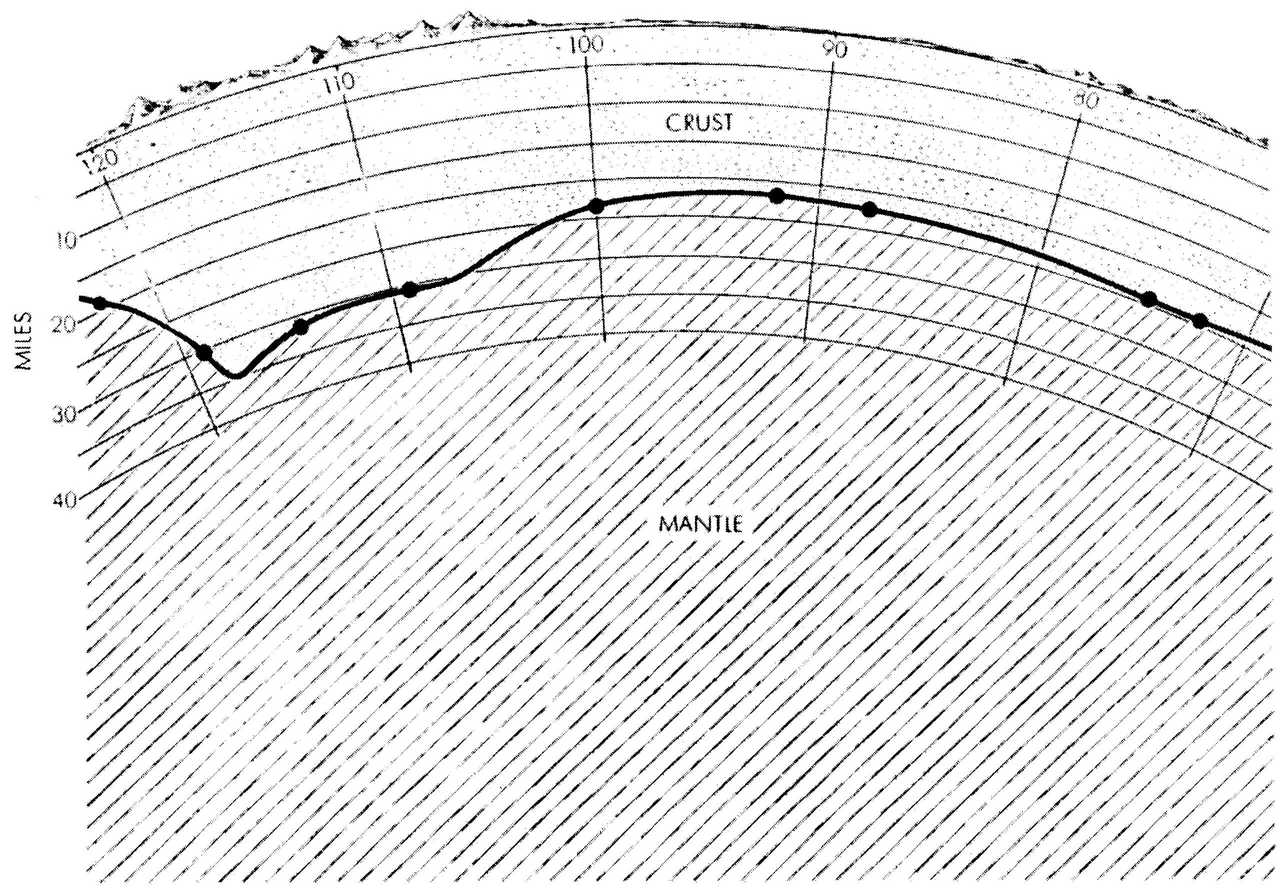
LONG EARTHQUAKE WAVES

Seismologists are tuning their instruments to record earth motions with periods of one minute to an hour and amplitudes of less than .01 inch. These waves tell much about the earth's crust and mantle

by Jack Oliver

The hi-fi enthusiast who has struggled to improve the response of his equipment in the bass range at around 20 cycles per second will have a fellow-feeling for the seismologist who is attempting to tune his instruments to the longest earthquake waves. But where hi-fi deals in cycles per second, seismology measures its frequencies in seconds per cycle. A relatively short earthquake wave has a duration of 10 seconds. At present the most informative long waves have periods of 15 to 75 seconds from crest to crest. With more advanced instruments seismologists hope soon to study 450-second waves. The longest so far detected had a period of 3,400 seconds—nearly an hour! Since these long earthquake waves travel at .9 to 2.8 miles per second, they range in length from 10 miles up to the 8,000-mile length of the earth's diameter. Their amplitude, however, is of an entirely different order: one of these waves displaces a point on the surface of the earth at some distance from the shock by no more than a hundredth of an inch, even when excited by



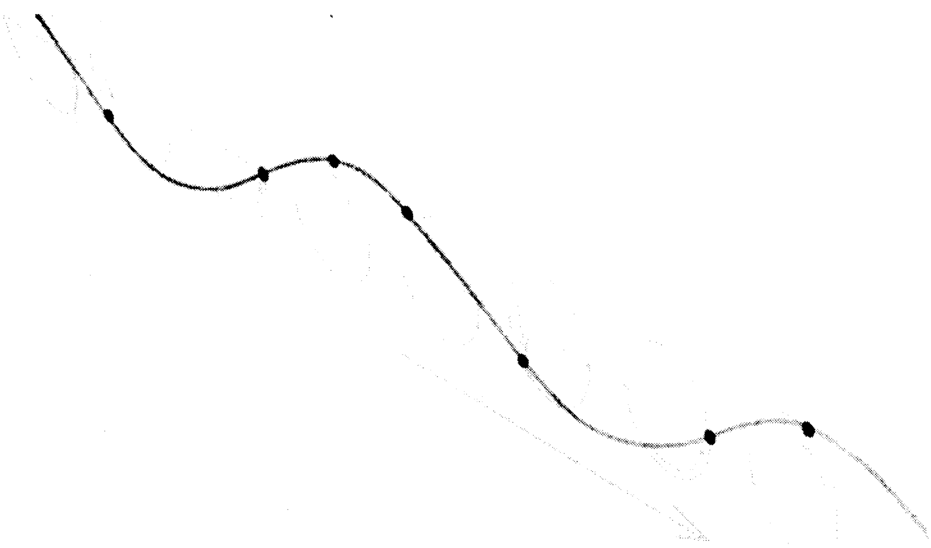
SUBTERRANEAN TOPOGRAPHY of the North American continent is being plotted by study of surface earthquake waves, which reveal boundary between the crust and the mantle. The heavy black curve traces this boundary. The dots on the curve are the points at which the depth of the boundary was established. The surface of the earth is curved in the scale of the diagram. The vertical scale of the topography and of the crust and mantle below, however, are exaggerated 20 times. The numbers at top are degrees of longitude.

earthquakes of the greatest magnitude. To measure such a small motion that takes such a long time presents one of the most challenging tasks of instrumentation to be encountered anywhere in modern science.

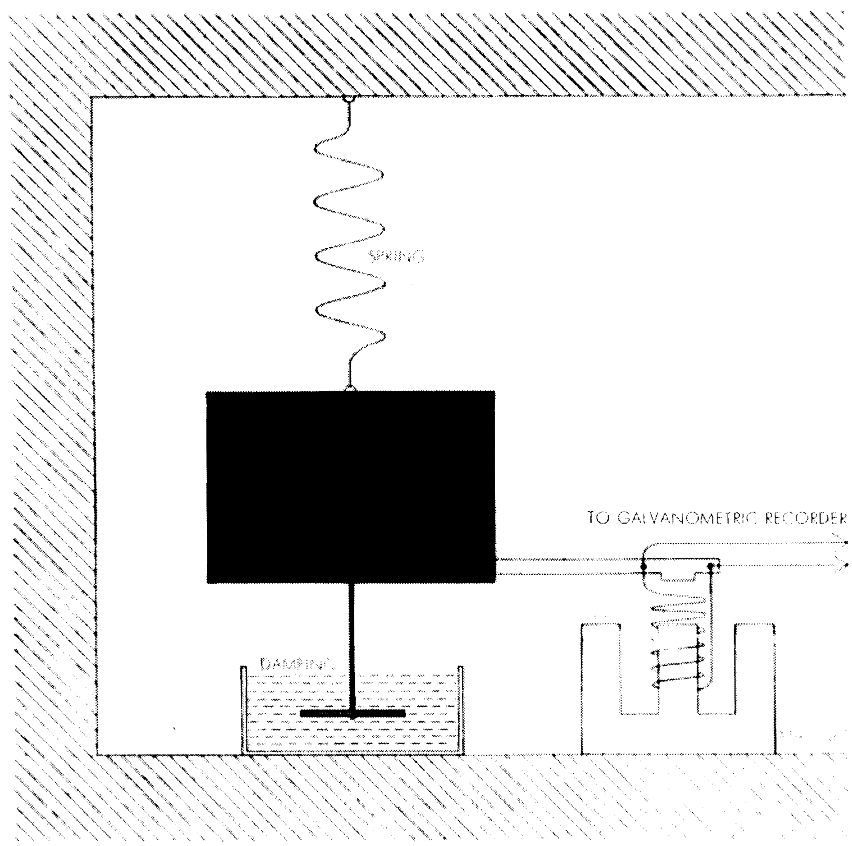
The seismologist has important reasons for going to all this trouble—in contrast, perhaps, to the true votary of hi-fi who regards frequencies as ends in themselves. From the shorter-period seismic waves that travel through the interior of the earth we have developed a remarkably full picture of the structure of our planet. These "body waves" have distinguished the concentric spheres of the earth's crust, mantle and core, and have given us a measure of the densities and the states of matter that prevail under the ascending pressures toward the earth's center. Now the long waves are adding a new dimension to our knowledge. They are surface waves, analogous to the ripples on the surface of a pond, which radiate from the center of a disturbance. Although only a few observatories are equipped to record them well, the long waves have already helped to establish one of the most significant facts we know about our planet. Under the continents the earth's crust reaches down to depths of 25 miles; under the oceans the crust is but three or four miles thick. Thus the boundary between crust and mantle, the so-called Mohorovicic discontinuity, lies only 9 or 10 miles below the surface of the ocean, and the "inside" of the earth is closer to the surface than we had thought [*see illustration on page 137*].

So far as we know all major earthquakes are associated with the rupturing or faulting of the rock in the crust or mantle. Except in the very strongest shocks the rupturing process takes but a few seconds and occurs along a fault line only 10 or 20 miles in length. The original disturbance is limited in space and time, yet it excites vibrations in a wide spectrum of wavelengths throughout the entire earth. The body waves, which penetrate deep into the earth and return to the surface bearing information about the interior, are of two types. One is the compression (P) wave, in which the particle motion is in the direction of the wave; the other is the shear (S) wave, in which the motion is at right angles to the line of travel. The fact that the core does not transmit this second wave constitutes the principal evidence for the idea that its structure is nonrigid or "liquid."

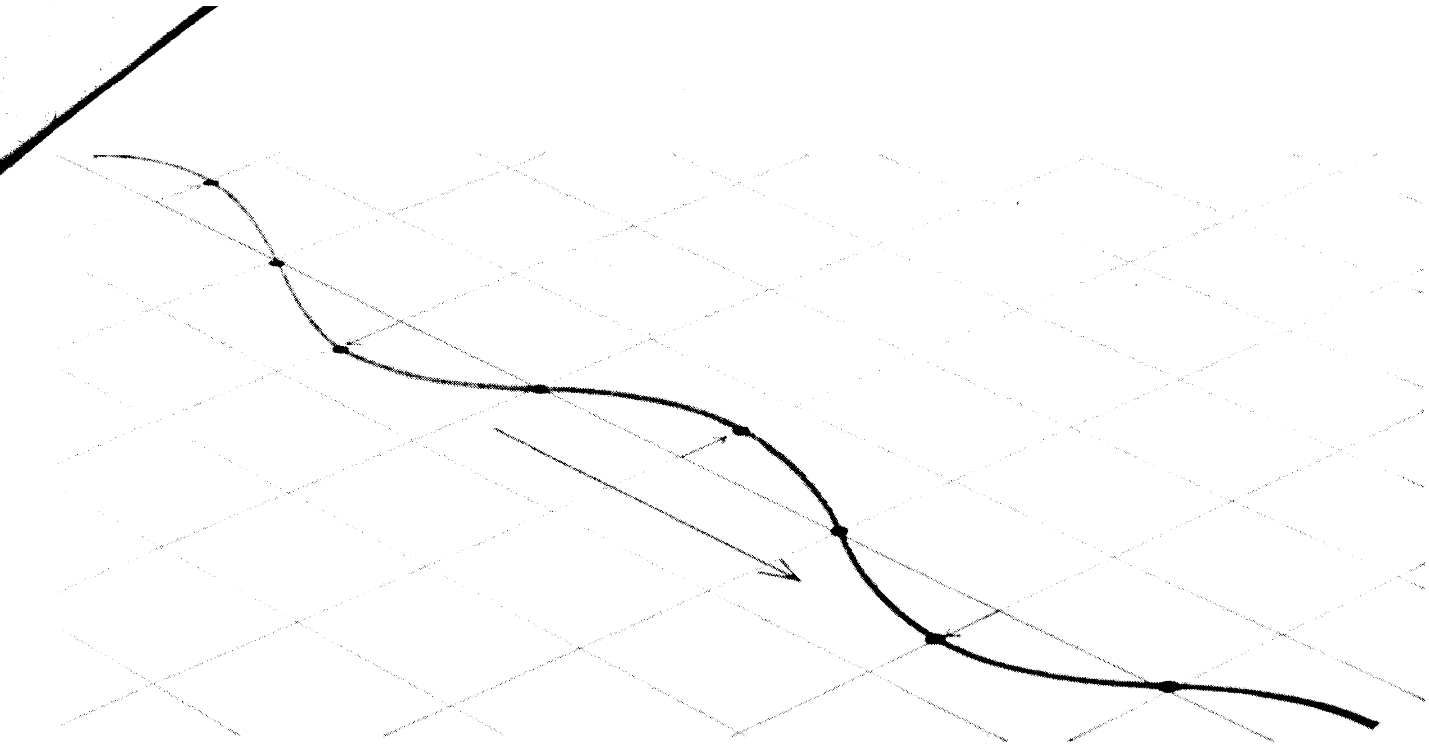
The surface waves are also of two



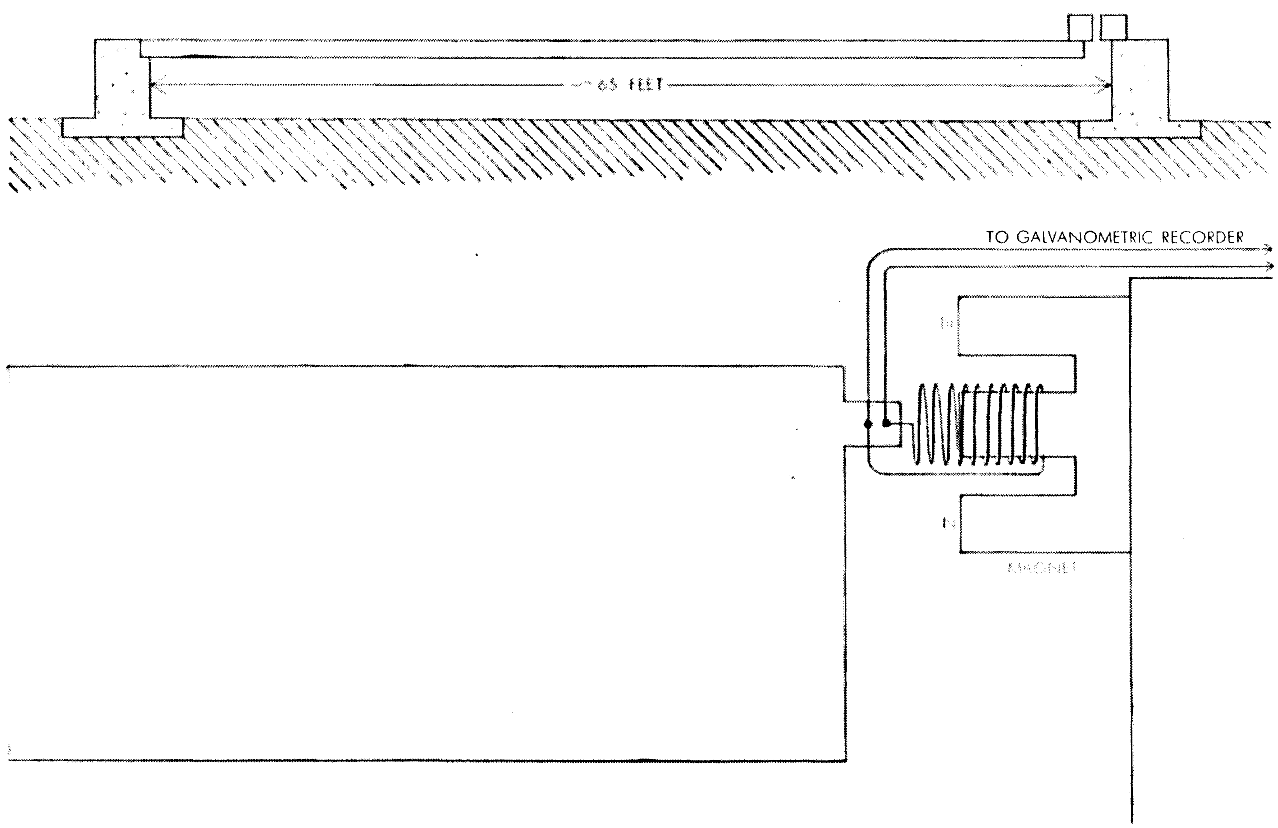
SEISMIC SURFACE WAVES are of two kinds, the Rayleigh wave (*above*) and the Love wave (*opposite page*). The Rayleigh wave moves in the vertical plane like a ripple, as shown



SEISMOMETERS are of two types, the inertial seismometer (*above*) and the strain-gauge seismometer (*opposite page*). The inertial instrument measures the motion of the earth with



by the gray line. The black dots on the ellipses show the retrograde motion of any given surface particle in the path of the wave. In the Love wave the motion of a particle is in the horizontal plane and at right angles to the direction in which the wave is propagated.

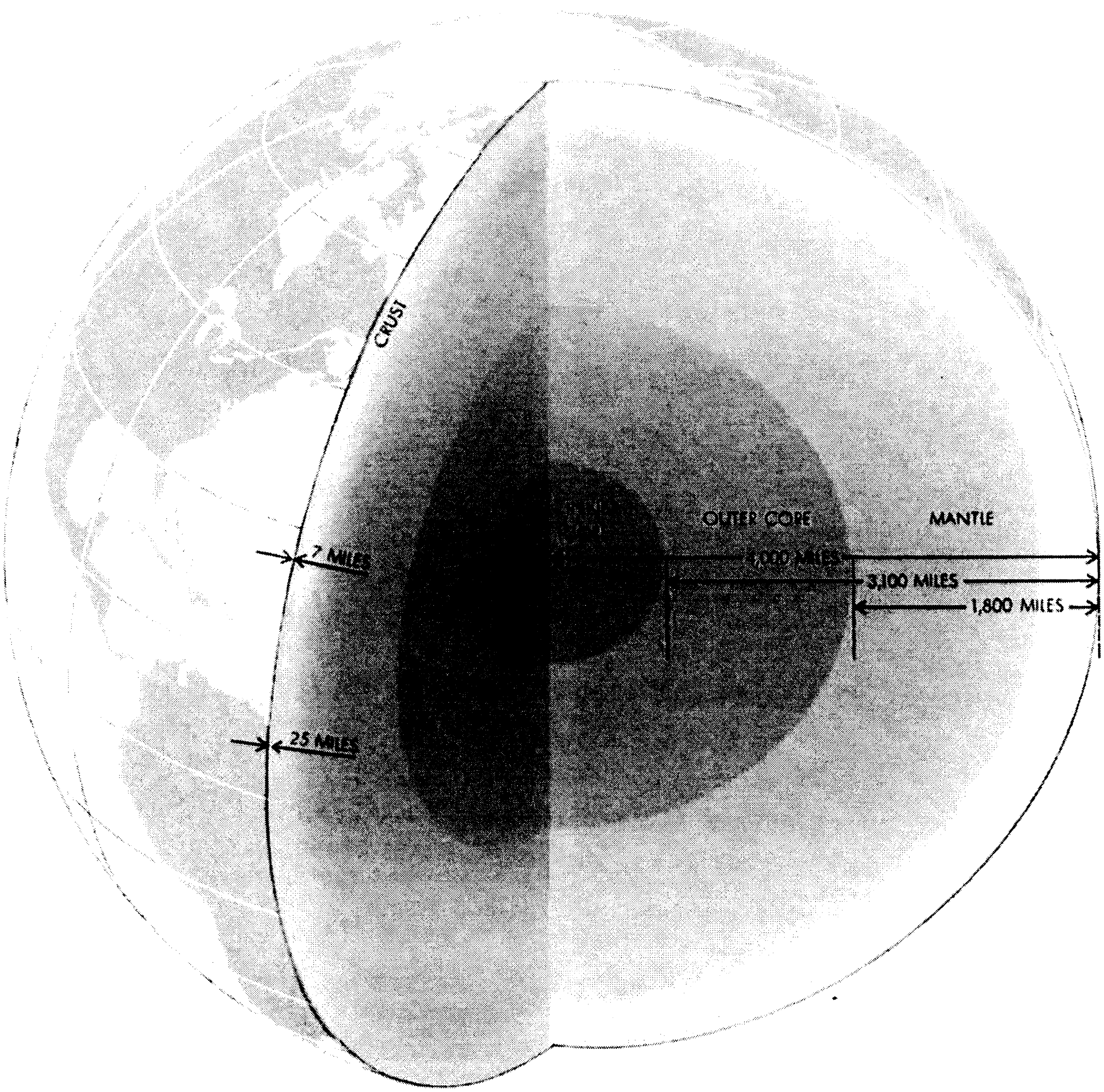


respect to a large mass (*black rectangle*) suspended from a weak spring. The strain seismometer measures the motion of one pier set in the earth with respect to the end of a rigid bar attached to a second pier that is fixed in the earth some distance from the first.

kinds. During the passage of the Love wave (named for its 19th-century discoverer A. E. H. Love) the ground vibrates horizontally at right angles to the direction of the wave. The Rayleigh wave (named for Lord Rayleigh, who described it in 1900) more closely resembles a ripple in water; when it passes a point on the surface, the point moves around an ellipse in a direction parallel but opposite to the direction of the wave [*see illustration at top of preceding two pages*]. Though they move through the earth simultaneously, the two waves are not associated in any way. On the seismograph they record complementary but different information.

The surface waves that reach the seismological observatory are characteristically long because the shorter-period components of the original disturbance tend to die out on the way. It is not known for sure whether body waves contain components of longer period. In any case, it was around the body waves of shorter period that most of the instrumentation of seismology was designed.

The basic instrument is a mass suspended from a weak spring [*see illustration at the bottom of page 132*]. With different suspensions the mass may be used to measure either the vertical or the horizontal component of a seismic wave. Because of its inertia the mass tends to remain fixed in space; when a seismic waves passes, the loose coupling

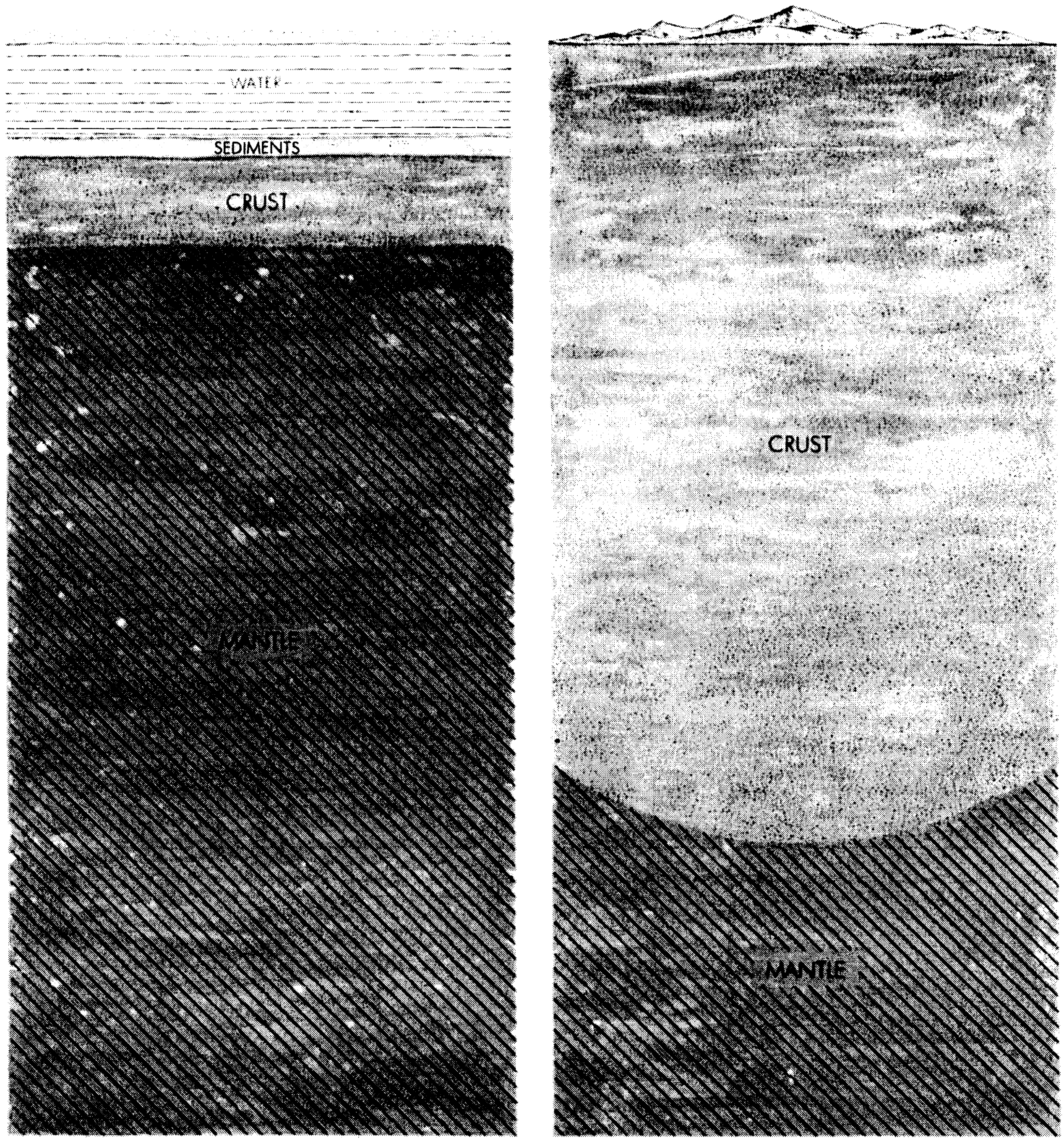


INTERNAL STRUCTURE OF THE EARTH is revealed by earthquake waves. Body waves penetrating deep into the earth have resolved the discontinuities in elasticity which differentiate the inner core, the outer core, the mantle and the crust. Surface waves have now established that the boundary between the mantle and crust lies much deeper below the continents than below the sea.

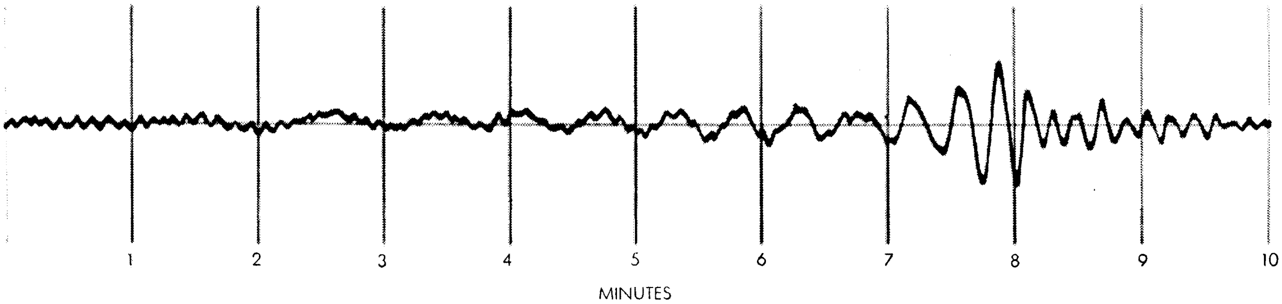
of the spring permits the earth to move with respect to the mass and insulates the mass from the disturbance. Measurement of the movement of the earth relative to the mass reveals the motion of the earth at the station.

Unfortunately the inertial seismometer works well only for those waves that are shorter than its own natural frequency, or free period of oscillation. It is difficult to construct an instrument with a free period greater than 15 seconds, especially for the measurement of the vertical component of ground motion. Recently ingenious new ways of suspending the inertial mass and damping its motion have doubled this figure, and a few test instruments have operated with a period of oscillation as great as 100 seconds.

The poor response of the seismometer at longer periods may be offset by using a device to detect the relative motion of the earth that favors those periods. The usual instrument is a galvanometer, which is coupled to the seismometer by a coil mounted on the mass and operating in the field of a magnet fixed to the



THICKNESS OF THE CRUST has been shown by surface waves to vary sharply as between the oceans and the continents. The boundary between the mantle and the crust lies only seven miles below sea level; subtracting average depth of the ocean and sediments the suboceanic crust is only 3.5 miles thick. Crust below continents goes down some 25 miles, plunging deeper under mountains.

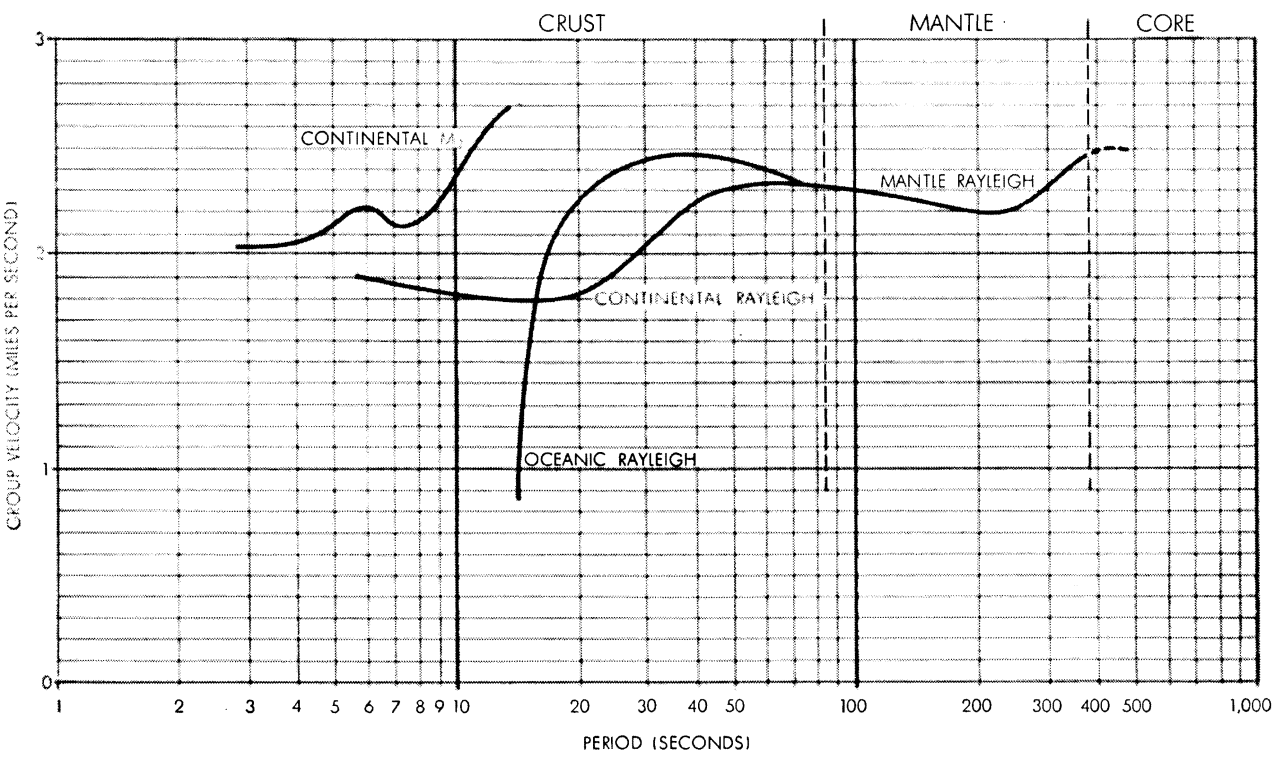


TRAIN OF SEISMIC WAVES, as recorded on seismometer, shows how variation in velocity disperses waves excited in same instant by earthquake. Longer waves (disregard high-frequency "noise") register in first minutes (*left*); shorter waves come minutes later.

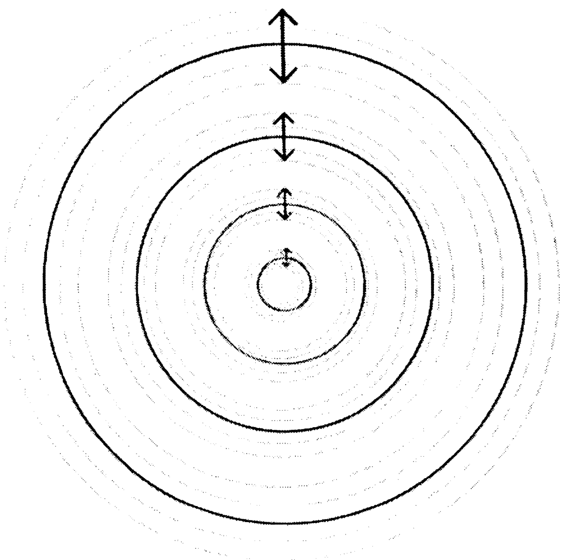
earth [*see illustration at bottom of page 132*]. In the galvanometer the current generated by the earth motion energizes an electromagnet which then rotates because it is suspended in the field of a permanent magnet. Here again large inertias, weak coupling and damping can achieve a long-period response. At the Lamont Geological Observatory of Columbia University an instrument designed by Maurice Ewing and Frank Press successfully combines a 15-second seismometer with a 75-second galvanometer; it has recorded seismic waves of 480 seconds' duration. The observatory now has 30-second and 100-second combinations in routine operation, awaiting signals from the next major earthquake.

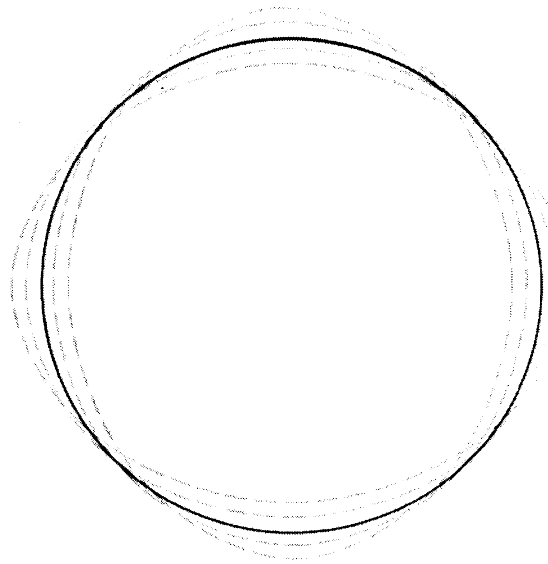
An entirely different type of instrument is the strain seismometer, developed by Hugo Benioff of the California Institute of Technology. This instrument measures the change in distance (the earth "strain") between a pier set in the earth and the free end of a long rigid rod attached to another pier set in the earth at a distance of 50 to 100 feet. Inherently more responsive to long waves, its performance may also be improved by a long-period galvanometer. The strain seismometer has traced waves of 300 seconds on several occasions, and has evidenced its great potential in registering the record 3,400-second wave. At present only a few stations have strain seismometers, and they are equipped to measure only the horizontal components of earth strain. A pair of strain seismometers arrayed at right angles to each other will produce a full record of a Love wave but will register only the horizontal motion set up in the earth by a Rayleigh wave.

Waves of different length become vehicles for information about the rock through which they pass because



RAYLEIGH-WAVE VELOCITIES (*vertical coordinate*) show variation in character of rock below the surface. Waves of greater length (*horizontal coordinate*) "feel" the rock at greater depth. The higher velocity of the shorter and shallower oceanic waves, of periods from 75 down to 15 seconds, shows that higher-density mantle rock comes closer to the surface. The lower velocity of continental surface waves starting at 75 seconds, and hence at great depth, reflects the lesser density of the thick layer of continental crustal rock.





FREE OSCILLATIONS of the earth in its natural frequencies of long period may set up standing waves of several forms. In the radial mode (*left*) the motion of particles is along the radii of the earth in a series of compression waves resembling sound waves. In the "football" mode (*right*) the earth alternately assumes the prolate (long axis vertical) and oblate (long axis horizontal) form.

they travel at different speeds and arrive at different times. An earthquake liberates waves of a wide range of periods at about the same time. Though the Love and Rayleigh waves are confined to the surface, they probe—in a sense, "feel"—a depth that depends upon their length. Since the elastic properties of the earth generally increase with depth, the waves of greater length travel faster, often more than twice as fast, as the shorter, shallower and slower waves. The seismograph station some distance from the quake thus records not a sudden impulsive disturbance, but a long train of waves, sometimes lasting several minutes and sometimes several hours, depending upon the paths the waves have severally traversed. The long-period waves, traveling quickest, arrive earliest. A seismogram recording a typical dispersed wave-train is reproduced at top of page 138.

The record of the arrival times of waves of different length, plus knowledge of the time and location of the earthquake, make it possible to compute the individual and average velocity of waves along the great-circle path between the quake and the station. The velocities provide a measure of the elasticity and, for practical purposes, the density of the rock. Thus seismic-wave spectroscopy yields an analysis of the earth's interior structure.

From observation of a number of earthquakes Ewing and his collaborators have charted the relation between wavelength, depth and velocity for Rayleigh waves of periods from 3 to 480 seconds [*see illustration at bottom of page 138*]. The chart shows a sharp divergence between the waves of 75 seconds or less duration that arrived at the Lamont Observatory, near New York City, via oceanic and continental pathways. From periods of 75 seconds down to 15 seconds the velocity of the oceanic waves maintains a high level and then drops off to an extremely low value. This curve shows us that mantle rock of relatively high elasticity, through which waves travel at high velocity, lies at a very shallow depth: only 10 miles below the surface of the ocean. The abrupt fall in the curve at 15 seconds marks the transition to the less elastic rock of the crust and shows the crust to be only three or four miles thick. Then, because compression waves in the ocean water and unconsolidated bottom sediments load the vertical component of the Rayleigh waves, the velocity decays to extremely low values. The slope of the curve in this region gives three miles as the combined depth of the water and the sediments. Subtracting the average depth of the ocean obtained from soundings, we find that the sediments range from a quarter to a half mile in thickness.

The velocity-period curve for waves traveling across continents falls below the ocean curve at 75 seconds. This curve tells us that the lower-velocity crustal rock must go down to great depths beneath the surface of the continents. The falling off of the continental velocity curve at 75 seconds places the boundary between mantle and crust—the Mohorovicic discontinuity—at a depth of 20 to 25 miles. This substantial contrast in the thickness of the crust as between the continents and the oceans had been indicated by explosion soundings at a few discrete points at sea and on land. Surface-wave records during the past few years have now established that the crust beneath the oceans is uniformly thin.

Though surface-wave seismology rests upon techniques for measurement of long waves, it is apparent that the shorter surface waves, which "feel" the rock at shallower depths, hold great interest for us. They can tell us much of what we want to know about the crust and upper mantle. Moreover, seismic waves, like waves of all other kinds, increase in resolving power with decrease in length; they can "see" structures of dimensions comparable to their own length. They should thus deliver quite fine-grained information, for example, on variations in the density of continental rock. Unfortunately their very capacity to resolve structural detail causes them to be scattered and absorbed when they encounter such details. The earth, in effect, filters out the short waves and transmits the long. However, just as an organ pipe or violin string may vibrate in harmonics as well as in a fundamental frequency, so may the surface waves. Such higher-mode, shorter-period harmonics accompany the longer waves on continental pathways. Their resolving power makes surface waves a potent medium for deduction of earth structure.

Love waves convey information in almost all respects parallel to that supplied by Rayleigh waves. Since their motion is restricted to the horizontal plane, however, they are not affected by the liquid layer of the ocean. In consequence they do not fall to the extremely low velocities attained by the suboceanic Rayleigh waves.

Surface waves with periods of 75 seconds or more are so long that they cannot "see" or "feel" such minor variations in the crust as the difference between oceans and continents. They derive their velocities primarily from the earth's mantle and may even be affected by the core. When better instrumentation is available, the long Rayleigh waves in particular may give us new information on the deep interior not available from body waves.

The long waves of both types have another useful quality in their resistance to attenuation and decay. Actually waves of all periods attenuate at the same rate, if attenuation is measured in loss per wavelength. Translated into distance this means that the very long waves can travel to much greater distances. Nearly 24 hours after the great Kamchatka earthquake of 1952 the Lamont Observatory was still able to record long waves that had traveled around the earth more than seven times —a total of 187,000 miles.

Waves of such length and longevity begin to verge on the dimensions of another class of wave: the free oscillations of the earth itself. In contrast to earthquake waves, which are propagated from centers of disturbance and travel to distant points, the earth's free oscillations are standing waves. By analogy, a long pipe between two floors in a building may be used as a speaking tube to carry traveling waves, provided all the waves are short with respect to the length of the tube. With wavelengths increased to a length comparable to that of the tube—that is, "tuned" to the tube —the sound waves become standing waves, and the tube becomes a whistle or an organ pipe. This is the nature of the 3,400-second wave detected by the strain seismometer at Cal Tech. That wave was excited by a tremendous earthquake. For regular observation of the free oscillations of lower amplitude excited by more frequent smaller earthquakes it will be necessary to make certain improvements in our instruments.

Though experimental data are scanty, the earth's free oscillations have intrigued mathematicians since the early 19th century. As a result a considerable body of theory is ready for the test of observation when that becomes possible. Theoreticians have usually considered a simple homogeneous sphere undergoing free vibration due to elastic and gravitational forces. They have shown that three major types of vibration are possible: a rotatory or torsional vibration in which motion of any point is always in the horizontal plane; a radial mode in which the sphere dilates and contracts uniformly; and a spheroidal mode in which the sphere oscillates from a prolate to an oblate spheroid, assuming the shape of a football during a part of the oscillation. For each type of oscillation there is an infinity of higher modes of increasing complexity.

Recently electronic calculators were put to work on the computation of the length of the vibration periods for the three long-period modes, based upon simplified data for a real earth with varying composition and elasticity. The calculator yielded an estimate of 42 minutes 30 seconds for the torsional mode, 20 minutes 44 seconds for the radial mode, and 56 minutes 44 seconds for the "football" mode. These figures are in close agreement with the consensus of calculations for purely theoretical spheres. Love himself had calculated the football mode at almost exactly one hour for a sphere of the earth's dimensions and the rigidity of steel.

Both calculations for the football mode come well within the limit of observational error for the longest-recorded, 3,400-second wave (56 minutes 40 seconds)! In this case, it happens, the close fit of theory and observation is a disappointment. The small difference means that the long waves lack the resolving power to enable us to learn much from them about the details of the earth's interior. We must content ourselves with what they tell us about the average properties of the earth. The higher-mode harmonics of the free oscillations, however, should possess the needed resolving power and promise to open a new field of investigation.

The hi-fi seismologist has good reason, therefore, for tuning in on the bass scale of the earth's oscillations. Lest anyone think that we are content with periods of one hour, let me say that space in the seismograph vault at Lamont Observatory is available immediately for an instrument responsive to earth motion of even greater periods.