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Earthquake Prediction

Recent developments reopen the question of the predictability of earthquakes.

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A few years ago the subject of earthquake prediction fell under the purview of astrologers, misguided amateurs, publicity seekers, and religious sects with doomsday philosophies. No wonder that the occasional scientist who ventured an opinion on the subject did so with trepidation and then with conservatism lest he be disowned by his colleagues.

The situation has changed dramatically in the past 3 years. Three of Japan's foremost earth scientists (*1*) proposed a program of research and concluded that "after ten years the amount of data should be fairly adequate for earthquake prediction." These experts were less certain about the prospects for organizing an efficient forecasting service and properly deferred consideration of this question until after the conclusion of the research program. In the United States an Ad Hoc Panel on Earthquake Prediction was appointed by the Office of Science and Technology. After some 15 months of deliberation the panel saw enough possibilities to justify a 10-year program of research, which they proposed to Donald Hornig, the director of OST (*2*). Their report is now under consideration by several government agencies.

In examining why earthquake prediction is now not only respectable but highly recommended, we review some of the exciting recent developments in seismology. In the United States these advances stem primarily from the International Geophysical Year and Vela projects, which stimulated the design and deployment of advanced instruments and led to new interpretation techniques based on extensive use of electronic computers. Systems procedures for large-scale field instrumentation were developed, and many new workers were brought into seismological research. In Japan the long history of observation of strains, tilts, displacements, and microseismicity in epicentral regions finally produced some evidence of changes in these parameters prior to and following earthquakes.

Seismology is an advanced science in Japan, and because of the severe earthquake hazard most workers in the field consider earthquake prediction the ultimate goal of their discipline. American seismologists are becoming inceasingly concerned about the catastrophic effects of a major shock in one of the populous Pacific coast states. Very few specialists doubt that California will eventually be visited by a major shock. Without going into an analysis here of the potential damage and casualties, one can appreciate the risk involved from Fig. 1, recently published by Allen *et al.* (*3*). It depicts the epicenters of the main shock, some 35 destructive aftershocks, and almost 100 damaging aftershocks of the 1960 Chilean earthquake, overlain to scale on a map of California.

The Earthquake Mechanism

Characteristics of the hypocentral region. The hypocentral region of an earthquake—the region where the earthquake is initiated—may never have been seen. Visible surface faulting sometimes occurs for large earthquakes. Deformation prior to and following the event has occasionally been documented, and strain jumps recorded at large distances have been reported. Precision locations of aftershocks and the radiation pattern of seismic waves are available for many large tremors. These diverse observations yield the following picture of the hypocentral region and fault zone.

Tocher's data on fault length as a function of magnitude, supplemented by additional data, are shown in Fig. 2 (*5*). It may be seen that the length of faulting associated with destructive earthquakes ranges from about 20 kilometers for events of magnitude 6 to about 1000 kilometers for shocks of magnitude 8¼. Rupture velocity has been measured for several large earthquakes, and values of 3 to 4 kilometers per second seem typical.

The vertical extent of faulting has been debated by seismologists. The best documented case for a great earthquake is that of the Alaskan shock of 1964, where it was found that faulting probably extended from near the surface to depths of 100 to 200 kilometers (*6*). The strain and stress jumps in the source region can be inferred approximately from the displacements revealed by surveying or by sea-level changes. By representing the fault as a dislocation sheet, the stress (and strain) distributions needed to match the observed displacements may be obtained. When one assumes a finite rectangular sheet and strain change of 10^{-4} to 10^{-3}, surprisingly small stress jumps of 10 to 100 bars are obtained (*7*). For the Alaskan earthquake, with magnitude of 8¼, numerical inte-

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gration of the strain energy density yielded an energy release of 10^{25} ergs, a value which is roughly checked by the empirical conversion of magnitude to equivalent seismic energy. This value is equivalent in seismic energy to 100 nuclear explosions, each of 100 megatons (*8*). Many seismologists accept Tsuboi's hypothesis that the strain energy density is independent of magnitude, the energy release and magnitude being determined by the source volume.

The seismic energy release generally decreases with depth, reaching zero at about 700 kilometers. Roughly 80 percent of the energy is released in the depth range 0 to 60 kilometers; the hydrostatic pressure range which corresponds to this depth range is 0 to 18 kilobars. A recent study (*9*) showed that, in southern California, most of the shocks occur at depths of about 5 kilometers (Fig. 3). It is important to note that the main focal region of California is accessible by drilling! In Japan, most major shocks occur at depths of less than 60 kilometers; more than half occur at less than 30 kilometers.

The earthquake mechanism. The focal region of earthquakes has remained inaccessible to direct observation, so theories of the earthquake mechanism are based on indirect observation—on (i) the movements of surface rocks above the actual focal region, (ii) the behavior of samples of rock stressed in the laboratory under the high-pressure and high-temperature conditions found in the earth, and (iii) the radiation pattern of seismic waves.

The great 1906 earthquake in California was accompanied by widespread surface movement of rocks. Study of these movements led to Reid's elastic rebound theory for the earthquake source mechanism (*10*). According to this theory an earthquake is the result of strain release caused by sudden shearing motion along a fault. This view of the origin of earthquakes is supported by the nature of seismic signals which emanate from many earthquakes, and is also reasonable in terms of laboratory behavior of rocks. If a rock such as diabase is stressed under the confining pressure appropriate to the shallow crust, it fails by faulting (Fig. 4). The failure is sudden and usually explosive. Jaeger (*11*) and Byerlee (*12*) have studied the characteristics of sliding on faults and artificial surfaces under pressure. These studies give some insight into the probable shallow-earthquake mechanism.

Typical sliding behavior of a fine-grained granite (*12*) is shown in Fig. 5. A cylindrical sample 12.5 millimeters in diameter and 31.7 millimeters long contained an artificial surface in an orientation close to the direction along which the rock would naturally fault. Compressive stress was applied to the ends of the sample, which was under confining pressure, until sliding occurred. A force-displacement curve (Fig. 5) shows that sliding was not smooth but was punctuated by sharp stress drops. Elastic shocks accompanied the stress drops, the magnitude of which ranged in this particular experiment from 50 bars to 2.5 kilobars.

The behavior of materials in experiments such as these recalls Bridgman's observations (*13*). He found that jerky stick-slip accompanied the shearing of a great variety of materials, even when normal pressures were as high as 50 kilobars.

These observations suggest that, in terms of the Reid mechanism, earthquakes could be produced by sudden stress drops during sliding. The stress drop probably does not represent complete release of stress at the earthquake focus. In other words, the stress released during an earthquake is only a small fraction of the total stress supported by the rock around the fault.

The Reid mechanism will probably be enhanced in regions of abnormally high pore pressure. High pore pressure tends to lower the frictional resistance to sliding; in an extreme case, when pore pressure equals overburden pressure, resistance on a horizontal fault would drop to the very

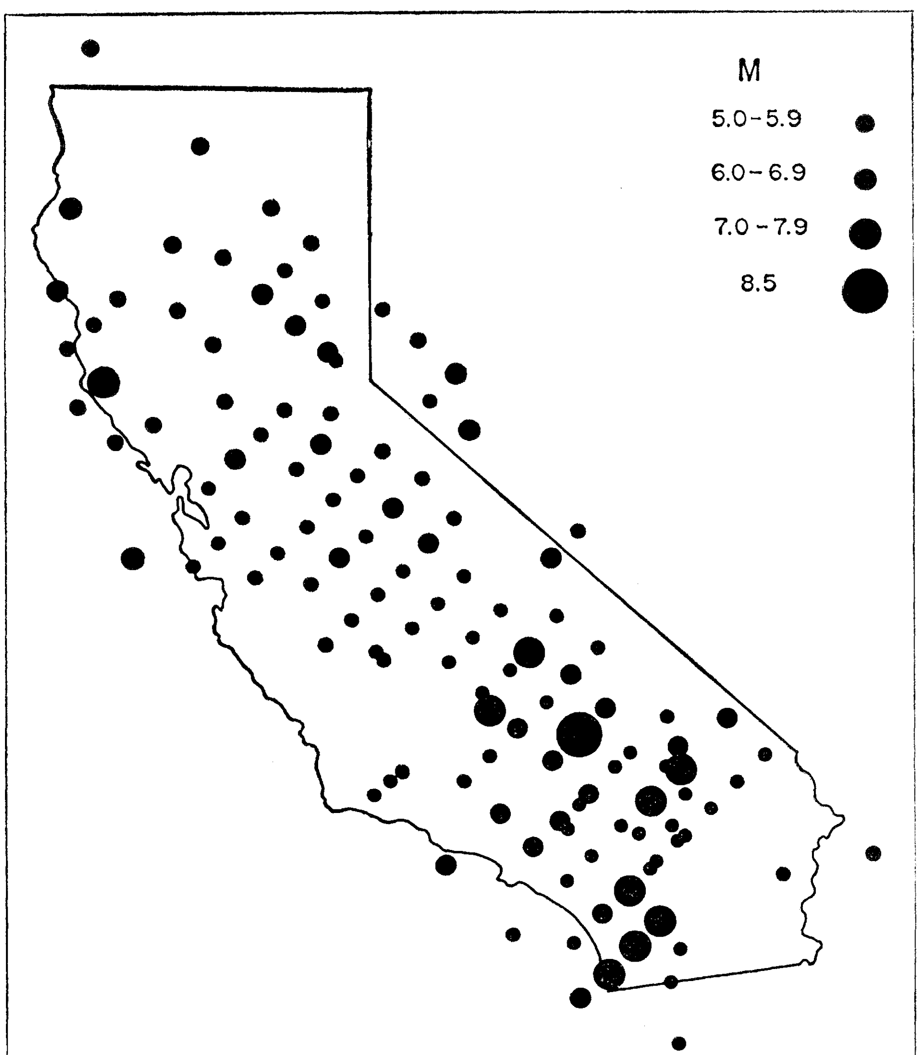


Fig. 1. Epicentral distribution of the major aftershocks (within 6 months) of the great Chilean earthquake of 1960, superimposed on a map of California at the same scale. (*M*) Magnitude. [From Allen *et al.* (*3*)]

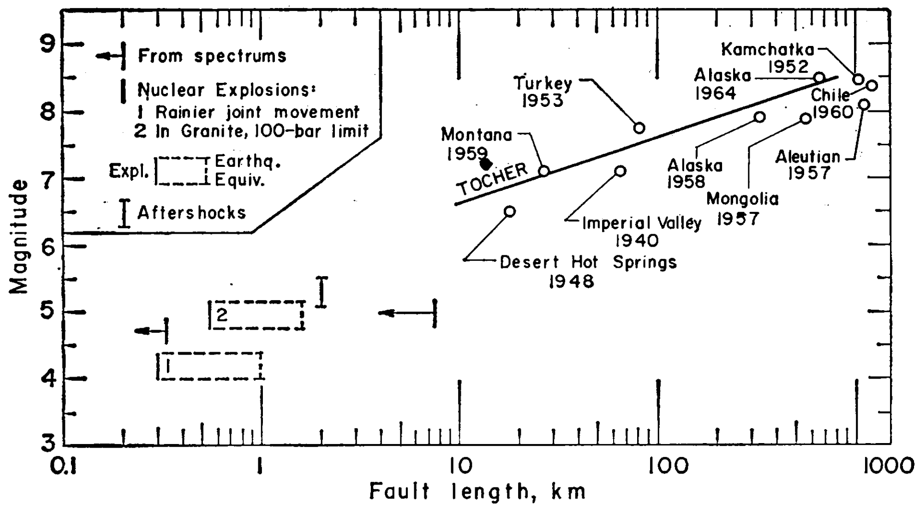


Fig. 2. Empirical data on fault length versus magnitude.

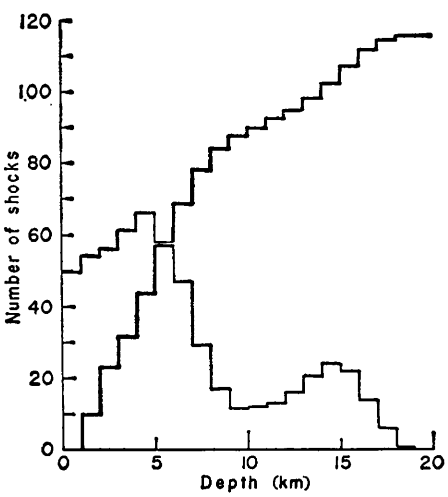


Fig. 3. Frequency distribution (top curve) and cumulative distribution (bottom curve) of focal depths for earthquakes in southern California. Magnitude range, 4 to 5; time interval, 1950 to 1963. Divide the vertical scale by 5 for frequency distribution.

small value of the intrinsic strength. In regions of high pore pressure, then, faulting could be caused by relatively low tectonic stresses. Pore pressures have actually been measured which are within a few percent of overburden pressure; however, these are typically in nonseismic regions.

The influence of pore pressure may have been revealed in the recent seismic activity near Denver, Colorado. Many small earthquakes which occurred at depths of about 5 kilometers seem to be correlated with pumping of liquid wastes down a well 4 kilometers deep. Although the role of the injected fluids is not entirely clear, one hypothesis consistent with available information associates the earthquakes with increased pore pressure in the vicinity of a fault (*14*).

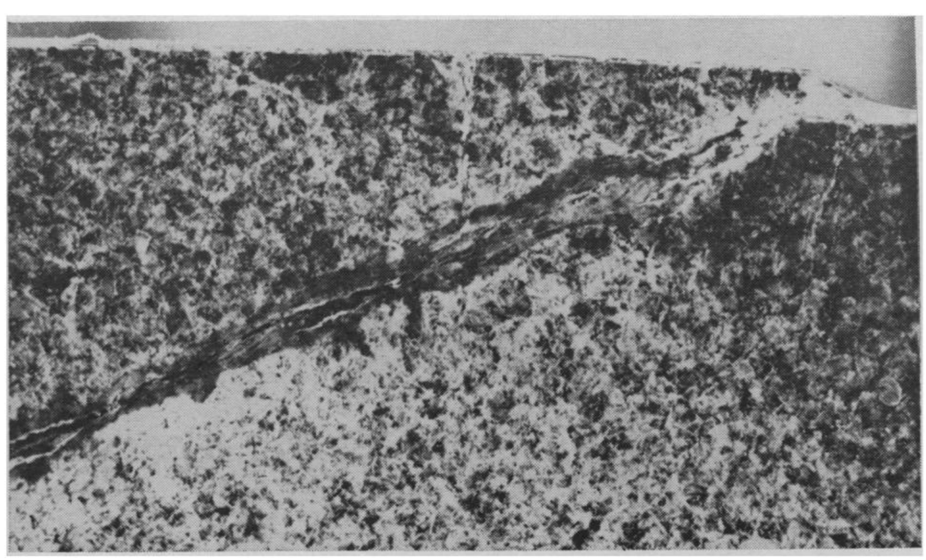


Fig. 4. Photograph of a fault in diabase produced in the laboratory at confining pressure of 3.2 kilobars. The direction of maximum compression was parallel to the long dimension of the sample; the short dimension is about 12.5 millimeters.

It is also possible that, in regions of abnormally high pore pressure, the Reid mechanism may be important to considerably greater depths than the shallow crustal depths considered here. Faulting and other behavior which one expects at shallow depths may occur at considerably deeper levels when pore pressure is abnormally high. This was shown in a recent study, by Raleigh and Paterson (*15*), of the effects of heating on rocks containing hydrous minerals. Embrittlement of serpentinite occurred at pressure of 5 kilobars and temperature of 700°C as water released during dehydration raised the pore pressure. Fractures formed in the course of these experiments and, therefore, might form in the earth, but it does not necessarily follow that jerky slip would also occur under these conditions. Nevertheless, it is tempting to consider the possibility, for breakdown of hydrous minerals must be a widespread phenomenon (*16*) throughout the deeper crust and even into the mantle. At these levels, however, the Reid mechanism would certainly be competing with some of the higher-temperature models which have been proposed.

Griggs and Blacic (*17*) have observed a drastic lowering of strength of silicate minerals which were deformed in the presence of water; the effect was attributed to the hydrolysis of the silicon-oxygen bonds. This weakening has been observed in a material as strong as quartz at temperatures as low as a few hundred degrees. In rocks under natural conditions, water weakening might so lower strength that sudden shearing failure would occur locally.

Orowan (*18*) has proposed an earthquake mechanism in which creep fracture plays an important role. According to his model, creep can occur through viscous grain-boundary sliding; at an advanced stage of creep, cavities form, a mechanical instability develops, and the rate of strain suddenly increases within certain narrow bands. This sudden increase in strain rate might produce an earthquake, much as the sudden frictional sliding in the Reid mechanism does. In the creep model, however, no actual frictional sliding takes place at the source. This model has not been tested experimentally for rocks.

Orowan (*18*) regards the Raleigh-Paterson effect (*15*) as due to lubrication by wet dehydrated serpentine and

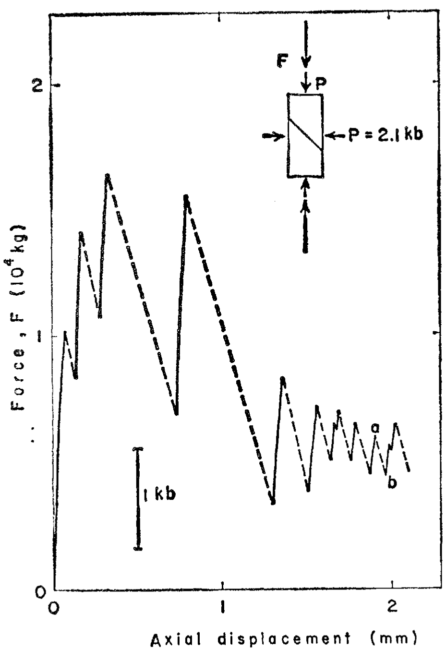


Fig. 5. Force-displacement during sliding on a ground surface in Westerly granite (*12*). Confining pressure, *P*, was 2.1 kilobars, and the loading arrangement used is suggested by the small figure at upper right. The position of the ground surface is shown by the inclined line in the small figure. The dashed parts of the curve are sudden stress drops; the true form of the curve during these stress drops is not known. The vertical bar shows axial stress difference in the specimen.

suggests that, throughout the region of 20 to 60 kilometers depth, strength could be lowered to a few hundred bars by decomposition of serpentine. Faulting might easily occur at these depths and cause earthquakes; stress drops would be of the correct magnitude. Only the largest faults which would originate at these depths would penetrate the upper 20 kilometers, where frictional resistance is high.

Premonitory Indications of an Earthquake

Forewarning of a large earthquake might come from three sources: from (i) tilts and strains in the epicentral region, (ii) the general increase in number of small seismic events, and (iii) changes in physical properties of rocks near the fault as they are strained. Some of these features have actually been observed prior to earthquakes, and others are suggested by consideration of possible source mechanisms.

Observed deformation before an earthquake. Anomalous deformations preceding earthquakes in Japan have been reported (*1*). These include anomalous changes in sea level (hours to days before the earthquakes), as reported by the public or as evidenced on tide gages. They include vertical and horizontal deformation, as revealed by precisely repeated surveys, the anomalous shifts being detected some months to years before the earthquake, depending on the frequency of the surveys.

Instrumental indications of anomalous tilt and strain changes preceding earthquakes by hours or days have also been reported (*19*). Perhaps the most remarkable and best documented case occurred prior to the Niigata earthquake of 1964. This earthquake of magnitude 7½ was not one of the great ones of recent years, but it was very destructive. An anomalous strain change was first detected some 9 hours before the earthquake. The sensing instruments were vertical strainmeters installed in shallow holes over a zone 10 kilometers long. This net was designed to measure subsidence associated with the withdrawal of natural gas. A vertical expansion of the ground of 0.3 to 0.4 millimeter was detected by 15 out of 20 instruments, corresponding to a strain change of 10^{-5} in the 9-hour period (*19*). A sample record is shown in Fig. 6. The instruments were approximately 70 kilometers from the epicenter and 20 kilometers from one end of the aftershock zone.

An example of anomalous ground tilting is shown in Fig. 7. The station at Ikuno was 60 kilometers from the epicenter, in this case the Tottori earthquake of 1943, magnitude 7½. This record, obtained by Sassa, is famous in Japanese seismology and represents the first instrumental indication of possible deformation prior to an earthquake (*19*). Both examples are suggestive but not uniquely indicative of the creep acceleration stage of creep fracture (*18*).

Tantalizing as the few records indicative of prior deformation are, their validity and reliability must be tested. This calls for a much larger number of case histories, involving use of large numbers of sensors, each operating continuously over long periods of time under conditions of minimal interference from extraneous sources. Both the Japanese and the U.S. programs incorporate proposals to test these preliminary results. These tests range in scale and technique from precise surveying with modern electronic devices to measuring with arrays of sensitive strainmeters and tiltmeters, some of them located in deep holes closer to, if not in, the possible focal region.

Seismic activity. A program of monitoring small shocks in seismic belts

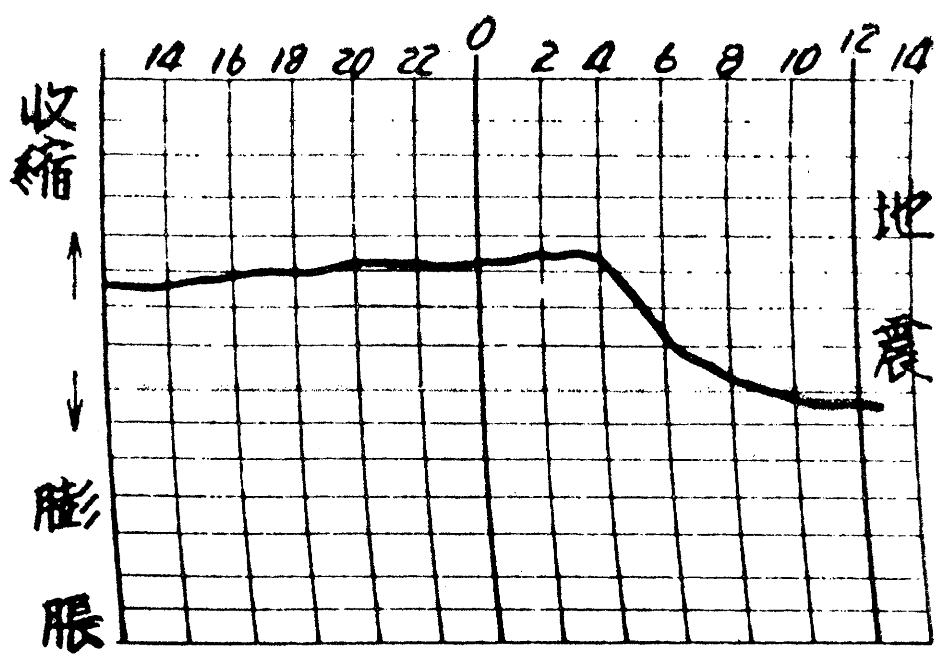


Fig. 6. Trace motion of vertical extensometer located in a well 40 meters deep near Niigata, showing anomalous expansion (downward direction on the ordinate) beginning 9 hours before the earthquake. Each vertical space represents 0.1 millimeter. The abscissa gives the time in hours, beginning at 12 hours on 15 June 1964. The trace is interrupted by the earthquake (*19*).

can be useful in different ways. The radiation pattern of seismic waves leads to a definition of the regional stress pattern. The shocks themselves define seismic belts and sometimes specific faults. Some recent field studies show that regional microseismicity is not necessarily positively correlated with earthquake probability in a tectonically active region (*3*). However, it is possible that the pattern of seismic release preceding an earthquake may be a forecasting element, as implied by the laboratory experiments described earlier.

As an example, we might cite Watanabe's study (see *20*) in which he subjected rocks to uniaxial compression and used ultrasonic detectors to sense the small shocks associated with local brittle fractures. He measured strain directly, and also the cumulative square root of the energy released by the shocks. These results are shown in Fig. 8 as a function of time. The strain of the specimen and the strain release by the shocks both show anomalous change prior to rupture. This is a well-known characteristic of fracture of rocks in compression, particularly at low confining pressure (*20, 21*).

A comparison of the release pattern of aftershocks and foreshocks, by Suyehiro *et al.* (*22*), for a small, perceptible earthquake suggests the possibility that different regimes may occur (Fig. 9). It may be seen that the foreshock activity has relatively fewer small shocks than the aftershock sequence—an observation which suggests a possible premonitory effect.

Physical changes in rock which precede faulting. Although rocks in laboratory experiments frequently fault without much warning, a number of subtle premonitory changes do in fact occur. Many of these changes can be detected in the laboratory at stresses well below the maximum stress a rock can support. In the natural situation, these changes might warn of an impending earthquake.

The most obvious effect which precedes faulting in laboratory experiments is the increase in microseismic activity (*20, 21*). This suggests either that the rock is cracking on a small scale or that surfaces within the rock are sliding on one another. Cracks have actually been observed to form on the surfaces of stressed samples (*23*), and sensitive measurements of volume change suggest that the new cracks are open (*24*).

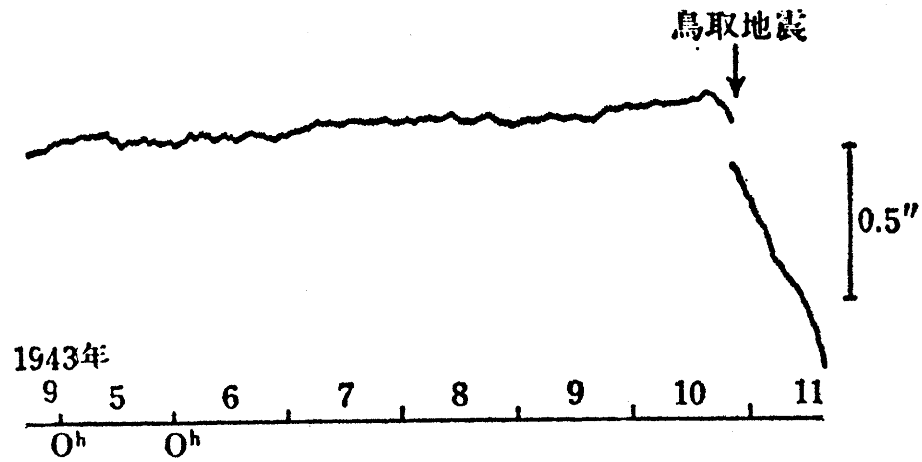


Fig. 7. Record, obtained at the Ikuno station, showing anomalous tilt that preceded the Tottori earhtquake of September 1943. The abscissa represents days; the arrow indicates occurrence of the earthquake (*19*).

Cracked rock has elastic and electric properties different from those of rock which is uncracked, particularly if the cracks are open. Moreover, the new cracks are apparently strongly aligned, in a direction parallel to the maximum compression. Thus, the changes in elastic and electrical properties ought to be markedly directional. Some observed changes in laboratory samples are shown in Fig. 10.

Compressional wave velocity was measured in two directions as a specimen of granite at 3.6 kilobars was stressed in compression (*25, 26*). Not much change in the two velocities occurred at stresses up to about half the maximum stress; beyond that point the velocity in a direction perpendicular to the maximum compression began to decrease. At the maximum stress, when the rock faulted, the velocities parallel and perpendicular to the maximum compression differed by about 20 percent. Electrical properties of rock saturated with water changed

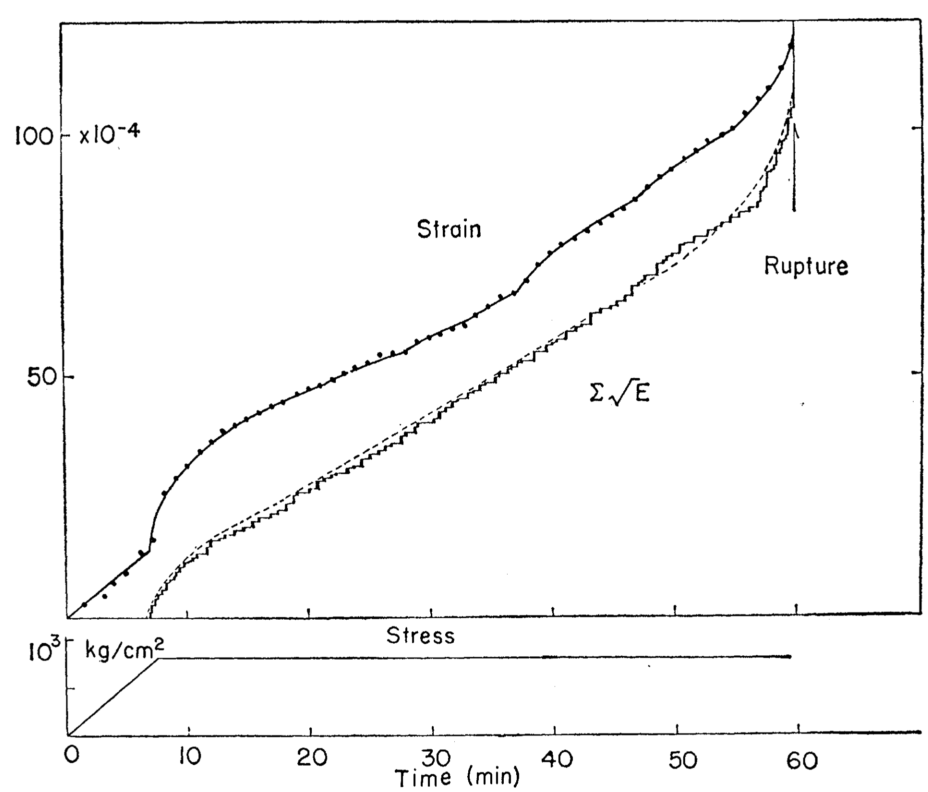


Fig. 8. Laboratory results for strain and cumulative square root of energy released by microfractures in granite, showing premonitory indication of rupture. [After Watanabe (*20*)]

a great deal more than this. Brace and Orange (*27*) found that resistivity in the direction of the maximum compression drops by a factor of 5 to 10; an example is given in Fig. 10 for granite fractured under an effective confining pressure of 1.6 kilobars.

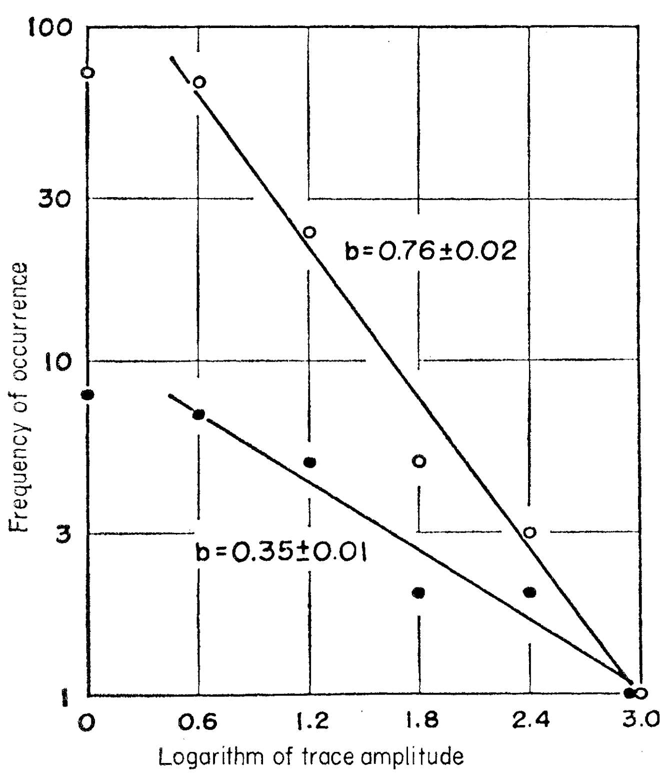
Whether these results can be extended to seismic regions of the earth is, of course, open to question. Although these phenomena are observed in the laboratory over a wide range of pressures, the temperatures and the physical character of rock in the natural environment may affect the changes in some unknown way. For example, the fissured, broken condition of rocks in the natural environment, particularly in seismic areas, may lower resistivity to such an extent that the changes that we observe in the laboratory for rock which is initially intact may not be detectable.

The small size of the stress drops associated with earthquakes may also make the task of detecting changes in physical properties quite difficult. The changes in elastic and electrical properties shown in Fig. 10 occurred for a stress change of several kilobars. This stress change is a great deal larger than the 100-bar stress drops associated with earthquakes. The changes in elastic or electrical properties due to a 100-bar stress change would be of the order of a tenth the size of the changes shown in Fig. 10. Velocity and other elastic properties might change by perhaps a few percent; it is doubtful that changes this small could be detected. On the other hand, with a 100-bar stress change, changes of resistivity might amount to 50 percent; this should be detectable.

It is interesting to note that earth resistivity changes much smaller than these have apparently been detected (*28*); the changes were associated with tidal deformation. Also, small velocity changes of the correct sign have been reported (*29*). Statistical treatment of a number of scattered observations suggested a correlation of anomalous seismic velocities with time intervals preceding strong earthquakes. The anomalous P-wave velocities were about 12 percent lower than normal, and the anomalous V_p to V_s ratio was 6 percent lower than normal.

Earthquakes caused by creep fracture should, one would think, be preceded by a period of accelerated surface strain which might be detectable. A severe and damaging earthquake due to a fault in a region of serpentinization might be preceded by a series of small shocks (*18*), which would represent the deep faults that do not penetrate the surface. Experience may reveal a critical surface-strain accumulation associated with damaging faults of this kind.

Other possible indicators. If one takes the point of view that an earthquake-prediction research program should test all physical parameters which are responsive to changes in stress, or the physical-chemical state of rocks, or to the mechanism of failure, then a number of additional observations suggest themselves. There may be a critical stress, strain, or strain rate associated with earthquakes in a given area. Water-well levels are sensitive to dilational strains as small as 10^{-9} to 10^{-8}. Permanent changes were produced in the water levels of wells in the southeastern United States following the great Alaskan earthquake of 1964. Variations in the geomagnetic field (or in its gradient) may occur in response to changes in magnetic susceptibility or electrical conductivity, or they might occur if the Curie point were shifted. Earth currents, either natural or artificial, might be even more sensitive indicators since they directly respond to resistivity changes which, in turn, may indicate the buildup of stress, as described above.



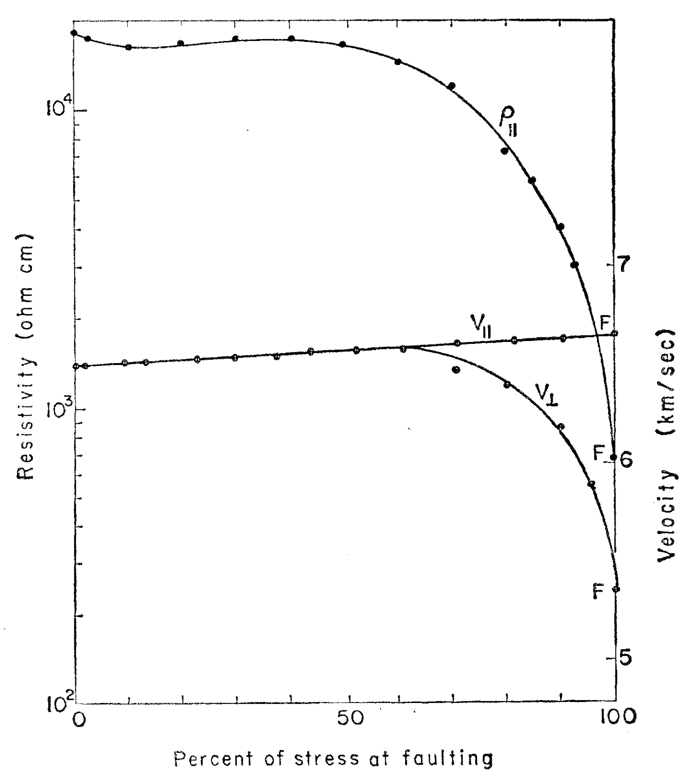


Fig. 9 (left). Relation between frequency of occurrence and magnitude in foreshocks (solid circles) and aftershocks (open circles) of a small earthquake in Central Japan (*22*). Fig. 10 (right). Electrical resistivity, ρ, and compressional wave volocity, V, as a function of stress. The resistivity data are for Westerly granite at pressure of 1.6 kilobars (*27*). [After Matsushima (*26*)]

Field Investigations

It is not our purpose, in this article, to design an earthquake prediction program. However, it seems obvious that a major feature of such a program would be the monitoring, with the greatest achievable sensitivity, of all possible indicators foretelling the occurrence of earthquakes. Networks of instruments would be deployed in seismic belts and would be operated continuously over long periods of time in such a way as to provide the greatest possible likelihood that many earthquakes would be "trapped" within the arrays. Although this is essentially an empirical and somewhat wasteful approach, the absence of a confirmed theory for the earthquake mechanism justifies it. All pre-earthquake activity predicted by the main theories of the earthquake mechanism would be tested, as would the phenomena reported, by reputable scientists, to have preceded earthquakes.

The new generation of instruments. Seismological observations are primarily concerned either with detecting propagating seismic waves or with sensing secular strains, tilts, and displacements associated with strain accumulation and release in seismic belts. For many years now the detection of seismic waves has been limited by background noise emanating from winds, sea waves, and industrial vibrations and not by instrumental sensitivity. At some frequencies of the seismic spectrum the noise can cause displacements as large as 10^5 angstroms. Seismometers are currently available which can detect ground motion as small as 1 to 10 angstroms. Advances in seismic wave detection have come through locating seismometers in deep wells at depths of several thousand meters, or through deploying arrays of large numbers of sensors. The instruments in deep wells gain in sensitivity because the "skin" effect associated with elastic surface waves (which comprise seismic noise) leads to a rapid decrease in noise amplitude with depth. Array processing leads to an increase in the signal-to-noise ratio by a factor somewhat better than the square root of the number of sensors. Optimum frequency filtering and wave-number filtering techniques are employed—an adaption of methods developed in radar signal processing. The filters exploit differences in frequency and phase velocity between signal and noise. The signal is typically coherent over the array; this is not always the case for the noise.

A large-aperture seismic array (LASA) was recently placed in operation under the auspices of Project Vela (*30*). This array, depicted in Fig. 11, consists of 21 clusters, each composed of 25 seismometers buried at depths of 30 meters. The array covers a circular area of about 200-kilometer diameter. The LASA contribution to the earthquake forecasting problem lies, perhaps even more than in the improved signal-to-noise ratio, in its systems approach to a large-scale multisensor field experiment. LASA demonstrates how hundreds of detectors distributed in a seismically active region can be tied together by microwave and telephone-line telemetry in which signal data are transmitted to on-line-monitoring computers. We discuss below how this systems concept can be modified to provide an essential tool in a prediction research program.

The tools for testing seismic activity as a forecasting element are arrays of extremely sensitive seismographs located in seismic belts, monitoring microearthquakes on an on-line basis. Some of the sensors would be in deep holes, possibly as deep as 5 kilometers. The procedure of monitoring the microearthquakes and characterizing the pattern of release would, of necessity, be automated, since the system would be dealing with thousands, if not millions, of events.

Observations of strain and tilt are also limited by ground noise. In this case the noise derives from thermal stresses associated with diurnal and seasonal temperature changes in the rock, from loading of the earth by rainfall, from changes in sea level, from varying barometric pressure, and from wind. These noise sources can produce strain changes of as much as 10^{-16} over a period of weeks to months. Smaller, short-period fluctuations (hours to days) also occur. The semidiurnal tides contribute strains of the order of 10^{-8}. Compare these disturbances with secular strain rates of 10^{-6} to 10^{-5} per year which might be expected in tectonic belts, or with the strain drop of 10^{-4} inferred for the epicentral region following an earthquake. Premonitory strain changes could be smaller than 10^{-4} by several orders of magnitude.

A strainmeter of the type developed by Benioff (*31*) can detect short-term strains, associated with seismic waves, as small as 10^{-10}. Strain changes of 10^{-9} to 10^{-8} corresponding to strain jumps in the epicentral region of large, distant earthquakes have also been detected (*32*). Whether this sensitivity is adequate for detecting premonitory strain changes is not yet known. If greater sensitivity is needed, noise reduction methods must be devised. These would include miniaturizing of strain seismographs so that they could be placed in deep holes. Deployment of arrays of strain seismographs is another possibility for improving signal-to-noise ratios.

The use of laser interferometers, modulated light beams, and microwave phase measurements over distances of 1 to 10 kilometers has often been proposed. The advantages of long-range observations have never been demonstrated, but is not unreasonable to suppose that local and extraneous strains which affect conventional strain seismographs would tend to be reduced, whereas regional strains associated with earthquakes would be emphasized. In a recent analysis, Gehrels (*33*) concluded that distances can be measured over 10 kilometers with a precision of 1 part in 10^7 by means of modulated-light-beam techniques which will soon be practicable. The effect of atmospheric temperatures is reduced by employing beams of different colors and by using the dispersion properties of the atmosphere. A strain of 10^{-7} is about 10^{-2} to 10^{-1} times the annual strain accumulation in a seismic belt and is 10^{-3} times the strain drop associated with major earthquakes.

Laser interferometers offer even greater sensitivity, but with certain restrictions. Long-term stability of the laser frequency of several parts in 10^{10} is currently attainable (*34*). This would be the instrumental accuracy of a laser strain seismograph in which the light beam traverses a controlled atmosphere, as in a buried light pipe. Practical considerations limit the use of such a device to strain sensing over a distance of a kilometer, and this may be all that is needed in a monitoring system. However, if sensing over distances between 1 and 10 kilometers is needed, this may be achieved by building arrays of shorter strainmeters or by using long atmospheric paths. In the latter case multicolor dispersion methods could be used to correct for atmospheric temperature, pressure, and humidity, and strain changes as small as 10^{-9} to 10^{-8} might be observed, provided coherence could be maintained through the open atmosphere. The uncertainty here pertains to beam coherence and the earth noise for these long paths.

Strains and tilts are both derivatives of a displacement field. Mechanical and liquid tiltmeters have been constructed, the latter on the basis of Michelson's concept of liquid flowing between two reservoirs. These instruments are capable of monitoring tilts as small as as 10^{-8} radian (*35*). In seismically active regions like Japan, tilt rates of as much as 10^{-5} radian per year are observed. Tiltmeters are subject to the same limitations as strainmeters—namely, excessive ground noise which restricts instrumental sensitivity. Liquid tiltmeters varying in length from 1 meter to 1 kilometer are currently being constructed.

The only instruments currently available for measuring ground displacements over long periods of time are tide gages and gravimeters. Because of the large, erratic sea-level fluctuations associated with swell, tides, and seasonal effects, it is difficult to see how tide gages could be used in a forecasting system, though they would provide valuable data on regional deformation before and after earthquakes. Gravimeters with long-term stability of 1 part in $10^{-9}g$ (1 microgal) are now available. These instruments are subject to semidiurnal tidal accelerations larger than this by a factor of 200. They respond to motion of the ground in the manner of accelerometers and displacement meters. The displacement effect predominates for periods somewhat longer than half an hour and is due to an equivalent "free-air" or Bouguer gravity change. A sensitivity of 1 microgal corresponds to a change in elevation of a few millimeters when the displacement is due to the addition of mass in the section. Arrays of gravimeters could provide useful information on vertical motions by reducing extraneous effects, such as loading of the crust by sea-level and atmospheric-pressure changes and solid-earth tides. Instruments for sensing horizontal displacements are not yet available, although proposals have been advanced for constructing such a device with the precise accelerometers of inertial navigation instruments as its main component.

Magnetometers operating by atomic processes show long-term stability in the range of ± 0.01 gamma (0.1 microoersted). They may be operated in a gradient configuration; this cancels micropulsations and provides great sensitivity to local changes in the magnetic field. The relation between magnetic field changes and tectonic processes is not yet established. It may involve susceptibility or Curie point changes due to variations in temperature and stress.

Strain-, tilt-, and displacement meters, by their nature, monitor variations subsequent to the installation of the instrument. Thus, their role in an earthquake prediction scheme depends on an anomalous premonitory effect. They could not sense critical values. Although some development is still needed, techniques exist for determining absolute stress *in situ*, a measurement which would be of great importance if critical stresses are involved in the earthquake mechanism. The methods were primarily developed to monitor stresses in mines and tunnels. One procedure involves isolating a specimen of rock (on which strain gages have been mounted) from the stress field in the surrounding rock by overdrilling. The stress relief sensed by strain gages is then related to the preexisting stress field (*36*).

The measurement of temporal variations in seismic velocity and electrical conductivity in earthquake regions should be mentioned to complete the picture. The measurement of conductivity may be of special importance in view of the laboratory results described earlier. The field procedure consists of deploying current sources and detecting voltmeters, with an electrode spacing appropriate for penetration to the focal region.

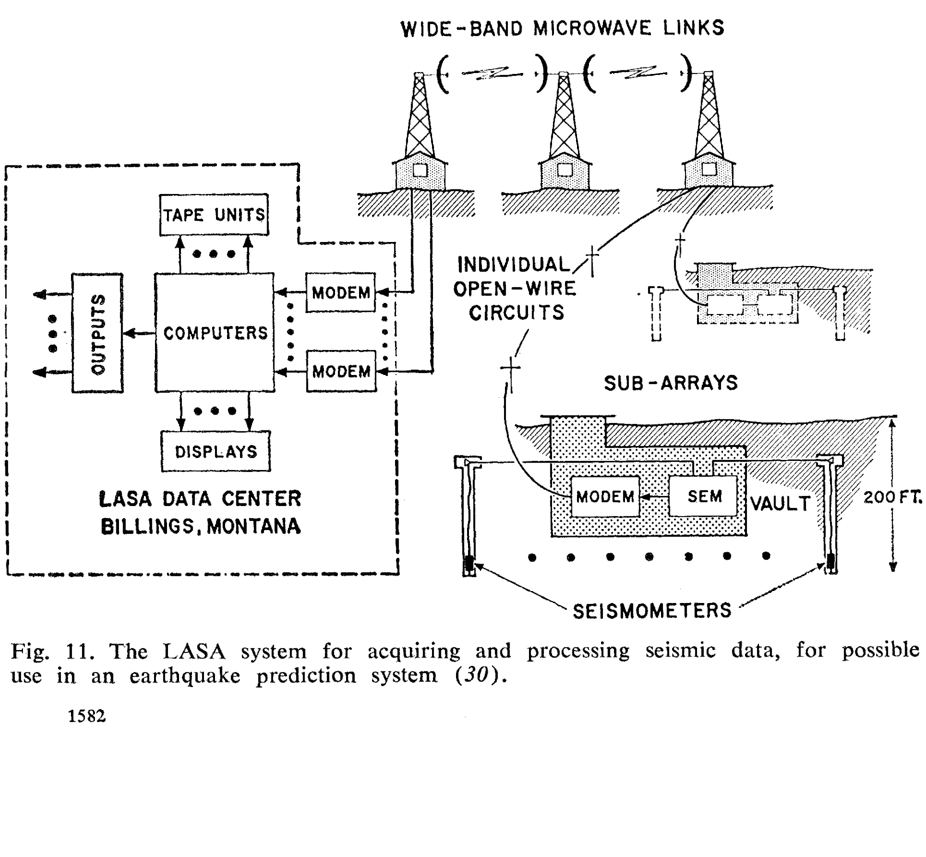


Fig. 11. The LASA system for acquiring and processing seismic data, for possible use in an earthquake prediction system (*30*).

Instrumenting Seismic Belts

The Ad Hoc Panel on Earthquake Prediction proposed that field observations be organized around permanently installed instrumental clusters, augmented by special surveying devices. The clusters primarily monitor local deformation, microseismicity, and gravitational, magnetic, and electric fields. The special surveying devices extend the point observations of the cluster element and also tie the cluster together by monitoring regional deformation. The clusters are sited along major earthquake belts, on the basis of low background noise, auxiliary evidence of a high probability of earthquake occurrence, and consistency with the overall coverage of the seismic belt.

A cluster would consist of the following instruments distributed over an area of 100 to 1000 square kilometers.

1) Microseismicity array: ten sensitive seismographs each buried in a hole about 30 meters deep. The frequency response will be somewhere within the short-period range (1 to 100 cycles per second) and will be selected on the basis of local noise and spectrum of microearthquakes; sensitivity will be the maximum that ground noise permits.

2) Tiltmeter array: ten small two-component tiltmeters. Tiltmeters will be temperature-compensated, will have sensitivity of 10^{-9} radian or better, and will be installed in trenches to

limit the effect of surface disturbances associated with atmospheric pressure and temperature changes.

3) Strainmeter array: ten small three-component strainmeters of strain sensitivity 10^{-9} or better will be installed in shallow holes or trenches. Background noise may make it difficult or impossible to achieve these goals for tilt and strain sensitivity.

4) Two magnetometers (rubidium-vapor or the equivalent); sensitivity, 0.1 gamma. The two horizontal components of the earth's natural telluric field will be measured at the same places. Artificially induced current will also be monitored as an indicator of resistivity changes.

5) Two recording gravimeters; sensitivity, 1 microgal; zero position must be recoverable after motion corresponding to $0.1g$. Two recording meters will permit measurement and comparison of gravity differences of two nearby locations.

6) Meteorological instruments and tide gages to monitor noise sources and forcing functions.

7) Deep-hole seismometers, tiltmeters, and strainmeters located in a borehole at depths of 3 to 5 kilometers. Since background noise is primarily generated at the surface, the operation of deephole instruments at tenfold to 100-fold gains may be possible. Furthermore, this approach offers the possibility (in California, at least) of siting instruments close to the seismic focus, where effects preceding an earthquake would probably be enhanced.

All of the instruments described, except for the deep-hole strainmeter and tiltmeter, will soon be available.

The special surveying devices consist of laser strainmeters permanently installed in each cluster, designed to measure strain changes of the order of 10^{-9} over distances of about a kilometer. These devices, which will soon be available, will probably involve the use of light paths in buried pipes which are evacuated or filled with dry nitrogen. In addition, optical or electronic strainmeters capable of observing regional strains over distances of 10 to 20 kilometers are highly desirable for tying clusters together and for facilitating regional surveying. To develop them will require development of a surveying device designed to operate in air in such a way as to achieve strain sensitivity of 10^{-8} to 10^{-7} after correction—that is, after compensation for changes in air pressure, temperature, and humidity.

Since strainmeters and tiltmeters only measure changes that occur subsequent to installation, it is proposed that absolute stress determinations be made on a daily basis. These observations may reveal the status of a region in the stress-accumulation, stress-release cycle.

The Ad Hoc Panel recommended that these clusters be deployed in Alaska and in the California-Nevada region, the two most seismic areas of the United States. In California-Nevada, a Y-shaped zone is defined by the major earthquakes which have occurred in

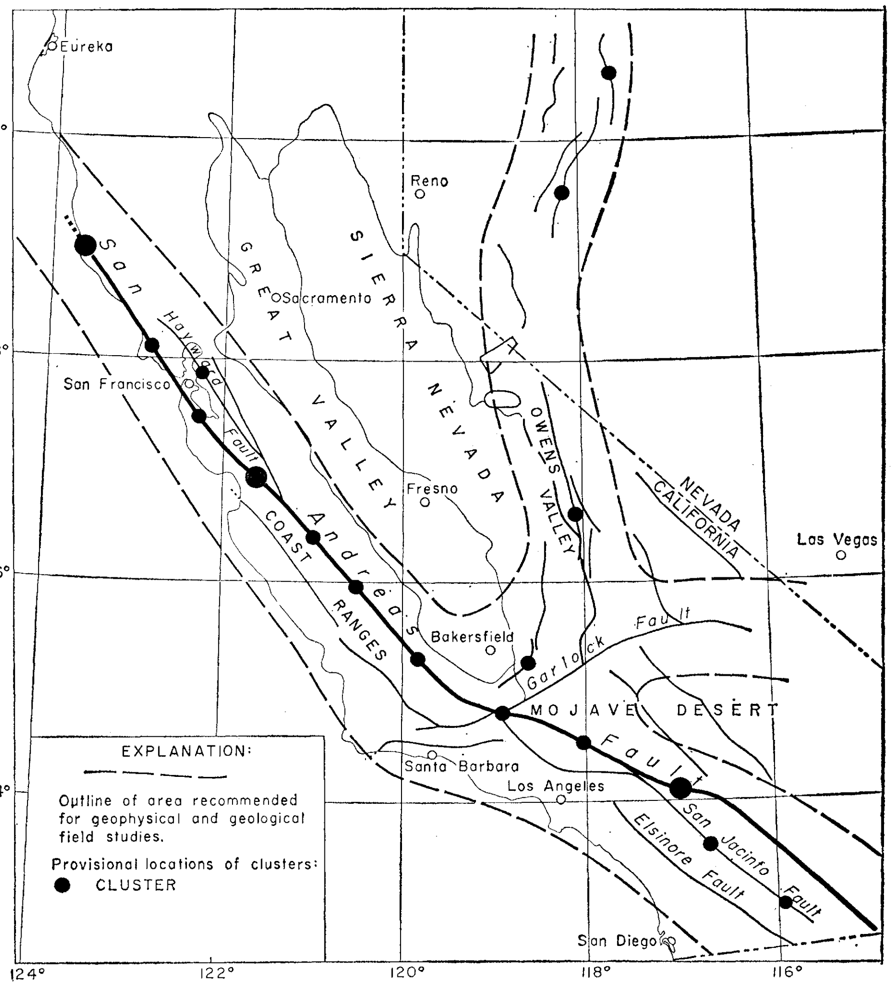


Fig. 12. Possible location of instrument clusters in fault zones of California and Nevada.

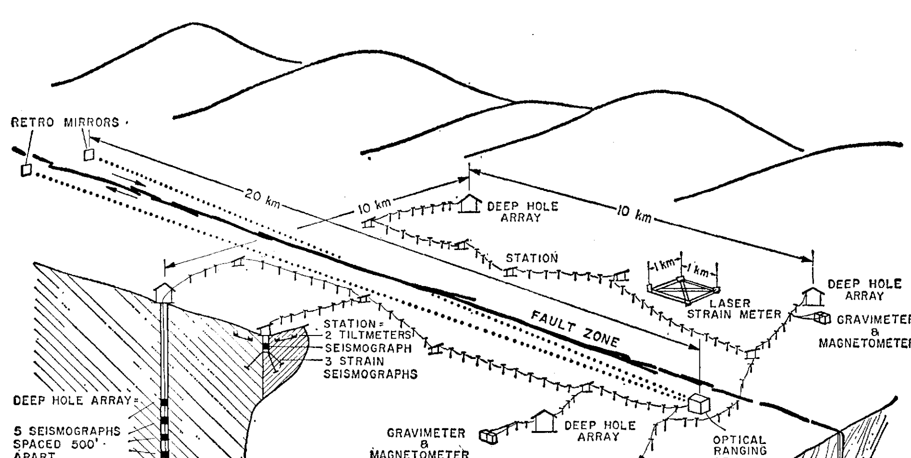


Fig. 13. Preliminary conception of an instrumental cluster.

this region over the last 100 years (Fig. 12). This zone includes such major fault systems as the San Andreas, Garlock, and Owens Valley systems in California and the Dixie Valley in Nevada (east of Reno). Some 15 permanent clusters and one movable one would be spaced at intervals of 50 to 100 kilometers along the San Andreas system and the other faults, as shown in Fig. 12. A schematic view of a cluster is shown in Fig. 13. Also shown in Fig. 13 are the special surveying devices designed to monitor regional strain.

All told, some 1000 to 1500 sensors would be used in the California-Nevada experiment. A system similar to that devised for LASA would be used to automate the data acquisition and analysis. The elements of the clusters and the clusters themselves would be linked by open-wire digital telephone circuitry and by microwave links, all feeding into a central computing and analysis facility. The computers would monitor microseismicity statistics, deformation, and variations in the several internal fields. The sensors would be correlated, among themselves, with external forcing functions such as atmospheric pressure, sea-level changes, tidal forces, diurnal heating and cooling, tectonic stress variations, and earthquakes. Numerical correlation and prediction techniques would be programmed into the computers.

The instrumental program proposed by the U.S. and Japanese panels covers regions which have experienced some of the most destructive earthquakes in recorded history. If these programs are fully implemented, it is reasonable to expect that many earthquakes will occur in localities where instruments to test for premonitory indications have been located. If earthquake forecasting elements occur in nature, these programs have a good chance of finding them.

Earthquake Engineering Research

A successful research program—one which leads to a forecasting system—would make the casualty problem less severe. It would reduce damage caused by fire by alerting electric, gas, and water utilities. However, the economic loss would still be significant, and only a special program of earthquake engineering research can be effective in reducing damage and destruction. Whereas earthquakes may not be predictable, there is no question that engineering research can reduce the vulnerability of urban areas to destruction by earthquakes. For this reason, the U.S. Ad Hoc Panel went beyond its guidelines and included a major program of earthquake engineering research in its proposal. In addition to basic engineering studies, expansion of the more practical aspects of antiseismic engineering would be needed.

In anticipation of advances in prediction capability or in the engineering aspects of the earthquake problem, studies evaluating the effectiveness of the response of urban populations to earthquake disasters are recommended. Even with a reasonably successful forecasting capability, the problems of evacuation, control, and reconstruction are severe ones.

In view of the public concern with earthquakes, it cannot be emphasized enough that the results of a research program such as this one cannot be anticipated. The goal is to go from the present capability of probabilistic prediction to specific forecasting. The problem is unquestionably a difficult one, yet specialists in Japan and the United States are anxious to attack it because they feel that significant contributions can be made. In joining earthquake prediction research with earthquake engineering research, they expect, as a minimum result, an increase in knowledge of fault tectonics and an improvement in construction practices in seismic belts, for reducing casualties and property damage.

Aside from forecasting, the instrumental development and field programs will contribute to some of the major unresolved problems of geology. The continental pattern of deformation, which will be monitored for the first time on a short-term basis, bears on such basic questions as orogeny and continental drift, and the stress sources which produce them.

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