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EARTHQUAKES AND SEISMIC ENERGY RELEASE IN OREGON

By Richard W. Couch* and Robert P. Lowell*

Introduction

The sector of the circumpacific earthquake belt which extends along the west coast of North America (Barazangi and Dorman, 1969) bifurcates in southern California; the western locus of earthquake activity passes out to sea in the vicinity of Northern California and is situated approximately 200 to 300 kilometers west of the coasts of Oregon and Washington. The eastern locus of activity extends northward through Nevada, Utah, Idaho, Wyoming and Montana. The eastern activity appears either to terminate near the border between the United States and Canada or to rejoin the earthquake activity in the vicinity of Puget Sound and Vancouver Island. Oregon appears as a relatively quiet island in this very active earthquake belt. Consequently, Oregon does not experience the level of activity of its neighboring states; however, Oregon is not aseismic--that is, it is not free from earthquakes.

A meaningful seismic history of Oregon extends back only to the late 1800's and is insufficient in length of observation to establish either the largest size earthquakes to be expected in Oregon or the frequency of occurrence of lesser shocks. The earliest reports extend from 1841 (Berg and Baker, 1963) but are clearly dependent on the size and distribution of the population. The population of Oregon approximately tripled between 1920 and 1970, and during those 50 years, approximately twice as many earthquakes were reported as during the preceding 50 years.

Over the past several years, the general populace has shown an increased awareness of the earthquake hazard and the general problem of earth stability, particularly when related to areas of rapid urbanization, or potential sites of large public construction projects. This awareness has greatly increased the demand for information concerning earthquakes.

It is the purpose of this paper to compile and review the available information concerning earthquake activity in Oregon. Although the available information is severely limited, it is hoped that it will allow those concerned to make more meaningful projections of anticipated earthquake activity in Oregon.

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Seismograph Stations in Oregon

Immediately after World War II, Dr. Harold R. Vinyard of the Physics Department of Oregon State College, Corvallis, Oregon began construction of two Wood-Anderson seismographs. These instruments produced the first seismograms at Oregon State College in 1946. In 1949, Dr. Vinyard in cooperation with Dr. Perry Byerly of the University of California emplaced three Schlicter seismographs in a hillside vault approximately three miles northwest of the OSC campus. In August 1962, the U. S. Coast and Geodetic Survey, on the request of Dr. Joseph W. Berg, Jr., of the Department of Oceanography of Oregon State University installed a World-Wide Standard Seismograph Station at Corvallis. This station consists of three short-period Benioff seismographs, three long-period Sprengnether seismographs, and a short-period vertical Geotech Corporation visual seismograph.

The Blue Mountains Seismological Observatory, a complex of 21 seismometers, is located at a seismically quiet site approximately 38 miles east of Baker, Oregon. The Air Force Cambridge Research Laboratories, operating under the Advanced Research Projects Agency's Vela Uniform Project, dedicated the Blue Mountains Observatory in September 1962. In January 1966, operation of the station was transferred to the U. S. Coast and Geodetic Survey. The station's seismometers consist of 13 short-period Johnson-Mathison instruments arranged as an array, one short-period Electro-Tech instrument, three intermediate-period Geotech instruments, and three long-period instruments. The seismic information is recorded on heat-sensitive paper, film, and magnetic tape and is transmitted to a central file in Rockville, Maryland. The station, under the direction of Mr. Lawrence Jacsha, Chief of the Observatory, Mr. Donald Newsome, Geophysicist, and Mr. James Myer, Electronics Technician of the National Ocean Survey, National Ocean and Atmosphere Administration, operates at a magnification of 750,000 and is Oregon's most sensitive station.

In 1963, Dr. Peter Dehlinger of the Department of Oceanography, Oregon State University, in cooperation with the Oregon Technical Institute, installed a short-period vertical Benioff seismograph at Oregon Technical Institute. Concurrently, Mr. Fred Brecken of the Oregon Museum of Science and Industry, installed a short-period Wilson-Lamison instrument at OMSI. All seismograph records obtained at Corvallis (OSU), Klamath Falls (OTI), and Portland (OMSI) since 1963 are catalogued and stored in the Department of Oceanography, Oregon State University.

During the summer of 1969, The Manned Spacecraft Center of the National Aeronautics and Space Administration under the direction of Dr. Richard Blank of the University of Oregon Center for Volcanology, installed a seismograph station at Pine Mountain, Oregon. This station, located approximately thirty miles southeast of Bend, Oregon is operated jointly by NASA and the University of Oregon. At present, the station's instrumentation consists of four short-period Geotech seismometers and associated 35 mm film recorders.

These five stations, Corvallis (OSU), Portland (OMSI), Klamath Falls (OTI), Blue Mountain (NOS), and Pine Mountain (UO-NASA), comprise the seismographic facilities of Oregon.

Earthquake Location

A shallow earthquake, in general, is the effect of a sudden release of elastic strain energy accumulated within the earth. The point within the earth from which the first energy is released is termed the focus or hypocenter, in reference to

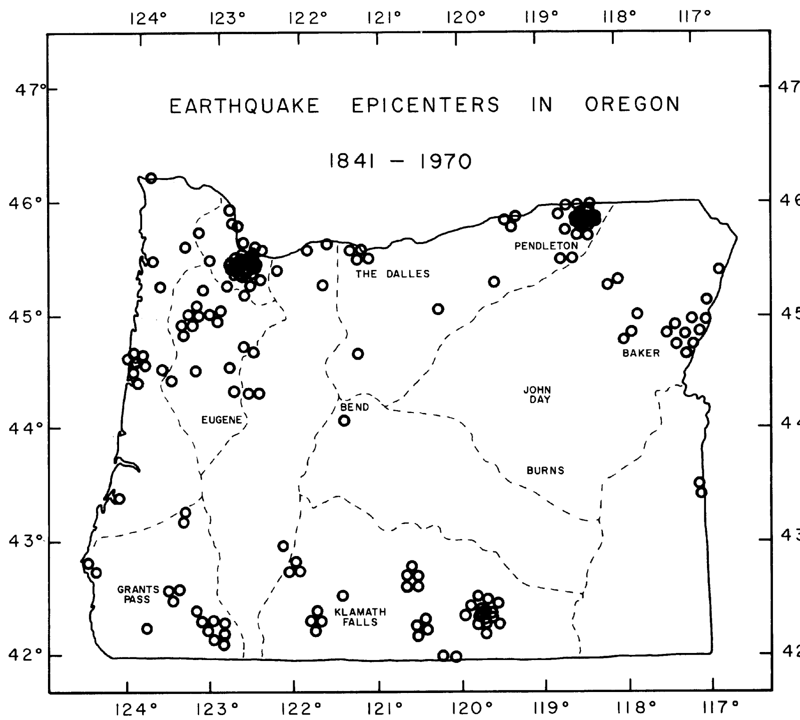

Figure 1. Earthquake epicenters in Oregon: 1841 through 1970. The dashed lines delineate physiographic areas as shown in Figure 3.

the radiated seismic waves the point is often termed the origin or source. The point on the earth's surface vertically above the focus is termed the epicenter.

Figure 1 shows the epicenters of earthquakes which occurred in Oregon from 1841 through 1970. Table 1 lists the date, time, location, intensity and magnitude of 44 earthquakes which occurred in Oregon from 1959 through 1970. Berg and Baker (1963) published a compilation of earthquakes which occurred in Oregon from 1841 through 1958.

In this paper, epicentral uncertainties are given as the radius of a circle of uncertainty whose center is the plotted epicenter and within which the true epicenter is expected to occur. Between 1841 and 1962 most earthquakes which occurred in Oregon were located by the felt effects reported by the people near the epicenter. The estimated uncertainty in epicenters located by felt effects is 8 km for the area west of the Cascade Range and 12 km east of and including the Cascade Range. These estimates are based on differences between epicenters located by instruments and by mapped intensities. The larger uncertainty east of the Cascade Range is due to the deeper focal depths and lower population density of the region.

From the late 1920's until 1962 some of the larger earthquakes in Oregon were located with seismographs of the University of California which at that time

TABLE 1. OREGON EARTHQUAKES: 1959 THROUGH 1970

Year	Month-Day	Origin Time(GMT)**	Location	Intensity +	Magnitude*
1959	Jan 21	07:15	Milton-Freewater	V	
	Nov 09	21:10	Heppner	IV	
1961	Aug 19	04:56:24.1	44°42'N 122°30'W	VI	4.5
	Nov 07	01:29:10	45°40'N 122°52'W	VI	
	Nov 07	21:30	Portland	V	
	Dec 15	11:35	Scappoose	III	
1962	Sep 05	05:37:06	Lebanon	IV	3.5
	Oct 17	08:33	West Linn	II	
	Nov 05	19:36:43.5	45°36.5'N 122°35.9'W	VII	5.0
1963	Mar 02	16:30	Portland	IV	
	Mar 07	22:53:25.0	44°54'N 123°30'W	V	4.6
	Dec 27	02:36:21.6	45°42'N 123°24'W	VI	4.5
1964	Oct 01	12:31:24.6	45°42'N 122°48'W	V	
1968	Jan 27	08:28:23.7	45°36.6'N 122°36.3'W	IV	3.7
	May 13	18:52:17.3	45°35.7'N 122°36.4'W	IV	3.8
	May 27	05:53:34	42°12'N 119°42'W		3.8
	May 28	00:08:48.0	42°15.0'N 119°40.1'W		4.4
	May 28	00:51:03	42°18'N 119°48'W		4.1
	May 28	12:55:42.8	42°15.0'N 119°48.6'W		4.4
	May 30	00:35:58.8	42°19.8'N 119°51.0'W	IV	5.1
	May 31	03:06:38	42°06'N 119°48'W		4.1
	Jun 03	13:27:39.7	42°15.0'N 119°48.0'W	V	5.0
	Jun 04	02:33:00	42°12'N 119°48'W		3.7
	Jun 04	02:34:14.5	42°14.4'N 119°52.2'W	VII	4.7
	Jun 04	02:38:29	42°18'N 119°48'W		4.0
	Jun 04	03:39:50	42°18'N 119°48'W		4.1
	Jun 04	05:52:22	42°18'N 119°48'W		4.0
	Jun 04	06:22:17.0	42°12.0'N 119°49.8'W		4.3
	Jun 04	10:58:22.4	42°15.6'N 119°46.2'W		4.2
	Jun 05	04:51:56.3	42°13.8'N 119°59.4'W		4.7
	Jun 05	05:12:35.4	42°18.0'N 119°46.2'W		4.4
	Jun 05	07:37:45	42°18'N 119°54'W		4.0
	Jun 05	08:04:40	42°18'N 119°48'W		3.3
	Jun 05	08:20:38	42°18'N 119°48'W		4.0
	Jun 05	14:08:40	42°18'N 119°54'W		3.8
	Jun 12	01:20:56	42°06'N 120°00'N		3.4
	Jun 12	01:46:21.9	42°07.8'N 119°47.4'W		4.3
	Jun 21	20:33:27.5	42°12.6'N 119°39.0'W		4.3
	Jun 22	09:39:52.9	42°10.8'N 119°43.2'W		4.3
	Jun 24	11:03:17.3	42°17.4'N 119°50.4'W		4.2
1969	Mar 05	11:43:07.3	45°37.8'N 122°49.0'W	III	3.5
	Aug 14	14:37:39.5	44°59'N 117°45'W	VI	3.6
1970	Feb 12	07:52:25.0	44°38.0'N 122°43.6'W	I	2.5
	Jun 25	07:48:20	W. Portland	IV	3.6

* Unified Magnitude Scale

\+ Modified Mercalli Scale (1956 Edition)

** GMT=Greenwich Mean Time (Pacific Standard + 8 hours)

Data compiled from the following sources:

U. S. Coast and Geodetic Survey (U. S. Earthquakes, 1959-1967 and Preliminary Determination of Epicenters, 1968-1970); Heinrichs and Pietrafesa (1968); Couch, Johnson, and Gallagher (1968); Couch and Johnson (1968); Couch and Whitsett (1969); Unpublished records of the Corvallis Seismograph Station and the Geophysics Group, Department of Oceanography, Oregon State University.

included those at Corvallis. However, because of the epicentral distances, station limitations, and uncertainties in travel-times, the instrumentally located epicenters of that period are probably as inaccurate as those estimated from felt effects. The available instrumental results during that period do suggest, however, that no earthquakes greater than magnitude 5 passed unnoticed in Oregon. After 1963, earthquakes in Oregon were located with seismograph stations located principally in the Pacific Northwest.

The accuracy of epicenters located by triangulation depends on the uncertainties in arrival times of the seismic waves and in the knowledge of the wave velocities between the source and observing station. The larger earthquakes are observed at more stations and show the first waves more distinctly; hence, they are generally located with greater precision. Wave velocities or transit times between the source and observing station are obtained from travel-time curves. Jeffreys and Bullen (1940), and Gutenberg and Richter (Richter, 1958), have published global travel-time curves and Dehlinger and others, (1965) have published local travel-time curves applicable to the Pacific Northwest.

An epicenter in Oregon can be located with equal precision using the global travel-time curves and the local travel-time curves; however, the accuracy will be different. For example, if an earthquake located in the Portland area, using the J - B travel-time curves and arrival times at Blue Mountain, Oregon, Tumwater, Washington, and Corvallis, Oregon, is relocated using the local travel-time curves, the computed epicenter will be repositioned approximately 5 to 6 km toward the west-northwest. This occurs because the local travel-time curves show an approximate 0.3 km/sec difference in wave velocity between eastern and western Oregon whereas, the global travel-time curves indicate a common velocity.

The difference in epicenter locations, computed with the global and local travel-time curves, depends on the locations of the observing stations and on the location of the earthquake. A difference in epicenter location usually implies also a difference in computed focal depth.

The systematic difference in velocities in Oregon as determined from the global and local travel-time curves also suggests the possibility of a systematic difference in located epicenters. Northrup (1970) has reported systematic differences in epicenters of earthquakes located near the Gorda Ridge and Blanco Fracture Zone off southern Oregon. He suggests the differences are caused by a restricted azimuthal distribution of seismograph stations and uncertainties in the assumed wave velocities. This is analogous to the situation within Oregon. The National Ocean Survey uses the Jeffreys-Bullen (1940) travel-time curves to locate earthquakes in Oregon and the Geophysics Group of the Department of Oceanography, Oregon State University, uses the local travel-time curves of Dehlinger and others, (1965).

The seismographic facilities in Oregon available since 1963 and the development of travel-time curves applicable to the Pacific Northwest (Dehlinger and others, 1965) have greatly reduced the uncertainty in the epicenter locations. Figure 2 shows the epicenters of 35 earthquakes which occurred in Oregon from 1963 through 1970. The filled circles indicate the superposition of three or more epicenters.

The uncertainty in epicenter location depends on the number and distribution of the seismograph stations recording the earthquake and the travel-time information used. An average uncertainty of 4 km is estimated for earthquakes larger than magnitude 3.5 located using the local travel-time curves applicable to the Pacific Northwest. This estimate is based on the precision or goodness-of-fit of individual epicentral arcs at well-determined epicenters.

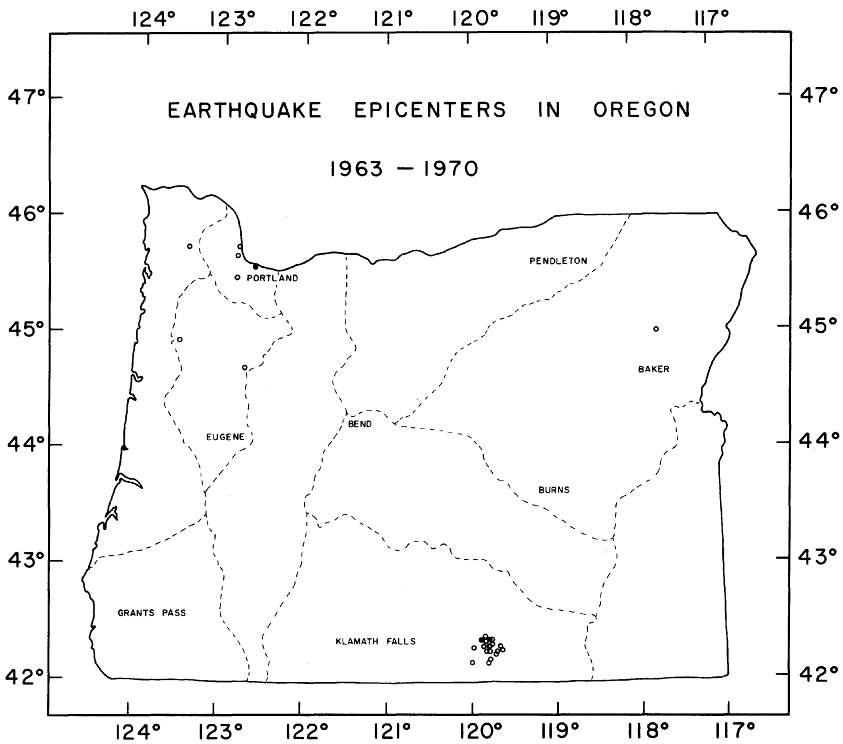

Figure 2. Earthquake epicenters in Oregon: 1963 through 1970. The earthquake data are listed in Table 1. The dashed lines delineate physiographic areas as shown in Figure 3.

The available facilities also suggest it is unlikely that any earthquakes greater than magnitude 3.5 have occurred in Oregon, post-1963, and passed unnoticed.

Focal Depths

The distance between the epicenter and the hypocenter is the depth of focus. Earthquake focal depths are termed "shallow" when they are less than 70 km deep, "intermediate" when they are between 70 km and 300 km deep, and "deep" when they are over 300 km deep. All reported depths of earthquakes in Oregon indicate shallow foci.

Earthquake focal depths in Oregon are currently determined by observing systematic shifts in projected origin times as a function of epicentral distance, by using a velocity profile and arrival times at a seismograph station located near the epicenter, or by the analysis of the arrival times of post-P_n phases. Earthquakes which occur in Oregon tend to be observed at a limited number of seismograph stations; hence, the uncertainty in focal depths determined by the analysis of shifts in the projected origin times can be large. When computed by this method the uncertainty in the depth of focus may be as great as ± 15 km for earthquakes in Oregon. The method of estimating the depth of focus using a velocity profile and the arrival time at a near station is dependent on both the availability and accuracy of the velocity profile and the availability of a seismic station near the epicenter. A ± 5 km uncertainty in focal depth is estimated for this method based on expected variations in crustal velocities. Recently, French (1970) developed a method of determining focal depths from the analysis of post-P_n phase arrivals. This method requires the ability to resolve the phases which closely follow the P_n phase. This is usually not possible with the common photographic seismograph because of slow recording speeds. A ± 3 km uncertainty is estimated for focal depths determined with French's method.

The U. S. Coast and Geodetic Survey (1963-1970), Dehlinger and others, (1963), Heinrichs and Pietrafesa (1968), Couch and others (1968), Couch and Johnson (1968), French (1970) and Couch and Whitsett (1969) reported the focal depths of sixteen Oregon earthquakes. Their reported focal depths indicate an average focal depth between 20 km and 25 km east of the Cascade Range and between 5 km and 15 km west of the Cascade Range.

The work of Dehlinger and others (1968), Thiruvathukal and others (1970) and Dehlinger and others (1971) suggests the earth's crust in Oregon is approximately 20 km thick west of the Cascade Range and between 35 km and 40 km thick east of the Cascade Range. This suggests further that the earthquakes in Oregon are predominantly crustal shocks.

Faults

Crustal earthquakes are generally attributable to the sudden fracturing or faulting of rocks, predominantly in shear. Faults associated with earthquakes do not always intersect the surface and, hence, are not always visible. Maps delineating faults show the visible surface traces of faults but do not indicate whether they are active or inactive. To the authors' knowledge, there are no confirmed observations of movement on a mapped fault in Oregon associated with an earthquake.

Focal mechanism studies by Couch and MacFarlane (1970) and Dehlinger and others (1971) characterize a regional stress field which produces the earthquakes of Oregon. The minimum compressive stress is aligned approximately east-west. The maximum compressive stress varies from an approximate north-south alignment to a vertical alignment. Faults, hence earthquakes, occur when the stress differential exceeds the strength of the crustal rocks. Three fault types are expected under a stress field aligned as above: 1) right-lateral strike-slip, oriented northwest-southeast, 2) left-lateral strike-slip oriented northeast-southwest and 3) normal, oriented north-south. Geologic heterogeneities and old lines of weakness may modify the anticipated fault directions and/or cause mixed faulting to occur. The stress field and consequent faults and associated earthquakes in Oregon suggest a gradual dilation or stretching of Oregon in general east-west direction.

Wells and Peck (1961) and Walker and King (1969) have mapped the surface traces of faults in Oregon; none are confirmed to be active. The orientation and offsets of the faults are, however, consistent with the postulated stress field.

Intensity and Magnitude

Intensity describes the amount of shaking or damage at a specific location; it is highest in the epicentral region and decreases away from the region. Intensities are based on observed or felt effects of the earthquake. Intensities range on the Modified Mercalli Scale (1956 Edition) (Richter, 1958), abbreviated M. M., from intensity I, which is not felt, to intensity XII in which damage is nearly total.

Magnitude is a rating that is essentially independent of the place of observation and that characterizes the amount of energy radiated from the source of an earthquake. Magnitudes are based on instrumental observations and range on a logarithmic scale from less than 1 for small shocks to over 8-3/4 for the largest earthquakes.

Richter (1958) has obtained an empirical relation between earthquake intensities and magnitudes based on observations of crustal shocks in California. An average depth of focus between 15 km and 20 km is estimated for California earthquakes. The average focal depths in Oregon as described above suggest that for a given intensity earthquake the empirical relation will yield a magnitude slightly high west of the Cascade Range where the hypocenters are shallower than in California and slightly low east of the Cascade Range where the hypocenters are deeper.

Table 1 lists the intensities and magnitudes of 44 earthquakes which occurred in Oregon from 1959 through 1970. The listed magnitudes were computed from seismograph measurements, and the intensities were estimated from reported or observed effects.

Observations indicate that intensities in Oregon are slightly greater than expected for a given magnitude earthquake. Enhanced ground motion in Oregon is due to relatively shallow foci, efficient energy transmission in the crustal layers, and the response of the surface layers. In most active areas of Oregon, the surface layers are composed of alluvial or fluviatile deposits. This type of material generally exhibits the greatest movement of any earth material during the passage of seismic waves.

Energy

Intensity, magnitude, and energy are used to characterize the severity of an earthquake. Of these quantities, seismic energy, the wave energy that is radiated from the source, is the most significant, but it is difficult to determine. Empirical equations have been established (Richter, 1958) to approximately relate intensity with magnitude and magnitude with energy.

In the United States, three definitions of magnitude are commonly used: Richter's magnitude (M_L) applicable to local earthquakes, Gutenberg and Richter's magnitude (M) based on teleseisms and applicable to distant earthquakes, and the unified magnitude scale (m) developed by Gutenberg and particularly suited to earthquakes at epicentral distances between 200 and 1000 km (Richter, 1958). Most seismograph stations report unified magnitudes.

The energy, E, of an earthquake may be calculated from Richter's magnitude M_L with the empirical equation $\log_{10}E = 9.9 + 1.9M_L - 0.024M_L^2$ or from the unified magnitude m with the equation, $\log_{10}E = 5.8 + 2.4\,m$. The unit of energy E is ergs. Energy may also be calculated from intensity by obtaining a magnitude M_L for the earthquake from the MM intensity using the empirical relation of Richter (1958).

Only a few magnitudes are available for earthquakes which occurred prior to 1963; hence, most of the energy computations are based on intensities. The intensities and magnitudes of the earthquakes used in the computations are those reported by Berg and Baker (1963) and those listed in Table 1. As noted above, the intensities reported for Oregon earthquakes are slightly greater than expected for a given magnitude earthquake. Consequently, the computations of energy may be as much as several orders of magnitude too high.

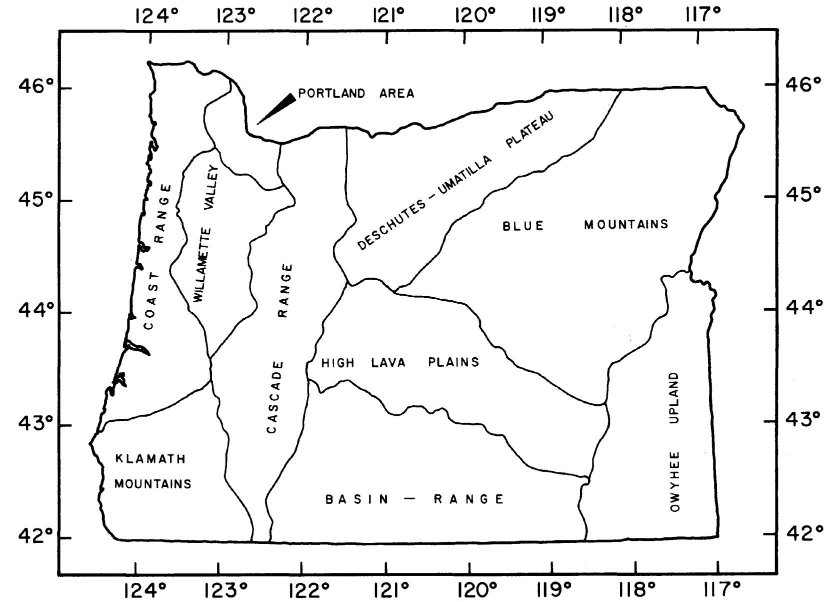

Figure 3. Physiographic divisions of Oregon (after Dicken, 1965).

Earthquakes and seismic energy release are discussed regionally following the physiographic divisions outlined by Dicken (1965) who divided Oregon into the following areas: Coast Range, Willamette Valley, Cascade Range, Klamath Mountains, Deschutes-Umatilla Plateau, Blue Mountains, High Lava Plains, Basin Range, and Owyhee Uplands. Because of the localized seismicity about Portland, the Willamette Valley division has been subdivided into the Willamette Valley area and the Portland area. Figure 3 delineates the nine areas. The low population density of historic Oregon may have caused some earthquakes to be missed or their intensity to be estimated too low. These omissions and inaccuracies would tend to reduce the error in the computed seismic energy release. Although significant uncertainties are associated with the available intensities, magnitudes and computed energies, a characteristic seismic level of each area of Oregon may be estimated and a comparison of areas may be made.

Using the relations between magnitude and intensity and magnitude and energy, a seismic energy release for every earthquake in Oregon from 1841 through 1970 has been computed. Benioff (1951) indicates that the square root of the energy ($\sqrt{E}$) is proportional to the elastic strain rebound, hence, the energies are plotted in Figures 4 through 11 as the cumulative square root of the energy. Each vertical bar represents the square root of the total seismic energy released for a given year; the year is indicated at the bottom of the graph.

Seismic Energy Release in the Portland Area

Of the areas of Oregon, the Portland area has the longest and most complete earthquake history. Figure 4 shows the cumulative seismic energy release in the Portland area for the period 1877 through 1970. The average seismic energy release

PORTLAND

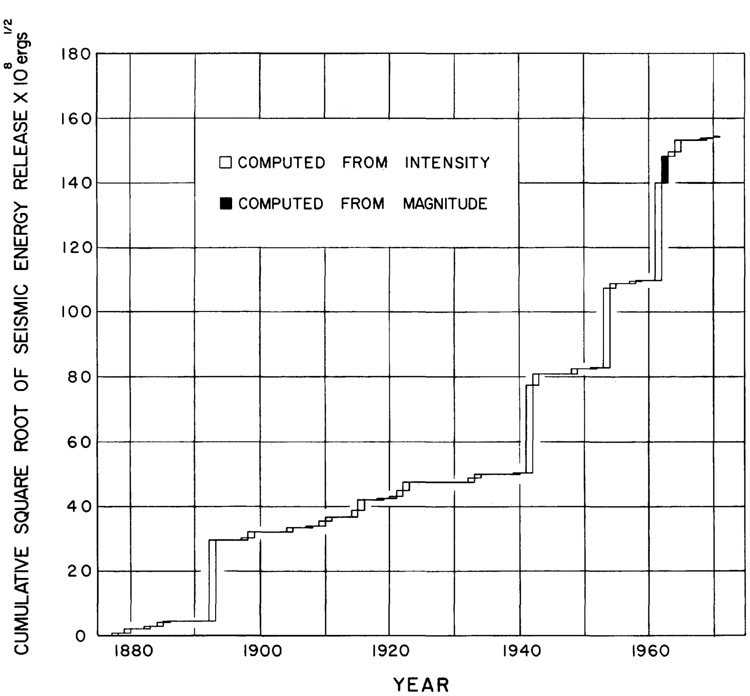

Figure 4. Cumulative seismic energy release in the Portland area.

rate during the 100 year period from 1870 through 1970 was 2.6 x 10^{17} ergs per year. This is approximately equivalent to one magnitude 4.8 (unified magnitude scale) (intensity MM V) earthquake each year. Couch and others (1968) noted that beginning about 1950 the rate of seismic energy release in the Portland area appeared to increase approximately ten times. The higher rate suggests a seismic level equivalent to one magnitude 5.2 earthquake (MM V-VI) approximately each decade. Historical records span too short a time period to indicate whether the change is a singular event or a cyclic change. Figure 4 does suggest, however, that seismic energy release in Portland is a continuing process and that the historical levels are quite likely indicative of future levels.

Figure 1 shows that (at least from the available records) the Portland area experiences more earthquakes than any other area of the state. The November 5, 1962, Portland earthquake was the largest of the recent earthquakes in that area. Dehlinger and others (1963) reported an average magnitude of 5 for this earthquake and indicated the observed maximum intensity in north Portland was VII. The U. S. Coast and Geodetic Survey strong motion seismographs recorded a maximum ground acceleration of 0.16 g (vertical component of 0.076 g and two horizontal components of 0.103 g and 0.096 g) (Dehlinger and others, 1968).

The epicenter was located between Vancouver, Washington, and Portland

COAST RANGE

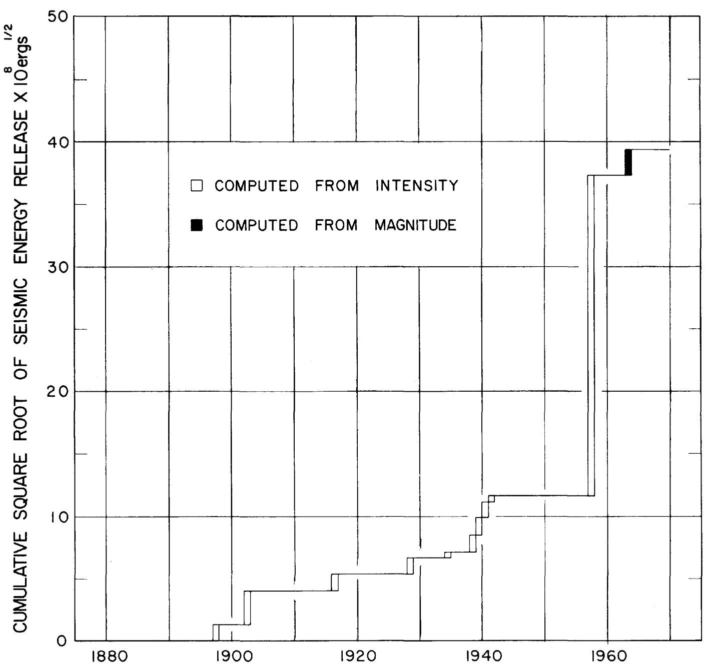

Figure 5. Cumulative seismic energy release in the Coast Range.

in the vicinity of the Columbia River (Dehlinger and others, 1963, Couch and others, 1968). No surface displacements or cracks were reported in the epicentral area (Westphal, 1962, Dehlinger and Berg, 1962, Dehlinger and others 1963). Westphal (1962) recorded 50 aftershocks associated with the Portland earthquake and suggested the seismic activity was related to motion on the Portland Hills fault. The epicentral locations of the principle shock and the subsequent aftershocks neither confirm nor deny this hypothesis. Figure 3 suggests the earthquake activity of the Portland area may occur in a broad fault zone with motion occurring on subsurface faults both in the vicinity of the Tualatin Mountains and the alluvium filled river valleys. Both right lateral motion along northwest-southeast trending faults and left lateral motion along northeast-southwest trending faults have been suggested as the cause of earthquakes in the Portland area (Dehlinger and others, 1963, Couch and others, 1968, Tobin and Sykes, 1968, Gallagher, 1969).

Seismic Energy Release in the Coast Range

Figure 5 shows the cumulative seismic energy release curve for the Coast Range area for the period 1897 through 1970. The average seismic energy release in

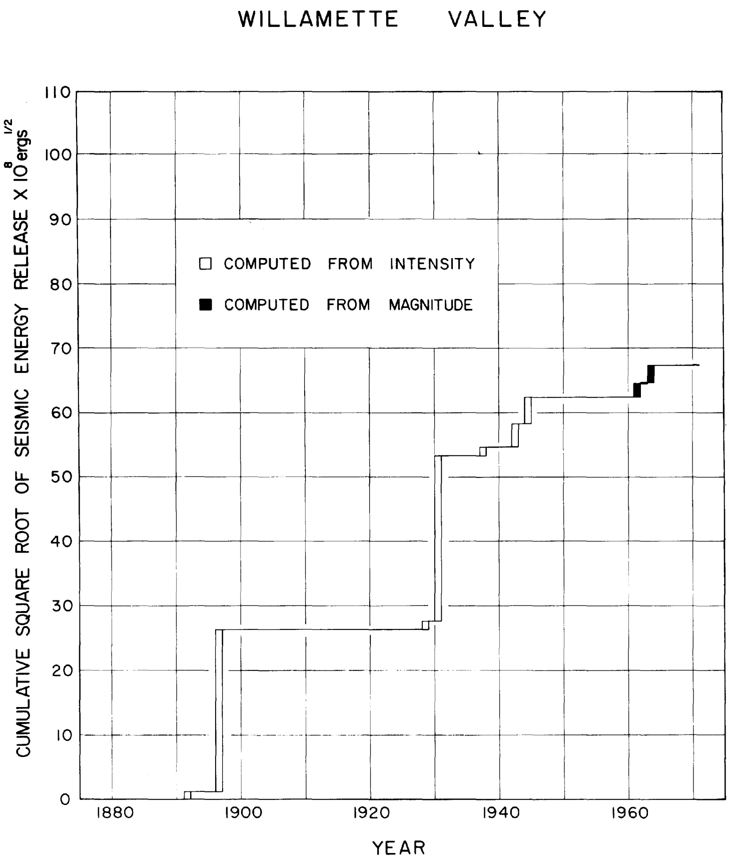

Figure 6. Cumulative seismic energy release in the Willamette Valley.

the Coast Range for the 100 year period (1870 through 1970) is 6.4×10^{16} ergs per year. This level of activity is approximately equivalent to one magnitude 5.0 earthquake (intensity V) each decade. In 1957 and 1963 earthquakes of intensity VI occurred in the region of the Coast Range between Tillamook and Salem and Tillamook and Portland, respectively. These recent earthquakes are the largest documented for the Coast Range area. The largest number of documented earthquakes for the area has occurred between Waldport and Newport and near Newport.

Mapped surface faults in the Coast Range (Wells and Peck, 1961, Walker and King, 1969) trend predominantly northeast-southwest; none are confirmed active faults. Observations at the Corvallis seismograph station indicate continuing minor activity in the area between Drain and Reedsport. It is not known at this time whether the activity is of tectonic origin, associated with downhill land movement, or due to quarrying.

Seismic Energy Release in the Willamette Valley

The earthquake activity in the Willamette Valley is distributed over the area with concentrations of epicenters occurring west of Salem and in the vicinity of the middle Santiam River. Figure 6 shows the cumulative seismic energy release for the Willamette Valley for the period 1891 through 1970. The average seismic energy release for the 100 year period from 1870 through 1970 is 1.3×10^{17} ergs per year. This level of activity is approximately equivalent to one magnitude 5.3 (intensity VI) quake each 30 years.

In 1963 a magnitude 4.6 earthquake occurred northwest of Corvallis, Oregon. Gallagher (1969) obtained a focal mechanism solution for this earthquake which suggests either motion on a northeast-southwest trending strike-slip fault or vertical motion on a northwest trending normal fault. The first solution is consistent with the mapped faults (Wells and Peck, 1961) in the vicinity of the epicenter. Analysis of the records of the Corvallis seismograph station indicates sporadic minor seismic activity occurring within short distances of the station.

Seismic Energy Release in the Klamath Mountains

The earthquake history of the Klamath Mountains extends from 1873 through 1970. Figure 7 shows the cumulative seismic energy release for this period. The average energy release rate for the 100 year period from 1870 through 1970 was 2.8×10^{18} ergs per year. As Figure 7 indicates, the total energy released during the 100 year period is clearly dependent on the intensity VIII earthquake which reportedly occurred near Port Orford in 1873. The intensity and location of this earthquake are questionable; consequently, the computed energy release rate may be much too high. No earthquakes have been reported in the Klamath Mountain area in the past twenty years. The mapped faults in this area trend predominantly northeast-southwest; none are considered tectonically active. Because of the high relief of the area, down-hill ground movement may occur with consequent minor earthquakes.

Seismic Energy Release in the Cascade Range

Figure 8 shows the cumulative seismic energy release in the Cascade Range area for the period 1877 to 1970. The average seismic energy release rate during the 100 year period from 1870 through 1970 was 2.7×10^{18} ergs per year. The computed energy release rate is largely dependent on the occurrence of an intensity VIII earthquake near Cascade Locks in 1877. The intensity and location of this earthquake are questionable; consequently, the computed energy release rate may be much too high. The Cascade Range, seismically, is a relatively quiet area in Oregon.

Decker and Harlow (1970) performed a reconnaissance survey of the microearthquake activity of the volcanic cones of the High Cascades during the summer of 1969. Microearthquake occurrence rates of 2 to 10 events per day were obtained in the vicinity of the larger cones. The surveys were of very short duration and consequently may not be indicative of the average seismic activity associated with the High Cascades. During the fall and winter of 1969 and summer of 1970, Dr. Tosimatu Matumoto, then affiliated with the University of Oregon Center for Volcanology, recorded microearthquakes in the vicinity of Crater Lake. Occurrence rates of 3 to 5 events per day were noted (Dr. Tosimatu Matumoto, personal communication).

KLAMATH MOUNTAINS

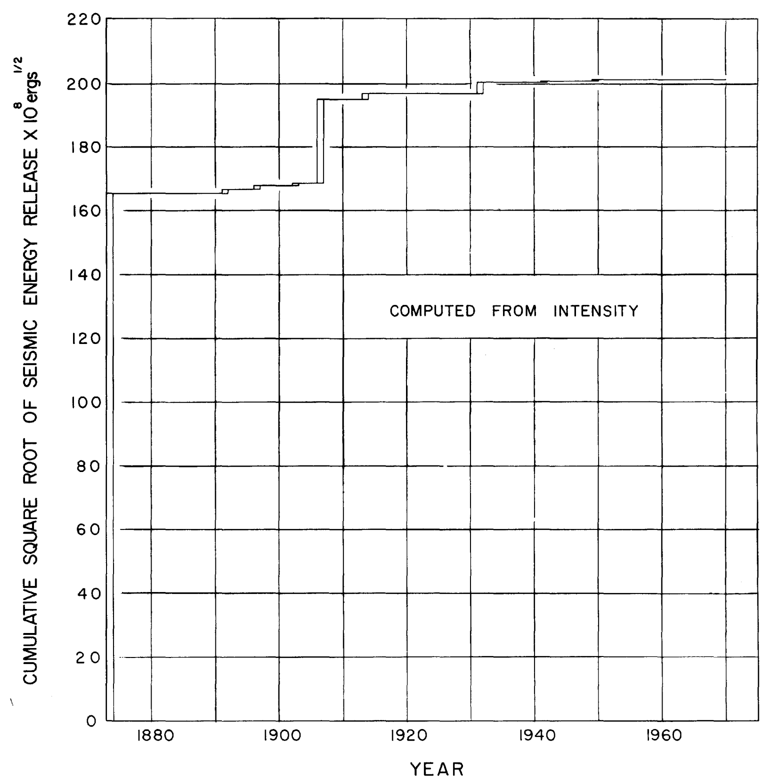

Figure 7. Cumulative seismic energy release in the Klamath Mountains.

Mr. Keith Westhusing, affiliated with the University of Oregon Center for Volcanology, emplaced a seismometer array on Mt. Hood during the summer of 1969 and Mr. Morris Brown under the direction of Dr. Richard Blank of the University of Oregon Center for Volcanology completed a microearthquake reconnaissance survey of the Cascade Range during the summer of 1970 (Dr. Richard Blank, personal communication). The analysis of their observations is continuing.

Seismic Energy Release in the Deschutes-Umatilla Plateau

Figure 9 shows the cumulative seismic energy released in the Deschutes-Umatilla Plateau area from 1892 through 1970. The average seismic energy release rate during the interval 1870 through 1970 was 8.4×10^{17} ergs per year. This is approximately equivalent to one magnitude 5.7 (intensity VI-VII) each 40 years.

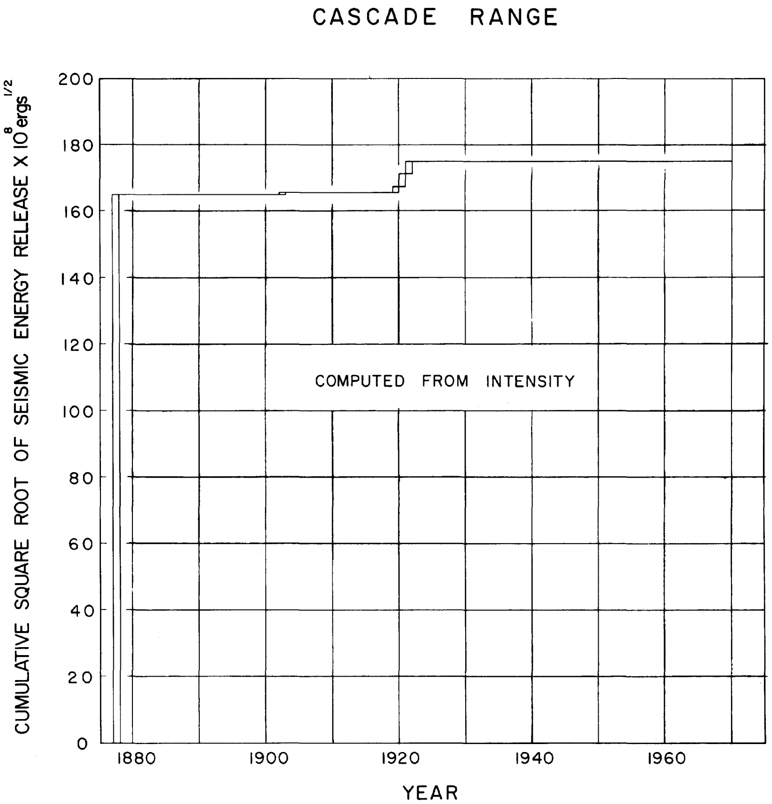

Figure 8. Cumulative seismic energy release in the Cascade Range.

The energy release in this area appears to occur in episodes spaced approximately 45 years apart. In 1936, an earthquake of intensity VII occurred near Milton-Freewater (Berg and Baker, 1963). This earthquake, the largest reported for the area was followed by twelve aftershocks with intensities from II to V. The aftershocks were located in the vicinities of Milton-Freewater, Athena, and Helix. Earthquakes have also been located near The Dalles and Hermiston. No earthquakes have been reported in the Deschutes-Umatilla Plateau area for the past ten years. Faults in this area are poorly mapped and no fault motions determined from earthquake analysis are available.

Seismic Energy Release in the Basin and Range

Figure 10 shows the cumulative seismic energy release in the Basin and Range area for the period 1906 through 1970. The average seismic energy release rate for the 100 year period from 1870 through 1970 was 8.8×10^{16} ergs per year. It is not known whether the absence of recorded earthquakes prior to 1906 is due to no seismic activity or because they were unnoticed. If earthquakes occurred in the area between 1870 and 1906 a slightly higher average seismic energy release rate is

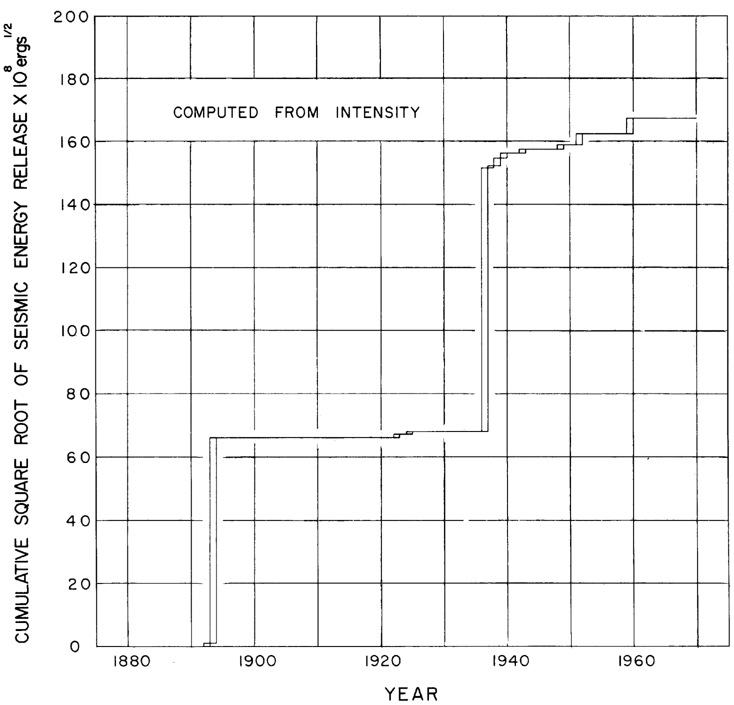

Figure 9. Cumulative seismic energy release in the Deschutes-Umatilla Plateau.

expected. The average seismic energy release as computed is equivalent to one magnitude 5.2 (intensity V-VI) earthquake per 20 years.

In 1968, a series of earthquakes occurred in the Warner Valley (Couch and others, 1968). Twenty-four earthquakes had magnitudes greater than 3.5. The largest magnitude was 5.1. These earthquakes occurred in a section of Oregon which had no previous history of earthquakes.

In the Warner Valley, normal faults which trend north-south are evident as are many northwest-southeast trending faults. Analysis of the first motions on seismograms of the earthquake series suggests fault motions which are consistent with both normal faulting along north-south trending faults and right-lateral strike-slip motion along northwest-southeast trending faults or left-lateral motion along northeast-southwest trending faults.

Seismic Energy Release in the Blue Mountains

Figure 11 shows the cumulative seismic energy release for the Blue Mountain area for the period 1906 through 1970. The average seismic energy release for

BASIN — RANGE PROVINCE

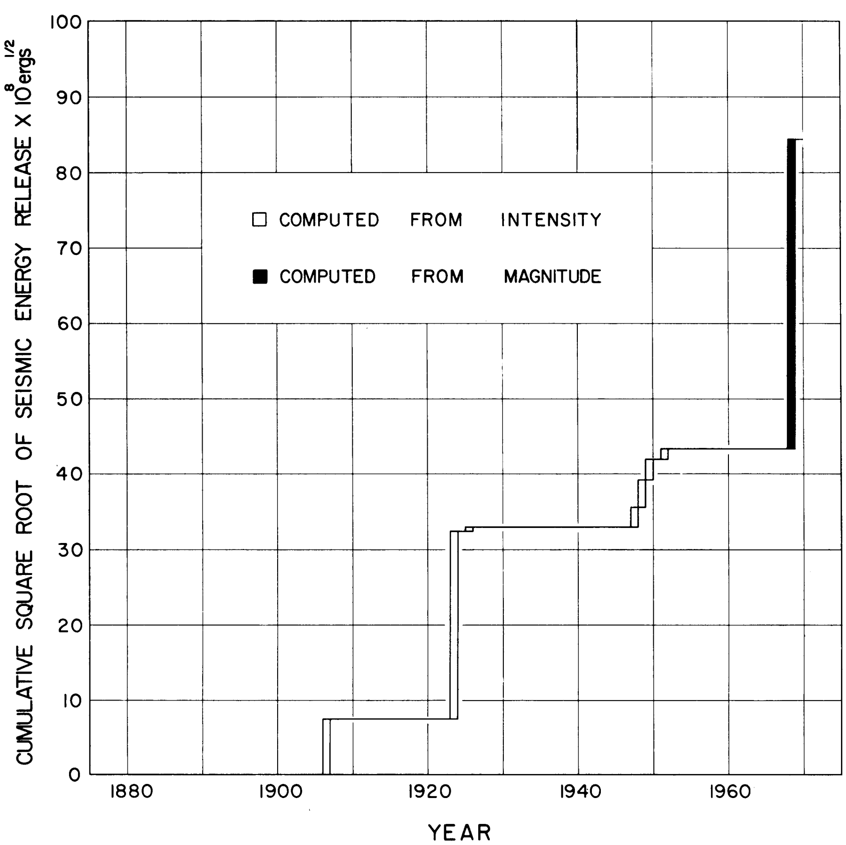

Figure 10. Cumulative seismic energy release in the Basin and Range.

the Blue Mountain area is 6.6×10^{16} ergs per year. This is approximately equivalent to one magnitude 5.1 (intensity V-VI) earthquake per 15 years. It is possible the slightly low energy release rate may be due to absence of reported earthquakes between 1870 and 1906.

The seismic activity of the Blue Mountains area is largely concentrated along the Snake River. A series of four earthquakes of intensity IV to V occurred in 1927 near Richland. Several earthquakes have also occurred in the vicinity of the Powder River north of Baker. Mapped faults in this area (Walker and King, 1969) trend predominantly northwest-southeast. The one earthquake motion study available for this area (Couch and Whitsett, 1969) is consistent with the mapped faults.

Seismic Energy Release in the High Lava Plains and the Owyhee Upland

The historical earthquake activity in the High Lava Plains and the Owyhee Upland areas is too low to compute an average cumulative seismic energy release. One intensity III earthquake occurred near Bend in 1943. Two earthquakes of intensity III

BLUE MOUNTAINS

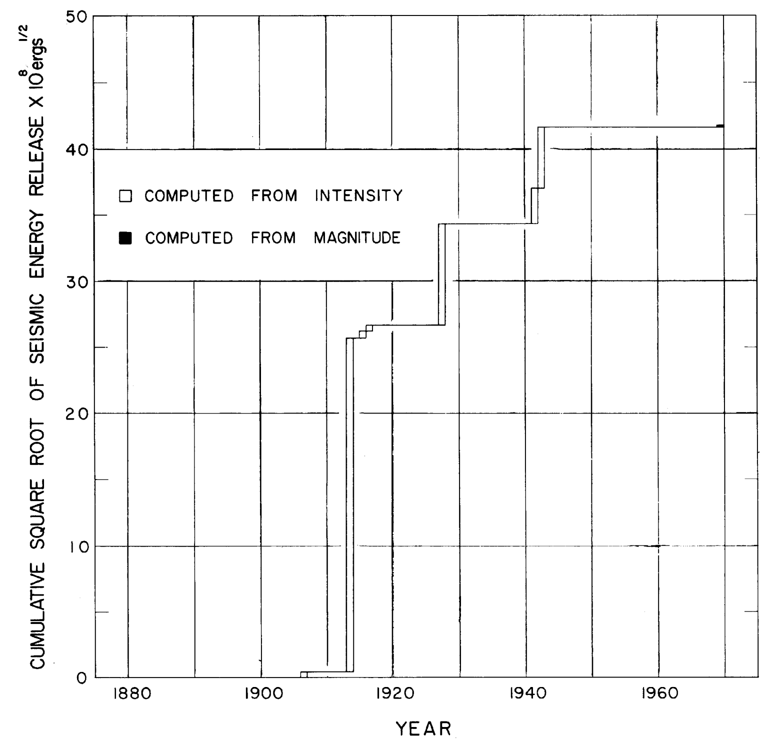

Figure 11. Cumulative seismic energy release in the Blue Mountains.

and IV occurred near Rockville, Oregon in 1943 and 1944 respectively. The Rockville earthquakes may be related to geothermal processes rather than to tectonic movements of the earth's crust (Berg and Baker, 1963). Because of the low population density and absence of a seismograph station in southeastern Oregon, it is possible that additional minor events may occur in these two areas without being detected. Mapped faults (Walker and King, 1969) trend predominantly northwest-southeast in the High Lava Plains and northwest-southeast and northeast-southwest in the Owyhee Upland. None of the mapped faults are known to be active. Decker and Harlow (1970) reported detecting no microearthquakes in the central Oregon lava fields during a 2-day recording period in 1969.

Earthquakes about Oregon

Earthquakes occur in the vicinity of Vancouver, Washington (Rasmussen, 1967, 1969) which are felt in Portland. It is quite likely the earthquakes in Vancouver are occurring along the same fault zone as those in Portland. The magnitude 7.1 earthquake which occurred between Olympia and Tacoma in 1949 exhibited an intensity of VII over the entire Portland area.

TABLE 2. EARTHQUAKE CHARACTERISTICS OF OREGON

Physiographic Area	Maximum Intensity+	Maximum Acceleration (cm/sec^2)	Years of Maximum Intensity	Average E/yr(E=ERG) 1870-1970	Average E/yr/km^2 1870-1970	Estimated Seismic Activity Level
Portland Area	VII	68.1	1962	2.6×10^{17}	8.7×10^{13}	One magnitude 4.8*(intensity V) quake per year; or One magnitude 5.3*(intensity VI) quake per ten years
Coast Range	VI	31.6	1957 1963	6.4×10^{16}	3.4×10^{12}	One magnitude 5.0*(intensity V) quake per ten years
Willamette Valley	VI	31.6	1896 1930 1961	1.3×10^{17}	9.6×10^{12}	One magnitude 5.3*(intensity VI) quake per thirty years
Klamath Mountains	VIII	147.0	1873	2.8×10^{18}	1.8×10^{14}	Insufficient Data
Cascade Range	VIII	147.0	1877	2.7×10^{17}	9.6×10^{12}	Insufficient Data
Deschutes-Umatilla Plateau	VII	68.1	1893 1936	8.4×10^{17}	4.4×10^{13}	One magnitude 5.7*(intensity VI-VII) quake per forty years
Basin and Range Province	VII	68.1	1968	8.8×10^{16}	3.3×10^{12}	One magnitude 5.2*(intensity V-VI) quake per twenty years
Blue Mountains	VI	31.6	1913 1969	6.6×10^{16}	1.1×10^{12}	One magnitude 5.1*(intensity V-VI) quake per fifteen years
High Lava Plains	III	3.2	1943	2.4×10^{13}	1.1×10^{9}	Insufficient Data
Owyhee Upland	IV	6.8	1944	2.0×10^{14}	6.9×10^{9}	Insufficient Data

+ Modified Mercalli Scale (1956 Edition)
* Unified Magnitude Scale

Earthquakes occur in the vicinity of Walla Walla, Washington which affect the area about Milton-Freewater, and earthquakes occur in Idaho which are felt in the vicinity of Richland. Earthquakes also occur offshore, along the very active Blanco Fracture Zone (Bolt and others, 1968, Couch and Pietrafesa, 1968) which are felt by the coastal inhabitants between Gold Beach and Reedsport. It is possible in some areas that earthquakes occurring outside Oregon may show larger intensities in Oregon than earthquakes which occur in Oregon. Well water level changes were noted in eastern Oregon after the great Alaskan earthquake in March 1964; an effect not previously noted with earthquakes occurring in Oregon.

Estimated Seismic Activity in Oregon

Table 2 summarizes the results of the preceeding sections. Column 1 lists the physiographic areas and column 2 lists the maximum intensity reported for an earthquake for the area. Column 3 lists the maximum accelerations expected for the reported intensity. The empirical relation between intensity and acceleration obtained by Richter (1958) was used to compute the acceleration. Accordingly, the acceleration, a, in cm/sec^2, is calculated from $\text{Log}_{10}\, a = \frac{I}{3} - 1/2$ where I is the intensity in units of the Modified Mercalli Scale (1956 edition). Column 4 lists the years of reported occurrence of the maximum intensity. Column 5 lists average energy release per year for each area. The Klamath Mountains and the Cascade Range are indicated as the two highest energy release areas but both are dependent on early questionable intensities. Recent observations suggest that the Portland area, the Deschutes-Umatilla Plateau, and the Basin and Range area are the most active areas within the state. Column 6 lists the average energy release per year divided by the total area of the physiographic area. Column 7 lists an estimated seismic activity level for each area. This is an anticipated typical maximum level of seismic activity based on the magnitude and characteristics of the energy release curve for each area. These results are based only on the limited data available and are accurate only in so far as the past earthquake activity is a good predictor of future activity.

Figure 12 shows the seismic risk map of Oregon adapted from a seismic risk map of the United States (Committee on Seismology, 1969) prepared by the National Ocean Survey (formerly ESSA/USC and GS). The 1969 edition is currently included in the Uniform Building Code (International Conference of Building Officials, 1970). The seismic risk indicated by the map is in agreement with the results summarized in Table 2. The differences between the 1948-52 and 1969 versions of the seismic risk map of Oregon are indicative of the improved seismographic facilities in the Pacific Northwest.

Microseisms

The earth's surface is in constant vibratory motion. These motions, the continuous seismic background noise of the earth, are called microseisms or earth noise. Microseisms of 2 to 10 second period are termed storm microseisms (Iyer, 1964) and are generally attributable to storms over the ocean, ocean waves, and extended fields acting on coastal regions, passage of cold fronts, and other meteorological disturbances. The typical spectrum of microseisms observed at the Corvallis seismograph station during a stormy period show microseism periods between 2 and 10 seconds with a peak near 7 seconds and amplitudes near 12 microns. During quiet periods the predominant period is approximately 6 seconds and amplitudes are less than 1 to 2 microns. The

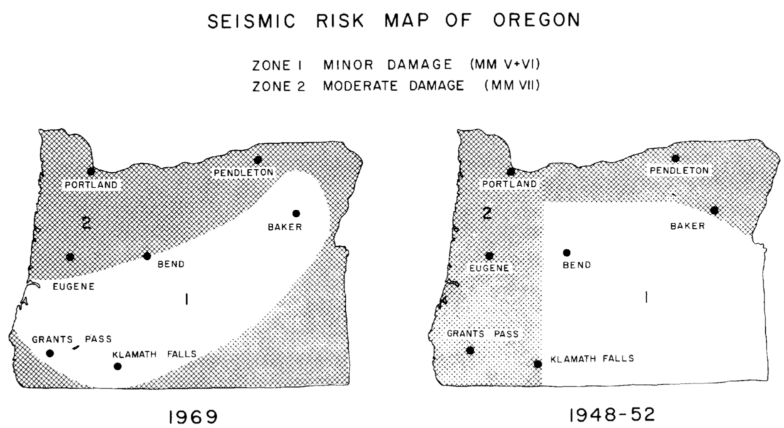

Figure 12. Seismic risk map of Oregon (after Committee on Seismology, 1969).

microseism amplitudes decrease from west to east but are detectable at the Blue Mountain Observatory near Baker, Oregon.

It is possible that microseisms may enhance movement of water-saturated potentially unstable earth particularly in coastal regions where amplitudes are several times larger than those observed in Corvallis, but in general, few effects are attributable to microseisms.

Surface Waves

Large earthquakes with shallow foci produce large surface waves which propagate great distances. Surface waves are of two types: 1) Rayleigh waves which in their passage cause the earth's surface to move in retrograde ellipses and 2) Love waves which cause the earth's surface to oscillate horizontally, normal to the direction of wave propagation. The ground displacements which occur during the passage of surface waves are relatively large but because the periods are long they are seldom noticed.

Surface waves, produced during the Alaska earthquake on March 28, 1964, exhibited amplitudes of approximately 0.5 cm with periods near 20 seconds during their passage across Oregon. The duration of these oscillations was approximately one hour. It is possible for waves of this type to set up a standing wave or seiche on the surface of an enclosed body of water (Richter, 1958) such as lakes, dams, or reservoirs, or on partially closed bodies such as harbors, channels, or estuaries. A seiche in an estuary may also be started by the arrival of a Tsunami.

Tsunamis

Tsunamis or seismic sea waves are usually generated by a vertical displacement of a large area of the sea floor. They are associated with earthquakes and usually occur as a consequence of surface or near surface normal or thrust faulting. Fault motions associated with the earthquakes of California, southwestern Oregon and the

Gorda Basin, and Mendocino Fracture Zone off the coasts of southern Oregon and northern California are predominantly horizontal. It is unlikely that such motions will generate a tsunami. It may be possible that normal faulting along the seismically active Gorda Ridge 200 km west of the southern Oregon coast could generate a minor tsunami but none attributable to this cause has been reported.

Tsunamis generated along other parts of the circumpacific earthquake belt, particularly those originating near Alaska, are of concern along the Oregon Coast. Shatz and others (1964) documented wave heights of 3 to 5 meters above high-tide along the Oregon coast following the great Alaskan earthquake of March 28, 1964. They have indicated that although the rugged open coast rapidly dissipated the waves, the estuaries and their environs were particularly susceptible to damage. Wilson and Tørum (1968) estimate the damage and loss in Oregon, due to the 1964 tsunami, was over $500,000.

Pattullo and others (1968) detected a series of seismic sea-waves at Newport, Oregon, 15 centimeters in height, generated by earthquake activity near Japan. The successful operation of the Seismic Sea-Wave Warning System, at that time, provided approximately 10 hours advance warning of wave arrivals along the Oregon Coast. The Seismic Sea-Wave Warning System provides, via the Oregon State Department of Emergency Services and state, county, and city police radio and teletype facilities, approximately 4 to over 15 hours advance warning of the arrival of a tsunami to residents of coastal Oregon.

Summary and Comments

Seismically active areas exist within Oregon. The seismic history of Oregon is too short to be used as an accurate predictor of earthquake size, number, and distribution. Continual monitoring of earthquakes by seismograph stations during the next several decades should provide a more accurate estimate of the seismicity of Oregon. To provide adequate and accurate coverage additional seismograph stations are needed, particularly in southeastern Oregon.

Rapid urbanization of the Willamette Valley and extensive public works planned for other areas of Oregon make it imperative that the stabilities of the areas be known. Specialized equipment such as portable microearthquake seismometer arrays, which have been developed to assist in local immediate problems, can accurately map active fault zones. In addition, they can locate active zones of continuing deformation which exist without producing the larger release of elastic strain energy of felt earthquakes. Studies of both historical and concurrent seismicity of an area should be of paramount importance when considering potential sites for nuclear generating stations or large construction projects such as dams, airports, large office buildings, and housing developments.

Acknowledgements

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