PACIFIC COAST EARTHQUAKES
Pacific Coast Earthquakes

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CONDON LECTURES

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ADDENDA AND ERRATA

ERRATA, page 25

Figure 9 legend should read: “Earth flow, 1906. Size of the flow can be judged by the man to which arrow points.

ADDENDA

THE CONDON LECTURES

The Condon Lectureship was established in 1944 by the Oregon State Board of Higher Education upon the recommendation of the late Dr. John C. Merriam who was, at that time, a member of the faculty of the University of Oregon. The Lectureship was named in honor of Dr. Thomas Condon, the first professor of geology at the University.

The subjects of the lectures relate to science and its application in the Pacific area. The lectures, usually two annually, are delivered three times in the state, namely, at Eugene, Corvallis, and Portland. They are then published in appropriately adapted form.

CONDON LECTURE PUBLICATIONS

The Ancient Volcanoes of Oregon. By Howel Williams, Chairman, Department of Geological Sciences, University of California. Jan., 1948. (Out of print.)


The China That Is To Be. By Kenneth Scott Latourette, D. Willis James Professor of Missions and Oriental History and Fellow of Berkeley College, Yale University. Mar., 1949. 75 cents.

The Pacific Island Peoples in the Postwar World. By Felix M. Keesing, Executive Head, Department of Sociology and Anthropology, Stanford University. Mar., 1950. 75 cents.

Pacific Coast Earthquakes. By Perry Byerly, Professor of Seismology, University of California. May, 1952. 75 cents.
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Foreword

The French seismologist, Montessus de Ballore, pointed out that the greater the gradient of the earth's surface in a region, the more liable that region is to have earthquakes. This is often called Ballore's Law. There are exceptions in that bit of earth's history which man has observed, i.e. the Alpine region has earthquakes but they are not very large. Nevertheless high mountains near deep seas usually must be paid for by shaking. During our short historical period Oregon has been most fortunate in the low price she has had to pay for her magnificent coast line. These lectures are a plea for earthquake-resistant building construction in Oregon as a matter of normal foresightedness. Be ready in case it turns out that Ballore's Law is more exacting than we hope it will be.
Pacific Coast Earthquakes

DISTRIBUTION

The history of the earthquakes of the Pacific Coast of the United States, 1769-1928, is set forth in a catalog compiled by S. D. Townley and Maxwell Allen and published as Volume 29, Number 1, of the Bulletin of the Seismological Society of America. This “catalog” had as its basis a “catalogue” published much earlier. The older publication was compiled by E. S. Holden and covered earthquakes from 1769 to 1897. It was published in Smithsonian Miscellaneous Collections No. 1087. The Townley-Allen catalog is a volume of 297 pages, of which California shocks occupy 252 pages, Oregon 5, Washington 9, Idaho 3, Nevada 16, Utah 5, Arizona 4.

The record is far too short for us to get a picture which might lead us to prediction. But we can study the California record to see what it offers.

The first entry in the earthquake catalog is dated April 11, 1769, in San Diego. It is based on a story told by the Indians to Father Serra that on the above date they felt an earthquake and had an eclipse of the sun. Astronomical catalogs deny the eclipse, so some doubt also the earthquake. Historical records for such early days are sparse, of course. There are only five more entries before 1800, but after that date they are numerous.

The year 1812 was called the year of earthquakes by the churchmen, for most of southern California was severely shaken. It is difficult from the old records to be sure whether reports from different localities describe the same or different earthquakes. So often the record is written some time after the occurrence, and dates are uncertain. Mission churches were wrecked at San Juan Capistrano, San Fernando, Santa Barbara, Santa Inez and Purisima.

Although many earthquakes, some no doubt very large, are entered for the early years, the scantiness of the record makes it hard to evaluate them.

However, in 1836 and again in 1838 there were very severe earthquakes in the San Francisco Bay region. On June 10, 1836 occurred a great shock on the Haywards Fault along the east side of San Francisco Bay. Large fissures were formed and there were “innumerable” after-shocks. It “caused havoc” in Monterey and Santa Clara.

One might have expected such a large earthquake to relieve the strain in the rocks of the earth’s crust of the Bay Region, but two years later
another great shock occurred, this time on the San Andreas Fault west of San Francisco Bay. A "great" fissure was formed from San Francisco to a point near Santa Clara. Walls were broken in the Presidio of San Francisco, at the Missions of Santa Clara, San Francisco and San Jose, and in Monterey. To show the nature of the records on which the accounts of these two earthquakes are based, we note that there were no newspapers in California at the time. The Church was undergoing forced secularization, therefore there were no mission reports. Sources are portions of old diaries and recollections of old men who were interviewed after the earthquakes of the 60's as to what they remembered of the 30's. One source was an article in the Honolulu Advertiser telling what a sea captain had to say regarding the effects of the shock in Monterey. The captain had arrived in Monterey a few days after the earthquake.

One of the greatest earthquakes in California history was that of January 9, 1857. It is known as the Fort Tejon earthquake. This old fort occupied a site which is now in northern Los Angeles County; and the San Andreas Fault runs close to it. The fault broke for many miles on each side of Fort Tejon; the report says the earth opened in a fissure 20 feet wide and 40 miles long and then came together with such violence that a ridge was formed 10 feet wide and several feet high. At the head of Terwilliger Valley near the Los Angeles County Playground this ridge is still in evidence. At Fort Tejon all buildings were thrown down. The shock was felt as far as Fort Yuma, Sacramento, and the Mexican border. One item in the record may appeal to some readers who know Los Angeles: "The Los Angeles River was thrown out of its bed."

In the 60's another pair of large earthquakes occurred in the San Francisco Bay region—one on the San Andreas Fault and one on the Haywards. It was on October 8, 1865 that the former struck. In San Francisco the old Merchants' Exchange was completely ruined. A number of other buildings were badly damaged. On marshy ground water and gas pipes were broken. In San Jose walls were thrown out of the Methodist church and the jail. In New Almaden in the Santa Cruz Mountains a building was shaken down and cracks were formed in the soil. The earthquake was felt as far as Sacramento and Stockton.

In 1868 on October 21 occurred the Haywards earthquake. Haywards (now Hayward) was a small town south of Oakland on the eastern shore of San Francisco Bay. The Haywards Fault cracked open for 20 miles from Warm Springs to San Leandro. Every building in Haywards was damaged and many were demolished. There was considerable damage in San Francisco to buildings on made land. The shock was felt at a distance of 175 miles from the central region, i.e., from Gilroy to Santa Rosa.
The next great California earthquake to claim our attention is that of March 26, 1872. This centered in the Owens Valley, Inyo County, east of the Sierra Nevada. The Owens Valley fault along the western flank of the Sierra Nevada broke for some 70 miles—not one continuous clean cut break, but many discrete breaks, some with displacements as great as 20 feet, both horizontal and vertical. In Lone Pine 52 out of the 59 houses there were thrown down, and 23 people were killed out of a population of 250 to 300. The shock was felt throughout most of California and Nevada and parts of Utah and Arizona.

The only truly great shock felt in California since 1872 was that of April 18, 1906. This earthquake is world famous because of the magnificent fault break and because of the great destruction by earthquake and fire in San Francisco.

Although many earlier geologists had associated faults and earthquakes, the conviction that the earth waves are caused by the fault rupture was founded in the example of the 1906 earthquake. At the time of this shock a great fault broke over a length of 270 miles, from Upper Mattole in Humboldt County to San Juan in San Benito County. This fault was named San Andreas after a lake through which it passed. Although parts of it had been mapped earlier, it had not been recognized as a continuous feature before 1906. The westerly side of the fault moved northerly relative to the easterly side. The maximum relative displacement was 21 feet at the head of Tomales Bay, where the displacement was measured on a road which crossed the fault at right angles. There is no question that the elastic energy released by such a break was sufficient to supply the energy in the earth waves. The energy involved has been variously estimated as between $10^{23}$ and $10^{26}$ ergs.

In San Francisco the estimated property loss due to the earthquake alone (exclusive of damage done by the fire which followed) was $20,000,000. San Jose and Santa Rosa, the latter some 20 miles from the fault trace, suffered terribly. It was this earthquake which pointed very clearly to the effect of the geologic foundation on building damage. Structures built on good rock fared well, while those built on made land fared badly. This earthquake was felt from Coos Bay to Los Angeles and as far east as Winnemucca (Nevada), a distance of some 300 miles from the fault.

We may note in passing that although several of these California earthquakes were exceedingly severe in their central regions, none of them shook areas of as great a magnitude as, for instance, the New Madrid (Missouri) earthquakes of 1811-12. These were felt all along the Atlantic seaboard, the Gulf Coast, and the Canadian border.

We have discussed the larger California earthquakes primarily because California is a neighbor of Oregon. Although the above record
Figure 1. Epicenters of Strong Earthquakes of the Northern Pacific Coast, 1901-1949.
lists no very large earthquakes in that part of California contiguous to Oregon, there have been a considerable number of quite strong shocks off the coast of northern California near the Oregon border. These have frequently been strong enough on shore to shake down chimneys in Humboldt County, California. The map, figure 1, indicates the epicenters of strong Pacific Coast earthquakes from northern California to Queen Charlotte Island, from 1901 to 1949.

These earthquakes are the ones deemed large enough to have done some damage to masonry and chimneys. For those off the coast we estimate this damage from (1) effects on shore and (2) the instrumentally determined "magnitude" of the shock. On this map one at once notices a lack. Oregon has been slighted!

The map in figure 2 covers a larger time, 1866-1949. There are four black spots in Oregon on this map and one off the coast near the California state line. These represent centers of larger quakes of the type plotted on figure 1. The crosses represent earthquakes only strong enough to rattle windows or less; and the squares those which moved small objects but did not damage masonry. There is a fair scattering of the little earthquakes. Reports of early shocks of course depend on population spread—no people, no reports. If the map extended farther west there would be a number of larger shocks well at sea.

Table 1 gives the data on which figure 2 is based. Table 1 includes...
some earthquakes centering in California but felt in Oregon. The sources of information are listed in the references. One must note that some of the early reports are vague and open to question. For example, the earthquake of October 12, 1877 is reported from “Portland, Marshfield and Cascades.” It “was of sufficient violence to overthrow chimneys.” At which place? In figure 2 it is plotted at Cascade Locks and that may be questioned.

After some of the entries in the table there are Roman numerals. These indicate the “intensity” of the earthquake as judged by its effects. The rating was done on the old Rossi-Forel scale which is as follows:

I. Microseismic shock—recorded by a single seismograph, or by seismographs of the same model, but not putting seismographs of different patterns in motion; reported by experienced observers only.

II. Shock recorded by several seismographs of different patterns; reported by a small number of persons who are at rest.

III. Shock reported by a number of persons at rest; duration or direction noted. A shock; a light shock.

IV. Shock reported by persons in motion; shaking of movable objects, doors and windows, cracking of ceilings. Moderate; sometimes strong; sharp; light.

V. Shock felt generally by everyone; furniture shaken, some bells rung.

VI. General awakening of sleepers; general ringing of bells, swinging of chandeliers; stopping of clocks, visible swaying of trees; some persons run out of buildings.

VII. Overturning of loose objects; fall of plaster; striking of church bells; general fright, without damage to buildings.

VIII. Fall of chimneys; cracks in the walls of buildings.

IX. Partial or total destruction of some buildings.

X. Great disasters; overturning of rocks; fissures in the surface of the earth; mountain slides.

In the latter part of the table ratings were made on the Modified Mercalli Scale which is given later in this paper. Arabic numerals are used for these intensities. The Modified Mercalli Scale is the one now in use by the U.S. Coast and Geodetic Survey which evaluates the intensities.

Every large earthquake is followed by a host of smaller ones as the earth settles back to rest. These are called aftershocks and this term is applied to a number of shocks in the table with a question mark after them. If one is sure that it is just an aftershock he is not likely to tabulate it. It is considered an essential part of the main earthquake. After a truly great shock, aftershocks may continue for years, i.e., there is an unusually great number of small shocks lessening in number and violence as time goes on.
Oregon, then, as is shown in figure 2, is not free of earthquakes—it has no immunity (as all fervently wish it had.) It lies between two states which have violent shocks. Moreover Oregon's history is very short. It will not be safe to ignore the possibility (even the probability) that Oregon may have violent shocks in the future. Requirement of earthquake resistant construction in Oregon is necessary, particularly for schools, churches and other public buildings.

There is one interesting entry in Holden's early catalog which Townley enters only to disqualify. A bored soldier named Warren at Fort Klamath wrote a lurid story of a terrible earthquake at the fort on January 8, 1865. He signed the letter L. Tennyson, Quartermaster's Clerk. Suspecting the story and the name, Professor Townley ascertained from the army that there had been no Tennyson in the United States military service at the time.

The state of Washington, although not so seismic as California in historic time, has exceeded Oregon in number and severity of its earthquakes. Those shocks which threw down chimneys, or appear to have been of sufficient intensity to have done so, total about nine.

1904, March 16. This shock was felt in Seattle, where bottles were thrown from shelves. It is rated as of high intensity because Indians reported changes in level and in water flow in the western part of the Olympic Peninsula.

1909, January 11. Felt for two hundred miles north and south of the Strait of Juan de Fuca. Cracked a sidewalk at Anacortes, threw down a lumber pile at Port Angeles, threw down chimneys in Vancouver (B.C.) and Westminster (B.C.).


1932, August 6. A few chimneys were demolished in Seattle and others badly damaged.

1939, November 12. Felt throughout western Washington and northwestern Oregon. Threw down chimneys at Auburn, Centralia, Elma. At Oakville and Tacoma chimneys were damaged.

1945, April 29. Felt throughout most of Washington and part of northwestern Oregon. Chimneys and brickwork damaged in Cle Elum, Leavenworth, North Bend, Snoqualmie.


1946, June 23. This earthquake centered in Georgia Strait. Chimneys fell at Eastsound and were cracked in Quinault. Bricks fell from some buildings in Seattle.

There have been a number of quite severe earthquakes in northern Vancouver Island and Queen Charlotte Island.

The distribution of Pacific Coast earthquakes outlined above, both in time and space, do not give any grounds for prediction. The distribution merely shows that the Pacific Coast is a region of earthquakes and that one would do well to prepare for them. It demonstrates that in the short period for which records have been kept, Washington has been luckier than California, and Oregon luckier than Washington. After another 200 years, say, scientists may be able to say that the mighty forces which cause earthquakes are less operative in the Oregon region than to the north and south of it. One may hope for that. But such a statement should not yet be made.

**CAUSE OF EARTHQUAKES**

American geologists and many, perhaps most, foreign geologists accept faulting as the source of earthquake waves. A fault is a break in the earth's rocky crust. The rocks become severely strained from the action of internal forces and eventually break along a fault. As one side of the fault moves relative to the other, the sides grate mightily. The vibrations so engendered travel out as elastic waves in the earth and wreak havoc on the works of man and some of the lesser works of nature.

The general acceptance of fault motion as the source of the earthquake was largely the result of the great California earthquake of 1906 with the accompanying break of the San Andreas Fault for 270 miles. Here at last was a visible fault break of such magnitude that there could be little question that it was a sufficient source. Long before this earthquake, of course, it had been noted that centers of shocks tended to lie more or less along linear zones and many had inferred that rupture of the rocks along internal surfaces caused earthquakes. A weakness in the faulting theory of earthquake causation is the observation that so very few earthquakes are accompanied by a surface fault break. Most of the time we must infer slipping at depth with no surface rupture.

It was in the report of the State Earthquake Investigation Committee on the 1906 earthquake that Harry Fielding Reid set forth clearly the Elastic Rebound Theory of the cause of earthquakes. Strains in the rock of the earth's crust are set up slowly by the action of interior forces probably applied from below. The energy of strain eventually becomes so large that the rocks cannot hold and rupture ensues. The energy which just before the shock was strain energy is transformed on break of the fault into (1) heat energy, and (2) elastic wave energy. The latter propagates away from the source and causes most of the damage.
Rarely has there been a building over a fault trace. Then, of course, the structure is torn in two.

The Elastic Rebound Theory shows how a large supply of energy (strain energy) which could be released catastrophically, might be concentrated along a narrow zone. The accumulation of this energy is slow, as are most geological processes; e.g., erosion and deposition, cooling and heating of the earth's crust.

![Figure 3. Displacement of Fence by Fault, offset nine feet; Calexico-Yuma Road, 200 feet west of Alamo River; Imperial Valley earthquake of May 18, 1940. Photo by J. P. Buwalda.](image)

In studying the distribution of the major California earthquakes, we are struck with a number of apparent inconsistencies. In the great earthquake of 1857, the San Andreas Fault broke in its southern part (the exact extent is not known), while in 1906 it broke in its northern part. There is a portion of the fault between these two breaks which has not ruptured. One would feel that the system was behaving more rationally if it tore along successive segments at successive times. The fault is easily traced by field evidence of prehistoric breaks in this intermediate region.

In the 1830's and again in the 1860's large earthquakes occurred successively on each side of San Francisco Bay at intervals of two or three years. From our theory as to their cause, it would appear con-
cerning faults as close together as the San Andreas and the Haywards, that relief of strain on the one should relieve strain on the other. However, the great 1906 shock in which the San Andreas Fault broke for 270 miles was neither preceded nor followed by a break on the Haywards Fault. Here was a shock large enough to relieve regional strain.

The data on which Reid based his elastic rebound theory were the changes in position of many triangulation stations of the United States Coast and Geodetic Survey. These had been observed at various times from 1850 to 1907. Between 1850 (about) and 1890 (about) it appeared that the stations to the west of a base line through Mt. Diablo and Mt. Mocho (about 40 miles east of the fault) had all been moving northerly relative to this line. At the time of the earthquake and faulting those stations quite near the fault behaved differently. Those west of the fault flipped farther north and those to the east flipped south, again relative to the base line. The data were not all that could be desired, but the design

Figure 4. The San Andreas Fault as a Geomorphic Feature. This airphoto shows the San Andreas Fault as a linear feature crossing the bottom of the picture near the left corner. Note the displaced stream valley.
was clear enough. Studying the data, Reid decided that the rate of drift of the Farallon Light west of the Golden Gate relative to Mt. Diablo on the edge of the Great Valley was about 10 feet in 50 years.

Since 1906 the Coast and Geodetic Survey has continued careful observations of the position of key triangulation stations on each side of the San Andreas Fault. The sensitive observation is that of sighting from a station about 20 miles on one side of the fault the position of a station about 20 miles on the other side. It is found quite consistently that the stations to the west are drifting slowly north relative to those to the east. The bearing of a westerly station from an easterly one always shows this slow drift. Its magnitude, according to Whitten, corresponds to a drift of about two inches per year in the region of the 1906 break. This is remarkably close to Reid's 20 feet in 100 years. Mighty forces are at work. The San Andreas Fault must break again some day to relieve the accumulating strain.

However, in one sense the Elastic Rebound Theory evades the issue, for we immediately inquire: What are the internal forces which cause the strain to accumulate?

Here we have a puzzler? Displacements along California faults have followed a definite pattern for a long time. The general trend of these faults more or less parallels the coast line and the displacements are regularly the Pacific side northerly relative to the continental side. This showed up beautifully in 1906 and also in the Imperial Valley earthquake of 1940, when a surface fault break also occurred.

It is borne out by physiographic features along many faults which have not broken in historic time. Stream valleys crossing the faults are displaced northward at the fault.

Reid thought that the nature of the displacements of the triangulation stations of the U.S. Coast and Geodetic Survey in 1906 indicated that the forces which caused them acted on the under side of the earth's crust. We then have the suggestion of subcrustal "currents" flowing plastically in a direction roughly parallel to the coast and flowing with different velocities at different distances from the coast.

The theory of Isostasy calls for subcrustal plastic flow. By plastic flow we mean that the rocks flowing are not liquid in the ordinary sense. These rocks can and do support the shear waves sent out by an earthquake; they are rigid and an ordinary liquid is not. Shoemakers' wax is a material which is rigid; a tuning fork can be made out of it, but if this fork sits for a day or so it flows down into a pancake.

The theory of Isostasy (Equal Standing) was called into being to explain certain observations of gravity which seemed anomalous. As long ago as two hundred years the French geodesist Bouguer went to Ecuador to measure the arc of a meridian. He expected the great mass
of the Andes to attract the plumb bob so that he would need to correct for it. He found that even the great mountain Chimborazo did not attract very much. Later others observed the same phenomena in other parts of the earth. The absolute value of gravity on top of a mountain is almost the same as the value at a similar height above the ocean. This means that the rocks of which mountains are made are lighter than those of which sea bottoms are made. The result of this would be a flotation. Mountains float in the heavier rocks below them like icebergs in the ocean. Mountains have roots of light rock extending deep into the substratum. As the mountains are eroded away and the products of erosion deposited in ocean basins, the former must rise (as an unloaded boat) and the latter sink. This calls for a (very slow) current at depth from under the loaded area to under the unloaded area. Here then is one possible source of deep flow. However, it would appear that such in California ought not to produce faults parallel to the coast line.

The earth is not a perfect sphere. It approximates an ellipsoid of revolution. The direction of the plumb bob at any point (save at the poles and along the equator) does not point to the geometric center of the ellipsoid. Two disturbing causes are present: centrifugal force acting on the plumb bob and the excess gravitational attraction due to the equatorial bulge. So the plumb bob direction, the direction of "gravity", points a little more toward the equator than does the radius to the center. This effect is less as one goes down into the earth. A floating continent is buoyed up by a force acting in the direction of gravity at the center of mass of the displaced subcrustal material. It is pulled down by a force acting in the direction of gravity at the center of mass of the continent. The latter center is higher than the former. So there is a tiny resultant force toward the equator. It is about one-millionth of the continent’s weight. This force is called the Eotvos force, and if the subcrust is not strong enough to resist it should eventually result in a clustering of continents around the equator. A little figuring makes it appear that so weak a subcrust could not support a hill much over 20 feet high. The ocean bottom, which has often been thought a sample of the subcrust, supports small islands which sometimes appear to be of the same general constitution as the bottom and thence not held up by flotation. However, the Eotvos force may be urging the North American continent toward the equator and the San Andreas Fault may be the rupture due to the Pacific holding back. It is fanciful and one could list many objections.

Some theories call for great convection currents within the earth’s mantle, between the fluid core (half way down toward the center) and the crust. These might drag different portions of the crust in different directions.
There remains the old theory. If the earth is cooling, the interior is lowering its temperature and getting smaller. The crust, which is in radiation equilibrium with the sun, maintains its temperature and volume and must therefore collapse occasionally to fit the shrinking interior.

Little is known of the underlying causes of earthquakes. None of the theories advanced, however, give us reason to expect to find a small region, such as Oregon, which enjoys relative seismic quiet between two active regions. High mountains beside deep seas usually mean large earthquakes.

**INSTRUMENTAL SEISMOLOGY**

Mention was made earlier of earthquake waves—elastic waves propagating in the earth. Seismology may be divided in three parts related to:

1. The cause (geologist's conclusion—it is faulting)
2. The propagation of the disturbance by elastic waves
3. The effects on the earth's surface when the earthquake waves arrive—i.e., the effects of the shaking.

Earthquake waves may be classified in two broad classes: (1) Body waves which penetrate the interior of the earth and are refracted back to the surface because their speed increases with depth in the earth (except for a drop at the core boundary); also there are body waves reflected at the core boundary, the earth's surface and occasionally at other boundaries; (2) surface waves which travel around the earth penetrating only to a slight depth which depends on their wave length.

**Body Waves**: There are two kinds of body waves. The fastest is a longitudinal wave, i.e., the particles of rock vibrate back and forth along the path of the wave. Waves of sound are also longitudinal. In fact some of the earthquake waves of this type are in the audible frequency range and occasionally are refracted into the air and are heard by people in regions where the shock is severe. Physically longitudinal waves are waves of compression-rarefaction or waves of dilatation. As they pass the rock changes volume but does not change shape.

The second kind of body wave is a transverse wave—the particles of rock vibrate at right angles to the path of the wave. One sees this type of wave when he shakes a rope—violin strings vibrate in this fashion. Physically these are shear waves or rotational waves or equivoluminal waves; the volume of the rock does not change as the wave passes, there is only a change in shape.

The longitudinal waves travel faster than the transverse waves. Their
speeds are related—the coefficient of incompressibility or bulk modulus of the rock $\kappa$, the coefficient of rigidity of shear modulus $\mu$, and the density $\rho$—by the relations:

$$v_p = \sqrt{\frac{\kappa + \frac{4}{3} \mu}{\rho}} \quad \cdots \cdots \cdots \cdots \cdots \cdots (1)$$

$$v_s = \sqrt{\frac{\mu}{\rho}} \quad \cdots \cdots \cdots \cdots \cdots \cdots (2)$$

Seismologists call the longitudinal wave $P$ and the transverse wave $S$ (for primary and secondary).

\[ \text{FIGURE 5. Berkeley seismogram of the Texas Earthquake of 1931, showing longitudinal (P), transverse (S), body waves and Love (L) and Rayleigh (R) surface waves.} \]

These body waves start on a path inclined downward. Since the speeds increase in general with depth, the rays are bent back toward the surface, i.e., refracted, and emerge again at the surface. Studies are made of the emerging waves to learn something of the conditions at depth in the earth. From such studies the speeds of $P$ and $S$ are known as a function of depth in the earth. Now a fluid will not support shear. Waves of the $S$ type cannot exist in a fluid. Below a depth of about 1800 miles $S$ waves are not transmitted. Also at this depth the speed of $P$ waves drops. So below this depth $\mu = 0$ and the earth is fluid. We call this fluid central part of the earth the core. It is probable that it consists
of two parts, an inner and an outer core and we know that at least the outer core is fluid.

However, $S$ waves travel freely at all points above the core. So there is no liquid layer at lesser depths, as scientists once assumed in order to explain volcanic phenomena. As we have noted before, the fact that rock is rigid, i.e. supports shear waves of periods a few minutes long or less, does not require that these same rocks may not flow plastically over periods of time measured in thousands of years.

Studies of earthquakes near their sources and of large blasts indicate that the outer crust of the earth, about the outer 20 miles under continents, is in layers. The rocks of the crust are lighter and the speeds of $P$ and $S$ lower than for the rock below. The speed of $P$ near the earth's surface is about 3.5 miles/sec. Under the oceans the thickness of the crust is smaller, perhaps about 5 miles or even less. The part of the earth between the crust and the core we call the mantle. The increase of speed with depth in the mantle is probably fairly gradual, the important sudden changes in the earth's constitution occurring at the top and bottom of the mantle.

$P$ and $S$ waves which have just grazed the core in their paths down and back emerge a little more than a quarter of the way around the earth from their source.

The source of the earthquake is called its focus. It is envisaged as a point on a fault surface where the fault begins to break. From the point it tears over a considerable area. That point on the earth's surface directly above the focus is called the epicenter.

Out to about 7000 miles from the epicenter both $P$ and $S$ waves are well recorded on seismographs. $P$ and $S$ travel closely the same paths in the mantle. But $P$ travels faster. We have curves and tables showing the relation between the difference in the times of arrival of $S$ and $P$ and the epicentral distance. These curves are empirical. They were originally constructed from data of earthquakes for which the epicenter was determined from field data, i.e., fault breaks, disturbance of soil, damage to structures. Gradually the curves were improved until nowadays computations based on distances from many seismographic stations give distances which intersect approximately at a point quite close to the region of damage.

So with distances of the epicenter from three stations, a seismologist can estimate the position of an epicenter with reasonable accuracy. More detailed studies take the observations of many stations and least square the solution for greater accuracy. Sometimes if the earthquake is very well recorded a good approximation of the direction of the epicenter can be obtained from the seismograms at one station. This
method makes use of the fact that \( P \) waves are longitudinal and the first motion is a vibration back and forth along the path. In such a case, with distance and direction from his station the seismologist can get a fairly good epicentral location without data from elsewhere. A good knowledge of where earthquakes have occurred before is also helpful—there are some regions of the earth’s surface where large earthquakes have not occurred in historical time, and it is a fair bet when trying for a quick location that a given shock did not come from one of these regions. However, one may get into trouble using this. Many years ago large earthquake was recorded at Berkeley which had its epicenter in Baffin Bay. The records gave the correct distance from the \( S-P \) interval and the first \( P \) waves indicated the direction to be northeast of Berkeley. But no large shock had ever been recorded from Baffin Bay, whereas many had been registered from the same distance in the Aleutian Islands. Remembering this, forgetting the science of his trade and applying the art, the author reported the epicenter as in the Aleutians. He was wrong.

But in another case the probability worked well. A large earthquake occurred in Turkey. The distance was correctly computed from \( S-P \). The direction, also, was found. Turkey seemed more likely than Iran because Turkey regularly has large shocks. It was reported as a Turkish earthquake to the United Press. They wired it to Ankara. It was several hours later that Ankara heard from the outlying Turkish provinces which had felt the shock. So the Turkish News Service wired congratulations!

When the epicentral distance gets much over a quarter of the way around the earth, the seismogram becomes more complicated. One enters the “shadow zone”, the core casts a shadow in the seismic rays, \( S \) ceases, \( P \) becomes weak, and the large early waves are reflections at the earth’s surface. For we record at many distances waves of both \( P \) and \( S \) types reflected both at the earth’s surface and at the core boundary. Frequently waves twice reflected are registered strongly. The core acts as a spherical lens of great aberration. About 10,000 miles from the epicenter we begin to get a strong \( P \) wave focused by the core. The seismogram gets very complicated and the theory more so.

**Surface Waves**: There are also two kinds of surface waves. The faster are called Love waves after the mathematician who first worked out their theory. They exist because the speed of shear waves increases with depth. They are shear or transverse waves with the vibrations horizontal. Their speeds depend on their wave length, i.e., on their depth of penetration. Waves of period 60 seconds travel about \( 2\frac{1}{2} \) miles per second. The second kind of surface waves are called Rayleigh waves after the Lord Rayleigh who first worked out their theory. As they pass
the earth's surface particles move in elliptical paths which lie in the vertical plane containing the path of the wave. They travel somewhat slower than Love waves. Both of these surface waves are dispersive—longer waves travel faster than shorter waves. Their dispersion depends on the way the speeds increase with depth and this fact is used to draw conclusions regarding the structure of the earth's surface layers.

In addition to the dispersion of surface waves which gives only a broad averaged idea of layering, seismological evidence for the layered nature of the earth's crust is based on two other kinds of observations. (1) Seismic waves penetrating the crust are refracted at the bottom of a layer (at the critical angle) and run along the boundary with the speed in the lower medium. These waves feed energy back into the layer. The path is not an "optical path". Although the path follows Snell's Law, it is not a strict Fermat path. Theoretically, the amplitude of such a wave should vary as the inverse square of the distance instead of the inverse first power. (2) Seismic waves may be reflected at a boundary between layers.

The second kind of observation is the mainstay of those who study the earth's layering by means of waves set up by explosives. This method requires particular distribution of seismographs in relation to the position of the source of the waves. Such distribution cannot be accomplished for earthquake observations since we do not know when and exactly where earthquakes will occur.

Seismic exploration is carried on extensively in search of geologic structures favorable to petroleum accumulation. The interest of these prospectors is in the sedimentary rocks of the upper crust. It has been found that to be certain of a reflection it is necessary to have an array of instruments fairly closely spaced. Only if the suspected reflection records on all the instruments can one be sure that there is a good reflecting surface which has stratigraphic significance. Sporadic large waves may record on one receiver and not on others—perhaps reflection from a small lens or even a large boulder. Such reflections are often said to be due to "scattering". An array of receivers is usually along a line. Six or twelve or even more receivers are set one hundred feet apart along this line. Recently in Montana Junger has observed reflections for which the travel time was as great as 7.5 seconds. This represents a reflecting surface at depth of the order of 20 kilometers—the bottom of the geologists' "granitic" layer perhaps. These reflections were observed in connection with a regular prospecting set-up—the records were merely kept running longer than is common practice. However, not in all localities were such reflections observed. And where late reflections were observed they were not always as late—sometimes came
in at 5 seconds or less. It would appear then that great irregularities in earth structure persist much deeper than seismologists sometimes have thought. That deep discontinuities should be irregular in mountainous regions, particularly, seems highly reasonable. That would mean that only in some set-ups would we be able to get a good reflecting surface which would return sharp waves to the vicinity of the source. Prospectors learned early to set up near the shot in order to record reflections, for the shot sets up very strong surface waves, known as "ground roll". A set-up near the source allows much of this "ground roll" to have passed the receivers before the reflections come in.

Early explorers tried setting up near the point where they expected the totally reflected wave to arrive—the place of arrival of the wave reflected at the "critical angle." The comparatively shallow depths at which they wished to find reflecting surfaces resulted in distances at which the ground roll was very bad.

Recently Tuve and Tatel have fired very large quantities of explosives seeking reflections at deep discontinuities. They find very large reflections, near the critical angle, from the Mohorovicic discontinuity at depths 30 to 40 km. For such depths the critical distance is around 100 km. and the ground roll trouble does not persist so far out. However, they have not found these reflections in their Pacific Coast experiments although refraction earthquake studies indicate their existence. Apparently then the Mohorovicic surface is very rough under this mountainous country.

Byerly, studying earthquake records, showed long ago that the Sierra Nevada in California had a great root extending well below the normal Mohorovicic surface. This root casts a shadow for the $Pn$ waves which travel just below the discontinuity. The seismographic stations at Tinemaha and Haiwee in Inyo County, California, are immediately in the eastern lee of the Sierra Nevada and do not record $Pn$ from coastal shocks.

From the lack of observation of waves reflected at the Mohorovicic surface in Washington state and in southern California (Tuve and Tatel), it appears that the surface is rough. But only under the great Sierra Nevada is it sufficiently rough to block the "refracted" boundary wave $Pn$. 
The Seismograph

Seismometers are instruments generally designed to measure very slight motions of the earth, although a few are designed to measure hard shaking. When a recording device is added to a seismometer the combination is a seismograph. These instruments measure only the ground motion where they are placed—there is no mystic premonition of events afar. A large earthquake shakes the whole earth. As long as the size of the shaking is larger than the “noise” background at the station, the seismic waves can be recorded. However, as the sensitivity of the seismometer is pushed up, small vibrations of the earth which are called microseisms are always encountered. Some are caused by traffic, and fall off at the hours when traffic is light. Some are caused by neighboring machinery. But the most troublesome have to do with weather. Great storms at sea can be correlated with heavy microseisms at points far distant in the interior of continents. It appears now that many emanate from a place near the “eye” of the hurricane. Recent theory indicates that stationary ocean waves could transmit considerable energy to the
bottom. Microseisms are thought by some to be connected intimately with "cold fronts."

Some investigators have found very satisfactory correlation between microseisms and surf height on nearby coasts or even distant coasts. Certain types have been blamed on frost and on rain and on heavy winds blowing on nearby mountains. When the desire is to record earthquakes, microseisms become very troublesome.

Seismometers are almost all pendulums of some type. Normally a clock is thought of when pendulums are mentioned. In the clock the pendulum mass swings while its support remains stationary. In the seismometer the reverse is the case. The support which is fastened to the earth shakes because of the earthquake, while the mass tends to remain stationary or at any rate to move differently from the support. This difference in the motion of support (earth) and mass is recorded on photographic paper or film. From the seismogram we can compute how the earth moved if the magnification system of the seismograph is not too complicated.

The simplest seismographs have mirrors on their masses. When the pendulum moves relative to the earth the mirror rotates. A beam of light plays on the mirror and is reflected and focused on photographic paper wrapped around a drum. The drum rotates and also translates. The spot of light writes a straight line on the drum when the earth is quiet, but when the earth moves the spot writes a record of its motion. The next most complicated system introduces a galvanometer which is fed by a transducer on the pendulum and its support. The transducer consists of a coil of wire, a magnet, and perhaps an iron core. Any two can be fastened to the pendulum with the third fastened to the support or vice versa. Relative motion of pendulum and support induces an electromotive force which operates the galvanometer. The mirror on the galvanometer records in the same way as the mirror on the simpler seismometer. As for the third degree of complication, electronic amplification may be introduced.

To a physicist reading the above, the suggestion that galvanometers and electronic systems are complicated may prove distasteful. However, he is accustomed to performing experiments at a time suitable to himself with all apparatus adjusted just before the experiment. For a seismologist, this situation occurs only when his source is a prearranged blast. In recording earthquakes, he does not know when the "experiment" is to be performed or just where. Therefore, the instruments must operate night and day, year in and year out. The cost of having a trained technician in constant attendance is prohibitive. A technical expert cannot adjust the setup just before the experiment. If such an expert could adjust the equipment every day, many recordings of shocks
which would occur while he was working on the seismographs would be lost. So the seismograph is in operation continuously, often in an out of the way place, its records changed daily by a man who may not understand its principles and may not care to learn. So the instrument needs to be very simple and foolproof.

SIZE OF EARTHQUAKES

There are a number of ways of evaluating the significance of an earthquake.

The oldest is to judge it by its effects on man and by the damage it does. Such a rating is called the “intensity” of the earthquake. On such a scale the least intensity represents a shock barely felt by a few experienced observers. Earthquakes are not phenomena for which familiarity breeds contempt. It is the “experienced observer,” one who has felt earthquakes before, who is most alert and easiest frightened. The greatest intensity would involve total destruction of buildings, great fissures in the earth and landslides. The most commonly used intensity scale in the United States at present is the Modified Mercalli scale. You may compare it with the older Rossi-Forel scale presented earlier. It reads as follows (in its abridged form):

1. Not felt except by a very few under especially favorable circumstances.
2. Felt only by a few persons at rest, especially on upper floors of buildings. Delicately suspended objects may swing.
3. Felt quite noticeably indoors, especially on upper floors of buildings, but many people do not recognize it as an earthquake. Standing motor cars may rock slightly. Vibration like passing of truck. Duration estimated.
4. During the day felt indoors by many, outdoors by few. At night some awakened. Dishes, windows, doors disturbed; walls made cracking sound. Sensation like heavy truck striking building. Standing motor cars rocked noticeably.
5. Felt by nearly everyone; many awakened. Some dishes, windows, and so forth broken; a few instances of cracked plaster; unstable objects overturned. Disturbance of trees, poles, and other tall objects sometimes noticed. Pendulum clocks may stop.
6. Felt by all; many frightened and run outdoors. Some heavy furniture moved; a few instances of fallen plaster or damaged chimneys. Damage slight.
7. Everybody runs outdoors. Damage negligible in buildings of good design and construction; slight to moderate in well-built ordinary structures; considerable in poorly built or badly designed structures; some chimneys broken. Noticed by persons driving motor cars.

10. Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations; ground badly cracked. Landslides considerable from river banks and steep slopes. Shifted sand and mud. Water splashed (slopped) over banks.


12. Damages total. Waves seen on ground surfaces. Lines of slight and level distorted. Objects thrown upward into the air.

There is some difficulty in rating the intensity—the more punctilious the observer, the greater his difficulty.

One trouble arises from the combination in the scale of effects on man and effects on inanimate objects. Much of our rating of intensities, particularly in the lower ranges, is based on post card reports sent to the U.S. Coast and Geodetic Survey by citizens in the area shaken. Someone may send a report which underlines the phrase “frightened all,” yet gives no further information. Is the earthquake then to be rated as VII leaving the reader of the report to infer that chimneys and walls were cracked? In developing an intensity scale, perhaps fright should be left out as a criterion.

Another difficulty in the application of the intensity scale is psychological. It goes back into the history of the use of these scales. There are two views as to why we wish to evaluate the intensity. The modern view is that there should be put on record in abbreviated form, the damage suffered in a given area. The old view was that the end was to find the epicenter of the earthquake. It is customary to display the results of intensity ratings on a map. A closed curve is drawn around the region of highest intensity, then another closed curve surrounding the first but including in addition all localities where the intensity was next highest, and so on. Such curves are called isoseismals. Very often these isoseismals are highly irregular. For example, in California the floor of Yosemite Valley seems quite sensitive to earthquakes. Many earthquakes centering in the Coast Ranges are reported felt in the Yosemite Valley but not elsewhere in the Sierran region. To a seismologist of the older school, there is a temptation to apply some sort of “foundation” coefficient to observed intensities in certain areas in order to iron out irregularities. To the other seismologists, such a procedure removes the main value of the rating—namely to show that certain localities are hazardous and that there particular care
should be exercised in construction. From the older viewpoint the "true intensity" at a locality was to be measured by what would have happened if the town had been built on solid rock and all the buildings well constructed.

Almost without exception we find that modern seismologists and engineers have a sort of contempt for intensity. This is hard to understand, for the intensity is a picture of damage done and that seems to be the most important item about an earthquake. It appears that seismologists and engineers desire only a measure of the total energy in the earthquake.

Attempts have been made to compute from a careful evaluation of all the larger waves on one or two seismograms the energy at the source. The difficulties are considerable. The first assumption is that the energy is sent out equally in all directions from the source. This is not so for P or for S or for L or for M, but perhaps for the sum total it may not be so bad. However, for P or S, waves which penetrate the interior, the warping of cones of energy due to refraction is so bad that we cannot hope for much. One may use only the surface waves and simplify the problem a little. These waves, being confined to the earth's surface, are probably not so badly distorted by refraction.

In the rare case, 1906, of a fault breaking to the surface in a region where geodetic surveys gave an estimate of the region involved in the strain which preceded the shock, Reid was able to make an estimate of the strain energy accumulated before the shock. He computed $10^{24}$ ergs. Benioff recently recomputed the energy in the 1906 earthquake using Reid's method but assuming that the rocks along the San Andreas Fault are weaker than normal (to explain why the break occurs there). He got $10^{28}$ ergs. Gutenberg and Richter, basing their estimation on the "magnitude" of the earthquake, obtain $10^{26}$ ergs.

The "magnitude" of an earthquake was first defined by Richter as the logarithm to the base 10 of the maximum trace amplitude in microns recorded on a Wood-Anderson short period seismometer at an epicentral distance of 100 kilometers. Since there is rarely a seismograph at this distance, let alone a Wood-Anderson, it was necessary to establish rules for interpolation and extrapolation. Richter and Gutenberg have pushed the matter using semi-empirical, semi-theoretical relations among intensity, magnitude, acceleration and energy, until they now estimate any of these quantities from the Pasadena seismograms. An earthquake of magnitude 8\(\frac{1}{4}\) is very large indeed (1906).

An example then of our present ability to evaluate the energy of a shock is then the 1906 earthquake in California. Its energy was between $10^{23}$ and $10^{26}$ ergs. The latter is one thousand times the first.
The severe shaking of a large earthquake often produces disastrous effects on loose soil and on structures built by man.

**Effects on Ground**

*Lurches.* Soft bottom land in the vicinity of river beds and man-made fill always suffers greatly in a large earthquake. The earth is thrown about, for it has little cohesion. Some areas may settle. Large cracks may open, usually parallel to the river bed in the case of bottom land. Struc-

![Figure 7. Earth Lurches, Imperial Valley, 1940.](image)

structures built on such a foundation are best built on piles or concrete mats. Sandy areas may suffer similarly. It is reported that after the 1663
earthquake in the St. Lawrence valley region of eastern Canada, acres of forest appeared as plowed ground.

*Slumps.* After a rainy season, hillsides on which there is a mantle of water-soaked decomposed rock often slip a little even without earthquakes. Such slumps can be seen on hillsides frequently. A large earthquake of course increases the number and magnitude of slumps. On occasion whole hillsides have "flowed" away.

*Avalanches.* A very severe shock may cause avalanches of dry rock and soil if the slope is near the angle of repose. Avalanches are less common than slumps.

*Mud flows.* It is common for a large earthquake to so disturb underground water-bearing strata that water is forced to the surface. If this occurs on a hillside and the water is voluminous enough, it may result in a large mud flow, the water mingling with the soil. Such a flow broke trees in the 1906 earthquake.

*Earthquake fountains and sand blows.* If the water forced up from below emerges on a level terrain, it creates a crater of sand about it. It gushes up in a fountain, may flow for several hours and bring up sand from considerable depths (80 feet in 1906 in the Salinas Valley). Sometimes merely gas is forced up without water. Such a "blow" often leaves a crater similar to a fountain. Steamboat Springs, south of Reno, Nevada, flowed muddy for several days after the 1906 earthquake.

*Auxiliary cracks.* Near the center of a great shock there may be many cracks even in good rock which are considered an effect of the shock.
because of their apparently haphazard distribution, rather than as faults which caused the shock.

**Effects on Structures of Man**

It is clear that man-made structures are likely to suffer especially if they are constructed on any geologic foundation which is shifted by an earthquake in one of the ways discussed above.

It is therefore of first importance to select a site for a residence which is not on man-made fill, preferably not on recent stream fill, and if on a steep hillside then on one whose rock is near the surface and where there is no evidence of earlier sliding. A filled-in swamp is about the worst foundation possible.

![Image of Church, Santa Barbara, 1925.](image-url)
Figure 12. Residence, San Francisco, 1906.

Figure 13. Residence, Helena, 1935.
Figure 14. Portland, 1949. (Courtesy The Oregonian).
Sometimes it seems necessary to construct commercial buildings on filled in ground, e.g., lower Market Street in San Francisco. Experience has indicated that buildings on piles stand up rather well. The Ferry Building in San Francisco weathered the storm of 1906. It is founded on some 5000 piles held up by skin friction. Concrete mats are presumably good. They require all parts of the foundation to move together. Of course if the mat were to tilt, the building might be less useful afterward. A structure on loose soil has an experience during a large shock which is much that of a boat at sea.

In 1906 a study of the effects in the city of San Francisco showed most clearly the effect of geologic foundations. On the rocky hills damage was least. Some people were not even awakened by the shock which occurred a little before 6 a.m. Next least severe was the damage in small valleys in the hills, then the sandy level stretches, with the great damage in stream valleys and on man-made filled ground. One old hotel sank into the fill until its second story was at ground level.

Important lessons for the householder to learn are that mud sills
should be bolted to the foundations—the house must be a unit. Many houses fall off their foundations in a large shock. Diagonal bracing is imperative. Diagonal flooring and sheathing is most effective bracing. Vertical members should be continuous. The second floor should not be another small house perched on top of the lower house. Roofs should be tied to the walls. In many one-story commercial buildings the roof slides off the side walls—in others it pounds the walls out.

Figure 16. Boy killed by falling bricks at Castle Rock, Washington High School, 1949.

Bricks should be tied together by a good cement mortar—not held apart by poor mortar. Bricks should be wet when laid. Use of two types of building material in conjunction has not proved successful—a wood frame with brick walls is not good in a structure of any size.

Tie everything together—use plenty of nails—and cross brace wherever possible. Let engineers of skyscrapers talk about flexible structures if they must; in building a dwelling or small building do not try it. It is no doubt true that a house of vulcanized rubber would not suffer struc-
tural damage during a hard shock. But what would happen to its contents, human and otherwise?

During the last twenty years the United States Coast and Geodetic Survey has developed a comprehensive program of studying the effects of earthquakes.

A systematic organization for collecting reports of the effects of earthquakes both for minor shocks and great ones has been developed. Postcard questionnaires are distributed in areas where earthquakes are frequently felt. Such questionnaires are sent out after shocks in regions where they have not been distributed earlier, and also in regions where reports lag. If the shock is of any size, i.e., if it damages structures, a geophysicist of the Survey goes into the field to make a detailed report. Such work is invaluable in keeping our earthquake history. We must know all aspects of the earthquake problem if we are to do something effective in preventing earthquake damage. Too often superficial students of earthquakes tend to depreciate careful observation of earthquake effects in their efforts to obtain by instrumental means some estimate of energy in the shock, or its maximum acceleration. The pertinent object is to prevent damage and one must have an accurate idea of what damage has been done in the past if he is to prevent it in the future.

The Coast and Geodetic Survey has also developed an excellent strong-motion seismograph—a very insensitive instrument designed to record only major earthquakes. This seismograph is called the accelerograph. There are dozens of these accelerographs placed in earthquake regions in the western United States, principally in California where the hazard is greatest. Records have been obtained from a number of strong earthquakes: Long Beach (1933), Helena (Montana) (1935), El Centro (1940), Seattle (1949). Several of these records have been integrated, a laborious process, to yield an accurate picture of just how the earth moved at the place where the accelerometer was located. Such information is invaluable to the theoretical engineer. He can analyze the effect of such motion on simple vibrators and gain some idea of what the effects may be on structures.
# Table I

## LIST OF EARTHQUAKES FELT IN OREGON

**1866-1949**

The following quakes are given Rossi Forel intensities, which are represented by Roman numerals.

<table>
<thead>
<tr>
<th>Date</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>1866 June</td>
<td>&quot;Gold Hill News&quot;, June 2, 1866, Jackson County.</td>
</tr>
<tr>
<td>1866 December</td>
<td>III. The Dalles, Wasco County.</td>
</tr>
<tr>
<td>December 15</td>
<td>Aftershock? Felt in Oregon.</td>
</tr>
<tr>
<td>December 16</td>
<td>Aftershock? Felt at Eugene, Oregon.</td>
</tr>
<tr>
<td>1873 November 22</td>
<td>Felt from Portland, Oregon to San Francisco, California. Maximum intensity of VIII at Crescent City, California and Port Orford, Oregon. Chimneys damaged in Crescent City and Port Orford. Nearly every brick building in Crescent City was damaged.</td>
</tr>
<tr>
<td>1876 August 16</td>
<td>41° 55' N 126° 25' W. (Off coast of southern Oregon). Heavy. It is not certain that this shock was felt in Oregon.</td>
</tr>
<tr>
<td>1877 October 12</td>
<td>Cascades. Chimneys overthrown. (9 a.m.)</td>
</tr>
<tr>
<td>October 12</td>
<td>VIII, Portland; Marshfield, Cascades. Chimneys overthrown. (1:53 p.m.)</td>
</tr>
<tr>
<td>October 26</td>
<td>Latitude 43° 13' N, longitude 128° W. Severe shock. No report that this shock was felt on land.</td>
</tr>
<tr>
<td>1879?</td>
<td>Smart shock in Portland. (George Ainsworth).</td>
</tr>
<tr>
<td>1880 December 12</td>
<td>Puget Sound. VII. Felt from Victoria to Portland.</td>
</tr>
<tr>
<td>May 1</td>
<td>Portland.</td>
</tr>
<tr>
<td>1883 September 28</td>
<td>Portland. Two shocks. (Not reported in Washington or California.)</td>
</tr>
<tr>
<td>1885 October 10</td>
<td>III East Portland. Three very light shocks.</td>
</tr>
<tr>
<td>1891 September 16</td>
<td>Salem. The shock was brief and distinct and was followed by a wave-like motion. Windows rattled.</td>
</tr>
<tr>
<td>1892 February 3</td>
<td>Portland: Brick buildings swayed and windows rattled, terrifying the inmates, who in many instances rushed into the street. Astoria: Houses shook. Salem: Windows rattled and buildings trembled. No damage.</td>
</tr>
<tr>
<td>April 17</td>
<td>Portland. Two heavy shocks. Many persons became frightened and rushed into the street when the buildings began to tremble. No damage.</td>
</tr>
</tbody>
</table>
1893 March 6 Umatilla. Succession of shocks. One wall of a large stone building was thrown down. Not reported elsewhere.
1896 April 2 Portland. McMinnville. People awakened. Two or three shocks with loud rumbling noise from the west.
June 5 Cape Blanco Lighthouse, Curry County. (Probably an earthquake.)
1897 January 26 Newport, Lincoln County. Sharp. Lasted about three seconds.
1897 December 6 Forest Grove, Washington County. Slight.
1902 June 14 Newport, Lincoln County. Sharp shock; no damage.
June 15 Newport. Sharp; no damage.
December 2 Kerby, Josephine County. Slight.
December 4 Hood River.
December 18 Fox Valley, Linn County.
1906 April 2 Ashland, Jackson County.
April 12 Ashland, Jackson County.
1906 April 18 San Francisco earthquake. Felt in southern Oregon.
April 19 Paisley, Lake County. Strong enough to awaken people generally. Three other shocks in the following one and one-half hours.
April 23 Grants Pass. Furniture moved. Several windows cracked.
April 29 Paisley, Lake County. Milk spilt.
1909 October 28 Felt in southwestern Oregon. IX in Humboldt County, California at Rohnerville and Fortuna.
1913 March 15 Medford, Jackson County. Three distinct shocks.
March 15 Roseburg, Douglas County. Dishes, etc. rattled for three seconds.
October 14 Seven Devils Region. Along the boundary between Oregon and Idaho. Broke windows and dishes. Felt at Homestead, Baker County.
1914 March 22 Portland. Jar, lasting five or ten seconds. Felt by hundreds.
1915 January 18 Summerville, Union County. Felt in surrounding district.
May 19 Portland. Reported as though localized in eastern part of Portland. Rattled dishes, rocked chairs, disarranged books in cases.
October 2 Baker, La Grande, Ontario. People frightened and hurriedly left hotels. This was the destructive shock of Pleasant Valley, Nevada.
October 19 Fruita, Wallowa County. One shock.
1916 January 4 Newport, Lincoln County. Rattled dishes.
1919 December 25 Bullrun, Clackamas County. First shock awakened people; other shocks through the night.

1920 April 14 V Crater Lake, Klamath County. Three shocks; felt at Fort Klamath.

1920 November 9 Portland. Felt by several; duration five seconds.

1920 November 28 Northern Oregon and southern Washington. Turner, from instrumental data, located the origin of a shock at this time off the northwestern shore of Vancouver Island.

1921 December 15 Cascadia, Linn County. Shock was felt by everyone.

1921 February 25 Cascadia. Felt by nearly all, over an area of six by twelve miles.

March 4 Portland. Several shocks; felt by several people.

September 22 Portland. Duration 10 to 20 seconds.

1922 January 31 Epicenter at seat off Cape Mendocino, California. Felt in southern Oregon.

May 15 III to IV Portland.

July 5 III near Portland. Duration 15 seconds.

October 15 Hermiston, Umatilla County. Three shocks. Felt by several.

December 12 Pendleton district. Distinct.

1923 January 10 VI+ Lake County. The reports of perception point to a source near or north of Goose Lake in Oregon. Strongest in Lakeview.

January 22 Epicenter off Cape Mendocino. Felt in northern California and southern Oregon.

1924 January 5 V Stanfield, Umatilla County. Duration three to five seconds. Felt by several. Probably same shock reported on January 6.

January 6 V Milton and Weston, Umatilla County.

1925 March 19 Press reports say shock was felt in Oregon on this date, but give no details.

June 27 Baker, Portland. The severe quakes in Gallatin County, Montana, on June 27 were felt in several of the neighboring states. Baker and Portland were the only places reported in Oregon.

July 1 Lakeview, Lake County. Slight. Felt by several.

1927 April 8 V Richland, Baker County. Four shocks reported at points in Pine and Eagle Valleys, eastern Baker County on this night and the following morning.

1927 August 20 Humboldt Bay, California. Probably felt in southern Oregon.

1928 September 4 Newport. Slight shock; felt over 10-mile radius.

1930 July 8 Perrydale. Slight.

July 18 Perrydale. Cracked plaster, etc. Crack appeared in roadbed one-half mile west of Perrydale. Also felt in McCoy.

The following quakes are rated by the Modified Mercalli intensity scale. (In intensities given by Arabic numerals.)
1931 August 16 5 at Talent, Oregon, where press reports lamp shaken from ceiling and one man thrown from chair. Also felt at Phoenix, Talbot, Ashland.
August 23 Quake off California coast near Eureka felt at Klamath Falls, Beagle, Curtin, and Hugo, Oregon.
September 3 Central Point. Feeble.
October 1 At sea, (ship report), heavy tremors at 42° 56' N, 124° 44' W. Sulfurous odors observed.
1933 November 23 Portland. 3. Several awakened.
1934 July 6 North Bend, Oregon. 3.
1936 May 8 Roseburg, Oregon. Meteor exploded, awakening many. (?)
July 15 45° 58' N, 118° 18' W. “State Line Earthquake” near Walla Walla, Washington and Milton, Oregon, 7+. Damage about $100,000. Area affected, about 105,000 square miles. The shock was strongest at Freewater, State Line, and Umapine, Oregon, where the intensities reached 7. The ground was badly cracked and there were marked changes in the flow of well water. In the cemeteries about 70 per cent of the stones rotated clockwise, viewed from above. Some stones in close proximity to each other rotated in opposite directions.
In Freewater, practically all chimneys over 10 years old were damaged at the roof level.
A fine new house about 4 miles west of Freewater was almost completely wrecked. Two cement houses about 20 years old, 7 miles west of Freewater, were practically demolished.
Four miles west of Freewater the ground was cracked over an area 1,200 to 1,500 feet long by 50 to 100 feet wide along the base of a hill running east to west. One crack some 200 to 300 feet long was from one to six feet wide.
At State Line concrete pavements were cracked.
At Umapine many walls and chimneys were cracked and a few were demolished. Several houses were badly damaged.
July 15 - November 17 Many aftershocks were felt at Athens, Milton, Freewater, Umapine, and Helix, Oregon, and at Walla Walla, etc., in Washington. Many of these were of intensity 4. Two, July 18, at Milton and Freewater, Oregon, and August 4, at Helix, Oregon, were of intensity 5.
1937 February 8 Ferndale, Oregon.
December 14 Dallas, Oregon. Seismic origin doubtful. Disturbances reported from Fall City to Tillamook; thought by some to have been due to a meteor.
1938 May 28 Probably off coast of northern Coos County, Oregon. Trees and bushes shaken at Gardiner. Pictures fell at North End.
October 27  Dry Creek (trees and bushes shaken) and Milton, Oregon. Slight.
April 13  Portland, Oregon. Distinct tremors felt by few.
November 12  Epicenter about 47°2 N, 123°9 W. Felt over most of Washington and in a small part of northwest Oregon.
1940 May 25  Oregon west coast. Lincoln County. Small objects moved at Waldport.
1941 February 16  Portland. Light shock.
July 6  Medford. Felt by fire lookouts.
October 19  Seal Rocks. Slight.
October 31  Portland. Doors rattled.
December 29  Portland. Store display windows shattered and other windows broken. Felt over 5,000 square miles.
1942 May 12  Maximum 5 at Corvallis. Felt by many.
November 1  Portland. Cracked plaster, rattled windows.
November 1  Madras, Oregon. Light shock. Rattled doors.
1943 June 24  Bend, Oregon. Slight. Felt by all upstairs.
August 4  Rockville, Oregon. Related to steam vent. Continuous, minor tremors.
1944 March 5  Dallas, Oregon. Sharp. Felt as if house struck by truck.
September 19  Rockville. Rapid motion; windows rattled.
1946 February 14  47°3 N, 122°9 W. 6 and 7 in Puget Sound Area. Most damage in Seattle. 6 at Portland.
June 23  49°9 N, 125°3 W. Georgia Strait, B.C. 8. 5 at Portland. 4 at Corvallis.
1947 December 24  Klamath Falls. Buildings creaked, etc. No damage.
1948 February 29  Buxton, 4.
March 21  Hildebrand (southeast section) 5.
1949 April 3  Grants Pass, 3.
April 3  Klamath Falls, 5.
April 13  Seattle earthquake. 7-8 in northern Oregon.
REFERENCES

HOLDEN, EDWARD S. Catalogue of Earthquakes on the Pacific Coast, 1769 to 1897. Smithsonian Miscellaneous Collections No. 1087, 1898.


