

CURRENT PROBLEMS IN RESEARCH

Low-Velocity Layers in the Earth, Ocean, and Atmosphere

These layers increase the difficulty of locating buried explosions and may cause sonar booms.

Beno Gutenberg

The most important "law" pertaining to the propagation of elastic waves (sometimes called sound waves) through solids, liquids, or gases is Fermat's principle. This states that the waves follow a path which makes their transmission time between two points a minimum. From this principle it follows that, along a given ray in the earth,

$$(r \sin i)/V$$

is constant, where r is the distance of a given point from the earth's center, V is the wave velocity at the point, and i is the angle between ray and radius. If the velocity V increases with depth, i also increases, and the ray is concave downward (Fig. 1a); similarly, if V decreases with depth, the ray is concave upward. In this case there are two possibilities. If in a layer the rate of decrease dV/dr is smaller than V/r , the radius of curvature of the ray is larger than that of the sphere through the point, and the ray returns to the surface (Fig. 1b); if the rate is larger, the ray goes downward (Fig. 1c). Near the earth's surface, V/r is close to zero.

Combination of layers with increasing and decreasing velocities (Fig. 2) leads to complicated wave paths which may

produce shadow zones. These are entered only by diffracted energy or other types of waves. The time-distance curves (Fig. 2, right) show that direct waves arriving beyond the shadow zone are delayed relative to those arriving at short distances. The shadow zones as well as the time-distance curves change with the depth of the source (H in Fig. 2). If the source is in a layer with relatively low velocity (Fig. 2d), waves leaving the source nearly horizontally may travel continuously up and down without ever reaching the surface and may form a "channel." On the other hand, waves arriving just beyond the shadow zone at the surface concentrate and form a "caustic" (see Fig. 2). In the neighborhood of caustics the arriving waves are especially large. Similar phenomena occur if the decrease in velocity is sudden at a discontinuity.

Since shadow zones are the main result of low-velocity layers, these layers may easily be overlooked, especially if waves arriving slightly later than the direct waves under consideration are incorrectly assumed to be direct waves arriving in the actual shadow zone. Thus, investigations of shadow zones are usually difficult, and frequently there is no proof that a low-velocity layer exists. Consequently, many results concerning low-velocity layers in the earth have been the subject of con-

troversy for years. However, there is no reasonable doubt that there are low-velocity layers in the solid earth, the ocean, and the atmosphere.

The Earth's Core

The portion of the earth first found to be a low-velocity layer was the outer portion of the core. Oldham had found in 1906, and Wiechert in 1907, that longitudinal waves through the deep interior of the earth are delayed, the delay indicating relatively low velocity somewhere deep in the earth. In 1913 I realized that the travel-time curve of these waves has the form shown in Fig. 2, *a* or *b*: There is a shadow zone, starting at a distance slightly more than halfway around the earth from the source, and at the distance where the delayed waves arrive, more than three-quarters of the way around the earth from the source, they are extremely strong, indicating a caustic. On the basis of these observations I concluded in 1913 that the core has a radius of 2150 miles (3470 kilometers) and that the wave velocity decreases at its boundary from 8.2 to 5.3 mi/sec (from 13.15 to 8.5 km/sec); the velocity outside the core boundary is now believed to be about 0.3 mi/sec more, that inside the core boundary about 0.3 mi/sec less; the figure for the radius of the core which is being used at present is the same as the earlier figure, within a few miles.

The relatively great decrease in the wave velocity at the core boundary is believed to be caused by the fact that the "mantle" of the earth surrounding the core is solid, while at least the outer portion of the core is not solid. This had been concluded first from the tides of the earth's body, which are noticeably greater than they would be if the earth were solid and very rigid throughout. The core is much less rigid than the mantle. No transverse waves which had been propagated through the core have been found. The small rigidity of the core results in longitudinal waves of relatively low

The late Dr. Gutenberg was director of the Seismological Laboratory, California Institute of Technology, Pasadena.

velocity. Another explanation for the decrease in the wave velocity at the core boundary is the probable sudden increase in density at the core boundary. Contrary to a frequently held belief, greater density results in lower wave velocity.

Asthenosphere Channel

In 1926 I made an investigation to determine whether the amplitudes of waves through the upper portion of the earth give any indication of whether or not the melting point of the material is reached somewhere near a depth of 50 miles, as had been expected by some volcanologists and geologists. I found that the amplitudes of longitudinal waves arriving at a distance of 1000 miles from the source of earthquakes are only about 1/100 of those arriving at a distance of 100 miles, and that this decrease in amplitude is gradual. At a distance of roughly 1000 miles the amplitudes increase very rapidly, to about the values which they have only 100 miles from the source. This I interpreted as an indication of a wave velocity at a depth of about 50 miles that was relatively low but not low enough to indicate a liquid material. This low-velocity layer and the consequences of its presence have been the subject of investigations ever since.

About the turn of the present century, it was concluded that near a depth of roughly 100 miles below the earth's surface there must be a layer which yields more easily to stresses of secular duration than the stronger crustal layers and permits the slow, gradual movements required to explain various observations. For example, accumulation or removal of loads at the surface of the earth, especially of ice during ice ages, is followed by gradual sinking or uplift, respectively, of the affected areas during the following centuries or even millennia; moreover, except for geologically disturbed areas, the earth's crust is nearly in equilibrium (isostasy) regardless of the elevation of the surface (mountains and oceans). These and other findings have been considered to be an indication that a layer exists below the crust in which a slow flow of the material is possible. This relatively "weak" layer has been called the "asthenosphere" since 1914. I have believed that it coincides with the asthenosphere low-velocity channel indicated by the seismic waves ever since I found the channel in 1926.

Results Based on Deep-Focus Earthquakes

Figure 2 shows that the shadow zone decreases in size as the source H moves toward the bottom of the low-velocity channel (Fig. 2e); no shadow zone should exist if the source is still deeper. Consequently, I decided, jointly with C. F. Richter, that a study of the records of earthquakes to compare the behavior of amplitudes of shocks originating at various depths should give some information about the depth of the channel. We found that in Peru the shadow zone for longitudinal waves disappears when the focal depth of the shocks exceeds about 100 miles. Later, I studied for different regions the amplitudes of waves recorded at various epicentral distances in deep-focus earthquakes in connection with the problem of developing graphs for finding the magnitude of deep shocks. Figure 3 is based on the results of these investigations. It shows that the minimum amplitude of longitudinal waves increases as the focal depth increases, and that, practically, the minimum is absent if the source is at depths in excess of about 150 miles. For transverse waves the shadow zone extends to slightly greater distances but is otherwise similar.

Results Based on Records of California Shocks

In Fig. 4, the beginnings of selected records obtained from the Kern County, California, earthquake of 21 July 1952 are reproduced. The first, recorded at an angular distance of 3.9° (269 miles along the earth's surface) shows a rather large, sharp beginning, produced by a longitudinal wave (a in the figure) which had been refracted (or diffracted) at the Mohorovičić discontinuity, which separates crust and "mantle," but had not penetrated far into the mantle. This wave is much smaller at a distance of 6.3° (435 miles) and has completely disappeared at a distance of 7.4° (511 miles); the

latter point, consequently, is in the shadow zone. It is obvious that a later wave (b) may be mistaken for the missing wave (a), and the shadow zone may be missed. The next two seismograms in Fig. 4 (for distance of 10.1° and 12.9° illustrate a similar situation. The seismogram (Fig. 4, middle) which first shows the relatively large beginning (c) of the delayed second branch near its caustic beyond the shadow zone was recorded at a distance of 15.2° (1050 miles). The last records, from distances of 18.6° and 19.3° (1283 and 1332 miles) respectively, still have rather large beginnings. The explanation of travel-time curves for earthquakes at epicentral distances between about 400 and 2000 miles is still a matter of controversy. However, the low-velocity layer is now very widely considered to be the cause of the complications.

In 1953 I pointed out that the apparent velocity (measured along the earth's surface) which the travel-time curve exhibits at its only point of inflection can be used to determine the velocity at the depth of the source for each given earthquake. Theoretically, this point of inflection corresponds to the ray which leaves the source horizontally, and the corresponding apparent velocity at the surface has its minimum. Even if there is a shadow zone, this minimum apparent velocity can be found with good approximation. If multiplied by the ratio of the radius at the source to the earth's radius, it gives the true wave velocity at the depth of the source.

In 1959 I studied information on the velocities immediately below the crust. The waves used for this purpose correspond to the beginning (a) of the first and second seismograms in Fig. 4. On the basis of results published by seismologists in many regions for shallow earthquakes and for artificial explosions, I found that, under continents, the following relationship exists: the deeper the extension of the crust into the mantle below—that is, the deeper the Mohorovičić discontinuity—the lower the velocity immediately be-

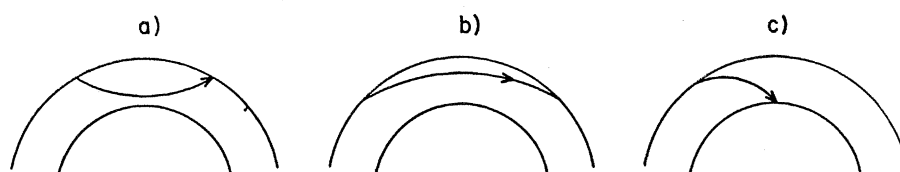


Fig. 1. Sound rays (schematic): (a) if the velocity V increases with depth h ; (b and c) if it decreases; (b) if dV/dh is smaller than V/h ; (c) if dV/dh is greater than V/h .

low the crust. Below continents, this depth varies between about 15 and 40 miles. Consequently, I concluded that immediately below the crust the velocity of both wave types decreases with increasing depth. I found that, on the average, this decrease exceeds the critical value $dV/dr = V/r$ mentioned at the beginning of this article. Thus, under average continents, the asthenosphere low-velocity channel begins at the bottom of the crust. In Fig. 5 my most recent findings for the velocities V of longitudinal, and v of transverse, waves in the upper portion of the mantle are reproduced. They show that the low-velocity channel for transverse waves extends to a greater depth than that for longitudinal waves, and that it has a minimum at a depth of almost 100 miles, while the minimum velocity of longitudinal waves in the earth's mantle is closer to 50 miles.

During the past two years additional information on the wave velocities as a function of depth has been gained from investigations of the dispersion of surface waves. There are two types of such waves, Love waves and Rayleigh waves. Their periods in records of distant earthquakes vary usually between about 10 and more than 100 seconds, and their velocities vary between about 2 and 2.75 mi/sec; consequently, most wavelengths are in the range between roughly 20 and more than 300 miles. Since most of the energy of surface waves is propagated within one wavelength below the earth's surface, the velocity of the longer waves is affected by much deeper portions of the earth's mantle than that of shorter waves. It is possible to calculate the velocities of surface waves if the velocities of longitudinal and transverse waves as a function of depth are given.

Until recently such calculations were very tedious, and not more than three layers could be assumed. With modern computers it is now possible to perform the calculation in a relatively short time, on the assumption of different velocities in 20 or more layers. Several groups of seismologists have performed such calculations of the velocities for both types of surface waves and have found that the calculated and observed dispersion curves agree best if an asthenosphere low-velocity channel of approximately the type given by the curves in Fig. 5 is assumed. This channel seems to be thicker under ocean bottoms than under continents. This results from the fact that the crust,

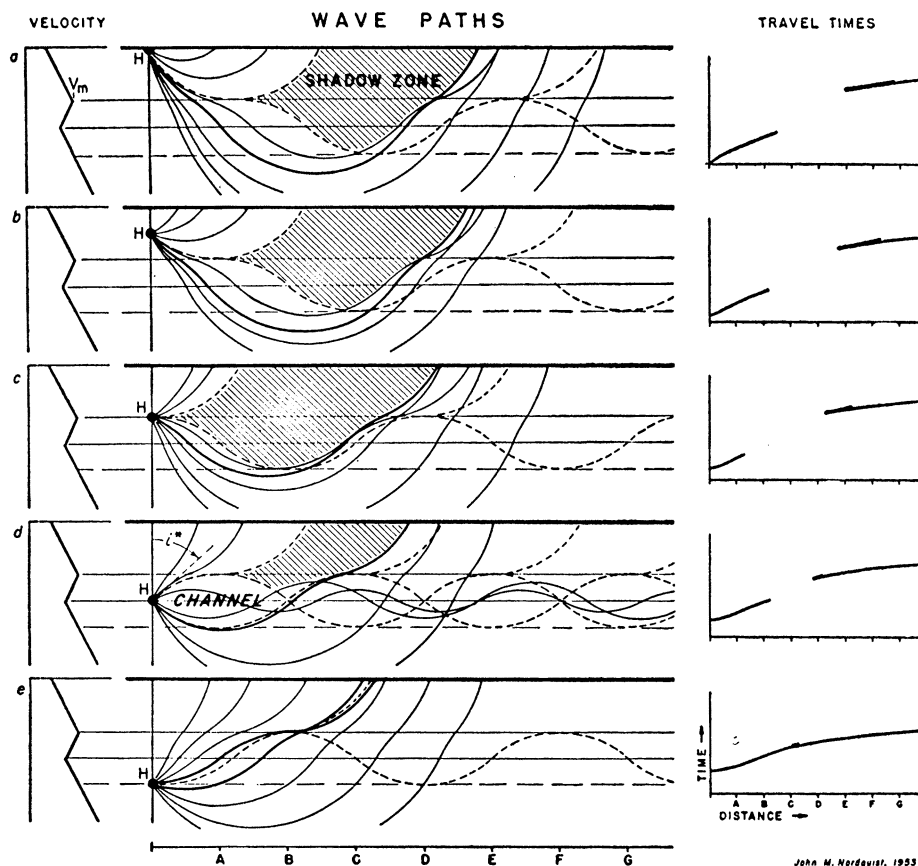


Fig. 2. Paths of sound waves (seismic waves), if the velocity changes with depth, as given at left. (Center) Wave paths for various depths of the source H . (Right) Corresponding transmission times as a function of the distance from the source.

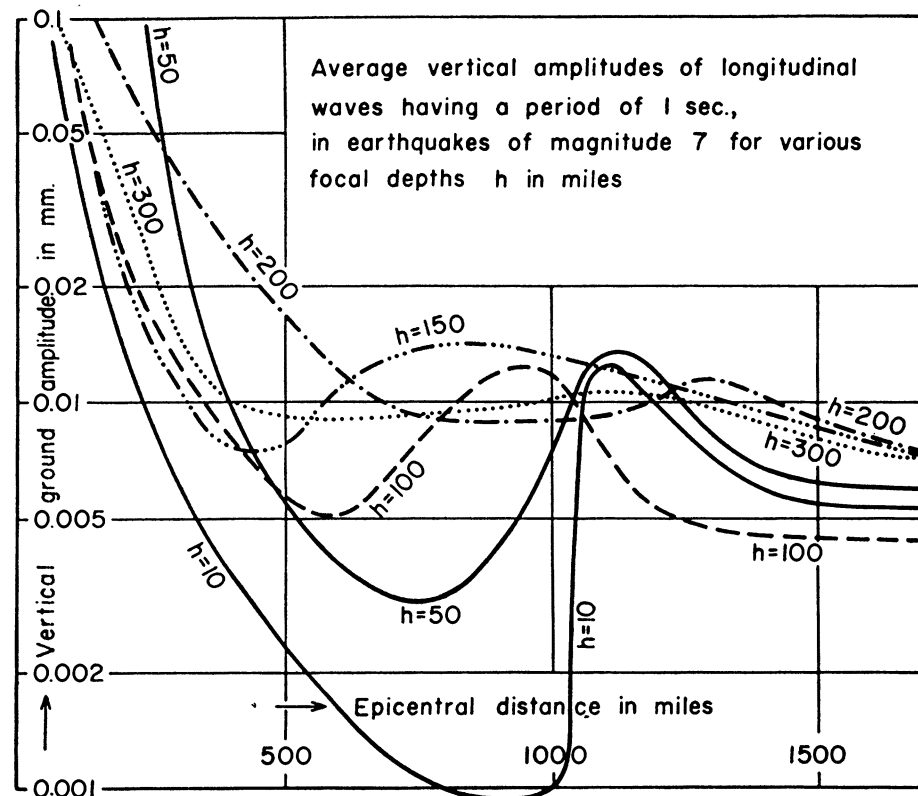


Fig. 3. Ground amplitudes of longitudinal waves in an earthquake of magnitude 7 as a function of the epicentral distance for various depths of the source; it is assumed that the wave periods are 1 second.

bounded by the Mohorovičić discontinuity, is frequently only 5 miles thick under ocean bottoms, as contrasted with about 15 miles (under some low lands) to 40 miles (under some high mountains) under continents. As we have seen, the deeper the discontinuity, the lower the velocity immediately under the discontinuity. Consequently, this velocity is reached at a shallower depth under high continents than under oceans (see Fig. 5).

Channel Waves

The Italian seismologist P. Caloi concluded in 1953 that "channel waves" should be propagated in the asthenosphere channel if the source of an earthquake is in the channel (see Fig. 2*d*). He found many instances where seismograms of deep-focus earthquakes

showed such waves as long as the source was at a depth of not over 150 miles. These waves travel to great distances, with a velocity of about 5 mi/sec (8 km/sec) for longitudinal, and 2.75 mi/sec (4.4 km/sec) for transverse, channel waves, regardless of whether they are propagated under continents or oceans. These velocities correspond to those found in the channel (Fig. 5). The waves affect the surface of the earth, since they travel only between one and a very few wavelengths below the surface. On the other hand, it has been suggested that all channel waves in or near the crust may be surface waves of higher modes. In any case, the fact that the waves considered here have been observed along continental and oceanic paths shows that the asthenosphere channel is not interrupted at the transition from continental to oceanic re-

gions. Such waves were observed independently in 1954 by Press and Ewing and later by others. In all instances, the depth ranges of the earthquakes involved and the velocities observed were the same, within small limits of error.

Cause of the Asthenosphere Channel

When I found the first indication of the asthenosphere channel in 1926 I was trying to determine whether there is any indication that molten material is present at a depth of 50 to 100 miles, where the estimated temperatures are close to the estimated melting point of the rocks expected there. The most recent estimates of the temperature at a depth of 60 miles vary between about 800° and 1500°C, those of temperatures at a depth of 180 miles, between about 1400° and 2000°C, while the corresponding melting points are estimated very roughly to be at 1500° and 1900°C, respectively. Thus, all we can say at present is that the melting point is fairly close to the actual temperature. However, at depths below 500 miles, practically all the estimated temperatures in the mantle are well below the estimated melting points.

A very interesting contribution to the questions of whether the melting point is reached in the upper mantle, and if so, where, was furnished by the Russian volcanologist G. S. Gorshkov in 1957. He found that no transverse waves are recorded in Kamchatka from Japanese earthquakes when a portion of their path follows the volcanic belt between Japan and Kamchatka, while they are well recorded at stations in Kamchatka to which the wave paths are only slightly different but not under the volcanic chain. Gorshkov concluded that there are foci of non-solid magma at a depth of about 35 miles under the volcanoes—that is, at a depth similar to that of the asthenosphere channel.

Laboratory experiments to find the velocities of waves in rocks at the temperatures and pressures in the earth's crust are not decisive. The conclusions depend on assumptions concerning the change in temperature with depth in the earth which are uncertain, as explained above. However, the results of the experiments do not exclude a decrease in velocity with depth in the asthenosphere channel. I believe that in the near future attempts should be made to

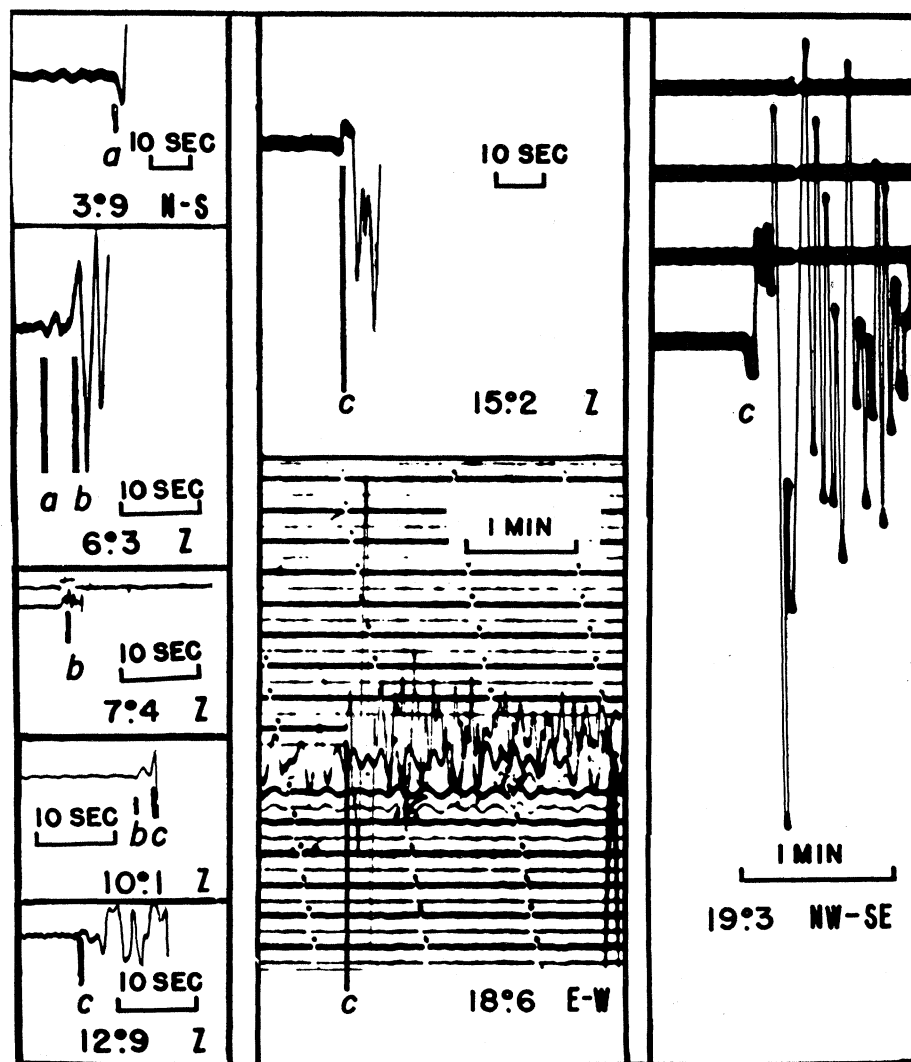


Fig. 4. Beginnings of selected seismograms of the Kern County, California, earthquake of 21 July 1952 ($1^\circ = 69$ miles). (N-S) The north-south component; (Z) the vertical component; (E-W) the east-west component; (NW-SE) a record obtained in the direction northwest-southeast.

reverse the calculations—that is, to assume values for the change of the wave velocities as a function of depth (for example, on the basis of data in Fig. 5) and to calculate the corresponding temperature in the upper portion of the mantle. A serious problem in all such calculations is that of deciding which rock type should be assumed.

The shadow zone produced by the asthenosphere channel (see in Fig. 3 the curve for a focal depth of 10 miles) has played an important role in the international discussions of the problem of how many stations are needed to locate atomic subsurface explosions. It has been realized during the discussions that stations have to be either fairly close to the source or over 1000 miles away. Figure 4 gives an idea of the differences in the amplitudes involved.

Lithosphere Low-Velocity Channels

Seismograms showing waves which have been transmitted only through the various layers of the earth's crust are rather complicated. About ten years ago it appeared that data from earthquake records were leading to different results, for the velocities involved, from data from artificial explosions. When I found, in addition, that the amplitudes of waves through a given layer may decrease faster with distance than would be expected if the velocity increases with depth, I concluded, in 1951, that in portions of at least two crustal layers the velocity may decrease with depth. Detailed investigation of these "lithosphere channels" is more difficult than that of the asthenosphere channel, since local differences in crustal structure are greater in the crust

than in the upper mantle, and it is frequently difficult to identify correctly the travel-time curve for waves through a given layer.

A break in this research came when Press and Ewing in 1952 discovered waves of two new types which travel with constant velocity. These waves have been observed only along continental paths. These, as well as additional waves of similar type, were investigated soon afterwards by Miss Lehmann in Denmark and M. Båth in Sweden. Båth established the fact that these waves disappear not only at the transition from continent to ocean but also under large mountain chains where the crustal layers are disturbed. These waves usually appear as a very striking short-period motion riding on top of the earliest, appreciably longer, surface waves. The regular "microseisms" which are visible continuously on records of sensitive instruments at most stations and are produced by high ocean waves near the coast may be lithosphere channel waves.

The cause of the lithosphere low-velocity layers is probably, again, the increase in temperature with depth. Close to the surface the effects of the increase in pressure (the closing of pores) prevail; below a depth of a few miles the effects of the increase in temperature may surpass those of the increase in pressure. Again, laboratory experiments are not decisive, for the same reasons which were cited in connection with the asthenosphere channel.

In the sedimentary layers of the crust, where the search for oil and minerals depends much on seismic methods, there are many "low-velocity" layers which interfere with the investigations of the structures still more than those in the deeper portion of the crust. For example, two types of layers may alternate with increasing depth; the one with the lower velocity creates a low-velocity layer each time it occurs. The resulting shadow zones are frequently a source of incorrect conclusions. Exploding the charge in a low-velocity layer may result in poor records.

Low-Velocity Channel in the Ocean

The velocity of sound waves in the ocean increases with increasing temperature, salinity, and pressure of the water. Usually the rather rapid decrease in temperature in the upper layers of the ocean results in a decrease in sound

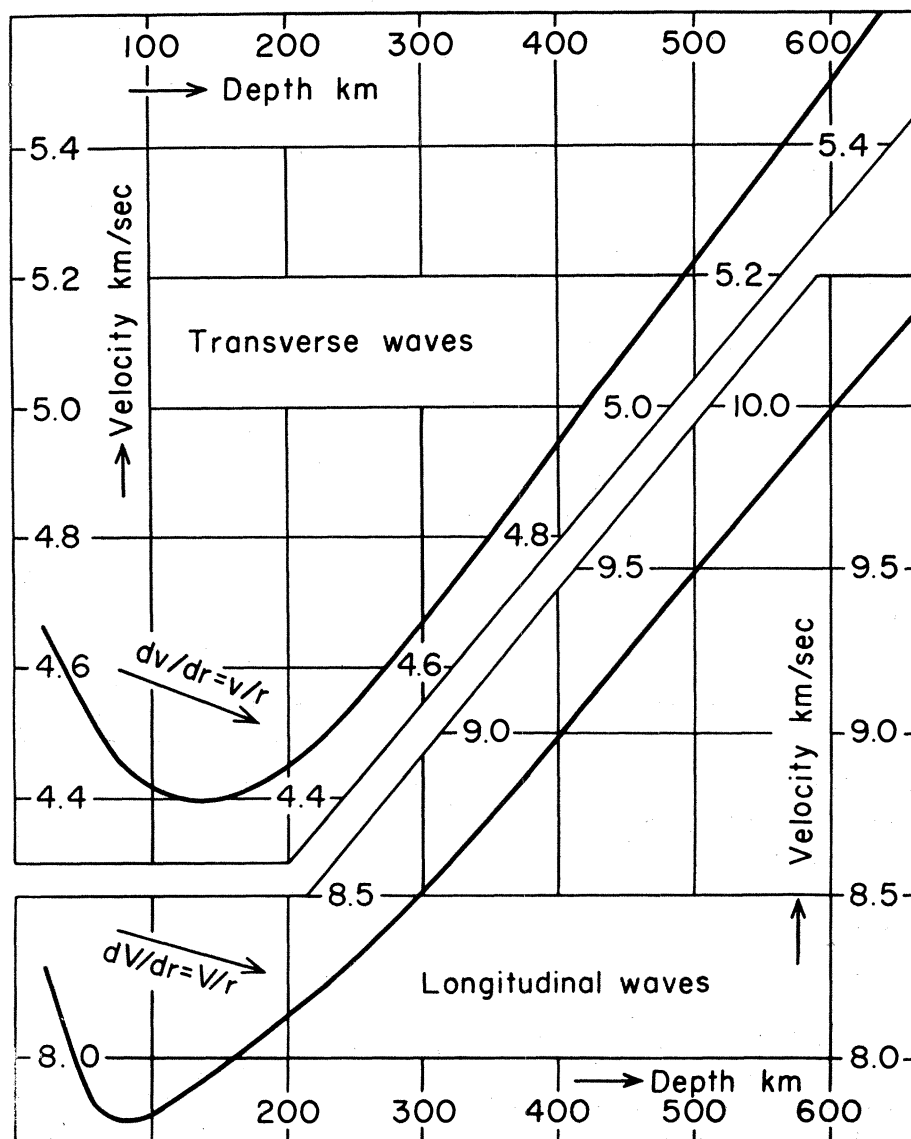


Fig. 5. Velocities V of longitudinal, and v of transverse, waves (in kilometers per second) as a function of the depth in the earth (in kilometers).

velocity with depth, while deeper down the effect of the increase in pressure prevails. Near the surface the sound velocity is nearly 5000 ft/sec (it depends on temperature and salinity), and it decreases usually by a few percent to a minimum at a depth of less than a

mile. Thus, we must normally expect in the ocean a low-velocity layer with all the phenomena indicated in Fig. 2. This conclusion was verified in 1934 by Dyk and Swainson of the United States Coast and Geodetic Survey, in little-publicized experiments. Independently,

Ewing and Worzel concluded in 1948 that the sound of an explosion of 4 pounds of TNT could be identified at a distance of 10,000 miles, if the TNT was exploded at a depth of 4000 feet in the low-velocity layer (see Fig. 2*d*). On the other hand, no sound from a submarine near the point *H* in Fig. 2*a* is audible aboard a ship in the shadow zone, which, under the conditions in the ocean, may begin immediately at *H* and may extend for 50 or even more miles around *H*.

Low-Velocity Channels in the Atmosphere

In the atmosphere, we may assume with sufficient accuracy a constant composition for the lowest 50 miles. In this case the sound velocity is proportional to the square root of the absolute temperature of the air. Since this decreases throughout the troposphere, except for relatively thin inversion layers, the sound velocity decreases, too, to a height of roughly 5 miles, depending on the latitude and the season. Since the temperature increases again at higher levels in the stratosphere, low-velocity layers are formed and we can apply the data in Fig. 2, which we now have to look at upside down. The effect of wind has to be considered in interpreting the observations.

In 1903 it was found that after an accidental explosion a zone of audibility surrounded the source, a "zone of

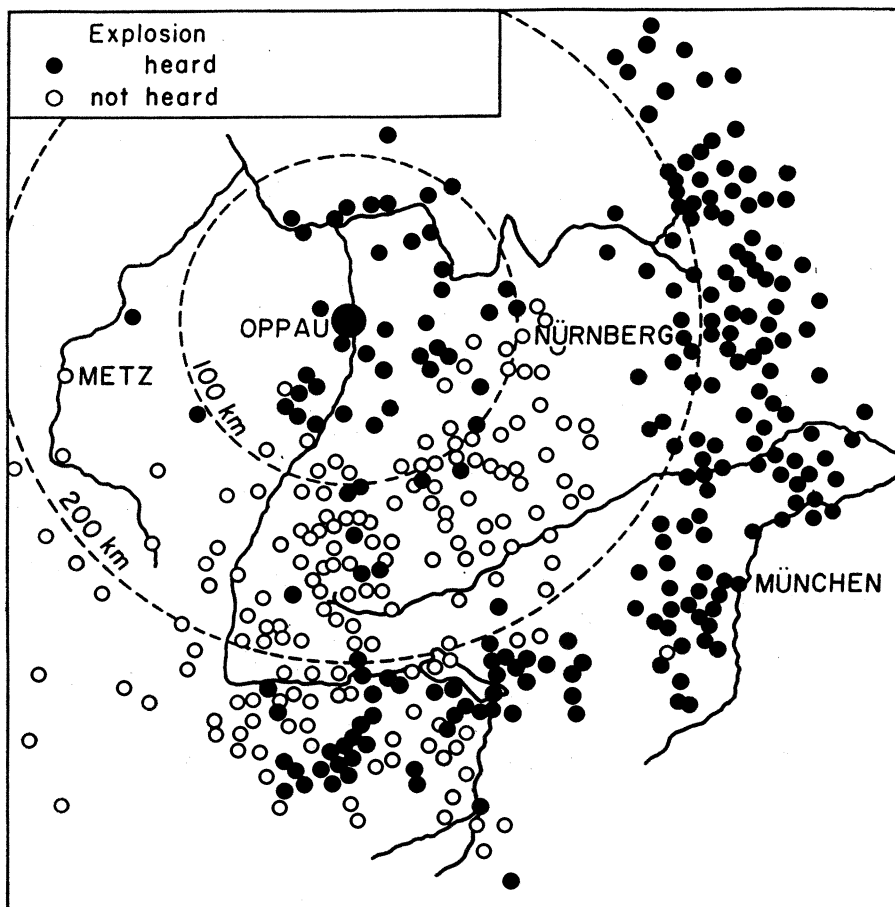


Fig. 6. Points at which the sound from the explosion at Oppau (Baden, Germany) on 21 September 1921 was heard and points where it was not heard (data from A. de Quervain). No observations were collected to the northwest and north of the source beyond a distance of about 100 km.

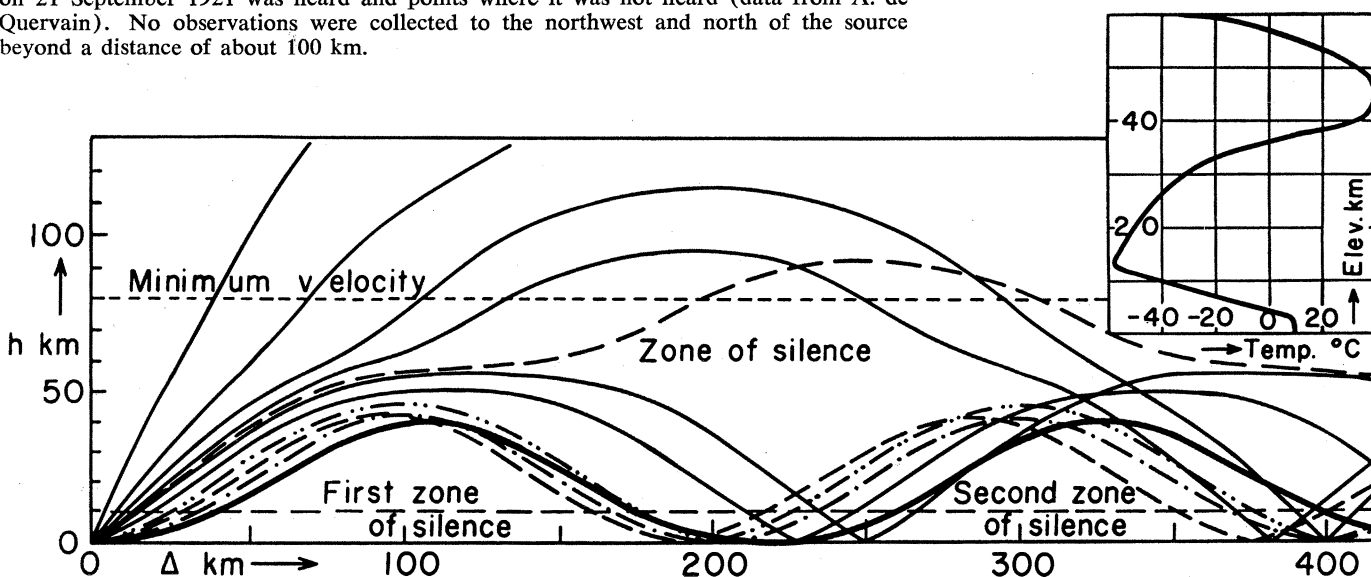


Fig. 7. Average paths of sound waves in the atmosphere [Δ , distance; h , elevation above ground (both in kilometers)]. (Insert) Characteristic temperature (in degrees centigrade) in middle latitudes as a function of the elevation h .

silence" followed, and at a distance of roughly 100 miles there was again a zone of strong sound. Figure 6 shows similar observations collected, after an accidental explosion in 1921, by the Swiss seismologist de Quervain. Richter and I reported an instance in 1930 in which target practice of the Navy off southern California was heard near the source; then came a zone of silence; then, at distances of about 70 miles from the source, the sound was so strong that windows rattled severely and, in one case, were even cracked. There is little doubt that sonic booms which sometimes irritate the population over fairly large areas may result from planes accelerating beyond the sound barrier; these planes may well be between 50 and 100 miles away (see, in Fig. 2, the caustic at the end of the shadow zone, and see Fig. 7). These large amplitudes correspond to those recorded in earthquakes about 1000 miles from the source (Fig. 4, middle and right).

Details of the sound propagation in the atmosphere were first investigated about 30 years ago, when the temperature in the atmosphere above an elevation of about 15 or 20 miles was still unknown. Discovery of the existence of the caustics beyond the zone of silence (Fig. 6) and the observed travel

times of the sound left no doubt that at greater elevations the sound velocity, and consequently the temperature, must increase again. Figure 7 shows results for the temperature and wave paths based on calculations which I made in 1925. At that time, great doubt was expressed about the calculated relatively high temperature at an elevation of about 30 miles. Recent direct observations show that the temperatures calculated from the observed sound waves arriving at the ground were substantially correct.

In 1949 Cox pointed out that a second minimum of the temperature and the sound velocity at an elevation of about 50 miles produces a second low-velocity layer. Its effect on the wave paths is indicated in Fig. 7. However, the corresponding waves arriving at the earth's surface are relatively small, since at elevations above 50 miles the absorption of sound increases rather rapidly; the molecules are too far apart to permit good transmission of energy.

Finally, sound waves have been observed to circle the earth repeatedly after large explosions—for example, that of Krakatoa or those of atomic blasts in the atmosphere. These waves correspond in principle to the channel waves (Fig. 2*d*), although other at-

mospheric processes affect the details.

We thus find that one single, little-publicized phenomenon—the low-velocity layer—which has been disregarded by some geophysicists and mentioned infrequently by others in explanation of observations, plays an important role in many instances of wave propagation through the earth's solid body, the oceans, and the atmosphere (1).

Note

1. I presented the first version of this article, "Low-velocity layers in the earth's interior, the ocean and the atmosphere," upon invitation at a meeting of the American Physical Society in Mexico City in June 1950. A revised version was my presidential address before the International Association of Seismology and Physics of the Earth's Interior, in Rome, 14 Sept. 1954. This was published in condensed form in *Geofisica pura e applicata* [28, 1 (1954)]. This version contains many references and the original of Fig. 2. The present version is contribution No. 941 of the Division of the Geological Sciences, California Institute of Technology, Pasadena. More detailed data on the low-velocity layers in the earth, and on temperature and pressure and other physical conditions in the earth, may be found in B. Gutenberg, *Physics of the Earth's Interior* (Academic Press, New York, 1959). Additional data are given in B. Gutenberg, "The asthenosphere low-velocity layer," *Annali di Geofisica* (in press). For data concerning the low-velocity layer in the ocean, see M. Ewing and J. L. Worzel, "Long range sound transmission," in "Propagation of Sound in the Ocean," [*Geol. Soc. Am. Mem. No. 27* (1948)]; also, K. Dyk and O. W. Swainson, "The velocity and ray paths of sound waves in deep sea water," *Geophysics* 18 (1953). For discussion of propagation of sound in the atmosphere, see B. Gutenberg, in *Compendium of Meteorology* (American Meteorological Society, 1951), pp. 366–375.

Beno Gutenberg, Geophysicist

The death of Beno Gutenberg in Pasadena, California, on 25 January 1960 marked the end of an era in seismology. That era had been dominated by him. A prodigious worker with broad interests, his was the era before computing machines and team attacks.

Working alone and with his colleague Charles Richter, he covered the gamut of seismology. His early interest in meteorology continued throughout his life, as did his broad interest in the gen-

eral problems of the physics of the earth.

Beno was born in Darmstadt in Hesse on 4 June 1889. As a student in the University of Göttingen he worked under Emil Wiechert. At the age of 22 he received the degree of doctor of philosophy, with a thesis on the subject of microseisms. For a year after receiving his degree he remained in Göttingen and completed his great work of computing the depth of the earth's core. His

value of 2900 kilometers still stands—one of the few numbers in seismology which is generally agreed upon.

Gutenberg served a year as meteorologist in the army. He then served on the staff of the International Seismological Central Station in Strasbourg. He was a professor at Frankfurt in 1930 when he was called to California Institute of Technology as professor of geophysics and meteorology. He later was appointed the first director of the Seismological Laboratory, a position which he occupied until his retirement in 1957.

It is difficult to select particular items from the contributions of so prolific a researcher. The several editions of the *Seismicity of the Earth* (which he wrote with Richter) are a monumental work of great value. When Beno received in 1953 the Bowie medal of the American Geophysical Union the president referred to this work as the "Gutenberg Bible." His collaboration with Richter