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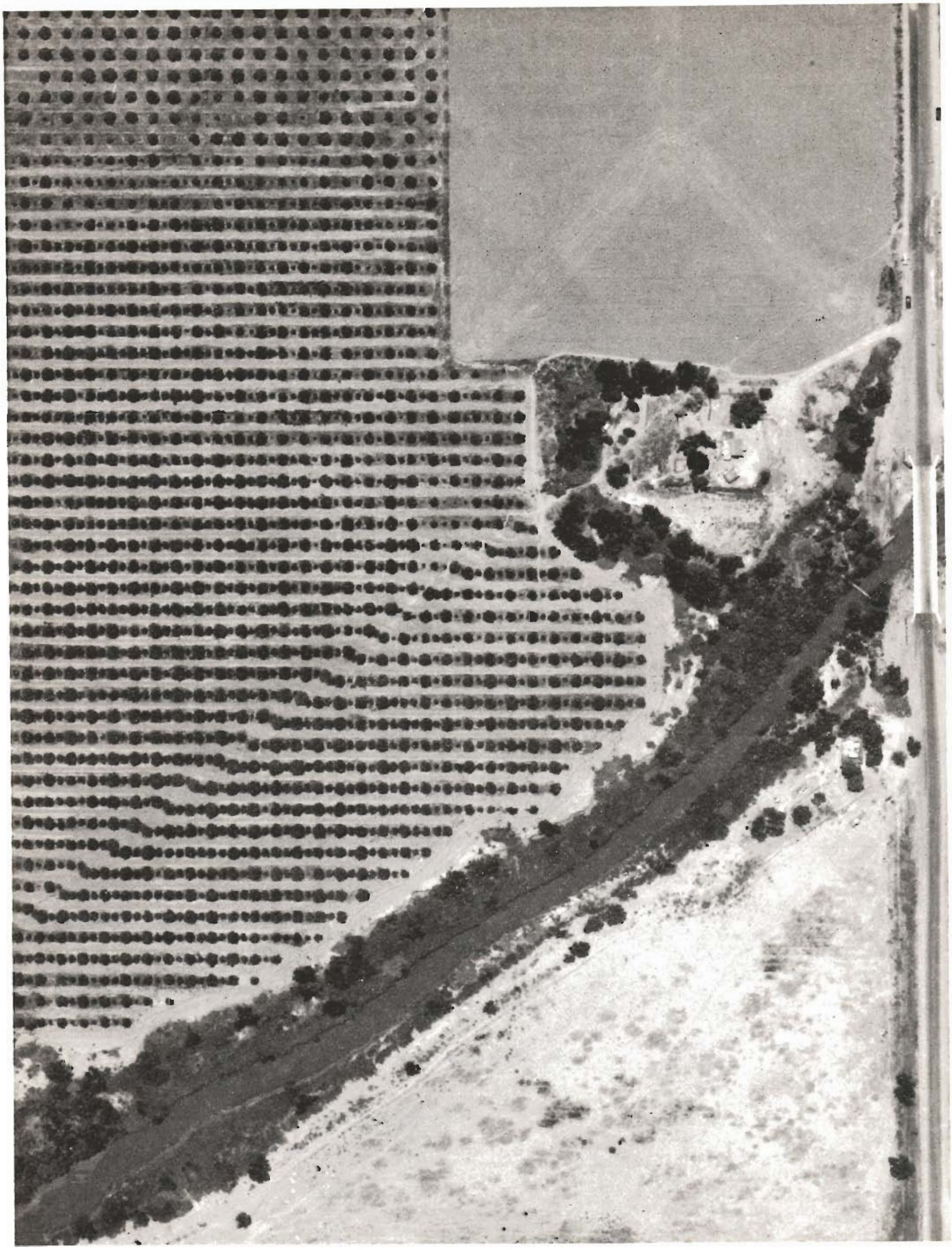


HORIZONTAL DISPLACEMENT of the ground during an earthquake in the Imperial Valley of California disrupted the regular pattern of trees in citrus groves. In this aerial photograph of an orchard seven miles east of Calexico, made shortly after the earthquake in 1940, the path of the San Andreas fault can be clearly traced diagonally across the groves west of the Alamo River. North is to the right.

The Motion of the Ground in Earthquakes

The slippage along a fault that produces an earthquake radiates seismic waves. Exactly how these waves shake the ground bears on the design of buildings and other structures in earthquake zones

by David M. Boore

Last year half a million people were killed by an earthquake that devastated the Chinese industrial city of Tangshan. In the western U.S. over the years earthquakes have caused considerable damage, although the number of fatalities has been relatively small. The low casualty rate has been partly due to the fact that many of the major earthquakes occurred either in sparsely populated areas or fortuitously quite early in the morning, when most large office and public buildings are almost empty. Over the past few decades, however, many earthquake-prone regions of the western U.S. have become further urbanized. In them more large buildings and facilities such as dams have been constructed or are being planned. If such structures were to fail during a future earthquake, large numbers of people could be killed or injured.

Today the attention of many seismologists is being focused on ways to reduce the hazards of earthquakes by learning how to predict their consequences. To many people the term earthquake prediction probably suggests determining the time, place and magnitude of future earthquakes. Equally important is determining which of many ways the ground is likely to shake during the earthquake, how strong the shaking will be and how long it will last. Knowledge of the ground motion that can be expected during an earthquake can make it possible to design structures that do not need unnecessary and uneconomic levels of strength in order to survive being shaken.

In order to predict both the occurrence of an earthquake and the ground motion it will generate it is essential to understand the characteristics of the earthquake source. So far most of our understanding of earthquake sources has come from measurements made during actual earthquakes at seismological stations some distance from the source. Such measurements yield information about certain average properties of the earthquake source, for example the dimensions of the original disturbance and the overall movement involved in it. Average properties are useful in elucidating how seismic energy is released and how it is transmitted over large areas; they have also been invaluable in probing the structure and nature of the earth's interior and in assessing the likelihood of large earthquakes in certain regions. Such average properties, however, yield little information about the details of the ground shaking in areas immediately surrounding the earthquake source. It is this kind of specific information structural engineers require. For that reason a number of seismologists are now beginning to investigate the details of earthquake sources. This important subject, which might be called strong-motion seismology, is still in its infancy but should grow rapidly.

Historically our understanding of the cause of earthquakes is relatively new. By the middle of the 19th century it had been observed that the damage caused by many earthquakes was concentrated in a narrow zone, which suggested that earthquakes had a localized source. It was not until the San Francisco earthquake of 1906, however, that it was recognized that earthquakes were caused by slippage along a fault in the earth's crust. In a classic study conducted shortly after the earthquake Harry F. Reid of Johns Hopkins University discovered that for several hundred kilometers along the San Andreas fault fences and roads crossing the fault had been displaced by as much as six meters. Moreover, precise geodetic surveys conducted before and after the earthquake demonstrated that the rocks parallel to the fault had been strained and sheared. On the basis of such observations Reid proposed the elastic-rebound theory of earthquakes.

According to the elastic-rebound theory, rocks are elastic, and mechanical energy can be stored in them just as it is stored in a compressed spring. When the two blocks forming the opposite sides of the fault move by a small amount, the motion elastically strains the rocks near the fault. When the stress becomes larger than the frictional strength of the fault, the frictional bond fails at its weakest point. That point of initial rupture, called the hypocenter, may be near the surface or deep below it.

From the hypocenter the rupture rapidly propagates along the surface of the fault, causing the rocks on opposite sides of the fault to begin to slip past each other. A portion of the frictional stress the rocks had exerted on each other before the rupture is suddenly and violently released; the rocks along the fault rebound, or spring back, to an equilibrium position in a matter of seconds. The elastic energy stored in the rocks is released as heat generated by friction and as seismic waves. The seismic waves radiate from the hypocenter in all directions, producing the earthquake. The point on the surface of the earth above the hypocenter is the epicenter of the earthquake.

In some cases the rocks rebound not in a period of seconds but over an interval of minutes, days or even years. The seismic energy radiated at any one time is then quite small. This slow process is known as aseismic slip or creep. Why the seismic energy is released violently in some cases and not in others is not well understood.

Although the physical details of the elastic-rebound theory are still uncertain, the conceptual model of the faulting process fits well with the current hypotheses of plate tectonics. Most earthquakes are generated in zones where the huge plates of the lithosphere, which make up the outer layer of the earth's surface, are shearing past each other.

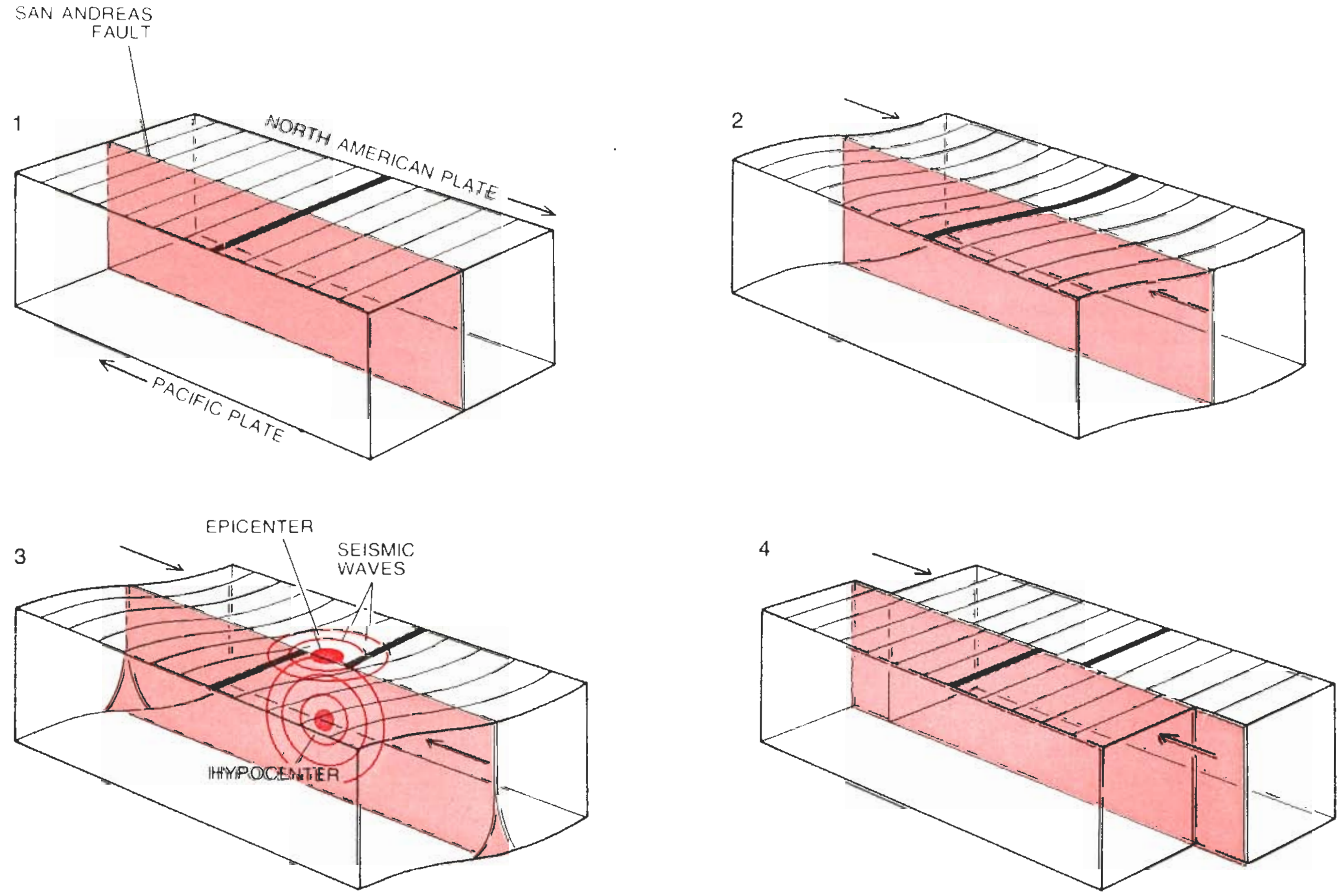
The concept of slip along a fault is at the heart of virtually all studies of earthquake sources. Indeed, the concept developed largely from investigations of earthquakes along the San Andreas fault. The San Andreas is a very long fault but not a deep one; earthquakes caused by its slippage are confined to about the upper 15 kilometers of the crust. Yet the study of this one shallow fault has led to a model that successfully explains the deformation of the ground and the radiation of seismic waves from all types of seismic sources, ranging from the shallowest slips to ruptures as deep as 700 kilometers along the advancing edge of a plate plunging below another plate.

The way the ground is deformed and the nature of the seismic waves that radiate during the earthquake provide basic information about the earthquake source: its dimensions, its shape and its orientation. The seismic waves have a wide range of period and amplitude. When a fault slips, the rupture process itself generally lasts between a fraction of a second (for a minor earthquake) and five minutes (for a major one). The waves generated by the fault's slippage can have periods ranging from essentially infinity down to less than a tenth of a second. The seismic waves with the longest period correspond to the quasi-permanent deformation of the ground around the fault. The waves with the shortest period actually fall into the low audible range. The waves with periods of about an hour have a frequency that coincides with the resonance frequency of the earth, and they cause the entire planet to ring like a giant bell.

The amplitude of the seismic waves can range from micrometers (millionths of a meter) to tens of meters. The amount by which the seismic waves deform the ground decreases with distance from the earthquake. In the great Chilean earthquake of 1960, for example, the total displacement of some points immediately adjacent to the fault ranged up to 20 meters. At Los Angeles, a quarter of the way around the world, the maximum displacement of the ground was about two millimeters.

Since seismic waves span such a broad spectrum of period and amplitude, many different kinds of instruments and experimental techniques are needed to capture all the information radiated by an earthquake source. Repeated geodetic surveys of the earth's surface can monitor deformations of the ground created by seismic waves with periods ranging from days to years. A variety of different seismographs have been designed to record seismic waves with periods ranging from an hour to a hundredth of a second. Some instruments are so sensitive that they can detect motions as minute as one micrometer, which they magnify tens of thousands of times in order to record them on paper. Other instruments are so rugged that they can withstand the jarring accelerations of the most violent earthquakes.

The record produced by the seismograph—a seismogram—holds a great deal of information; even with the aid of a computer, however, deciphering that information is neither simple nor straightforward. The waves recorded on a seismogram after passing through the earth can be thought of as violin music

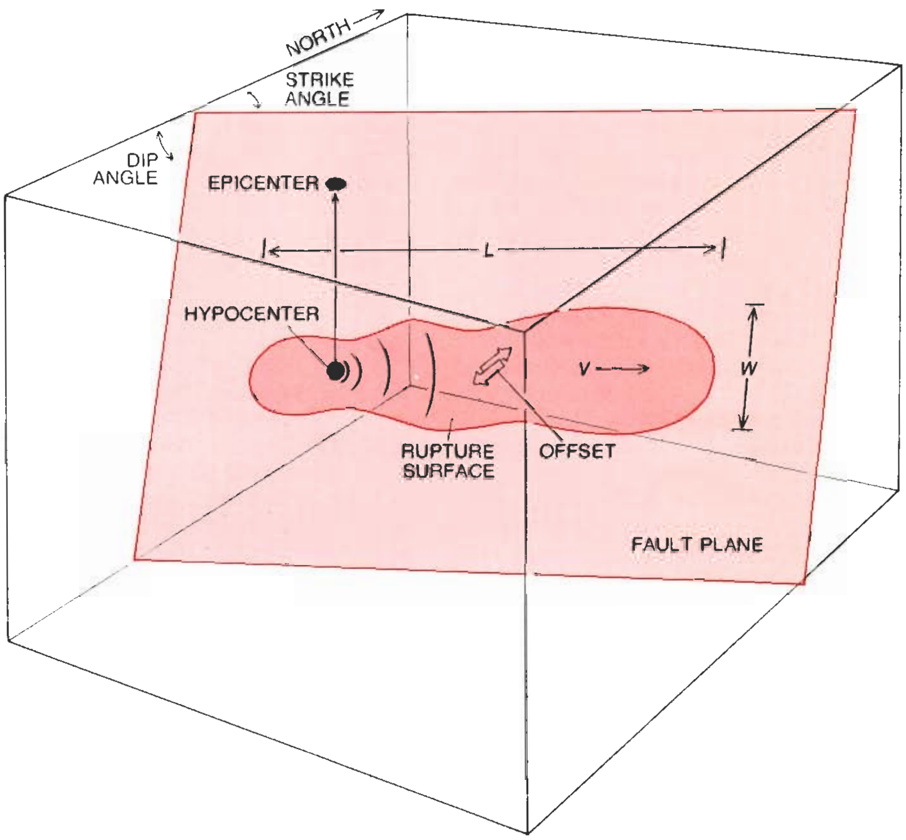


ELASTIC-REBOUND MODEL OF EARTHQUAKES assumes that two moving blocks of the earth's crust, each of which is part of a different tectonic plate in the earth's lithosphere, meet at a fault (*1*). Friction between the plates along the surface of the fault at first keeps them from slipping past each other, but the material around the fault is deformed by the stress (*2*). The deformation builds up until the frictional lock is ruptured at its weakest point, usually well below the surface (*3*). The rupture spreads out from that point, the hypocenter, radiating seismic waves as it does so. The point vertically above the hypocenter, where the seismic waves first reach the surface, is the epicenter of the earthquake. As the rupture spreads along the surface of the fault the blocks slip past each other, usually in a few seconds, coming to rest in a new equilibrium position (*4*). The stress around fault is relieved and ground rebounds to earlier state.

recorded on magnetic tape after first being transmitted over a telephone line that distorts the music. In this analogy the violin corresponds to the seismic source, the telephone line corresponds to the inhomogeneous elastic earth that distorts the signal passing through it and the tape recorder corresponds to the seismograph (which further distorts the signal as it is being recorded).

It is easy to correct for the distortion of the tape recorder. The challenge lies in trying to deduce something about the nature of the violin on the basis of the distorted sound received at the end of the telephone line. If one assumes that the telephone line is free of distortion, one might then reasonably conclude that a violin intrinsically produces a harsh sound. On the other hand, if one knows how a violin sounds when it is heard "live," one could use that knowledge to discover how the telephone line filters and distorts the music.

A similar problem faces the seismologist examining the record of an earthquake. In seismology the earth filter that distorts the seismic waves is complex because the internal structure of the earth is complex. As a result of decades of geological research, however, we now know much more about the earth's internal structure and how it distorts a seismic signal than we know about the earthquake source. Because the earthquake source is usually deep underground its seismic radiation cannot be "heard" firsthand. Seismologists must deduce the nature of the source by the indirect procedure of constructing a theoretical model of it, calculating the pattern of seismic radiation produced by the model, estimating how the seismic signal would be distorted as it propagated through the earth to the seismograph and comparing the synthetic seismogram with the actual seismogram recorded. By repeating the procedure several times with better information it is possible to refine the description of the earthquake source. Current models thus constructed attempt to describe the complex rupture process with relatively few parameters.



IDEALIZED MODEL OF EARTHQUAKE SOURCE suffices to describe most earthquakes with about a dozen variables. In the model the rupture begins at the hypocenter h kilometers below the surface, spreads across a fault plane at a velocity V and finally stops after growing into a region with an average length L and an average width W. The orientation of the fault plane is specified by its strike angle and dip angle. The slip between the two fault surfaces (*large arrows*) can have any orientation in the plane. On the average the slip requires τ seconds to reach its final offset. All these parameters are determined from recordings of the seismic waves.

At the simplest level a model specifies the location of the hypocenter and the magnitude of the earthquake. At a more complex level the model includes the orientation of the fault surface underground and the direction of slip across the surface. The model can be made even more realistic by adding the dimensions of the entire area that ruptured, the average amount of slip across that area and the average length of time required for a point on the fault surface to be offset by the maximum amount. Since friction opposes the motion of the two sides of the fault past each other, it is believed that once a fault begins to slip, its direction cannot reverse. Such models are quite successful in predicting the different types of seismic waves actually observed, particularly in predicting seismic waves with wavelengths at least as long as the dimensions of the fault.

The location of an earthquake can be determined by a procedure akin to triangulation, taking advantage of the fact that different types of seismic waves travel at different speeds. Seismic waves are of two general types: P waves and S waves. The P waves are longitudinal compression waves that travel through the deep interior of the earth, even propagating through the lower mantle and the liquid core. The S waves are transverse shear waves that travel through the solid portions of the earth.

P waves travel significantly faster than S waves. At a location close to the earthquake source the two types of waves will arrive fairly close together, but at one farther away the S wave will lag significantly behind the P wave. By observing the difference in arrival time between the two types of waves at any one station it is possible to calculate the distance of the earthquake from the station. Such a calculation from a single station does not determine the direction of the earthquake, but when observations from three or more stations are combined, the precise location of the earthquake can be determined. If there are enough data, it is also possible to locate earthquakes from the P waves alone. In fact, this is the technique used by the National Earthquake Information Service in Golden, Colo., which collates earthquake data recorded all over the world and issues information about the position of an earthquake as soon as possible after each event.

The most widely recognized measure of the strength of an earthquake is the scale of magnitudes developed in the 1930's and 1940's by Charles F. Richter and Beno Gutenberg of the California Institute of Technology. The scale is based on the notion that ideally the magnitude determined should be an absolute measure of the energy released by the earthquake itself and should not be affected by the location of the seismographic station or the particular seismograph employed. The Richter method for determining the magnitude of an earthquake is quite simple. First, the seismologist measures the amplitude of the ground motion recorded in a certain specified part of the train of seismic waves. Second, he divides the recorded amplitude of the ground motion by the magnification of the particular seismograph to estimate the true ground motion at the seismographic station. Third, he calculates the common logarithm (the logarithm to base 10) of that ground motion. Fourth, he applies certain empirical corrections to that number to compensate both for the attenuation of the ground motion as it spreads out from the earthquake source and for the degree to which the response of the particular seismograph is influenced by local geological conditions.

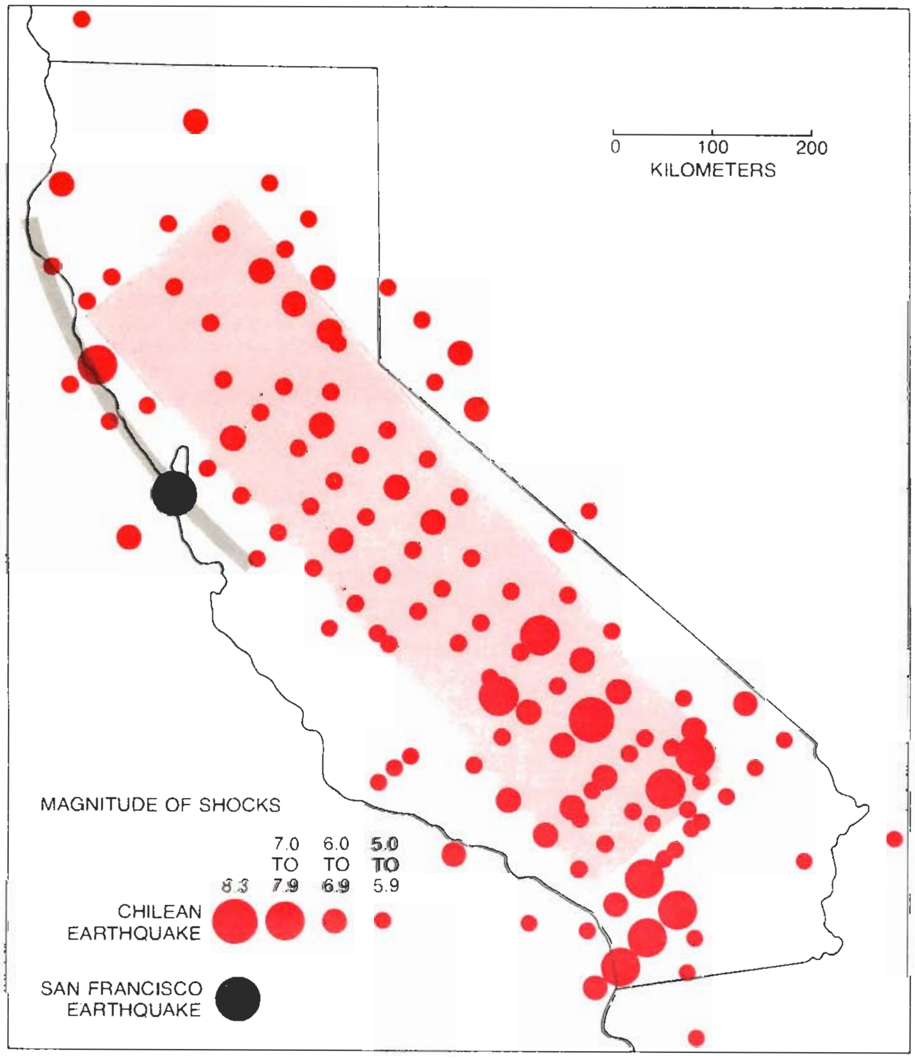
The empirical corrections are applied so that for any given earthquake the same magnitude should be determined at all seismographic stations. In practice the magnitudes differ from one station to another, and an average magnitude is calculated from all of them. On the Richter magnitude scale larger numbers correspond to larger events. Since the scale is based on the common logarithm of the corrected ground displacement, each increase of one magnitude unit implies an increase of a factor of 10 in the amplitude of the ground motion. The magnitude scale is open-ended, and negative magnitudes have been measured.

Actually there are several magnitude scales in common use, each based on a different part of the seismic wave train. One is the scale of body-wave magnitude, measured from the *P* waves that travel through the body of the earth and reach the seismograph before any other waves. By convention, *P* waves with a period near one second are used in the magnitude determination. Another scale is the scale of surface-wave magnitude, measured from the dispersed waves that travel over the surface of the earth and reach the seismograph somewhat later. The surface waves employed have periods of 20 seconds. The two magnitude scales are cross-calibrated so that on the average both will yield the same magnitude when the earthquake being recorded has a magnitude of 6.75. By measuring the two magnitudes for a particular earthquake one obtains an estimate of the overall amount of seismic energy radiated in two quite different regions of the seismic spectrum. For a large earthquake the surface-wave magnitude is generally greater than the body-wave magnitude. This fact implies that the excitation in the long-period part of the spectrum increases faster with earthquake size than the excitation in the short-period part of the spectrum.

After the location and the magnitude of an earthquake have been determined from seismograms, the kind of information that can next be most readily obtained is the geometry of the earthquake source: the orientation of the fault in the earth, the dimensions of the portion of the fault plane that has slipped and the direction of the slip in the fault plane. Just as an array of radar antennas has a defined pattern of radiation, with large amounts of energy being beamed in some directions and small amounts in other directions, so also does an earthquake source have a defined pattern in which it radiates seismic energy. The radiation pattern not only determines the amplitude of the seismic signal in different directions but also determines how the seismic waves are polarized.

The radiation pattern can be understood by means of a simple experiment with a cube of foam rubber. Slit the top of the cube and push the two sides horizontally in opposite directions parallel to the slit. You will notice that the foam is compressed in two diametrically opposed quadrants and dilated in the other two quadrants. When a fault slips, the material around it is similarly compressed and dilated. The first waves emitted from an earthquake fault display the same distribution of compressions and dilations. The distribution of those waves on the surface thus reveals the orientation of the fault plane and the relative direction of the slip.

In the experiment with the cube of foam rubber, however, the quadrants of compression and dilation are clearly separated by two orthogonal lines; one line is the fault and the other line is perpendicular to the fault. Observations of the radiation pattern from an earthquake determine the orientation of two similar orthogonal planes, either one of which may be the earthquake fault. The ambiguity can be resolved if the orientation of the true fault plane is known

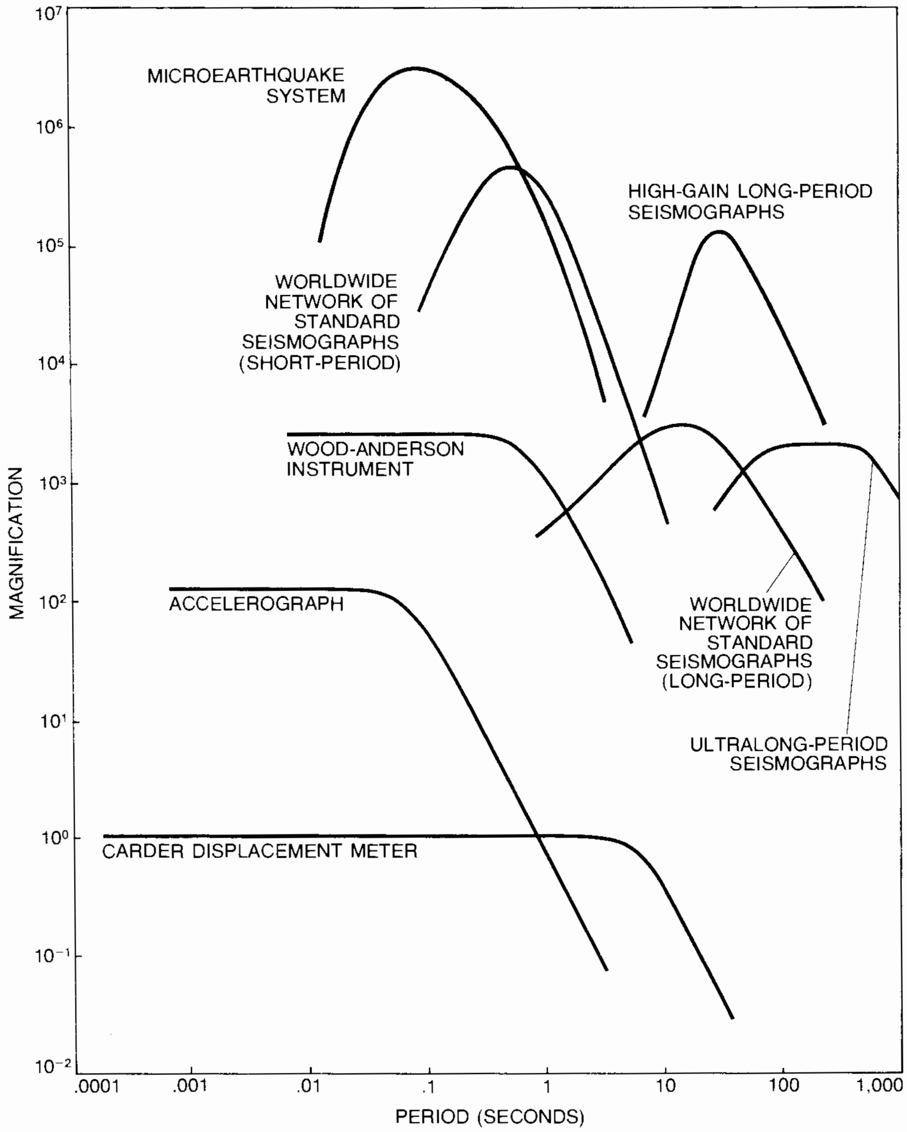


MAGNITUDE OF EARTHQUAKES is an inadequate measure of the actual size of large earthquakes. Both the San Francisco earthquake of 1906 and the Chilean earthquake of 1960 had a magnitude of 8.3. The area that ruptured in the San Francisco earthquake (*gray*), however, was approximately 15 kilometers deep and 400 kilometers long whereas the area that ruptured in the Chilean earthquake (*color*) extended to a depth equal to half the width of the state of California. The black dot represents the location of the epicenter of the San Francisco earthquake; the dots in color represent the locations of aftershocks of the Chilean earthquake with respect to its epicenter (*largest color dot*), superposed on the map of California for scale. The diameter of each dot represents the magnitude of each shock. Because earthquakes in California are caused by plates sliding past each other horizontally and not by plates subducting over each other as in Chile, no earthquake in California will be as great as earthquakes in Chile.

from the local geology. Alternatively, the orientation of the true fault plane can be determined from the pattern of aftershocks, smaller tremors that generally follow an earthquake, because the hypocenters of the aftershocks are usually scattered along the fault plane.

The information about the geometry of faults that has been amassed from earthquakes has been invaluable in developing the theory of plate tectonics. It has played a key role in identifying the faults between plates of the lithosphere and the relative motions of the plates. The seismic waves from an earthquake also yield information about the dimensions of the area that ruptured along the plane of the fault. The detail of the rupture area it is possible to resolve depends on the wavelength of the seismic radiation, just as in optics the wavelength of light limits the resolution of visual observation.

The area that ruptured in 1906, causing the magnitude-8.3 San Francisco earthquake, was 15 kilometers deep and 400 kilometers long; the area that ruptured in 1971, causing the magnitude-6.5 San Fernando earthquake in the Los Angeles area, was also 15 kilometers deep, but it was only 15 kilometers long. Seismic waves travel about four kilometers per second. The surface waves with a period of 20 seconds hence have a wavelength of some 80 kilometers. The 20-second waves might have provided a certain amount of detailed information about the source of the San Francisco earthquake, but with such waves the source of the smaller San Fernando earthquake would have appeared to be a point. By the same token, with seismic radiation having a period of several hundred seconds even the source of the San Francisco earthquake would have seemed to be a point.



RESPONSE OF SEISMOGRAPHS of different types has been tailored to monitor seismic waves over a broad spectrum of period and amplitude. The magnification of the instrument is the number of times the instrument amplifies the ground motion so that it can be recorded. The amplitude of the ground motion in centimeters is approximately equal to the inverse of the magnification. For the most sensitive instruments the magnification is limited by ambient vibrations of the ground produced by wind and surf. The microearthquake system records small earthquakes within about 100 kilometers of the instrument. The Wood-Anderson instrument records moderate earthquakes at distances of several hundred kilometers. Moderate-sized earthquakes occurring almost anywhere in the world can be recorded on the short-period and the long-period systems of the Worldwide Network of Standard Seismographs or on special instruments such as the ultralong-period seismograph or high-gain long-period seismograph. Carder displacement meter and accelerograph record strong shaking close to fault.

Clearly a fault is not a point source. As a rupture propagates over the surface of the fault the point from which the seismic radiation is being emitted moves and causes the seismic waves emitted from one portion of the fault to destructively interfere with the waves emitted from another portion. The shorter the period, the more important the destructive interference. The period at which the interference first becomes noticeable can be used to estimate the dimensions of the fault. For example, the period might be about six seconds for a fault with dimensions of 10 kilometers by 10 kilometers or 60 seconds for one with dimensions of 100 kilometers by 100 kilometers.

In the 20th century about 55 earthquakes have been observed with surface-wave magnitudes ranging between 8.0 and 8.7, and no earthquakes have been observed with a surface-wave magnitude greater than 8.7. Actually two earthquakes near the upper end of the magnitude range may have the same surface-wave or body-wave magnitude and yet radiate vastly different amounts of seismic energy. In other words, for large earthquakes the magnitude scale becomes saturated.

The reason for this saturation is easily understood. The largest earthquakes rupture faults hundreds of kilometers long. If a fault is very long, it takes more time for a wave emitted from the farther end of the fault to reach the seismograph than it does for a wave emitted from the nearer end of the fault. Since the wavelength of a surface wave can be much shorter than the length of a very long fault, the part of the wave train from which the earthquake's magnitude is measured will be emitted from only a fraction of the fault's area rather than from the entire fault. The result is that the strength of the earthquake appears to be less than it actually is, and the magnitude scale cannot accurately measure very large earthquakes.

A new measure of the strength of an earthquake, known as seismic moment, has recently come to the fore. Seismic moment is not as easy to measure as seismic magnitude, but it is a more physical measure of the size of an earthquake source. The seismic moment is determined by the Fourier analysis of seismic waves of such long period that the details of the rupture are smoothed out and the entire fault appears to be a point source. (The periods at which the seismic moment is determined increase with the size of the fault.) If the fault is "viewed" by such long-period waves, the slip from the unruptured state to the ruptured one appears to be instantaneous. The actual pattern of the seismic radiation emitted by the instantaneous rupture is mathematically equivalent to the theoretical pattern of radiation emitted by a model consisting of two hypothetical torque couples embedded in an unruptured elastic medium.

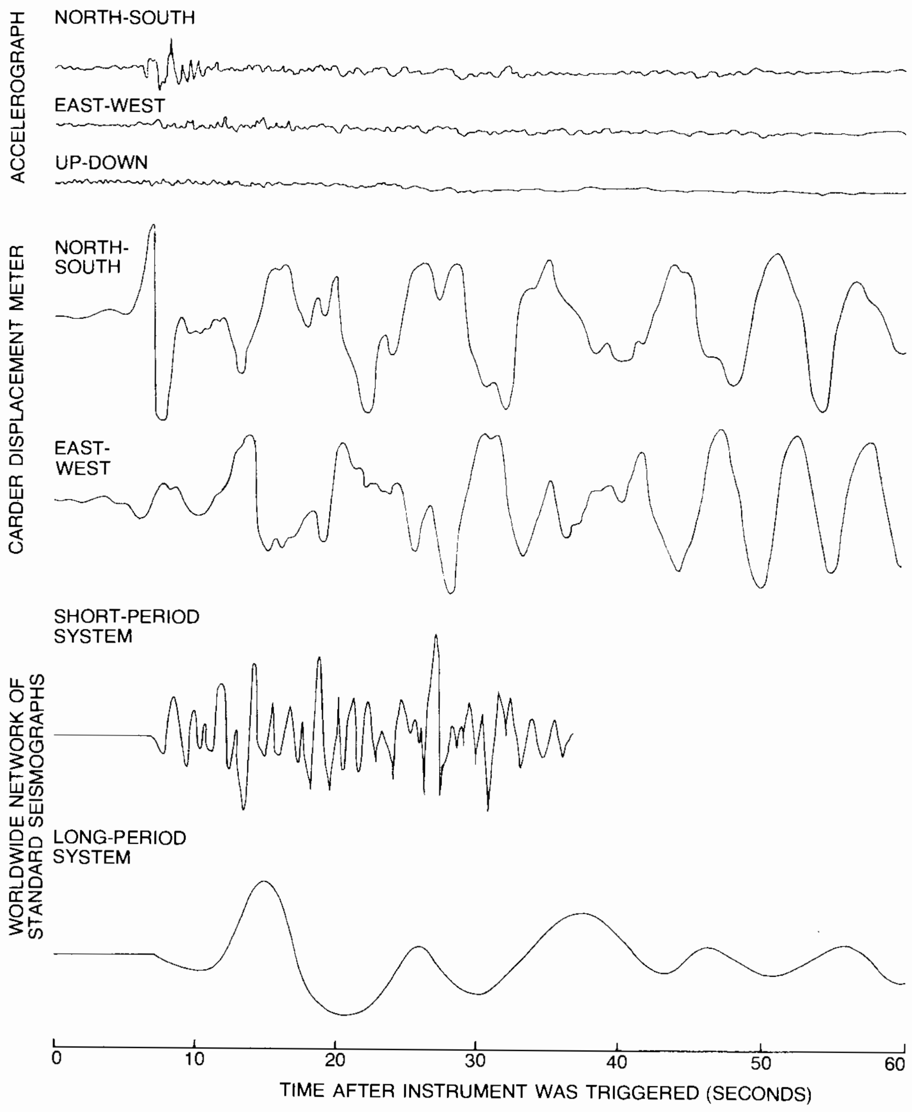
Each of the two torque couples can be visualized as a pair of small spheres, with a thin wire attached to each sphere. The wires are pulled with equal force in such a way that one pair of spheres rotates in one direction while the other pair rotates in the opposite direction. The magnitude of the rotary force—the torque—exerted by each pair of spheres on the elastic medium is the moment. Since the two torque couples rotate in opposite directions, however, no net torque is applied to the medium. The two torques nonetheless deform the medium, radiating elastic waves in a characteristic pattern: a pattern identical with the one in which an earthquake source radiates seismic waves. From this model the moment of the seismic radiation emitted by earthquakes can be calculated. The model has been named the double-couple source model.

The seismic moment measures the seismic energy emitted from the entire fault and not from just a portion of the fault, so that it is a fundamental measure of the magnitude of an earthquake. Hiroo Kanamori of Cal Tech has developed a new magnitude scale based on the seismic moment. The new scale extends the standard Richter scale so that it can accurately measure the strongest earthquakes without becoming saturated. For example, both the San Francisco earthquake of 1906 and the Alaskan earthquake of 1964 had a surface-wave magnitude of 8.3, but the seismic moment of the Alaskan earthquake was 100 times greater than that of the San Francisco one. On Kanamori's scale the magnitude of the San Francisco earthquake has been demoted to 7.9 and that of the Alaskan earthquake has been advanced to 9.2. The strongest earthquake on record is the Chilean earthquake of 1960, with a surface-wave magnitude of 8.3 and a seismic-moment magnitude of 9.5.

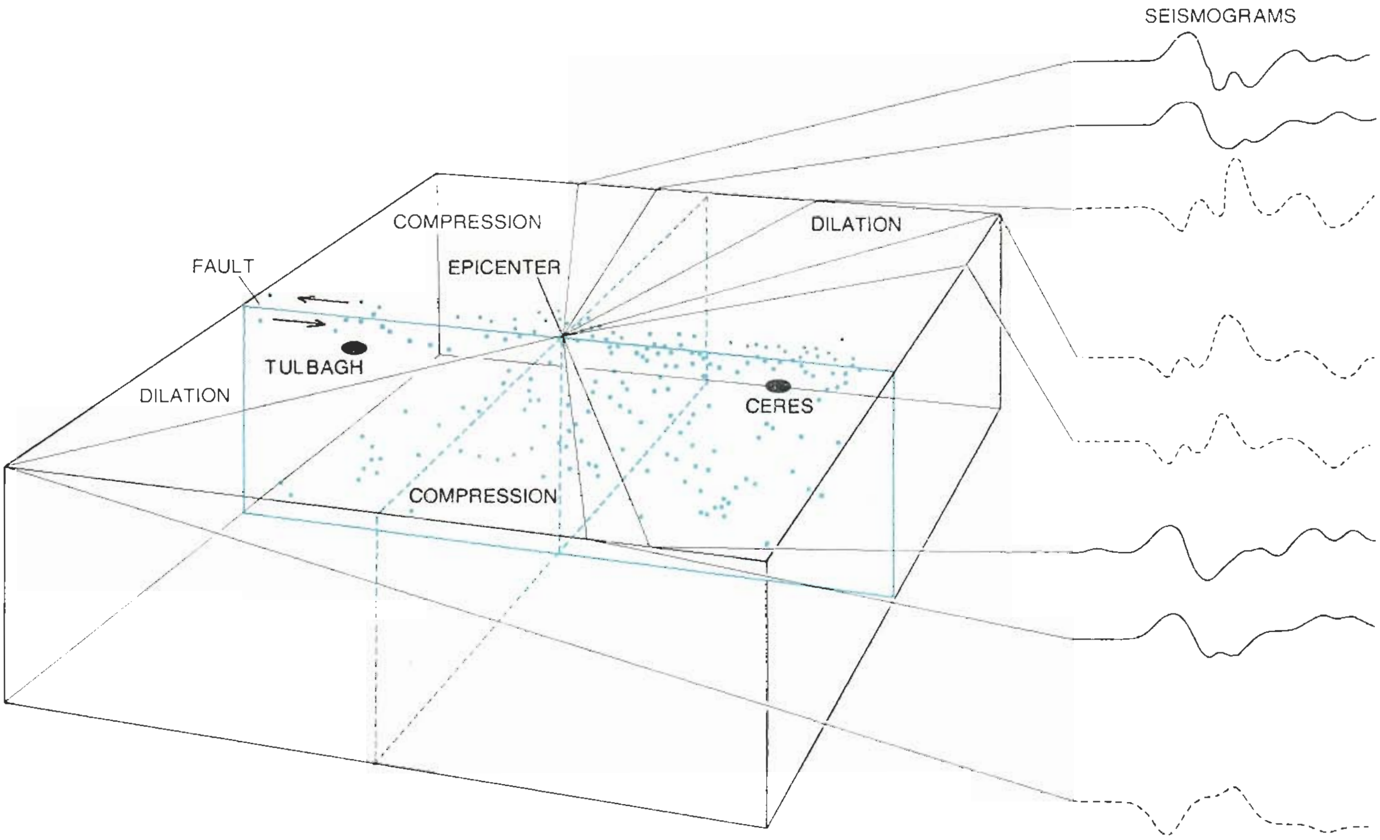
Seismic moment is more than just a convenient scale by which to rank earthquakes according to their magnitude. In 1966 Keiiti Aki of the Massachusetts Institute of Technology showed that the seismic moment is equal to the product of three factors: the average slip of the fault, the area of the rupture and the rigidity of the material that is faulted. Thus if one has independent measurements of the area of the rupture and the rigidity of the material, one can determine the average slip of the fault. The correlation between the average slip of a fault and the average strength of the resulting earthquake provides useful criteria for designing structures such as highways and pipelines that must cross active fault zones.

The total amount of slip accumulated from a number of earthquakes over a period of time also enables one to estimate the velocity at which the tectonic plates bounding the fault are moving past each other. By comparing that velocity with the velocity computed from independent geological, magnetic and geodetic evidence, it is possible to determine how much of the relative motion of the plates gives rise to earthquakes and how much gives rise to aseismic creep. It seems that in some areas, for example Chile, all the motion between plates is accomplished by earthquake slippage, and that in other regions, for example the Marianas arc in the western Pacific, the motion is accomplished by long-term steady creep.

The seismic moment and the dimensions of the fault also yield information about the amount of stress across the fault that is released during the earthquake. The drop in stress is only weakly



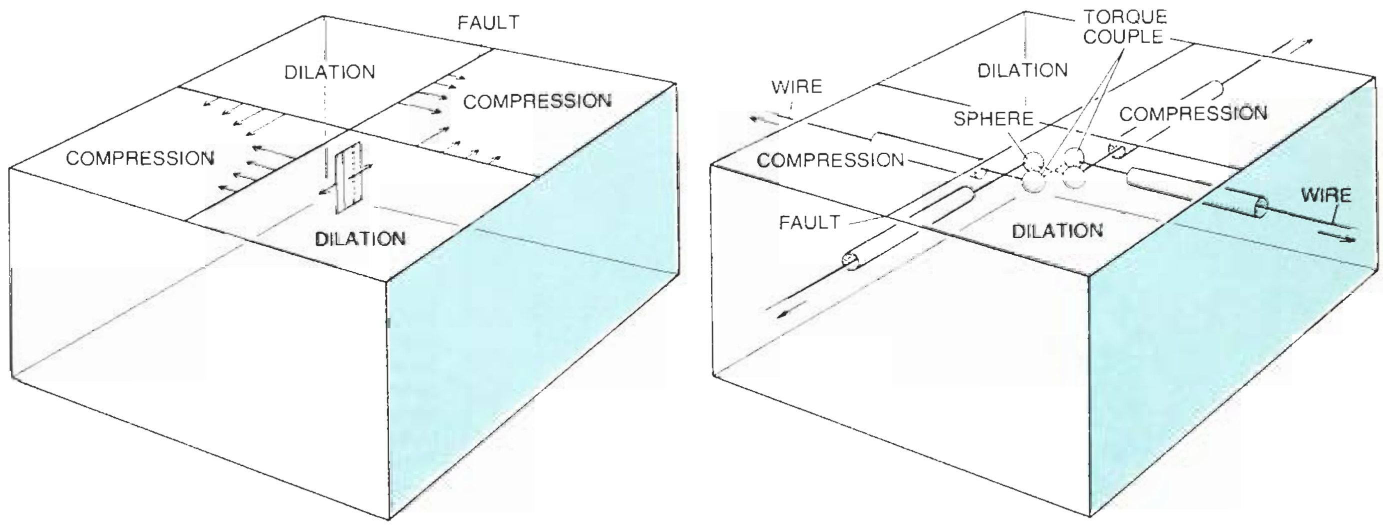
TYPICAL SEISMOGRAMS recorded by different instruments at the same site during the same earthquake can be remarkably different. The top two sets of curves are the recordings of an accelerograph and a Carder displacement meter at El Centro, Calif., from an earthquake at Borrego Mountain, some 60 kilometers away. Both instruments were triggered by the initial *P* wave, or compression wave, from the earthquake; the first strong pulse on each recording is the slower-traveling *S* wave, or shear wave, which arrived seconds later. The prominent reverberations on the recording from the Carder displacement meter are resonances of the seismic waves in the thick blanket of sediments in the Imperial Valley. The bottom pair of curves is the recording made at La Paz in Bolivia of the vertical component of the initial *P* wave from the same earthquake that was recorded by a short-period seismograph and a long-period seismograph in the Worldwide Network. By the time seismic waves had traveled to La Paz, a fifth of the way around the world, *S* waves (*not shown*) arrived approximately nine minutes later than *P* waves.



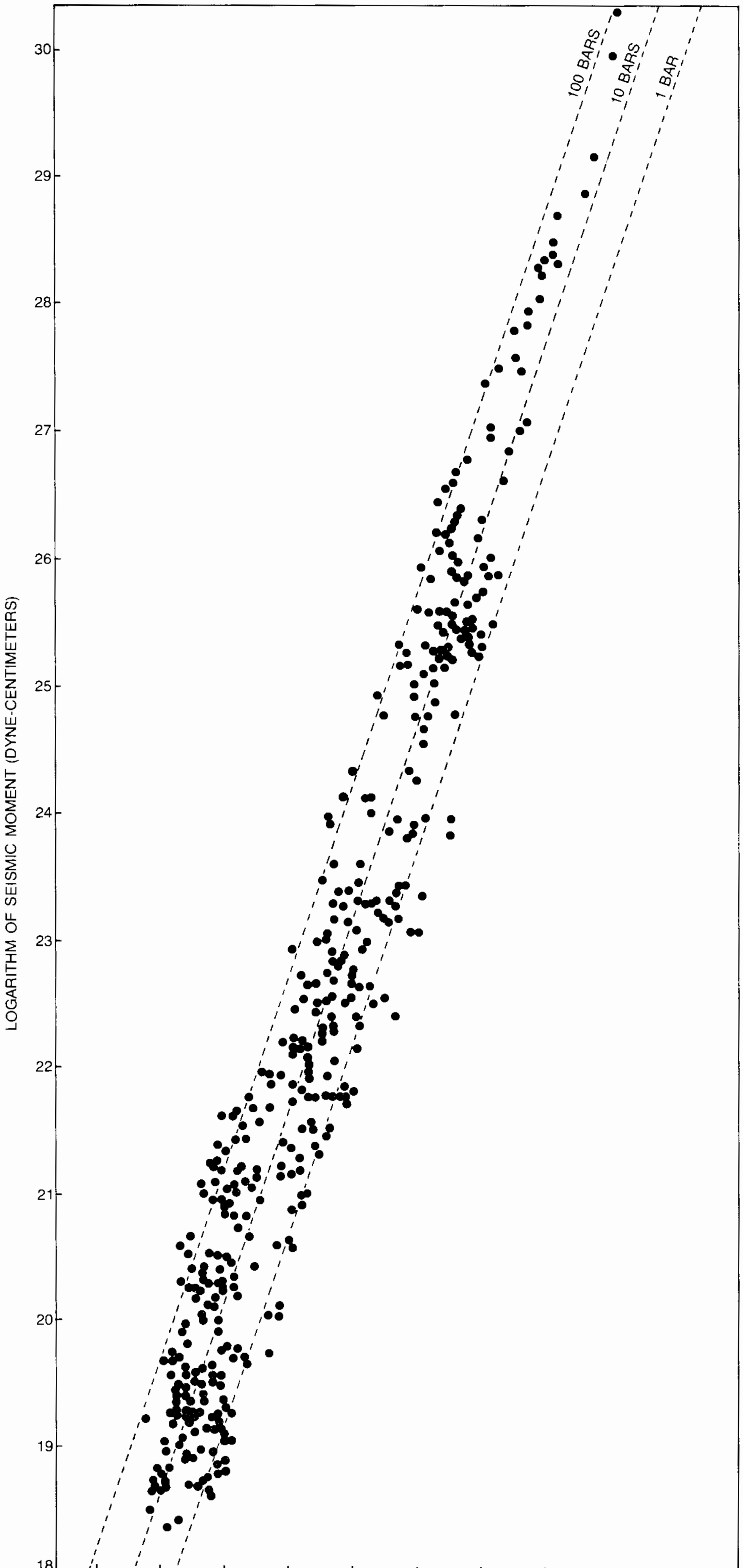
ORIENTATION OF A FAULT below the surface can be detected from the way the ground is initially compressed and dilated around the epicenter of an earthquake. This pattern of compressions and dilations is preserved in the seismic waves that are radiated by the earthquake source. In the illustration portions of seismograms (*right*) recorded during an earthquake near Ceres in South Africa show how the phase of the initial seismic waves received was shifted with azimuth between the source and the recording station. From this information alone the fault could be either of two orthogonal planes: the actual fault (*dark color*) or an imaginary plane perpendicular to it (*light color*). The path of actual fault can be determined from the location of earthquake aftershocks (*dots*) which lie along a single plane.

dependent on the magnitude of the earthquake. Most measurements during large earthquakes indicate that the drop in stress is between 10 and 100 bars. (A bar is 15 pounds per square inch.) The absolute, or total, stress on the faulted material could be considerably higher, but the radiated seismic waves are influenced only by the change in the stress across the fault and not by the absolute stress. Why the drop in stress should be essentially constant for earthquakes spanning such a great range of magnitude is under active debate: the explanation probably lies in the physical properties of the materials within the fault zone and in the forces driving the lithospheric plates.

The properties of the earthquake sources I have discussed so far have been deduced from seismograms made



DOUBLE-COUPLE SOURCE MODEL is mathematically equivalent to the slippage of an earthquake fault. When a small fault slips (*left*), the material closest to it slips more (*longer arrows*) than the material farther away (*shorter arrows*). Thus the material around the fault is compressed and dilated. The same deformation pattern can also be obtained if opposite torques are exerted on two torque couples embedded in an elastic medium (*right*). A torque couple can be visualized as a pair of spheres with a wire attached to each sphere running through a frictionless tube to exterior of medium. When wires are pulled with equal force, elastic medium is deformed in same way as material around a fault. Moment, or amount of torque exerted, is a good measure of strength of earthquake producing the deformation.



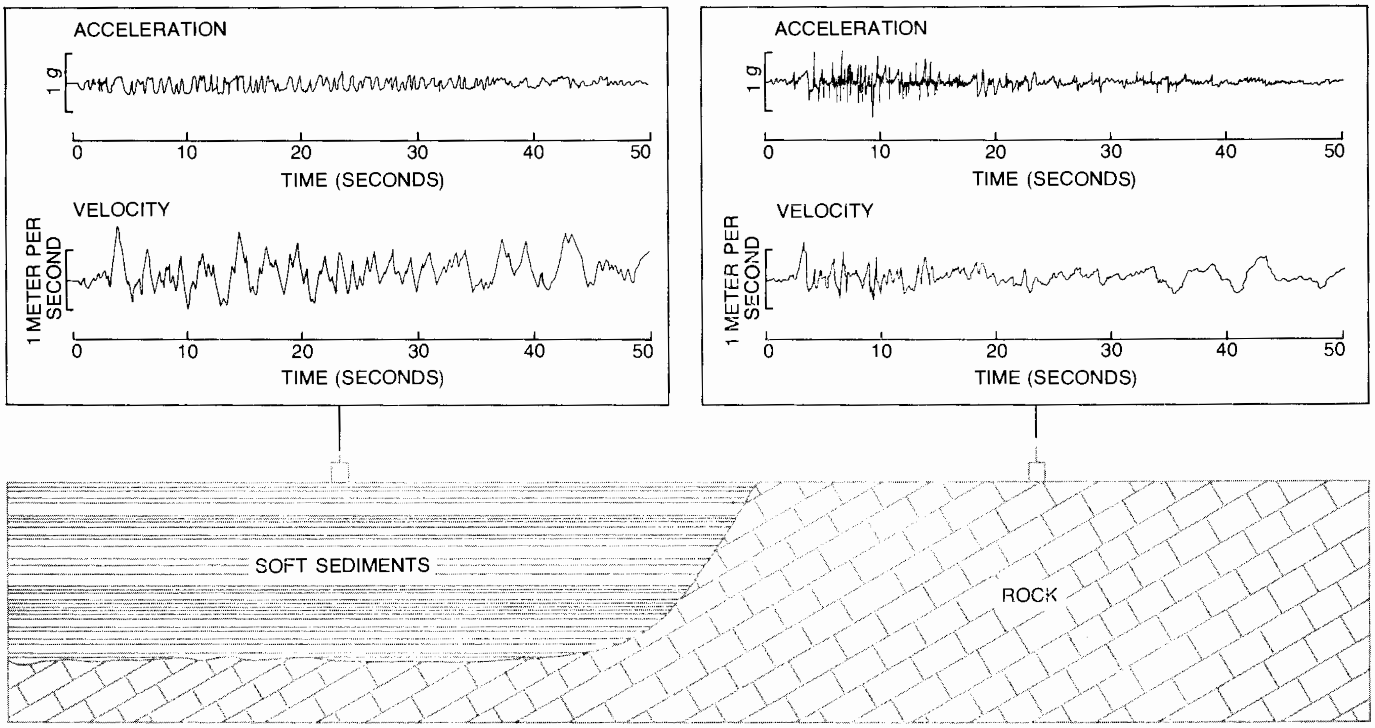
at stations far from the source. Observations at a distance, however, rarely make it possible to resolve the detailed structure of the source. Recently seismologists have been devoting an increasing effort to gaining an understanding of the intricate effects involved in the propagation of the rupture along the fault and in the distortion of the seismic radiation by geological heterogeneities near the fault. This understanding is essential for the design of structures to withstand ground shaking.

Man-made structures are particularly susceptible to earthquakes because the seismic waves have frequencies that coincide with the resonant frequencies of the structures (which range from a tenth of a hertz for large structures such as the Empire State Building up to 30 hertz or even higher for small structures such as systems of pipes in an industrial plant) and because the largest ground motions are usually in the horizontal plane. All buildings are inherently capable of withstanding large vertical forces (at least 1 *g*, or the force exerted on them by the earth's gravity) but special precautions must be followed in earthquake country to ensure adequate resistance to large horizontal forces.

In general the most destructive ground motions have wavelengths smaller than the dimensions of the earthquake fault. Therefore the ground motions are strongly influenced by the details of the rupture process, such as the speed at which the rupture travels over the fault surface, the frictional strength of the fault and the drop in stress across the fault. Geological heterogeneities in the path of the seismic waves can also affect the waves' amplitude and frequency; a seismogram recorded at two stations close to each other may differ significantly. In the past seismologists have rarely been lucky enough to have a good distribution of seismographs close to the source of a major earthquake, and the few seismographs that have been close to the fault have usually been shaken so violently that the recording pen was thrown off the paper. Accordingly the short-period seismic waves are not as well understood as the long-period ones.

In recent years several types of inex-

DROP IN STRESS across a fault during a large earthquake seems to be independent of the strength of the earthquake. The dots represent measurements of seismic moment obtained during many earthquakes with respect to the size of the rupture in square kilometers. Stress drop is inferred from measurements. Lines of constant stress drop are shown. Scatter in measurements for smaller earthquakes may be due in part to experimental error. A bar is a unit of pressure equal to 15 pounds per square inch; a dyne is a unit of force required to impart an acceleration of one centimeter per second per second to a mass of one gram.



LOCAL GEOLOGY AFFECTS GROUND MOTION near the recording site. The waves propagating from the hypocenter up to the earth's surface slow down as they encounter the deformable rocks near the surface, and in general their amplitude increases in much the same way that the amplitude of an ocean wave increases as it approaches the shore. When soft sediments are subjected to strong shaking, however, the amplitude of the motion can actually be reduced. Seismograms at the right are hypothetical recordings of acceleration and velocity of ground for an area underlain by hard rock. Seismograms for a nearby area underlain by sediments (*left*) show that the ground moves faster but amplitude of its acceleration is less. Acceleration is given in terms of *g*, acceleration of gravity at earth's surface.

pensive, rugged and reliable low-magnification instruments have been designed and installed in large numbers near many earthquake faults. The most widely used instrument is the accelerograph, which measures the acceleration of the shaking ground. There are now more than 1,200 accelerographs on station in California alone. Even with so many instruments now in operation we still do not know much about the ground motions close to a fault during a severe earthquake. So far only two useful recordings of an earthquake of magnitude 7.0 or greater have been obtained within 40 kilometers of a fault, and one of them was obtained during an earthquake in the U.S.S.R. To a large extent this lack of data is due to the fact that there have been no large earthquakes in the U.S. in the four or five years since most of the accelerographs were installed.

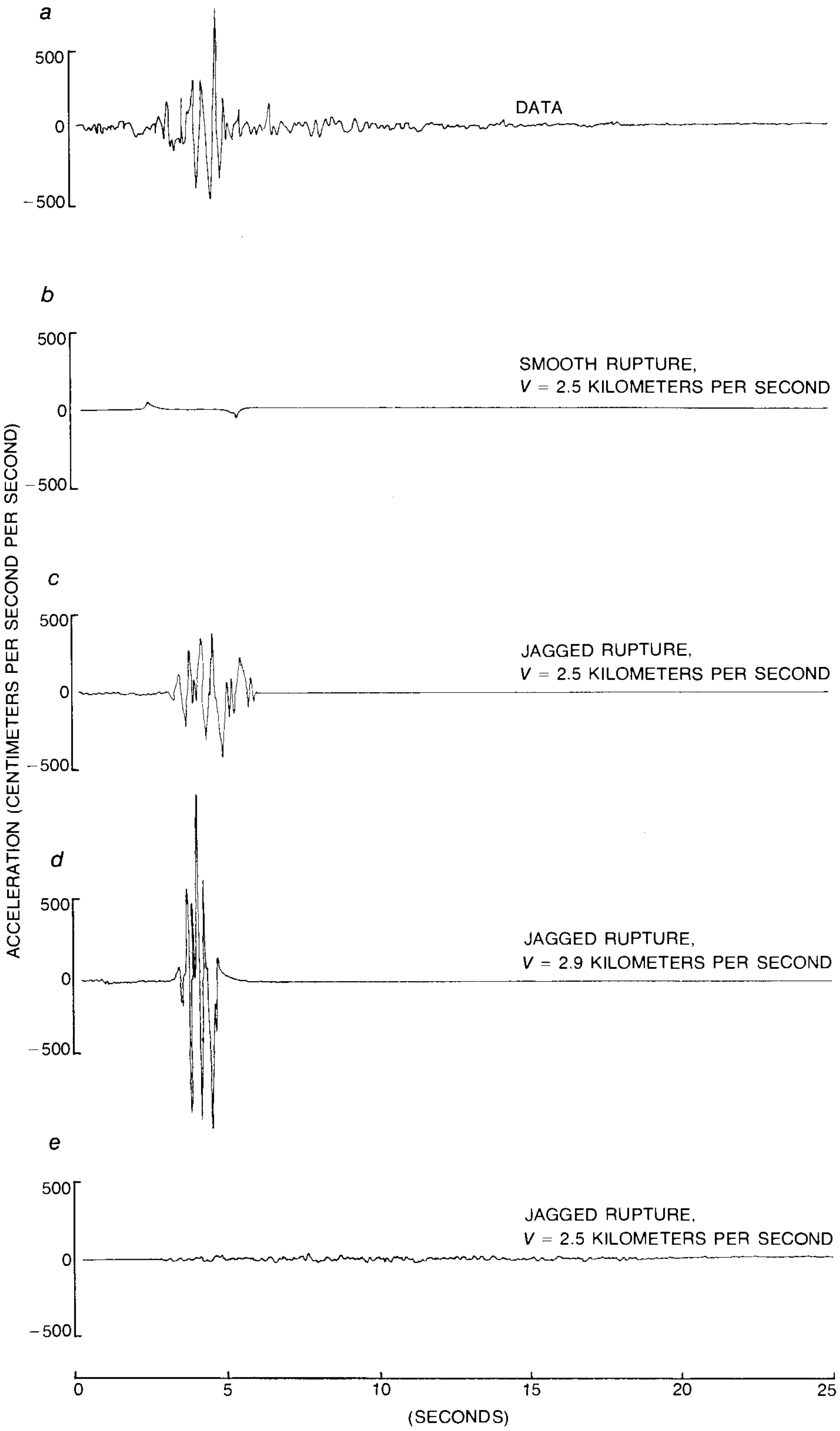
The measurements that do exist have been the main resource for estimating the strength of the ground motion. The few recordings close to faults have had a disproportionate influence on earthquake engineering design, even though these data may not be truly representative of the motions close to future earthquakes. As might be expected, the few close-in recordings have received intensive scrutiny. For example, an accelerograph on a rock abutment near the Pacoima Dam in California during the San Fernando earthquake recorded a peak acceleration of nearly 1.5 *g*, the largest acceleration yet recorded near an earthquake. The record was obtained in a region of exceptionally rugged cliffs and hills, and numerical simulations of the propagation of the seismic waves suggest that the topography may have amplified the ground acceleration by as much as 50 percent with respect to the motions that would be expected on flat ground.

At distances beyond 10 or 20 kilometers from the fault there are a fair number of recordings for earthquakes of magnitude less than 7.0. It is convenient to study the peak acceleration of the ground, expressed in terms of the acceleration of gravity at the earth's surface (*g*), which can be measured directly from the accelerograph records. The peak acceleration expected is widely used by engineers to specify the ground motion a structure should be able to withstand. The peak acceleration of the ground decreases with distance from the fault, both because the seismic waves spread out as they propagate away from the source and because their energy is attenuated by the slight inelasticity of the rocks through which they propagate. Between 20 and 200 kilometers from the fault the peak acceleration decreases approximately as the inverse square of the distance from the fault.

The data that have been obtained 20 kilometers or more from the fault imply that the peak acceleration of the ground is correlated with the earthquake's magnitude. Contrary to what one might expect from the definition of magnitude, however, earthquakes differing by one unit of magnitude do not generate peak accelerations differing by a factor of 10. Moreover, the few available data obtained some 10 kilometers from the fault indicate that very close to an earthquake, peak acceleration is hardly correlated with magnitude at all. For example, an accelerograph close to a fault near Oroville, Calif., recorded a peak acceleration of .6 *g* during an earthquake of magnitude 3.4 but another instrument near a fault in the Imperial Valley recorded a peak acceleration of only .4 *g* during an earthquake of magnitude 7.1.

The lack of correlation between peak acceleration and magnitude is easily understood. The seismic waves measured by accelerographs have a dominant frequency of about four hertz, much higher than the frequency at which the magnitude of the earthquake is measured. For all earthquakes but the smallest, seismic waves with a frequency of four hertz have a wavelength much shorter than the dimensions of the fault. Thus peak acceleration is not a good measure of the strength of large earthquakes, and for the same reason that magnitude is not. The duration of the ground motion is probably much better correlated with earthquake strength.

Strong-motion seismology is a new



SYNTHETIC ACCELEROGRAMS were constructed on a computer by the author and William B. Joyner of the U.S. Geological Survey in order to determine experimentally how an earthquake generated observed ground shaking. An actual accelerogram is shown at the top (*a*). If the earthquake were produced by a smooth rupture of the fault propagating toward the theoretical seismographic station, its accelerogram would consist of a few simple isolated peaks corresponding to the radiation emitted as the rupture started and stopped (*b*). The peaks are small because the rupture was constrained to have a gradual acceleration and deceleration at the ends of the fault. Actual data, however, generally show a more continuous shaking. To simulate this shaking random fluctuations were added to the amount by which the fault slipped. The resulting theoretical curve looked more like the actual data (*c*). Next the author let the rupture propagate toward the theoretical accelerograph at a velocity close to the velocity of seismic waves in the surrounding material. The seismic radiation from the fault then arrived in a sharp peak (*d*). When rupture propagated away from theoretical accelerograph, however, the seismic radiation was spread over a longer time interval and its amplitude was reduced (*e*).

disiscipline, and there are many unknowns in it. In the future data obtained by means of accelerographs and other instruments close to earthquake faults should provide information about both the complexities of earthquake sources and the ground motions they generate. The theoretical and computational models of the seismic source should also improve. Such information will be of direct value to engineers designing major structures. Until that information is available, however, architects and engineers must continue to design buildings on the basis of the few data that do exist, some simple theoretical scaling arguments and plain educated guesses.

Seismologists studying models of the strong ground motions near faults are just beginning to recognize that many of the problems facing them have an essentially statistical character. For years engineers designing major structures in earthquake zones have treated accelerograms of short-period motions as recordings of random noise. On that basis they have devised many ways to generate random series of short-period motions that look much like the accelerograph recordings. The random series were generated in such a way that they matched certain constraints derived from existing data, but they paid scant attention to the physics of the earthquake source. This engineering approach is certainly a reasonable first approximation on which to base the design of a building, but it is of little value in determining what is actually happening below the ground.

Seismologists, on the other hand, have tried to predict the ground motion from earthquakes purely on the basis of deterministic models. In these models earthquake sources have been idealized as simple faults in layer-cake geological structures. Such deterministic models have been relatively successful in predicting only the long-period components of the ground motions.

Clearly the time has come to merge the engineer's statistical view with the seismologist's deterministic one. A number of seismologists are now attempting such a synthesis. Predictions of the ground motion are, however, only as good as the statistical distributions incorporated into the model and physical knowledge of the earthquake source: the properties of the fault surface and of the surrounding rocks and soil. For that information we must not only study existing strong-motion recordings but also draw on other fields such as rock and soil mechanics. I foresee an exciting future in which the skills and the learning of many disciplines, ranging from classical seismology to soil engineering, are combined to gain a better understanding of the nature of earthquakes and to reduce the hazards they create.