A look back at Oregon's earthquake history, 1841–1994

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ABSTRACT

More than 6,000 earthquakes, the vast majority smaller than local (Richter) magnitude (M_L) 3, have occurred in Oregon dating back to 1841. About 75 percent of these events have been recorded since March 1993 as part of the 1993–1994 Klamath Falls and 1993 Scotts Mills sequences. The state's largest earthquakes have been the 1873 M_L 6¾ Crescent City, 1936 M_L 6.1 Milton-Freewater, 1962 M_L 5½ Portland (epicenter actually in Vancouver, Washington), 1993 M_L 5.6 Scotts Mills, and 1993 M_L 5.9 and 6.0 Klamath Falls earthquakes. Significant historical seismicity has also occurred near the town of Adel in south-central Oregon in 1968 and in the Deschutes Valley in north-central Oregon in 1976. Persistent areas of seismicity have been the Portland region, probably the state's most active, and the Pine Valley graben-Cuddy Mountain area near the Oregon-Idaho border. Because the historical record for Oregon is so brief, the earthquake potential of the state has not been fully revealed. Recent and future paleoseismic studies have and will likely show that the state was shaken by prehistoric crustal earthquakes up to M_L 7+ in many more areas than was previously believed. The damages from the moderate-size 1993 Klamath Falls and Scotts Mills main shocks indicate that compared to larger but more distant earthquakes occurring within the Cascadia subduction zone, shallow crustal earthquakes may pose the greatest hazard to the Willamette Valley and eastern Oregon.

INTRODUCTION

A look back at 1993 reveals that it was one of the most significant years in terms of seismic energy release in Oregon based on the state's brief historical earthquake record which dates back to 1841. This 1993 energy release was dominated by the very active Klamath Falls main shock-aftershock sequence, which began in late September and was highlighted by events of M_L 5.9 and 6.0 and, to a lesser extent, the March 25 M_L 5.6 Scotts Mills earthquake. A heightened interest in seismic hazards in Oregon in the past decade has led to increased interest and involvement in earthquake research and has resulted in a greater understanding of Oregon's seismic potential, particularly that of the Cascadia subduction zone[1]. In this paper, we review the state's historical seismicity and some recent research on some of its most significant events and discuss their implications for future earthquake occurrence. Because very few earthquakes are known to have occurred in the Cascadia subduction zone[2] beneath westernmost Oregon, our review focuses on seismicity generated by faults located within the North American crust beneath the state.

Three previous significant studies have provided the basis for constructing the historical earthquake record of Oregon: a compilation by Holden (1898) for the Pacific Coast and the period 1769–1897; the well-known Townley and Allen (1939) catalogue, also for the Pacific Coast (1769–1928); and an Oregon catalogue by Berg and Baker (1963) for 1841–1958. Subsequent efforts by the National Earthquake Information Center (NEIC), the University of Washington (UW), Oregon State University (OSU), and Woodward-Clyde Consultants have further refined the state's historical record.

A historical earthquake record usually provides the sole basis for assessing earthquake frequency or recurrence in a region, which is crucial in evaluating seismic hazards on a probabilistic basis. However, a significant aspect of historical records in the western United States, including California, is that they are all very brief in the context of geologic time. The frequency of large earthquakes occurring on a particular fault in the western United States can range from a few hundred to more than 100,000 years; so observing an earthquake from a specific seismic source in a period of 154 years is, more often than not, fortuitous. Hence the historical record seldom totally reveals a region's full seismic potential. This is where the relatively new science of paleoseismology[3] has become so very important in extending the earthquake history of a region back into prehistoric times. Significant advancements in the instrumental detection of "microearthquakes" (events smaller than M_L 3), which are much more frequent than larger events, have also provided important information on previously unknown seismic sources. Such sources would likely have gone undetected without the use of sensitive instruments and expanded seismographic coverage.

PREHISTORIC EARTHQUAKES

In the past decade, geologic evidence has been mounting that great megathrust earthquakes have repeatedly occurred

[1] The region encompassing the boundary between the descending or subducting Juan de Fuca (oceanic) plate (or the Gorda plate at the southern end) and the overlying North American (continental) plate. The zone stretches for a distance of 1,000 km from northwest of Vancouver Island in British Columbia southward to Cape Mendocino in northern California.

[2] Two types of earthquakes have occurred and can occur within the Cascadia subduction zone: (1) great "interplate" events of moment magnitude (M_W) 8 and larger which rupture along the megathrust boundary between the two plates and (2) "intraplate" events which rupture within the subducting plate. An example of the latter includes the 54-km deep 1949 magnitude 7.1 Olympia, Washington, earthquake.

[3] Geologic studies of prehistoric earthquakes through evaluating their surface faulting effects or their impacts on the environment (e.g., liquefaction features, tsunami deposits, buried marshes, etc.).

within the Cascadia subduction zone beneath the coasts of Oregon and Washington. The evidence has been developed from a variety of studies (Rogers and others, 1991; Atwater and others, 1995), but the most convincing has been the discovery of multiple buried soils in coastal intertidal lowlands (e.g., Atwater, 1987, 1992; Darienzo and Peterson, 1990). The presence of these soils suggests that these areas have been subjected to sudden subsidence resulting in submergence of the land surface. The most plausible explanation for these observations is that large megathrust earthquakes have ruptured the Cascadia subduction zone, which resulted in coastal subsidence.

Multiple lines of evidence including buried peats, tsunami sands, and trees killed by salt-water inundation have been observed at several locations along the Oregon, Washington, and northern California coasts. This evidence also indicates the date of the most recent subduction-zone earthquake to be about A.D. 1700 (Atwater and others, 1995). Possibly the most dramatic evidence for this earthquake is the recent discovery by Satake and others (1995) that a 2- to 3-m-high tsunami reached the shores of Japan on January 27–28, 1700. On the basis of the historical accounts of the event and computer modeling, these authors believe the tsunami was the result of an earthquake of about moment magnitude (M_W)[4] 9 that ruptured the Cascadia subduction zone at about 9 p.m. on January 26. The size of the event implies that most, if not all, of the subduction zone was ruptured. Because of the difficulties and uncertainties in dating deposits in the coastal marshes, the ages of earlier events are not as well known. The available data indicate, however, that large earthquakes appear to have struck the Pacific Northwest coast at intervals ranging from a few centuries to about 1,000 years (Atwater and others, 1995).

HISTORICAL RECORD AND EARTHQUAKE DETECTION

The time span covered by the historical earthquake record can be divided into pre-instrumental and instrumental periods. Prior to adequate seismographic coverage, the detection of earthquakes was generally based on human observations and reported effects. This capability is strongly dependent on the geographic distribution and density of population. Both have generally increased with time in Oregon. The Modified Mercalli (MM) intensity scale, described in Table 1, is the best known of several attempts to quantify earthquake effects. Written documentation of pre-instrumental observations, particularly for events from before the turn of the century, is crucial in piecing together the historical record. Oregon, like of much of the western United States, was sparsely populated in the 1800s, and thus the detection of pre-instrumental earthquakes has been of variable completeness as described below.

Comprehensive and recoverable written documentation of historical events did not begin in Oregon until about the 1840s. This was about 35 years after the Lewis and Clark expedition reached the Pacific coast at the mouth of the Columbia River and opened up the Pacific Northwest to set-

Table 1. *Abridged Modified Mercalli intensity scale. Equivalent Rossi-Forel (RF) intensities in parentheses*

I	Not felt except by a few under especially favorable circumstances. (RF I).
II	Felt only by a few persons at rest, especially on upper floors of buildings. Delicately suspended objects may swing. (RF I–II).
III	Felt quite noticeably indoors, especially on upper floors of buildings, but many people do not recognize it as an earthquake. Standing motorcars may rock slightly. Vibration like passing of truck. Duration estimated. (RF III).
IV	Felt indoors by many, outdoors by few during the day. Some awakened at night. Dishes, windows, doors disturbed; walls make creaking sound. Sensation like heavy truck striking building. Standing motorcars rocked noticeably. (RF IV–V).
V	Felt by nearly everyone, many awakened. Some dishes, windows, and other fragile objects broken; plaster cracked in a few places; unstable objects overturned. Disturbances of trees, poles, and other tall objects sometimes noticed. Pendulum clocks may stop. (RF V–VI).
VI	Felt by all, many are frightened and run outdoors. Some heavy furniture moved; a few instances of fallen plaster and damaged chimneys. Damage slight. (RF VI–VII).
VII	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 cars. (RF VIII).
VIII	Damage slight in specially designed structures; considerable in ordinary substantial buildings with partial collapse; great in poorly built structures. Panel walls thrown out of frame structures. Falling of chimneys, factory stacks, columns, monuments, walls. Heavy furniture overturned. Sand and mud ejected in small amounts. Changes in well-water levels. Persons driving cars disturbed. (RF VIII+ to IX).
IX	Damage considerable in specially designed structures, great in substantial buildings; with partial collapse; well-designed frame structures thrown out of plumb. Buildings shifted off foundations. Ground cracked conspicuously. Underground pipes broken. (RF IX+).
X	Some well-built structures destroyed; most masonry and frame structures destroyed with foundations; ground badly cracked. Rails bent. Landslides considerable from river banks and steep slopes. Sand and mud shifted. Water splashed, slopped over banks. (RF X).
XI	Few, if any, [masonry] structures remain standing. Bridges destroyed. Broad fissures in ground. Underground pipelines completely out of service. Earth slumps and land slips in soft ground. Rails bent greatly.
XII	Damage total. Waves seen on ground surface. Lines of sight and level distorted. Objects thrown into the air.

[4] The moment magnitude scale has become the scale of choice among seismologists because it is based on seismic moment and is the best measure of earthquake size. Seismic moment is a function of the area of the fault which ruptures, the average displacement on the fault, and the shear modulus, a parameter which is related to the rigidity of the rocks in the fault zone. The units of seismic moment are dyne-cm (grams-cm^2/sec^2).

tlement from pioneers traveling the Oregon Trail. Oregon's first towns were established principally west of the Cascades, including Salem in 1844, Portland around 1845, Roseburg in 1851, Eugene in 1852, Coos Bay in 1854, and Medford in 1883. In central and eastern Oregon, settlement began later: Pendleton in 1851, The Dalles in 1857, Baker and Klamath Falls in 1866, and Burns and Lakeview in 1884. The publication of newspapers, which are a major source of documentation, began soon after the establishment of these major towns. Based on this population growth and distribution, we estimate that the pre-instrumental historical record for earthquakes of $M_L \geq 5.0$ is complete for western Oregon since about 1850 and for most of eastern Oregon since around 1890.

Although seismograph stations were established in the Pacific Northwest as early as 1906 in Seattle, adequate seismographic coverage of Oregon, at least for smaller events ($M_L \leq 4$ to 5), did not begin until 1979, when the UW expanded its regional network into northwestern Oregon (Figure 1). Few stations operated in Oregon before this time. The seismograph station at OSU in Corvallis (COR) appears to have been the first station installed in the state in 1946.

In 1962, stations at Klamath Falls (KFO) and Pine Mountain (PMT) were also installed by OSU together with the Blue Mountains Seismological Observatory (BMO), originally a 10-station array operated by Teledyne Geotech Corporation and transferred to the U.S. Geological Survey (USGS) in 1966 (Figure 1). The Longmire (LON) station in Washington, installed in 1958, became part of the Worldwide Network of Standard Seismograph (WWNSS) stations in late 1962 as did COR. Both stations have been significant in recording some of Oregon's larger events. We estimate the level of detection from the late 1920s to 1962 was about M_L 4 in western Oregon and M_L 5 in eastern Oregon.

The most significant improvement in instrumental monitoring of Oregon's earthquakes was the installation of stations by UW. By 1980, five UW stations were operating in northernmost Oregon. Today, 44 stations are operating in the state, including those operated by UW, OSU, Boise State University (BSU), and the University of Oregon (UO) (Figure 1). The current detection threshold in northwestern and north-central Oregon is about M_L 1.5 to 2.0 (T. Yelin, USGS, personal communication, 1994). Small-magnitude seismicity in much of eastern Oregon, however, remains relatively unmonitored, with the exception of the northernmost portion of the state adjacent to the Columbia Plateau and the area around Hells Canyon that is being monitored by BSU (Figure 1).

The improvement in earthquake detection is dramatically illustrated in Figure 2. From 1841 to the mid-1960s, fewer than five earthquakes per year were typically recorded in the state, with many years having no known events. As the UW network expanded, the number of recorded and located earthquakes (principally microearthquakes) increased exponentially, culminating in the year 1993 when more than 2,500 events were recorded, most of which were aftershocks of the Klamath Falls and Scotts Mills main shocks. In 1994, more than 2,100 earthquakes were recorded and located as Klamath Falls aftershocks continued to occur but at a steadily decreasing rate.

The historical catalogue for Oregon from 1841 through the end of 1994 contains more than 6,000 earthquakes ranging in

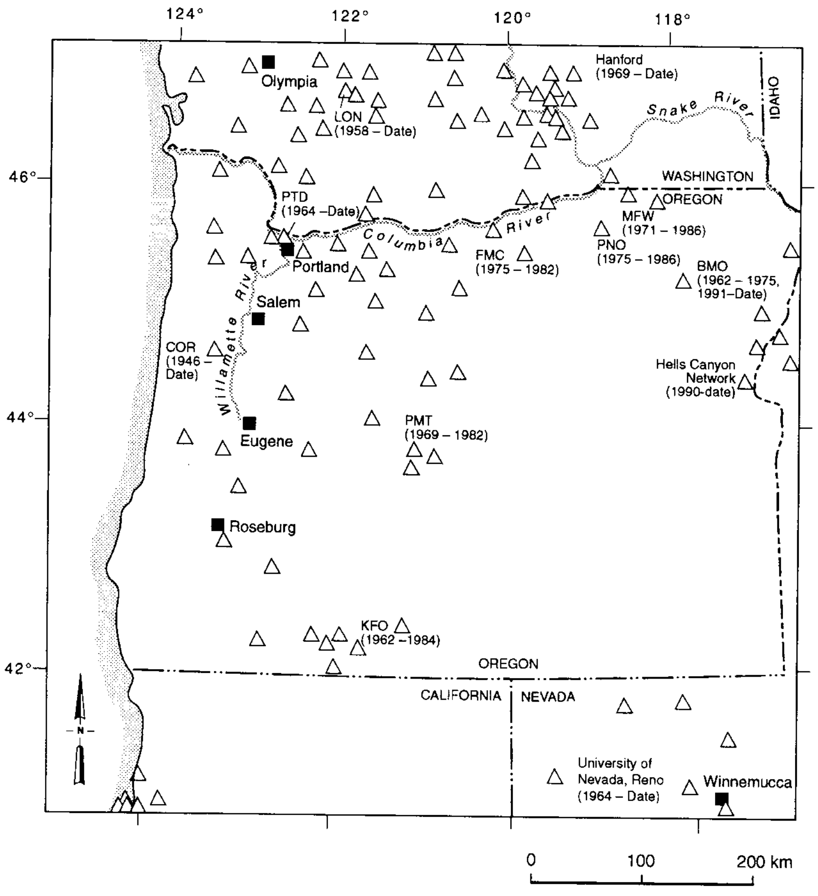

Figure 1. Historical seismographic coverage of Oregon. Stations or networks are represented by open triangles. Abbreviations for these: BMO = Blue Mountains Seismological Observatory; COR = Corvallis; FMC = Fourmile Canyon; KFO = Klamath Falls; LON = Longmire; MFW = Milton-Freewater; PMT = Pine Mountain; PNO = Pendleton; PTD = Portland. Periods of operation are also shown. Unlabeled stations are part of the University of Washington network, which was initially installed in Oregon in 1979.

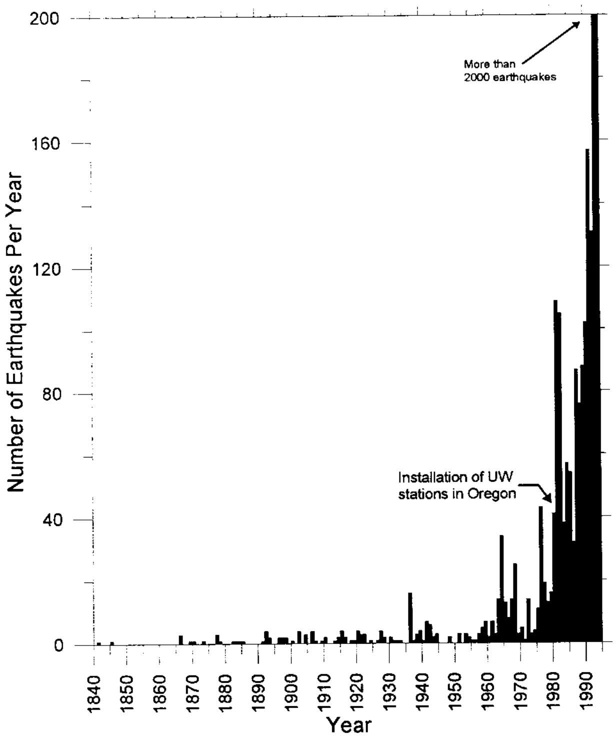

Figure 2. Histogram of recorded and located earthquakes in Oregon, 1841 through 1994. The dramatic increase in seismicity in 1993–1994 was due to the Klamath Falls sequence.

size from less than M_L 1 up to 6¾. Data sources for this catalogue include compilations from Townley and Allen (1939); Berg and Baker (1963); the Decade of North American Geology (Engdahl and Rinehart, 1988); NEIC; UW; Woodward-Clyde Consultants; and BSU (e.g., Zollweg and Wood, 1993).

Within the historical catalogue, a range of magnitudes assigned to a single earthquake is not unusual. For example, the 1962 Portland earthquake has been assigned values of M_L 5.0 to 5.5 and M_W 5.2 (Bott and Wong, 1993). In addition to the existence of several magnitude scales, differences in instrumentation and seismic-wave travel-path effects between events and their recording stations can lead to different magnitude estimates for the same earthquake.

SIGNIFICANT HISTORICAL EARTHQUAKES

The first earthquake in Oregon's historical record was felt with a maximum intensity of MM III and occurred at 4:00 p.m. on December 2, 1841, near Fort Vancouver along the Oregon-Washington border (Berg and Baker, 1963). The first known earthquake in the eastern half of the state reportedly occurred near Umatilla on March 6, 1893 (Figure 3a). This event, described as a "succession of shocks," knocked down one wall of a large stone building (MM VI or VII?) (Townley and Allen, 1939). The 1893 earthquake may have been one of Oregon's largest events, according to its reported maximum intensity. However, very little is really known about this earthquake. Reports of the event outside Umatilla are apparently unknown, which suggests that it was only locally felt and thus not that large.

Since 1841, five earthquakes larger than M_L 5.5 are known to have occurred in Oregon (Table 2). There have been an additional six events of about M_L 5 to 5.5 in size (excluding the 1962 Portland earthquake; see following discussion). Three earthquakes of approximate M_L 5, whose source was near Portland (1877, 1892, and 1961), have been recently described by Bott and Wong (1993). Twenty-eight events have occurred within the state in the approximate range of M_L 4.5 to 5.0 (MM V or VI, if no magnitude assigned) (Table 2). The following describes the most significant releases of seismic moment in Oregon during historical times.

1873 Crescent City earthquake

On November 23, 1873, at about 9:00 p.m., an earthquake of estimated M_L 6¾ occurred near the Oregon-California border east-southeast of Brookings (Toppozada and others, 1981) (Figure 3a). The maximum reported intensity of the event was MM VIII in the Smith River Valley north of Crescent City, California (Figure 4). Chimneys were knocked down in Crescent City, Port Orford, Grants Pass, and Jacksonville. Ground cracking was observed east of Crescent City. The earthquake was felt as far north as Portland (MM III–IV) and as far south as San Francisco (Townley and Allen, 1939). Because the location of the 1873 earthquake can only be estimated from the center of the isoseismal contours (Figure 4), its uncertainty is large, and the event could have occurred in northernmost California or southernmost Oregon. The lack of aftershocks led Ludwin and others (1991) to suggest that the earthquake may have occurred within the subducting Gorda (or Juan de Fuca) plate of the Cascadia subduction zone. Such intraplate earthquakes are rare in Oregon. Alternatively, the event may have been crustal in origin and occurred far enough offshore such that no aftershocks were felt (Ludwin and others, 1991).

1936 Milton-Freewater earthquake

The largest and most significant earthquake in northeastern Oregon, known as the Milton-Freewater or State-line earthquake, occurred at 11:08 p.m. on the night of July 15, 1936 (Neumann, 1938) (Figure 3a). The maximum intensity was MM VII+, and it was felt over an area of 275,000 km^2 (Figure 5). In a reevaluation of the event, Woodward-Clyde Consultants (1980) (also Foxall and Turcotte, 1979) calculated a magnitude of M_L 6.1, as recorded at 17 seismographic stations. Based on the isoseismal map (Figure 5) and an empirical relationship between magnitude and total felt area developed by Toppozada (1975) (see Bott and Wong [1993] for further discussion), the event was estimated to be a M_L 6.4. The main shock was preceded by

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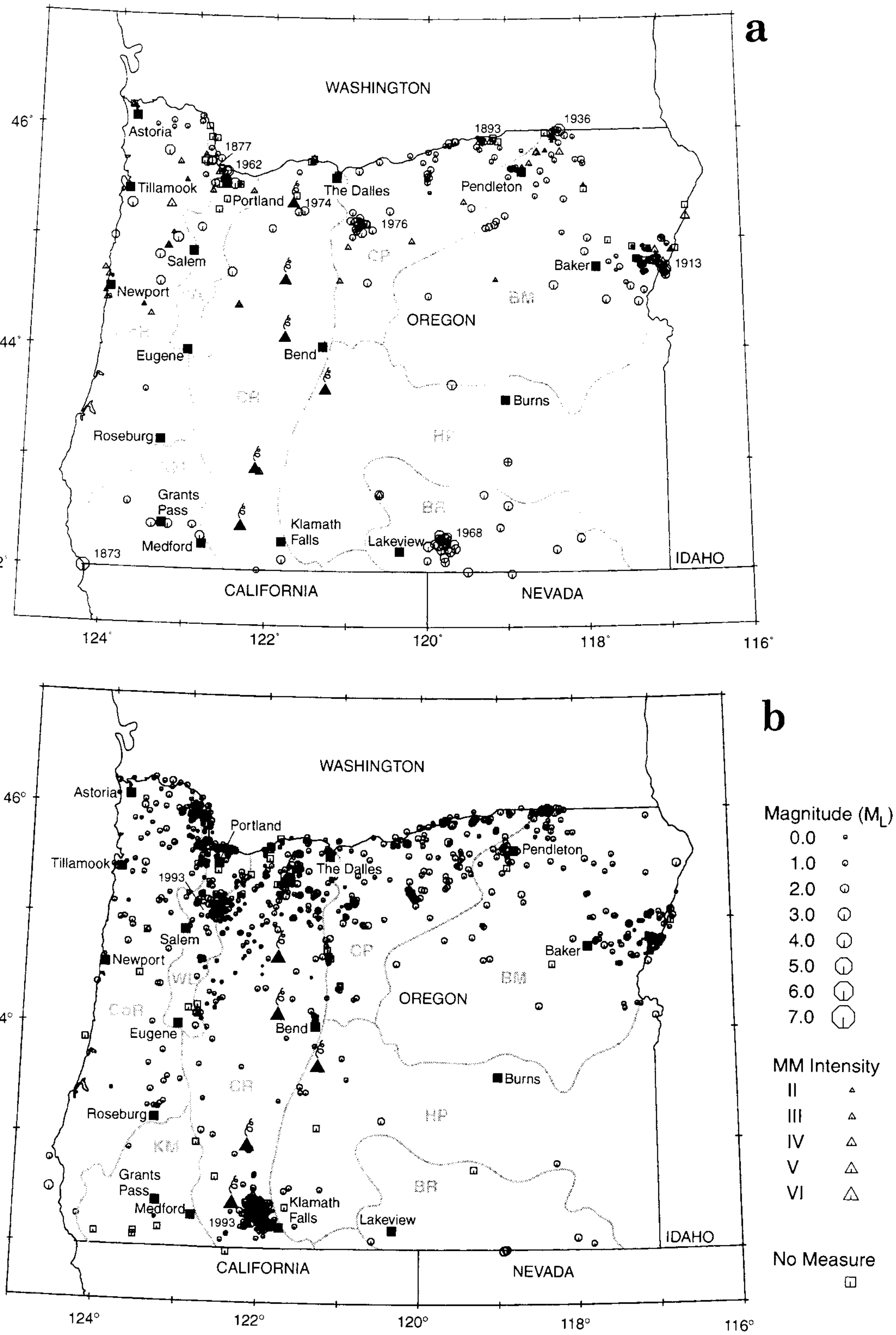

Figure 3. Historical seismicity and geologic provinces of Oregon: (a) 1841 through 1978 and (b) 1979 through 1994. Major Cascade volcanoes are shown as erupting triangles. Abbreviations: CoR = Coast Range; WL = Willamette Lowlands; KM = Klamath Mountains; CR = Cascade Range; CP = Columbia Plateau; BM = Blue Mountains; HP = High Lava Plains; BR = Basin and Range.

Table 2. *Significant historical earthquakes in Oregon, magnitude 4.5 or MM intensity V and greater*

Date	Time (GMT)	Magnitude	Maximum MM intensity	Location	Comments
Nov. 23, 1873	05:00	M_L 6¾	VIII	Crescent City	Largest historic event
Oct. 12, 1877	17:00	M_L 5¼	VII	Portland	Portland's second largest event
Feb. 4, 1892	04:30	M_L 5	VI	Portland	"Severe shock"
Mar. 5, 1893	?	—	VI or VII	Umatilla	
Apr. 2, 1896	11:17	M_L 4	VI	McMinnville	
Apr. 19, 1906	09:30	—	V	N of Lakeview	Three felt aftershocks
Oct. 14, 1913	23:00	—	VI	Hells Canyon	
May. 18, 1915	03:00	—	V	Portland	One of three shocks
Apr. 14, 1920	23:45	—	V	Crater Lake	One of three shocks
Feb. 25, 1921	20:00	—	V	E of Sweetwater	
Jan. 11, 1923	04:29	—	VI	Lakeview	
Jan. 6, 1924	23:10	—	V	Milton-Freewater	
Apr. 9, 1927	05:00	—	V	Pine Valley-Cuddy Mountain	
Jul. 19, 1930	02:38	M_L 4	V-VI	20 km NW of Salem	Cracked plaster
Jul. 16, 1936	07:07	M_L 6.1	VII+	Milton-Freewater	Eastern Oregon's largest event
Jul. 18, 1936	16:30	—	V	Milton-Freewater	Aftershock
Aug. 4, 1936	09:19	—	V	Milton-Freewater	Aftershock
Aug. 28, 1936	04:39	—	V	Milton-Freewater	Aftershock
Dec. 29, 1941	18:37	M_L 4½	VI	Portland	Minor damage
Jun. 12, 1942	09:30	—	V	Pine Valley-Cuddy Mountain	Minor damage
Nov. 1, 1942	17:00	—	V	Portland	
Jan. 7, 1951	22:45	—	V	Hermiston	
Dec. 16, 1953	04:32	M_L 4½	VI	Portland	Minor damage in Portland
Nov. 17, 1957	06:00	M_L 4½	VI	S of Tillamook	Felt strongest near Salem
Mar. 12, 1958	12:09	M_L 4.5	—	SE of Adel	
Jun. 2, 1959	18:49	M_L 4.7	—	NW of Burns	
Aug. 19, 1961	04:56	M_L 4½	VI	SE of Salem	Minor damage in Albany
Nov. 7, 1961	01:29	M_L 5	VI	NW of Portland	Minor damage in Portland
Nov. 6, 1962	03:36	M_W 5.2, M_L 5½	VII	Vancouver-Portland	Damage in Portland
Mar. 7, 1963	23:53	Body wave (m_b) 4.6	V	West of Salem	Minor damage in Salem
Dec. 27, 1963	02:36	M_L 4½	VI	Vernonia NW of Portland	Minor damage near epicenter
May. 30, 1968	00:35	M_L 5.1	V	Adel	Swarm
Jun. 3, 1968	13:27	M_L 5.0	V	Adel	Damage
Jun. 4, 1968	02:34	M_L 4.7	VI	Adel	Swarm
Jun. 5, 1968	04:51	m_b 4.7	—	Adel	Swarm
Apr. 13, 1976	00:47	M_L 4.8	V-VI	Deschutes Valley	Minor damage
Mar. 25, 1993	13:34	M_L 5.6	VII	Scotts Mills	$28 million in damage
Sep. 21, 1993	03:28	M_L 5.9	VII	Klamath Falls	Two deaths
Sep. 21, 1993	05:45	M_L 6.0	VII-VIII	Klamath Falls	$7.5 million in damage
Dec. 4, 1993	22:15	M_L 5.1	VII	Klamath Falls	Aftershock

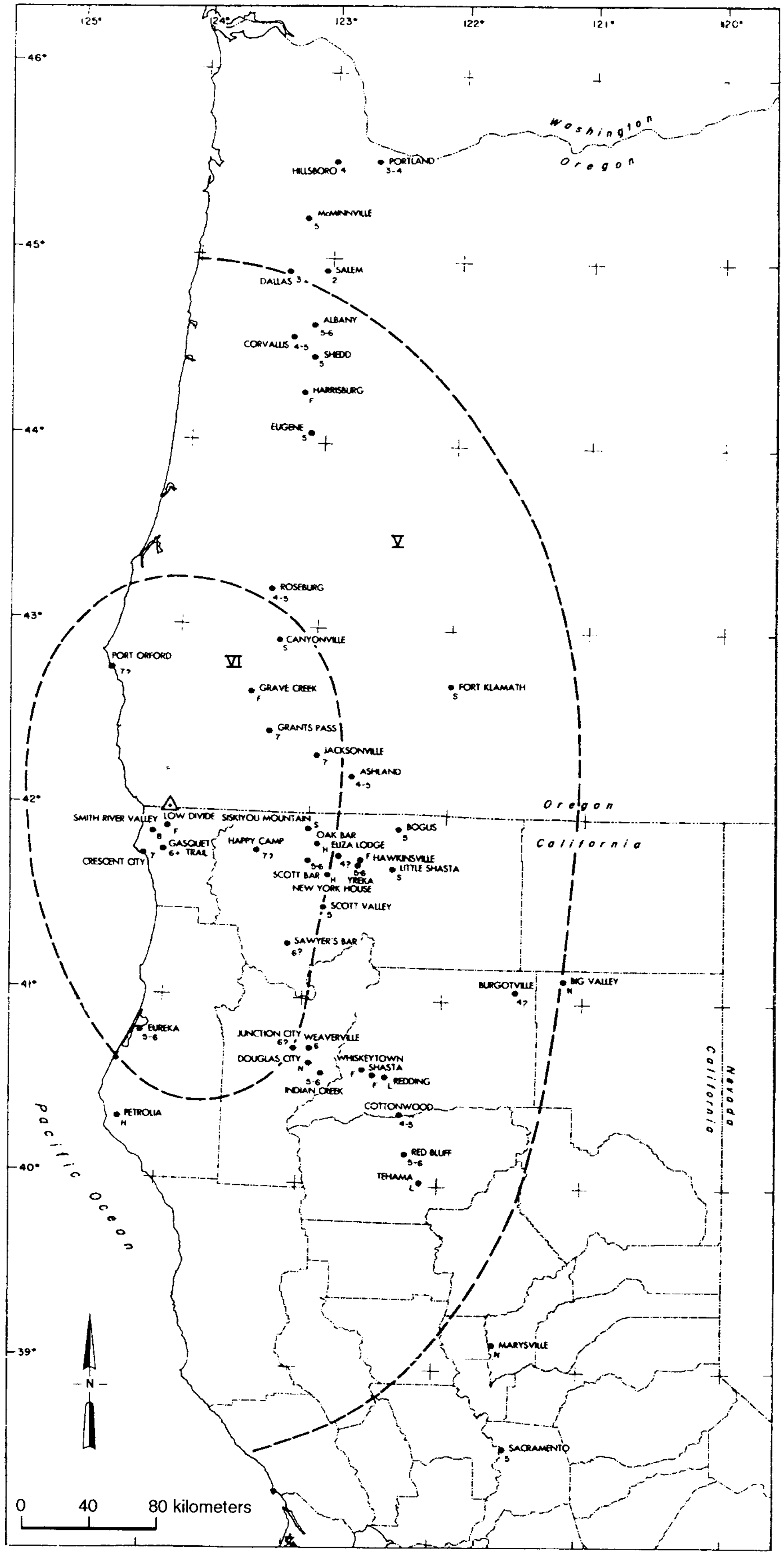

Figure 4. Isoseismal map of the 1873 Crescent City earthquake. Smoothed isoseismal contours, centered around estimated epicenter (triangle), define zones with reported MM intensities V and VI (see Table 1). Individual sites reporting effects either show Arabic numerals for (equivalent Roman numeral) intensities or letters for indeterminate intensities as follows: N = not felt, F = felt, L = light, H = heavy, S = severe. From Toppozada and others (1981).

two felt foreshocks at 10:30 p.m. and 11:20 p.m. and was followed by numerous aftershocks (Neumann, 1938).

The main shock was felt most strongly and caused damage in and around Milton-Freewater, Umapine, and Stateline, Oregon (Figure 5). It was also strongly felt in Walla Walla, Washington, just north of the border. Total damage amounted to $100,000 in 1936 dollars. Many chimneys were damaged, houses were moved off their foundations, canned goods were scattered in a cannery, plaster cracked, windows broke, and school buildings were damaged (Neumann, 1938).

Intense ground cracking occurred in a zone 25 m wide and 500 m long extending west-northwest along the base of a hill west of Milton-Freewater. Some cracks were 1–2 m wide, and in one place the ground dropped by 2.4 m. Water emerged from some of these cracks, indicating that liquefaction as well as ground slumping and landsliding had occurred. Ground-water flow generally increased in wells, and several springs were revived (Brown, 1937).

The epicentral location of this earthquake has been difficult to determine. An epicenter based on the isoseismal data gave a location about 10 km northeast of Milton-Freewater (Neumann, 1938). The International Seismological Centre and the U.S. Coast and Geodetic Survey calculated instrumental epicenters in southeastern Washington north-northeast of Walla Walla. Woodward-Clyde Consultants (1980) relocated the event after rereading arrival times and determined a similarly placed epicenter. They suggested that the 1936 earthquake may have occurred on the Hite fault. In contrast, Mann and Meyer (1993) suggested the source of the 1936 earthquake was the Wallula fault zone near the zone of ground cracking just south of Umapine and west of Milton-Freewater, based on their reassessment of the maximum reported intensities.

We believe, however, the large epicentral uncertainties of both the felt and instrumental locations of the 1936 earthquake make any interpretations of its source tenuous. Earthquake locations based on maximum intensities can be erroneous by tens of kilometers because of site or seismic-wave propagation effects on ground shaking (e.g., Wong and Savage, 1978). Errors are also often large (also tens of kilometers) when locating pre-1960 instrumentally-recorded earthquakes, because clock errors were common, and recording stations few and distant. For example, the closest station recording the 1936 event was in Spokane, about 250 km away, and clock error appears to have been large. The majority of the stations recording the earthquake were in California, which resulted in possible azimuthal bias in the seismographic coverage.

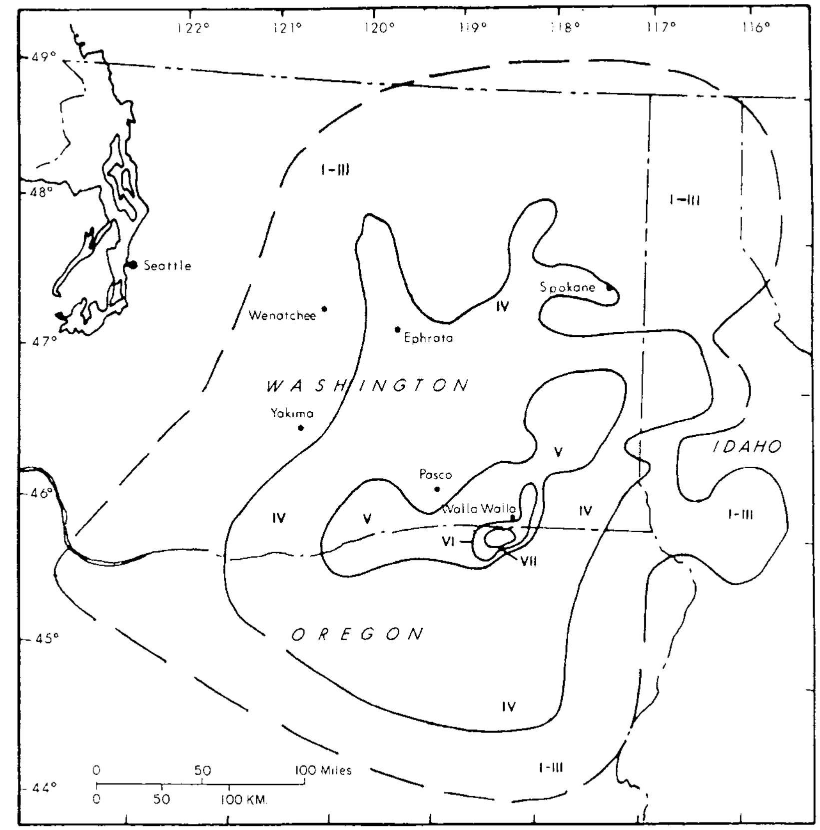

Figure 5. Isoseismal map of the 1936 Milton-Freewater earthquake. From Washington Public Power Supply System (1974).

1962 Portland earthquake

One of the best-known earthquakes in Oregon history occurred at 7:36 p.m. on November 5, 1962, near Portland (M_L 5.2 to 5.5) with a maximum intensity of MM VII (Coffman and others, 1982; Yelin and Patton, 1991; Bott and Wong, 1993) (Figures 3a and 6). The earthquake was felt over a large area of 70,000 km^2 in northwestern Oregon and southwestern Washington (Figure 6). In Portland, many chimneys cracked or fell down, windows broke, tile ceilings fell, and plaster cracked (Dehlinger and Berg, 1962). In Vancouver and Battleground, Washington, furnishings and small objects shifted. Rumbling sounds were heard just before the earthquake was felt, and the shaking lasted from a few to 30 seconds (Dehlinger and Berg, 1962). Numerous aftershocks occurred, but none were large enough to be felt in Portland.

Yelin and Patton (1991) recomputed the location of the 1962 earthquake and placed it 15 km northeast of downtown Portland and at a depth of 16 km. This epicenter is 7 to 8 km northeast of the original location of Dehlinger and others (1963). This location makes this earthquake a Washington event, although it clearly warrants inclusion in any discussion of Oregon seismicity. Yelin and Patton (1991) calculated several possible earth-

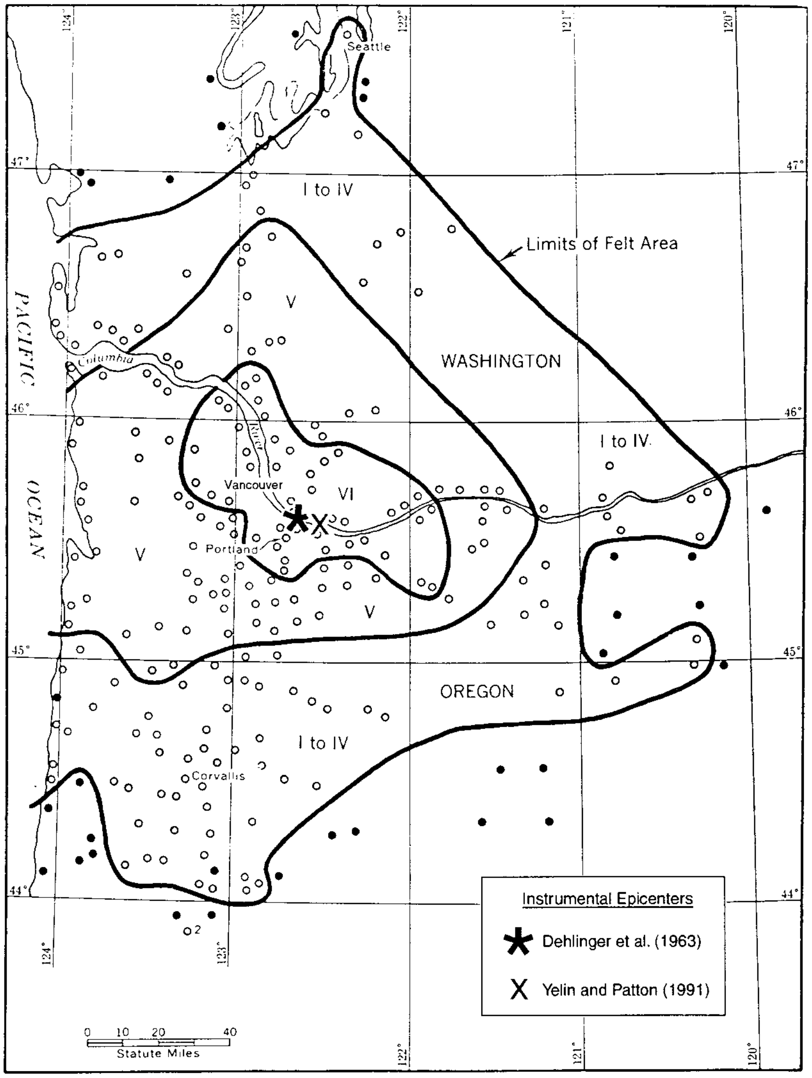

Figure 6. Isoseismal map of the 1962 Portland earthquake. Modified after Lander and Cloud (1964)

quake focal mechanisms[5] for the event but favor the mechanism that exhibits normal faulting on a northeast-trending fault plane. This is consistent with the model of a Portland pull-apart basin (Beeson and others, 1985), although the stress directions exhibited by the focal mechanism are not consistent with the regional stress pattern of approximately north-south maximum compression. Alternatively, Unruh and others (1994) have suggested that the Portland area lies within the Portland fold belt, a region characterized by northwest-trending anticlines[6], possibly indicative of fault motion at depth. We favor the reverse faulting focal mechanism of Yelin and Patton (1991), because it is more consistent with the contemporary stress field and other focal mechanisms in the region.

1968 Adel earthquake swarm

An intense sequence of earthquakes began on May 26, 1968, in the vicinity of the town of Adel near the Oregon-Nevada border (Coffman and Cloud, 1970) (Figure 3a). No historical events are known to have occurred in the epicentral area before this sequence. The activity reflected typical characteristics of an earthquake swarm: a sequence of events concentrated in space and time with no single dominant event. The largest events of the swarm occurred within the first 10 days of activity; the largest earthquake (M_L 5.1) occurred at 4:36 p.m. on May 29 (Schaff, 1976). A M_L 5.0 event occurred at 5:27 a.m. on June 3 and was felt with a MM V intensity in Lakeview, Oregon, and Fort Bidwell, California. The strongest felt earthquake (M_L 4.7) reported at Adel (Coffman and Cloud, 1970) occurred at 6:34 p.m. also on June 3 with a maximum intensity of MM VI. It was felt over an area of 18,400 km^2, causing damage to many chimneys in Adel and producing ground fissures near Fort Bidwell (Couch and Johnson, 1968). Grocery items were thrown to the floor in a local store, and a rock wall of a storage building collapsed. In the library, books on the west wall were thrown to the floor. Geologic effects included rockfalls from the western wall of Warner Valley and cracks in State Highway 140, about 2 km west of Adel (Coffman and Cloud, 1970). Increased flow at a hot spring was also reported (Couch and Johnson, 1968). The swarm continued through June and July, decaying exponentially in intensity and occurrence with time.

Portable seismographs were deployed in the vicinity of Adel from June 6 to July 25 by the University of Nevada at Reno, to monitor the swarm activity. Analysis of 169 aftershocks (Schaff, 1976) indicates a 15-km-long, 6-km-wide, north-trending zone located northwest of Adel between the depths of 3 and 12.5 km (Schaff, 1976). A focal mechanism computed by Patton (1985) indicates that the largest earthquake was the result of normal faulting on an approximately north-striking plane. That plane is consistent with the trend of aftershocks and the structural grain of the northern Basin and Range province. Thus an unmapped fault near Adel appears to have been the source of the 1968 swarm.

[5] A stereographic projection plot resulting from an analysis of seismic waves as recorded on seismograph stations. A focal mechanism displays two possible orientations for the causative fault and associated slip directions and the pattern of the tectonic stresses initiating the earthquake rupture.

[6] A geologic structure where layered rocks have been folded or arched upwards by compressive forces. The Portland Hills may be an anticline.

1976 Deschutes Valley earthquake

On April 12, 1976, a M_L 4.8 earthquake shook an area of 35,000 km^2 centered near the town of Maupin in north-central Oregon (Figure 3a). Although the epicentral area had exhibited no prior seismicity, historical earthquakes were reported in adjacent areas such as The Dalles as early as 1866 (Figure 3a). Maximum intensities of MM V–VI were observed along the Deschutes River Valley, where houses were shaken, resulting in cracked plaster (Couch and others, 1976) (Figure 7). Some objects were thrown to the floor in Maupin, South Junction, and Warm Springs. Sounds described as "distant thunder, sonic booms, and strong wind" were also reported in this event.

An epicentral location was determined by Couch and others (1976) based on the P-wave readings from 48 seismograph stations in the western United States and Canada (Figure 7). The focal depth of the main shock was estimated to be 15 km. The main shock was preceded by nine foreshocks and followed by 13 aftershocks, of which the largest was a M_L 4.2 (Figure 3a). A composite focal mechanism of the main shock and other events in the sequence suggest that the source of the Deschutes Valley earthquakes was a west-northwest-striking reverse fault (Couch and others, 1976). The presence of such a fault is consistent with several similarly oriented anticlines in the epicentral area.

1993 Scotts Mills earthquake

On March 25, 1993, at 5:35 a.m., a strong earthquake (M_L 5.6, maximum MM VII intensity) struck western Oregon and Washington (Madin and others, 1993; Dewey and others, 1994). This event, which is the fourth largest earthquake to occur in western Oregon in historical times, was felt over an area of 97,000 km^2 (Dewey and others, 1994). The earthquake occurred 5 km east of the town of Scotts Mills (Figure 3b). Damage to property was estimated at \$28 million, mostly to older, unreinforced masonry structures such as the Molalla High School (Dewey and others, 1994). Numerous chimneys were damaged, and broken plaster and bricks were common. Despite the strength of this earthquake, only minor injuries were incurred.

The earthquake was recorded throughout the UW regional seismographic network. A focal depth of 15 km was determined for the main shock, although this value is poorly constrained. Aftershocks recorded by a portable network installed in the epicentral region within 12 hours of the main shock align along a northwest-striking, moderately north-northeast-dipping plane at depths of 8 to 15 km (Thomas and others, in preparation). The main-shock focal mechanism indicates oblique-reverse slip on a northwest-striking, northeast-dipping nodal plane in response to a north-south compressive stress. The earthquake locations and focal mechanism are all consistent with the sequence occurring on the Mount Angel fault (Thomas and others, in preparation). Werner and others (1992) earlier suggested that a sequence of six small earthquakes ($M_L \leq 2.5$) in August 1990 near Woodburn could have occurred on the northern end of the Mount Angel fault. Through 1994, a total of about 300 aftershocks had been recorded by either the UW regional or portable networks.

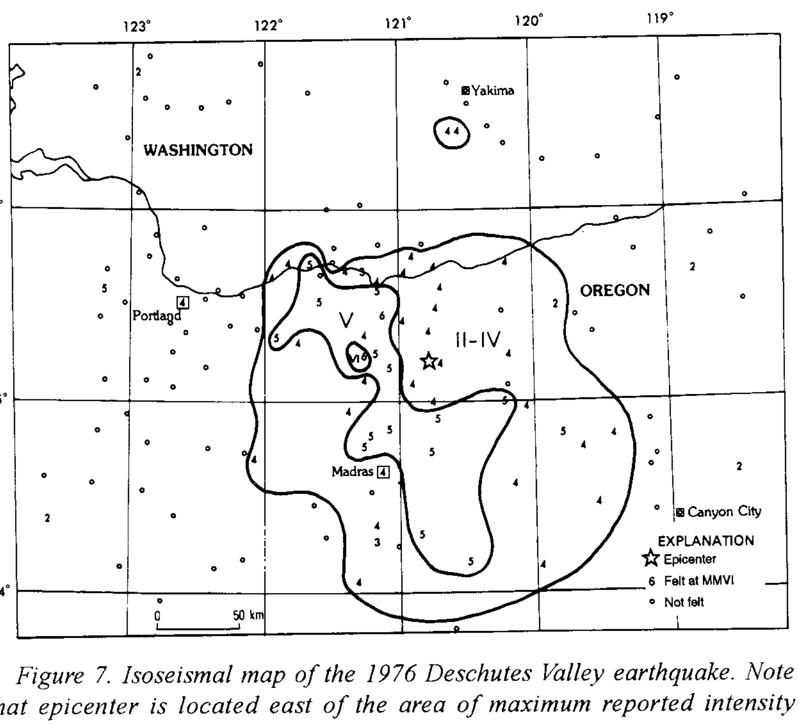

Figure 7. Isoseismal map of the 1976 Deschutes Valley earthquake. Note that epicenter is located east of the area of maximum reported intensity (MM VI). From Coffman and Stover (1978).

1993 Klamath Falls earthquakes

On the evening of the September 20, 1993, two moderate-sized earthquakes struck the Klamath Falls area (Figure 3b), causing two deaths and extensive damage. The casualties were the first reported to result from an earthquake in Oregon. The two largest events, M_L 5.9 at 8:28 p.m. and M_L 6.0 at 10:45 p.m., and the ensuing aftershocks were centered approximately 20 km northwest of Klamath Falls (Figure 8). (Braunmiller and others [1995] estimated M_W 6.0 for both events.) Focal depths of the events were generally less than 12 km (Braunmiller and others, 1995). A foreshock of M_L 3.9 occurred 13 minutes before the M_L 5.9 event. A portable network was deployed in the epicentral area by the USGS, OSU, and UO within days of the largest events.

Damage from the Klamath Falls earthquakes amounted to about \$7.5 million, mostly to various residential, commercial, and government buildings, including the Klamath County Courthouse buildings

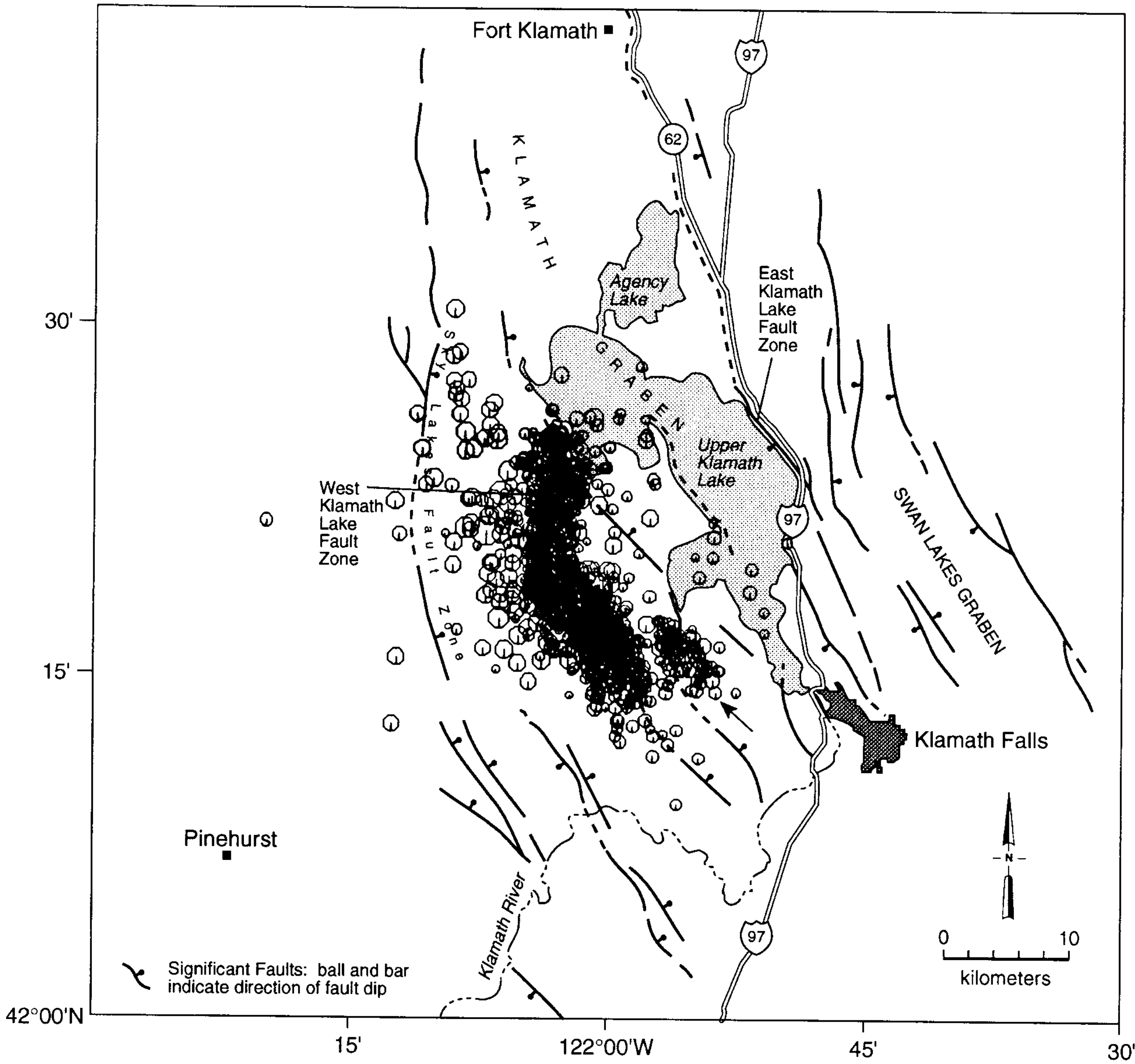

Figure 8. Epicentral map of the 1993 Klamath Falls sequence. Seismicity data courtesy of the University of Washington. Faults taken from Hawkins and others (1992).

(Wiley and others, 1993). Some unreinforced masonry buildings were severely damaged, while wood-frame buildings sustained little or no damage (Wiley and others, 1993). Modern buildings, including two reinforced concrete buildings on the Oregon Institute of Technology campus, also sustained considerable damage. Surface cracking within artificial fill along roads and landslides/rockfalls were induced by the earthquakes. No evidence for surface faulting was found.

The Klamath Falls area has experienced only low levels of historical seismicity, but geologic evidence shows late Quaternary (past 500,000 years) fault activity in the epicentral area (Hawkins and others, 1992) (Figure 8). The earthquake sequence was located between the fault bounding the western side of the Klamath graben and the Sky Lakes fault zone, both of which are east-dipping normal faults. Analyses to date indicate three source zones of seismicity: a northwest-trending zone that included the M_L 6.0 earthquake; a north-trending zone including the M_L 5.9 event; and a shallow northwest-trending zone to the east near Klamath Lake (Braunmiller and others, 1995).

A focal mechanism for the M_L 6.0 event exhibits north-northwest-striking planes; the preferred plane being the east-dipping normal fault with a small component of left-lateral motion (Wiley and others, 1993; Braunmiller and others, 1995). The focal mechanism is consistent with approximately northeast-southwest-oriented tectonic extension typical of the northern Basin and Range province. Based on several focal mechanisms and the aftershock distribution, Braunmiller and others (1995) suggested that the Lake of the Woods fault zone (part of the Sky Lakes fault zone; Figure 8) may have been the source of the two main shocks.

A relatively late aftershock (M_L 5.1) in the Klamath Falls sequence occurred on December 4, 1993, causing minor damage. It was the largest of more than 4,200 aftershocks that were recorded and located through 1994; it caused a parapet to fall onto an adjacent building on Main Street, resulting in the collapse of the roof of a new comic book store.

AREAS OF SIGNIFICANT SEISMICITY

Before 1979, the historical record shows seismicity, consisting principally of earthquakes of $M_L \geq 3$, occurring rather sparsely throughout Oregon. The few concentrations were centered on Portland, in the epicentral areas of the 1968 Adel and 1976 Deschutes Valley earthquakes, in the Pine Valley graben-Cuddy Mountain area east of Baker, and the area between Hermiston and Milton-Freewater, northwest and northeast of Pendleton (Figure 3a). Since 1979 and the southward expansion of the UW seismographic network, seismicity, particularly microearthquake activity, appears to be concentrated in northernmost Oregon and in the Pine Valley graben-Cuddy Mountain area (Figure 3b). To some degree, this pattern may be an artifact of station coverage (Figure 1).

A map of the seismic moment released in Oregon based on the 153-year historical record dramatically reveals the sites of the state's largest earthquakes as well as its most active areas (cover illustration). The 1873 earthquake, which may have occurred within the Cascadia subduction zone as an intraplate event, has dominated the seismic moment release in Oregon. Areas of significant crustal seismic moment release include, in order of decreasing moment, the epicentral areas of the 1993–1994 Klamath Falls and 1936 Milton-Freewater earthquakes and the Portland region (cover illustration).

The apparent low level of seismicity in most of southeastern Oregon may be largely due to the absence of seismograph stations (Figure 1). As previously stated, a region of notably few earthquakes has been the Cascadia subduction zone beneath western Oregon (Ludwin and others, 1991). This lack of seismicity is in contrast with the seismically active subducting Juan de Fuca plate beneath western Washington. The reason for this quiescence in Oregon is perplexing.

Portland region

As recognized by Couch and Lowell (1971) and Bott and Wong (1993), the Portland region (the 100-by-100-km area centered on Portland, including Vancouver) is the most seismically active area in Oregon (Figures 3a and 3b). At least 17 earthquakes of M_L 4 and larger and six events of M_L 5 and larger (including the 1877 and 1962 earthquakes) have occurred in the region in historic time. As noted earlier for the 1962 earthquake, Yelin and Patton (1991) explain seismicity in the area within the context of the Portland pull-apart basin. The basin is bounded on the southwest by the right-lateral, strike-slip Portland Hills fault zone and on the northeast by a postulated right-lateral, strike-slip Frontal fault zone (Yelin and Patton, 1991).

Since 1982, when seismographic coverage of the Portland basin became sufficient to detect events as small as M_L 1, the Portland Hills fault zone has been nearly aseismic (Yelin, 1992). However, a composite focal mechanism for four events that occurred at the south end of Sauvie Island between 1982 and 1985 exhibited predominantly strike-slip faulting on a northwest-striking plane consistent with the Portland Hills fault zone (Yelin and Patton, 1991). From July to October 1991, a small earthquake swarm of about 40 events (three of M_L 3.0 to 3.5) occurred at depths of 15 to 18 km at the northern end of the Portland Hills fault zone. Focal mechanisms for two of the largest earthquakes exhibited mixed reverse and right-lateral strike-slip faulting along a plane that coincides with the postulated Portland Hills fault, similar to what is observed elsewhere in the Portland basin (Yelin, 1992). A composite mechanism for several small events, 4 km south of Battleground, Washington, exhibited oblique-strike-slip faulting on a near-vertical, northwest-striking plane possibly associated with the Frontal fault zone (Yelin and Patton, 1991). In none of these cases, however, is the evidence definitive that the Portland Hills or the Frontal fault zones are seismically active because of the uncertainties in associating the relatively deep crustal seismicity in the Portland area with these structures whose depth extent is unknown (Blakely and others, 1995).

Mount Hood

Seismicity has occurred at Mount Hood, a Cascade volcano east of Portland (Figure 3b). The largest known earthquake was a M_L 4.0 event that occurred on December 13, 1974. Based on a 16-station temporary network operated at Mount Hood from November 1977 to December 1978, a total of 10 earthquakes were recorded and located, with the largest event reaching approximately M_L 3.4 (Weaver and others, 1982). All events occurred above a depth of 15 km. Focal mechanisms for five of six events exhibited predominantly right-lateral, strike-slip faulting on a north-northwest-striking plane (Weaver and others, 1982). Weaver and others (1990) suggested that some of this activity and earthquakes in 1989 and 1990 may be associated with a north-northwest-striking seismic zone beneath Mount Hood, similar to one under Mount St. Helens in Washington. The 90-km-long St. Helens seismic zone has been one of the most seismically active areas in the Pacific Northwest in historical times (Weaver and Smith, 1983; Ludwin and others, 1991). Geomatrix Consultants (1990) suggested that the Mount Hood seismic zone may coincide with the Mount Hood fault.

Pine Valley graben-Cuddy Mountain

The most active area in eastern Oregon appears to be the Pine Valley graben-Cuddy Mountain area along the Oregon-Idaho border (Figures 3a and 3b). The first recorded earthquake in the area occurred at 3:00 p.m. on October 14, 1913 (Figure 3a). The earthquake was assigned a maximum intensity of MM VI and was felt most severely in Landore, Idaho, where windows broke, furniture rocked, and dishes were thrown from shelves. Zollweg (BSU, personal communication, 1992) estimates the size of the event was M_L 4¾, based on a review of historical seismograms.

A number of additional small earthquakes (MM IV–V) have occurred in the Pine Valley graben area (Mann, 1989) (Figures 3a and 3b). The abundant microseismicity shown on Figure 3a is the result of an analysis of events recorded at the BMO array from 1962 to 1967 (Zollweg and Wood, 1993). None of these events can be definitively associated with any mapped faults.

Zollweg and Jacobson (1986) operated a portable five-station microearthquake network, which recorded 15 aftershocks of two M_L 3.8 earthquakes that occurred on August 10 and September 19, 1984, in the Cuddy Mountain area (Figure 3b). A composite focal mechanism exhibited normal faulting on north- to northwest-striking planes, suggesting that the area is being subjected to Basin and Range-like extensional stresses (Zollweg and Jacobson, 1986). Mann and Meyer (1993) suggested that the seismicity in the area is associated with a portion of the Olympic-Wallowa lineament which includes the Pine Valley graben and Brownlee fault. Conversely, we speculate that the Pine Valley graben-Cuddy Mountain area may represent the westernmost extent of the east-west-trending Centennial Tectonic Belt (Stickney and Bartholomew, 1987).

EARTHQUAKES AND ACTIVE FAULTS

All earthquakes of tectonic origin, no matter their size, are the result of sudden displacement on a fault. The larger the fault area that is displaced or ruptured, the larger the event. For example, a M_W 7 earthquake will typically rupture a fault or portion of a fault that is about 1,000 km^2 in area, such as a fault 50 km long and 20 km wide.

Few late Quaternary crustal faults have been identified in Oregon, particularly in the western half of the state (Pezzopane and Weldon, 1993). The dense vegetation and rapid erosion rates make it difficult to find evidence of young faulting in western Oregon. Active faults may also be more deeply seated west of the Cascades because of a thicker seismogenic crust (Wong and others, 1994). Thus they would not be as well expressed at the earth's surface as in other regions in the western United States (e.g., Basin and Range province) where the seismogenic crust is on the order of 15 km thick. To many, the 1993 Scotts Mills earthquake was an unexpected event for western Oregon, possibly because it occurred on a "blind" or hidden fault. Earthquakes of similar or larger magnitude, however, will likely occur on other blind structures elsewhere in this half of the state.

In eastern Oregon, like much of the western United States, late Quaternary faults are more prevalent or more visible at the surface, although few have been studied in detail. Faults that have been investigated include, for example, two Basin and Range-like structures: the Alvord fault along Steens Mountain (Hemphill-Haley and others, 1993) and the Goose Lake graben faults near Lakeview (Pezzopane and Weldon, 1993).

Because there have been so few large historical earthquakes in Oregon and seismic monitoring has been generally sparse, only a small number of events has been associated with known active faults in the state. These include possibly the 1936 Milton-Freewater, 1993 Scotts Mills, and 1993–1994 Klamath Falls earthquakes, which may have been associated with the Wallula or Hite faults, Mount Angel fault, and Lake of the Woods fault zone, respectively. As previously stated, the Portland Hills and Frontal fault zones may also have associated seismicity (Blakely and others, 1995).

Because of the incomplete historical record for the state, our understanding of the earthquake potential for Oregon will probably be quantified only through future paleoseismic fault studies and continued earthquake monitoring. Unfortunately for much of western Oregon, this quantification may never be complete because of its few known faults.

CONCLUSIONS

Although Oregon is not generally thought of as being "earthquake country," especially compared to the neighboring states to the north and south, the historical record clearly indicates that the state faces a level of earthquake hazard that requires further quantification. Realistic estimates of the state's earthquake potential must rely heavily on future geologic studies. Crustal earthquakes as large as M_L 6 and possibly intraplate earthquakes up to M_L 6¾ within the Cascadia subduction zone have occurred in Oregon according to the observations of the past 153 years and will undoubtedly occur in the future. Additionally, paleoseismic studies along the Oregon coast indicate that the state has been shaken in the past by great Cascadia subduction zone megathrust earthquakes, possibly as large as M_W 9. Paleoseismic investigations along late Quaternary faults also indicate that events as large as M_L 7 have occurred repeatedly in several locations in eastern Oregon (e.g., Hemphill-Haley and others, 1993; Pezzopane and Weldon, 1993).

Although very few detailed paleoseismic studies have been performed on the few known late Quaternary faults in western Oregon, circumstantial evidence suggests that crustal earthquakes as large as M_L 7 are possible in some areas, such as near Portland (Wong and others, 1994). Such events may pose the greatest hazard to the urban areas in the Willamette Valley and eastern Oregon because of the severe ground shaking that could result from such relatively nearby earthquakes, as compared to larger earthquakes which may occur at greater distances within the Cascadia subduction zone (e.g., Weaver and Shedlock, 1991; Wong and others, 1993).

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