSP'	Longitudinal waves transformed on reflection from the outer surface of the core, and then transformed on reflection at the earth's surface into *P'* waves.
P'P'	*P'* waves reflected from the earth's surface, passing twice through the core.
PKKP	Longitudinal waves reflected one time from the inner surface of the core, passing twice through the core.
PS	Longitudinal waves transformed on reflection from the earth's surface. (Have almost same travel time as *SP*.)
SP	Transverse waves transformed on reflection from the earth's surface. (Have almost same travel time as *PS*.)
PPS	Longitudinal waves reflected one time at surface of earth and transformed on second reflection at same surface. *PPS*, *PSP* and *SPP* have approximately same theoretical travel times.
PSPS	Longitudinal waves transformed at each of three reflections from the earth's surface. *PPSS*, *SPSP*, *SPPS*, and *PSSP* all have approximately the same theoretical travel times.
P'_1 and P'_2	The two branches of *P'* waves that have traversed the core and emerged beyond 145° epicentral distance. See text on page 12.
SS or SR_1	Transverse waves reflected one time at the earth's surface.
SSS or SR_2	Transverse waves reflected two times at the earth's surface.
ScS	Transverse waves reflected from the outer surface of the core.
ScSP'	Transverse waves reflected from the outer surface of the core and then transformed on reflection from the earth's surface into *P'* waves.
ScSScS	Transverse waves reflected successively from the outer surface of the core, the surface of the earth, and the outer surface of the core.
SKS	Transverse waves transformed into compressional waves on refraction into the core and transformed back to transverse waves on refraction out of the core to the earth's surface.
SKKS	Transverse waves transformed on refraction into the core into *P* waves that are reflected from the inner surface of the core and then transformed back into transverse waves on the last leg to the surface.
SKP	Transverse waves transformed into compressional waves on refraction into the core and remaining compressional after refraction out of the core to the earth's surface.
SKPP'	*SKP* waves reflected at the earth's surface as *P'* waves.
SKSP'	*SKS* waves transformed into *P'* waves on reflection at the earth's surface.

3. *Body wave phases for deep-focus earthquakes:*

In addition to the preceding phases listed for normal depth earthquakes, deep-focus earthquakes also register many duplicate phases generated when the waves from deep foci strike the earth's surface near the epicenter and are reflected back again into the earth. Each symbol for a duplicate phase generated in this manner is preceded by a small p or s, symbolizing the short path from focus to the earth's surface and subsequent reflection.

Examples:

- *pP* — Compressional waves from a deep-focus shock reflected at the earth's surface near the epicenter and propagated thereafter as direct waves from the point of reflection to the station.
- *pSP* — Compressional waves transformed into transverse waves on reflection at the earth's surface near the epicenter and transformed back into compressional waves on a second reflection from the earth's surface as in the case of a normal *SP* wave.

4. *Mohorovicic's nomenclature for crustal layer phases:*

- *Pn* — Same as normal first preliminary tremors *P*.
- $\overline{P}$ — Individual or upper first preliminary tremors whose path lies entirely in the continental layer.
- $\overline{S}$ — Individual or upper second preliminary tremors whose path lies entirely in the continental layer.
- $R_1\overline{S}$ — Upper second preliminary tremors reflected from the lower boundary of the continental layer.
- $Rs\overline{S}$ — Upper second preliminary tremors reflected from the upper surface of the continental layer.
- $Rs_2\overline{S}$ — Upper second preliminary tremors reflected two times from the upper surface of the continental layer.

5. *Jeffreys' nomenclature for crustal layer phases:*

- *Pn* — Same as normal first preliminary tremors *P*.
- *Pg* — Same as $\overline{P}$ of Mohorovicic. The path is entirely in the upper layer.
- *Sg* — Same as $\overline{S}$ of Mohorovicic. The path is entirely in the upper layer.
- *P** — Upper first preliminary tremor whose path lies along the top of the intermediate layer except for short distances in the surface layer.
- *S** — Upper second preliminary tremor whose path lies along the top of the intermediate layer except for short distances in the surface layer.

6. *Surface waves:*

- *L* — Long waves, unidentified, at the beginning of the surface phase.
- *Lq* or *G* — Surface shear (Love) waves in which the earth particle oscillates in the horizontal plane normal to the direction of propagation. The shorter period waves are channel waves in the crust. The *G* waves, at beginning of the *L* group with waves of long period (1 to 4 minutes) and large amplitudes with characteristics of *Lq* are essentially channel waves in the mantle.
- *Lr* — Surface (Rayleigh) waves in which the earth particle describes a retrograde elliptical orbit in a vertical plane in the line of propagation. The shorter period waves are channel waves in the crust.
- *T* — Short-period wave group that travels with the speed of sound in water from its source and is propagated from shorelines into continental interior with ordinary seismic velocities. Upon im-

pinging on the shore it gives rise to P, S, and L waves that are recorded as a complex and irregular T phase.

Lg A short-period wave with transverse characteristics similar to L, superposed on long-period surface waves, which travels over long continental paths with relatively little loss of energy but is cut off abruptly when its path has even a small oceanic segment. It is a channel wave in the crust.

M Waves of maximum amplitude in the main phase.

C Waves of the end portion.

W_2 Surface waves traveling around the major arc to the station.

W_3 Surface waves returning to the station after a complete circuit of the globe.

F End of discernible movement.

7. *Nature of the motion and deduced data:*

i Impulsive and sharply defined beginning of a phase.

e Poorly defined emergence of a phase.

( ) Parentheses in "seconds" column of arrival times indicate that the time is doubtful.

A Either (a) amplitude of earth motion measured from the position of equilibrium in microns (1 micron is 0.001 mm.), + toward the north, east, or zenith, −toward the south, west, or nadir; or (b), when so marked, the trace amplitude or half range of movement of the pen on the record measured from the median line in millimeters.

A_E E-W component of A.

A_N N-S component of A.

A_Z Vertical component of A.

T Period of wave.

H Time of earthquake at focus, generally expressed in G. C. T.

Δ Arcual distance from station to epicenter.

8. *Constants of seismograph:*

T_o or T_1 Free or undamped period of the seismograph.

T_g or T_2 Free or undamped period of the galvanometer.

V Static magnification.

ϵ Ratio of successive damped amplitudes of pendulum decay curve.

r Half width of zone within which friction will completely arrest the recording pen.

h Altitude of observatory above sea level. (Not to be confused with the damping coefficient used in equation for forced vibration of a pendulum.)

EXAMPLES OF SEISMOGRAM INTERPRETATIONS

The seismogram illustrations in figures 11, 16, and 17 will give only a moderately satisfactory idea of the problem of identifying phases on seismograms that are in accord with accepted travel-time charts. The most the reader may expect to gain from a study

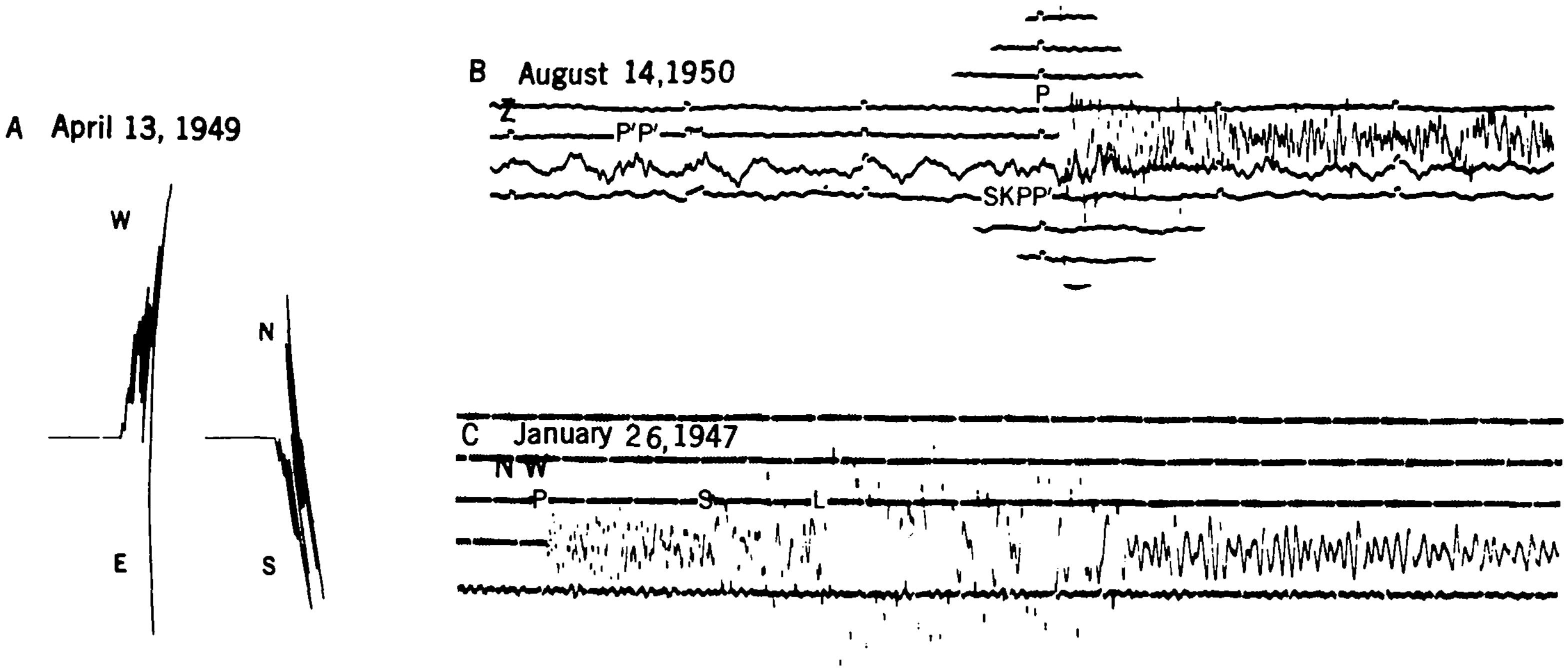

FIGURE 11.—Seismograms from Seattle, Bermuda, and Tucson. See pp. 47 and 48 for discussion.

of these records will be emphasis on the tremendous influence of pendulum period on the appearance of a record, and the fact that sometimes the more obscure dependent phases will show up well and sometimes they will not. In the present state of seismology there is no substitute for actual experience in studying scores of seismograms if one wishes to familiarize himself with them and make an accurate appraisal of interpretation techniques. As stated elsewhere the primary objective in reading seismograms is to report what is on the record, and not attempt to find obscure and perhaps doubtful activities that will fit some theoretical phase on a travel-time chart.

The following are the approximate instrumental constants of the instruments whose sample records are discussed in this section:

Bermuda.—Two horizontal Milne-Shaw seismographs have periods of 12 seconds and lever magnification of 250.

Butte, Mont.—The Benioff short-period vertical (Z–SP) seismometer period (T_o) is 1.1 seconds and the galvanometer period (T_g) is 0.5 second. The peak magnification, near the T_o, is about 50,000.

College, Alaska.—The Wenner N-S horizontal component has a T_o of 9.3 seconds and a T_g of 16 seconds. The magnification is perhaps 1,000; the magnification curve is rather broad with a peak between the above constants.

Eureka, Nev.—The Benioff short-period vertical has a peak magnification of 200,000 at 0.7 second.

Honolulu, Hawaii.—The Houston Technical Laboratory (HTL) short-period vertical with a peak of 0.5 second has a magnification of 20,000.

Philadelphia, Pa.—The two Wenner horizontals have periods of 9 seconds and magnification of 1,400. The galvanometer periods are about one-fourth of the College Wenner.

Tucson, Ariz.—The Benioff short-period vertical has a 1.0 second period with a peak magnification near 50,000. A 77-second galvanometer operating in conjunction with the same pendulum registers long-period results, the peak magnification being on the order of 2,000. A Benioff horizontal (E-W) seismometer with a galvanometer period of 1.6 seconds has a probable magnification of 3,500.

The magnitudes of the teleseismic shocks treated in this section, unless otherwise stated, are about 6–6½ on the Gutenberg-Richter magnitude scale. All times are G.C.T.

Figure 11. A–April 13, 1949. Seattle, Wash. (University of Washington) record from Bosch-Omori seismograph magnifying 25 times on smoked paper. This is a good illustration of the value of insensitive seismographs in recording strong local shocks. The epicenter, 47.1° N., 122.7° W., was based on reports from many stations and is about 70 km. S. 30° W. of the seismograph station. The first impulsive motions on the original Seattle records (2.4 mm. up on the E-W component and 3.7 mm. down on the N-S) indicate an origin either S. 33° W. of the station or N. 33° E. of it since the southwesterly motion of the ground could be either a rarefactional wave from the southwest or a compressional wave from the northeast. A vertical-motion record would have indicated a downward (rarefactional) ground impulse since the origin is known to be southwest of the station.

The $S–P$ time interval may be read as 13 seconds (on the E-W component) and there is also an $L–P$ or a second $S–P$ interval of 21 seconds. If normal P and S velocities in continental structure are assumed, respectively 6.0 and 3.5 km/sec., the distance according to Omori's formula would be 110 km. Since the epicentral distance is known to be only about 70 km., and since neither the local velocities nor crustal structure is known, it is seen that one could only cautiously surmise from the Seattle data alone that the focal distance was of the order of 100 km. (63 miles) and that the epicentral distance could be considerably less than 100 km. depending upon the depth of focus. In this case a deeper than normal focus is indicated by the data obtained from distant stations. It is important that station directors report the *S–P* or *L–P* time intervals because they assist materially in locating the epicenter and focal depth.

Figure 11. B–August 14, 1950. A Tucson long-period Benioff vertical record of a northern Argentina earthquake, 27° S., 62½° W. Focal depth about 700 km. Distance 8,250 km. H=22h 51m 28s G.C.T. The record illustrates the prominent *P* waves associated with deep-focus earthquakes and shows the character of *P′P′* and *SKPP′* waves as registered on a long-period system.

Figure 11. C–January 26, 1947. A Bermuda Milne-Shaw horizontal component (northwest-southeast) record of a western Nicaragua earthquake, 13° N., 86.5° W. Normal depth. Distance 3,100 km. H=10h 06.7m G.C.T. T_o=12 seconds. This is a typical record of a moderately distant earthquake with well-defined *P*, *S*, and *L* phases.

Figure 16 (Pocket). A–October 13, 1959. A Tucson Benioff short-period vertical record of an earthquake at Flagstaff, Arizona, 35.5° N., 111.5° W. Normal depth. Distance 320 km. H=08h 15m 00s. The record illustrates modified use of Jeffreys' nomenclature that postulates a two-layered continental structure, namely, *Pn*, *Pg*, etc. (See fig. 5 and nomenclature.)

Figure 16 (Pocket). B–September 12, 1959. A Tucson Benioff short-period vertical record of an earthquake located in the Gulf of California at 27° N., 110½° W. Normal depth. Distance 622 km. H=07h 25m 10s. The record is a typical example of the phase nomenclature as used by Jeffreys, showing *Pn*, *Pg*, *Sn*, *S**, and *Sg*.

Figure 16 (Pocket). C–September 26, 1959. A Tucson Benioff long-period (T_g=77 sec.) vertical instrument gives a typical record for the distance of 1,960 km. Located at 43½° N., 128½° W., off the coast of Oregon. H=08h 20m 51s. The long-period waves are well represented by this record wherein the *P*, *S*, and *L* are denoted.

Figure 16 (Pocket). D–August 29, 1959. The College Wenner horizontal (N-S) record is of an earthquake in the Lake Baikal region of the USSR, 52° N., 106½° E. H=17h 03m 10s. Distance 5,540 km. Magnitude 6½–6¾, Pasadena. From data on short-period vertical records of more distant stations it appears that the focus is slightly deeper than normal. This accounts for the sharp impulsive character of the *PP*, *S*, and *SS* phases. Generally speaking, phases are more impulsive and are more easily recognized with increasing depth.

Figure 16 (Pocket). E–February 23, 1958. Eureka Benioff short-period vertical record of two different deep shocks. The upper record depicts a shock located in the Santiago del Estero Province, Argentina, 27½° S., 63° W. H=08h 14m 48s. Depth about 600 km. Distance 9,210 km. This typical record of a deep shock illustrates the depth interpretation possibilities of the *pP*, as well as the more complex phases associated with deep foci (*SKPP′*, *pPKKP*). The other shock, with a *P* approximately under the *SKPP′* of the Argentina one, occurred in the Bonin Islands, 28½° N., 139½° E. H=09h 12m 20s. Depth about 400 km. Distance 9,144 km. The magnitude of these shocks is slightly above 6.

Figure 16 (Pocket). F–December 10, 1958. The Tucson Benioff records of the New Zealand shock, 37° S., 176½° E., demonstrate the recording characertistics of different instruments. H=07h 02m 59s. Depth about 300 km. Distance 10,750 km. Magnitude 6¾, Pasadena. The short-period vertical record shows the principal longitudinal phases that are detected considering the conditions above. Note the impulsive nature of the wave groups. The short-period horizontal shows the transverse waves as typically recorded on this component. This record, as well as the long-period vertical, shows the characteristic diminution of the surface waves with increasing depth.

Figure 17 (Pocket). G–February 7, 1959. The Philadelphia Wenner horizontals illustrate the directional nature of the particle motion of seismic waves. The earthquake, near the coast of northern Peru, 4° S., 81½° W., is almost due south of the recording station. H=09h 36m 51s. Distance 4,950 km. Magnitude 7¼–7½, Pasadena. Note the larger amplitudes of the P and S waves on the N-S component, and the emphasis of surface waves on the E-W component. The emergence of the ScP on the N-S record strongly implies a deeper than normal focus.

Figure 17 (Pocket). H–July 20, 1959. The Eureka short-period Benioff vertical record is of an earthquake in the Java Sea, 6° S., 111° E. H=02h 41m 04s. Distance 13,900 km. Depth about 500 km. At this distance the $SKKP$, as well as the $PKKP$, is well recorded on sensitive short-period vertical instruments. Note also the low amplitude, rather uniform-appearing preliminary group of the PKP (P'). Gutenberg and others call this $PKIKP$, assuming that this slightly higher velocity wave has penetrated the higher velocity medium of the inner core. The impulsive character of the waves are accentuated owing to the depth of the focus.

Figure 17 (Pocket). I–February 15, 1959. This Eureka short-period vertical record contains waves of three earthquakes. Two are P' shocks and the other is within P distance. The P (P=04h 15m 49s) on the record is of an earthquake in the Sinkiang Province, China, 44½° N., 83½° E. H=04h 02m 22s. Distance 10,550 km. The first P' comes about 2½ minutes after the P cited above. The shock was in the Sandwich Islands, 59½° S., 25° W. H=03h 59m 25s. Distance 13,800 km. P', PP, $PKKP$, and $SKKP$ are shown. The second Sandwich Islands shock has a P' at 05h 01m 33s. Location 59½° S., 26° W. H=04h 42m 35s. Distance 13,750 km. The latter shock is about ¼ magnitude larger, and the focus is slightly deeper than normal, which accounts for the more effective emergence of $PKKP$ and $SKKP$.

Figure 17 (Pocket). J–April 10, 1957. The Honolulu HTL short-period vertical record illustrates the waterborne T wave of a shock near Kodiak Island, 56° N., 154° W. H=11h 29m 58s. Distance 3,870 km. Magnitude 7.1, Pasadena.

Figure 17 (Pocket). K–November 6, 1959. The Butte short-period vertical record is of a chemical explosion located about 170 km. from the station. Note the compressional first motion.

Figure 17 (Pocket). L–May 11, 1958. The Eureka short-period vertical record shows a typical longitudinal wave of a distance nuclear blast from the Marshall Islands. H=17h 50m 00s. Distance 8,200 km. Note the compressional first motion.

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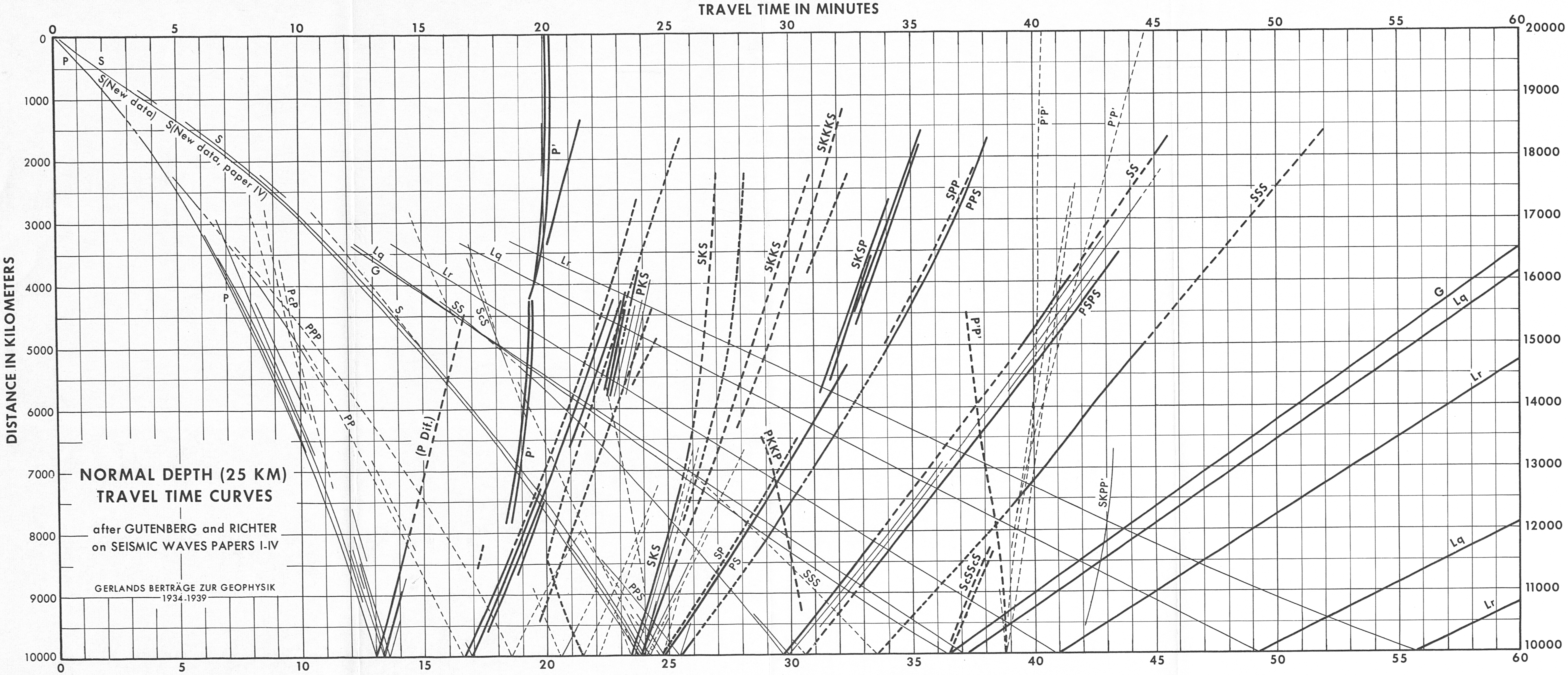

FIGURE 12.—Gutenberg-Richter travel-time chart for normal focal depths.

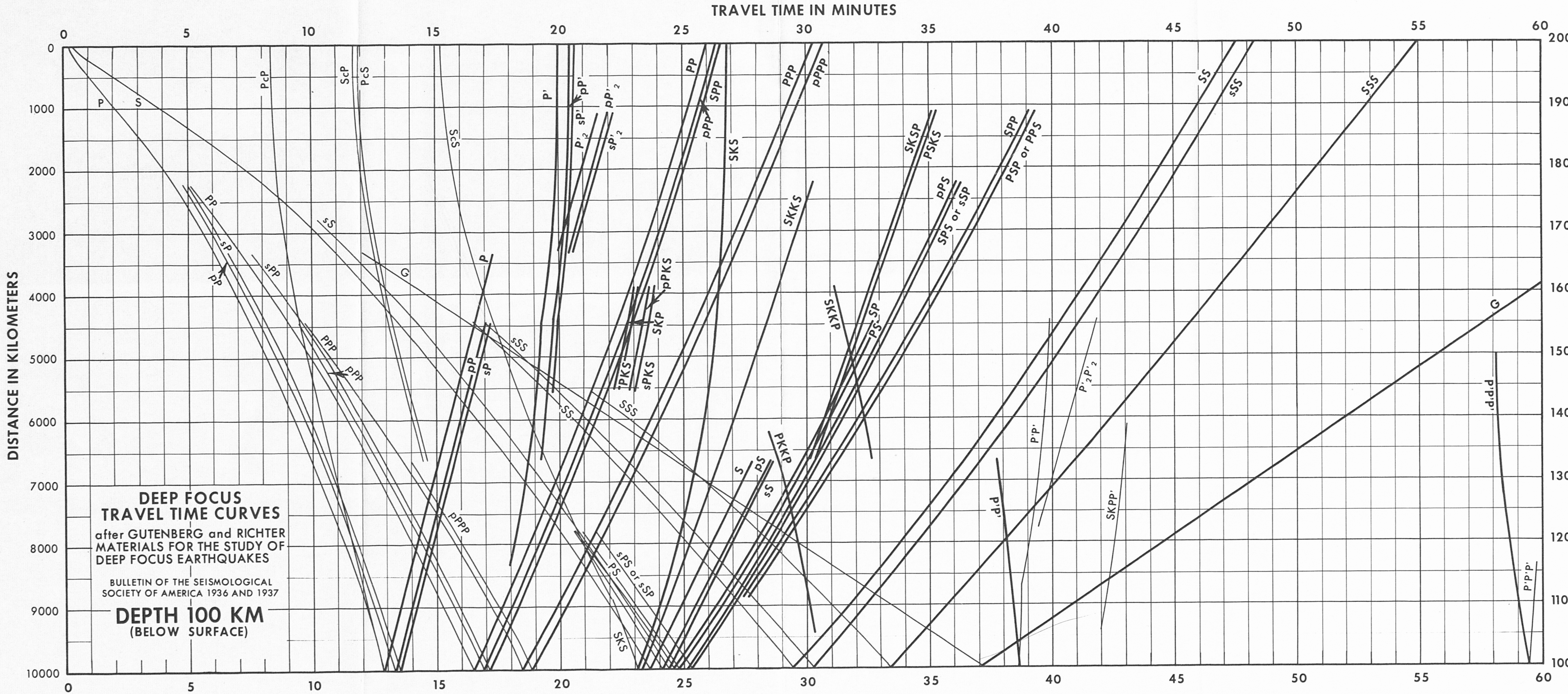

FIGURE 13.—Gutenberg-Richter travel-time chart for 100 km. focal depth.

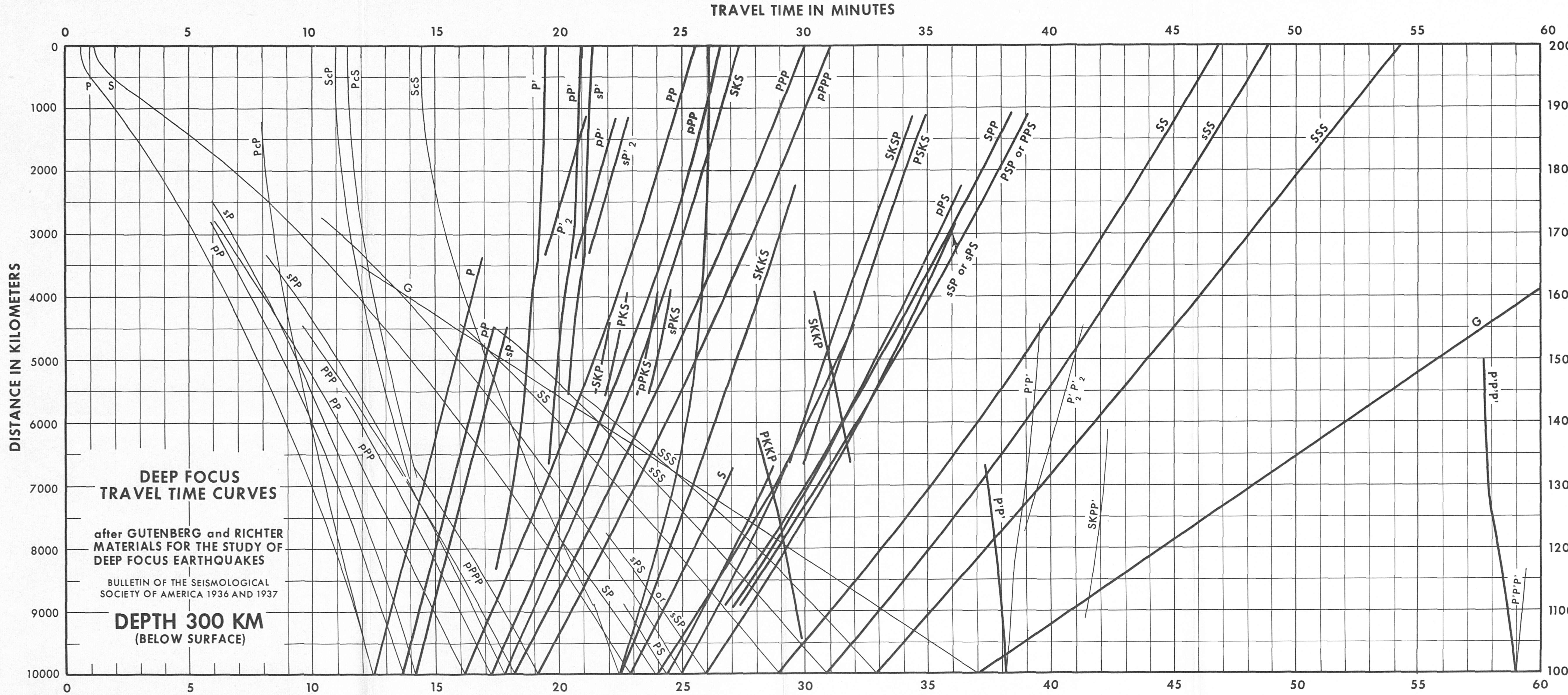

FIGURE 14.—Gutenberg-Richter travel-time chart for 300 km. focal depth.

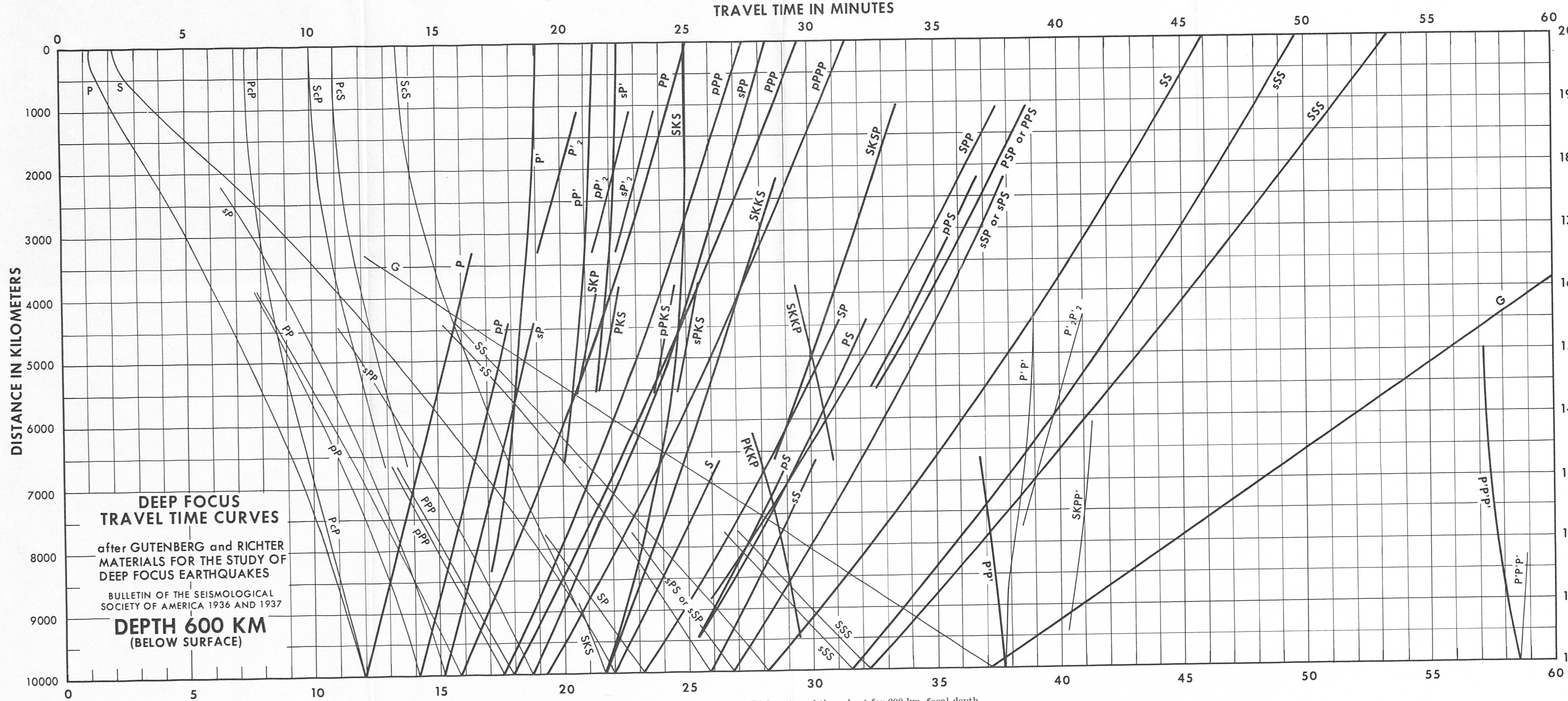

FIGURE 15.—Gutenberg-Richter travel-time chart for 600 km. focal depth.

Figure 16.—Seismograms from College, Eureka, and Tucson.

FIGURE 17.—Seismograms from Butte, Eureka, Honolulu, and Philadelphia.