Earthquakes in North America

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DURING THE PAST TEN YEARS considerable progress has been made in determining the seismicity in a given area—the frequency of occurrence, and distribution of earthquakes. Earlier investigations were based almost completely on field observations, but now extensive use of instrumental records is possible. This assures much more uniform results for the whole earth. The use of seismograms in investigations of seismicity was made possible by the development of methods which permit a rapid calculation of a function of the earthquake energy from instrumental observations. The first seismogram of a distant earthquake that was recognized as such was made on April 17, 1889, when an instrument at Potsdam wrote a record identified as that of a shock in Japan. During the following years instruments were designed which gave fairly good records of distant earthquakes. In 1897, a committee of the British Association for the Advancement of Science called attention to the desirability of observing earthquake waves that had traveled great distances. By 1899, thirteen stations provided such observations and the results were analyzed. In 1904 the number of stations reporting had increased beyond one hundred, but less than half of them reported wave arrival times reliable within about a quarter-minute. From that time on, however, it has been possible to locate within a few hundred miles all great earthquakes and most major shocks. In 1907 the International Central Station at Strasbourg issued the first catalogue giving all readings for the larger shocks reported for 1904. Thus, starting with 1904, research on seismicity could be based on instrumental observations. The systematic publication of such data was discontinued during the first world war (when the catalogue for 1908 was in press) and later was resumed, starting with the data for 1918. For the years 1912 to 1917 summaries for selected shocks were published. Detailed data concerning arrival times of waves at the reporting seismological observatories are printed in the "International Seismological Summaries." In addition, these volumes contain calculated values of the coordinates and depths of the earthquake foci and the origin times of the shocks. They were formerly compiled at Oxford, England, and now at Kew. The summaries are based on the bulletins that are issued by most seismological stations. Some of these station bulletins contain in addition to observed times of various phases the calculated amplitudes of the ground motion. With this information, it is possible to determine the size of the earthquakes. The great importance for research of all such station bulletins and international catalogues is obvious.

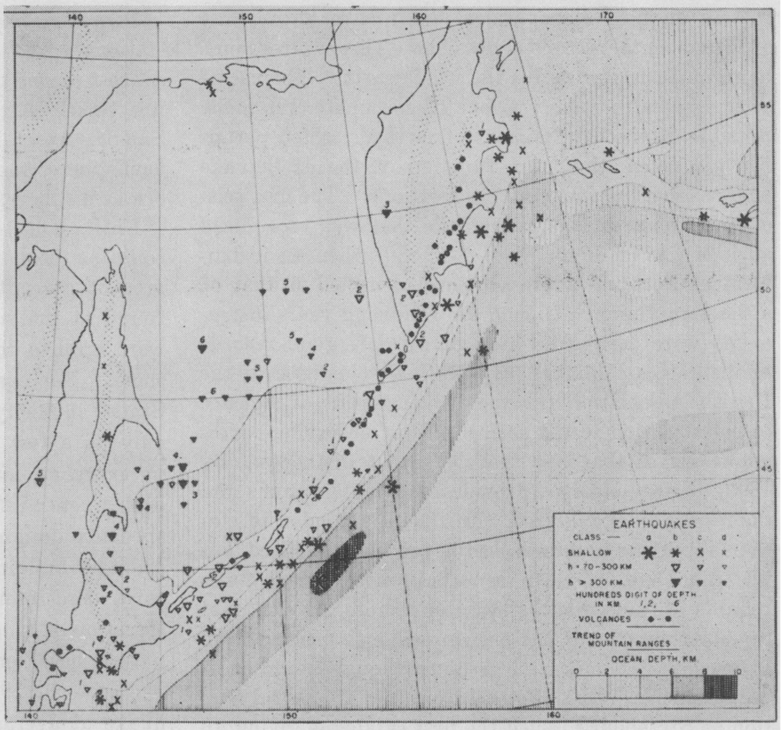
There are now roughly three hundred seismological stations with accurate time service (at least to the nearest second) practically all over the world, including South Africa, South America, New Zealand, Samoa, Australia, and Madagascar in the southern hemisphere, and a much denser network in the northern hemisphere.

Until about ten years ago the size of an earthquake could be estimated only from the observed size of the area of perceptibility or of damage or from changes found at the surface of the earth. Arbitrary scales were applied to such data to find the *intensity* of a shock. For example, in the scale used in the United States, intensity II indicates that the shock was felt only by a few persons; intensity V, that it was felt by everyone, many were awakened, some dishes were broken, etc.; intensity VIII indicates slight damage in specially designed structures, considerable damage in ordinary buildings, great damage in poorly built structures; and intensity XII, the maximum, indicates destruction of all structures. A scale of wholly different nature, based on instrumental data, was devised by C. F. Richter in 1935. He defined *magnitude* of an earthquake at average (shallow) depth in southern California as the common logarithm of the maximum trace amplitude expressed in thousandths of a millimeter, with which the standard short period torsion seismometer (period 0.8 seconds, magnification 2800, damping nearly critical) would register that earthquake at an epicentral distance of 100 kilometers. Magnitude $M=2$ corresponds in shallow earthquakes to a shock barely felt; a shock of magnitude 5 causes minor damage; magnitude 7 is the lower limit of major earthquakes; $8\frac{1}{2}$ is the highest magnitude that has been determined from amplitude data given in individual bulletins of seismological stations since 1904. This magnitude scale was later extended by Gutenberg and Richter to apply to shallow earthquakes occurring in other localities and recorded by other types of instruments. In 1945 Gutenberg devised means for determining magnitudes of shallow earthquakes using amplitudes and periods of waves that had traveled through the interior of the earth. He also extended the scale to include deep focus earthquakes. It is now possible to determine the magnitude of larger earthquakes within a few tenths of the scale from seismograms at any well-equipped station. The relationship between magnitude M of an earthquake and its energy E in ergs is given roughly by the approximate equation $\log E = 12 + 1.8\ M$. This holds for any focal depth. The data concerning the magnitude and the instrumentally determined epicenters and depths of foci of earthquakes provide the basis for seismicity studies.

The results of such an investigation of earthquakes recorded over the period from 1904 to 1947 has been published by Gutenberg and Richter in a book, *Seismicity of the Earth,* published by Princeton Press in 1949. Much of the following information is taken from this book.

The use of magnitudes for the first time provides reliable information concerning the relative seismicity of all regions of the earth. It eliminates the effects of density of population and of communication facilities on the determination of intensities of reported earthquakes, as well as effects of uneven distribution of seismological observatories on seismicity patterns. If the magnitude of the earthquakes is not considered, distorted appearance of seismicity maps may result from an accumulation of many small shocks, which are plotted only in regions well covered by stations with sensitive instruments. Thus, Europe—which, except for the Mediterranean area, has a low actual seismicity—has appeared on maps in the past as a region of relatively high seismic activity. There are now five stations reporting magnitudes of earthquakes in their routine bulletins, but many more regularly furnish amplitude data required for the magnitude determination. Magnitude can be determined from a seismogram at any station where instrumental constants are known and where a clear record of an earthquake has been written, regardless of the distance or depth of the shock. Magnitudes determined at different stations rarely differ by more than one fourth from the average for a given earthquake.

The outer part of the earth consists of relatively inactive blocks, separated by active zones falling into four groups: (1) the circum-Pacific zone, which includes about 80 percent of all shocks with origins at a depth not exceeding 60 kilometers (about 40 miles), 90 percent of the so-called intermediate shocks, which have their sources at depths between 60 and 300 kilometers (about 40 and 190 miles), and all deeper shocks (maximum observed depth approximately 400 miles). (2) The Mediterranean and trans-Asiatic zone, which includes nearly all remaining intermediate and large shallow shocks. (3) Narrow belts of shallow shocks, which follow the principal ridges in the Atlantic, Arctic, and Indian Oceans. (4) Moderate activity associated with rift structures such as those of East Africa and the Hawaiian Islands.



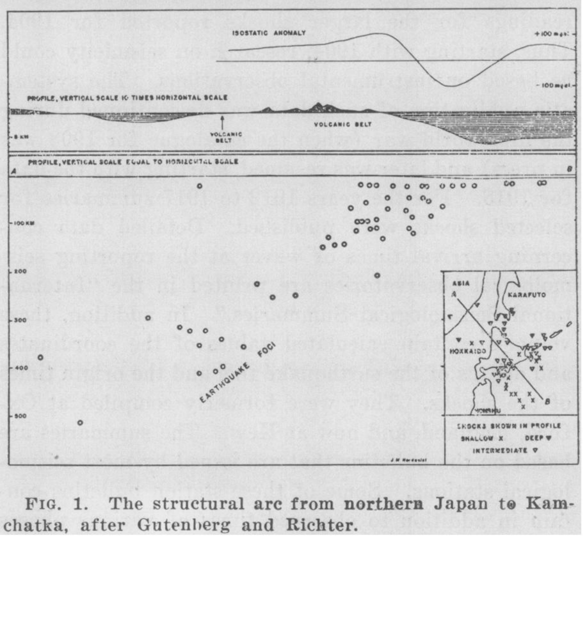


FIG. 1. The structural arc from northern Japan to Kamchatka, after Gutenberg and Richter.

The most extensive inactive block is the Pacific basin (excluding the Hawaiian Islands). On the continents, most of the ancient shields are quite inactive. Between the stable shields and the active belts are regions of minor to moderate activity having occasional large shocks. Small shocks (magnitude 5 and less) apparently occur everywhere.

A structural arc of the Pacific region—for example the Tonga arc, the Marianas arc, or the northern Japan arc—exhibits the following typical features in order, beginning at the convex side: (A) a foredeep; (B) shallow earthquakes and negative gravity anomalies along anticlines; (C) positive gravity anomalies and slightly deeper shocks; (D) the principal mountain arc (Tertiary or older), with active volcanoes and shocks about 100 kilometers deep; (E) an older structural arc with volcanism in a late stage or extinct, and shocks about 200–300 kilometers deep; (F) a belt of deep shocks (below 300 kilometers). In some arcs only a few of these features can be identified; this is true of the similar structural arcs along the southern Alpid front of the trans-Asiatic zone. In parts of the Pacific belt (for example, along the coast of the continental United States and British Columbia) structural arcs and the accompanying features are absent. In many such sectors (as in California) there is strong evidence of block faulting in place of the folding characteristic of the arcs.

The seismicity of North America is mainly associated with the Pacific belt. Relatively high activity occurs in the area of the Aleutian Islands. The Aleutian arc is a typical Pacific arc; it extends from the Commander Islands into central Alaska. Seismic and volcanic activity is relatively high. In general, shallow seismic activity follows the northern concave side of the Aleutian trench. Intermediate shocks at depths down to about 100 miles occur along the north side of the island arc. No shocks originating deeper than 200 miles are known in the area of the North American continent. The shocks having depths of approximately 60 miles occur near the line of volcanoes, as usual. Shallow shocks in the interior of Alaska represent an interior structure.

Another sector of the Pacific belt extends from southeastern Alaska to Puget Sound and includes the rather active area of the Queen Charlotte Islands, where a great earthquake occurred in August, 1949. There are neither well-developed ocean deeps nor shocks at intermediate or greater depth in this area. The seismic activity decreases considerably in the vicinity of the state of Washington. There is a clear gap between this and the next seismic zone, which begins about 200 miles off the coast of Oregon. Thence, an uninterrupted belt of earthquake foci extends in a southeasterly direction. It reaches the coast of northern California, then follows the coastal area to the region of San Francisco and continues inland following the well-known San Andreas fault zone. This zone has been traced at the surface as far south as the Salton Sea, but the earthquake belt continues along the Gulf of California at least as far as the southern tip of Lower California. Volcanic activity is low along this zone; the few volcanoes, such as Mount Lassen, and Tres Virgenes in Lower California, appear to be in a late state of activity.

The next sector to the southeast is one of noticeably higher activity. It follows the Pacific Coast from Colima in Mexico to Panama. There are two lines of active volcanoes, one extending west–east across central Mexico from Colima to Vera Cruz, the other beginning in Guatemala and extending southeastward through Central America. Accompanying the line of active volcanoes, once more earthquakes are found at depths of somewhat less than 100 miles. Mexico City is in the west–east belt of intermediate shocks and consequently experiences rather frequent earthquakes; however, they usually cause relatively little damage as a result of their considerable depth below the surface. Ocean deeps off the Mexican coast are well developed and include the Acapulco deep and the Guatemala trench. Unfortunately, gravity measurements are very scarce off the whole Pacific Coast of North America but the few data available indicate appreciable negative gravity anomalies, at least off the coast of Mexico in the neighborhood of the ocean deeps.

The earthquake belts mentioned thus far are responsible for most of the seismic activity in North America. In the United States, for example, the California–Nevada region contains about 90 percent of the whole seismic activity. This result is based mainly on instrumental data covering the past 40 years, but is in good agreement with historical information. The remaining shocks are partly situated in areas marginal to stable masses, partly in regions which have undergone higher tectonic activity in the not too distant geological past. The Rocky Mountains and related structures, which are of the same age as others belonging to the circum-Pacific belt, show relatively low seismicity.

Other regions with occasional earthquakes include the area between Siberia and Alaska. This is transgressed by the Bering Sea, whose coasts are probably of structural significance, since practically all the known shocks of the regions are close to them. Activity marginal to the Canadian shield includes a major earthquake off Newfoundland in 1929; to the northeast, marginal shocks have occurred in Davis Strait and Baffin Bay. There is notable activity along the St. Lawrence River.

The Appalachian belt is a region of fairly frequent minor activity. The northeastern part of it is shaken occasionally by marginal shocks of the Canadian shield, and some moderate earthquakes originate within the Appalachian area. Near the Atlantic coast is the epicenter of the Charleston (North Carolina) earthquake of 1886. Historically the greatest shocks in the United States outside the Pacific area are the earthquakes of 1811 and 1812 in the Mississippi Valley, which originated near New Madrid (Missouri); their magnitudes possibly surpassed magnitude 8. It is of interest to note that shocks east of the Rocky Mountains seem to originate occasionally at depths of about 30 or 40 miles below the surface, which is near the lower limit of "shallow" earthquakes. As a consequence, these shocks are sometimes felt over a wide area without doing any serious damage anywhere. An earthquake near Charleston (Missouri), in 1895 occasioned only minor damage near its epicenter and yet was felt from the District of Columbia to New Mexico, from Canada to Louisiana. Contrasting with these shocks, California earthquakes usually originate at a depth of approximately 10 miles; even when they cause considerable damage, they have much smaller areas of perceptibility.

The instrumental data furnish information as to the contemporary seismicity of any given region. However, the historic records where available indicate that in most areas the seismicity changes only relatively little with time; on the other hand, a few regions are known to have shown a much higher seismic activity in earlier periods, and in some instances major earthquakes have occurred in regions which have been considered inactive. Of the roughly one million earthquakes per year which are potentially strong enough to be felt somewhere on earth (magnitude 2 and more) about 2 percent occur in the earthquake belt of California and Nevada (including the shocks off the coast of northern California and Oregon).

It is possible to make certain statistical statements about the frequency and the probability of the occurrence of earthquakes within relatively large areas over long periods of time. For example, of the present average of about 220 great shocks ($M \geqq 7\frac{3}{4}$) and about 1,200 additional major earthquakes ($M = 7.0$–7.7) per century over the whole earth, about 5 and 18 respectively can be expected to occur in the Pacific United States, about 14 great shocks and 65 major shocks in Alaska and the Aleutian Islands, and about 11 major shocks in the remainder of central, eastern, and northern North America. It is not possible, however, to predict the approximate location or time of larger earthquakes, since too little information is available on the sources of energy and the processes involved in the building up of strain leading to an earthquake.

Some information on tectonic processes is being furnished by geodetic measurements. The U. S. Coast and Geodetic Survey has installed a special system of triangulation stations and bench marks in California, which are checked from time to time. In this way, changes in elevation as well as horizontal movements over larger areas are found. Such measurements have indicated, for example, that during the past 60 years the region on the west side of the San Andreas fault between San Francisco and San Jose has moved roughly 10 feet north relative to the east side. This is not a new type of movement; geological evidence indicates that this type of movement has persisted during many centuries, at least. Wherever rivers flow across the fault system the river bed has been displaced in the same direction—the western side northward relative to the eastern side. The total amount of these displacements is not known. In the neighborhood of the San Andreas fault some offsets exceed one mile; however, no information concerning displacements in excess of the distances between successive valleys can be found in this way. Thus far no definite correlation has been found between rocks corresponding to each other on the two sides of the San Andreas fault.

Records of earthquakes have been used to find the direction of the movement at the source and to draw conclusions as to the fault movement during a shock. It is possible to determine whether the first motion of the longitudinal waves is from the source toward the station or in the opposite direction. Thus, for example, earthquakes along the San Andreas fault to the north of Pasadena begin on the Pasadena records with a dilatation toward the source, whereas earthquakes from the San Andreas fault to the east of Pasadena start with a compression toward Pasadena. The motion in the shear waves can be investigated in a similar way. Studies of this type, which have been undertaken in California during recent years, have fully confirmed the persistence of the movements just described. They throw some light on the details of the processes in earthquakes.

It is of interest that similar investigations seem to indicate that in Japan, in the Philippines, and in New Zealand the prevailing movement is such that the continental side is also moving southward relative to the Pacific side. However, the data are insufficient for a more general conclusion as to movements of the continents relative to the Pacific block.

Additional information on processes leading to earthquakes can be expected from investigations on the relation of earthquakes and aftershock sequences to rock creep, which Hugo Benioff is undertaking. On the basis of the elastic rebound theory, the fault rock strain relief which produces an earthquake is proportional to the square root of the energy. Consequently, in a sequence derived from a single fault system the square root of the energy of each shock represents a strain release (or increase) increment, and a plot of the accumulated sum of such increments against time represents the motion of a fault as a function of time. The method thus appears to provide a means for observing tectonic movements in progress. The energy is derived from magnitudes of earthquakes as determined by Gutenberg and Richter. In the case of aftershock sequences, Dr. Benioff has found that creep curves exhibit either simple compressional elastic creep of the fault rock or compressional elastic creep release followed by shearing elastic creep release.

Study of a number of earthquake sequences occurring in all the active regions of the world has revealed that most of them form creep series. Many types of creep are represented, such as constant velocity creep, exponential velocity creep, elastic creep, and elastic flow creep. Individual sequences may have linear extents of 20° to 30° of latitude, as in the case of the South American and Tonga sequence, and the evidence strongly suggests that they are derived from movements of single mechanical units. The deep focus Tonga sequence exhibits elastic creep which continued some 25 years, thus demonstrating that at depths of 650 kilometers rock masses can support elastic creep stresses without appreciable flow for many years. Unfortunately, data available for this type of research cover too short a period to permit conclusions to be drawn as to the exact processes involved. In some of the series investigated by Dr. Benioff, discontinuities in the rate of movement were observed within the short interval of time during which instrumental records are available. Other sequences appear to exhibit no evidence of a discontinuous change in rate since 1904.

Still another type of approach could start with stresses to be expected from theoretically derived forces—such as those connected with contraction or expansion of the earth, or subcrustal currents—and their effect on geological structures. All ideas dealing with forces and structures are by far too controversial to permit the drawing of conclusions concerning expected seismicity, but data obtained from earthquakes and artificial explosions are at present the most reliable basis for hypotheses connected with tectonic and structural problems. The distribution of earthquakes and the other observed phenomena in the Pacific arc structures leave little doubt that there is a great difference between the structure of the Pacific crustal layers and the structure of crustal layers in the continents and in the Atlantic.

A change in surface structure occurs at some point off all Pacific coasts. In the western and southwestern Pacific the boundary is given by the so-called Marshall line or andesite line, which separates the more andesitic material on the continental side from the less andesitic on the Pacific side. This line is known to run to the east of the Japanese, Marianas, and Philippine Islands and crosses the Caroline Islands leaving Yap and Palau on the continental side. It turns sharply to the east near the northwestern end of New Guinea and later passes between Samoa (which is on the Pacific side) and the Tonga Islands (on the continental side). Near Samoa it turns southward and remains to the east of the Kermadec Islands and of New Zealand. Its location in the eastern Pacific is not known, since no islands can be used there for locating the line, but it appears to follow along the coast of North America at a distance which varies from place to place. The andesite line is the intersection of a deepgoing surface of discontinuity with the surface of the earth. The difference in structure on its two sides provides one of the reasons for the accumulation of earthquakes along the line. The fact that in many areas a belt of large negative gravity anomalies parallels the andesite line indicates tectonic processes extending to rather large depth. The deep focus earthquakes are connected with these processes, and the fact that the very deep shocks occur nowhere on earth except near and inland of the andesite line is another indication of the unique structure of the Pacific basin.

All information available for the Atlantic side of North America indicates that the transition from the continent to the bottom of the Atlantic is rather gradual. Although granitic material is probably missing in the deeper parts of the Atlantic basin, as shown by recent seismic explorations by W. Maurice Ewing and his collaborators, *deeper* continental material may be present throughout the bottom of the Atlantic Ocean. In contrast with the Pacific coasts, there are no earthquake belts surrounding the Atlantic or the Indian Ocean. However, earthquakes and volcanoes occur along the mid-Atlantic ridge. In contrast with this ridge and similar ridges with seismic activity in the Indian Ocean, no ridges of the Pacific show any earthquake activity, with the exception of the area near the Hawaiian Islands.

There is no agreement on any hypothesis as to the ultimate source that furnishes the energy for earthquakes. Perhaps they are connected with differences in heat production in the various units of the earth's crust. Laboratory experiments indicate that much more radioactive heat is generated in granitic material than in the more basic material (simatic rocks) of the deeper layers and much less in the ultrabasic material which probably extends relatively close to the surface in the Pacific area. There is, in addition, the effect of the temperature difference between ocean bottom (which is kept at a temperature near 0° C by the deep water in the oceans) and the temperature of roughly 200° C at the corresponding depth under the continents. Subcrustal currents may be a consequence of this horizontal temperature gradient. This may be combined with the fact that the structural arcs along the Pacific boundary are usually interpreted as due to forces either pushing or drawing subcrustal material downward toward the foredeeps, with compensating movements elsewhere. However, we know too little about the details of these processes.

During recent years it has been a very common experience in geophysics that hypotheses concerning the structure and the processes in the earth's crust have become less and less certain as data accumulate; frequently the fact is revealed that the approximations used were not as good as was believed. There is little doubt that the number of recognized unsolved problems is increasing rather than decreasing in practically all fields of geophysics. The hope of explaining and predicting earthquakes seems to be more remote now than at any previous time.

Ray Lyman Wilbur: 1875-1949

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RAY LYMAN WILBUR, chancellor of Stanford University, died on June 26, 1949. A heart infection contracted about five years before had impaired his health and caused him to lead a more cautious life, yet up to within a few days of his death he went regularly to his office in the Hoover Library, maintained an active interest in public affairs and in Stanford University, and kept up with an extensive correspondence.

Dr. Wilbur was of English ancestry. His forebears came to Massachusetts in early colonial days, fought in the Revolutionary War, and were prominent in the life of the Commonwealth. One branch migrated to the Roger Williams Colony in Rhode Island; the next generation moved on to western New York; and the grandfather of Ray Lyman Wilbur went thence to the Western Reserve, which became Ohio. His father was born there, grew up in time to serve in the Civil War, studied law at the University of Michigan, married and, in keeping with the ancestral tradition, moved on westward to Boonesboro (now Boone), Iowa, where Ray Lyman Wilbur was born on April 13, 1875. Eight years later the family went to the new settlement at Jamestown in the Dakota Territory, and thence to Riverside, California, in 1887.

Graduating from the Riverside High School at seventeen, young Ray entered Stanford University with the class of '96, having chosen Stanford "because it had no ivy on its walls." He was tall and lank, six feet four inches in height, and had a marked resemblance in stature and countenance to young Abraham Lincoln. The nationwide depression of the nineties, and a failure in the orange crop in southern California, threw him largely on his own resources to work his way through the university. An early employment as a laboratory assistant in his major department, physiology, helped out.

He soon became acquainted with a young sophomore majoring in geology, named Herbert Hoover, who was also working his way through Stanford and roomed in the same men's dormitory, Encina Hall. Both had native qualities for leadership and teamed up on various student body issues, and thus began a close friendship which went on through life.

Wilbur graduated from Stanford in 1896, was awarded the master's degree the following year in physiology, and received the M.D. degree in 1899 from the Cooper Medical College (later to become the Stanford Medical School) in San Francisco. He was an instructor in physiology at Stanford (1896–'97) and an assistant professor (1900–'03). He engaged in postdoctorate work in medicine in Frankfurt-am-Main (where he worked for a time in Paul Ehrlich's laboratory) and in London (1903–'04), and at the University of Munich in 1909–'10. After a few years of private practice in Palo Alto he became professor of medicine in 1909 and the first dean of the new Stanford University Medical School in 1911—a position which he resigned to accept the presidency of Stanford in 1916.

In the ensuing three decades he rose to eminence in three broad fields of human endeavor: in the medical sciences, in higher education, and in public affairs. In each separate field, the service he rendered was distinguished enough to give him high rank among the men of his time.

In medicine he rose rapidly in influence as a teacher, practitioner, and administrator. His lectures were models of clear, precise statements, backed by a sound knowledge of the fundamental sciences, and sparkling