ELEMENTARY SEISMOLOGY AND SEISMIC ZONING

by

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Seismology is the study of earthquakes; sometimes we make our own earthquakes with high explosives or nuclear bombs but for the most part we are dealing with natural earthquakes. While natural earthquakes may occasionally be caused by such action as the impact of meteorites or the collapse of caverns or by landslides, the earthquakes with which you, as earthquake engineers, are concerned are all due to tectonic forces. These forces, acting within the earth are those same forces which are responsible for mountain building and similar geological processes.

Seismologists are not in complete agreement about how tectonic energy is released when an earthquake occurs. Most North American seismologists have been strongly influenced by observations in California, particularly at the time of the 1906 San Francisco earthquake. It will be worth our while to look at what happened there. The San Francisco earthquake was the result of slipping along the San Andreas fault over a length of something more than 200 miles. This slipping was almost purely horizontal and attained a maximum displacement of slightly more than 20 feet. Fortunately, the area of the fault had been surveyed by precise triangulation on two occasions before the earthquake. The survey was repeated immediately afterwards. The story can be reconstructed with the help of Figure 1, which shows a plan view of the fault. It should be pointed out the scale of this diagram is very much distorted, the vertical dimension representing about 20 feet and horizontal dimension about 20 miles. The line O'O" represents the fault, dipping vertically and striking in a north-south direction. Suppose that a line AOB had been drawn at right angles to the fault about 100 years before the earthquake, and suppose that we had been able to watch this line constantly over the years. We would have seen it gradually distorted until just before the earthquake in 1906, it had arrived at the position A'OB'. At that instant the stresses set up exceeded the strength of the fault; the fault broke, the line A'O sprang up to A'O', the line OB' became O"B', and the displacement of the Pacific (western) side of the fault with respect to the continental side, which had taken 100 years to accumulate, suddenly appeared on the fault as the displacement O'O". The strain within the region was completely relieved by this displacement.

North American seismologists and geologists think that this sequence, of stress gradually accumulated and suddenly relieved by the breaking

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of fault, is a reasonable picture of what happens in most earthquakes. It is not necessary that the fault should be vertical nor that the motion be horizontal; this process of stresses gradually accumulated and suddenly released can be regarded as perfectly general. This theory is called the elastic rebound theory.

It should be stated that North American seismologists have probably given too much attention to the phenomena of the San Francisco earthquake because surface-faulting is an exception in earthquakes. Richter has published a list of earthquakes which were accompanied by visible faulting. For the hundred-year period 1850 and 1950 he was only able to find about 55 events. When we realize that in this 100-year period there were somewhere between 1,000 and 2,000 major earthquakes, it is clear that the percentage of earthquakes accompanied by surface-faulting is small. Most earthquakes originate at depths of about 10-20 miles and in certain areas earthquake depths of as much as 450 miles occur. It is difficult to see how stresses can accumulate to large amounts at these depths, one would not expect the rocks to have sufficient strength, and for this reason many seismologists do not accept the elastic rebound theory as applicable -- certainly not at all depths. The question is academic from the point of view of earthquake engineers. We may accept the elastic rebound theory as a satisfactory explanation of what happens in an earthquake.

In passing we may discuss this question of the depth of earthquakes. As was stated above, earthquakes may occur to depths of as much as 450 miles, and what seismologists call a "normal" earthquake has a depth of about 20 miles. The point at which the earthquake energy is released -- the centre of the earthquake -- is called the focus. The point at the surface vertically above the focus is called the epicentre of the earthquake.

The damage which an earthquake causes is clearly related to its focal depth. If an earthquake is deep, no point on the surface is close to the point of energy release but a large surface area is about equidistant from this point. Consequently the damage will be slight but it will be stretched over a considerable area. On the other hand, if an earthquake focus is very shallow the point immediately above it will be very close to the point of energy release and damage will be very heavy here. As we go away from this epicentral area, however, the distance from the point of energy release increases rapidly and, therefore, the energy will drop off very rapidly with distance.

You may find it inconsistent that I talked about a movement along 200 miles of the San Andreas fault, and at the same time talk about the focus as if it were a point. How do we rationalize these two concepts? When a fault breaks, the motion must begin at some point. If this is the main break it is this point that seismic measurements will define as the focus. It may, on the other hand be a minor slip, triggering the main break at some other point. The second, main, break would probably be the focus located, although near-by stations might locate the smaller, triggering shock.

In many earthquakes the main shock is preceded by smaller foreshocks, which may occur hours, or days, before, and all large earthquakes are followed by many aftershocks, usually with gradually decreasing intensity. The number of these aftershocks may range into the thousands and if one plots them they usually block out, in area and in depth, the faulted area involved in the earthquake.

The Character of Elastic Waves

Material such as a rock is said to be elastic if, having been deformed slightly, it returns to its original shape when the deforming forces are removed. For short-term forces, such as those involved in earthquake waves, the earth is a perfect elastic solid. Such a body can transmit elastic waves in a variety of ways. First of all, we have waves which travel through the body of material. There are two sorts of body waves. In the first kind of wave the particles pushed by the disturbance, push their neighbours, which, in turn, push their neighbours. The disturbance travels out from the source with each individual particle moving back and forth in the direction in which the wave is progressing. This wave is called a longitudinal wave. In the second kind of wave a particle which is pulled sideways by the disturbance pulls its neighbour in turn and the disturbance propagates as a wave in which the individual particles move at right angles to the path of propagation. This wave is called a shear wave. Obviously particles can't pull each other aside in the fluid and there is, therefore, no shear wave in fluid. In addition to waves transmitted through the body of the earth, there are other waves transmitted along the free surface. The physics of these surface waves is very complicated, and we must content ourselves with a simple summary. They are of two sorts, Rayleigh waves, in which the ground particle moves vertically and radially in the plane of propagation, and Love waves in which it moves at right angles to this plane. Both types of surface waves are dispersed, that is, their velocities are dependent on their wave lengths.

What are the properties of these various waves? Let us say at the start that I am speaking as a seismologist, not as an engineer. We seismologists are interested in the study of earthquakes at some distance from the source. We normally use high-magnification instruments because we want to record as many distant earthquakes as possible, and we record at low paper speed because we have to record continuously. The frequency response of our instruments is also rather special; we normally operate one narrow-band set of instruments peaked at about 1 second, and a broad-band set covering the range from about 15 to 100 seconds. These instruments do not get the sort of information the engineers want -- this information comes from strong-motion seismographs about which you will hear later.

Within these limitations this is what we find. Close to the epicentre the longitudinal waves which are recorded on conventional seismographs usually have periods of about 1/10th to 1/5th of a second; the traverse waves are of longer period, perhaps of the order of a 1/4 second to 1/2 second, and the surface waves may have periods ranging up to 2 or 3 seconds. When we are at a greater distance from an earthquake all of these periods are considerably increased and when we are at a great distance from a major earthquake we will usually find that the longitudinal wave has a period in the range 1 to 5 seconds, the shear waves have periods that range from 7 to 15 seconds, and the surface waves may have periods ranging all the way from a few seconds to as much as an hour. The large distant earthquake is of no concern to the earthquake engineer. For your purposes you may expect to have to contend with wave periods ranging from a few tenths of a second up to a few seconds. As for the amplitudes of these waves, this had best be left until the discussion on "strong motion" seismographs. If a conventional seismograph is close to the centre of the earthquake its high magnification usually results in a jumbled trace or a broken instrument.

Earthquake Magnitude, Intensity and Energy

There are two terms which seismologists and earthquake engineers use which you may find confusing. These are intensity and magnitude. It is important to understand the difference.

When a seismologist or an engineer is examining the results of an earthquake he attempts to define these in terms of an empirical scale based on observed effects on population, buildings and ground. This is known as an intensity scale. Intensity, therefore, is a measure of the effects of an earthquake. On the other hand, when a seismologist records an earthquake on his seismographs, he is able to measure the exact amplitude and period of the waves which the earthquake causes at his station. This is a measure of the absolute size of the earthquake and by taking into account the distance of the station from the earthquake, it is possible to express this absolute size; the term for this is magnitude. When the magnitude is determined from the records of several stations there is usually quite good agreement between the values and the average can be accepted with confidence.

We are now in a position to understand the difference between magnitude and intensity. An earthquake might have a reasonably low magnitude but because of shallow focus, poor soil conditions or poor building standards, it might cause a great deal of damage; it would, therefore, have a very high intensity. On the other hand, an earthquake of very large magnitude might have a great focal depth, or might strike in an area where there was very little man-made structure to damage. It would not, therefore, have a high intensity. It is important that we understand the difference between these two concepts.

A great deal of effort has been expended by seismologists in trying to determine the relationship between the magnitude of an earthquake and the energy which it releases. The equation currently accepted is

$$\log E = 11.4 + 1.5M$$

where the energy E is measured in ergs. If we want to compare the energies E_1 and E_2 of two earthquakes of magnitudes M_1 and M_2, we must write the above equations for each earthquake and subtract to obtain

$$\log E_1/E_2 = 1.5\ (M_1 - M_2)$$

This equation will give us the ratio of energies of any two earthquakes. In Figure 2 I have taken the Long Beach, California, earthquake of 1933 as a standard earthquake and I have related the energy of other North American earthquakes to it. The Long Beach earthquake did $50,000,000 worth of damage. It damaged every school in the city, some to the point of total collapse, and started a number of fires which fortunately were controlled. The Long Beach earthquake had a magnitude of 6.2. If we put $M_2 = 6.2$ and E_2 equal to unity, the foregoing equation will tell us the ratio of the energy released in any other earthquake to that of the Long Beach one. The graph of this equation is shown in the figure. Two general points have been indicated on the graph, the threshold of damage provided by an earthquake of magnitude 5, and the largest known earthquake of magnitude 8.9. We see that the former released only about 1/100 the energy of the Long Beach earthquake, the latter more than 10,000 times as much. This emphasizes the fact of the magnitude scale being, a log arithmic one, is very much compressed.

What is a relationship between intensity and magnitude? Obviously this question cannot be answered rigorously because the intensity depends on so many factors other than the absolute strength of the earthquake. It depends, for example, on the focal depth of the earthquake, on the nature of the soil and of the buildings, and of the care with which buildings have been constructed.

It is only in limited areas that it is possible to give any relationship. Figure 3 shows the relationship which is believed to exist in California; experience suggests that this graph applies reasonably well in Canadian earthquakes, but we must always remember that no close relationship can exist.

Tsunami and Seiches

In 1929 an earthquake occurred on the Grand Bank of Newfoundland, about 250 miles from shore. It did practically no direct damage on land. However, because of some change in the ocean bottom, a large wave was set up on the ocean which travelled out in all directions from the epicentre. It was noted on tide gauges as far away as Bermuda and in the Azores but its principal damage was confined to Placentia Bay on the south coast of Newfoundland. Here the wave built up to a height of 30 feet, inundated several villages, and drowned 27 people.

Seismic sea waves have been particularly damaging in Japan and we employ a Japanese word - tsunami - in describing them. The cause of tsunami is open to some question but there is always a possibility that one may occur when an earthquake occurs under the ocean. Once the wave has been set up it propagates with the velocity appropriate to the water depth. The approximate formula in deep water is

$V = \sqrt{gD}$, where g is the acceleration of gravity and D is the depth of water.

If this formula is applied to the Pacific Ocean, where the mean depth is about 3 1/2 miles, one obtains a velocity of some 500 mph. The period of the wave is about one hour and the wavelength is, therefore, about 500 miles. A wave with such a wavelength, even if the amplitude is very high, is quite indistinguishable at sea. However, as the wave runs up on a shelving shore, with consequent reduction in velocity, the wave height builds up in order to conserve the energy in the wave and heights as great as 50 feet have been observed in Hawaii. If, in addition, the wave is running into a funnel-shaped channel, so that the volume of water is confined between narrowing walls, much greater heights may occur; heights as great as 90 feet have been observed in Japan.

The Pacific Ocean is a big place and despite the high velocity of the tsunamis they take a long time to cross the ocean. This allows time for adequate warning. The United States Coast and Geodetic Survey maintains a 24-hour watch from a central station in Hawaii. When an earthquake is recorded which appears to have originated under the ocean, they radio to certain key seismic and tidal stations, locate the epicentre accurately, keep check on any developing tsunami and compute its arrival time for various points. The most spectacular success of the service was in the case of the Chilean tsunami of 1960; the arrival time of this at Hawaii was forecast within 1 minute. Despite this, 61 persons were killed and 282 injured. They had ignored the warning.

Earthquakes can affect closed bodies of water as well, by setting them into oscillation. The phenomenon is known as a seiche. The best known recent seiche was that caused by the Hebgen Lake, Montana, earthquake of 1959. A seiche with a period of 17 minutes and an amplitude of some tens of feet was set up in the lake.

Where do Earthquakes Occur?

I am going to leave to a later speaker the question of where earthquakes occur in Canada. Figures 4 and 5 show where the major earth quakes of the earth occur. There are two zones -- a circum-Pacific zone, beginning in Chile, completely encircling the Pacific to end in New Zealand, and an "Alpide zone", starting in the Azores, stretching through the Mediterranean, the Hindu Kush, the Himalayas and Indonesia, to join the circum-Pacific zone in New Guinea. The circum-Pacific zone contains 90% of the earth's normal earthquakes, almost all of its deep ones, and the Alpide zone takes care of most of the rest.

Our Pacific coast is part of the circum-Pacific zone, and as such it has its share of major earthquakes. You will see one or two epicentres in eastern Canada. You will hear about both of those areas later from Mr. Milne.

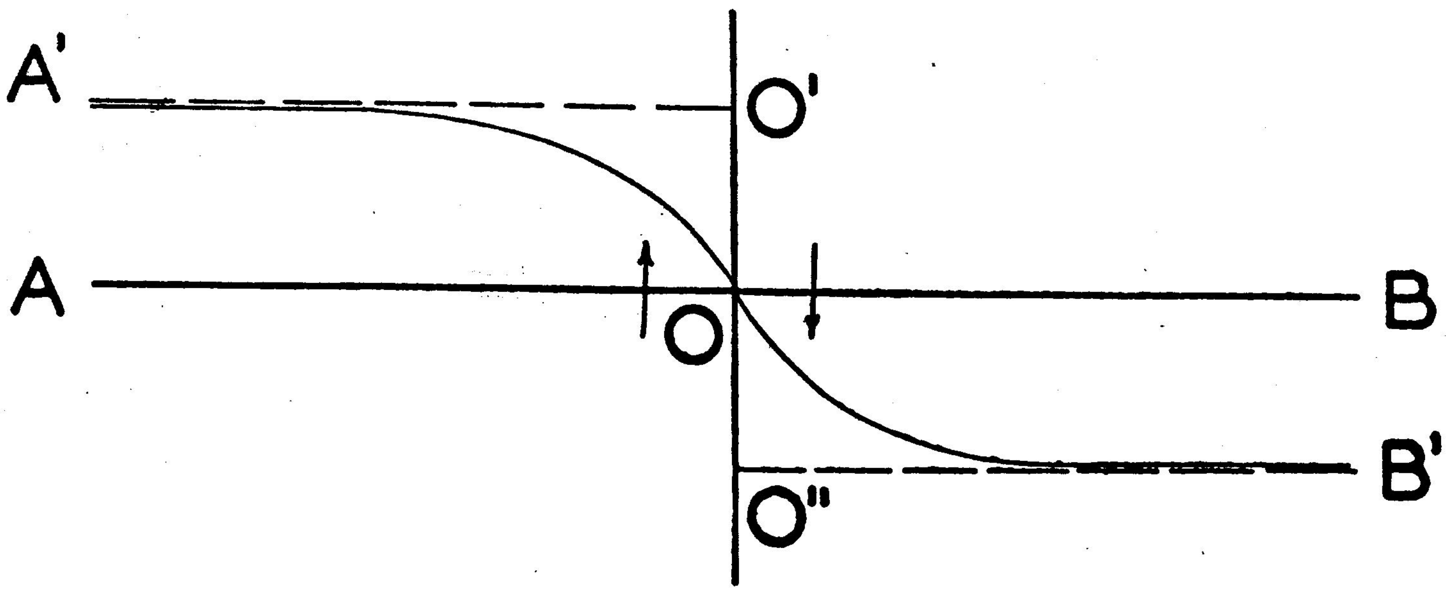

Fig. 1 Illustration of Elastic Rebound.

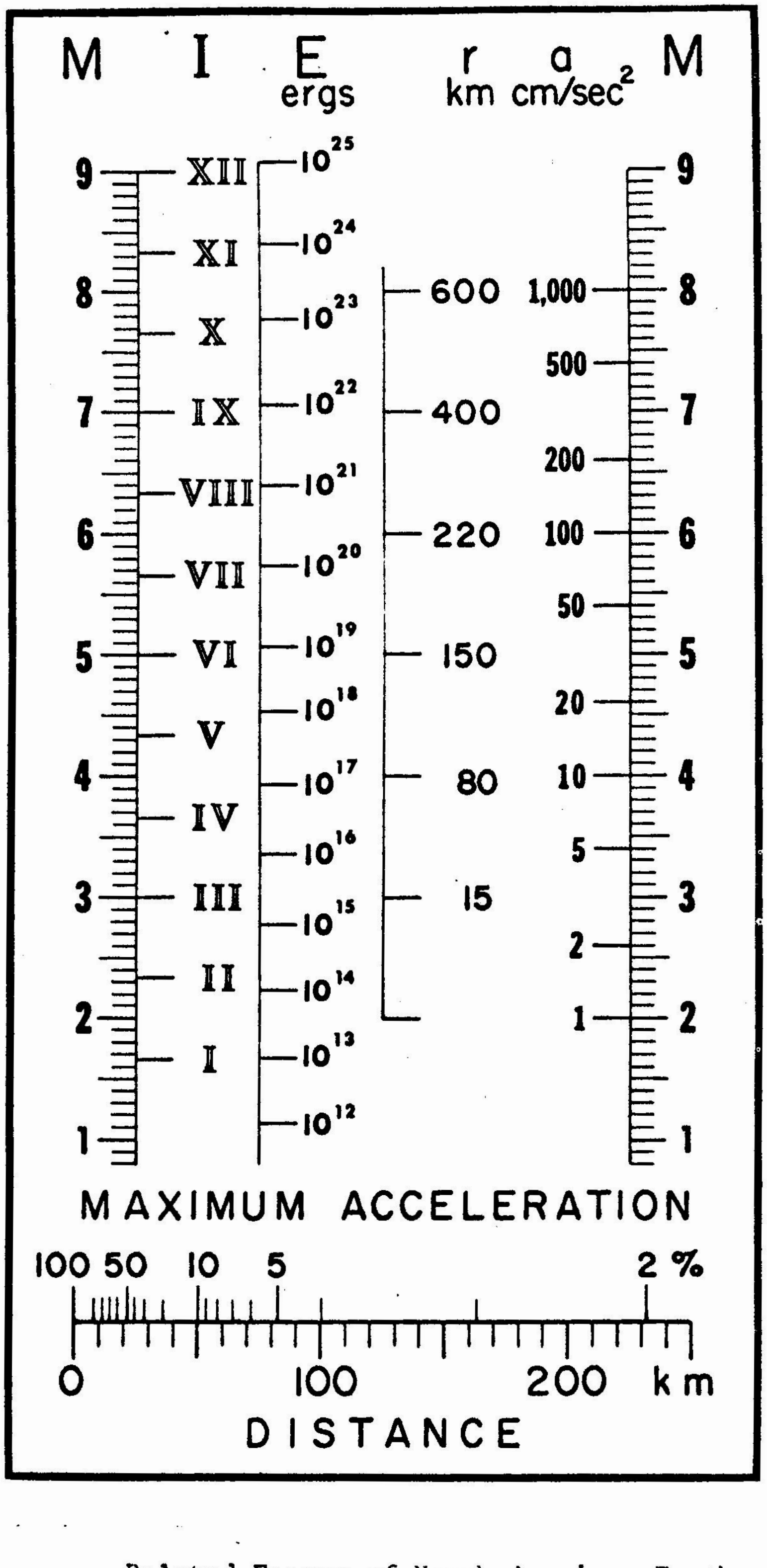

Fig. 2 Related Energy of North American Earthquakes

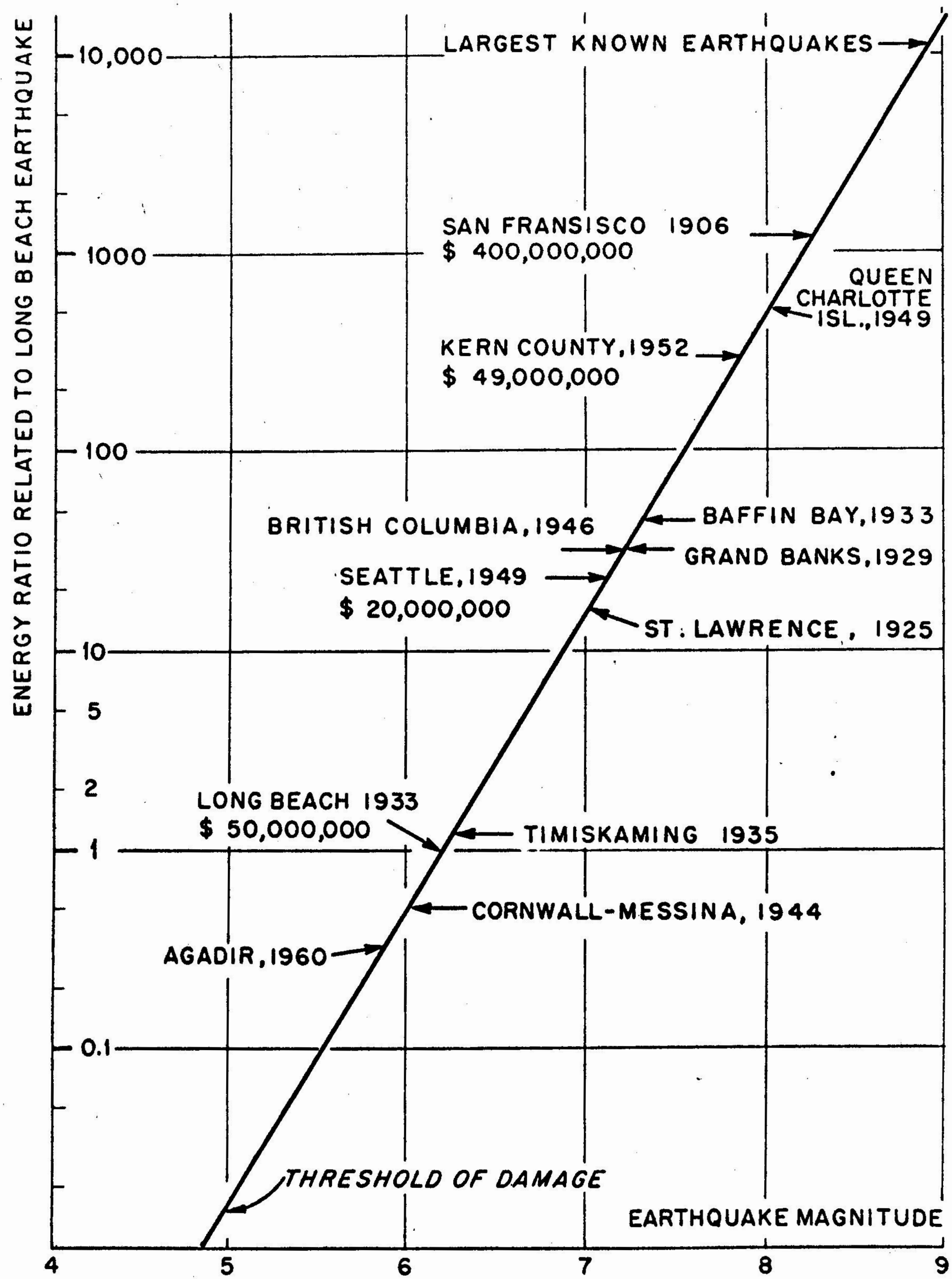

Fig. 3 Relationship Between Intensity and Magnitude (California)

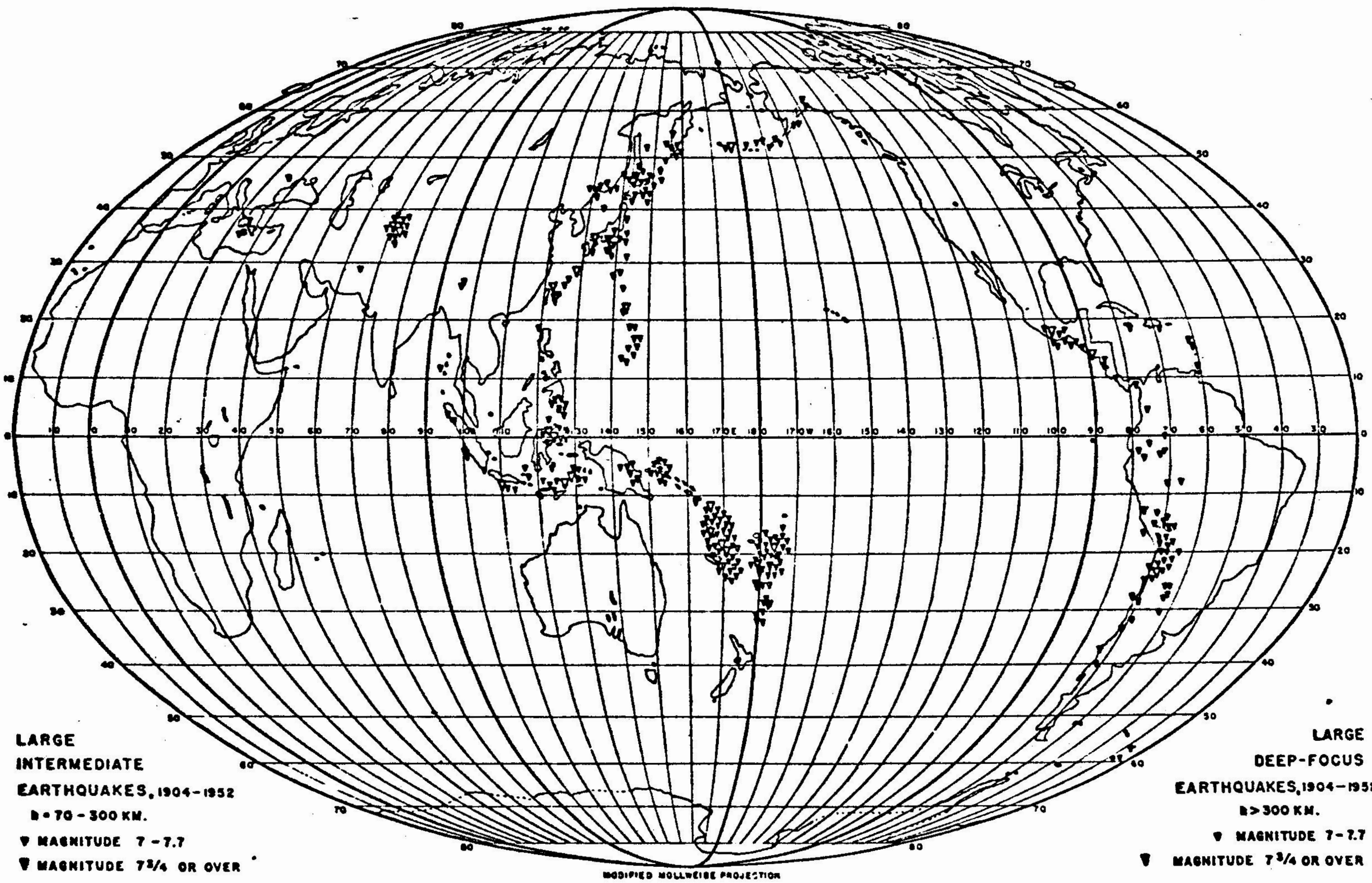

Fig. 5 World map of deep-focus earthquakes: all known shocks of magnitude 7 and over, 1905-1952.

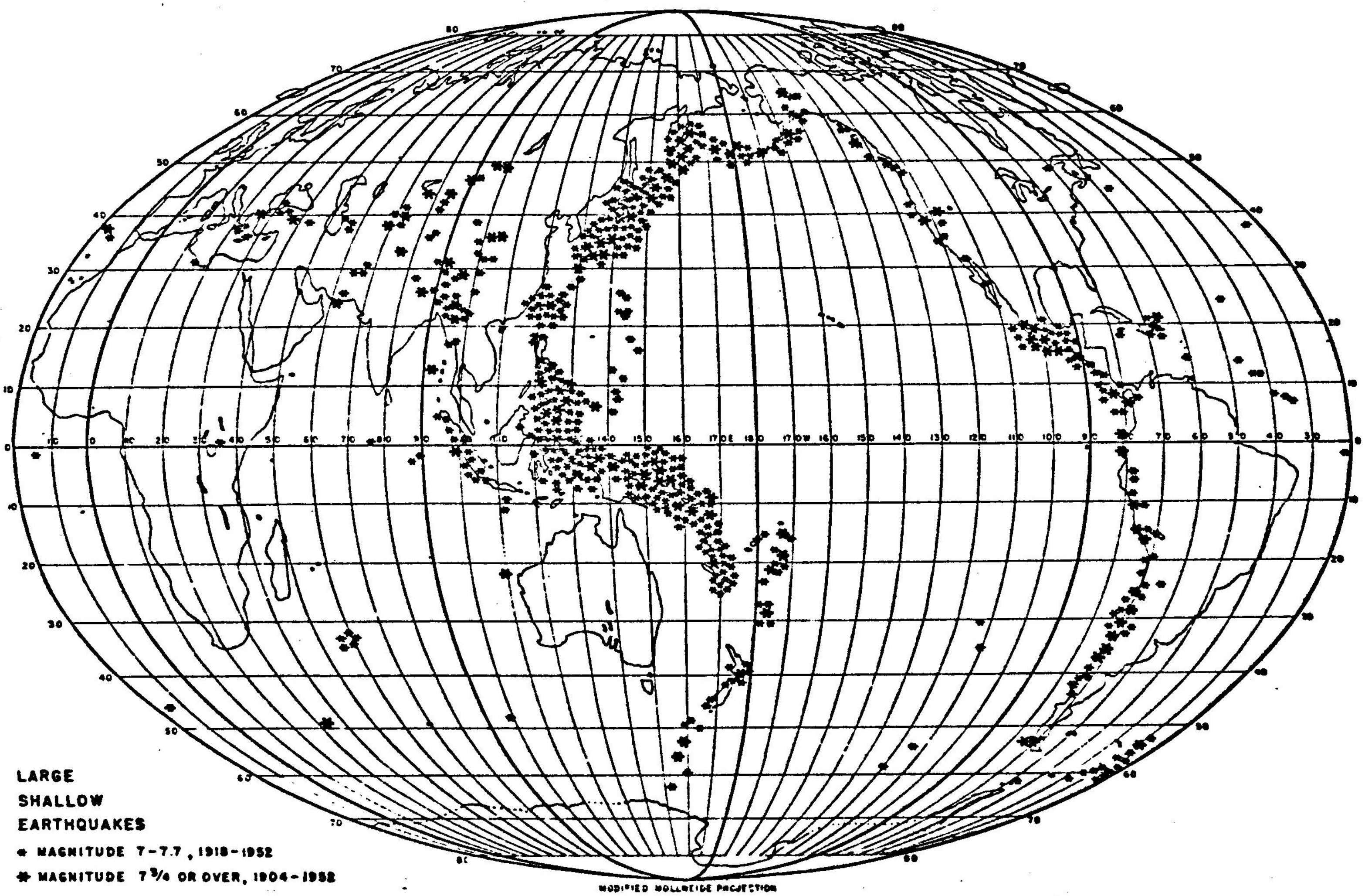

Fig. 4 World map of shallow earthquakes: class *a* 1904-1952; class *b* 1918-1952.